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GEOLOGY OF THE UPPER SEGRE AND VALIRA VALLEYS, CENTRAL PYRENEES, ANDORRA / SPAIN

BY

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ABSTRACT

Sheet 10 of the 1:50,000 geological map series of the Geological Institute, Leiden University, is presented accompanied by a survey of structural, stratigraphic and morphologic features. The region includes part of the southern Axial Zone, the eastern end of the Nogueras Zone and part of the marginal trough adjacent to the south.

Formations mapped range in age from Cambro-Ordovician to Pliocene. Detailed lithostratigraphic studies of the Cambro-Ordovician and Devonian rocks enabled correlations to be made with other regions of the Pyrenees.

The Hercynian orogeny formed structures of variable shapes and orientation, which, however, have all be ascribed to stress accumulating in a N-S direction causing consecutive deformations. These can, to a certain extent, be correlated with the deformation phases known in other regions of the Axial Zone to the north. A first phase formed the largest folds of an order of 20 km down to less than 1 km wide. The slaty cleavage is generally parallel to the bedding and may possibly be related to this phase, in which case it is to be interpreted as a "concentric cleavage". Second phase structures, generally characterized by an axial plane crenulation cleavage, indicate further compression. Temporary relief of the main stress after the first and second phases thought to be a result of accelerated strain rates due to folding, gave rise to cross folds and cascade folds. Gravity sliding of unstable structures in Devonian rocks from the Rabassa dome into the Arcabell syncline and Segre area, resulted in thick accumulations of Devonian material in these depressions. Strong deviations from the general E-W strike of first and second phase structures in the eastern part of the Orri dome and in the Segre unit are attributed to the influence of an inferred infrastructure rise thought to exist below the Orri dome.

Epirogenic movements along the northern border of the marginal trough, were particularly important during the Permian and Cretaceous. The Devonian massif of the Monsech de Tost, amidst post-Hercynian deposits in the Nogueras Zone, is interpreted as a Hercynian gravity structure upthrusted during the Pyrenean (Middle Alpine) deformation. The Pedraforca structure is also explained as an upthrusted block. Later Alpine movements during the Neogene formed grabens in the Segre valley.

CONTENTS

INTRODUCTION

The present report describes the geology of a part of the southern axial zone and Nogueras zone (Fig. 1) in the Spanish and Andorran Pyrenees. The publication is one of a series dealing with the structure and stratigraphy of the Central Pyrenees, which have been subject of investigation for the Department of Structural Geology of Leiden University under direction of Prof. Dr. L. U. de Sitter and Prof. Dr. H. J. Zwart. The field work was carried out during the summers of 1961–1967.

Topographic maps
For the Spanish part of the map area use was made of the new 1:25,000 topographic maps of the Servicio Geográfico del Ejército including the sheets Tirvia (182) II, III, Seo de Urgel (215) I–IV, Bellver (216) I–IV and Gosól (254) I–IV. The old 1:50,000 map of Orgaña (253), covering the southwest corner of the area, was inadequate for geological mapping and for this zone a new topographic map has been drawn from aerial photographs and partly from a photogrammetric map made by von Glasenapp (1967) and corrected with field observations by the author. For the Andorran part of the area also a new map has been drawn from aerial photographs as the existing topographic maps of Chevalier, Kucera and Llopis Lladó are inadequate for detailed geological mapping.

Aerial photographs
Aerial photographs have been available for Andorra the runs of Hospitalet-Andorre, 1948, nr. 113–194 and for Spain from north to south the runs 589, 170, 604, 268, 550, 539, 539b, 611, 596, 554, 597 and 541.

Geological maps
Relatively detailed geological maps of parts of the area have been consulted, namely those of Boissevain (1934) Upper Segre valley 1:40,000, Solé Sabaris & Lopis Lladó (1946) Sheet Bellver (216) 1:50,000, Solé Sabaris & Llopis Lladó (1947) Andorra 1:50,000 and Guérin-Desjardins & Latreille (1961) Segre-Llobregat 1:100,000. Recently, as this report was going to press, a series of 1:25,000 sheets was published posthumously of a geological map of Andorra made by Llopis Lladó (1970). This map is very divergent from the map presented by the present author. Use has

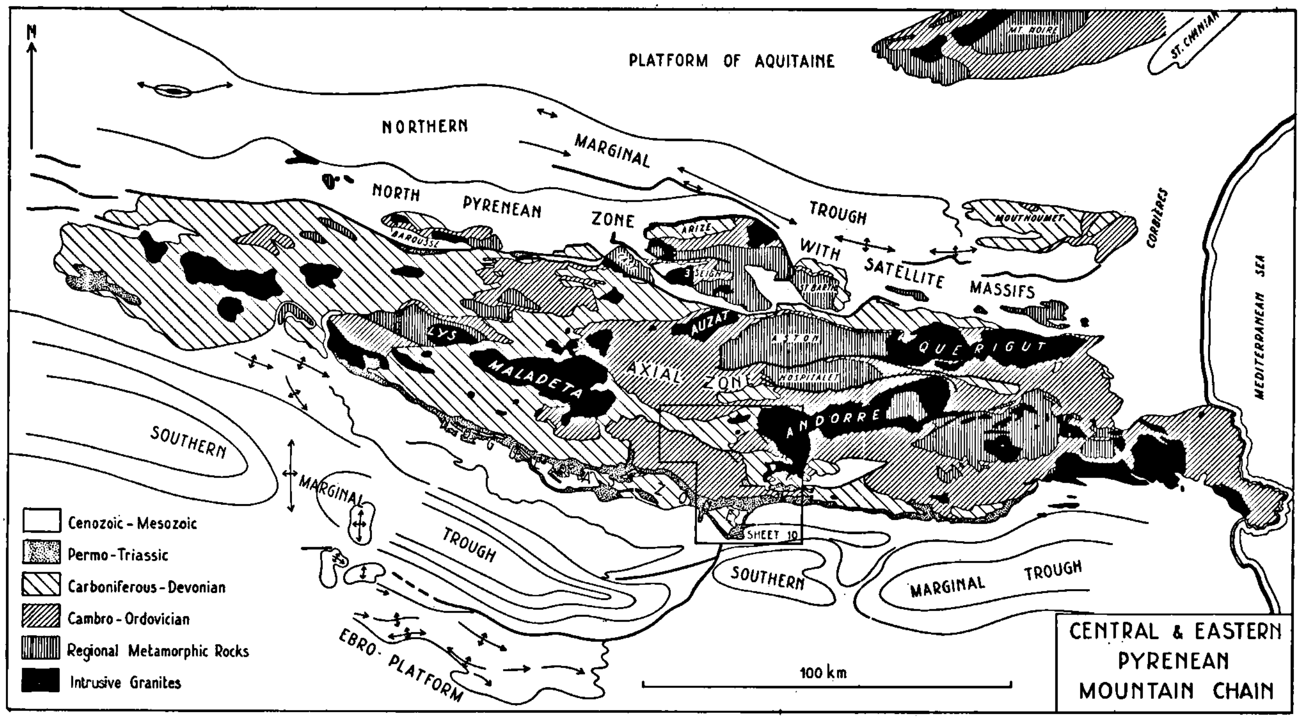

Fig. 1. Geological sketch-map of the central and eastern Pyrenees (slightly modified after de Sitter, 1964), with location of map area.

also been made of geological maps in internal reports, all in Dutch, held at the Geological Institute in Leiden, of Dessauvagie (1960), Diederix (1963) and van Wees (1970).

Acknowledgements

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CHAPTER I

STRATIGRAPHY

INTRODUCTION

The geological mapping of this area is based mainly on a lithostratigraphic subdivision in formations. In the past, however, time-stratigraphic boundaries have often been provisionally placed at formation boundaries if palaeontological evidence to the contrary was lacking e.g. the boundaries between Ordovician and Silurian and those between Carboniferous, Permian, Triassic and Jurassic. Although this is obviously incorrect such subdivisions have here been retained in order to facilitate reference to other work. However, where positive biostratigraphic evidence has rendered earlier usage untenable the new time-stratigraphic boundaries have been indicated.

PRE-HERCYNIAN FORMATIONS

A very low grade of metamorphism of the greenschist facies, displaying minerals like sericite, chlorite, chloritoid, quartz and calcite is developed in pre-Hercynian formations. Folding and cleavage development have often obscured sedimentary structures and the original thickness of beds.

CAMBRO-ORDOVICIAN

In the Cambro-Ordovician of this area several formations can be distinguished (Hartevelt 1965). The wide extent of these formations is rather exceptional since in most other areas of the Pyrenees the equivalents of these formations can only be traced over relatively short distances in a monotonous sequence of slates. Moreover many more fossils have been found in this area than in other comparable areas of the Pyrenees.

To establish a correlation with the stratigraphy of the surrounding areas a number of sections have been studied outside the map area, such as those along the north flank of the Massana anticline and in the Marimaña area (Hartevelt in press). All sections studied by the author and those copied from other authors or compiled from their data are gathered in Appendix II. Correlations have been facilitated by the widespread occurrence of two distinctive marker horizons, viz. the coarse Rabassa Conglomerate and the fossiliferous marls of the Estana Formation.

The stratigraphy presented for this area has been developed from data obtained from a schematic section by Schmidt (1931) of the rock sequence along the Segre valley east of Seo de Urgel and from descriptions by Boissevain (1934) of the Ordovician south of the Rio Segre. Brouwer (1968b) gave a detailed sedimentological description of a section of the upper Cambro-Ordovician near Talltendre, just east of the map area. His interpretation has been incorporated in the conclusions concerning depositional environment.

Seo Formation

The Seo Formation, named after the town of Seo de Urgel, is the lowest rock unit of the Cambro-Ordovician and crops out in many large anticlinal structures covering wide areas of the Pyrenees. The formation is defined as the rock sequence underlying the coarse Rabassa Conglomerate and comprises a wide diversity of rocks among which slates are most common and in which arenitic sediments and limestones can locally be abundant. A lower boundary of this formation is postulated by Autran, Fonteilles & Guitard (1966), who interpreted the gneisses in the Pyrenees as a pre-Cambrian basement below these rocks.

Descriptions have been given by Cavet (1957), Clin (1959), Kleinsmiede (1960), Zandvliet (1960), de Sitter & Zwart (1962), Zwart (1965), Mirouse (1966), Brouwer (1968a), Bloemraad (1969) and Trouw (1969).

Lithology. – In this area the Seo Formation consists almost completely of a monotonous alternation of thin slate, silt and sand layers, generally varying in thickness from 1 mm to several cms (Fig. 2), whereas quartzitic layers up to 10 cm may be observed locally. These so-called "schistes rubanés" have a grey to light green colour. Quartzitic layers may show cross-bedding and load-casts. Graded bedding may sometimes be observed, but contacts between quartzitic and slate layers as a rule remain sharp.

Fig. 2. Typical slate-siltstone alternation of Seo Formation.

North of the Llavorsi syncline, between the Saloria and Syspony, light-grey quartzite beds ranging in thickness from 1–30 m occur in the monotonous slate series (Appendix II, Sections 18–20). To the east as well as to the west these quartzites wedge out. Locally a thin layer of quartz conglomerate has been observed with well-rounded pebbles up to 1 cm in diameter.

Fossils. – Trilobites and ostracods have been recorded in other areas of the Pyrenees by Roussel (1893, quoted by Zandvliet, 1960) and Schmidt (1931), but recent studies have not confirmed these observations and this formation is generally described as non fossiliferous.

Correlation. – The Seo Formation being defined by its position below the Rabassa Conglomerate can easily be correlated in all sections containing these (Appendix II).

The eastern Pyrenees differ from the present area as there the rock sequence consists of a lower "Série de Canaveilles" characterized by intercalations of limestone beds and black carbonaceous slates in a thick slate sequence and an upper monotonous slate sequence, the "schistes de Jujols", about 2000 m thick (Cavet, 1957).

Zandvliet (1960) divided the comparable rock sequence, in the Salat-Pallaresa anticlinorium, into two units viz. the "Lleret-Bayau series" to be correlated with the Série de Canaveilles and the "Pilas-Estats series" with the Schistes de Jujols.

Zwart (1965) found that the continuation of the Lleret-Bayau series in Andorra occur less than 400 m below the conglomerates (App. II, section 12) and considered a correlation with the very much deeper buried Série de Canaveilles further east unlikely and introduced the Ransol Formation containing quartzites, limestones and carbonaceous slates. Quartzite beds observed in the Massana anticline (App. II, sections 8 and 18–20) can most probably be correlated with the Ransol Quartzites, occurring as they do at a comparable depth below the conglomerates. Because Zwart's Ransol Formation only comprises the limestone, carbonaceous slate and quartzite intercalations in the Seo Formation, it is preferred to rank this unit as a member.

Kleinsmiede (1960) and de Sitter & Zwart (1962) described, in the eastern part of the Garonne dome (Northern anticline), a thick limestone unit known as the "Calcaire métallifère" 50–200 m below the coarse conglomerate level (App. II, Sections 2, 3 and 4). Zandvliet (1960), who mapped the eastern continuation of this same "Calcaire métallifère", correlated it with the limestones of the Marimaña area, which occur above a coarse conglomerate level. According to the present author, however, the limestones of the Marimaña area are most probably Devonian and do not represent the "Calcaire métallifère", which is not developed in this more southerly area (Hartevelt in press). Zandvliet's sections of the Moredo and Bonaigua have therefore been modified as shown in Sections 5 and 6 in Appendix II.

More recent investigations by Bloemraad (1969) and Trouw (1969) in the area between Cerdaña and Camprodon showed that limestones which have been attributed to the Série de Canaveilles occur only 200 m below the coarse conglomerate (App. II, Sections 30 and 31) and are not to be correlated with the Série de Canaveilles of Cavet.

Summarizing, one may conclude that the Seo Formation in the Central Pyrenees is composed of a thick sequence of slates which may locally display a more quartzitic development and in which thick quartzite beds may occur, viz. the Ransol Member. Limestone members may be present at varying levels: the Calcaire métallifère and the limestones between Cerdaña and Camprodon at a relatively short distance, the Ransol Member at an intermediate distance and the limestones of the Série de Canaveilles at a large distance below the conglomerate.

Due to the presence of limestones at various levels, the Seo Formation should not be subdivided into one unit with limestones and one without, a view recently also shared by Cavet (pers. comm.).

Age. – Cavet (1957) made a lithological correlation with the fossiliferous Cambro-Ordovician of the Montagne Noir and concluded with some reserve that the lower part (Série de Canaveilles) can best be attributed to the Cambrian while the higher part (Schistes de Jujols) possibly represents the lower Ordovician.

Depositional environment. – Mirouse (1966, p. 34) suggests a marine environment in which a rhythmic variation of influx of argillaceous and quartzitic material may be due to variations in seasons and climate or to epirogenic pulsations.

Rabassa Conglomerate

The Rabassa Conglomerate, named after the Riu de la Rabassa in southern Andorra, outcrops continuously throughout almost the entire map area and can also be observed in many other areas of the Pyrenees, the Mouthoumet massif and the Montagne Noir.

Descriptions have been given by Cavet (1957), Clin (1959), Kleinsmiede (1960), Zandvliet (1960), de Sitter & Zwart (1962), Zwart (1965), Mirouse (1966), Brouwer (1968b), Bloemraad (1969), Trouw (1969) and Hartevelt (in press).

Lithology. – The Rabassa Conglomerate conformably overlies the Seo Formation. The unconformity recorded by Schmidt (1931) in the southern part of the Orri dome, proved to be the unconformity below the Bunter conglomerate. The unconformity described by Llopis Lladó (1965) in Andorra was not found by the present author; only some very local erosional channels were observed.

The conglomerate is composed of quartz, quartzite and slate pebbles in a purple to greenish matrix of slate, sandy slate or siltstone (Fig. 3). Pebbles of black metamorphic schist only occur in the western part of the area (Fig. 4-B). The pebbles in general possess a random orientation and often do not touch each other.

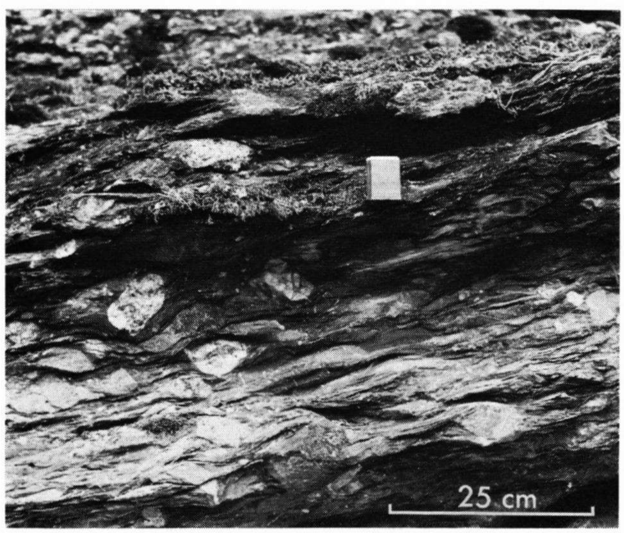


Fig. 3. Rabassa Conglomerate, north of Burch. Note the refraction of the cleavage in the slates around the pebbles.

The size of the pebbles varies greatly, quartz and quartzite pebbles often attaining diameters of 20 cm and quartzite blocks with a length of 50 cm having occasionally been observed. As can be seen in Fig. 4-B, the size distribution of the pebbles over the area is quite irregular, but very fine conglomerate is only developed north of the Llavorsi syncline where the formation is even absent in places. The general

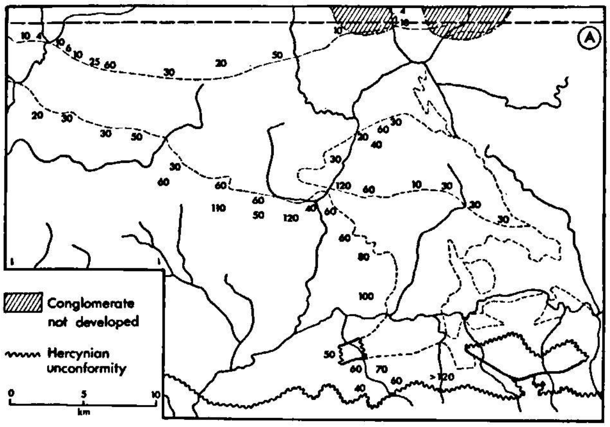


Fig. 4A. Thickness distribution of the Rabassa Conglomerate (metres).

wedging out of the conglomerate to the north is in accordance with the occurrence of this formation in lenses in the Hospitalet massif (Zwart, 1965) and the very poor development in the north flank of the Salat-Pallaresa anticlinorium (Zandvliet, 1960).

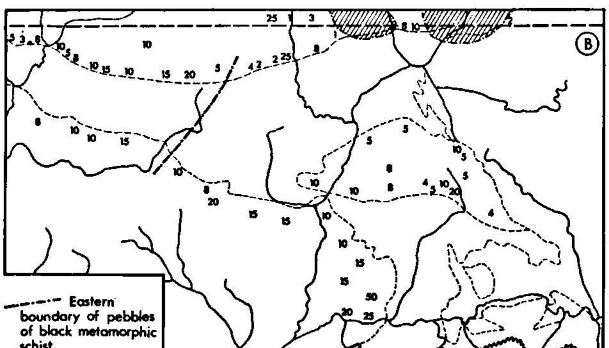


Fig. 4B. Distribution of maximum pebble size of the Rabassa Conglomerate (centimetres).

Thickness. – A maximum thickness of more than 100 m, observed in the Segre valley, generally decreases towards the north (Fig. 4-A).

Correlation. – In most sections of the Pyrenees only one level of coarse conglomerate is to be seen (Appendix II).

Depositional environment. – The occurrence of pebbles of material often unknown in the Seo Formation indicates that the conglomerates must have been transported over long distances from a source area outside the axial zone of the Pyrenees. Large pebbles floating in slate have often been observed in certain zones; such pebbly mudstones were probably deposited from mud-flows. The general decrease in the pebble size and the wedging out of the formation towards the north indicate that the conglomerate originated in the south, a view already expressed by Schmidt (1931). N-S directed channels found by Brouwer (1968b) are also in accordance with a transport from south to north.

Cavá Formation

The Cavá Formation, named after the village of Cavá, is defined as a mainly coarse detritic, partly fossiliferous, rock unit overlying the Rabassa Conglomerates and overlain by either the Estana Formation or the Ansobell Formation. The formation is more or less restricted to the southern part of the Pyrenees.

Accounts of these rocks are given by Cavet (1957), Bloemraad (1969) and Trouw (1969). Brouwer (1968b) gives a very detailed sedimentological description.

Lithology. – The lower boundary of the Cavá Formation is drawn below the first coarse sandstone beds and above the last large pebbles of the Rabassa Conglomerate. The formation can be divided into four members some of which have only locally developed. These members from bottom to top, are: greywackes, red and green slates, siltstones and purple quartzite.

The greywacke member consists of a thick succession of mainly coarse detritic rocks comprising greenish to purple greywackes, sandstones, microconglomerates, siltstones and slates. In the south this member attains a thickness of more than 750 m (App. II, Section 37) and gradually wedges out towards the north. Coarse-grained layers contain mainly detritic grains of rock fragments (schist and quartzite), quartz, chert and feldspar (plagioclase) in a slate matrix. The grains are badly sorted, ranging from a silt fraction to a diameter of 1 cm, and are also badly rounded. The relatively high feldspar content, constituting 15–20% of the detritic grains, suggests that these sediments were partly derived from tuffs. True tuffs as recorded by Schmidt (1931) in the Segre valley have, however, never been found. The sandstone beds generally show a fining upwards, the rock sequence as a whole also becoming more finegrained towards the top. Beds generally range in thickness from very thin to 50 cm, but larger units of sandstone and slate up to 10 m thick may also occur, such as east of Bar and Estana.

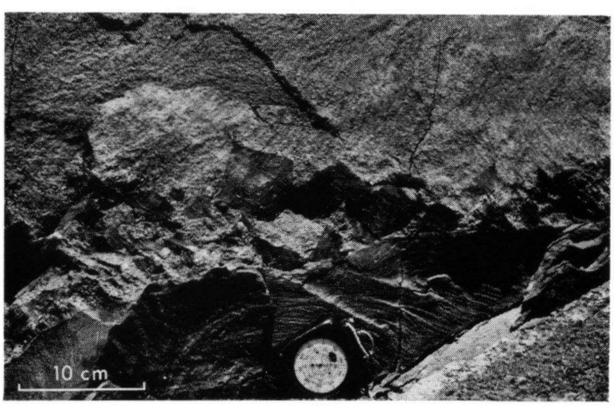


Fig. 5. Clay fragments in a channel in the Cavá Formation, southeast of Cavá.

Bedding is often very irregular, especially near the bottom where beds frequently wedge out and channels and cross-bedding are often observed. Clay fragments, often still angular, frequently occur embedded in coarse material (Fig. 5). Loadcasts are very frequently observed (Fig. 6).

Fig. 6. Load-casts in the Cavá Formation, southwest of Ansobell.

Reddish and greenish silty slates developed locally as a distinct level on top of the coarse greywackes and sandstones of the lower member west of Queforadat and in the eastern part of the Orri dome (App. II, Sections 32–35). The thickness of this member ranges from 35 to 90 m. Locally these slates contain small lenses of coarse sand and microconglomerate. In Andorra this level has only been observed east of Sant Julia (App. II, Section 27). East of Bar and Estana (App. II, Sections 36, 37) no distinct level of red-green slates can be found, but only thin beds of such slates intercalated in the greywacke member.

The siltstone member, which also contains some sandstone and slate, rests conformably either on the coarse greywackes or on the red and green slates. The siltstone member has a typical dull brownish colour. The thickness ranges from 40 to 100 m. The sand and silt fraction consists mainly of quartz and some chert grains, feldspar and rock fragments being almost absent. The grains are subrounded to rounded, gradually decreasing in size towards the top of the member.

The most conspicuous feature of the siltstone member is the occurrence of the first fossils, mainly brachiopods. In the lower part of the member only a few scattered fossils are to be found, but towards the top they increase in number until beds, up to 10 cm thick, almost constituting a coquina, can be observed (Fig. 7). Most fossils are completely decalcified, occurring only as casts.

Bedding is generally parallel and only small flaser and linzen structures occur. Rhythmic graded bedding is frequent although burrows often occur and the bedding is partly destroyed. Loadcasts of small dimensions occur as well.

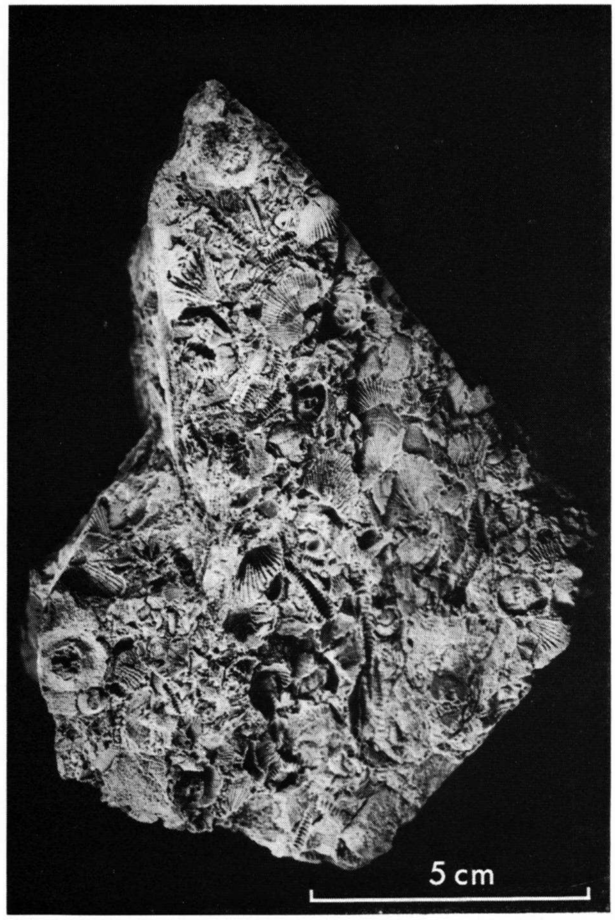


Fig. 7. Fossils seen on a bedding plane in the Cavá Formation, northeast of Seo de Urgel.

The purple quartzite member is only locally developed (App. II, Sections 32, 36), occurring between the siltstone member and the marls of the Estana Formation. The thickness of this member does not exceed 12 m. A very distinct cross-bedding is frequently developed.

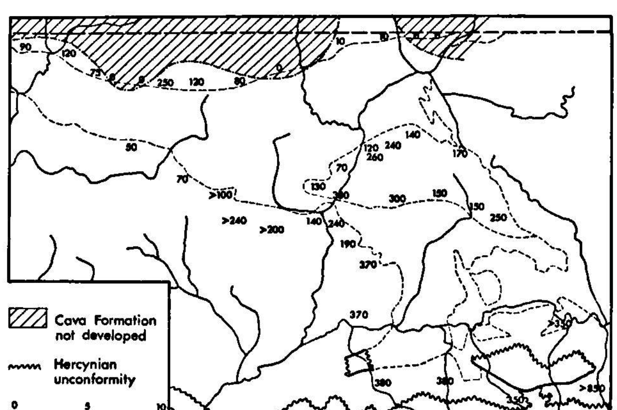


Fig. 8. Thickness distribution of the Cavá Formation (m).

Thickness. – The Cavá Formation is very thick in the south (over 850 m north of Bestanis) and gradually wedges out towards the north where it is replaced by the black slates of the Ansobell Formation (Fig. 8 and 9).

Fossils and age. – Fossils collected from various localities in the siltstone member of the Cavá Formation were kindly determined by Professor Dr. N. Spjeldnaes (University of Aarhus, Denmark), who dated this level as high Caradoc. The following determinations could be made:
brachiopods: *Svobodaina* sp., *Rostricellula* sp., *Rafinesquina* sp. and further tentaculitids, cystoids (plates and stems) and bryozoans are present.

Earlier authors also collected many fossils in this level and extensive lists are given by Schmidt (1931, p. 33), Boissevain (1934, p. 41–42) and Solé Sabaris & Llopis Lladó (1946).

Correlation. – As the Cavá Formation grades, towards the north, into the black slates of the Ansobell Formation, it can be correlated with the lower part of this formation below the Estana Formation (Fig. 9). To the west the formation also grades into slates (App. II, Section 23), to the east it can be followed at least as far as Camprodon (App. II, Section 31). A correlation of the greywacke member with Cavet's tuffaceous "Schistes de Jujols supérieurs" (App. II, Section 14) seems probable because of its pyroclastic elements. The fossiliferous siltstone member should probably be correlated with the lower part of Cavet's "Schistes troués" or "Grauwacke à Orthis".

Depositional environment. – From the occurrence of mudcracks in the greywacke member near Talltendre (App. II, Section 29) and from the channels and clay fragments frequently observed in this member, Brouwer (1968b) concluded a fluviatile deposition on a flood plain. The fluviatile influence would have gradually decreased towards the top of the formation and have been replaced by a shallow marine environment and deposition of the siltstone member on a tidal flat as may be concluded from the occurrence of marine fossils, the flaser and linzen structures and the rhythmically graded bedding. Since the Cavá Formation grades, towards the north, into the marine slates of the Ansobell Formation, it seems logical to assume a source area in the south as was also concluded for the Rabassa Conglomerate. Important basin subsidence must have occurred in the southeastern part of the area, in view of the considerable thickness of the formation in that region.

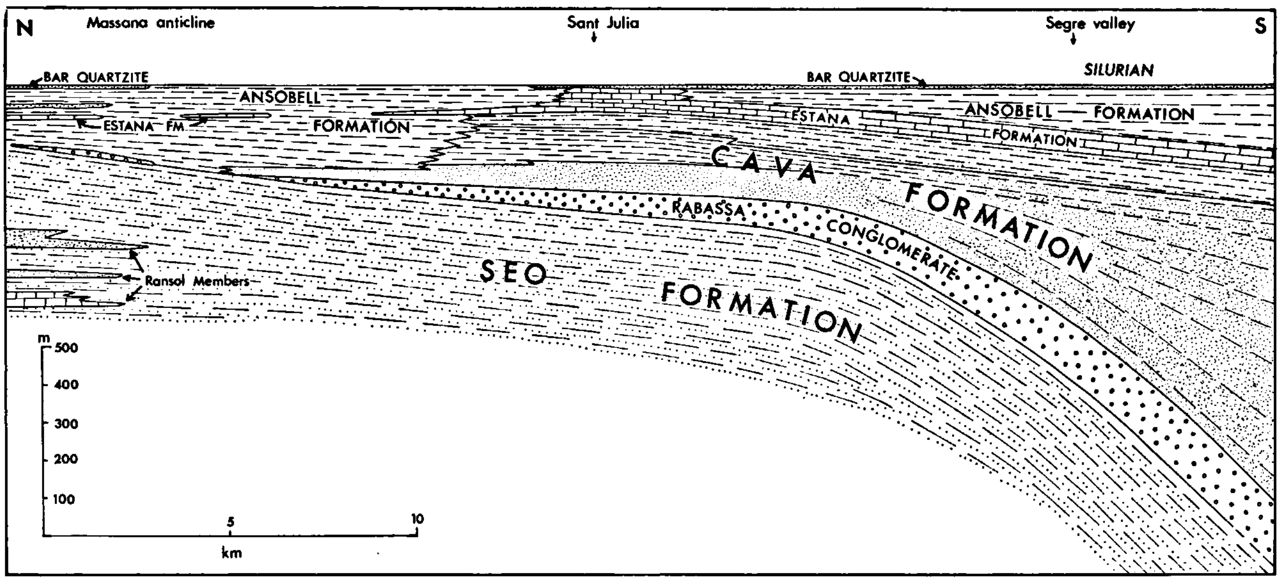


Fig. 9. Schematic N-S cross-section through Cambro-Ordovician formations.

Estana Formation

The Estana Formation, named after the village of Estana, is defined as the marl and limestone unit occurring in the upper part of the Cambro-Ordovician sequence somewhere above the Rabassa Conglomerate. In the southern part of the map area the formation can be more precisely defined as a marl and limestone unit situated between the Cavá and Ansobell Formation. The formation, often known as "the Caradoc limestone", is widespread throughout the Pyrenees, though not always deposited continuously.

Recent publications are: Cavet (1957), Clin (1959), Kleinsmiede (1960), Zandvliet (1960), de Sitter & Zwart (1962), Zwart (1965), Mirouse (1966), Brouwer (1968a, b), Bloemraad (1969) and Trouw (1969).

Lithology. – The lower boundary can be sharp if the formation overlies the purple uppermost quartzite of the Cavá Formation, it has, however, more often to be drawn between non-calcareous and slightly calcareous slates and is largely arbitrary. The lime content generally increases towards the top and locally pure light-grey limestones are developed, formed from a carbonate mud with fossil fragments. Fossils occur scattered throughout the formation and can locally be abundant (Fig. 10). The weathered surfaces of the limestones often have a red-brown appearance.

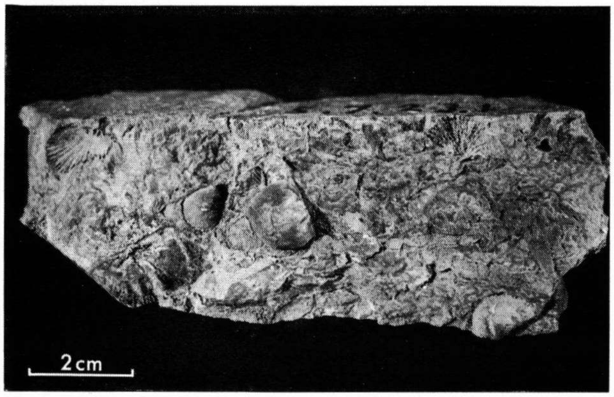


Fig. 10. Fossils seen on a bedding plane in the Estana Formation, south of Musa.

The upper boundary is almost exclusively formed by the contact with the black slates of the Ansobell Formation. Only in the area of Sant Julia and north of Argolell is the formation directly overlain by the Bar Quartzite (App. II, Section 27). Here intercalations of black slate a few cm thick occur within the limestones.

Thickness. – In the map area there is a general decrease in thickness from the south (App. II, Section 37, 60 m) towards the north where the formation is often absent. Locally, however, as near Sant Julia, the thickness again increases (Appendix II and Fig. 9).

A very thick development has been observed by Trouw (1969) in the Fresser Valley (App. II, Section 38, at least 200 m) and by Zandvliet (1960) in the north flank of the Salat-Pallaresa anticlinorium.

Fossils and age. – Macrofossils collected from various localities in the Estana Formation were kindly determinated by Professor Dr. N. Spjeldnaes ,(p. 00) who arrived at an age ranging from topmost Caradoc to (low) Ashgill. The following determinations could be made:
brachiopods: *Sampo* sp., *Hedstroemina* sp., *Nicolella* sp., aff. *Glyptorthis* sp., aff. *Harknessella*
bryozoans: *Chasmatoporella* sp., *Graptodichtya* sp., *Hallopora* sp., *Homotrypa* sp., *Rhinopora* sp.?
cystids: aff. *Pleurocystites* sp.
further rugose corals, gastropods, pelecypods and trilobites are present.

Conodonts, also found in this formation, were kindly determinated by Mr. K. T. Boersma (University of Leiden), who also arrived at an upper Caradoc-lower Ashgill age. The following conodonts were found: *Oistodus excelsus*, *Oistodus* sp., *Amballodus triangularis*, *Panderodus gracilis*, *Amorphognathus ordovicica*, *Amorphognathus* sp., *Drepandus* sp., *Icrodella superba*, *Plectodina* sp., *Acodus* sp.

Extensive lists of other fossils from this formation are given by Schmidt (1931, p. 34), Boissevain (1934, p. 41–42) and Solé Sabaris & Llopis Lladó (1946).

Correlation. – The correlation of this widely occurring formation in general poses no problems and is shown in Appendix II. To the northeast the Estana Formation should be correlated with Cavet's 'Schistes troués" or "Grauwacke à Orthis" which contains the same fauna as the Estana Formation and the siltstone member of the Cavá Formation together.

Depositional environment. – The fossiliferous limestones, indicating a limited influx of detrital material, must have been deposited in a more or less sheltered shallow marine environment (Brouwer 1968b).

Ansobell Formation

The Ansobell Formation, named after the village of Ansobell, has developed all over the Pyrenees.

Lithology. – In the northern part of the map area the Ansobell Formation often rests upon the Rabassa Conglomerate, but more to the south it is restricted to a level between the Estana Formation and the Bar Quartzite.

The black slates of this formation are practically homogeneous and the bedding is generally very vague, a few sandy layers being visible only near the top and bottom. Locally small sandy lenses have developed. In the transition zone to the Cavá Formation, north of the Llavorsi syncline, more sandy intercalations can be observed. Pyrite, generally altered into limonite, occurs frequently.

Thickness. – The thickness varies from 20 to 320 m. The greatest thicknesses generally occur in the northern region where the Ansobell Formation replaces the Cavá Formation (Appendix II and Fig. 9).

Fossils. – Fossils are extremely rare, only one fragment of a trilobite and a badly preserved cystide or crinoid stem have been found.

Correlation. – Correlation (Appendix II) provides no problems wherever the Rabassa Conglomerate is developed. If these conglomerates are absent (App. II, Section 10), the boundary with the underlying Seo Formation can still be drawn as the typical banded slates have been developed. However, if the two formations are lithologically similar, as may occur north of this area, no boundary can be drawn and correlation becomes difficult.

Depositional environment. – The black colour and the pyrite crystals indicate a deposition in an euxinic environment. The almost complete absence of fossils and the position of the formation between the shallow marine Estana Formation and the coastal barrier deposits of the Bar Quartzite suggest a lagoonal environment.

Age. – Since determinable fossils have never been found, the age of this formation remains a matter of discussion. Ages suggested by other authors are:
Dalloni (1930): Caradoc s.l.
Schmidt (1931): Llandovery
Boissevain (1934): Llandovery
Solé Sabaris & Llopis Lladó (1946): Silurian.

The present author provisionally placed this formation in the Ashgill.

Bar Quartzite
The Bar Quartzite, named by Spjeldnaes (pers. comm.) after the village of Bar, has almost exclusively developed in a restricted area of the southern Pyrenees (Fig. 11).

Previous work has been done east of the map area by Brouwer (1968a, b), Bloemraad (1969) and Trouw (1969).

Lithology. – The Bar Quartzite is a sheet sand practically always resting on the black slates of the Ansobell Formation, but in the region of Sant Julia and Argolell resting on the Estana Formation. It is always overlain by the black carbonaceous slates of the Silurian.

The quartzite generally has a light-grey colour, but locally north of Bar and northeast of Bescarán it may be black. It is composed mainly of well-rounded quartz grains and only in the coarse layers are rock fragments (schist and quartzite) of any importance. Limited amounts of chert also occur.

The bedding is parallel and well-developed, heavy minerals (rutile and zircon) may be concentrated parallel to the bedding. The thickness of the beds is symmetrically distributed over the formation, thinly bedded at the top and bottom and increasingly thick towards the middle of the formation (Brouwer 1968b).

A thin (10–30 cm) conglomerate layer often occurs in the middle of the formation, containing pebbles of quartzite and schist and badly preserved fossils and fossil fragments (Fig. 12). Near the top of the formation thin black slate layers may interfinger which locally, such as southeast of Estana, may reach a thickness of 50 cm. Ripplemarks have frequently developed on top of these quartzite beds (Fig. 13).

Thickness. – The quartzite sheet generally varies in thickness between 8 and 18 m, wedges out towards the north (Figs. 9 and 11), but reappears in a small area around Merixell in Andorra (App. II, Section 10). To the east and west the formation also wedges out.

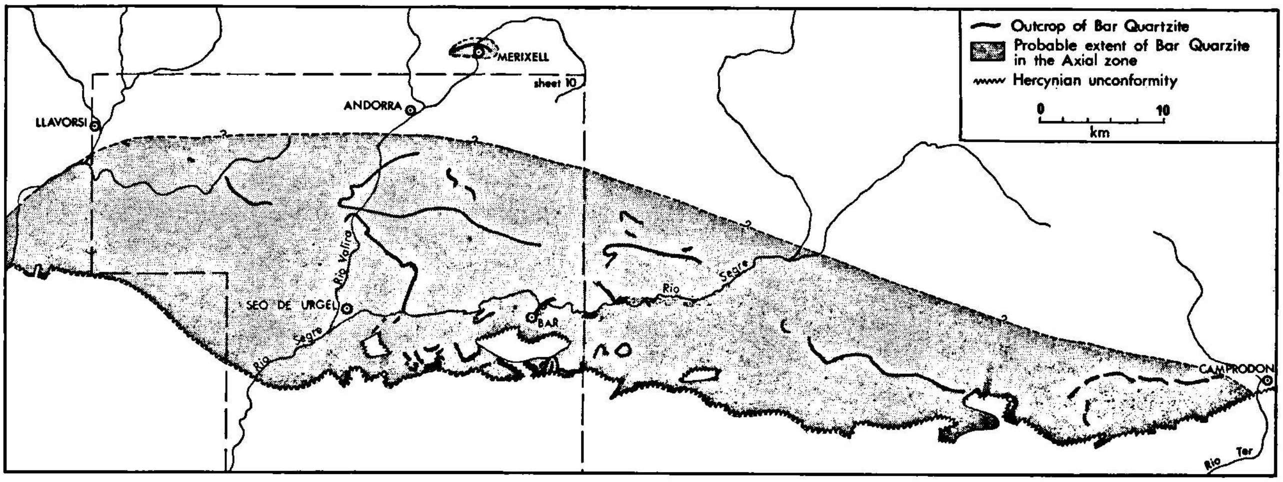


Fig. 11. Distribution of the Bar Quartzite.

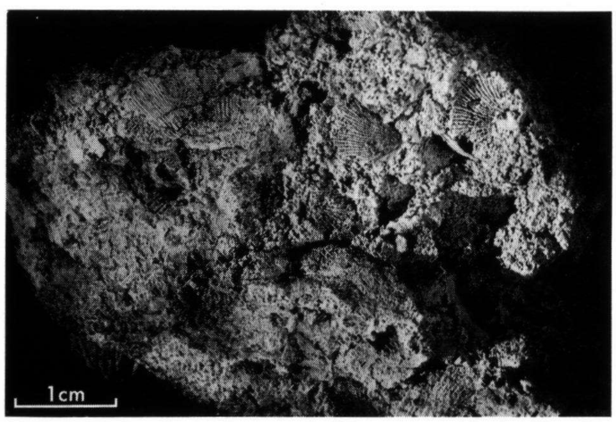


Fig. 12. Fossils from the Bar Quartzite, northwest of Estimaríu.

Fossils. – Badly preserved fossils were mainly collected north of Estimaríu along the road to Arcabell, a second but poorer fossil locality was found near Bordas del Ras.

Fig. 13. Ripple-marks on the Bar Quartzite.

The following fossils were collected: brachiopods (Dalmanellids and Strophomenids), cystoid plates and stems and rugose corals.

Correlation. – Correlation of this continuously traceable rock unit provides no problems. The quartzite described by Cavet (1964) at the top of the Ordovician of the Massif de l'Agly in the northern Pyrenees is possibly also related to the Bar Quartzite (App. II, Section 15).

Depositional environment. – The clean mainly light-coloured quartzite showing a parallel bedding, well-rounded quartz grains, well-sorted grains in the individual beds but badly sorted over the formation as a whole, the concentration of heavy minerals in layers and the occurrence of ripple marks, all indicate a beach environment (Brouwer 1968b). The absence of coarsening upwards in the Bar Quartzite, and the occurrence of probably lagoonal slates of the Ansobell Formation below, strongly suggest a barrier beach environment. The sheet form of the formation may have been caused by the slow transgression restoring the marine environment in the Silurian.

Age. – Since fossils were not suitable for determination purposes, the age of the Bar Quartzite can only be estimated. Ages suggested by other authors are:
Dalloni (1930): Caradoc s.l.
Schmidt (1931): Llandovery
Boissevain (1934): Llandovery
Solé Sabaris & Llopis Lladó: Silurian

The present author provisionally placed this formation in the top of the Ashgill. It should, however, be realized that littoral and sub-littoral formations thus far described may well be strongly diachronic.

Paleogeography and concluding remarks

After the deposition of the very uniform marine deposits of the Seo Formation, a sudden instability caused extensive mud flows from a source area somewhere south of the present axial zone depositing the Rabassa Conglomerate. The southern part of the map area was uplifted and fluviatile sediments of the lower part of the Cavá Formation were deposited on a flood plain. Volcanic activity must have accompanied this uplift, tuffaceous material being abundant. Important basin subsidence in the southernmost part of the axial zone permitted extremely thick deposits to accumulate. The upper part of the Cavá Formation is marked by a gradual transgression towards the south and deposition on tidal flats. A shallow marine environment is indicated by the limestones and marls of the Estana Formation. A small regression becomes apparent from the development of the coastal barrier of the Bar Quartzite and the lagoons of the Ansobell Formation. The Silurian is marked by a general transgression.

SILURIAN

The black carbonaceous slates, conformably overlying the Bar and Ansobell Formations and conformably covered by the Devonian Rueda Formation, belong to a rock unit of almost completely Silurian age which has a very uniform development in the Pyrenees and also far beyond these mountains. For this reason the introduction of a local geographic name for this rock unit would be undesirable and the term Silurian is used, as is done by most authors.

The Silurian rocks have already been extensively described by Destombes (1953), Zwart (1954), Cavet (1957), Clin (1958), Kleinsmiede (1960), Zandvliet (1960), de Sitter & Zwart (1962), Mirouse (1966) and Mey (1967). For further details of the general rock characteristics reference is made to these publications.

Lithology. – The Silurian generally consists of a sequence of black carbonaceous slates in which thin black limestone beds are intercalated near the top. Slates often contain graptolites, some trilobites and calcareous concretions are frequent in which orthocerids may often be found. Limestones frequently contain orthocerids, crinoid ossicles and lamellibranchs.

In a restricted area, south of Tolorfu, a 10 m thick nodular limestone is found between the black slates (Fig. 14 and App. II, Section 39). This Tolorfu limestone shows grey limestone nodules surrounded by black slate. Except for conodonts, no fossils have been found in it.

The most important characteristic of the Silurian is its extreme incompetency due to which these slates may have been squeezed out in one place and accumulated in another. Diapiric movements along faults have also frequently been observed.

Thickness. – Because of the incompetency of the Silurian slates the thickness can only roughly be estimated at about 100 m in the south to 250 m in the northern part of the map area.

Fig. 14. Nodular limestone in slates of the Silurian southeast of Toloríu.

Fossils. – Graptolites are abundant in the slates, orthocerids, lamellibranchs, crinoids and trilobites are mainly found in the black limestones near the top. Conodonts may be found in all limestones. Extensive lists of fossils (mainly graptolites) found in this area are given by Roussel (1904, quoted by Boissevain, 1934, p. 51), Dalloni (1930, p. 62–65), Schmidt (1931, p. 36), Boissevain (1934, p. 45–47), Solé Sabaris & Llopis Llopis Lladó (1946, p. 87–89) and Llopis Lladó (1966).

Two fossil localities deserve special attention as they are of great structural importance occurring in diapirs of Silurian slates amid the Carboniferous slates in the centre of the Llavorsi syncline. Graptolites from the locality east of B. de Corves are shown in Fig. 15.

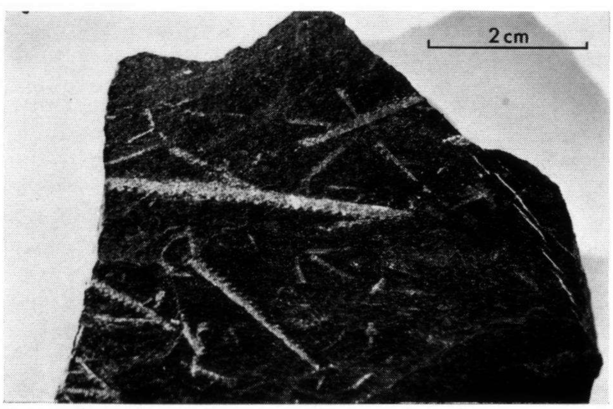


Fig. 15. Graptolites from a slate diapir in the Llavorsi syncline, east of B. de Corves.

A second locality east of Bixesarri was found by Oliver (1967) who recorded: *Monograptus cf. spiralis*, Geinitz, *Monograptus cf. priodon*, Bronn, *Retriolites* sp., indicating a Llandovery age.

Age. – The following ages have been deduced from graptolite faunas by other authors:
Llandovery, Wenlock, Ludlow; near Estana; Roussel (1904).
Llandovery, Wenlock; between Ortedó and Ansobell; Dalloni (1930).
Ludlow; near Torres; Schmidt (1931).
Llandovery; S of Rio Segre; Boissevain (1934).
Llandovery (?), Wenlock, Ludlow; W of Encamp; Llopis Lladó (1966).
Llandovery; E of Bixesarri; Oliver (1967).

Determinations by Boersma (1968a) of conodonts indicate a Lower Gedinnian age in the Orthoceras limestones from the Nogueras zone west of this area. Conodonts from the Toloríu limestone were not conclusive for the age.

Depositional environment. – Very calm sedimentation in a shallow marine euxinic environment (Mirouse, 1966, p. 46–47).

DEVONIAN

Rueda Formation

The Rueda Formation, defined by Mey (1967) in the southern Pyrenees between the Esera and Mañanet valley, also occurs in our area and conformably overlies the black carbonaceous slates of the Silurian.

A detailed description of this formation in the Ribagorzana and Baliera valleys was given by Mey (1967).

Lithology. – As shown in Fig. 16, the Rueda Formation is not uniformly developed in the area of sheet 10. In the north, in the Llavorsi syncline (App. II, Sections 40 and 41), the formation is made up of an alternation of impure limestone beds and black slates that best fits Mey's description. The thickness of the individual beds generally varies between 1 and 50 cm. The rocks are usually dull brown and contain abundant crinoid stems and orthocerids.

Locally (viz. north of Sant Joan Fumat) a dark-grey, rather pure 20 m thick limestone is present at the base of the formation, displaying typical orange-brown blots on the weathered surface. This limestone is more abundant to the west in the Mañanet-Flamisell basin (Roberti in prep.).

Towards the southeast the limestones of the Rueda Formation become increasingly pure and slate intercalations become thinner and less frequent. In the Segre valley (App. II, Sections 42 and 43) the clay content of the limestones is often so small that the boundary between this formation and the overlying Basibé Formation becomes very indefinite. Here limestones vary from dark to light grey and may locally be nodular. Crinoid stems and orthocerids are abundant.

The most aberrant development of the Rueda Formation occurs in the Monsech de Tost (App. II, Section 44). Here the lower part consists of a thick (approx. 100 m) siltstone sequence in which locally thin impure limestone beds which often yield an abundance of brachiopods. At the top a 20 m thick impure limestone level with some slate intercalations resembles the formation in the Segre valley.

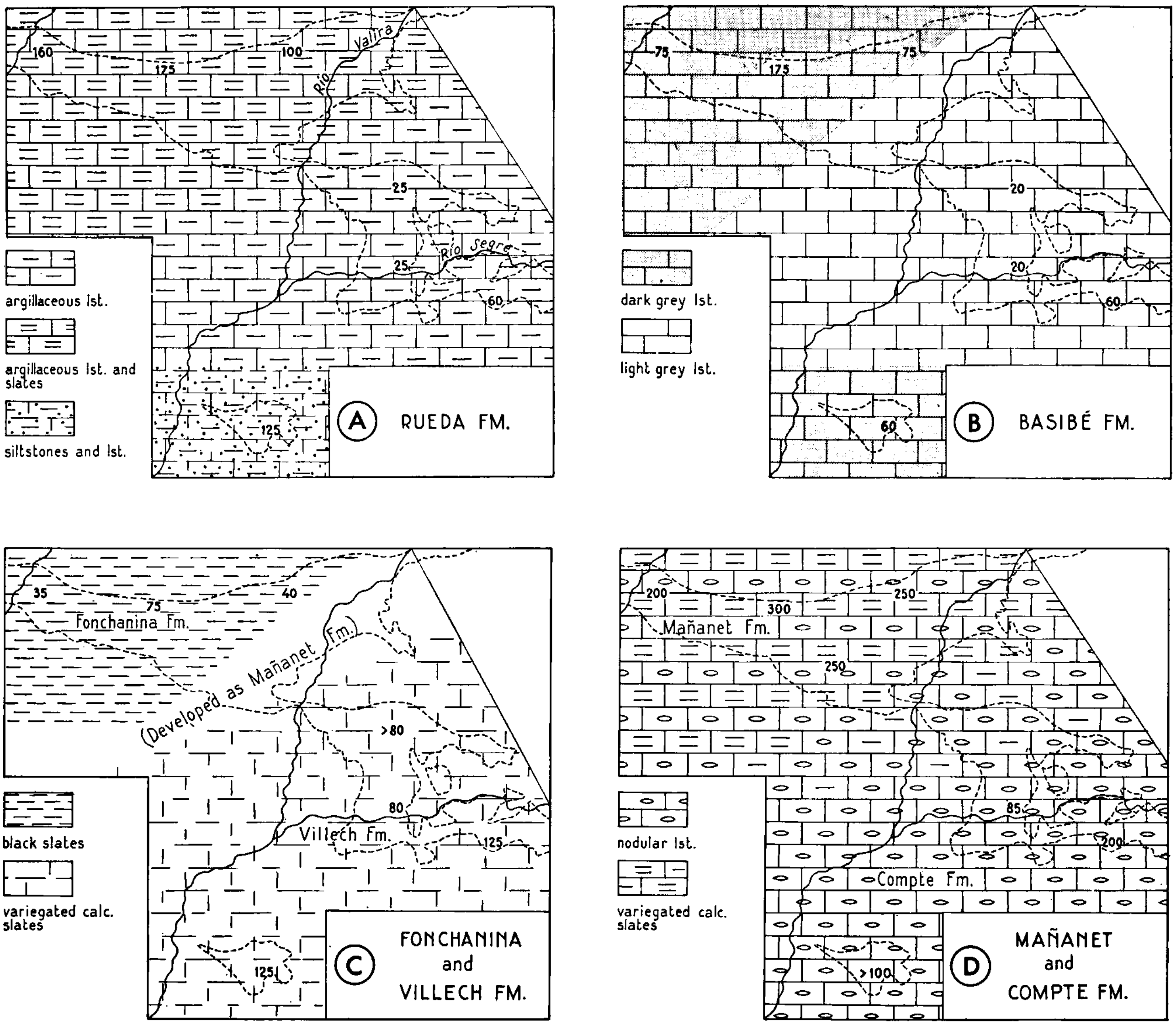


Fig. 16. Lithofacies maps of the Devonian and distribution of thickness in metres.

Thickness. - The thickness varies between 25 and 175 m (Fig. 16-A).

Fossils and age. – Crinoid stems, orthocerids, corals, tentaculitids and conodonts are generally present. In the Monsech de Tost brachiopods are particularly abundant and trilobites may also occur. Poor preservations of the fossils generally prohibits determination, however. Fossil lists were provided by Dalloni (1930, p. 84) and Solé Sabaris & Llopis Lladó (1946, p. 89–90), but as the exact locality in the severely folded and faulted Devonian sequence can generally not be read from their descritptions, the lithostratigraphic position is not known with certainty. Schmidt (1931) found in this formation near Torres: *Phacops cf. fecundus* Barrd. and *Thylacocrinus vanniosti* Oehl., from which he concluded a Gedinnian age. Boissevain (1934, p. 53) recorded in a locality near Pont de Bar: *Phacops fecundus* Barr., *Capulus* sp., *Ambocoelia* sp. Boersma (1968a) found a lower Gedinnian age for conodonts from the lowermost part of this formation, in the Nogueras zone west of the map area.

Basibé Formation

The Basibé Formation, defined by Mey (1967) in the southern Pyrenees between the Esera and Mañanet valleys, conformably overlies the Rueda Formation in our area.

A detailed description of this formation in these more westerly areas is given by Mey (1967).

Lithology. – The Basibé Formation has developed uniformly as a massive pure limestone unit and only variations in thickness and colour occur (Fig. 16-B). In the western and northern part of the Llavorsi syncline the limestone is dark-grey with a light-grey weathering and pyrite may locally be abundant. To the southeast the limestones gradually change from dark to light-grey, locally becoming slightly nodular. In the Segre valley the boundary between the light-grey Basibé limestones and the rather pure limestones of the Rueda Formation is often difficult to draw. In the Monsech de Tost the Basibé Limestones become dark-grey to almost black and can easily be recognized in the field.

The upper boundary of the formation is always sharp, yet conformable. In the western part of the Llavorsi syncline the limestones are overlain by the black slates of the Fonchanina Formation and in the eastern part of this syncline they are directly covered by the multicoloured nodular limestones and calcareous slates of the Mañanet Formation (Fig. 16-C). In the Segre valley and in the Monsech de Tost the formation is overlain by the multicoloured calcareous slates of the Villech Formation.

Thickness. – The thickness varies between 20 and 175 m (Fig. 16-B).

Fossils and age. – Crinoid stems, conodonts and tentaculitids generally occur. Boersma (1968a) found an Emsian age for conodonts from the top of this formation in the Nogueras zone west of our area.

Fonchanina Formation

The Fonchanina Formation defined and described by Mey (1967) in the southern Pyrenees between the Esera and Mañanet valleys to the west, is only to be seen in the western and northern parts of the Llavorsi syncline in our area (Fig. 16-C).

Lithology. – The Fonchanina Formation is a sequence of dark-grey to black slates in which a few thin black limestone beds may occur. The formation is always underlain by the dark-grey limestone of the Basibé Formation. At the top a rather gradual transition takes place to the overlying variegated slates and nodular limestones of the Mañanet Formation. To the southeast the black slates of the formation grade into the variegated slates and calcareous slates of the lower part of the Mañanet Formation. Fig. 15 B and C show that this lateral transition takes place in the same zone where the colour changes from dark to light-grey in the Basibé Limestone below.

Thickness. – Thickness estimates of the relatively incompetent Fonchanina Formation are necessarily uncertain. The maximum thickness does not exceed 75 m (Fig. 16-C).

Fossils. – Only a few crinoid ossicles have been found.

Correlation. – To the southeast the formation should be correlated with the lowermost part of the Mañanet Formation and the lower part of the Villech Formation (App. II, Sections 41, 42).

Age. – Boersma (1968a) found an Emsian age for conodonts from the Fonchanina Formation in the Nogueras zone west of the map area.

Villech Formation

The Villech Formation, named after the Tossal de Villech south of the Rio Segre, is a sequence of calcareous slates and griottic limestones bounded by the Basibé Formation below and the Compte Formation above. The formation has developed in the southern Pyrenees, from the Pallaresa valley in the west up to the Ter valley in the east.

Descriptions of this formation have been given by Hartevelt (1965), Rijnsburger (1967), Boersma (1968a, Brouwer (1968a), Bloemraad (1969) and Trouw (1969).

Lithology. – The Villech Formation is a mainly reddish slate and limestone sequence which always rests with a sharp conformable contact on the Basibé Limestones. In the area of Bescarán, the Beneidó and in the Monsech de Tost (App. II, Sections 42, 44) the formation consists largely of reddish fissile slates,calcareous slates and some thin yellowish limestone beds. Intercalations of greenish calcareous slates frequently occur. Towards the east the formation becomes more calcareous and pink griottic limestone intercalations become very frequent, even predominating in the Beixech area. The pink griotte-like layers generally have a relatively high slate content which makes them easily distinguishable from the massive dark-red griottes of the overlying Compte Formation. Locally, especially near the granodiorite contact, the formation may be discoloured to greyish cacareous slates and limestones and no longer be differentiated from the Rueda Formation.

The formation is conformably overlain by the light-grey nodular limestone of the Compte Formation. To the northwest there is a very gradual transition into the Mañanet Formation (App. II, Sections 41, 42 and Fig. 16-C), and the boundary drawn through Farga de Moles is arbitrary.

Thickness. – The thickness of the relatively incompetent Villech Formation varies between 80 and 125 m (Fig. 16-C).

Fossils and age. – Crinoid stems, orthocerids, ammonites, corals, tentaculitids and conodonts can generally be found. Dalloni (1930, p. 84–85, 90) found in this formation north of Arseguell: *Phacops occitanicus* Trom-Grass., *Orthoceras* sp., *Anarcestes* sp., *Bacrites* sp., *Atripa reticularis* Linn., *Orthis*, *Cyathocrinus pinnatus* Goldf., *Cyathophyllum*, *Favosites*., from which he concluded an Eifelian age. Boersma (1968a, pers. comm.) found an Emsian age for conodonts in the lower part of the formation, an Eifelian age in the middle and upper part and a Givetian age at the top.

Correlation. – To the northwest the formation should be correlated with the Fonchanina Formation and the lower part of the Mañanet Formation.

Mañanet Griotte
The Mañanet Griotte, defined by Roberti (pers. comm.) in the Mañanet valley, has in our area only developed in the Llavorsi syncline. A detailed description of this formation in the area between the Rio Esera and Mañanet is given by Mey (1967).

Lithology. – The Mañanet Formation is a sequence of multicoloured nodular limestones in which calcareous slates are frequently intercalated. Colours of the fresh rock may be reddish, greenish, violet, white or grey, whereas the weathered surface has a greenish-brown or grey appearance. In the northern and western part of the Llavorsi syncline there is a gradual transition into the black slates of the Fonchanina Formation below, whereas in the south-eastern part of the Llavorsi syncline and in the Arcabell syncline west of the Valira valley the formation directly overlies the Basibé Limestone with a sharp but conformable contact. To the southeast the lower part of the Mañanet Formation grades into the similarly developed Villech Formation, whereas the upper part is the equivalent of the massive nodular limestones of the Compte Formation. At the top there is a gradual transition into the black slates of the Civis Formation.

Thickness. – The thickness of the Mañanet Formation varies between 200 and 300 m (Fig. 16-D).

Fossils. – Crinoid ossicles.

Correlation. – The lower part of the Mañanet Formation should be correlated with the Villech Formation and the upper part with the Compte Formation. The lowermost part of the Mañanet Formation, in the southeastern part of the Llavorsi syncline and the western part of the Arcabell syncline, can be correlated with the Fonchanina Formation (App. II, Sections 41,42).

Age. – As no determinable fossils have been found in this formation there is no direct evidence concerning of age. The Mañanet Formation overlies the Fonchanina Formation (Emsian) and is correlated with the Villech Formation (Emsian-Givetian) plus the Compte Formation (Givetian-Tournaisian), so that the age of the formation may range from Emsian to Tournaisian.

Compte Formation
The Compte Formation, named after the hamlet of Compte in the Pallaresa valley west of our area, is a sequence of massive nodular limestones typically developed in the southern Pyrenees from the Pallaresa valley in the west up to the Ter valley in the east.

Descriptions of this formation have been given by Schmidt (1931), Boissevain (1934), Hartevelt (1965), Rijnsburger (1967), Boersma (1968a), Brouwer (1968a), Bloemraad (1969) and Trouw (1969).

Lithology. – The limestones of the Compte Formation always rest, with a sharp conformable contact, on the Villech Formation. The formation can generally be subdivided into three members: a light-grey often nodular limestone below, a dark-red nodular limestone in the middle and a second light-grey often nodular limestone at the top (App. II, Sections 42, 43).

The two light-grey limestone members show no clear lithological difference and can generally only be distinguished from each other by their stratigraphic position. Where nodules are developed the middle of each is often a red spot. Intraformational breccias occurring in some places in the lower member show that the partly lithified sediment was broken up after deposition. The dark-red griotte member consists of limestone nodules, very often enclosing goniatites, surrounded by a darker slaty material. The subdivision into three members is not possible in the Monsech de Tost and north of Coll de Ser because the reddish and grey nodular limestones are distributed more irregularly here.

The Compte Formation is overlain by the Bellver Formation (App. II, Section 43). This contact may be sharp and conformable if overlain by a chert, slate and limestone sequence. An unconformable contact is generally observed if the overlying rocks are conglomeratic. The unconformity can be important as locally the Compte Formation is even absent e.g. ENE of Beixech.

Thickness. – The Compte Formation at the Beneidó (App. II, Section 42) can be estimated to be 85 m thick, as remnants of the chert layers of the Bellver Formation can still be observed almost in situ. The thickness (Fig. 16-D) increases to 200 m in the Beixech area (App. II, Section 43) where the lower member has a very thick development of 100 m in the B^{co} de Quer.

Fossils. – Ammonites, conodonts, tentaculitids and crinoid remains are generally found and goniatites are especially abundant in the red griotte member.

Age. – A Famennian age for the two upper members was concluded by Schmidt (1931, p. 50) from ammonite fauna near Isobol, east of this area. Mr. K. Th. Boersma (University of Leiden) kindly determined the conodonts from the lower member (Dc1) near Villech as Givetian-Frasnian, the middle member (Dc2) as Frasnian-Famennian and the upper member as Famennian. At several other localities east and west of our area the uppermost part of the upper member has been dated as Tournaisian (Boersma, 1968b; pers. comm.).

Correlation. – The Compte Formation should be correlated with the upper part of the Mañanet Formation.

Depositional environment in the Devonian
Mirouse (1966; 1967, p. 167–168) suggests deposition in a shallow marine environment in unstable basins with differential subsidence.

PRE-HERCYNIAN CARBONIFEROUS

Bellver Formation

The Bellver Formation (Brouwer, 1968a) shows a characteristic development in the southern Pyrenees from the Pallaresa valley in the west to the Ter valley in the east. Descriptions of this formation have been given by Boissevain (1934), Solé Sabaris & Llopis Lladó (1946), Rijnsburger (1967), Boersma (1968a), Brouwer (1968a), Bloemraad (1969), Trouw (1969) and Waterlot (1969).

In the map area the formation is only present in some synclines in the Beixech area, whereas southwest of the Beneidó some fragments of chert layers have been found and along the road between Ortedó and Vilanova some conglomerate of this formation is present in a fault.

Lithology. – Where conformable with the Compte Formation the sequence may locally begin with a thin black chert layer of about 30 cm which is often followed by a thin yellowish-grey nodular limestone about 1 m thick. A thick sequence of black or brownish slate with sandstone intercalations rests either upon this limestone or directly upon the Compte Formation. In these slates and sandstones sedimentary structures such as graded bedding, cross-bedding, loadcasts and slumps may occur. Conglomerate lenses, which sometimes have unconformable contacts with the other deposits, frequently occur. The conglomerates mainly consist of wel-rounded pebbles, with a maximum diameter of 15 cm, in a slaty matrix. Usually they are of quartz, quartzite, chert and gneiss but limestone pebbles also occur, especially near the contact with the limestones of the Compte Formation. Boissevain also recorded some pebbles of granitic and dyke material. The conglomerates may laterally grade into pebbly mudstones. Between Coll de Ser and the B^co de Quer the conglomerates comprise the entire formation and rest directly on the Compte Formation, often displaying local unconformities. Buried topography below these conglomerates may be seen northeast of Beixech, where locally the Compte Formation is even completely eroded away.

Thickness. – No section is complete in the Beixech area, but the Bellver Formation is at least 200 m thick.

Fossils and age. – No fossils have been found in the Beixech area, but in Cerdaña Boissevain (1934, p. 56–59) found goniatites of Viséan age, a crinoidal limestone and indeterminable plant remains. In the Nogueras zone west of the map area Waterlot (1969, p. 122–125) found goniatites and a crinoidal limestone and Boersma (1968a) recorded some brachiopods. Conodonts, generally found in limestone intercalations east and west of the map area, have been dated by Boersma (1968b; pers. comm.) as Tournaisian-Viséan and possibly Namurian.

Correlation. – The Bellver Formation can best be correlated with the Civis Formation developed in the Llavorsi syncline.

Civis Formation

The Civis Formation, named after the village of Civis west of Andorra, is defined as the black slate sequence which overlies the Mañanet Formation. The formation is developed north of the Bellver Formation and in the map area it only occurs in the Llavorsi syncline. Recent descriptions have been given by Zandvliet (1960), Mey (1967) and Waterlot (1969).

Lithology. – The Civis Formation consists mainly of monotonous greyish-black slates in which locally some thin sandy layers and limestone beds may occur. The transition between the greenish calcareous slates of the Mañanet Formation and the grey-black slates of the Civis Formation is always gradual. The slates are usually slightly sandy and often contain large mica flakes. Bedding is generally very vague but becomes clearer where the beds are more sandy or calcareous.

Limestone beds are relatively abundant in the area around Civis and in Andorra; they are generally impure, with a brownish-grey appearance somewhat resembling the limestones of the Rueda Formation. These limestones can only be followed over short distances (up to 1500 m) as they grade laterally into the black slates. Only these calcareous deposits are fossiliferous.

Real sandstones are very scarce and have only been found along the Andorran border southeast of the Caumer.

Thickness. – The thickness cannot be estimated as the top of the formation is unknown, but should be at least 400 m.

Fossils. – Tentaculitids, gastropods and crinoid stems (Fig. 17).

Correlation. – Lithological similarities and stratigraphic position make a correlation with the Bellver Formation plausible (App. II, Sections 40, 41, 43).

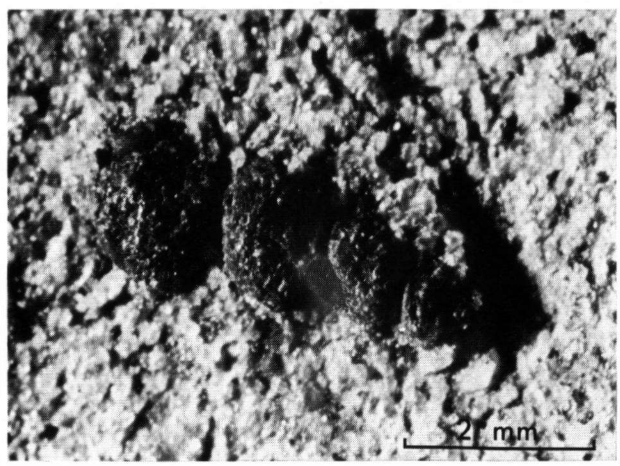


Fig. 17. Gastropod from the Civis Formation west of Civis.

Age. – Fossils found in this formation cannot be specifically determined so that no direct age determinations can be given. The correlation with the Bellver Formation and the fact that they are pre-orogenic suggest that the age may range from Tournaisian to Westphalian. The subject is extensively discussed by Zandvliet (1960, p. 58–59).

Depositional environment in the pre-Hercynian Carboniferous
The lower part of the conformable sequence of the Bellver Formation with its chert and limestone beds is characteristic of a very slow influx of detritic material in a shallow marine environment. The higher part of the sequence with its erosional features and conglomeratic piemont and littoral deposits (Waterlot, 1969, p. 176–178) indicates the instability which was the prelude to the Hercynian orogeny. In the more northerly area, where the Civis Formation is developed, evidence of such instability has not been observed.

POST-HERCYNIAN CARBONIFEROUS

Aguiró Formation
The Aguiró Formation (Mey et al., 1968) is a thin sequence of slope breccias which is only locally developed between the post-Hercynian unconformity and the Erill-Castell Volcanics. Previous work has been carried out in the map area by Schmidt (1931) and Diederix (1963). Other detailed descriptions of this rock unit in more westerly areas have been given by Dalloni (1930), Mey (1968) and especially Nagtegaal (1969).

Lithology. – The Aguiró Formation mainly consists of coarse breccias, with fragments up to 50 cm, of material derived from the Hercynian basement which it unconformably overlies. The breccias have colours varying from grey to purplish, whereas Cambro-Ordovician rocks directly below the unconformity have been reddened. In the Bco. de Piedra the breccias grade into greywackes, sandstones and shales with identifiable plant remains and a coal seam near the top. The formation is unconformably overlain by the Erill-Castell Volcanics or by the Malpas Formation.

Thickness. – A maximum thickness of 15 m has been observed south of Ges. In several other localities, not indicated on the geological map, very thin slope breccias have been found, e.g. west of Bestanis where some carbonaceous material was also observed below the Volcanics.

Correlation. – The position in the stratigraphic succession and the age determinations confirm the correlation of the deposits in our area with those at Aguiró. Schmidt (1931, p. 9–10, 60) incorrectly correlated these deposits in the Bco de Piedra (Bco de Colo or Bco. de Bastida) with the Malpas Formation at Las Minas and placed all coal-bearing formations below the volcanics. Boissevain, misled by this error, suggested a Permian age for the Erill-Castell Volcanics.

Depositional environment. – The formation most probably represents a fossil slope breccia deposited on a steep relief of a land area (Nagtegaal, 1969).

Fossils and age. – Plant casts found by Diederix (1963) in the Bco. de Piedra have kindly been determined by Mr. H. W. J. van Amerom (Rijks Geologische Dienst, Heerlen), who concluded an upper Westphalian D – lower Stephanian A age. The material yielded the following imprints: *Annularia stellata* v. Schloth., *Annularia sphenophylloides* (Zenk.) v. Gutb., *Calamites* sp., cf. *Sphenophyllum* sp., *Neuropteris* sp., *Callipteridium* cf. *pteridium* v. Schloth., *Pecopteris cyathea* v. Schloth., *Pecopteris candolleana* Brgt., *Asteroctheca* sp., cf. *Pecopteris* sp. In addition, Dalloni (1930, p. 99) collected imprints of: *Calamites* sp., *Pecopteris cyathea* Schloth., *Pecopteris hemitelioides* Brongn., *Odontopteris Reichi* Gutb. and concluded a Stephanian age. In the region of

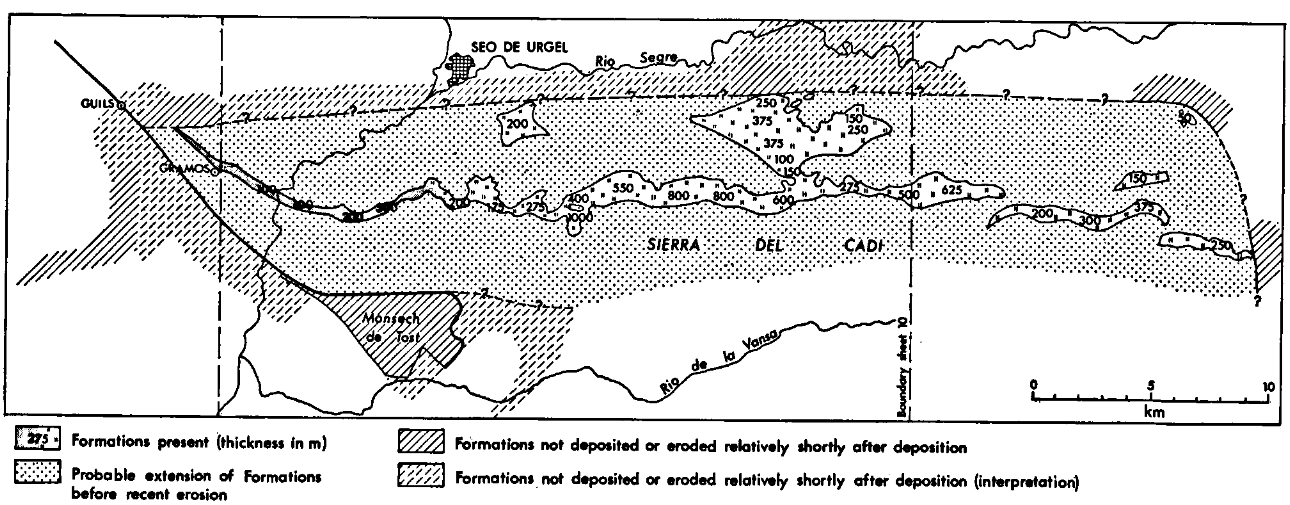


Fig. 18. Distribution and thickness of the Erill Castell Volcanics and the Malpas Formation.

Aguiró, Dalloni (1930, p. 95–96) reported an age ranging from Westphalian D to lower Stephanian.

Erill Castell Volcanics

The Erill Castell Volcanics (Mey et al., 1968) are mainly composed of tuffs in which andesite sheets may occur. The formation, which in the map area is restricted to an E-W trending zone east of Las Minas (Fig. 18), overlies the Aguiró Formation or the Hercynian basement and is covered either by the Malpas Formation or the Peranera Formation. Previous work in the map area was carried out by Boissevain (1934) and Morre & Thiébaut (1964). Other detailed descriptions of this rock unit in more westerly areas have been given by Mey (1968) and especially Nagtegaal (1969).

Lithology. – The grey, greenish or reddish tuffs of the Erill Castell Volcanics unconformably overlie the slope breccias of the Aguiró Formation or the Hercynian basement. The lower boundary is often very irregular and the volcanics probably buried a still pronounced relief. Slope breccias identical to those observed in the Aguiró Formation, may be intercalated in the lower part of the formation. Darkgrey lava sheets may occur locally which are generally more resistant and project from the surrounding weathered tuffs. These lava sheets may have a composition of andesite, dacite, albitophyre or spilite as described by Boissevain (1934, p. 59–65) and Morre & Thiébaut (1964).

Thickness. – As the Erill Castell Volcanics often fill up old valleys and the bedding is usually very poorly developed, estimates of the thickness are difficult. Great thicknesses of up to 800 m can be observed between the meridians of Ges and Queforadat (Fig. 18). A very thick development has been observed 2 km south-east of Ges (cross section 8) where a high ridge of volcanic rocks projects above the red beds of the Peranera Formation displaying an extreme example of the steep relief that existed locally still during the deposition of the Peranera Formation.

Depositional environment. – The irregular lower boundary and the intercalation of slope breccias in the tuffs suggest a deposition on a still pronounced topography of a land area (Nagtegaal, 1969).

Age. – Stephanian floras of the Aguiró Formation below and the Malpas Formation above indicate that the Erill Castell Volcanics are also of this age.

Malpas Formation

The Malpas Formation (Mey et al., 1968) is, in the map area, mainly developed in a narrow E-W trending zone between the B^{co}. de Piedra in the east and Guils in the west (Fig. 18). The formation overlies the Erill Castell Volcanics or the Hercynian basement and is always conformably overlain by the Peranera Formation. Previous work has been carried out by Dalloni (1930), Schmidt (1931) and Diederix (1963). Detailed descriptions of this formation in more westerly areas have been given by Mey (1968) and especially by Nagtegaal (1969).

Lithology. – West of the Can Franch, the Malpas Formation, with an unconformable contact, rests upon the Hercynian basement; east of it it rests upon the Erill Castell Volcanics. At the base of the formation a breccia of Cambro-Ordovician material may be locally developed which west of the Segre near Gramós (just outside the map), displays a dark-red colour. If the formation rests on the Erill Castell Volcanics a breccia of Cambro-Ordovician material may also be found, as, for example, northeast of the Can Franch. The nature of the contact with the Erill Castell Volcanics remains uncertain as the stratification of the volcanic rocks is generally poorly developed.

The main mass of the formation consists of an alternation of yellowish-brown sandstones and greywackes with much reworked tuffaceous material, well-rounded conglomerates, grey-black mudstones and shales often containing siderite concretions and thin coal beds (App. III, Section 45). Locally a rhythmic alternation may be observed. The coal seams are mainly concentrated in the lower part of the sequence and reach a maximum thickness of about 80 cm at Las Minas, where they have mainly been exploited in the past. Plant imprints are also chiefly restricted to the lower part of the formation.

Limestones are well developed near Gramós where numerous thin (<30 cm) limestone beds intercalate in a black slate sequence in the lower part of the formation, and at the top a 2 m thick limestone bed is developed which is conformably overlain by the red beds of the Peranera Formation. Between the Segre and Las Minas, some thin limestone beds are also intercalated.

Some traces of carbonaceous material probably representing the Malpas Formation were found east of the Grau on top of the tuffs, while Boissevain (1934, p. 67) recorded some of this material in the Serrat de Mosbé between the Erill Castell Volcanics and the Bunter (not indicated on the map).

The upper boundary is usually a gradual transition from the greyish and black deposits of the Malpas Formation to the reddish and grey deposits of the Peranera Formation.

Thickness. – A maximum thickness of about 300 m is observed near Las Minas (Fig. 18).

Fossils and age. – Plant imprints are generally found in the Malpas Formation and extensive lists were given by Dalloni (1930, p. 98–99) of fossils from localities near Gramós, Las Minas and north of the Can Franch and by Schmidt (1931, p. 60) for Las Minas. All determinations indicate a Stephanian age.

Depositional environment. – The deposition of coal, shale, mudstone and limestone indicate a backswamp and lacustrine environment, the rhythmic alternation of sandstone indicates a fluviatile influence. For a detailed analysis reference is made to Nagtegaal (1969).

PERMIAN AND LOWER-TRIASSIC

Peranera Formation

The Peranera Formation (Nagtegaal, 1962) is a very thick sequence of red beds overlying the Malpas and Erill Castell Formations and covered by the Bunter. The formation exposed in the map area is geographically restricted to a narrow E-W trending zone between Guils in the west and Prat de Aguillo in the east (Fig. 19).

Previous work in the map area has been carried out by Dalloni (1930), Boissevain (1934) and Diederix (1963). Detailed descriptions of this formation in more westerly areas have been given by Roger (1965), Mey (1968) and especially Nagtegaal (1969).

Formation (App. III, Section 46) may be roughly subdivided into three members. The lower member mainly consists of coarse breccious deposits, greywackes and sandstones mainly composed of transported tuffaceous material. Some thin silty and sandy beds may intercalate and locally a thin limestone bed may be observed just above the zone of transition into the Malpas Formation.

The middle member is characterized by the occurrence of calcareous concretions and concretionary limestone beds intercalated in a sandstone and siltstone sequence.

The upper member is mainly composed of silt and mudstones in which mud-cracks are very frequent. In the lower part of this member a light coloured coarse sandstone bed occurs, in which plant imprints are often to be found. Very thin limestone beds may be interbedded in the uppermost part of the member. The formation is generally unconformably overlain by the coarse conglomerate of the Bunter Formation.

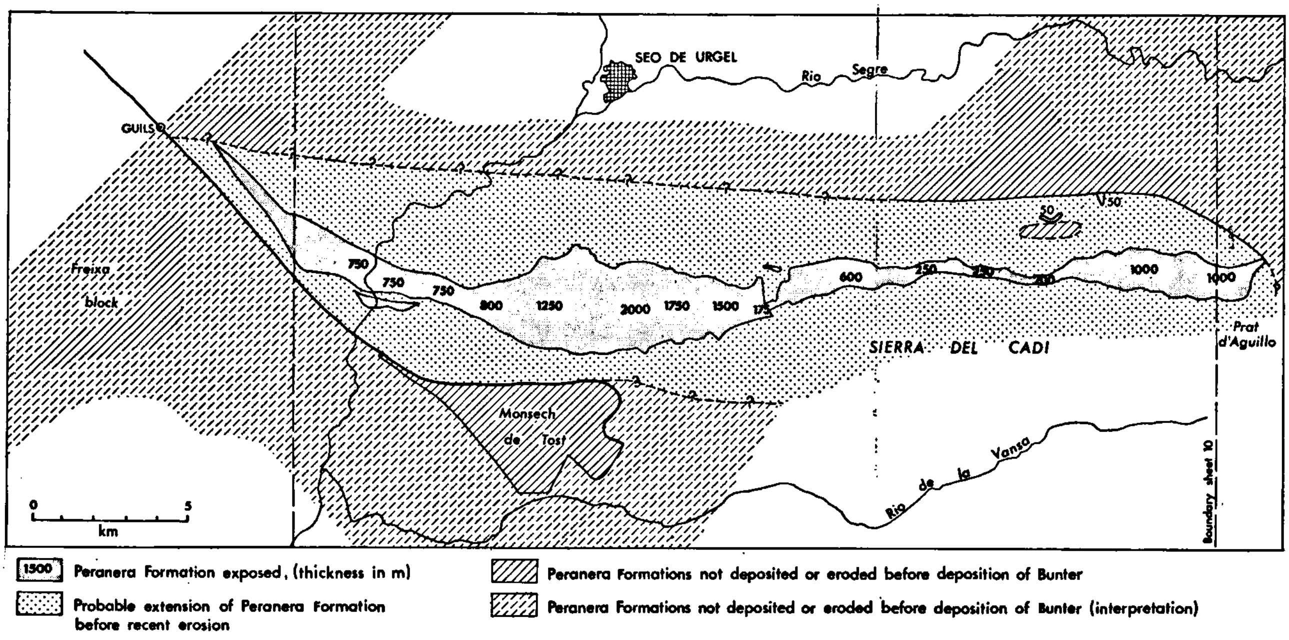


Fig. 19. Distribution and thickness of the Peranera Formation.

Lithology. – The transition from the Malpas Formation to the Peranera Formation is generally very gradual, taking place within a 10–15 m thick zone in which reddish shales and sandstones alternate with black shales and grey-brown sandstones. East of the B^{co}. de Piedra, the Peranera Formation, with a sharp contact, rests directly upon the Erill Castell Volcanics. There the nature of the boundary becomes uncertain because of the poorly developed stratification in the volcanic deposits. Southeast of Ges the lower contact is clearly unconformable upon the paleorelief of the volcanics.

The thick greyish-red rock sequence of the Peranera

Thickness. – Considerable variations occur over short distances, a maximum thickness of 2000 m may be observed south of Seo de Urgel (Fig. 19) and important basin subsidence must have taken place.

Fossils. – Badly preserved plant imprints generally found in the upper part of the formation (App. III, Section 46) show a resemblance with *Walchia* sp. which is commonly found at that level.

Age. – A Permian age was inferred by Dalloni (1930, p. 114), Virgili (1960) and Roger (1965).

Depositional environment. – Nagtegaal (1969) concluded that deposition took place on lowlands bordering on alluvial fans under arid climate conditions.

Bunter Formation

The Bunter Formation (Mey et al., 1968) is a sequence of red beds unconformably overlying rocks ranging from Cambro-Ordovician to Permian age and always overlain by the Pont de Suert Formation. The formation is developed almost everywhere.

Previous work in the map area has been carried out by Dalloni (1930), Boissevain (1934) and Diederix (1963). Detailed descriptions of this formation in more westerly areas have been given by Roger (1965), Mey (1968) and especially Nagtegaal (1969).

Lithology. – The lower contact of the formation is always sharp and in general unconformable. Contacts between the Peranera Formation and the Bunter, observable in the map area are, however, usually para-conformable, but to the west near Guils (Fig. 19) this contact becomes an angular unconformity of about 35° (Roberti pers. comm.) completely cutting off the Peranera Formation towards the north. A similar relationship should be expected east of Bestanis and also between the Sierra del Cadi and the Turo de Call Pubill where the Peranera Formation is almost completely absent. As the formation is not developed in the Monsech de Tost, it is also believed to have been cut off towards the south by the Bunter unconformity, a feature which was earlier reported in the Nogueras zone to the west, by Mey (1968, p. 274–279).

The basal layers of the Bunter vary considerably from place to place. When overlying the Peranera Formation they almost invariably consist of very coarse conglomerates, not exceeding a thickness of 10 m, with well-rounded pebbles of quartz, quartzite and chert up to 20 cm in size. Boissevain also found pebbles of volcanic material derived from the Erill Castell Volcanics and granitic pebbles which, however, differed in composition from the Andorra granodiorite. The lower contact is always strongly loadcasted in the underlying silststones. The same conglomerate may occur locally, as the formation overlies the Erill Castell Volcanics. At the unconformity with the Devonian limestones, as may be observed north of Coll de Ser and in the Monsech de Tost, the basal layers consist mainly of limestone breccias intercalated in the red mudstones and always reflecting the local lithology below the unconformity. North of Coll de Ser, it is especially noticeable that during the deposition of the Bunter the relief locally remained steep since large limestone boulders up to one meter in diameter occur amid mudstones bounding small but steep buried cliffs of Devonian limestone (Fig. 20), (Hartevelt & Roger, 1968). Locally these Devonian limestones formed elevations high enough to project above the Bunter deposits, and here the limestones of the Pont de Suert Formation rest directly upon those of Devonian age. Topographic highs become more prominent towards the east where all Triassic deposits finally wedge out 7 km east of Bestanis (Boissevain, 1934).

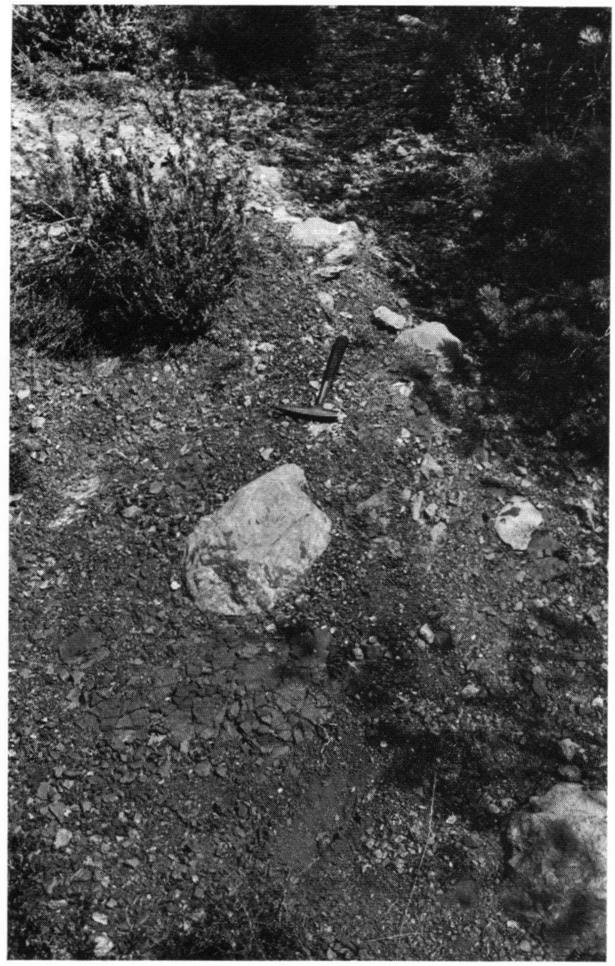


Fig. 20. Boulders of Devonian limestone in mudstone of the Bunter Formation northwest of Coll de Ser.

Following upon the basal layers, the main mass of the Bunter Formation consists of an alternation of red micaceous sandstones, mudstones and siltstones and a few shales. Coarse conglomeratic sandstones are more frequent in the lower part of the formation, whereas silty material abounds near the top (App. III, Section 47). Sedimentary structures such as cross-bedding, graded bedding, ripple marks and load-casts frequently occur.

The uppermost layers of the formation consist of a thin but typical layer of reddish, greenish and grey-black shales which are conformably overlain by the gypsum or limestones of the Pont de Suert Formation.

Thickness. – The thickness of the Bunter may range from 0 m (locally, north of Coll de Ser) to 350 m (Segre Valley).

Fossils and age. – *Esquisetum arenaceum*, the only recognizable fossil ever reported in this formation, found by

Dalloni (1930, p. 115) near Guils, 5 km west of our area, indicates a Triassic age.

Depositional environment. – Nagtegaal (1969) concluded deposition on a pediment under semi-arid climatological conditions. Low relief provided fossil slope breccias locally.

MIDDLE AND UPPER TRIASSIC

Pont de Suert Formation

In the map area, the Pont de Suert Formation (Mey et al., 1968) is either developed as an alternation of mainly limestone and gypsum beds or as a complete limestone sequence. The formation almost always overlies the Bunter Formation and is generally covered either by the Bonanza Formation or the Adrahent Formation. Previous work in the map area has been carried out by Dalloni (1930), Boissevain (1934), Guérin-Desjardins & Latreille (1961) and Diederix (1963). A detailed description of this formation in a more westerly area was given by Mey (1968).

Lithology. – The contact between the Pont de Suert Formation and the Bunter is always sharp but conformable; only locally, north of Coll de Ser, does the formation unconformably rest upon Devonian limestones (Hartevelt & Roger, 1968). This contact, which is unique in the map area, is comparable to those described by Mey (1967, p. 178–179) in the Ribagorzana and Baliera valleys.

In general, two facies areas may be distinguished: in the north the entire formation is made up of limestone unconformably overlain by the Adrahent Formation; only a very thin layer of gypsum may occur locally. In the south (App. III, Section 48), the formation generally consists of an alternation of gypsum, limestones and some dolomite beds, together with variegated marls in the middle of the formation. As a result of the extreme incompetency of the gypsum, the formation is generally very much disturbed, limestone slabs being distributed chaotically in the other rocks.

The limestones are generally black on the fresh surface, but always weather to light grey. Burrowing tracks have been frequently observed (Fig. 21). The gypsum deposits are generally white or grey and may locally display reddish colours. Marls and shales may show reddish, green or black colours. Dolomites are often cavernous and are not restricted to any particular stratigraphic level.

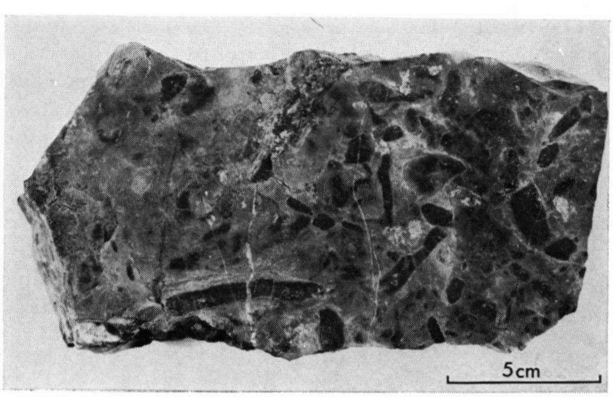


Fig. 21. Burrows seen on the bedding plane of limestone of the Pont de Suert Formation at Coll dé Arnat.

In the area between Tora and Noves de Segres, dark green bodies of crystalline basic rocks consisting of plagioclase and clinopyroxene and alteration products of epidote, prehnite, aerinite and chlorite are to be found. Because of their ophitic texture, these rocks are commonly referred to in literature as ophite. The ophite bodies are typically restricted to the Pont de Suert Formation, they never rise into the overlying Jurassic or Cretaceous formations nor have they been observed in older rocks. The ophites occur either as round bodies suggesting an intrusive origin or, as may be observed east of Hostalets, as thin sheets completely parallel to the bedding (App. III, Section 48). The lower sheet of igneous material illustrated in the section consists of loose crystalline debris in which rounded pebbles of black limestone occur, arranged more or less parallel to the general stratification (Fig. 22). This type of mixture of igneous and country rock suggests possibly pyroclastic deposition on a partly emerged and eroded surface and subsequent redeposition from a mudflow (possibly a lahar). The limestone clasts almost certainly appear to have been derived from the same formation and, as ophites are the only igneous rocks known in the Triassic, the matrix is most probably their tuffaceous equivalent. We may therefore conclude that the ophites are very probably partly intrusives and partly extrusives of Middle to Upper Triassic age.

Fig. 22. Limestone conglomerate in ophite debris in the Pont de Suert Formation, Tora valley.

In the southern part of the map area the upper boundary of the formation is formed by a conformable contact between the gypsum below and the limestones of the Bonanza Formation above. However, this boundary is usually tectonically disturbed and the exact relation with section 48 cannot be given.

Thickness. – A thickness of about 450 m is estimated in the area east of Hostalets. The limestone sequence in the Sierra del Cadi reaches a maximum thickness of 230 m.

Fossils and age. – Burrows may occur in the limestones. A dacicladacae is recorded by Guérin-Desjardins & Latreille (1961, p. 924) in one of the upper limestones, indicating a Triassic age. The Middle and Upper Triassic age which is commonly suggested for the Pont de Suert Formation is, however, mainly based on the lithological resemblances with the Middle and Upper Triassic of germanic facies of other mountain chains such as the Catalan Mountains (Virgili, 1958).

Correlation. – The Pont de Suert Formation should in general be correlated with deposits described by others as Muschelkalk and Keuper. As the sequence in our area does not show a predominantly limestone unit below one of variegated marls and evaporites the classical subdivision would be misleading here.

Depositional environment. – Shallow marine, partly evaporitic.

JURASSIC AND LOWER CRETACEOUS

Bonansa Formation

The Bonansa Formation (Mey et al., 1968) is a limestone, marl and dolomite sequence only developed in the La Vansa area. It always rests upon the Pont de Suert Formation and is overlain by the Prada, Adrahent or Bona Formation. Previous work in the map area has been carried out by Dalloni (1930), Guérin-Desjardins & Latreille (1961) and Diederix (1963).

Lithology. – The formation can always be subdivided into two or three members. In the Segre valley (App. III, Section 49) and at the Coll d'Arnat the lower part consists of very fossiliferous yellowish-brown argillaceous limestones and dark-grey marls with thin limestone intercalations. The lower contact with the gypsum of the Pont de Suert Formation is generally strongly disturbed. In the eastern part of the La Vansa valley, the fossiliferous marls and limestones are generally underlain by light-grey thinly bedded limestones, breccious limestones and dolomites (Guérin-Desjardins & Latreille, 1961, p. 924). Locally grey limestones and dolomites may also be present at this level in the Segre valley (Dalloni, 1930, p. 156).

The upper member of the formation is a black coarse-grained dolomite which rests, with a sharp but conformable contact, upon the fossiliferous marls. A black limestone bed may occur in the middle of the dolomite member.

The upper boundary of the formation may be sharp but conformable if overlain by the Prada Limestone. A sharp and unconformable contact occurs if the formation is overlain by the Adrahent or Bona Formations in which case the formation may be very incomplete. Bauxites were reported from below this unconformity, by Bataller (1943).

Thickness. – In the map area a maximum thickness of 375 m is reached near Josa del Cadi, further south it increases to 550 m (Guérin-Desjardins & Latreille, 1961).

Correlation. – Guérin-Desjardins & Latreille (1961) applied the following names to the three members described:

black dolomites (J_2)	– Dogger (J_{1-2})
fossiliferous marls and lst. (J_1)	– Lias (L_{3-6})
grey lst. and breccious lst. (J_1)	– Infra Lias (L_{1-2})

Fossils and age. – Brachiopods, lamellibranchs, belemnites and ammonites are abundant in the marls and argillaceous limestones. Fossil lists were given by Dalloni (1930, p. 157–158). The age of the fossiliferous member varies from Domerian to Aalenian. The lower non-fossiliferous limestones and breccious limestones are commonly considered to be lowermost Lias. The black dolomite member ranges from Bajocian to Lower Oxfordian (Dalloni, 1930, p. 170; Peybernès, 1968).

Prada Limestone

The Prada Limestone Formation (Mey et al., 1968) is named after the Sierra de Prada, a mountain ridge southwest of Noves de Segre. The formation is always developed between the Bonansa Formation and the Llusa Marls, all three of which are restricted to the La Vansa area. Previous work in the map area has been carried out by Dalloni (1930), Rios (1951, 1956), Grekoff et al. (1961), Guérin-Desjardins & Latreille (1961), Rat (1966) and Peybernès (1968).

Lithology. – A well-exposed very thick sequence may be observed in the narrow gorges of the Rio Segre south of Hostalets (App. III, Section 49). The limestones are generally fossiliferous and are black or dark-grey; at the weathered surface, however, they have a light-grey appearance.

The Prada Formation always rests, with a sharp but conformable contact, upon the black dolomites of the Bonansa Formation. The lower part of the formation as exposed in the northern gorge, consists mainly of black limestones in which thin black dolomite intercalations may occur near the base.

The middle part of the formation consists of black breccious limestones followed by light-grey argillaceous limestones and a thin layer of black carbonaceous shales. These rocks are exposed in the depression around the confluence of the Segre and La Vansa rivers.

The upper part of the rock sequence, in which the southern gorge is eroded, begins with a number of light-coloured coarse bioclastic limestones followed by a very thick sequence of black micritic limestones. The upper boundary is a sharp but conformable contact with the Llusa Marls.

Thickness and lateral extension. – The Prada Limestone is only developed in two separate basins generally bounded by Cretaceous fault or flexure zones (Fig. 23).

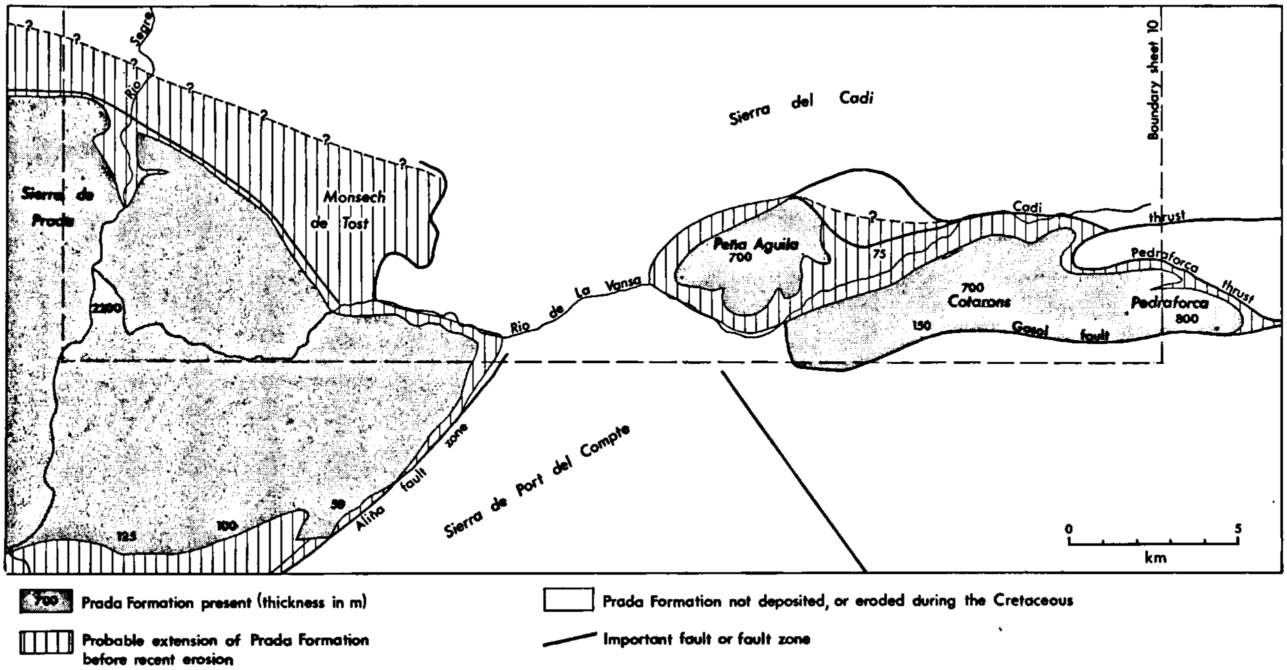


Fig. 23. Distribution and thickness of the Prada Formation.

In the large western basin, in the southwestern corner of the map, the limestones attain a maximum thickness of 2000 m. In the small eastern basin of Gosól the thickness does not exceed 800 m. Towards the borders of these basins the limestones may decrease in thickness, but they are often cut off by the boundary faults and are absent in the bounding areas. A discussion of the limited occurrence of these deposits is given by Ashauer (1934, p. 226–230), Rios (1956), Rat (1966) and Souquet (1967, p. 35–42). Considerable differences in thickness show that important basin subsidence took place during the Malm and Lower Cretaceous.

Fossils and age. – Foraminifera, algae, ostracods, lamellibranchs, brachiopods, bryozoans and crinoids are to be found. Fossil lists were given by Grekoff et al. (1961) relating to localities in the Segre valley, south of Josa del Cadi and north of Gosól, and by Peybernès (1968) relating to the Segre valley. The ages concluded range from (upper) Oxfordian to Aptian and the stages distinguished by Peybernès (1968) along the Segre valley are indicated in section 49 (App. III).

Correlation. – The Prada Limestone, which up to 1961 was regarded as an Aptian rock unit, has often been referred to as "Urgonian" or "Urgo-Aptian". Grekoff et al. (1961), who discovered the Neocomian and the Malm in the lower reaches of the formation, called this lower part "Série Intermédiaire". Guérin-Desjardins & Llatreille (1961) called the rock unit below the breccious limestones "Série de Passage" or "Série de Transition" (J_3–n_3), whereas the rest of the formation was referred to as "Urgonien" (n_4). Peybernès (1968), who investigated the Segre section in more detail, described only the upper part of the Neocomian limestone as being of Urgonian facies.

Depositional environment. – The presence of green algae and the absence of detritic material indicate a shallow marine environment and a very moderate relief of possibly nearby land areas (Rat, 1966, p. 125).

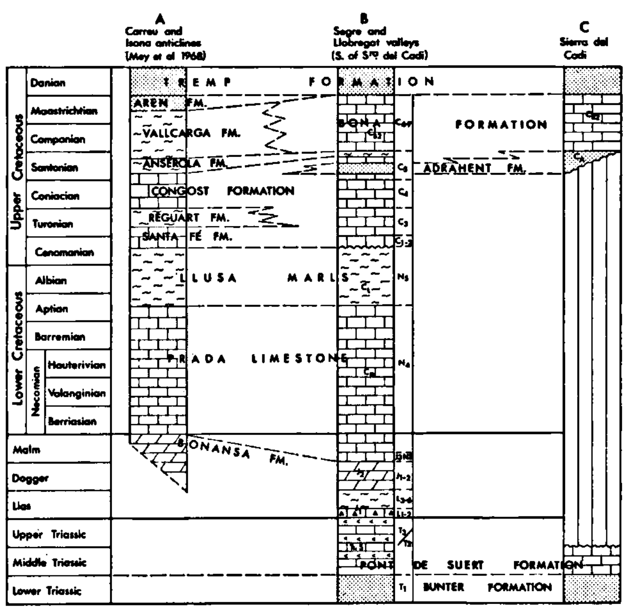


Fig. 24. Correlation diagram for Triassic, Jurassic and Cretaceous formations.

Llusa Marls
The Llusa Marl Formation, (Mey et al., 1968) is a very thick marl sequence developed between the Prada and Santa Fé Limestones (Fig. 24). On the map only the lowermost part of these deposits is shown. Previous work south of the map area has been carried out by Dalloni (1930) and Souquet (1967).

Lithology. – The Llusa Marls rest, with a sharp but conformable contact, on the Prada Limestone. The lower part of the formation consists of a monotonous sequence of greyish marls in which ammonites may be found. A detailed description of the formation in the Sant Fé syncline is given by Souquet (1967, p. 48–53).

Thickness. – The thickness is about 1500 m in the Segre valley, further west it increases to more than 2500 m (Souquet, 1967).

Fossils and age. – Ammonites occur in the lower part of the formation. Higher levels, exposed south of the map area, are very fossiliferous and yield foraminifera, echinoids, lamellibranchs, ostracods and spicules of sponges. For further details reference is made to Souquet (1967, p. 50–51), who concluded an age ranging from Aptian to lower Cenomanian.

Correlation. – The Llusa Marls should be correlated with the "Aptien Albien" (n_5) of Guérin-Desjardins & Latreille (1961) and with the "Marnes noires" or "Marnes à spicules" (C_{4-1}) of Souquet (1967).

Depositional environment. – Marine environment with considerable influx of detritic material from emerging land areas into the subsiding basin (Souquet, 1967, p. 60–66).

UPPER CRETACEOUS, PALEOCENE AND EOCENE

Cenomanian, Turonian and Coniacian deposits are absent in the map area, which hiatus represents an important period of emergence that probably already began in the Lower Cretaceous, as is indicated by the development of bauxite in the Bonansa Formation directly below the unconformity with the Upper Cretaceous (Battaller, 1943; Souquet 1967). The deposition of quartz conglomerates above it announces a general transgression.

Adrahent Formation
The Adrahent Formation, named after the village of Adrahent (Mey et al., 1968), is a sequence of quartz conglomerates and sandstones, in general overlying rocks ranging from Lower Triassic to Santonian age and covered by the Bona Formation or Tremp Formation. Guérin-Desjardins & Latreille (1961) used the term "Série d'Adrahent" for this rock unit.

Previous work in the map area has been carried out by Ashauer (1934), Boissevain (1934), Guérin-Desjardins & Latreille (1961), Diederix (1963), Souquet (1967) and van Hoorn (1970).

Lithology. – In the map area the formation always rests, with sharp and unconformable contact, on the Pont de Suert Formation in the Cadi area and on the Bonansa Formation in the La Vansa area. The formation consists of white quartz conglomerates and sandstones in which some quartz-wacke and black shales may be intercalated. Feldspar is present among the smaller grains. Plant remains mainly occur concentrated in black streaks in the more fine-grained sandstones (App. III, Section 50). A thin yellow-brown dolomite with some quartz and feldspar pebbles may occur locally at the base (Diederix, 1963). At the top a gradual transition exists into the sandy limestones of the Bona Formation, the boundary is chosen where the carbonate content is visible in the field.

Thickness and lateral extension. – A maximum thickness of 180 m is observed near Adrahent, but the formation is generally much thinner. The formation occurs in a narrow zone extending from the Fresser valley in the east, where a thin layer of these conglomerates can be found between the Bunter below and the Tremp Formation above (Ashauer, 1934, p. 234; Trouw, 1969), to the Sierras Zone in the southwest where these deposits rest on the Jurassic strata (Souquet, 1967, p. 271–273). In the Cadi area the formation is generally present (except northwest of Cavá). In the La Vansa area, however, the development of the formation becomes irregular and south of it and in Sierra de Aubens the formation wedges out in a Santonian limestone sequence (Fig. 24-B), (Guérin-Desjardins & Latreille, 1961, p. 927; Souquet, 1967, p. 270–271).

Fossils and age. – Plant remains, the only fossil material observed in the map area, did not allow of any determinations. Foraminifera recorded from calcareous material in this formation and below and above it in the Sierra de Aubens, the Sierras Zone and south of the La Vansa valley, all indicate a Santonian age (Guérin-Desjardins & Latreille, 1961, p. 927; Souquet, 1967, p. 270–273). Boissevain (1934, p. 70–73), who dated the entire overlying Bona Formation as Maastrichtian, suggested a Campanian age for the Adrahent Formation. The conglomerates may very well be strongly diachronic.

Correlation. – The Adrahent Formation should be correlated with the "Campanien" of Boissevain (1934, p. 71) and with part of the "Santonien" (C_5) of Guérin-Desjardins & Latreille (1961, p. 927), (Fig. 24). Souquet (1967, p. 270–273) described similar deposits in the Sierra de Aubens and the Sierras Zone as "Santonien-c7G", and van Hoorn (1970, p. 129–132), who also studied these deposits, correlated them with the conglomerates at Adrahent. The correlation with the formations established by Mey et al. (1968) in more westerly regions is shown in Fig. 24.

Depositional environment. – Van Hoorn (1970, p. 129–132, 146–147) suggested a non-marine deposition and southwest of the map area, where foraminifera have been found in the material, deposition from a longshore current. The coarse terrigenous material (mainly quartz and some feldspar) was probably derived from the Andorra granodiorite as this massif most probably cropped out at that time. The southwesterly current direction deduced by van Hoorn from cross-bedding is in accordance with such a source area.

Bona Formation

The Bona Formation, named after the Rio Bona (Mey et al., 1968), is a limestone unit generally overlying rocks ranging in age from Stephanian to Santonian and covered by the Tremp Formation.

Previous work in the map area has been carried out by Dalloni (1930), Ashauer (1934), Boissevain (1934), Guérin-Desjardins & Latreille (1961), Diederix (1963), Souquet (1967) and van Hoorn (1970).

Lithology. – In the Cadi area, only the upper part of the formation (C_{B2}) is developed. Here the quartz conglomerates and sandstones of the Adrahent Formation grade into the sandy limestones of the Bona Formation (App. III, Section 51). The formation consists mainly of biomicritic and argillaceous limestones and some marl intercalations. The generally almost black limestones near the top may locally also be sandy. The boundary with the overlying Tremp Formation is generally sharp but conformable.

In the La Vansa area an older rock unit (C_{B1}) has also been locally found. This rock unit, which is described by Guérin-Desjardins & Latreille (1961, p. 927), comprises from top to bottom: marls with echinoids, black limestones, sandy limestones, light-grey limestones and represents the Santonian. The varying lithology of these rocks shows that several formations may be distinguished, but as these rocks have not been investigated by the author, they have been provisionally ranked with the Bona Formation (C_{B1}).

Thickness. – The formation reaches a maximum thickness of 2000 m in the region southwest of Gosól, in the Sierra del Cadi the thickness is generally less than 300 m (Fig. 25).

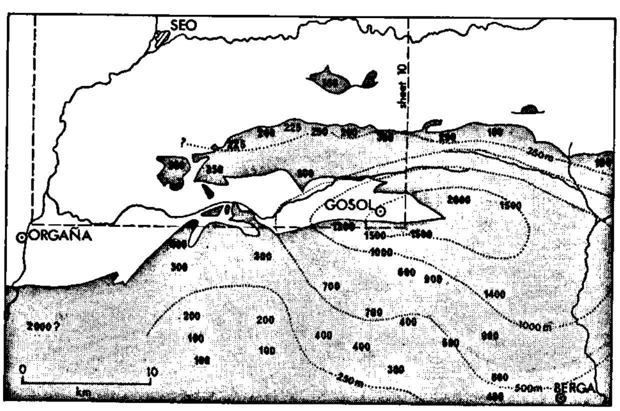


Fig. 25. Isopach map of the Bona Formation, compiled from data of Guérin-Desjardins & Latreille (1961).

Fossils and age. – The very fossiliferous limestones of the Bona Formation (upper part, C_{B2}) may yield: lamellibranchs (mainly rudists), corals, foraminifera, bryozoans and algae. Extensive fossil lists were given by Dalloni (1930, p. 207) and Boissevain (1934, p. 72–73). Dalloni and Guérin-Desjardins & Latreille concluded a Campanian-Maastrichtian age, whereas Boissevain dated these limestones in the Cadi area as Maastrichtian. The lower part of the formation (C_{B1}) is of Santonian age (Guérin-Desjardins & Latreille, 1961).

Correlation. – As may be seen in Fig. 24, the upper part of the formation (C_{B2}) should be correlated with the Aren Sandstone and the Valcarga Formation to the west. The C_{B2} unit is the equivalent of the "Campanien-Maestrichtien" (C_{6-7}) of Guérin-Desjardins & Latreille (1961) and the "Campano Maestrichtien" (c9–8C) of Souquet (1967).

The C_{B1} unit is the equivalent of the "Santonien" (C_{4-5}) of Guérin-Desjardins & Latreille (1961). The marls with echinoids in this lower unit can probably be correlated with the Anserola Formation in the west (Mey et al., 1968) and the sandy limestones with the Adrahent Formation.

Depositional environment. – The fauna in the Bona Formation suggests a shallow marine environment (Boissevain, 1934, p. 70; van Hoorn, 1970, p. 132). Considerable basin subsidence permitted thick deposits to be formed south of the map area.

Tremp Formation

The Tremp Formation (Mey et al., 1968) is a sequence of largely continental deposits which in the map area is always developed between the Bona Formation and the Cadi Formation.

Previous work in the map area has been carried out by Boissevain (1934), Guérin-Desjardins & Latreille (1961) and Diederix (1963).

Lithology. – The formation rests, with a sharp but conformable contact, on the limestones of the Bona Formation; black shales are often developed directly above this contact.

Two members are distinguished in the Tremp Formation (App. III, Sections 52 and 53). In the Cadi area the lower member (CT_1) consists mainly of red and violet shales, cross-bedded sandstones and some conglomerates. Locally calcareous sandstones and shales and limestone nodules may occur. South of the map area marls, limestone and lignites have also been reported in this member (Guérin-Desjardins & Latreille, 1961).

The upper member (CT_2) always begins with a grey limestone bed above which predominantly marls and calcareous deposits occur, whereas sandy deposits are very scarce. The marls have mainly reddish or violet colours, the limestones being grey or slightly reddish. In the northern part of the Sierra del Cadi and in the Turo de Call Pubill the uppermost part of the formation consists almost entirely of limestone in which only thin streaks of reddish shale occur.

The upper boundary is taken at the top of the uppermost red shales or, if limestones predominate in the upper part, at the transition from limestones in which foraminifera are absent or rare into limestones in which they abound.

Thickness. – A thickness of 1500 m was recorded by Guérin-Desjardins & Latreille (1961) southeast of the map area. Towards the north the thickness decreases to less then 500 m in the Sierra del Cadi and 200 m in the Turo de Call Pubill.

Fossils. – No fossils have been found, although to the south of the map area *Microcodium* and *Chara* were recorded in the lower member (Guérin-Desjardins & Latreille, 1961, p. 927).

Correlation. – The lower member (CT_1) should be correlated with the "Garumnien inférieur" (C_8) of Guérin-Desjardins & Latreille (1961), the upper member (CT_2) with the "Garumnien supérieur" (e_{1-3}) and the limestone at the base of the upper member with the "Calcaire de Rognac" (r).

Depositional environment. – The majority of the deposits are continental, but the important limestone development at the top in the Cadi area is probably marine.

Age. – Because of its position between the Maastrichtian of the Bona Formation and the Upper Thanetian of the Cadi Limestones, the age of the Tremp Formation may range from Danian to Lower Thanetian.

Cadi Alveolina Limestone

The Cadi Alveolina Limestone Formation (Mey et al., 1968), is named after the Sierra del Cadi of which this formation forms the crest. The formation, which is largely composed of a thick sequence of limestones with an abundance of foraminifera, is always developed between the Tremp Formation and the Roda Formation.

Previous work in the map area has been carried out by Dalloni (1930), Boissevain (1934), and Guérin-Desjardins & Latreille (1961).

Lithology. – The lower boundary of this transgressive formation is generally sharp. In the northern part of the Cadi area the foraminifera bearing limestones rest directly on the limestones of the Tremp Formation (App. III, Section 53), but in the region between the Monsech de Tost and Fornols the sequence begins with limestone conglomerates and locally quartz conglomerates with a foraminifera bearing limestone matrix. Several conglomerate levels can be seen intercalated in the lower reaches of the formation (App. III, Section 52). Owing to their slightly reddish colour, the pebbles must have derived from the limestones of the Tremp Formation.

The main mass of the formation in the Sierra del Cadi, described by Boissevain (1934, p. 76–77), consists of grey limestones with occasional marls at the base and sandy limestones at the top (Section 54).

The formation is conformably overlain by the Roda Formation.

Thickness. – A maximum thickness of 1600 m was recorded southeast of the map area by Guérin-Desjardins & Latreille (1961). Towards the north it decreases to less than 1000 m in the Sierra del Cadi.

Fossils and age. – Foraminifera (mainly *Alveolina* and *Nummulites*) are very abundant throughout the formation, oysters occur in the lower reaches and gastropods have been recorded in the upper part of the sequence. An extensive fossil list was given by Boissevain (1934, p. 76–77). The age of the Cadi Limestone may range from Upper Thanetian to Lower Lutetian (Guérin-Desjardins & Latreille, 1961, p. 298; Hottinger, 1960, quoted by these authors).

Correlation. – The Cadi Alveolina Limestone is the equivalent of the "Calcaire à Alvéolines" (e_4) of Guérin-Desjardins & Latreille (1961).

Depositional environment. – Shallow marine.

Roda Formation

The Roda Formation (Mey et al., 1968) is a sequence of grey marls developed between the Cadi Alveolina Limestone and the Collegats Conglomerates (south of the map area). A short description is given by Guérin-Desjardins & Latreille (1961, p. 928), and partly quoted below.

Lithology. – The formation, which rests conformably on the Cadi Limestones, consists mainly of grey, often sandy, marls in which sandstone intercalations, some conglomerate layers and argillaceous limestones may occur. Black bituminous material may also be present in the marls. Coquina may occur, rich in oysters, *Nummulites* or small gastropods. Gypsum may be developed in the lower part of the formation.

Thickness. – Ranges between 200 and 1000 m.

Fossils and age. – Oysters, foraminifera and gastropods. Foraminifera are of Upper Lutetian to Bartonian (Upper Ledian) age.

Correlation. – The Roda Formation is the equivalent of the "Série bleue" and the "Lutétien Priabonien inférieur" (e_5) of Guérin-Desjardins & Latreille (1961)

Depositional environment. – Shallow marine, partly evaporitic.

OLIGOCENE

During and after the Pyrenean folding phase in the late Eocene, involving the uplift of the axial zone, thick piemont deposits have accumulated north and south of the mountain chain.

Collegats Conglomerate
The Collegats Conglomerates (Mey et al., 1968) consist of a sequence of conglomeratic piemont deposits partly covering a pronounced topography and partly conformably overlying the Roda Formation.

Previous work in the map area has been carried out by Birot (1937), Guérin-Desjardins & Latreille (1961) and Diederix (1963).

Lithology. – On the Purrédon, north of Montant, where the deposits have their widest extent, the conglomerates unconformably overlie the Prada Limestone, badly rounded boulders and pebbles of which occur in the well-lithified basal layers. Above these basal layers the conglomerates are poorly cemented and contain fragments of the Cadi Alveolina Limestone, Devonian limestone, Ophite and Bunter, all derived from nearby outcropping formations. A small occurrence of conglomerates at a much lower topographic level south of Castellá, shows that deep valleys existed during deposition, which were filled up with these conglomerates.

South of the map area, two different sequences of these conglomerates may be observed: a lower sequence which conformably overlies the Roda Formation and a higher sequence which unconformably overlies the lower conglomerates and all older formations. This intraformational unconformity indicates that the deposits are partly syntectonic (Birot, 1937, p. 176–199; Guérin-Desjardins & Latreille, 1961, p. 927).

Thickness. – As the formation covers a pronounced topography and the upper boundary is formed by the present topographical surface, no thickness can be estimated. A thickness of more than 2000 m can be read from sections of Guérin-Desjardins & Latreille (1961) south of the map area.

Fossils and age. – No fossils were found in the map area. The only fossil locality known in the Pyrenees, near Sosis, indicates a Ludian age for the base of the formation (Dalloni, 1930, p. 246). The upper Ledian age of the Roda Formation below is in accordance with this age. It is generally assumed that the major part of the formation was deposited during the Oligocene (Misch, 1934, p. 60–65; Ashauer, 1934, p. 268).

Correlation. – The Collegats Conglomerates are the equivalents of the "Priabonien supérieur Oligocène" (g) of Guérin-Desjardins & Latreille (1961).

Depositional environment. – Fluvial piemont deposits (Dalloni, 1930; Misch, 1934; Birot, 1934; Nagtegaal, 1966).

NEOGENE

In Neogene times the pronounced Oligocene topography was eroded to a gently undulating landscape. Tectonic activity caused the intramontane graben of the Conflent basin in France, the Cerdaña and Bellver basins just east of the map area and the Seo de Urgel basin to subside. In the basins lacustrine, fluviatile and slope deposits were deposited; in the Seo de Urgel basin two formations can be recognized: the Piedra and Ballestá Formations.

Previous work on the Neogene deposits of the Seo de Urgel basin has been carried out by Chevalier (1909, 1910), Dalloni (1930), Boissevain (1934), Birot (1937), Solé Sabaris & Llopis Lladó (1946) and Diederix (1963).

Piedra Formation
The Piedra Formation, named after the Bco. de Piedra, is defined as a sequence of poorly lithified argillaceous sediments, locally sandy or conglomeratic, in which characteristic lignite seams may occur. The formation, which is of Miocene age, rests on Paleozoic rocks and is always overlain by the Ballestá Formation.

Lithology. – Chevalier (1909) described, along the Bco. de Piedra (Rio de la Bastida), the only section in which almost the complete formation was exposed. As the clays are exploited the outcrop has changed, but the rock units of Chevalier can still be fragmentarily recognized. The lower contact is poorly exposed, but the coarse basal conglomerate is probably in contact with a fault plane. Chevalier found from base to top (App. III, Section 59):1. Basal conglomerates, overlain by red, white and grey clays with lignite seams and plant remains; 2. Red sandy clays and fine conglomerates; 3. Dark-red sandy clay; 4. Alternation of coarse sands, sandy clay and fine conglomerates; 5. Red, white and grey clays with thin lignite seams and plant remains. In the Bco. de Piedra, the contact with the overlying Ballestá Formation, which is not exposed, is probably an angular unconformity.

Other exposures of the formation only show the upper part of the sequence. At Firal (App. III, Section 58, northeast of the junction of the main roads in Seo de Urgel and now almost completely built up) and to the north (App. III, Section 57), the red clays of the top of the formation yielded a rich fauna of mammals (Chevalier 1909, 1910; Vidal 1875, quoted by Chevalier). Just east of Alás, dark-red clays with some lignite layers also yielded plant remains and mammals (Chevalier, 1910; Solé Sabaris & Llopis Lladó, 1946, p. 44).

Thickness. – About 100 m in the Bco. de Piedra.

Fossils and age. – Plant remains may be found throughout the formation. Remains of mammals have been recorded by Chevalier (1909, 1910) and Vidal (quoted by Chevalier, 1909) from the upper part of the formation. Chevalier (1909) dated the formation as

Tortonian, whereas Birot (1937) considered it to be Sarmatian.

Correlation. – Comparable deposits in the Bellver basin are the "Assise profonde" (Sarmatien) and the "Assise inférieure" (Pontien) (Astre, 1927; Boissevain, 1934, p. 79).

Depositional environment. – Intramontane lacustrine and paludal environment (Chevalier, 1909; Boissevain, 1934, p. 82; Birot, 1937; Solé Sabaris & Llopis Lladó, 1946).

Ballestá Formation

The Ballestá Formation, named after the village of Ballestá, is defined as a sequence of poorly lithified conglomeratic and breccious sediments generally resting unconformably upon the Piedra Formation or older rocks.

Lithology. – In the Seo de Urgel basin the formation rests on a gently undulating surface, comparable to the present topography of the basin. In the basin sedimentary zones may be distinguished (Appendix V), characterized by a distinct mode of deposition and source area. Breccias with an orange clay matrix are often developed at the base. West of Ballestá the formation consists completely of these deposits which contain only badly rounded fragments from the underlying Seo Formation and from the Rabassa Conglomerates outcropping nearby and probably represent fossil slope deposits. Southeast of Seo de Urgel mainly clays and fine conglomerates with much red Permian and Lower Triassic material from the south were deposited. Apparently this part of the basin was not reached by the main streams of the Segre river to the north. A N-S trending central zone is characterized by very coarse conglomerates often with well-rounded pebbles and with boulders up to 1 m in size. The conglomerates are for a large part composed of granodioritic material and some Devonian limestone, which was presumably transported from the Upper Valira basin by a torrential river. Along the sides of the central zone the conglomerates may overlie the slope deposits, but more towards the centre they generally rest directly on the Cambro-Ordovician rocks.

Between Pla de San Tirs and Hostalets two terraces occur along the Segre river in which conglomerates identical to those of the Ballestá Formation may be observed. As they lie somewhat isolated they may very well represent Quaternary terraces.

In the La Vansa Valley, small isolated occurrences of conglomeratic and breccious material can also be found, often with a reddish or yellowish clay matrix. Because of their lithologic similarity they are correlated with the Ballestá Formation as was also done by Guérin-Desjardins & Latreille (1961) who grouped them in the "Mio-Pliocène".

Thickness. – A maximum thickness of 80 m was observed east of Ballestá (App. III, Section 55).

Fossils and age. – Remains of mammals found in the basal layers of the formation (App. III, Section 55, 58) indicate a Pontian age (Chevalier, 1909). Higher levels did not yield any fossils, but are generally considered as Pliocene (Chevalier, 1909; Boissevain, 1934; Solé Sabaris & Llopis Lladó, 1946).

Correlation. – In the Bellver basin the "Assise supérieure" (argiles rutilantes) is comparable with the clayey breccia of the Ballestá Formation, whereas the "Pliocène supérieure" or "Sicilien" is similar to the coarse conglomerate (Boissevain, 1934). In the Conflent basin the Codalet Formation (Pontien-early Pliocene) and the Escaro Formation (late Pliocene-Quaternary) show a great similarity to the clayey breccia, whereas the Ternère Formation (late Pliocene-Quaternary) may best be compared with the coarse conglomerate in the Seo de Urgel basin (Oele et al., 1963). In the northern Pyrenees the Lannemezan Formation (Pliocene?) may be compared with the Ballestá Formation (Taillefer, 1951).

Depositional environment. – The clayey breccias were probably deposited from a mud flow descending from slopes into the basin, whereas the coarse conglomerates were deposited in a torrential river (Oele et al., 1963).

QUATERNARY

Quaternary deposits have only been mapped when obscuring older deposits and structures over large surfaces. Glacial deposits, rock glaciers (p. 231), landslides, alluvial fans, scree and alluvium are indicated on the map. Scree indicated in the granodiorite massif may partly represent glacial deposits, as these have only been mapped from aerial photographs. Areas indicated as not exposed generally represent cultivated or wooded slopes.

CHAPTER II

INTRUSIVE AND METAMORPHIC ROCKS

The southern part of the Andorra-Mont Louis granodiorite in the region of Lles-Aristot has been described in detail by Roggeveen (1929). This work presents much data about the igneous rocks in the main body and the dykes as well as the rocks in the contact metamorphic aureole.

Granodiorites

Four bodies of granodiorite occur in the map area: the western part of the large Andorra-Mont Louis granodiorite and three smaller ones in the Llavorsi syncline: the Santa Coloma, Fontaneda and Montescladó granodiorites.

The granodiorites are quite homogeneous masses of light-grey, unoriented rocks of medium grain-size. The main constituents are quartz, plagioclase (often strongly zoned, labradorite-oligoclase), potassium feldspar, biotite and locally hornblende and pyroxene. Zircone, apatite, titanite and ore occur as accessories. Plagioclase, biotite and hornblende are idiomorphic, whereas quartz and potassium feldspar fill up the interstices. Sericite, chlorite and clinozoisite often occur as alteration products.

Dark fine-grained xenoliths of varying shape and size are generally abundant. They consist mainly of biotite, hornblende and plagioclase and some quartz and potassium feldspar.

Fine-grained granodiorite with a slightly porphyritic texture may often be observed near the contacts. The small Montescladó granodiorite is entirely fine-grained and, in addition to the four normally observed constituents, it contains hornblende and pyroxene which also occur elsewhere close to the contacts (Roggeveen, 1929; Zwart, 1965).

Dykes

Aplites. – Aplite dykes occur almost exclusively in the granodiorite massifs and their immediate surroundings. The dykes, which are light-grey or slightly pink, mainly consist of quartz and potassium feldspar. Some dykes also contain albite or tourmaline, sericite and biotite.

Lamprophyres. – Lamprophyry dykes occur abundantly in the granodiorite bodies as well as in the pre-Hercynian sedimentary rocks. The dykes, which may be grey, dark-green or brown, are fine-grained and often porphyritic. The lamprophyres may range in composition from kersantite containing plagioclase, biotite and some quartz to spessartite displaying plagioclase, hornblende, pyroxene and some quartz. The dykes are often considerably altered and secondary minerals such as chlorite, sericite and calcite are generally present.

Granodiorite porphyries. – Porphyritic dykes of about the same composition as the larger granodiorite bodies are abundant in the map area. The dykes are generally grey, brownish or greenish and are often strongly altered displaying sericite, chlorite, clinozoisite and calcite.

Quartz-porphyries. – Quartz-porphyry dykes have occasionally been found in the northern part of the map area. These dykes, which are generally almost white, consist mainly of some large quartz crystals in a very fine-grained largely indeterminable ground mass, in which quartz and muscovite may be recognized.

Contact metamorphic rocks

The granodiorite bodies always caused a contact metamorphic aureole which may vary in width from 150 to 1500 m. In general, however, the smaller bodies have only narrow contact zones (e.g. the Montescladó granodiorite, max. 200 m) and as a rule dykes usually hardly show any contact metamorphic zones.

Slates have been converted into spotted slates in the outer part of the contact zone, whereas hornfelses only may occur close to the granodiorite contact. The spotted slates, in which altered andalusite and cordierite occurs, still show their original texture and cleavages and bedding can be recognized. In the hornfelses the texture became a completely unoriented intergrowth of quartz, cordierite, andalusite, biotite and muscovite. Andalusite and cordierite have generally been altered into sericite and muscovite and biotite has often been chloritized.

In the carbonaceous Silurian slates contact metamorphism caused chiastolite and occasionally andalusite crystals. These slates are never converted to real hornfelses and often show remarkably little recrystallization in the contact zone.

Pure limestones have been recrystallized into marbles with coarse calcite crystals. Impure limestones and calcareous slates have often been converted into lime-silicate rocks in which garnets (grossularite-andradite), vesuvianite, diopside-hedenbergite and epidote are most frequent.

CHAPTER III

STRUCTURE

INTRODUCTION

The Pyrenean mountain chain is an essentially E-W trending structure, consisting of an axial zone, deformed mainly during the Hercynian orogeny, flanked on both sides by marginal troughs, deformed mainly during the Pyrenean folding phase (Middle Alpine). The northern marginal trough is bounded to the north by the Aquitanean basin, the southern marginal trough is bounded to the south by the Ebro basin (Fig. 1).

Along the southern border of the axial zone, between the Esera valley in the west and the La Vansa valley in the map area, a narrow zone, in which isolated Hercynian blocks surrounded by post-Hercynian rocks were folded together during the Pyrenean folding phase, is known as the "Nogueras zone".

Stable and unstable periods of varying intensity occurred in the area. Folding and faulting which took place during the Upper Carboniferous (probably during the Lower Westphalian) will be referred to as the "Hercynian deformation". The term Hercynian has been used in this restricted sense, as no appropriate name for an orogenic phase in the Lower Westphalian exists and because it has become common usage in literature on the Pyrenees. The deformation during the Eocene is the Pyrenean folding phase (Middle Alpine) and the deformations during the Oligocene and Neogene are referred to as the post-Pyrenean deformations (Late Alpine). Important epirogenic movements as during the Permian and Lower Cretaceous have been indicated with the corresponding period names.

HERCYNIAN STRUCTURES

The Hercynian orogeny in the Pyrenees can be split up in several deformation phases. This has been proved by many investigations in other regions of the axial zone, among others by Zwart (1963), Boschma (1963), Oele (1966) and Mey (1967), who mainly used the trends of folds to differentiate between a number of folding phases. In the area investigated there is good evidence that multiple folding is also involved in the development of the structures. This is based mainly on several sets of cleavages, on the occurrence of folded folds and interference structures and on the distribution of lineations. The final scheme of structural events worked out is only partly similar to those published previously by Zwart (1963) and Boschma (1963).

TABLE 1

Terms used:

ss	sedimentary stratification
s_0	cleavage predating F_1
s_1	cleavage belonging to F_1
s_2	cleavage belonging to F_2
s_3	cleavage belonginh to F_3
s_4	cleavage belonging to F_4
F_1, F_1', F_1''	1st folding phase and subphases
F_2	2nd folding phase
F_3	3rd folding phase
F_4	4th folding phase
δ_2	intersection of ss and s_2
B_1	fold axis of F_1
B_2	fold axis of F_2
B_3	fold axis of F_3
B_4	fold axis of F_4
a	direction perpendicular to the fold axis and in the axial plane
b	direction parallel to the fold axis
c	direction perpendicular to the axial plane
X	longest axis of the deformation ellipsoid
Y	intermediate axis of the deformation ellipsoid
Z	shortest axis of the deformation ellipsoid

Cleavages and folds

In argillaceous rock units of all pre-Hercynian formations a strong slaty cleavage (s_0) occurs, which is invariably parallel to the bedding (ss) (see Table 1). The slaty cleavage consists of a parallel arrangement of micaceous minerals, mainly sericite, muscovite and chlorite. The distance between the cleavage planes is determined by the basal cleavage of these micas. The parallel orientation of the mica flakes causes a shiny surface which can easily be recognized in the field. The parallelism of s_0 and ss can best be observed in the Seo Formation in which the bedding is perfectly expressed by the regular alternation of thin sandy and argillaceous layers (Fig. 26).

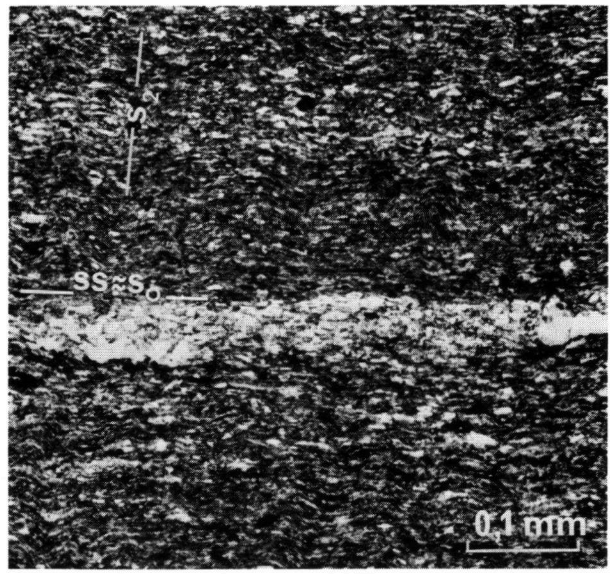


Fig. 26. Thin section of slates from the Seo Formation with slaty cleavage minerals statistically parallel with the bedding, gently crenulated by F_2 deformation (crossed nicols).

At first sight there appears to be a small angle between ss and s_0 but more careful examination of many thin sections reveals that this is due to the effects of later crenulations. Folds with axial planes parallel to s_0 have not been observed.

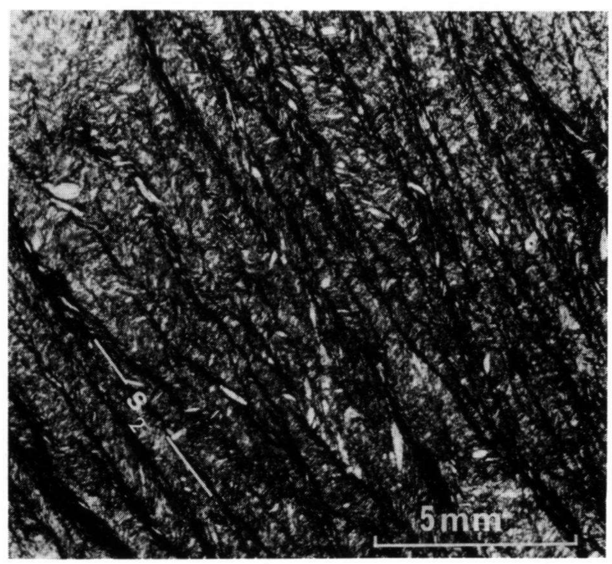


Fig. 27. Thin section showing well-developed s_2 crenulation cleavage in slates of Seo Formation (crossed nicols).

Crenulations in s_0 are always to be attributed to a later folding phase, which will be associated with later folds called F_2 because they can be shown to refold older (F_1) folds, as can be deduced from steeply plunging fold axes and lineations. This crenulation or microfolding of s_0 is very vague in some areas (Fig. 26) but is generally accompanied by cleavage more or less parallel to the axial planes (Fig. 27) which will henceforth be called s_2. This crenulation cleavage generally has spaces of less than 1 mm between the planes which often show lateral movement parallel to them. They conform precisely to the definition of crenulation cleavage by Rickard (1961).

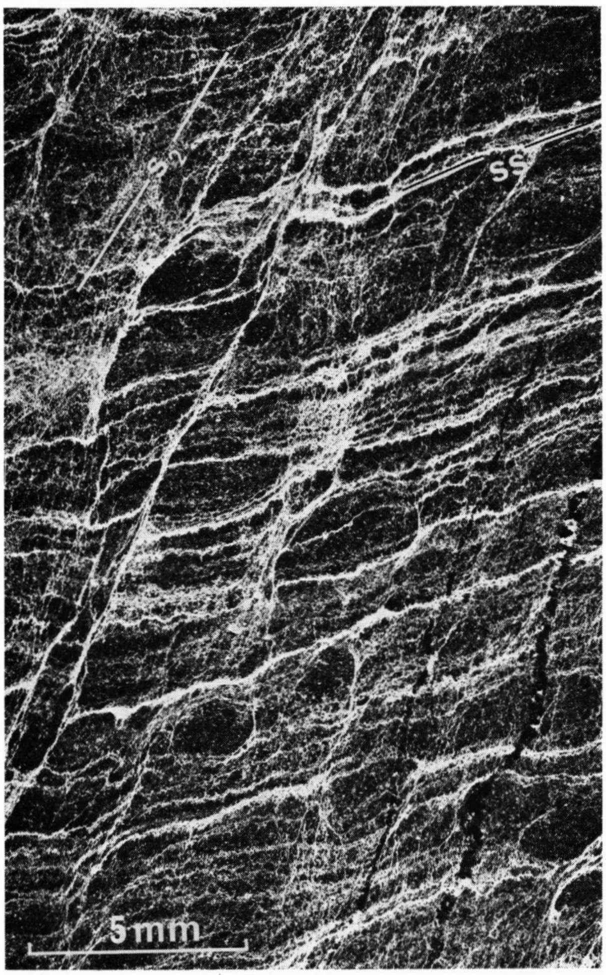


Fig. 28. Thin section showing s_2 fracture cleavage in fine-grained sandstone of Civis Formation (negative print).

Locally the s_2 cleavage is folded and consequently these folds should belong to a later generation. They are only developed in well-cleaved rock units and always occur in narrow zones, dipping 45°–60° to the northwest. A fracture cleavage (s_3) may occur in these folds, which often developed as kink-bands (Fig. 29). Fold axes plunge steeply to the north or northwest.

Another set of kink-bands with E-W axes and south-dipping axial planes also folds the s_2 cleavage. Locally these kink-bands occur as a conjugate set with a horizontal and a vertical component. Sometimes these structures look more like concentric folds and have south-dipping axial planes.

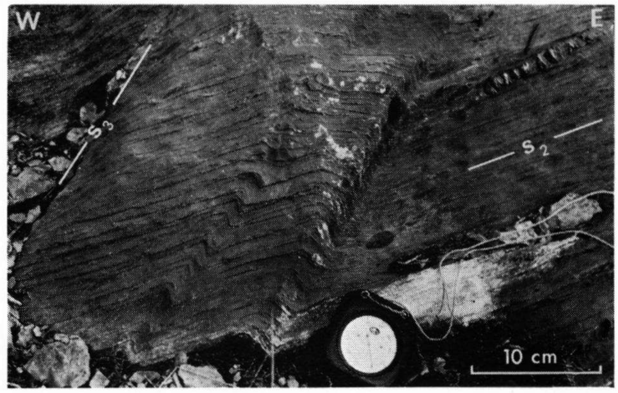


Fig. 29. F_3 fold in Rueda Formation west of Asnurri.

Recumbent folds accompanied by numerous sub-horizontal faults and fractures are interpreted as gravity structures later than F_2.

Fig. 30. Disharmonic F_2 folds in Seo Formation.

The development of folds and s_2 cleavage in the rocks in this area is clearly related to the lithology of the various rock units: A well-developed s_2 crenulation cleavage can be observed in the pure slates of the Ansobell Formation, the Fonchanina Formation, the Civis Formation and in the Silurian. In these homogeneous rock units, which usually show little or no bedding, folding can only be seen in thin section as a crenulation of s_0.

Fig. 31. Thin section showing mixed F_2 folds and deviations of s_2 in the Seo Formation (negative print).

In the Seo Formation, compression of the regular alternation of thin slate and sandstone beds caused a well-developed folding (Fig. 30 and 31), with amplitudes dependent on the thickness of the beds, mainly up to 15 cm. Generally mixed (parallel and similar) folds are developed, zones with many sandstone beds show more affinities to concentric folds, whereas in more slaty zones the folds are more frequently of the similar type. The crenulation cleavage, developed in slate beds between parallel folded sandstone beds, can often be seen deviating strongly from the attitude of the axial plane (Fig. 31). Such deformation patterns have already been extensively discussed by many others, e.g. de Sitter (1954), Kleinsmiede (1960), Hara (1966, 1967) and Dieterich (1969).

In the more coarse-grained and thick-bedded rocks, as often observed in the Cavá Formation and occasionally in the Bellver and Civis Formations, only a fracture cleavage developed with no small-scale folding (Fig. 28). The fracture cleavage observed conforms to the definition given by Rickard (1961).

In the Rabassa conglomerate, some cleavage developed in the matrix, which tends to curve around the pebbles. Many pebbles are still spherical or angular and do not seem to have been flattened (Fig. 3). However, a number of flat pebbles are oriented in the cleavage, but according to the author they have retained their original shape and have merely been rotated into the cleavage. Small-scale folding is absent from this formation.

In the Bar Quartzite no cleavage occurs. As a result of its position between the relatively incompetent Ansobell slates and the highly incompetent Silurian slates, this thin quartzite band has been able to develop folds with wavelengths of about 100 m, disharmonious with the adjoining formations.

The limestones, especially of the Devonian, are particularly strongly folded into folds with wavelengths mostly between 50–200 m. The development of the s_2 cleavage in these folded limestones depends on their clay content. When this is high crenulation cleavage is well-developed, with a low clay content there are s_2 "solution bands", along which the limestone has been dissolved, leaving an insoluble residue of parallel mica flakes. Similar limestone cleavages in other regions are described in more detail by Plessmann (1964, 1966), Richter (1965), Langheinrich & Plessmann (1967) and the same phenomenon will be described in detail by Roberti (in press) from the Devonian of the Pyrenees.

Structural units

A very important structural feature, as always in the Pyrenees, is the disharmony between the folding style in the Cambro-Ordovician rocks below and the Devonian and older Carboniferous rocks above the Silurian, a disharmony made possible by the extreme plastic behaviour of the Silurian shales, acting as a "plan de décollement" between the two rock sequences mentioned above. The Cambro-Ordovician rocks form the largest structures which normally attain wavelengths of 6 to more than 20 km. In the large synclines the relatively thin Devonian and Carboniferous rock sequence is intensively folded into folds with wavelengths of 50 to 200 m. In some synclinal areas there are Devonian rock masses that must have slid down from a higher topographic position because they can be seen to bear little relationship to the underlying structures. These differences in structural style are the basis of the subdivision into structural units as shown in Fig. 32.

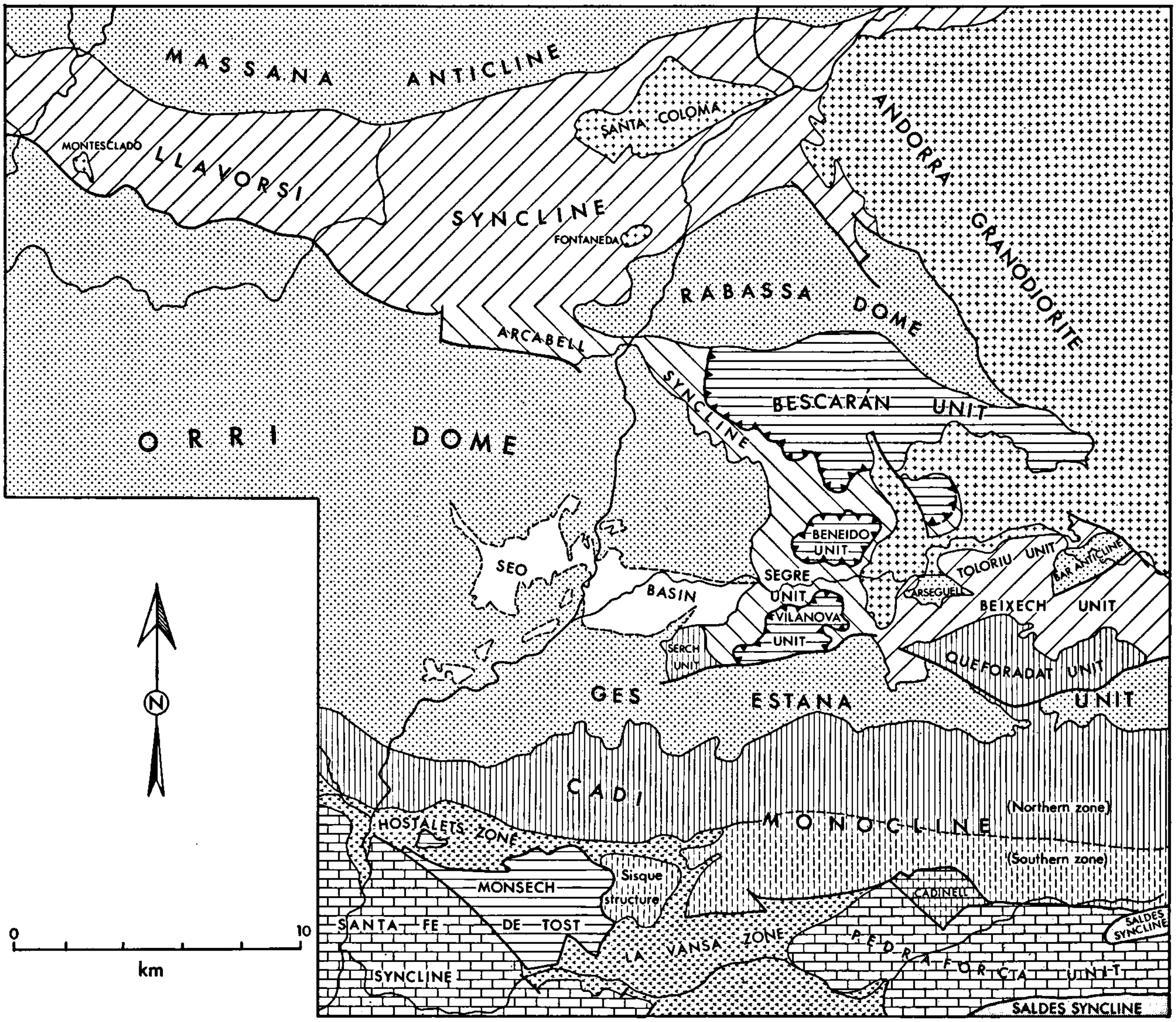


Fig. 32. Structural units.

Massana anticline

The Massana anticline (Zwart, 1965) is not completely covered by our map. It is overturned to the south, the overturned south flank being bounded by the Llavorsi syncline and the north flank by the Tor-Casamanya syncline (Fig. 33). To the west the Massana anticline continues into the Salat-Pallaresa anticlinorium (Oele, 1966) or Central anticline (Zandvliet, 1960). The overturned south flank can be observed in the low-lying areas in the valleys of the Rio Lladorre, Rio d'Os and Riu Valira del Nort. In the higher regions, north of Conflens, the southern part of the crest of the anticline is visible, showing an overall dip of 35° to the south.

The s_0 slaty cleavage is, as always, parallel to the bedding while the strike of the s_2 cleavage in the southern border of the anticline also follows the broad curvature of the anticline itself, varying from a direction of 115° in the west to 80° in the east (Appendix IV).

The folds related to s_2 only occur in the crestal part of the anticline (cross sections 15–24), which indicates that the large anticlinal structure predates the small folds. These folds are parasitic and could only develop in the crest of the large anticline, where the angle between bedding and the largest stress direction was still small enough to allow folding. In the overturned flank this angle was too large, so that flattening of the bedding plane took place instead and folds could not develop here (de Sitter, 1964).

It might be expected that $\delta_2$lineations (intersection between ss and s_2) would be parallel to the general direction of the large folds, but this is generally not realised. Steep plunging N-S $\delta_2$lineations are common and small N-S folds also occur locally in the Seo Formation (Fig. 34A). Hence two folding phases have

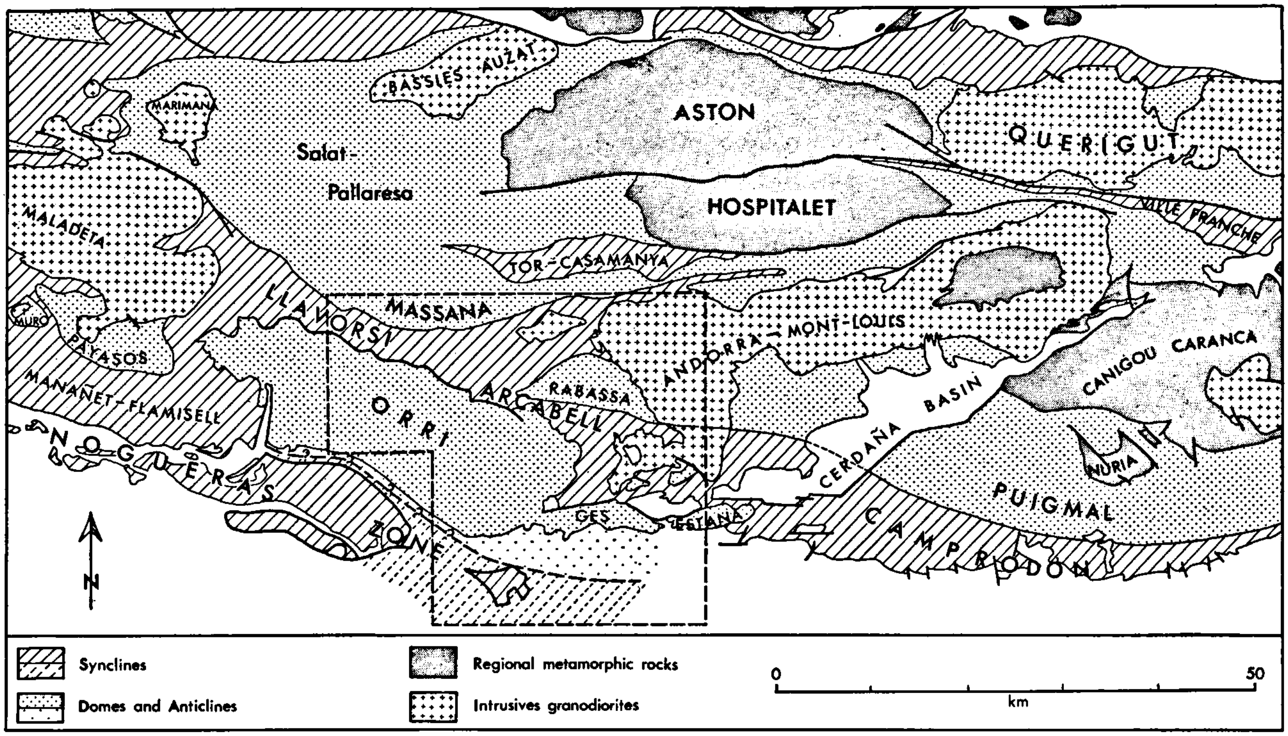


Fig. 33. Structural setting of Hercynian structures of sheet 10.

been involved with divergent axial planes. In the first phase, folds with limbs must have formed which were later intersected by a steep s_2 cleavage of a slightly different strike, thus causing steeply plunging δ_2-lineations. Only locally was the angle between s_2 and ss large enough for the formation of small N-S F_2 folds.

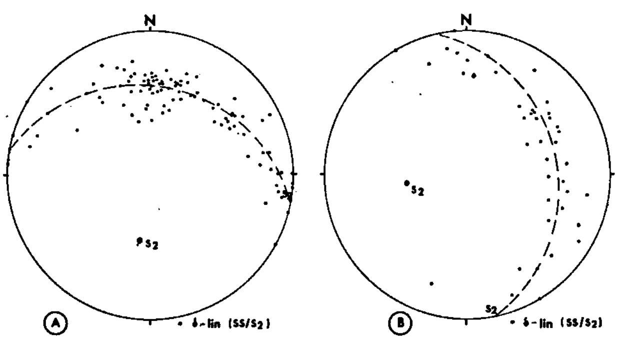


Fig. 34. Orientation diagrams of δ_2 lineations for areas of N-dipping s_2 planes (A) and E-dipping s_2 planes (B).

Some minor folds trending NE-SW with an axial plane fracture or crenulation cleavage (s_3) dipping about 45–60° NW, intersect and fold s_2 confirming that these are later.

South-dipping kink-bands (s_4) are particulary abundant in this anticline (see also Zandvliet, 1960).

N-S to NW-SE vertical crossfaults in the south flank and crest of the anticline can possibly be explained as the result of stretching which took place in the curved south flank of the structure.

Rabassa dome

The Rabassa dome is in fact the western termination of the largest Cambro-Ordovician dome structure of the eastern Pyrenees, comprising large gneiss domes as, for instance, the Canigou-Carança massif. The Rabassa dome is separated from its main mass by the Andorra granodiorite (Fig. 33).

The overall structure of this part of the dome shows a rather gentle north flank dipping 20°–40° to the NNW and a vertical to slightly overturned E-W trending south flank. North of Argolell the dome punges about 30° to the west.

A well-developed slaty cleavage (s_0) is seen parallel to the bedding and the s_2 cleavage generally dips 50°–55° to the north with some local deviations. In this dome the s_2 cleavage converges to the west, but does not completely curve around to a N-S position as in the Orri dome.

Steeply plunging N-S δ_2 lineations and occasional minor N-S folds indicate a slight difference in strike between two folding phases, which, as in the Massana anticline, can be referred to as F_1 and F_2.

Faulting parallel to s_2 is usually reverse upwards towards the south but some normal faulting has also taken place.

The up thrusting west of the Valira is particularly interesting, as it is restricted to the Cambro-Ordovician formations below the Bar-Quartzite and displacements of about 150 m in the Rabassa Conglomerate completely disappear in the slates of the Ansobell Formation and leave the Bar Quartzite undisturbed.

E-W kink-bands are frequently seen folding the s_2

planes into vertical or steeply north- or south-dipping attitudes. In the south flank of the dome a system of vertical N-S cross-faults occurs, with systematic apparent throw down to the east of as much as 500 m. The eastern part of the dome is largely cut off by a system of vertical NW-SE faults, with downward movements of the north-east block of as much as 1000 m, which brings the Seo Formation into contact with the Silurian, or the Upper Ordovician with the Upper Devonian.

Orri dome

The Orri dome, only the north-eastern part of which is shown on the map, was named by Schmidt (1931). In the north it is bordered partly by the Llavorsi syncline and partly by the Arcabell syncline, in the east by the Silurian and Devonian of the Segre unit and in the south-east it has its continuation in the Ges-Estana unit. In the south the dome is unconformably covered by post-Hercynian formations.

To the north this Cambro-Ordovician structure is mostly bounded by the large WNW-ESE striking Llavorsi fault, which considerably obscures the original form of this contact. In the Valira valley, however, the fault runs into the dome itself, and north of the fault the normal contact with the Silurian of the Arcabell syncline can be observed. Here it becomes apparent that the north flank of the dome has in fact a gentle dip (20°–25°), which only very gradually grows steeper towards the north. The normal contact with the Llavorsi syncline must also have been very gentle, as suggested by the dip of the Upper Cambro-Ordovician formations just south of the Llavorsi fault.

Bedding and s_0, parallel as usual, are folded together into folds generally showing an axial plane crenulation cleavage (s_2). These F_2 folds may be of varying size and order. The large mappable folds, approx. 1000 m in wavelength, can mainly be traced in the upper Cambro-Ordovician formations. Smaller folds, ranging from 100 m to microscopic scale, can be observed in the banded slates of the Seo Formation.

The trend of s_2 is approximately parallel to the outline of the dome itself (Appendix IV). In the northern part of the dome we find an E-W trending s_2 which gradually swings into a N-S trend northeast of Seo de Urgel, whereas the dip varies between 30° and 65°. Fold axes of large F_2 folds also swing around from an E-W to a N-S trend and folds in the Devonian of the Arcabell syncline and Segre unit swing around in the same manner. Fold axes of smaller F_2 folds and δ_2 lineations, however, may have any orientation within the s_2 plane (Fig. 34). These relationships between structures show that there is no structural difference between the northern and eastern borders of this Cambro-Ordovician unit, and suggests a dome structure in the area between the Segre and Valira.

Repetitions in the stratigraphic succession, northeast of Seo de Urgel (Fig. 35), are due to thrusting from east to west, more or less parallel to s_2. Presumably folding could not easily develop in the thick conglomerate and sandstone sequence of the Rabassa and Cavá Formations.

Locally, west of the Valira, a NE-SW oriented cleavage with a NW dip has been observed, which presumably represents s_3.

Morphological features illustrate that an important NW-SE fault system has developed in the Orri dome as is especially clear in the region of Castellbó, where most of the large valleys are in this direction. This fault system is possibly related to a verical NW-SE oriented cleavage, associated with folds of the same trend, observed south of Seo de Urgel.

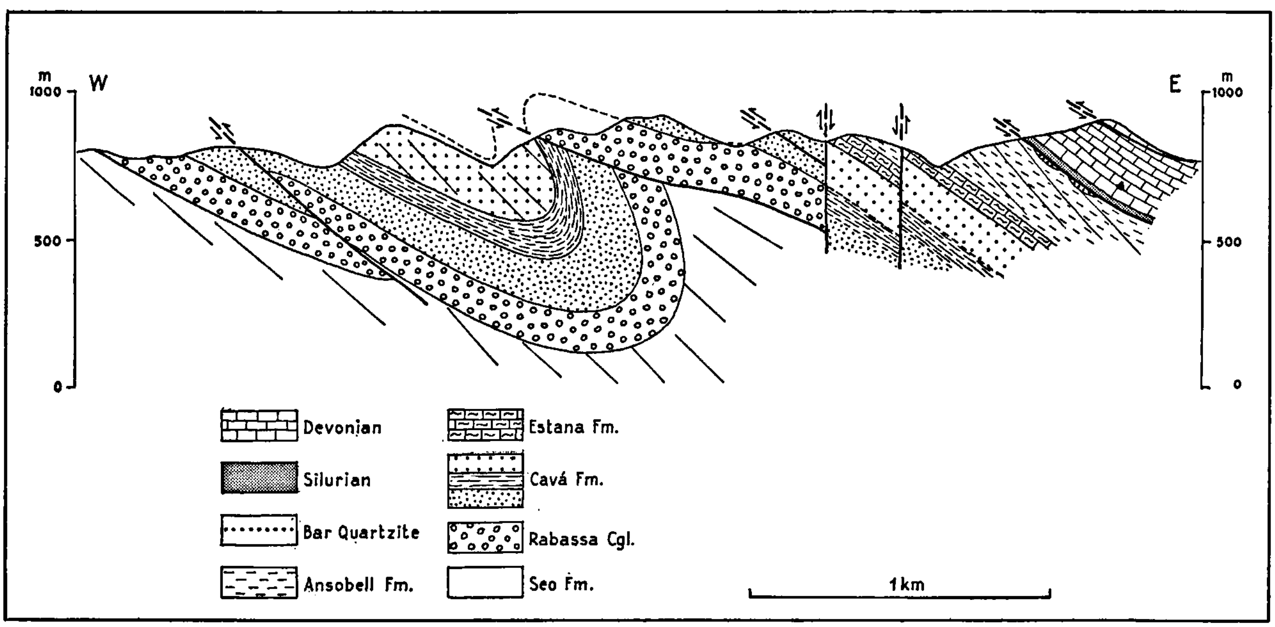


Fig. 35. E-W cross-section through the Cambro-Ordovician along the Segre valley between Seo de Urgel and Torres.

Ges-Estana unit

The Ges-Estana unit is the eastern continuation of the Orri dome. It can be described as a separate unit, lying as it does somewhat outside the main part of the Orri dome. In the north the unit is bounded by the Ortedó and Cavá faults, both probably of Neogene age. In the south the Cambro-Ordovician is unconformably overlain by post-Hercynian rocks. The Ges-Estana unit as a whole shows a very gentle axial plunge to the east, and about 5 km east of our map it plunges below Silurian and Devonian rocks.

The slaty cleavage s_0, when present, is parallel to the bedding; often, however, the sediments are too coarse to have allowed its development. Bedding and s_0 are folded together in approximately E-W trending folds with an s_2 axial plane crenulation or fracture cleavage dipping about 45° to the north. As can be seen on Appendix IV, fairly large variations in strike and dip of s_2 occur, although frequently no s_2 cleavage is developed at all. This may partly be due to the lithology, as most rocks belong to the coarse-grained Cavá Formation, but it can also be partly explained by the less severe compression that took place in these southern areas. F_2 folds, well-developed in the region of Ges, are asymmetrical and show anticlines with steep short south flanks and long gentle north flanks which together with the north dipping s_2 cleavage suggest a parasitic relation to the north flank of a larger anticline. More to the east the folds die away gradually so that finally, east of Estana, only one large south-dipping flank is left. Occasional upthrusting to the south occurs parallel to s_2 cleavage planes. Very gentle N-S folds shown on the E-W cross section of Fig. 36 may either represent axial plunges of F_2 folds or may be related to post-Hercynian N-S faults, as developed extensively in the eastern part of the Arcabell-Camprodon syncline, east of the map area (Fig. 33) (Brouwer, 1968; Bloemraad, 1969; Trouw, 1969).

Arseguell-Bar anticline

The relatively small Arseguell-Bar anticline is exposed at three different places (Fig. 32).

The westernmost outcrop is found north of Arseguell where practically only the Bar Quartzite and the Ansobell Slates are exposed. The Ansobell Slates, which occupy the major part of the visible structure, are largely covered with scree and along the road north of Arseguell they have been hornfelsed so that no structures or cleavages can be seen. The Bar Quartzite, however, shows the steep south flank of the anticline (cross section 4) and the steep plunge of the structure below the contact-metamorphosed limestones of the Devonian to the west. The large outcrop of Ansobell Slates suggests a gentle north flank cut off by faults to the north and northeast.

South of Pont de Bar a small outcrop of Bar Quartzite was observed which is most probably a part of the north flank of the Arseguell-Bar anticline.

The easternmost part of the anticline, northeast of Bar, exposes formations as deep as the Cavá Formation. The trend of this part of the structure is no longer E-W, but has swung into a SW-NE orientation with s_2 dipping 45° to the northwest. The northwest flank of the anticline has a gentle dip of about 30°, whereas the southeastern border is formed by the subvertical Bar fault bringing Devonian and Carboniferous formations into contact with the Ordovician. In the northeast the structure is cut off by the subvertical Aguila fault, which brings Cambro-Ordovician rocks without visible contact-metamorphism into contact with the granodiorite.

In the fine-grained rocks, slaty cleavage s_0 is well-developed. A poorly developed fracture or crenulation cleavage can also be observed which, because of its orientation, can best be correlated with s_2.

Llavorsi syncline

The Llavorsi syncline forms part of one of the largest synclinal structures of the Pyrenees. The entire syncline is about 100 km long and runs from the Esera valley in the west (sheet 7) to the Ariège valley (France) in the east (sheet 6). The name Llavorsi syncline is

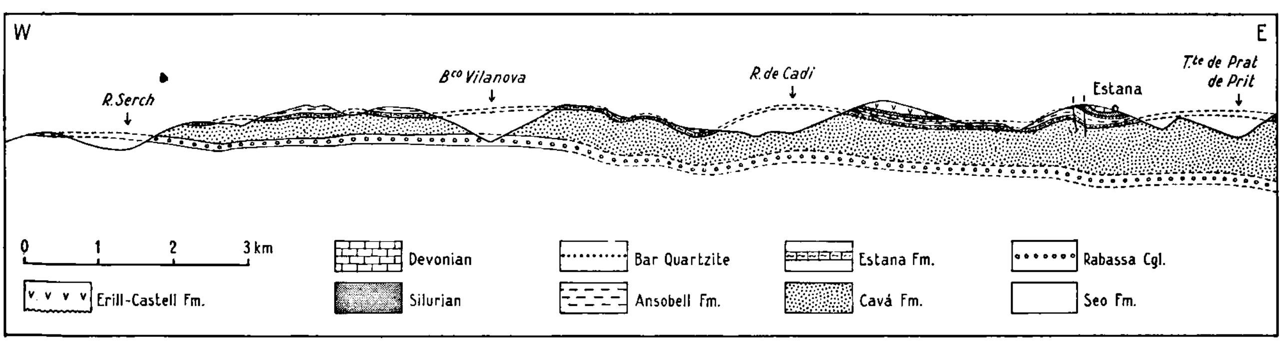


Fig. 36. E-W cross-section through the Ges-Estana unit.

The Hercynian unconformity dipping approx. 45° to the south, shows that a post-Hercynian deformation must have played its part in the structure described. This deformation will be dealt with later.

most commonly used for the eastern portion between the Maladeta batholith and the Ariège valley (Fig. 33), whereas some authors have referred to it as Tirvia-Espot syncline.

Structurally this relatively narrow synclinal zone is completely controlled by the bordering anticlinal structures: in the north the Sallat Pallaresa anticlinorium and Massana anticline, in the south the Orri and Rabassa domes. The northern flank is generally overturned towards the south, dipping about 45°–60° to the north. The south flank, bordering on the Rabassa dome, whows a gentle dip of about 30° to the northwest. The western part of the south flank, however, is very incomplete as is clearly demonstrated in the area around the Bco de Tressot (cross-sections 23, 24). Here the Carboniferous Civís Formation is in contact with Cambro-Ordovician rocks.

The Llavorsi fault, forming the boundary with the Orri dome, has cut off the gentle northward dipping south flank of the Llavorsi syncline.

West of the plunging nose of the Rabassa dome, there is a transition to the Arcabell syncline. To the east the syncline is largely cut off by the Andorra batholith.

The depth of the Llavorsi syncline is still a subject of discussion. A depth of 1 km below sea-level has been suggested by Zandvliet in his cross-sections, but since no traceable horizon has been found in the Carboniferous rocks, any estimate is necessarily hazardous.

Large E-W folds. – Just as in the Cambro-Ordovician rocks, the pelitic rock units of the Silurian, Devonian and Carboniferous have a slaty cleavage (s_0) orientated subparallel to the bedding, which has been folded and crenulated in a later folding phase. The numerous large and small mappable folds which can be found in the Devonian and Carboniferous of the Llavorsi syncline generally have the s_2 cleavage roughly parallel to their axial planes, dipping 35°–55° to the north.

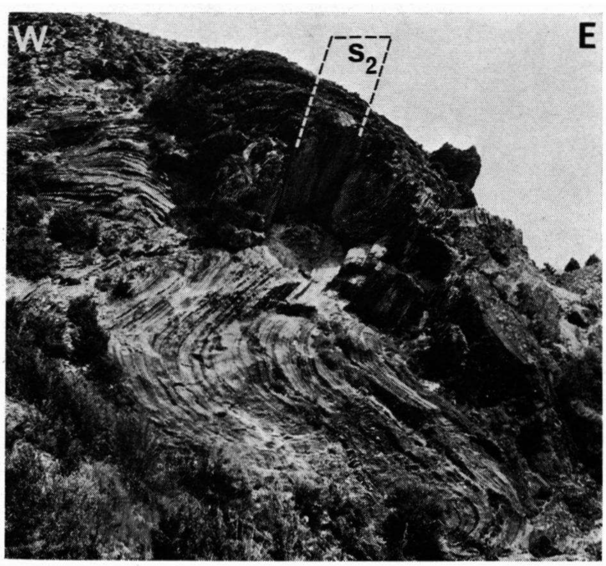


Fig. 37. F_2 fold in a flank of a first phase fold, west of Ars. The F_2 fold axis plunges 45° north.

There are many δ_2 lineations forming a large angle with the fold axes of the large folds. The most likely interpretation of this phenomenon is that, before the F_2 folds and the s_2 cleavage developed, the bedding and s_0 were already deformed in folds having a slightly different orientation. This is confirmed in the area around Ars and Asnurri and SE of Bordas de Conflens, where, in local zones, clearly two sets of folds occur, the latest of which, having s_2 parallel to the axial plane, can be correlated with F_2 (Fig. 37). Consequently, the folded folds must belong to an earlier generation which will be called F_1. As far as could be ascertained F_1 was not accompanied by an axial plane cleavage. The geometry of these folds, which have N-plunging axes, is shown in Fig. 38.

In addition to the folded F_1 folds some fold axes plunging steeply northwards are clearly not related to folding of an F_1 fold about an F_2 axis. Such axes can be seen at all angles to the horizontal and even in inverted sequences so that synclinal antiforms and anticlinal synforms occur. While gentle plunges of axes are to be found throughout the Llavorsi syncline the steeper northern plunges are largely concentrated around Ars and Asnurri as well as east of Conflens. The origin of these axial plunges and their probable relation to the F_1 folding will be dealt with later (p. 217).

In the Llavorsi syncline the s_2 cleavage trends only roughly E-W, in general it is parallel to the boundaries of the large anticlines and domes (Appendix IV). Thus, here, s_2 swings around in approximately the same way as in the Massana anticline which is considered to be an original feature. The strong local deviation of s_2 northwest of the Santa Coloma stock, however, is considered to be a deformation due to a shouldering-aside effect of the granodiorite.

The strong compression, related to F_1 and F_2 ,as can be measured from the folds by means of the ratio between the thickness of strata in fold-hinges and flanks (de Sitter 1964), is more or less uniform throughout the entire Llavorsi syncline, the total shortening amounting to about 70%. The tightness of the folds, however, shows marked differences, which are due to differences in the original thicknesses of the strata involved. This difference in fold style can be observed by comparing the tight folds, formed in the thin strata at Llavorsi (cross-section 27) and south of Os (Fig. 39), with the large, less tight folds developed in the thick strata between Farrera and Conflens (cross section 21 and 22).

NE-SW kink-bands. – In well-cleaved rocks of the Rueda Formation in the area northwest of Asnurri, NE-SW striking kink-bands or kink-bandlike folds are common (Fig. 29). Sometimes a fracture cleavage (s_3) is developed parallel to the kink-bands. Kink-bands and s_3 cleavage dip 55° to the northwest, and fold axes plunge between north and northwest.

E-W kink-bands. – In the Llavorsi syncline E-W kink-bands are widespread, usually dipping to the south. Especially intensive development can be observed near

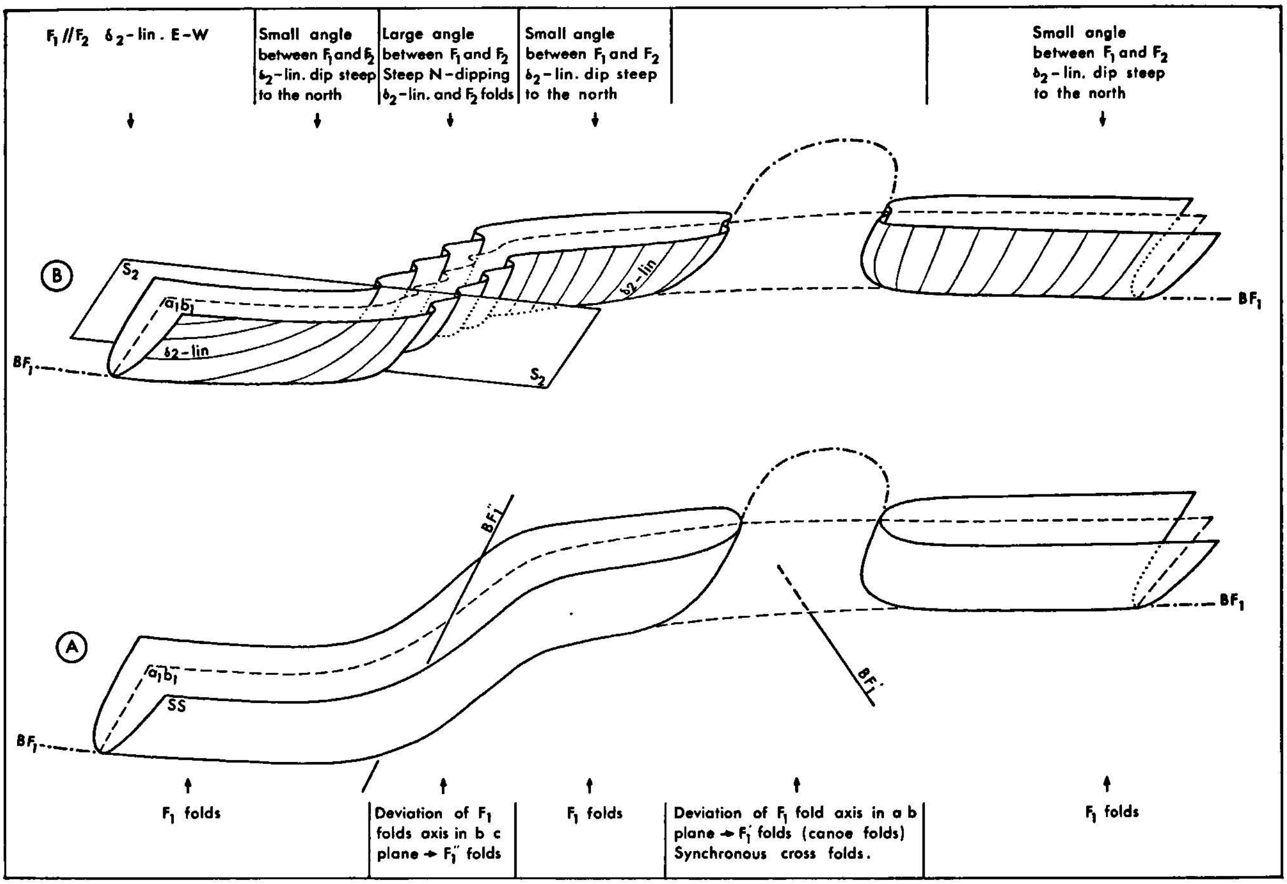


Fig. 38. Superposition of folding phases in the Llavorsi syncline. (A) Condition after first folding phases, (B) condition after the second folding phase.

Fig. 39. Tight fold in the Basibé Limestone, southeast of Os de Civis.

Burch, where conjugate sets have been formed with varying southerly dips (Fig. 40).

Fig. 40. Conjugate set of F_4 kink-bands in Rueda Formation east of Burch.

Slate-diapirs. – In the Llavorsi syncline, between Farrera and the Andorra granodiorite, several large

E-W trending fault zones occur containing Silurian slates. These fault zones, that can locally attain widths of more than 500 m, are more or less parallel to s_2. Normal contacts of the Silurian with the Devonian rocks are only occasionally found and the occurrence of the Silurian slates can therefore generally not be explained by thrusting. As is to be seen on the map, most of these zones are situated in the centre of the syncline and therefore occur in contact with Carboniferous rocks. Since both formations consist mainly of black slates, these fault zones appear in the field primarily as depressions (Fig. 41).

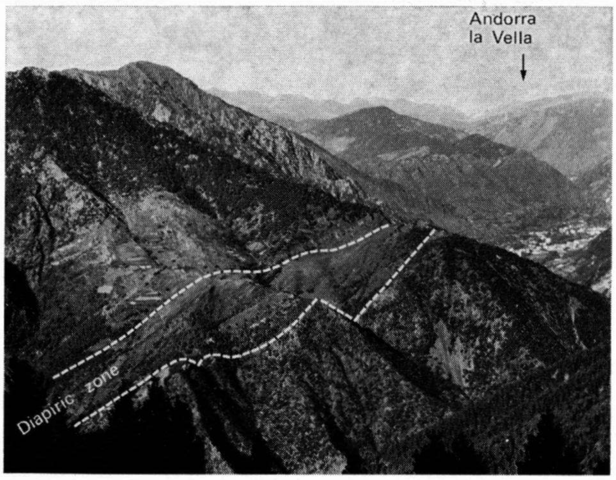


Fig. 41. Diapir of Silurian slates amidst Carboniferous slates between Bixesarri and Andorra la Vella.

In the western part of the Arcabell syncline, a similar fault zone occurs in which a large block of Bar Quartzite was found, northeast of Sant Joan Fumat. This block must have been transported over a vertical distance of at least 350 m. The only reasonable explanation for the emplacement of the quartzite is a diapiric movement of the incompetent Silurian states. Probably these slates were forced upwards from the deepest parts of the syncline, where most of the Silurian material was initially accumulated.

Since these diapiric movements took place approximately parallel to s_2, they must have been contemporaneous with or later than F_2. In some contact-metamorphic areas, not all Silurian slates are metamorphosed. This may on the one hand be due to the high carbon content, making them a metamorphic resistor, or on the other hand to diapiric movements after the granodiorite intrusion.

Sant Julia Flexure zone. – In the Valira valley in Andorra, just north of the Rabassa dome, the ENE-WNW trending structures of the Llavorsi syncline show a distinct kink in their trend, outcropping as they do parallel to the N-S river valley for about 1 km before sharply resuming their regional orientation (Appendix IV). Near Sant Julia, where the Rabassa dome is not affected by this flexure, the disharmony is accommodated in the large masses of Silurian slates which probably intruded diapirically.

NW-SE Faults. – Vertical faults striking NW-SE, systematically throwing the northeasterly blocks downwards, only occur in the eastern-most part of the syncline and constitute the continuation of the essentially similar system that cut off the Rabassa dome.

Arcabell syncline and Segre unit

The Arcabell syncline (Schmidt, 1931), shown on the map, merely forms part of a very large synclinal zone, extending approximately 85 km long from the area of Ars where it splits off from the Llavorsi syncline all the way eastwards to Camprodón (Fig. 33). This large syncline, which will be referred to as the Arcabell-Camprodón syncline, is cut into three separate parts by the Andorra granodiorite and the Cerdaña Basin.

To the north the syncline shows a vertical to overturned contact with the Rabassa dome and east of this area the north flank also remains rather steep. West of the plunging nose of the Rabassa dome there is a gradual transition into the Llavorsi syncline characterized by a zone with large outcrops of Silurian and older Devonian formations. The eastern part of this zone, around Argolell, provides an excellent example of the disharmony between the structures in the Cambro-Ordovician and Devonian formations (Cross sections 15 and 17).

The contact with the Orri dome in the Valira valley, at Farga de Moles, dips steeply north, but at higher levels the attitude of this contact becomes much more gentle. West of the Valira, the Llavorsi fault largely constitutes the southern border of the Arcabell syncline. East of the Valira, the southern contact of the syncline gradually swings around with the structures of the Orri dome until a N-S position is reached in the Devonian of the Segre unit. Arcabell and Segre units grade into one another and no sharp boundary can be drawn. In the Arcabell syncline the E-W direction of F_1 and F_2 predominates, whereas in the Segre unit the N-S direction is more important.

In the south the Segre unit is bounded by the Ortedó fault which forms the contact with the Cambro-Ordovician of the Ges-Estana unit. To the east the Segre unit is cut off by the granodiorite.

On top of the Arcabell and Segre units we find three units, the Bescarán, Beneidó and Vilanova units, which show no direct relationship with the underlying structures and will therefore be dealt with separately.

The structures in the Arcabell syncline, west of the Valira, show the same characteristics as those in the Llavorsi syncline, with E-W trending folds and a well-developed s_2 cleavage. A Silurian slate diapir is developed here as well.

The Devonian of the Arcabell syncline crosses the Valira river at Farga de Moles as a single isoclinal syncline of Rueda Limestone; east of this river and in the region of Arcabell, however, exposures are scarce. Southeast of Arcabell, Silurian and Devonian rocks can be seen which form part of the south flank of the syncline with isoclinal folds comparable to those west of the Valira. The Devonian of the north flank is not exposed here, further to the east it is covered by the Bescarán unit.

Southeast of Arcabell the general trend of structures is WNW-ESE, but also N-S folds occur comparable to those described in the Ars-Asnurri region, presumably produced by the F_2 refolding of F_1 folds. Further to the southeast the structures gradually swing around to a NW-SE direction. Here fold-hinges are broken through by thrusts approximately parallel to their axial planes which dip about 20°–35° to the northeast. More to the south, in the Segre unit, the dip is generally to the east. Folds are rather rare in the eastern part of the Arcabell syncline and in most cases thrusting has caused the repetitions in the stratigraphy. These repetitions almost invariably occur in the same manner, always showing a normal stratigraphic succession from the bottom upwards: Silurian, Rueda, Basibé, Villech, the thrust and then again the same succession etc.

In some larger thrust sheets, however, folds do occur which clearly preceded the thrusting, but which have been preserved in their entirety within the sheets. A good example of such a fold is to be seen in the west slope of the valley of the Tte. de Bescarán shown in Fig. 42.

The incompetent Silurian slates, which acted as a lubricant in the thrust, are often squeezed out altogether in one place and accumulated diapirically in another. Mylonitized limestones often occur in the thrusts and many folds developed in relation with the thrusting (Fig. 43). The Villech Formation, which is

Fig. 43. E-W recumbent folds above thrust plane, east of Estimariu.

also an incompetent stratum, acted as a secondary detachment plane since the competent limestones of the Compte Formation never take part in the thrusting, being generally overthrust as a plate. Folding and thrusting must therefore have transported the Compte Formation to structurally higher levels where it was eroded away (Fig. 44).

The fact that the broad core of the Arcabell syncline

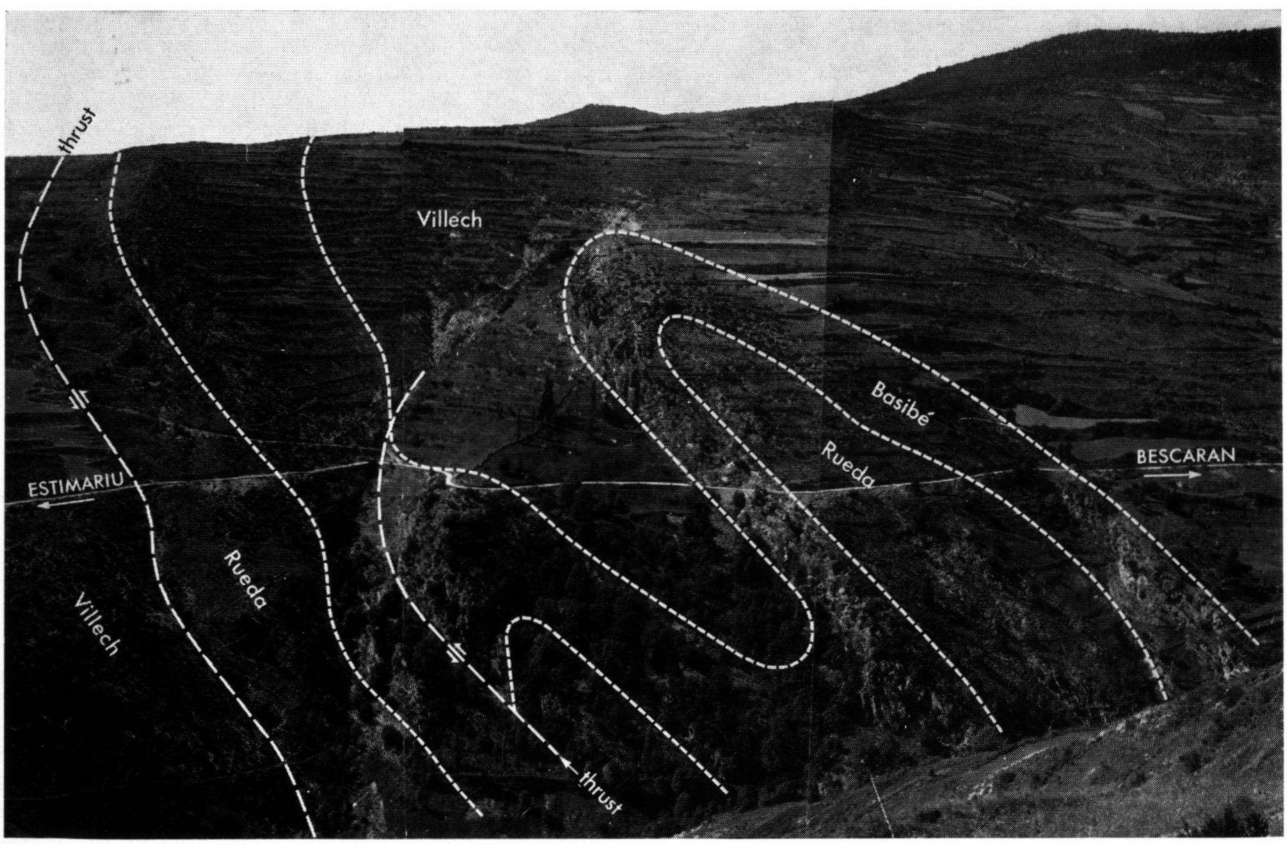


Fig. 42. NW-SE trending fold in the Devonian between Estimaríu and Bescarán.

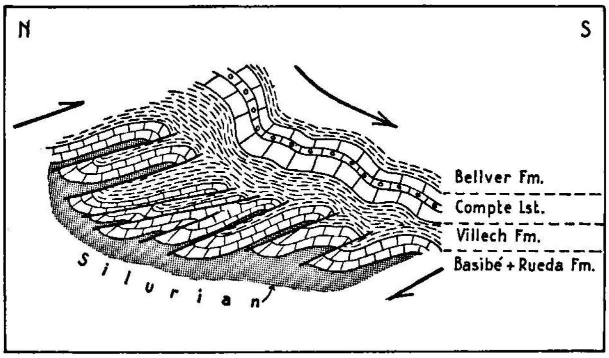


Fig. 44. Sketch section showing the thrusting in the Rueda and Basibé Formations and the related gliding of the Compte and Bellver Formations.

contains only Villech rocks suggests that the depth of this wide syncline in the area of Bescarán must be rather limited and will not be any deeper than at Farga de Moles.

In the Segre unit, in the Segre valley itself, the large folds trend N-S and fewer thrusts occur. The axial planes of these folds dip about 45° to the east. s_2, which can still be found north of Estimaríu, is absent in the Segre valley, so that a definite correlation is not possible with the deformation phases already established. The structural trends shown in these folds, however, are parallel to the nearest known F_1 and F_2 folds north of Estimaríu and to s_2 in the Orri dome west of the Segre unit (Appendix IV). Hence it is likely that these N-S folds can be correlated with these two early fold phases.

These N-S structures have been partly refolded about ESE-WNW axes with subhorizontal or gently north-dipping axial planes which locally develop into thrusts. A coarse fracturing can be seen, approximately parallel to these thrusts. All E-W folds show an asymmetry indicating a north to south movement. In the Segre valley itself, N-S folds are well-developed and the superposed E-W folds caused a complicated interference pattern which is excellently exposed along the road in the Segre valley (Fig. 45 and 46). At higher levels above the Segre, the N-S folds are much scarcer and finally disappear altogether, so that only E-W recumbent folds remain. The structures are shown diagrammatically in Fig. 47.

In the southeastern part of the Segre unit, east of Vilanova, E-W folds occur with axial planes dipping 45° to the north. It is not very likely that these folds are related to the E-W recumbent folds in the Segre valley. A gradual change in trend of the folds in the Segre unit from N-S in the Segre valley to E-W east of Vilanova seems more likely, as s_2 in the Ges-Estana unit in the south and structures of the Toloríu and Beixech units in the east all show an E-W trend. In the Segre unit itself no correlation can be proved, as the N-S folds in the Segre valley cannot be traced to the south in the monotonous Villech Formation. Bad exposured and contact metamorphism in the valley of

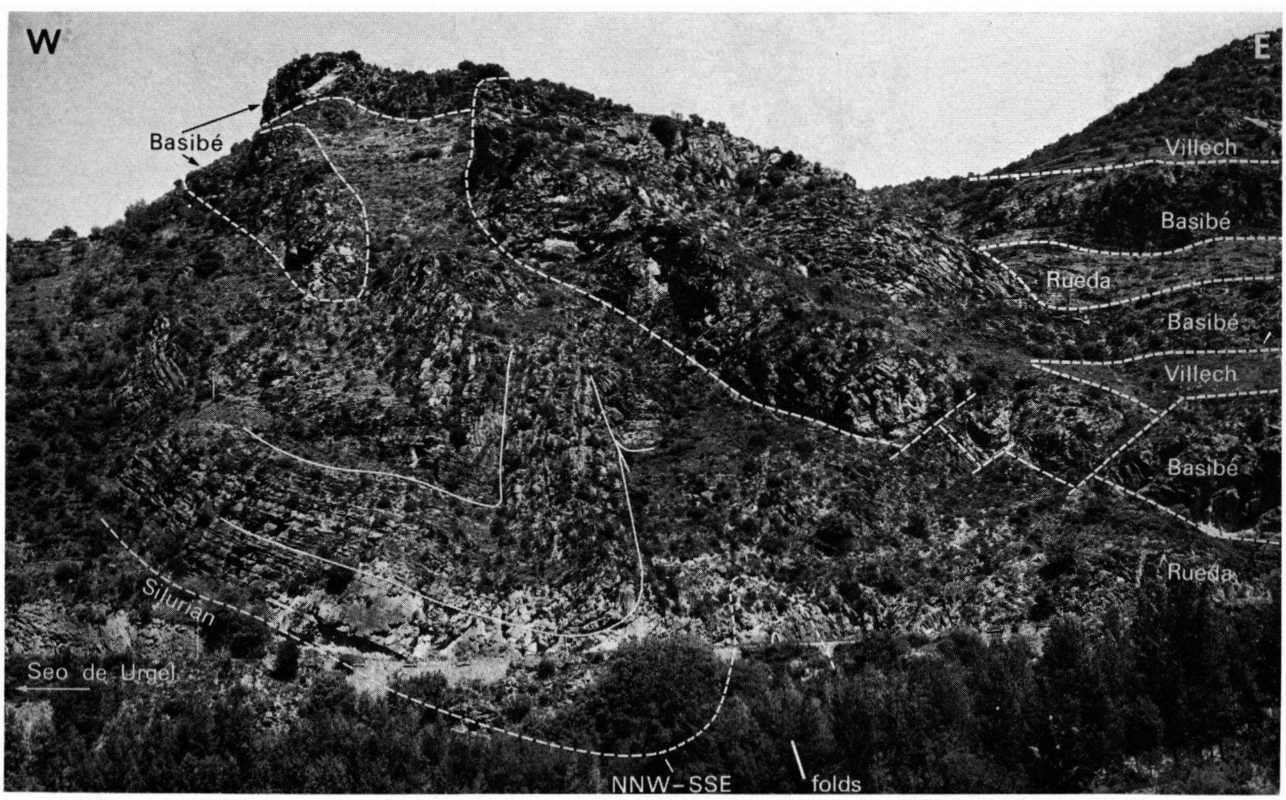


Fig. 45. N-S folds with superimposed E-W recumbent folds, along the Segre valley southeast of Torres.

the B^{co} de Vilanova alse made it difficult to trace these structures.

In the easternmost part of the Segre unit, thrust planes and most of the bedding planes dip to the west. This is clearly in contrast with the overall eastern dip of thrust and bedding planes in the western part of

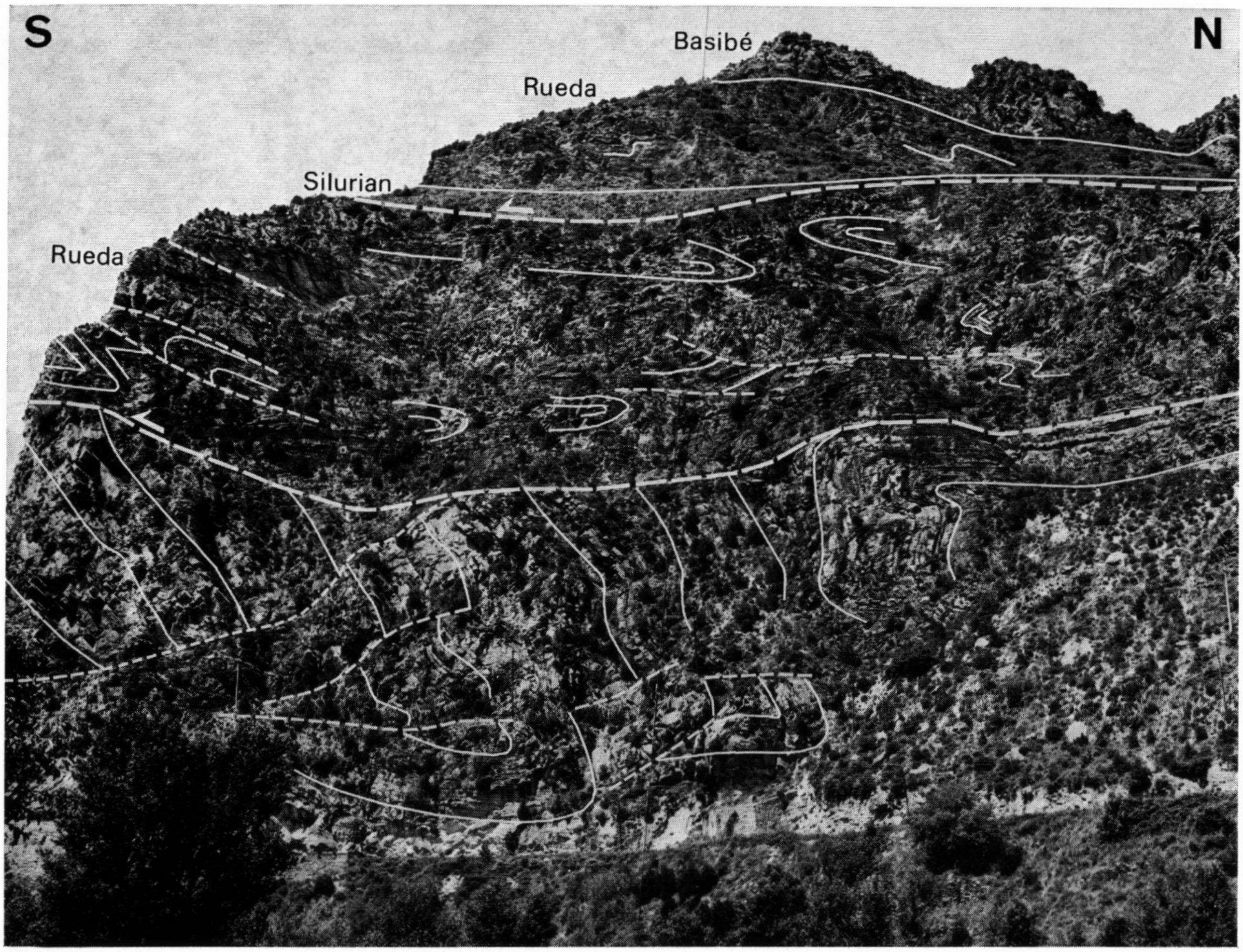


Fig. 46. E-W recumbent folds and subhorizontal thrusts in the Rueda Formation in the Segre valley west of Torres.

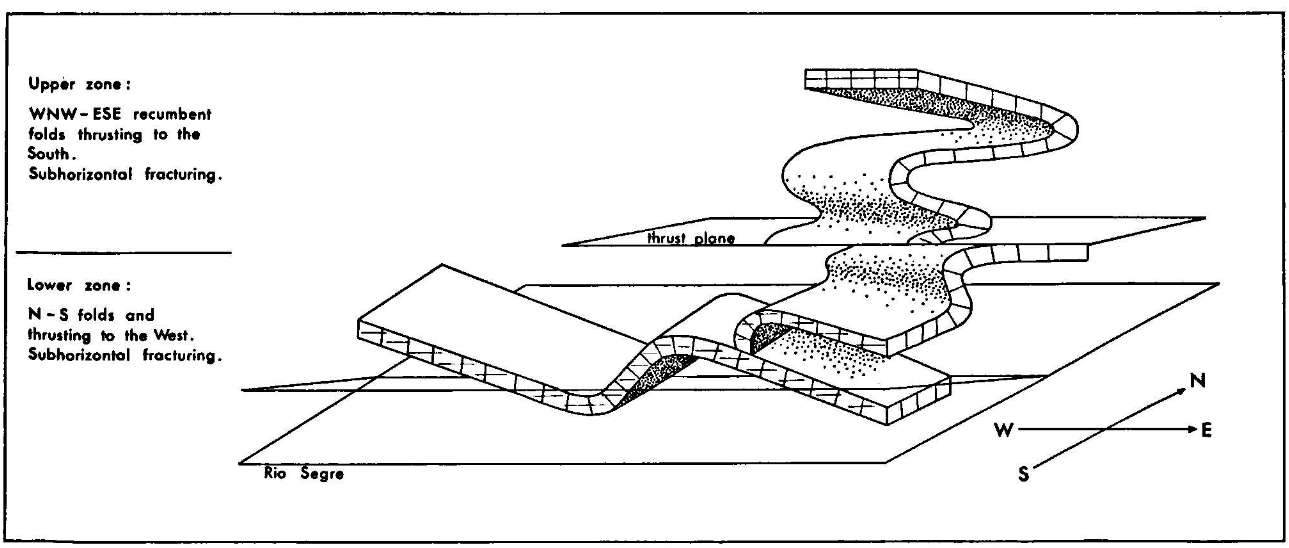


Fig. 47. Sketch of superposition of N-S folds and WNW-ESE recumbent folds in the Segre unit.

this unit. The line separating these two areas is called the Bescarán axis. Since the Devonian east of the axis forms the roof of the granodiorite, the dip to the west most likely has been caused by the intrusion.

Bescarán, Beneidó and Vilanova units

Overlying the Arcabell and Segre units, three units are exposed which have no direct relationship to the structures in the units below and which are composed of clearly allochthonous, subhorizontal thrust sheets of Devonian and Silurian rock units.

The Bescarán unit largely covers the eastern part of the Arcabell syncline. It consists of a large number of thrust sheets in which only occasional folds can be observed, all of which face south. Along the southern border of this unit these thrust sheets are approximately parallel to the thrust sheets in the Arcabell syncline, and there is no clear separation between the two units. Along the western border, however, the relationships are clearer. South of Arcabell in the Arcabell syncline, folds can be observed with axial planes dipping 40° to the north. Higher up and more to the east, in the Bescarán unit, thrust planes generally dip 25°–30° between east and northeast and cut off the structures of the Arcabell syncline (cross-sections 12 and 13).

In the Bescarán unit the tilting of the Devonian east of the Bescarán axis is very pronounced. East of Bescarán the thrust sheets dip to the northwest and continue in a similar manner in the area of the Pla de Lles and in the roof pendant of the Sierra de la Tuta, which at least partly belong to the Bescarán unit. In the northernmost and highest part of the unit there appears to be no influence of the warping, and axial planes of E-W folds and thrusts show a moderate dip to the north or are subhorizontal.

On top of the northern part of the Segre unit an allochthonous unit forms the summit of the Beneidó. This Beneidó unit consists of only one slightly south-dipping sheet comprising a large horizontal ESE-WNW striking syncline which, as all others, faces south. Since this unit contains the entire Devonian sequence, including the Compte Formation and even some loose remnants of the Bellver Formation, the detachment generally observed along the Villech Formation did not develop here.

The southern part of the Segre unit is largely covered by the Vilanova unit. This unit is composed of three or four thrust sheets dipping gently to the east, except in the southernmost part, against the Ortedó fault, where a northward dip is more prominent. Folds developed on the thrust planes face south, showing a direct relationship to the thrusts, and indicate movement toward the south. The thrust sheets of the Vilanova unit, mainly composed of Silurian, Rueda, Basibé and Villech Formations, show that here, as usual, the detachment zone in the Villech Formation was rather important having allowed the higher formations to move elsewhere.

In the Devonian directly east of Vilanova, which is considered as autochthonous and a part of the Segre unit, the Compte Formation is also present. This is the only example in the Segre unit where the Compte Formation was not detached from the rest of the Devonian.

Tolorïú unit

The Devonian and Silurian of the north flank of the Arseguell-Bar anticline, grouped in the Tolorïu unit, lie largely in the contact-metamorphic aureole, and structures remain rather uncertain. A small area southeast of Baños de S. Vicente, however, which lies outside the aureole, shows E-W folds with axial planes dipping 45° to the north which should probably be correlated with the folds of the same attitude in the Segre unit east of Vilanova.

Beixech unit

The Silurian, Devonian and pre-Hercynian Carboniferous rocks, south of the Arseguell-Bar anticline, have been grouped in the Beixech unit.

West of Coll de Ser, where a normal contact with the Arseguell-Bar anticline may be assumed, the folds have E-W axes with axial planes dipping 45° to the north even in the region between the Grau and Ansobell, where the Silurian and Devonian rest upon the north-dipping flank of the Ges-Estana unit. The only exceptions are in the narrow gorge of the Rio de Cadi, just north of the Hercynian unconformity, where folds with vertical axial planes have developed.

In the area west of Arseguell N-S folds are frequent, and become more so near the granodiorite contact. Since the axes trend parallel with this contact a shouldering-aside effect during the intrusion is evident for their origin, but contact metamorphism and bad exposures obscure the exact relationships with the E-W folds elsewhere.

East of Coll de Ser the fold axes gradually swing into a NE-SW direction and axial planes become vertical. The differences in orientation between these vertical folds and the structure of the eastern part of the Arseguell-Bar anticline with an s_2 cleavage, generally dipping 45° to the northwest, can possibly be explained by a rotation caused by the down-faulting of the Beixech unit along the Bar fault.

To the northeast the structures are cut off by the Aguila fault which down-faulted the Beixech unit and brought the unmetamorphosed Devonian and Carboniferous into contact with granodiorite and strongly metamorphosed Devonian limestones.

The steepening of the Hercynian unconformity during later deformations, appears to have had no influence on the Hercynian structures west of Coll de Ser. Further east, however, in the B^{co}. de Quer the syncline of Beixech became overturned to the south, and in the T^{te}. Capiscot the anticline is deformed into a domelike structure.

In the steep or vertical flanks of the large folds, smaller recumbent folds are common (Fig. 48). These folds, which occur in the flanks of the Hercynian NE-SW folds as well as in strata that have been rotated into a vertical position together with the Hercynian unconformity, should be interpreted as cascade folds.

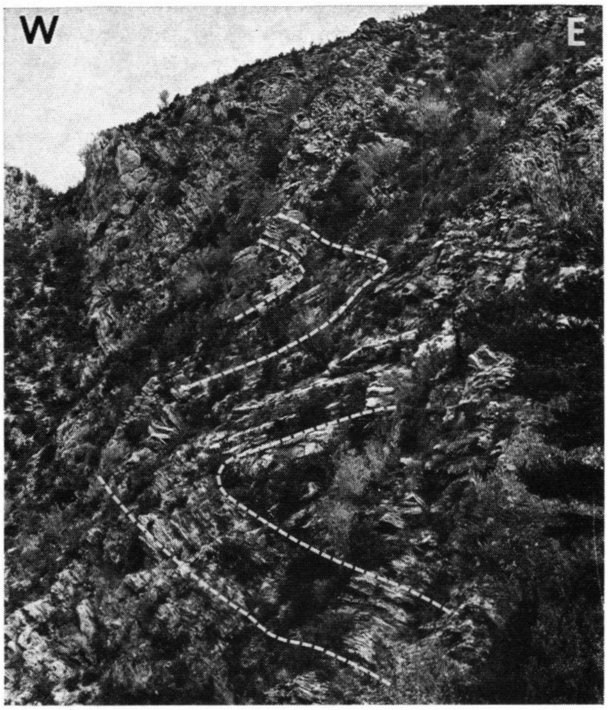


Fig. 48. Cascade folds in the Basibé Limestone, in the T^{te}. Capiscot.

Similar structures, but of a larger size, occur in the SE flank of the syncline of Beixech, which is folded into one large gentle cascade fold. Northeast of Beixech, in the middle of the Carboniferous slates of this syncline, a large block is found, consisting mainly of Compte limestone. Since this block is clearly situated on top of the Carboniferous slates it should be interpreted either as a slip-sheet that slid down into the partly eroded syncline, or as an olistolith emplaced relatively early. The overturned southeast flank could easily have provided a slip-sheet.

Minor unconformities, as described below the Bellver Formation (p. 182) in the Beixech unit, show that the earliest movements occurred during the deposition of the older Carboniferous. No clear relationship has been established between these movements and the structures described.

Monsech de Tost Unit

The Monsech de Tost unit is an isolated structure composed of Devonian and Silurian rocks and completely surrounded by post-Hercynian rocks. Although the post-Hercynian deformations involving the emplacement of the Monsech de Tost are of great importance, the internal structures of this unit are largely Hercynian. Only these Hercynian structures will be described here; younger structures will be dealt with on p. 222 and 226.

The Monsech de Tost unit may be divided into three sub-units: a central triangular part, comprising the Monsech de Tost itself, a northern zone east of Torá and north of Tost and an eastern zone (Fig. 62, p. 226). The internal structures of the central triangle provide the clearest evidence of their Hercynian age.

In the central triangle two units of different structure can be found: a) The northwestern slope, consisting of numerous thrust sheets dipping about 30° to the south; the few folds developed in these sheets generally face south; in some sheets N-S folds occur facing west; to the southwest these thrusts are unconformably covered by the Bunter, indicating their Hercynian age. b) The summit and southern slope of the Monsech de Tost consist of one large unit comprising the entire Devonian sequence. On the summit these strata are subhorizontal and rest upon the thrust sheets visible in the northwest slope. In the southern slope E-W folds have developed facing south. In the deepest part of this structure the Rueda Formation crops out showing a somewhat irregular fracture-cleavage dipping 30°–40° NNE.

Structural setting of intrusive bodies

Granodiorites. – Four granodiorite bodies occur within the map area the very large Andorra granodiorite in the northeast and three smaller stocks in the Llavorsi syncline, viz. the Santa Coloma, Fontaneda and Montescladó granodiorites (Fig. 32).

The Andorra granodiorite occupies one sixth of the area of the map extending beyond the northeast margin. It forms the western part of the Andorra-Mont Louis granodiorite, the largest intrusive body of the Pyrenees (576 km^2).

The attitude of the contact with the surrounding sedimentary rocks is variable although a relatively large proportion is subhorizontal constituting the roof of the granodiorite body (cross-sections 1–6, 10) It can often be shown that the roof has been tilted as, for example, the eastern parts of the Segre, Bescarán and Beneidó units (p. 208, 209). Sedimentary rocks along steep contacts may locally show folds, with axial planes parallel to the contact, which are possibly related to a shouldering-aside during the intrusion, as they do not occur outside the contact zone (e.g. east and north of Arseguell). Steep fault contacts occur only locally such as north of Arseguell, south of Musa and in the valley of the Valira d'Orient where the contact is strongly sheared (Zwart, 1965).

Faults and macro-joints are very frequent in the granodiorite itself. From aerial photographs it can be observed that the NW-SE system is by far the most important. This was also concluded by Roggeveen (1929), who measured many joints in the southern part of the granodiorite (Lles-Aristot massif) (Fig. 49), which was not sufficiently exposed for an analysis by means of aerial photographs. The Castellnou fault, affecting the roof, and the faults that cut off the Rabassa dome, all trend NW-SE and are considered to belong to the same system as the faults and macro-joints in the granodiorite itself.

The three small granodiorite bodies, Santa Coloma, Fontaneda and Montescladó granodiorite intrusives in

Fig. 49. Diagram showing the orientation of joints in the southern part of the Andorra granodiorite, compiled from data of Roggeveen (1929).

the Llavorsi syncline, all show a somewhat different relationship to the surrounding rocks. The Santa Coloma stock (11 km^2 in area) has steep contacts which are only partly parallel to the structures in the enclosing rocks (cross-sections 12–15). The structures along the NW contact seem to have been pushed aside into a NE-SW trend. The Fontaneda stock (0.5 km^2) does not show any structural feature that could indicate a shouldering-aside effect. The Montescladó intrusive (0.5 km^2) is a lens-shaped mass subparallel to the existing isoclinal folds.

The contact-metamorphic aureole which developed around the granodiorites varies greatly in width. The width of the aureole in the Devonian rocks around the Segre valley varies mainly between 500 and 700 m (cross-sections 1, 2, 3, 5, 6), in the Valira valley between Sant Julia and Andorra la Vella, however, this zone appears to be about 1500 m wide (cross-sections 10, 12–15). This great width is possibly partly due to hidden intrusive bodies. The very thin aureole around the Montescladó granodiorite, not exceeding 200 m, is certainly due to the small size of the body.

East of Conflens, traces of contact-metamorphism have been observed in the Ordovician, Silurian and Devonian rocks. In the field this contact-metamorphism is not very obvious, but thin sections revealed a growth of epidote and chiastolite in random orientation with respect to s_2. Although no exposures have been found, a granodiorite body is probably present in this area because of the large amount of dyke material found in the debris southeast of Conflens. Such an intrusive could conceivably explain the relatively high position of the Devonian north of the Sierra Plana fault.

Dykes. – The distribution of dykes in the area seems to be related to that of the granodiorites, they are frequent in the granodiorite bodies themselves as well as in the adjoining structural units.

In the Andorra granodiorite most dykes are oriented NW-SE, parallel to the main joint direction (Roggeveen, 1929). Beyond the southern boundary of this granodiorite, they only occur in a relatively narrow zone 2 km in width, whereas in the Segre valley to the west of this granodiorite, dykes have been observed to a distance of 6 km from the contact, even in the eastern part of the Orri dome. In the Valira valley lamprophyres can be observed which were intruded along the Llavorsi fault. In the Segre unit and Arcabell syncline and in the related gravity units, dykes are very frequent and generally intruded along the s_2 cleavage and the thrust and gliding planes. In the Rabassa dome dykes have only been observed in the eastern part, generally intruding along the steep NW-SE striking faults. In the Llavorsi syncline they generally intruded along s_2 planes and occasionally also along the diapiric zones. An important E-W dyke-swarm occurs between the Montescladó granodiorite and the Tressot valley and is possibly related to this intrusive body. Southeast of Conflens an important concentration of dykes occurs which is probably related to contact-metamorphism, occasionally observed in this area.

In the Ges-Estana unit, between Querforadat and Estana, an isolated concentration of lamprophyres can be observed, one of which has been observed to be cut off by the Hercynian unconformity.

Dykes which intruded along s_2 planes in the Llavorsi and Arcabell syncline often themselves show a strong cleavage which is parallel to s_2; a compressive stress must therefore have operated after their intrusion.

Succession of the Hercynian deformation phases

The structures described in the units will now be discussed and correlated. The final scheme to which this discussion will lead is given in Fig. 50.

Genesis of the slaty cleavage (s_0). – As we have seen the analysis of the large structural units yields a relatively simple pattern of large domes, anticlines and synclines for pre-F_2 structures e.g. the Massana anticline (p. 199) and the Llavorsi syncline (p. 203). The slaty cleavage s_0, being parallel to ss, follows these large structures. In view of this situation there are two possibilities for the formation of s_0: s_0 either predates F_1 or is related to it.

If s_0 predates F_1, there is no structural feature to which it can be related and s_0 can only be attributed to a synsedimentary orientation of mica flakes, a mica orientation by compaction, load-metamorphism or flattening perpendicular to the bedding during a shearing movement but without the formation of recumbent folds. The process of a slaty cleavage formation by a shear movement was proposed by Zwart (1963) for the flat-lying cleavage or schistosity in the infrastructure of the large Cambro-Ordovician domes and anticlines. If this concept should be used

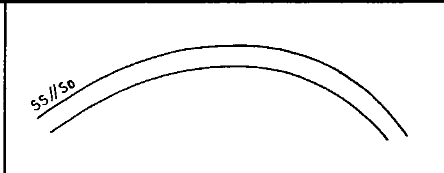
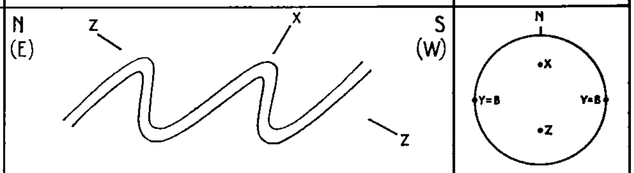
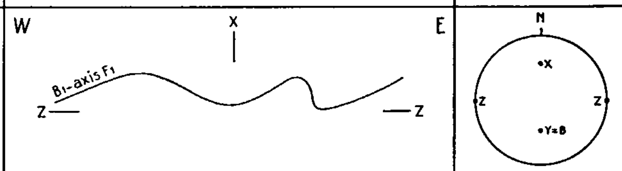
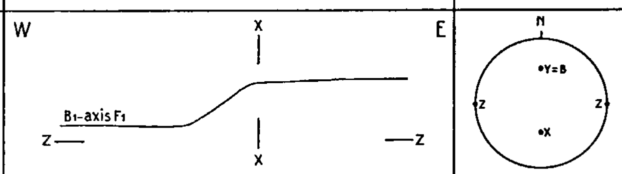
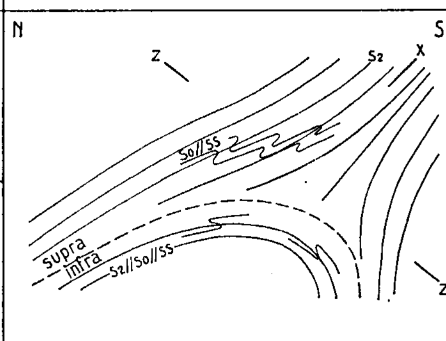
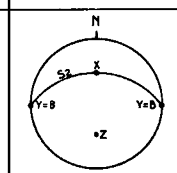
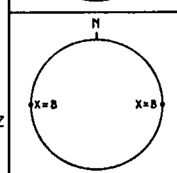
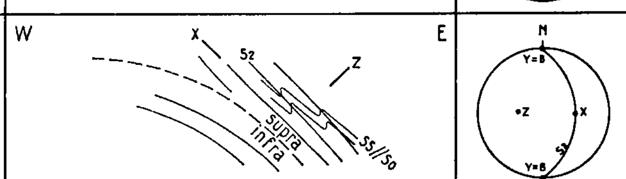
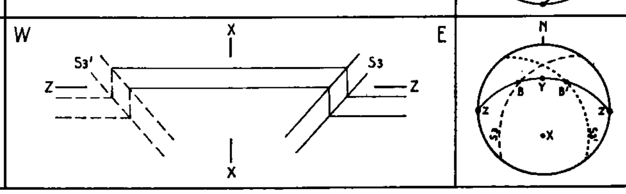
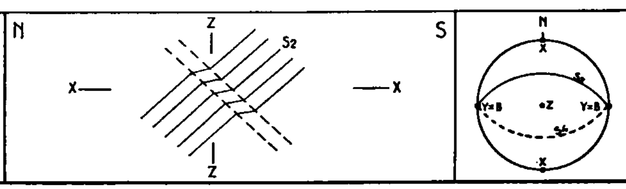
DEFORMATION PHASE		FOLD AXIS	S-PLANE	SECTION ⊥ B-AXIS	STEREOGRAPHIC PROJECTION	DEFORMATION PHASES ZWART 1963
FIRST PHASE	F_1	Domes, largest folds	? If $S_0=S_1$ ↓ S_0 concentric slaty cleavage		No specific orientation of axes	Pre-cleavage or pre-schistosity folding
		Parasitic E-W folds N-S folds				
	F_1'	Folding of B_1 axis in a-b plane				
	F_1''	Folding of B_1 axis in b-c plane				
SECOND PHASE	F_2 Supra	E-W folds	S_2 axial-plane crenulation cleavage			1st phase or mainphase
	F_2 Infra	E-W recumbent folds	Reactivated S_0 slaty cleavage			
	F_2 Supra (N-S)	N-S folds	S_2 axial-plane crenulation cleavage			2nd phase (only in infra-structure)
THIRD PHASE	F_3	Conjugate set, steep folds	S_3 kink-bands NE-SW and NW-SE			3rd phase
(not observed)						4th phase
FOURTH PHASE	F_4	Conjugate set of S_4 and S_2 E-W folds	S_4 kink-bands E-W			

Fig. 50. Tectonic scheme of deformation phases.

it would imply that before the F_1 folding an infrastructural regime existed which extended into the formations now exposed in this area.

If s_0 is related to F_1 this slaty cleavage, being subparallel to the bedding, could have formed as a concentric cleavage (de Sitter, 1964).

Savage (1961, 1967) described in great detail such a concentric relationship of slaty cleavage and a large syncline in the Cantabrian mountains. He concluded that the development of this slaty cleavage was caused by flattening perpendicular to the bedding and related to the formation of the large syncline, relying on the fact that slaty cleavage is generally considered as a plane of flattening (Furtak, 1962; Hellermann, 1965); see also Ramsay (1964, p. 180). Concentric orientation of crenulation cleavages in relation to smaller folds, described and discussed by Hara (1966, 1967) and Dieterich (1969), and the stress directions perpendicular to the convex portion of fold hinges deduced from them, support this concept of concentric cleavage. In our area, in which the slaty cleavage is also subparallel to ss, the genesis of this cleavage could be related to the doming and the formation of the largest anticlines and synclines, involving a flattening of the bedding plane. If the relationship of the slaty cleavage and F_1 should be accepted, this cleavage could be indicated as s_1.

First phase structures (F_1). – First phase structures are defined in this area as all structures that predate F_2, and can best be recognized as such from the relationship to s_2. Two different types of F_1 structures exist in this area: 1. Large domes, anticlines and synclines comprising complete structural units. 2. Large parasitic folds superposed upon these major structures. Since F_1 and F_2 are usually subparallel, the main problem is to differentiate between these two phases.

The general indication that these two phases do exist is the widespread occurrence of δ_2 lineations showing important deviations from the F_1 fold axis, which clearly demonstrates that two non-parallel folding phases are involved (Fig. 34, p. 200). This phenomenon, however, does not reveal the shape or trend of the F_1 folds.

From the relations of folds and cleavage in the crest and the south flank of the Massana anticline, it became clear that this large structure antedates F_2 and must have been formed during F_1 (p. 199). It seems logical to consider the bordering Llavorsi syncline and all other large domes, anticlines and synclines as first phase structures as well. Zwart (1963) also suggested an early origin for the largest structures in his description of the large Ransol syncline and the related large Hospitalet gneiss anticline, north of this area in Andorra. This oldest folding phase was termed by Zwart the "Pre-cleavage or pre-schistosity folding".

The first phase origin of the large parasitic folds, as for instance developed in the Llavorsi syncline, is deduced from the following evidence:

a. Folded F_1 folds in zones where the deviation between F_1 and F_2 is sufficiently large, e.g. in the region around Ars and Asnurri.

b. The relation of the large folds as parasitic to the structure of the Massana anticline.

It has already been stated in the description of the Massana anticline that the first phase folding created a steep north-dipping south flank in the Ordovician rocks of this anticline, which formed such a small angle with the later s_2, that no second phase folds could develop here. Likewise the attitude of Devonian and Carboniferous rocks of the north flank of the Llavorsi syncline should not have allowed F_2 folds to develop. Hence the large folds observed in the Llavorsi syncline must have originated during the F_1 phase, whereas the F_2 phase was only responsible for their further compression.

In the Segre unit s_2 cleavage is poorly developed so that only a few of the large folds can be definitely assigned to F_1 or F_2 phases. Since these folds form the continuation of those in the Arcabell syncline, both F_1 and F_2 may be represented in the N-S folds of the Segre unit. If the large parasitic folds that swing around from an E-W orientation in the Arcabell syncline into a N-S orientation in the Segre unit partly belong to F_1, the stress field along the border of the Orri dome must have been a radiating stress between N-S and E-W. Such a stress distribution could be caused by the doming of the Cambro-Ordovician in that area.

Deviations of F_1 fold axis (B_1) of the large folds, as developed in the Llavorsi and Arcabell synclines (Figs. 37 and 38, p. 204), are of two different types: deviations of B_1 within the axial plane causing "canoe folds" (Lahee, 1961) and deviations of B_1 within the bc-plane[1] of the F_1 folds. Both deviations seem to indicate a compression in b, only the direction of extension differs in the two types. The genesis and relationship of these first phase folds will be discussed on p. 217.

Second phase structures (F_2). – Second phase structures are characterized in most of the area by a strong s_2 cleavage, but in the southern part s_2 is in general less well developed and even absent in a few places. The manner of the folding also varies from tight in the north to more open in the south.

In the Llavorsi syncline and the southern zone of the Massana anticline the s_2 cleavage is parallel to the axial plane slaty cleavage, which developed in the Salat-Pallaresa anticlinorium north of the map area. This slaty cleavage was described by Zwart (1963, 1965) and Oele (1966) as the "s_1 cleavage", which developed during the "first folding phase or main phase". Because of this parallelism the first phase of Zwart is correlated with the second phase (F_2) of the present author. This correlation could be explained as follows: in a south to north direction the s_2 crenulation cleavage is gradually transformed into a slaty cleavage by a more severe compression possibly accompanied

1) a, b and c axis of reference are used here to indicate fold geometry as defined by Dennis (1967).

by more intensive recrystallization. In the Salat-Pallaresa anticlinorium, the slaty cleavage (s_0) parallel to ss as in our area, became parallel to the new cleavage s_2 (s_1 of Zwart) in the isoclinal folds and is reorientated in fold hinges.

Another possibility is that s_0, as developed in our area, never existed in the Salat-Pallaresa anticlinorium and that s_2 (s_1 of Zwart) did develop in that area as an original steep axial plane slaty cleavage.

It can be seen from Appendix IV that the trend of s_2 behaves in two different ways with respect to the large Cambro-Ordovician domes and anticlines: Along the southern border of the Massana anticline, the northern and southern border of the Rabassa dome and the northern as well as the eastern border of the Orri dome, the strike of s_2 is sub-parallel to the border of the dome. On the other hand the western border of the Rabassa dome clearly shows a cross-cutting strike of s_2 as is probably also the case in the Ges-Estana unit.

between dome-boundaries and the strike of s_2, especially of the large curvature of s_2 in the eastern part of the Orri dome, the Arcabell syncline and Segre unit, requires some explanation. The main question is whether this dome structure should be described as an original structure formed by first and second deformation, or as a structure formed by a later deformation of an s_2 with an originally E-W strike.

We shall first discuss the possibility of a secondary deformation of s_2. Fig. 51 represents a schematic stereographic projection of the positions of the s_2 cleavage and the fold axes of the large folds related to s_2 in three different areas (P, Q and R) of the Orri dome and the bordering zones of the Arcabell syncline and Segre unit. Fold axes of minor folds and lineations are not used here because they may, as already mentioned, generally possess a random orientation in the s_2 plane. In the diagram (s_2) P would represent a postulated "original" E-W striking s_2 and related E-W

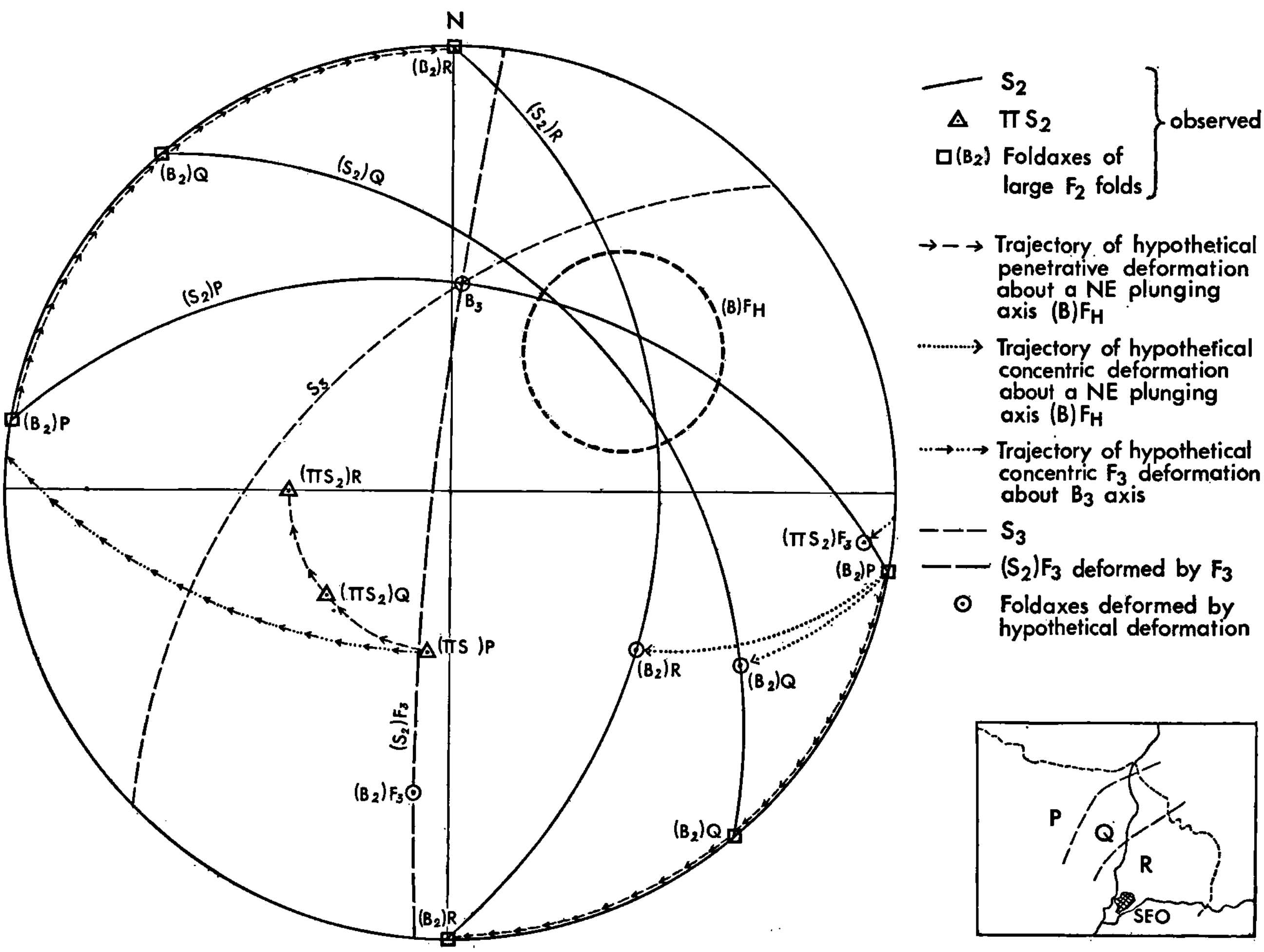


Fig. 51. Composite stereographic diagram demonstrating the impossibility of a secondary deformation of s_2 in the eastern part of the Orri dome.

The E-W strike of the s_2 cleavages and the way they cut across eastern and western terminations of large Cambro-Ordovician structures can simply be explained as a result of N-S compression. However, the parallelism

fold axes, whereas (s_2) Q and (s_2) R would represent the result of the secondary regional deformation of s_2 and related fold axes. Here the following considerations are of importance: Folding of structures from orientation P into structures of orientation Q and R would require a deformation without vertical tectonic transport of any importance. A folding process in which a well-developed s_2 plane yields in simple shear in the horizontal plane, would require a penetrative deformation involving extensive movements in a NE-SW direction in the Valira and Segre valleys. Such a deformation would certainly have been accompanied by visible evidence such as small-scale folding or fracturing of s_2, nothing of which has ever been observed.

A concentric folding of s_2 about a NE plunging axis, (B) F_H in diagram, could result in east-dipping s_2 planes, but this yields a deformation path for the F_2 fold axis curving out of the horizontal by up to 40°.

A third possibility for a deformation of s_2 might be provided by the F_3 folding phase. Since only the east flank of the Orri dome could be considered as the actually deformed part of the structure, the corresponding stress field should be approximately E-W. F_3 folds also possess only one deformed flank and have a corresponding E-W stress field (see discussion of F_3 folds, but the deformed flank generally has a vertical N-S attitude and, besides, in these folds the F_2 fold axis would have curved out of the horizontal. Furthermore, F_3 folds do not occur in the area considered (Appendix IV), and in areas where they do occur they are of small dimensions and the limited occurrence could not represent substantial compression. If the eastern termination of the Orri dome is considered as a large fold, the E-W compression must have been considerable in which case the absence of F_3 folds is inexlpicable.

Hence we may conclude that the dome structure northeast of Seo de Urgel is not due to a secondary deformation of s_2, but should be regarded as an original dome with a symmetry concentric between north and east, formed during the second and possibly already during the first deformation phase. The centre of symmetry of this dome should then be situated somewhere inside the dome, southwest of Seo de Urgel. We shall call this affinity of s_2 to the dome structure the concentric tendency, which genesis will be discussed on p. 218.

Large folds parasitic on the major structures are generally related to F_2 as they nearly always have s_2 parallel to their axial planes. If no s_2 is developed in the folds themselves, the axial planes are parallel to s_2 in a bordering structural unit; N-S folds in the Segre unit, for instance, have their axial planes parallel to s_2 in the Orri dome, and E-W folds in the western part of the Beixech unit have their axial planes parallel to s_2 in the Ges-Estana unit. In the discussion of the first phase structures it has been argued that at least in the Llavorsi and Arcabell synclines the large parasitic folds are primarily F_1 folds and were merely compressed more during F_2. In the Segre and Beixech units, however, this is more doubtful.

As has been shown on p. 199 and p. 203 the steeply plunging δ_2 lineations (Fig. 34, p. 200) resulted from a small angle between the trends of the two folding phases F_1 and F_2. The development of F_2 folds depended on the angle between the two phases being large enough, where the angle is too small only δ_2 lineations caused by F_2 marks the existence of two folding phases.

The attitude of the s_2 cleavage, consistently dipping north in most of the southern part of the axial zone, has been discussed by a number of authors (de Sitter, 1956; Zandvliet, 1960; Zwart, 1963; Boschma, 1963), who attributed it to a relatively late fanning-out process initiated by an uplift of the axial zone. In this area, however, it can be demonstrated that a fanning-out process is unlikely. The process, suggested by Zandvliet, involved normal faulting along s_2, but in the Rabassa dome it has been demonstrated that most of the movement along s_2 consisted of upthrusting from N to S. In the eastern part of the Orri dome, where s_2 dips east, such a fanning out of the cleavage toward the south would not even explain this fact so that such a mechanism cannot have operated in any case.

Third phase structures (F_3). – Third phase folds or kink-bands only occur in areas with well-developed E-W striking s_2 planes (Appendix IV). The extreme asymmetry of the folds, the very long north flank being the undeformed s_2 and the short east flank the actually deformed part, indicates an approximately E-W stress field, forming a slight angle with s_2 (Weiss, 1968).

The geometry and orientation of these folds (Fig. 29 and 50), strongly suggest a relationship with the "third phase folds" described by Zwart (1963). Zwart found a conjugate set of these folds, with folds trending NW-SE and NE-SW. In the Aston and Hospitalet massifs, north of this area, the NW-SE folds were more frequently observed (Zwart, 1963, 1965; Oele, 1966). In our area, however, the NW-SE folds are almost unknown. The small dimension and the scarcity of the folds indicate that the E-W compression related to this folding was very limited. The genesis and relationship of the F_3 folds with other folding phases will be discussed on p. 217.

Fourth phase structures (F_4). – Fourth phase kink bands result from the deformation of well-developed E-W striking s_2 planes and therefore only occur in the northern part of the map area.

Two varieties of F_4 kink-bands have been observed: south dipping kink-bands and conjugate sets of kink-bands. The south dipping kink-bands are most extensively developed and are also described by Kleinsmiede (1960), Zandvliet (1960) and Zwart (1963). Zandvliet correctly ascribed these kinks to a very limited dilatation in a horizontal sense, i.e. to a vertical compression. Zwart suggested a conjugate set of the kink-band and cleavage as shown in Fig. 52A. A conjugate set implies that beside the relative movement along the kink-band a slip occurred along s_2. This opinion is confirmed by the experiments of Weiss (1968), who showed that not only did the s-planes between the kink planes rotate, but so did the s-planes outside. This rotation implies a slip along these planes, as is required in the conjugate set. The compression required for this fold system may vary between vertical and inclined to the north, but is slightly steeper than s_2.

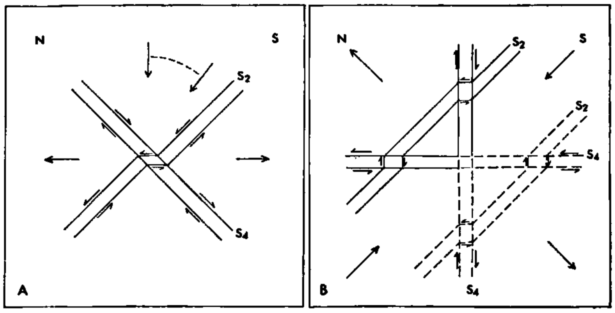


Fig. 52. Geometry of F_4 kink-bands.

Conjugate sets of kink-bands as shown in Fig. 52B and Fig. 40 (p. 204) occur only locally .They have developed in a stress field with a N inclined main stress parallel to s_2. Outside these s_4 kink planes no slip is required. The genesis and relationship of the F_4 folds with other folding phases will be discussed on p. 217.

pushed to the south as one structural unit. The incompetent calc-slates of the Villech Formation acted here as a detachment plane in which all the thrusts were concentrated.

The north to south sliding movement can also be demonstrated by means of structures from below the allochthonous units. In the underlying Segre unit subhorizontal fractures, thrusts and ESE-WNW recumbent folds are very important and here, too, a north to south thrust direction was deduced. Hence it is very probable that the thrusting in the Segre unit, the thrusting and sliding in the Bescarán unit and the sliding of the Beneidó unit are all related to a gravitational sliding movement of the Devonian from the south flank of the Rabassa dome to the south. In the Arcabell-Camprodón syncline east of the Andorra granodiorite large recumbent folds and thrusts have also been observed (Brouwer, 1968a; Bloemraad, 1969), and obviously gravitational deformation is the leading principle in this large syncline.

In the foregoing reasoning about the thrust and sliding directions the orientation of folds was used as an indication. It should, however, be stated, that it is not impossible that after the formation of these folds the

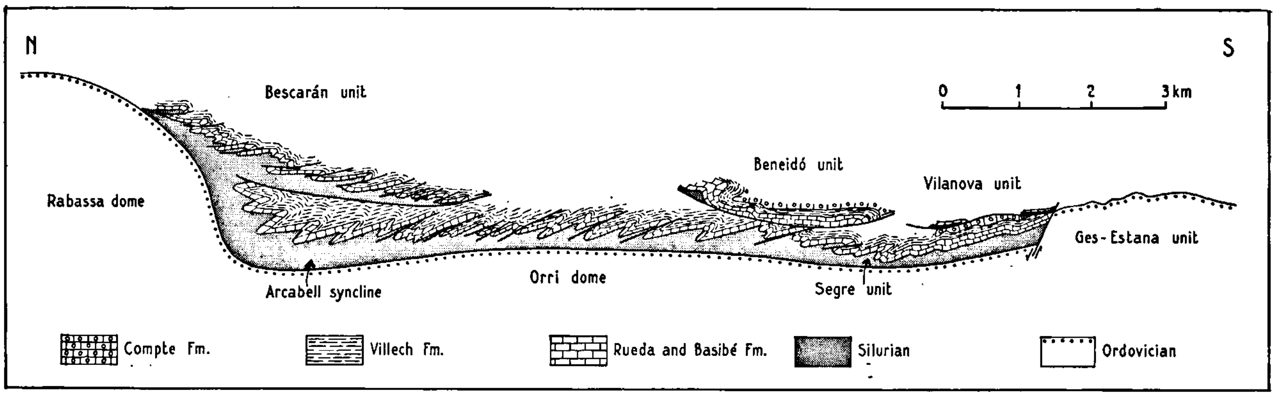


Fig. 53. Schematic N-S cross-section through the Arcabell syncline, Segre unit and related gravity structures.

Large scale gravity structures. – Since all the folds in the Bescarán, Beneidó and Vilanova units face south and all thrusts show a displacement to the south, it is assumed that the entire mass of these units slid down from a higher topographic position, the Rabassa dome, into the depression of the Arcabell syncline and Segre unit (Fig. 53). Since Compte and Bellver Formations have hardly ever been observed outcropping below these gravity units (except, locally, the Compte Formation below the Vilanova unit), and even small fragments of these rock units are absent from the thrust planes, it seems highly improbable that younger formations were overridden by these gravity structures. The numerous thrusts which facilitated this southward movement must have been largely restricted to the Rueda, Basibé and Villech Formations, whereas the competent limestones of the Compte Formation with the slates of the Bellver Formation on top were

thrust sheets slid backwards or in some other direction without folding during subsequent movements. The possibility should not be excluded that some thrust sheets of the Vilanova unit finally slid down from the Ges-Estana unit to the north. It is also conceivable that some sheets did slide down from the Orri dome to the east or northeast. In many cases this would shorten the necessary distance of transport. Since the observations on folds always indicate a transportation in a roughly southerly direction, the solution given above is, however, preferred.

The Hercynian structures of the Monsech de Tost are largely comparable to the structures developed in the Bescarán and Beneidó units. Since these two units have been interpreted as allochthonous structures that slid down from the south flank of a dome into a synclinal area, a similar solution is suggested for the Monsech de Tost unit. That is to say that the Devonian of this unit must have slid down from the south flank of the Orri dome into a synclinal zone where the Monsech de Tost is now located (Fig. 63A, p. 227). Most probably this synclinal zone extends along the entire southern border of the Orri dome and thus also comprises the Devonian and Carboniferous of the Freixa-Castells area west of the Rio Segre (Fig. 33). The present position of these structures amid post-Hercynian rocks will be dealt with on p. 222 and 226.

Folds with subhorizontal axial planes which chiefly developed in the autochthonous Segre and Beixech units can be described as cascade folds, developed in unstable flanks of older folds (Figs. 45, 46 and 48). The orientation of these folds only depends on the original orientation of these deformed flanks.

Granodiorite intrusions. – The contact-metamorphism related to the granodiorite intrusions has been definitely established as post-F_2 because of its having affected the s_2 cleavage.

The contact-metamorphic aureole developed in the Bescarán and Beneidó units is very regular and from the map picture no thrust or sliding movements can be identified that postdate the contact-metamorphism. Other late deformation phases have not been observed in the contact-metamorphic zone and the time relationship to these phases cannot be established. Zwart (1963) found that the contact-metamorphism of the Andorra granodiorite postdates his fourth phase (E-W refolding) (Fig. 50); this phase is not developed in this area, which implies that the contact-metamorphism postdates F_3.

The time relationship between the contact-metamorphism caused by the intrusion and the folding phases, however, is not necessarily the same as the relationship between the deformative stress caused by the intrusive and the same deformation phases. It is believed that during the process of intrusion structures can be formed which can eventually be metamorphosed by the same intrusive body. There is evidence in this area that F_2 could have developed simultaneously with the intrusion of the granodiorite. This can possibly be read from the aberrant strike of s_2 northeast of the Santa Coloma stock. Dykes, which are thought to be related to the granodiorites, have often intruded parallel to s_2 and are themselves cleaved. This also suggests that the intrusion process took place during F_2. The large scale gravity structures, as for instance the Bescarán and Beneidó units, could also be contemporaneous with the intrusion in which case these units need not be derived from the Rabassa dome but could also have slid down from an uplifted area where the Andorra granodiorite is now located.

Dyke intrusions. – The relationship of the intrusion of dykes and the deformation phases seems more complex than the larger granodiorite masses. Cleaved dykes which are parallel to the s_2 plane, as can be observed in the Llavorsi and Arcabell synclines, possibly intruded during F_2. A cleavage development in these dykes during a post-F_2 compression perpendicular to s_2 is not very likely since such a deformation phase is not known in this area. E-W kink bands occasionally observed in these dykes indicate a pre-F_4 age.

More or less undisturbed dykes parallel to s_2, dykes occurring in the thrust of sliding planes of the Arcabell syncline, the Segre unit and their related gravitational units, and dykes occurring in the steep block faults only indicate that they postdate these features. The lamprophyre west of Estana which is cut off by the post-Hercynian unconformity demonstrates its Hercynian age.

Summarizing we may say that the dykes probably started to intrude during F_2, possibly in relation with an intrusion of the granodiorites. The last dyke intrusions took place after the block faulting and predate the Hercynian unconformity.

Block faulting. – Block faulting mainly occurs in the Cambro-Ordovician and in the granodiorites. The most important fault direction is NW-SE. A downthrown NE block is observed most frequently. The faults postdate F_2 since they cut off these folds. Intrusions of dykes along these faults indicate that they predate the Hercynian unconformity.

Relationship and genesis of the Hercynian deformation phases
To explain the occurrence of successive folding phases the following considerations are of importance. It is assumed that, because of continuous overall shortening of the orogene, tectonic stress builds up more or less continuously at a low rate as has been recorded in tectonically active regions, but is released at certain intervals as folding takes place thus creating an oscillating stress (Fig. 54). Furthermore, it relies on the

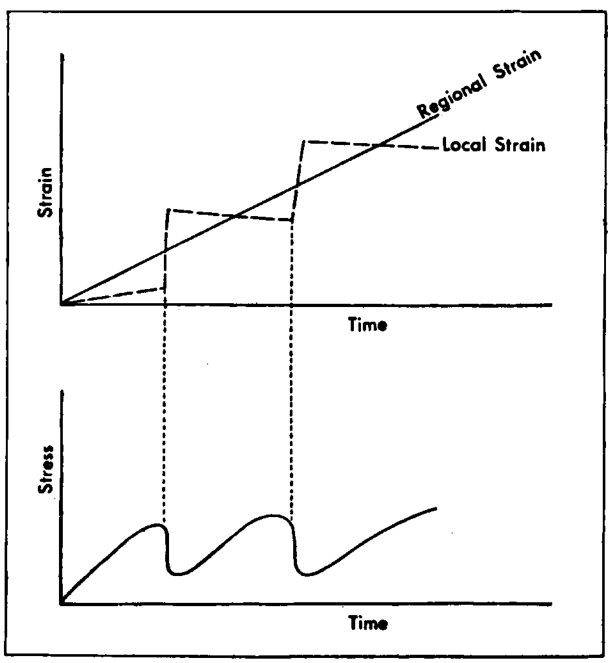


Fig. 54. Sketch diagram of stress-time and strain-time relationships.

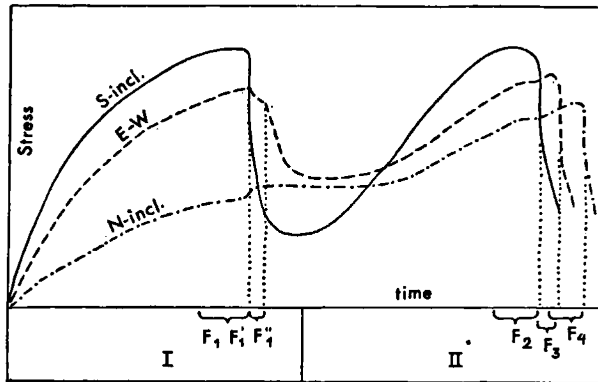
principle that plastic strain achieved by folding would probably take place much more rapidly than elastico-plastic strain recovery. The retarded recovery of elastico-plastic strain is well known from rock-mechanical experiments.

An increasing stress in one specific direction will generally result in folding involving shortening in c, a large extension in a and a small extension in b. The stress built up in b may be released by folding of the B axis within the ab plane forming "synchronous cross-folds" (Bhattacharji, 1958). As a result of the folding the stress in c is relatively rapidly relieved and drops below the value of stresses built up in the other axial directions b and a, as the sharp decrease in c would only slowly be transferred in the other axial directions. The stress remaining in b, being the largest, could be released by folding of the B axis within the bc plane, and also the stress in a could be released by folding involving an extension in c.

The general idea of successive differently orientated folding phases developing from one major compressive stress has also been proposed by Knill (1960).

The successive folding phases described and listed in Fig. 50 are of different trend, dimension and frequency. As the stress responsible for the major compression (F_1 and F_2 folding phase) was inclined to the south relative to the units as we see them and other directions of important regional compression are unknown, the general proccess described above provides a solution for the genesis of these phases. For sake of simplicity only the northern part of the map area, where the major structures (F_1 and F_2) show an E-W trend, will be used in the following discussion.

Two major folding phases can be distinghuished (Fig. 55): Major folding phase I (see also Fig. 38) consists of F_1 E-W trending folds and two reaction phases resulting from the stress in b of which the F'_1 folds are synchronous cross-folds, as they may have formed simultaneously with F_1, and the F''_1 folds which necessarily postdate the F_1 folds. The synchronous cross-folds (F'_1), however, could also be explained by differential flattening. Major folding phase II consists of F_2 E-W trending cleavage folds and two reaction phases F_3 and F_4. F_3 folds may result from the stress built up in E-W direction during the F_2 folding and are the equivalent of the F''_1 folds of the major phase I. The very limited E-W compression, deduced from the scarceness and small size of the F_3 folds, just may compensate the small E-W extension caused by the F_2 folding. F_4 folds could develop as soon as the S inclined stress dropped below the N inclined stress (F_4 B) or the N-S stress dropped below the vertical stress (F_4 A) and therefore can be partly synchronous with F_3.



Folding phase	Stress directions	Direction of shortening	Direction of main extension
I F_1	S incl. > E-W > N incl.	S incl.	N incl.
F'_1	S incl. > E-W > N incl.	E-W	N incl.
F''_1	E-W > N incl. > S incl.	E-W	S incl.
II F_2	S incl. > E-W > N incl.	S incl.	N incl.
F_3	E-W > N incl. > S incl.	E-W	S incl.
(B)	E-W > N incl. > S incl.	N incl.	S incl.
or (A)	E-W > Vertic. > N-S	Vertic.	N-S
$^+F_4$ (B)	N incl. > E-W > S incl.	N incl.	S incl.
or (A)	Vertic. > E-W > N-S	Vertic.	N-S

$^+F_4$(A): conjugate set s_4 and s_2
$^+F_4$(B): conjugate set s_4 kinkbands

Fig. 55. Schematic stress-time diagram showing the relation between successive Hercynian deformation phases.

Origin of the concentric tendency of s_2

If we compare the large Cambro-Ordovician domes of the Pyrenees (Fig. 1) with each other, it is striking that they generally have an exposed core of gneisses mantled by micaschists and phyllites in which flat-lying schistosity or cleavage planes have developed, characteristic of the infrastructure (Zwart 1963), e.g. the Salat-Pallaresa anticlinorium, containing the Aston and Hospitalet gneiss massifs with their mantling metamorphic rocks as infrastructural core.

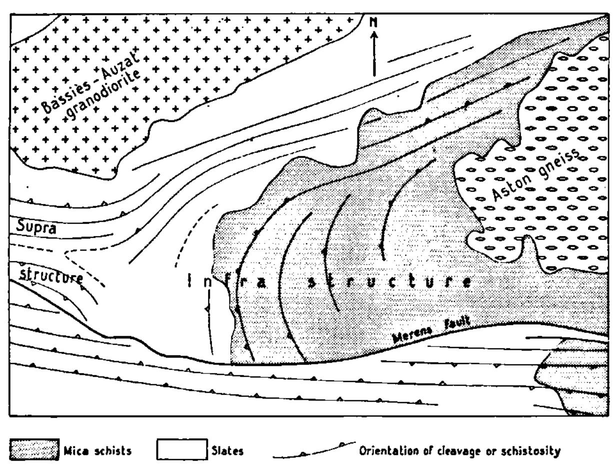


Fig. 56. Trend of cleavage or schistosity in infra- and supra-structure in the transition zone between the Aston massif and the Salat-Pallaresa anticlinorium.

Outside the infrastructure, in the suprastructure, steep, roughly E-W striking cleavage planes have developed. There is a transition zone of attitudes between the flat-lying cleavages of the infrastructure and the steeply dipping cleavages of the suprastructure. The latter often show the tendency to strike parallel to the infrastructural core as shown in Fig. 56 and also parallel to the general outline of the entire Cambro-Ordovician dome. This affinity has been called the concentric tendency. With increasing distance from the infrastructural core, however, this tendency weakens and is gradually replaced by a more strict E-W position of the steep cleavage plane. We shall call this affinity to the E-W position the linear tendency. Both tendencies occur in the map area.

subsequent N-S folds. This general idea of a N-S compression being transferred in the infrastructure into an E-W compression is schematised in Fig. 57. In this figure the cleavage in the "zone of linear tendency" is vertical as is the case west of the Aston massif. In our area the corresponding s_2 cleavage dips about 45° north and the entire structure is also thought to be inclined. This process provides a solution for the concentric arrangement of the strike of the s_2 cleavage in the suprastructure observed in this area and especially in the Orri dome. It is suggested here that in

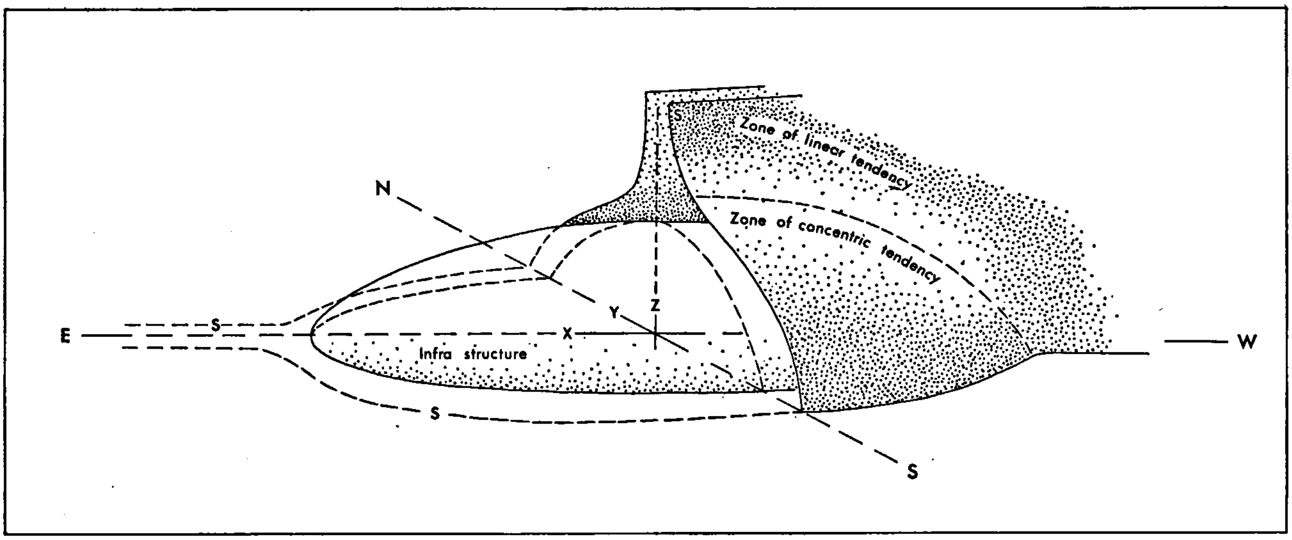


Fig. 57. Schematic diagram showing the relation between the infrastructure and the zones of concentric and linear tendency in the suprastructure.

The origin of the concentric tendency in the suprastructure can be found in Oele's (1966) description of the transition zone between the Aston massif and the Salat-Pallaresa anticlinorium. Oele found that in the infrastructure of the Aston massif the general N-S compression gave rise to an E-W elongated fabric in the micaschists. According to Oele (1966), the E-W elongation caused an underthrust movement and

the Orri dome too, an infrastructural core was developed during the N-S compression of the second folding phase. This N-S compression caused an E-W extension in the infrastructure and consequently a compression east and west of it. The total result of the process would have been a radial compression around the entire infrastructure which could account for the concentric tendency of s_2 in the suprastructure.

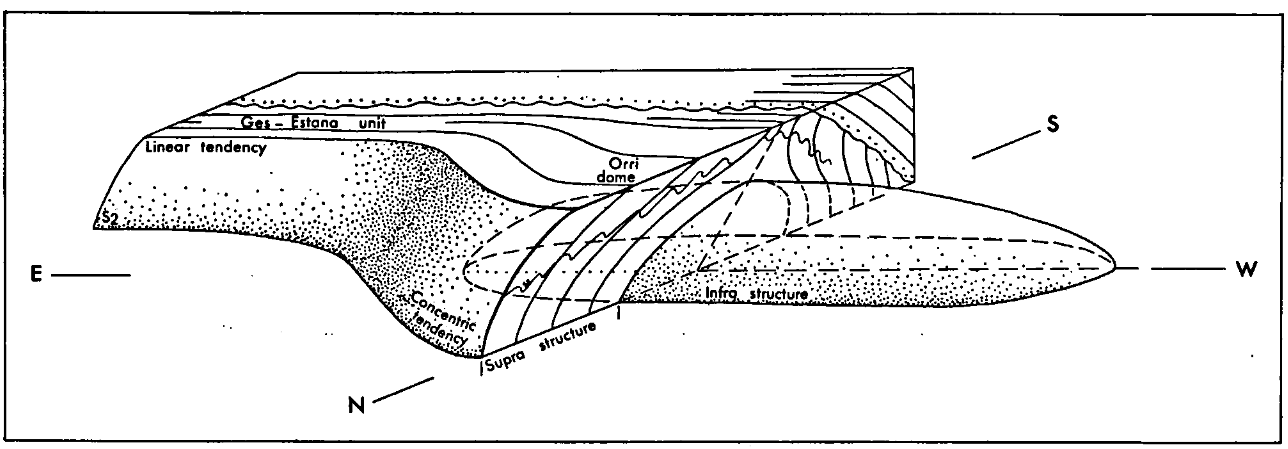


Fig. 58. Diagram showing the relation between the Orri dome, the Ges-Estana unit and the inferred infrastructure.

The concentric tendency of s_2 prevails in the anticlinal areas of the Orri dome, the southern border of the Massana anticline and the northern and southern borders of the Rabassa dome. The linear tendency prevails in the western termination of the Rabassa dome and in the Ges-Estana unit. The Ges-Estana unit is considered to be the linear continuation of the concentric Orri dome structure (Fig. 58). In the synclinal areas the s_2 orientation of the adjacent anticlinal area is adopted.

POST-HERCYNIAN STRUCTURES

Post-Hercynian structures are of a completely different type to the Hercynian structures described above. Folds are usually very gentle and often only a monocline is visible, while the rocks are not cleaved or metamorphosed.

Post-Pyrenean structures, comprising the deformation of Oligocene and Neogene sediments will be dealt with separately in Chapter IV, together with the geomorphology.

Structural units

Post-Hercynian structures have been divided into two zones of different structural type (Figs. 32 and 59):
The northern zone with structures directly attached to the Hercynian basement, comprising:
Querforadat unit
Serch unit
Cadi monocline (northern zone).

The southern zone with structures detached from the Hercynian basement by the highly incompetent gypsum of the Pont de Suert Formation, comprising:
Cadi monocline (southern zone)
Saldes syncline
Pedraforca structure
Santa Fé syncline
Monsech de Tost structure
La Vansa zone
Hostalets zone

The presence or absence of gypsum has enabled the two zones to be satisfactorily demarcated everywhere except in the Sierra del Cadi where the choice is arbitrary.

The two major zones of different structural units thus defined differ essentially in character. Structural units of the northern zone exactly follow the deformation of the Hercynian basement, whereas structures of the southern zone follow this deformation to a limited extent, since diapiric movements play a very important role.

Querforadat unit

The Querforadat unit is largely a southward dipping monocline, bounded in the north by a gently to steeply dipping unconformity plane and in the south by the steep northward dipping Cavá fault. Near this fault the strate often show a gentle dip to the north, probably related to drag.

The displacement along this fault has been rotational since the apparent throw is about 500 m east of Cavá and gradually decreases towards the west until the fault dies out at the Grau. This rotational movement along the fault face has caused the eastward dip of the strata west of Cavá. Vertical faults occurring in the Turo de Call Pubill can generally be explained as step faults emphasising the general structure of the monocline and the Cavá fault. The E-W fault in the Serrat de Mosbé is antithetic to the Cavá fault.

Serch unit

The Serch unit consists only of the volcanic rocks of the Erill-Castell Formation and, because of the lack of layering, internal structures can not be detected in it although it is possible to demonstrate the steep topography of the Hercynian unconformity below. This unit occurs between two faults of the E-W trending step fault system of probable Neogene age.

Cadi monocline (northern zone)

The large southward dipping Cadi monocline is, in analogy with the Querforadat unit, sub-horizontal in the north and develops a steep southward dip towards the south. This general southward dip is sometimes disturbed, as, for instance, east of the Can Franch as a result of a NE-SW trending fault with a downthrown southeastern block (Appendix IV).

Vertical E-W faults, as may frequently be observed between Novés de Segre and Sabaña, generally show a downthrown southern block. A large upthrusted southern block may be observed crossing the valleys of the Rio Segre and the Rio Palleróls. Here, the combination of these two opposite movements brought about folding in the Bunter and in the limestones of the Pont de Suert Formation (cross-section 20).

Vertical NW-SE faults frequently occur, often cut off the E-W faults, and usually show a downthrown southwestern block.

The description of the southeastern part of the Cadi monocline, the Saldes syncline and the Pedraforca structure has largely been based on data from Guérin-Desjardins and Latreille (1961).

Cadi monocline (southern zone)

The southern zone of the large monocline of the Sierra del Cadi rests on the thick gypsum deposits of the Pont de Suert Formation (seen in the Rio de Bona valley) which Triassic, Jurassic and Cretaceous formations are structure and the Hercynian basement with its pre-Pont de Suert cover. To the south the monocline is bounded by the large south dipping Cadi thrust, along which Triassic, Jurassic and Cretaceous formations are thrust over the Eocene rocks. The western part of the Sierra del Cadi is intersected by vertical NW-SE trending faults, the westernmost fault of which clearly offsets the Cadi thrust. The strike of the westernmost segment of the structure curves markedly to the south, which is most probably due to a diapiric movement between Adrahent and Sisque.

North of Sisque and more or less in the continuation of the Cadi monocline a somewhat isolated block is

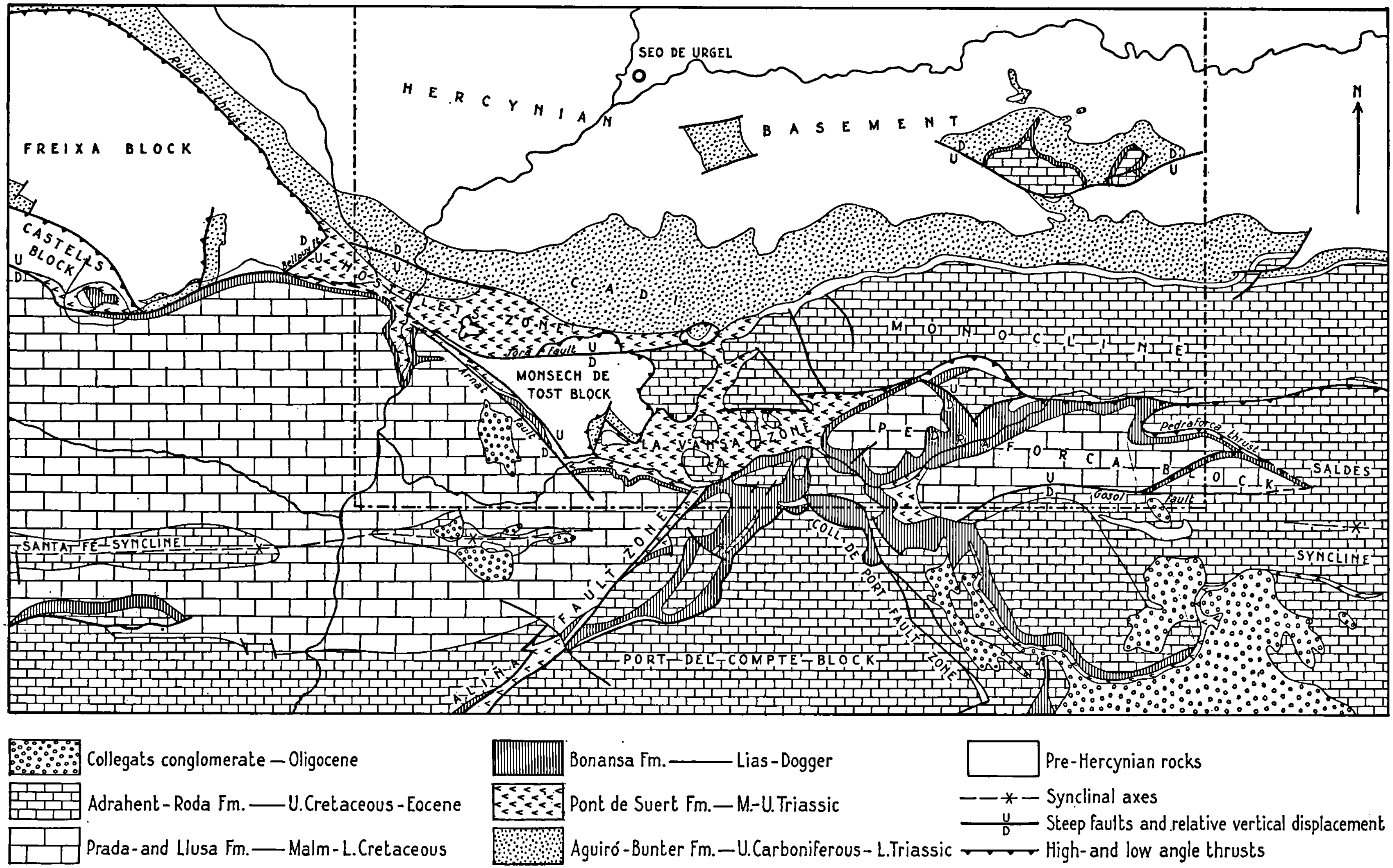


Fig. 59. Structural setting of post-Hercynian structures of sheet 10.

found, showing the same stratigraphic sequence as the Cadi. This Sisque structure has been intensively folded into a north-dipping anticline (cross-sections 14 and 15), most probably formed by its southward sliding over the gypsum of the Pont de Suert Formation. To the east this structure is overthrust by the Silurian and Devonian of the Monsech de Tost.

Saldes syncline
The Saldes syncline (Fig. 59), only a very small portion of which is shown on the map, is a large basin-like structure extending from the Rio Llobregat up to the area of Gosól. The syncline is thrust northwards over the Cadi monocline along the Cadi thrust.

Pedraforca structure
The structures of the Lower Cretaceous and Jurassic formations of the Pedraforca (just west of this map, see Fig. 59), Cotaróns and Peña del Aguila are almost completely surrounded by structural units in which no Lower Cretaceous deposits are present.

The Peña del Aguila (cross-section 6) consists of a roughly E-W striking syncline cut off in the northeast by a NW-SE trending fault zone which separates it from the Upper Cretaceous of the Cadinell.

The Cotaróns (cross-sections 3 and 4) consists of a syncline in the north which in its eastern part is thrust over the Eocene of the Cadi monocline. In the southern part an anticline has developed which is bounded by the steep E-W trending Gosól fault, bringing the structure into contact with the Saldes syncline.

The Pedraforca (cross-section 1) is composed of two south-dipping blocks of Lower Cretaceous and Jurassic formations thrust up northwards on the Upper Cretaceous of the Saldes syncline. In the south the structure is bounded by the steep Gosól fault, which also bounds the Cotaróns. The Pedraforca thrust dips southwards and must intersect the Gosól fault in the large unexposed area east of the Pedraforca. These structures then die out, no disturbances having been observed in the Saldes syncline further to the east.

Santa Fé syncline
The Santa Fé syncline (Fig. 59) of which only a small part of the north flank is shown on the map, is a very large E-W trending syncline which can be traced from much further west in the Pallaresa valley to the Segre valley. To the north, the syncline is down-faulted along steep faults which bring this structure into contact with the Castells and Freixa blocks west of the Segre and with the Monsech de Tost block east of the Segre. In the Segre valley south of Hostalets and Noves de Segre and in the La Vansa valley south of Barceloneta, normal contacts exist with the Pont de Suert Formation. To the southeast the syncline is bounded by the Aliña fault zone.

In the deeply incised valleys of the Segre and La Vansa rivers in the north flank of this syncline, several parasitic folds may be observed (cross sections 17 and 20).

Monsech de Tost
The Monsech de Tost is a block of Hercynian tectogene (p. 210), later separated from the rest of the mountain chain, which underwent strong deformation in the process. The boundaries between this block and the surrounding younger rocks are almost invariably faulted. As shown in Fig. 62 (p. 226), the block is more or less triangular, surrounded by other types of structures (Santa Fé, Hostalets, La Vansa and Sisque). Internally the block can also be subdivided into three distinct zones, again separated by faults.

Central triangle. – Along the SW side the unconformity plane, the Bunter and the Pont de Suert Formation are bent into a vertical or steep SW dip by movements of the sub-vertical Arnat fault. The Devonian formations below the unconformity also take part in this deformation as can be observed east of Coll de Arnat, where the relatively flat-lying cap of the Monsech de Tost is bent into a steep SW dip.

To the south, in the valley of the Rio de la Vansa, the Central triangel is cut off by a vertical E-W trending fault which brings it into contact with the Pont de Suert Formation.

To the east the Devonian of the Central triangle is thrust over the westward dipping strata of the Bunter and Pont de Suert Formation. More to the north this thrust separates the Central triangle from the Eastern zone. The Bunter below the thrust is cut off in the east by a vertical fault.

To the north the Central triangle is downfaulted along the vertical Tora fault and only locally, south of the Tora fault between Tora and Tost, can the Devonian be observed thrust over the Pont de Suert Formation.

Northern zone. – North of the Tora fault several slabs of Devonian, Silurian and Bunter may be observed, clearly resting on the Pont de Suert Formation of the Hostalets zone. The orientation of the unconformity plane and the Bunter in the different slabs varies greatly. The easternmost slab of Devonian is unconformably overlain by Bunter dipping SW which is overthrust by a second slab of Devonian. Directly east of Tora the unconformity is overturned and dips about 30° to the north (cross-section 17). The isolated sheet of Devonian north of Tost shows a steep N-S trending unconformity. Several small slabs of Bunter rest upon the gypsum and limestone of the Hostalets zone. The relatively large slab of Bunter and Pont de Suert limestone west of Adrahent is in a clearly inverted position and forms a gentle synform of Pont de Suert limestone with Bunter in the core, the Bunter conglomerates are just exposed at the top of the hill (cross-section 14).

Eastern zone. – The Eastern zone, which is separated from the Central triangle by a westward dipping thrust, also bears in its interior thrusts in which the Bunter locally takes part. The structure as a whole is thrust to the east over the Cadi Limestone of the Sisque structure

The unconformity plane with the Bunter, in the southern part of this zone, is overturned to the south (cross-section 16). The Bunter shows a conformable overturned contact with the Pont de Suert Formation and both formations are thrust over the La Vansa zone which here includes some slabs of Cadi Limestone. This southern part of the structure strongly suggests a sliding movement towards the south.

La Vansa zone

The La Vansa zone is the area where four important structural elements intersect: the Monsech de Tost thrust, the Cadi thrust, the Aliña fault zone and the Coll de Port fault zone (Fig. 59). The surrounding structural units are generally bounded by these faults and only locally can relatively undisturbed sedimentary contacts be observed. Strata in the adjoining units usually dip away from the La Vansa zone irrespective of their regional trend. This can best be observed from the curvature in the Cadi monocline south of Adrahent, and also from the eastern dip in the western part of the Peña del Aguila. These phenomena are an indication of the diapiric nature of the La Vansa zone.

Internally the zone consists of a large mass of gypsum and some limestone of the Pont de Suert Formation, which is folded and faulted in a rather chaotic way. In the field the limestones of this formation can only be followed over relatively short distances and only occasionally can large regular folds be observed in the gypsum.

North of the Rio de La Vansa several slabs of younger material, including beds ranging from Adrahent to Cadi Formation but usually with a very incomplete stratigraphic sequence, rest on top of the Pont de Suert Formation. These slabs usually dip gently southwards (cross-sections 12 and 14).

South of the Rio de La Vansa, between La Vansa and Tuxent, the Pont de Suert Formation is largely covered with Jurassic and Upper Cretaceous deposits also displaying many large and small gaps in the sequence.

Hostalets zone

The large outcrop of Pont de Suert Formation in the valleys of the Rio Segre and the Rio Tora de Tost and the area around Novés de Segre is defined as the Hostalets zone. This zone is bounded in the north by the Bunter of the Cadi monocline, partly by sedimentary and partly by steep fault contacts. West and south of Hostalets the normal stratigraphic succession to the north flank of the Santa Fé syncline can be found. East of the Segre the southern boundary is formed by the Tora fault and the Monsech de Tost thrust.

The Pont de Suert Formation, which in this zone too, consists largely of gypsum, is strongly folded and faulted, but in the area north of Hostalets and Tost disturbances are limited and structures can still be mapped. A series of E-W trending folds with northward dipping axial planes can be observed, suggesting a collapse of the relatively steep general southward dip (cross-section 18).

Northeast of Hostalets a N-S trending fold can be observed with an eastward dipping axial plane. South of Hostalets the beds of the adjoining Santa Fé syncline generally deviate away from the Segre valley to an eastward or westward dip. These two features are probably part of a N-S elongated diapir in the valley itself.

Historical review

The post-Hercynian structures so far described have previously been studied by many authors and because of the complexity of the structures this has resulted in widely diverse theories. The two major topics of discussion were the emplacement of the isolated Hercynian blocks, such as the Monsech de Tost, in a zone of mainly Triassic sediments called the Nogueras zone and the question of autochthony of the various facies of the Cretaceous. A brief review of all theories that have been put forward concerning these structures will therefore be useful.

Dalloni (1913, 1930) who was the first to map the southern Pyrenees introduced the theory of the "nappe de Nogueras" in which he included all Hercynian blocks found in a narrow E-W trending zone between the Esera and the Segre valleys and which is separated from the axial zone in the north by a steep flexure and bounded in the south by the Jurassic and Cretaceous strata. In this zone, where Hercynian blocks are intimately folded together with post-Hercynian rocks, Dalloni found recumbent southward facing folds from which he concluded that the Hercynian blocks, including the Monsech de Tost, formed part of a large nappe which thrust from the axial zone to the south. No doubts were expressed by this author concerning the autochthony of the Cretaceous.

Jacob et al. (1926) suggested a para-autochthonous origin for most of the Hercynian massifs of the "nappe de Nogueras" of Dalloni, but considered the Monsech de Tost to be a fragment of Hercynian basement that took part in a very large northward thrust of the more completely developed sequence of the Jurassic and Cretaceous the "Serie de Pedraforca" over the Tertiary and less completely developed Cretaceous sequence the "Serie de Cadi". This thrust was thought to have originated in the Ebro basin. The structure of the Pedraforca itself was not investigated by these authors.

Misch (1934) introduced the term "Nogueras zone" for the zone in which Dalloni situated his "nappe de Nogueras". All Hercynian blocks in the Nogueras zone are considered autochthonous or para-autochthonous by this author, but he did not extend his study east of the Segre to the Monsech de Tost. The Cretaceous was generally also considered as autochthonous.

Ashauer (1934) who mainly studied the region east of the Segre considered the Monsech de Tost as well as the Freixa and Castells blocks (Fig. 59) to be autochthonous and suggested an emplacement by upthrusts to the north. The Cretaceous of the "Serie de Pedraforca" was also considered autochthonous by this author and the fault contacts with the Cretaceous and

Tertiary of the Sierra del Cadi were considered as steep upthrusts.

The views of Misch and Ashauer were shared by Llopis Lladó (1936, quoted by Rios, 1959) and also by Jacob (1935).

Birot (1937) independently arrived at the same conclusion with regard to the autochthonity of the Cretaceous and Almela & Rios (1947) too, accepted these ideas.

De Sitter (1959) suggested that some of the Hercynian blocks, such as the Freixa and Castells blocks (Coma de To), are autochthonous, whereas others are drifting slabs.

Guérin-Desjardins & Latreille (1961), who made the first detailed map of the area south of the Cadi between the Rio Segre and the Rio Llobregat, considered the Monsech de Tost thrust and the Cadi thrust to be one steep thrust along which the Monsech de Tost and the "Serie de Pedraforca" were thrust up to the north. The Monsech de Tost is considered to be completely detached from its basement. Two different solutions for the emplacement of the Pedraforca structure were proposed by these authors. In the first solution, which they accepted, the Lower Cretaceous and Jurassic strata of the Pedraforca structure thrust up to the north along the Cadi fault, and subsequently the part east of Gosól slid down to the south over the Upper Cretaceous formations. The upper part of the Saldes syncline was cut off by this sliding movement (Fig. 60A). In this complex process of emplacement the relationship with the Cotaróns, west of Gosól, which was only upthrust, remains obscure.

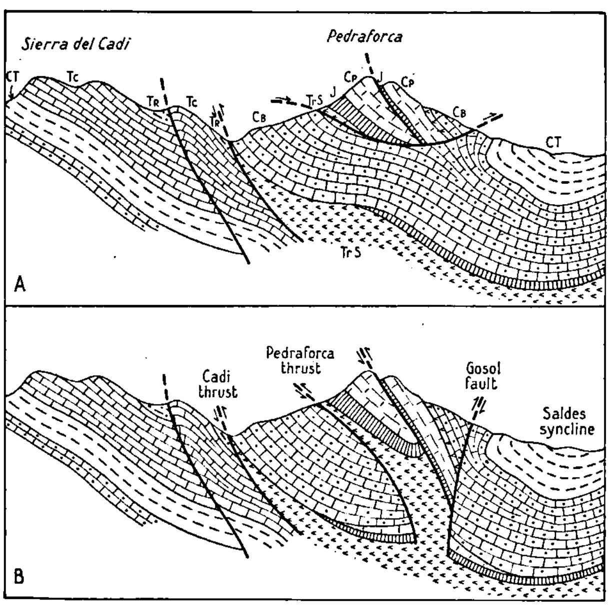


Fig. 60. Cross-sections showing two different interpretations of the Pedraforca structure: (A) according to Guerin-Desjardins & Latreille (1961), (B) according to the present author.

The second solution, which was rejected by these authors, implies that the Pedraforca structure was thrust up to the north as one unit, partly along the Cadi thrust over the Eocene of the Cadi monocline and partly along the Pedraforca thrust over the Bona Formation of the Saldes syncline. To the south the structure was thrust up along the steep northward dipping Gosól fault, thus forming the wedgelike structure of the Pedraforca. It is this explanation that will be adopted by the present author (Fig. 60B).

Diederix (1963), who first made a detailed map of the Monsech de Tost, took up the ideas of Dalloni and for this Hercynian block proposed a sliding movement from the north from the uplifted axial zone. His arguments were based mainly on the overturned unconformity planes and inverted Bunter. De Sitter (1964, 1965) accepted this sliding theory for the Monsech de Tost, but considered the Freixa and Castells blocks as autochthonous. Based upon the same arguments of overturned unconformity planes which can be interpreted as parts of overturned folds that slid down from the north ("têtes plongeantes"), Seguret (1964) also took up Dalloni's theory of the "nappe de Nogueras". Recently Solé Sugrañes & Santanach Prat (1970) also concluded on gliding from the north.

For the Cretaceous Seguret (1969) also proposed a nappe theory which is the exact counterpart of the theory of Jacob et al (1926). Seguret suggests a sliding down of the Cretaceous of the "Serie de Pedraforca", comprising the Saldes and Santa Fé synclines, from the axial zone over the autochthonous "Serie de Cadi" to the south. In this theory the "Serie de Cadi" is also thought to extend below the "Serie de Pedraforca" up to the Ebro basin. The Pedraforca structure is considered to be a higher nappe resting upon the first one. On the axial zone north of the Sierra del Cadi, however, no Jurassic or Lower Cretaceous was developed as far as can be ascertained and therefore this theory would seem to require unlikely factors making it rather improbable.

Post-Hercynian deformation phases

The structures observed in post-Hercynian formations are often merely called Alpine or Pyrenean structures. Although the most important deformation undoubtedly took place during the Pyrenean phase it is by no means the only deformation phase. With the description of the post-Hercynian formations it is made clear that the successive longitudinal E-W trending sedimentary basins showed a gradual shift towards the south and that the deformation of these rocks must have commenced with the sinking of these basins. The basin subsidence and shift towards the south can be read from the sections in Fig. 61.

Upper Carboniferous, Permian and Triassic deformations

From the sections in Fig. 61 it can be read that the subsidence of the basins caused a steepening of their northern border. In the western part of the area the sinking of the Upper Carboniferous and Permian basins caused a flexure in the unconformity with a dip of 30°–35° to the south. Subsidence during the Middle

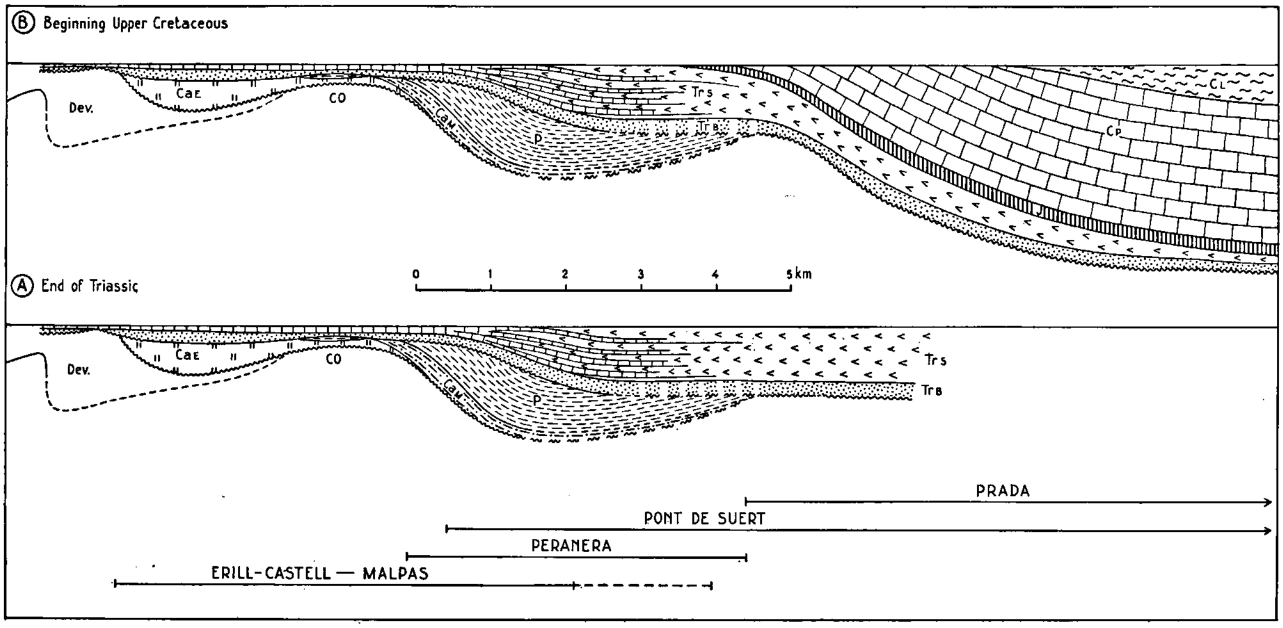


Fig. 61. Schematic N-S cross-sections through post-Hercynian formations showing the basin subsidence shifting to the south.

and Upper Triassic resulted in an additional dip of 10°–15°. The observed southward dip of the unconformity of 45° can thus largely be explained by basin subsidence before the Jurassic.

In the east, in the Sierra del Cadi, a dip of 25° can be explained by basin subsidence during the Upper Carboniferous and Permian, whereas the remaining 5°–15° are of Upper Triassic or younger age. The age of this remaining deformation in the east is not known, as there are no data relating to the thickness of the Pont de Suert Formation south of the Sierra del Cadi.

The southern boundaries of the Upper Carboniferous and Permian basins are not well known. The absence of the Upper Carboniferous and Permian Formations in the Monsech de Tost structure, however, indicates that these formations either wedged out toward the south against the Monsech de Tost or were cut off by a fault of Permian age which was possibly the predecessor of the Pyrenean Monsech de Tost thrust. The former possibility is shown in Fig. 61.

Upper Jurassic and Lower Cretaceous deformations
In the north flank of the Santa Fé syncline the strong wedging out to the north of the Prada and Llusa Formations (Fig. 61 and Fig. 23 p. 189), indicates that at the beginning of the Upper Cretaceous this flank had a maximum dip of 33°. Since the Jurassic and Lower Cretaceous formations rest upon the gypsum deposits of the Pont de Suert Formation, structures in general will probably be strongly influenced by diapirism. Little is, however, known, about the thickness of the Pont de Suert Formation below the Jurassic, and it is difficult to decide to what extent the Lower Cretaceous basin should be attributed to diapiric movements on the one hand or to movements in the basement on the other. Whichever the exact explanation may be, it seems logical to suppose that at least part of the subsidence of the Lower Cretaceous basin was due to the older sediments sinking into the highly incompetent mass of gypsum, squeezing out all mobile material until the Jurassic practically rested upon the Bunter. The squeezed out gypsum must have migrated to areas of lesser sedimentation, reinforcing that tendency. Such a process could explain the very pronounced Lower Cretaceous basins (e.g. Pedraforca structure and Santa Fé syncline), bounded by areas where Prada or Llusa Formations are absent (e.g. Saldes syncline and Port del Compte block, Fig. 59). The remarkably straight basin boundaries, often coinciding with observable fault zones such as the Cadi and Pedraforca thrusts, the Gosól fault and the Aliña fault zone, probably represent synsedimentary fault zones.

Upper Cretaceous and Lower Tertiary (pre-Pyrenean) deformations
The differences in thickness between Upper Cretaceous formations north and south of the Sierra del Cadi (Fig. 25, p. 191), indicate that in this period, too, some differential subsidence occurred that could again somewhat steepen the southward dip.

Irregular sedimentation due to diapiric movements may be observed in the La Vansa zone. The very incomplete sequence of Upper Cretaceous and Tertiary formations, found in the slabs that rest on the Pont de Suert Formation in this zone, is most probably caused by diapirism. These differences, the dip of strata of the surrounding structures away from the La Vansa zone, the chaotic internal structure of this zone and its location at the intersection of several important partly synsedimentary faults, make it highly probable that the La Vansa zone is one large diapir, which has been active at least during the Upper Cretaceous and Lower Tertiary and probably even during the Lower Cretaceous.

Pyrenean deformation

The main folding phase, that deformed the early structures in the unstable sedimentary basins of the southern Pyrenees, began in the Upper Eocene and continued in several zones during the Oligocene (Dalloni, 1930; Birot, 1937; Guérin-Desjardins & Latreille, 1961).

As can be understood from the description of the structural units and their relationship, large thrusts and vertical faults were formed, often accompanied by diapirism. Three important fault directions can be observed in this area: E-W, NW-SE and NE-SW.

These fault directions frequently occur in the Hercynian basement in the axial zone and they are therefore also believed to exist in the basement below the post-Hercynian structures, which can be proved by the occurrence of the Monsech de Tost amid these structures. It is not believed, however, that the post-Triassic rocks will follow the structures in the basement as closely as suggested by Guérin-Desjardins & Latreille. It is rather suggested that fault movements in the basement initiated diapiric movements which caused more complicated structures on top of the Pont de Suert Formation.

Emplacement of the Monsech de Tost. – The Monsech de Tost block must have been emplaced during the Pyrenean deformation, since it rests partly on the Lower Tertiary Cadi Formation.

As can be concluded from the historical review of the theories on the post-Hercynian deformation, there remain two conflicting theories which deal with the emplacement of this structure, one theory in favour of a slide from the north and one theory that implies an emplacement by a steep upthrust to the north. To decide what mode of emplacement is most likely we shall now compare significant features of this structure and of the axial zone.

The Devonian of the Monsech de Tost shows marked differences from the nearest Devonian of the axial zone in the Segre valley (p. 178 and Appendix II). In the Monsech de Tost the lower part of the Rueda Formation consists mainly of silty material and is poor in limestone, brachiopods are remarkably abundant. Only the uppermost part of this formation is developed as impure limestone. The Basibé Formation is developed as an almost black limestone. In the Segre valley the Rueda Formation consists mainly of impure limestones, and brachiopods have only occasionally been found. The Basibé Formation is developed in this area as a light grey limestone. Dykes, which are a common feature in the Devonian of the Segre valley, are completely absent in the Monsech de Tost.

It could be argued that the Devonian of the Monsech de Tost need not come from the area of the Segre valley where Devonian is still found, but could have slid down, during the Pyrenean deformation, from any location on the Orri dome where a differently developed

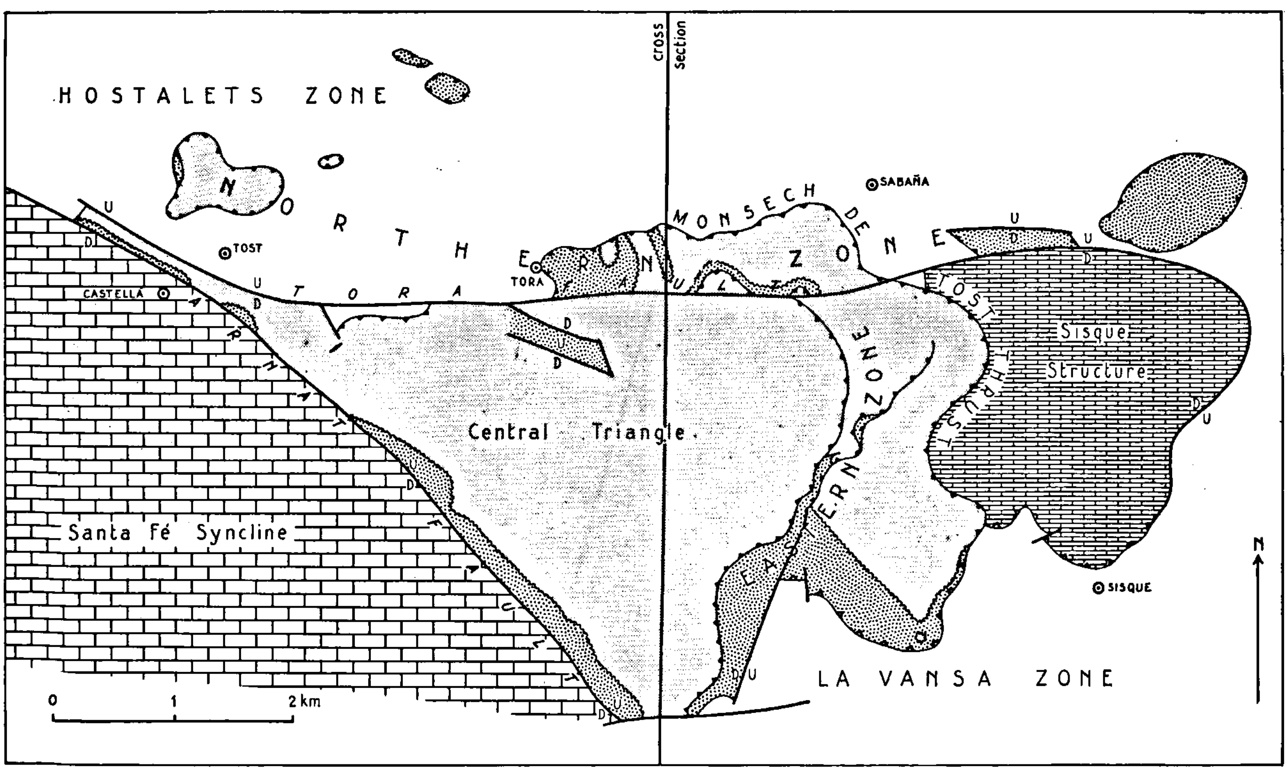


Fig. 62. Structural units of the Monsech de Tost area.

Devonian could have existed. However, for the following reasons this is very unlikely: In the first place a more southern origin is improbable because these areas must have been covered with Upper Carboniferous formations and with the Peranera Formation, while the Monsech de Tost is directly covered with Bunter. Secondly, the Hercynian structure of the Monsech de Tost has been analysed as a gravity structure that, already during the Hercynian orogeny slid down in a synclinal area (p. 216) (Fig. 63A).

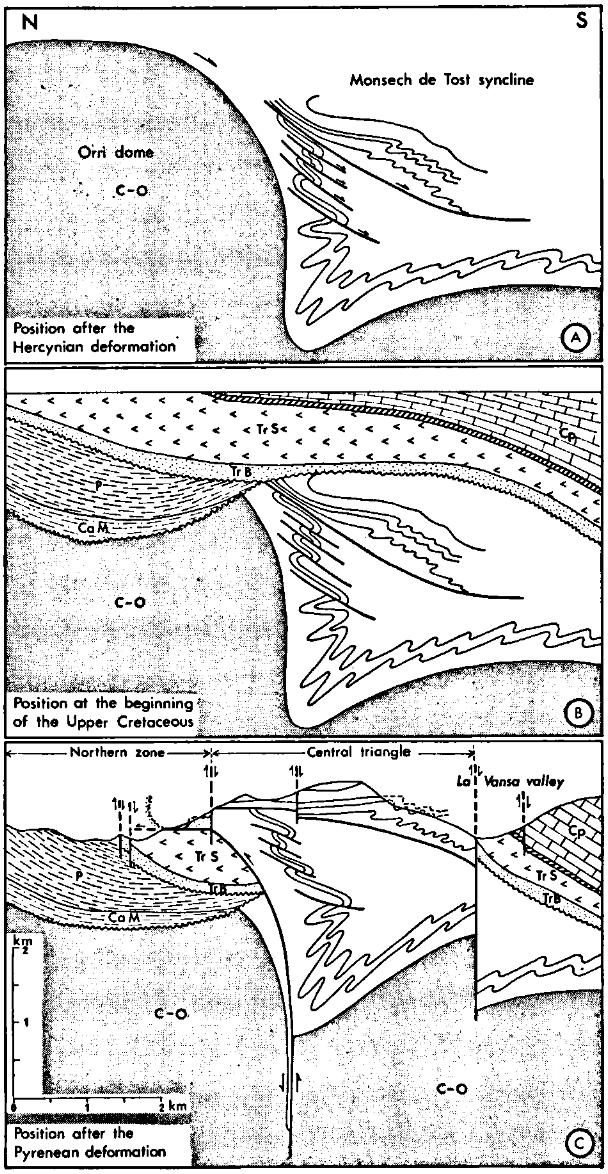


Fig. 63. Schematic cross-sections demonstrating three stages in the development of the Monsech de Tost.

The absence of Upper Carboniferous formations and the Peranera Formation in the Monsech de Tost cannot be used as proof of a northern origin, since it is known, from the western part of the Nogueras zone, that these formations have been deposited in a narrow E-W elongated basin there (Mey, 1968), Fig. 18, 19 and 61).

The structural features of the post-Hercynian structures of the Monsech de Tost are less convincing and are not conclusive either one way or the other. Along the southwestern boundary of the Central triangle the SW dipping unconformity plane has been turned into a vertical position parallel to the Arnat fault and nowhere becomes overturned as is sometimes suggested by other authors. This implies that the Central triangle has no "tête plongeante" suggesting a sliding movement. The vertical position of the unconformity plane can be explained completely by drag along the Arnat fault.

The stratigraphic hiatus between the Monsech de Tost and the Santa Fé syncline is very limited. If the sliding hypotheses were followed this very small hiatus would be pure coincidence, which seems unacceptable.

In the Northern zone inverted unconformity planes and strata of Bunter often occur and could be interpreted as parts of recumbent folds occurring at the front of the northward thrust. These structures are, however, very fragmentary.

The Eastern zone provides little information about the thrust direction over the Sisque structure. The southern part of this zone, however, shows an unconformity plane which is overturned to the south (cross section 16). Undoubtedly this part of the structure slid somewhat to the south, but as can be seen from the comparable Sisque structure, which certainly constitutes the continuation of the Cadi monocline, the necessary movement can be very limited. Probably the location of the La Vansa diapir in this area stimulated the slide to the south.

Considering the above-mentioned features, in the opinion of the author there is no doubt concerning the (para-) autochthonous position of the Monsech de Tost so that the following tectonic events must have taken place (Figs. 62 and 63):

1. A steep northward upthrust of the basement, flattening into a gentle southward dip as the level of the Pont de Suert Formation is reached. Locally the Monsech de Tost is thrust over the "Serie de Cadi". At the front of the thrust, in the northern zone, recumbent folds and secondary thrusts developed. A thrust movement to the northeast must also have taken place, causing the Central triangle to be thrust over the Eastern zone.

2. Steep faults developed: a NE-SW trending fault cut into the southern part of the Eastern zone and the eastern part slid down to the south together with the Sisque structure. The Central triangle, the Eastern zone and the Sisque structure were all downfaulted along the Tora fault, which obscures the western continuation of the Monsech de Tost thrust. The Santa Fé syncline was downfaulted along the Arnat fault.

It is not unlikely that the two easternmost slabs of Bunter and Pont de Suert limestone have anything to do with the Monsech de Tost thrust, since their location north of the Sisque structure implies that these slabs were covered by this structure at the time the Monsech de Tost was thrust to the north. The two slabs probably slid down to the south together with the Sisque structure, in the process of which they became inverted.

The relationship between the Monsech de Tost and the Freixa and Castells blocks. – The structure of the large Freixa and Castells blocks west of the Segre valley (Boersma, 1968a) is in many respects comparable to the structure of the Monsech de Tost (Fig. 59). (The structures of these two blocks west of the Segre will be dealt with in detail in the description of sheet 9 which is in preparation). The Freixa block is thought to have thrust to the north along the steep Rubio thrust, which is the equivalent of the Monsech de Tost thrust east of the Segre. The continuation of the Rubio thrust to the east is cut off by the NE-SW trending Bellpuy fault along which the Hostalets zone is uplifted. To the south the Freixa and Castells blocks are bounded by the steep La Guardia fault which is considered to be the continuation of the Arnat fault, thus forming one large curved vertical fault generally separating the Jurassic and Lower Cretaceous in the south from the Hercynian blocks in the north. The interaction of the La Guardia-Arnat fault and the Tora and Bellpuy faults explains the absence of pre-Hercynian rocks in the area of Hostalets and Novés de Segre.

The Monsech de Tost and Rubio thrusts form the boundary between the Orri dome with its post-Hercynian cover in the north and the upthrusted Devonian blocks in the south. Since Cambro-Ordovician and Devonian structural units are always separated by a zone of weakness, i.e. the Silurian, it is plausible to suggest that the two thrusts were determined by the Silurian zone of the southern border of the Orri dome which is now covered by post-Hercynian rocks. This view is also expressed in Fig. 63.

Emplacement of the Pedraforca structure. – For the emplacement of the Pedraforca structure amidst the Upper Cretaceous and Tertiary formations of the surrounding structural units, the second solution, proposed and rejected by Guérin-Desjardins & Latreille (1961) (Fig. 60-B), effecting a simple upthrust is favoured for the following reasons:

The Pedraforca and Cotaróns belong to one indivisible structure, of which the Cotaróns is a simply upthrusted block, and therefore the same holds for the Pedraforca. The complicated aberrant mode of emplacement suggested for the Pedraforca by Guérin-Desjardins & Latreille (see p. 224) should have created complicated structures in the area around Gosól; such structures have not, however, been observed.

The main objection of these authors against a simple upthrust, viz. that no disturbance could be observed in the Saldes syncline further to the east, can be disproved because the Pedraforca thrust and the Gosól fault have opposite movements and will neutralize each other in the area of intersection east of the Pedraforca.

We therefore conclude that the entire Pedraforca structure was upthrusted as one unit. This movement could have been facilitated by diapirism of the underlying gypsum deposits of the Pont de Suert Formation.

Post-Pyrenean deformations

As already mentioned, the Pyrenean folding phase is not restricted to the Upper-Eocene but also continued during the Oligocene as can be observed south of this area where the Collegats Conglomerates are folded together with the Eocene and older rocks and the unconformity lies within these conglomerates. In the map area these conglomerates, which unconformably overlie the Prada Limestone, are only slightly tilted to the north.

Younger deformations have been of more importance in this area. Since these deformations should be discussed together with the disturbance of the Neogene deposits and the related erosion surfaces, they will be dealt with together in the chapter on geomorphology

CHAPTER IV

GEOMORPHOLOGY

PRE-GLACIAL EROSION SURFACES, THE SEO DE URGEL BASIN AND THEIR DEFORMATION

During the Pyrenean orogeny in late Eocene times, the Pyrenees were folded and uplifted above sea-level allowing severe erosion to form the steep relief now buried below the fluvial piemont deposits of Oligocene conglomerates. The topography was gradually subdued, to become gently undulating in the Miocene. According to current opinion the axial zone was, however, domed up during late Miocene, causing a general dip of erosion surfaces toward either side of the mountain chain (de Sitter, 1956). Fault movements accompanied this uplift, causing intramontane sedimentary basins.

Pre-glacial erosion surfaces

Pre-glacial erosion surfaces, quite abundant in this area (Appendix V, morphological map), show beautifully on aerial photographs and can be recognized by their gently undulating topography with slopes of up to a maximum of 20° in various directions. They range in altitude from 2900 m (Tossa Plana) to 800 m in the Seo de Urgel basin, but are not entirely continuous over

PRESENT AUTHOR	BOISSEVAIN 1934	PANNEKOEK 1937	BIROT 1937	NUSSBAUM 1946
Tossa Plana surface 2900 - 2100m	Niveau des Crêtes 3100 - 2700m Upper Eocene	Gipfelflurniveau ±2900m Upper Miocene (Pontien)	Haute Surface de Carlit 2900 - 2700m Triassic ? Cretaceous ? Post - Pyrenean ?	Gipfelflur 2900 - 2150m Upper Eocene-Lower Oligocene
	?	?		
Sierra de Arcabell surface 2200 - 1900m	Niveau des fonds des cirques 2300 - 1950m Upper Miocene (Pontien)	Karbodenniveau 2400 - 2000 Upper Miocene (Pontien)	Niveau de Raset de Puig Pedros 2250 - 2000m	Camporeils - Niveau 2300 - 1900m Oligocene
Vilanova surface 1700 - 1000m Miocene	Niveau de la Percha 1712 - 1577 m Pliocene	Perche - System ±1600m Pliocene (Plaisancien)	Niveau de la Perche 1700 - 1150m Upper Miocene (Pontien or pre - Pontien)	Perche - Niveau 1600 - 1300m Upper Oligocene - Lower Miocene
Calvinyá surface 1400 - 800m Upper Miocene	Terrasse principale ±1230m Pleistocene (Sicilian)			Pliocäne Terrassen 200m above valley floor

Fig. 64. Interpretation of erosion surfaces by various authors.

this interval because of later erosion. Discontinuities in the surfaces are caused partly by lithological differences and partly by successive erosion phases of which the Pleistocene glaciation is the most important.

Four groups of correlatable surfaces can be distinguished, occurring in a certain altitude range. These surfaces, which are listed in Fig. 64, have been described before by several authors and interpreted and correlated in various ways. Especially the surfaces east of the map area, around the Neogene Cerdaña basin, have been extensively studied by other authors who made correlations with the deposits in this basin.

Tossa Plana surface. – The Tossa Plana surface is the highest, being found in an altitude range of 2900–2100 m mainly in the northeast. At this altitude it is mainly grass-covered and dissected by steep walls of cirques and glacial valleys, few residual mountain peaks rising above it. South of Andorra the surface generally dips less than 15° S, whereas in Andorra it is almost horizontal or dips northwest towards the Valira valley, suggesting a relationship with the present river system.

Boissevain (1934), who did not recognize this surface as such, interpreted the highest peaks ranging from 2700–3100 m, including the highest levels of the Tossa Plana surface, as remnants of a "Niveau des Crêtes" and suggested an Upper Eocene age. Pannekoek (1937) suggested an Upper Miocene age for this same surface ("Gipfelflurniveau").

East of the map area, Birot (1937) defined the "Haute Surface de Carlit" ranging from 2900–2100 m which, as may be confirmed from aerial photographs, is the equivalent of the Tossa Plana surface. He suggested several possible ages for this surface; Triassic, Cretaceous or post-Pyrenean. Nussbaum (1946), who extended the morphologic study as far as the Valira valley, called this uppermost surface "Gipfelflur" and correlated it with the lower (conformable) conglomerate sequence, possibly of Upper Eocene-Lower Oligocene age, south of the Sierra del Cadi.

The erosion surfaces mapped by Zwart (1965), north of the map area, are at least partly the equivalent of the Tossa Plana surface.

Sierra de Arcabell surface. – In the northern part of the map area the Sierra de Arcabell surface ranges in altitude from 2300–1900 m and is separated from the higher Tossa Plana surface by a steeper slope, 100–200 m in height. It is dissected by glacial cirques and river valley slopes. It generally dips gently towards the main valleys and is often somewhat rougher than the Tossa Plana surface, which can be accounted for by the lithological differences in the Devonian formations into which it is often eroded.

Boissevain (1934), who called this surface "Niveau des fonds des cirques" and Pannekoek (1937) "Karbodenniveau", both suggested an Upper Miocene age.

Nussbaum (1946), who called this surface "Camporeils-Niveau", correlated it with the upper (unconformable) conglomerate sequence, possibly of Oligocene age, south of the Sierra del Cadi.

The erosion surface near Montant (1300–1200 m), which since Oligocene conglomerates rest upon it, is undoubtedly of Oligocene age, should either be correlated with the Sierra de Arcabell surface or with the Tossa Plana surface. The nearby surface forming the summit of the Monsech de Tost (1600–1350 m) and that of the western part of the Sierra del Cadi (2300–1800 m) and near the Cotaróns (2100–1800 m) should probably be similarly correlated.

Vilanova surface. – The Vilanova surface, roughly 400–500 m above the Segre valley floor (Fig. 65), dips gently downstream from 1300 m near Beixech to 1000 m near the Seo de Urgel basin. It occurs both north and south of the Rio Segre and usually shows, in addition to the general dip downstream, a dip towards this river or occasionally towards a tributary. The surface is separated from the Sierra de Arcabell surface by a slope at least 200 m in height and is often incised by recent river valleys.

A correlation of this west-dipping surface with the almost completely horizontal sequence of levels defined by Boissevain (1934) is difficult. For example, he interpreted the surface at Llés as forming part of the "Niveau de la Percha" which he dated as Pliocene, whereas he named the surface at Vilanova "Terrasse principale" for which he suggested a Pleistocene (Sicilian?) age. A better correlation can be made with the "Niveau de la Perche" as defined by Birot (1937), since he suggested a gentle westerly dip more or less parallel to the river profile. For this revised "Niveau de la Perche", in which the surfaces at Vilanova and Toloríu are included, Birot (1937) suggested a Pontian or pre-Pontian age, a view which he supported by quoting a tradition, reported by Penck in 1894, according to which Miocene (?) lignite had been exploited on the surfaces at Vilanova and Toloríu. Nussbaum (1946) defined his "Perche-Niveau" in the same manner as Birot and suggested an even older age of Oligocene-Lower Miocene.

Seo de Urgel basin. – In the Seo de Urgel basin, situated in the depression around the confluence of the Segre and Valira rivers, Neogene sediments have been deposited (p. 193). The unconformity plane below the Neogene Ballestá Formation, which may be observed in several localities west of Seo de Urgel, displays a gently undulating surface in the centre of the basin; along the borders it gently slopes about 5–15° towards the centre. Around the Neogene deposits a subdued surface (Calvinya surface) slopes towards the basin and extends into the unconformity plane. This can be seen south of Calvinya, but north of Ballestá the erosion of

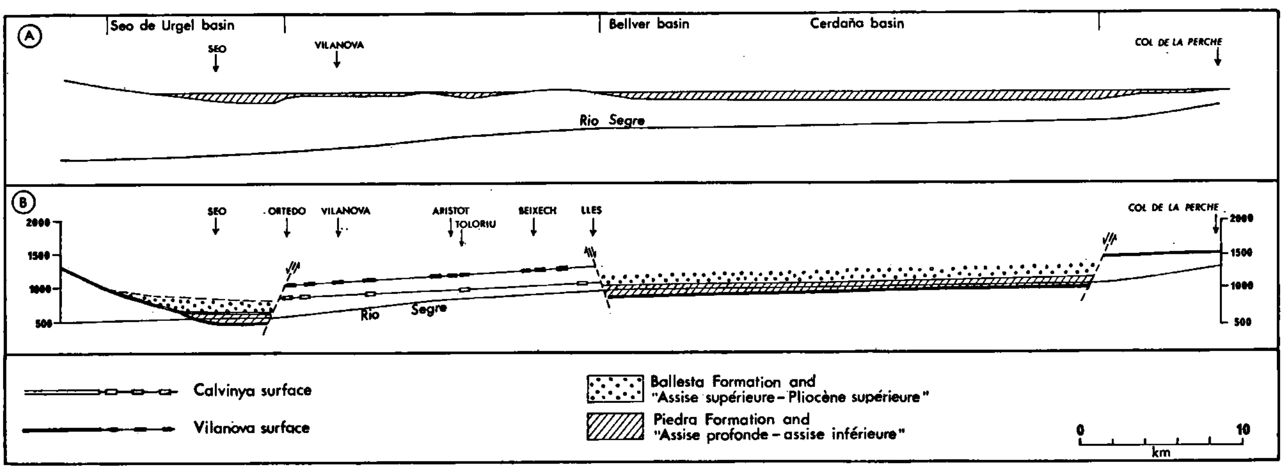


Fig. 65. Section along the Rio Segre between Coll de la Perche and Seo de Urgel, showing condition after the development of the Vilanova surface, (A), and the position of the Vilanova and Calvinya surfaces at present (B).

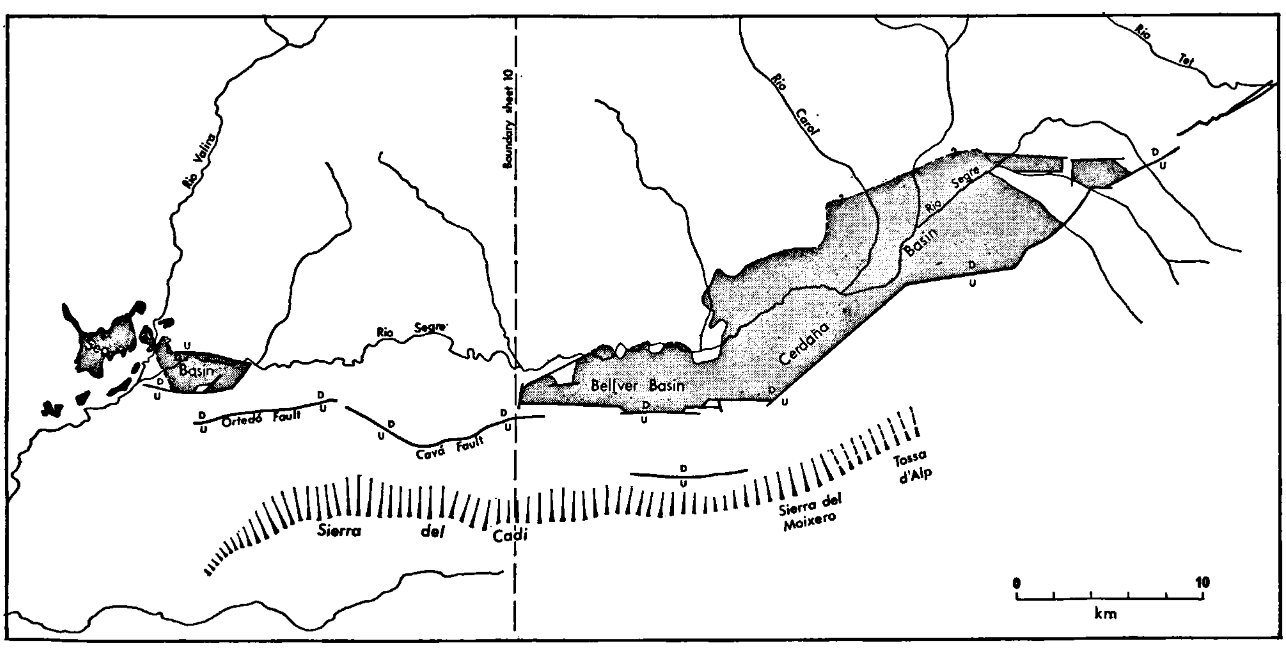


Fig. 66. Structural setting of the Neogene basins of the Segre valley.

the numerous gullies has almost completely destroyed this surface. In the region of Vilamitjana an extensive surface descends from the higher Vilanova surface (?) around 1400–1500 m to 900 or 800 m in the Seo de Urgel basin; south of Seo de Urgel remnants of this same surface also occur.

East of Seo de Urgel the basin is often bounded by faults which can be observed along the southern boundary and which must be of Miocene and Pliocene age. Along the northern boundary a fault can only be deduced from the straight row of facets along the road east of Seo de Urgel. To the east the basin is probably also bounded by a fault. The absence of contacts between erosion surfaces and Neogene deposits in this part of the basin can be explained by subsidence of the basin along these faults.

Erosion surfaces found 200–300 m above the Segre valley floor east of the Seo de Urgel basin should possibly be correlated with the Calvinya surface.

The age of the Calvinya surface is probably Upper Miocene or Pliocene, as it appears to be the continuation of the unconformity plane below the Ballestá Formation.

Upper Tertiary (post-Pyrenean) deformations

From the descriptions of the Upper Tertiary deposits, the structural setting of the Cerdaña and Seo de Urgel basins and the erosion surfaces it may be deduced that several deformations occurred after the Pyrenean phase.

The Upper Eocene-Oligocene conglomerates rest conformably upon the Eocene Roda Formation south of the Sierra del Cadi (Birot, 1937; Nussbaum, 1946; Guérin-Desjardins & Latreille, 1961) which indicates that the axial zone was already uplifted before the real folding in the marginal zones began.

This can also be proved from granodiorite boulders found in these conglomerates and which could only have come from the axial zone. These conformable conglomerates were folded together with the older Tertiary and Mesozoic formations and unconformably covered with younger conglomerates. Nussbaum (1946) suggested that these two conglomerate sequences resulted from the two earliest phases of uplift and should therefore be reflected by the two highest erosion surfaces, viz. the Tossa Plana and Sierra de Arcabell surfaces. It should, however, be realized that there is no proof for this detailed correlation.

The transport of the granodiorite boulders from the axial zone to the area south of the Sierra del Cadi implies that the high barrier of the Sierra del Cadi, Sierra Moixero and Tossa d'Alp (Fig. 66) did not exist during the deposition of these conglomerates and that at the time of this southward transport the E-W trending part of the Segre valley was not yet developed. This important conclusion deduced by Birot (1937) means that the Sierra del Cadi was uplifted with relation to the Segre valley during the Upper Oligocene (Nussbaum, 1946), Mio-Pliocene (Pannekoek, 1937) or Pliocene (Boissevain, 1934; Birot, 1937). This block faulting must have taken place mainly along the boundary faults of the Cerdaña and Seo de Urgel basins, the Cavá fault and the Ortedó fault (Fig. 66). Further west no fault movements can be observed and the Sierra del Cadi also terminates at this meridian.

The Vilanova surface, showing a clear relationship with the Segre valley could have formed as soon as a small subsidence had occurred. These movements were probably very limited during the Miocene as the deposits were generally not very coarse. The Vilanova surface, on which, according to Birot (1937) the lacustrine Upper Miocene sediments were deposited, was downfaulted further during the Pliocene along the faults mentioned above. Uplifted surfaces south of these faults were easily eroded away and during this time basins were filled up with coarse Pliocene deposits.

GLACIATIONS

During the Quaternary glaciations many glaciers were formed in the more elevated valleys of this area. A map showing the largest extent of these glaciations and the related snow line (Fig. 67) was compiled from data of Chevalier (1906), Nussbaum (1946) and Zandvliet (1960) combined with the author's observations. The remodelling of the pre-glacial landscape by the glaciers resulted in the typical U-shaped valleys, hanging tributary valleys, rock steps, cirques, cirque lakes, roches moutonnées, striae and moraines. It is not known how many glaciations are involved and recessional moraines, such as those of the Carol glacier near Puigcerda, which in previous literature were regarded as an indication of successive glaciations, merely represent stages in the last retreat of the glacier (Viers, 1963).

In the thick masses of coarse rock debris found in the high cirques, garlands and curved ridges, classified as rock glaciers, occur abundantly. The current opinion concerning these phenomena is that they are related to the glaciations, occurring as they do exclusively in formerly glaciated areas and have developed mainly from moraine material which slowly moved downwards with the glaciers. Movements which still occur after disappearance of the glaciers are most probably made possible by freezing of the interstitial water so that these rock glaciers can move more or less as real glaciers (Wahrhaftig & Cox, 1959; Blagbrough & Farkas, 1968; Embleton & King, 1968, p. 523–526).

POST-GLACIAL RELIEF

Post-glacial erosion is mainly evident in the steep incision of rivers in the valley floor of U-shaped, stepped and hanging valleys. Good examples of such gorges in rock steps can, for instance, be observed north of Os de Civis in the Rio Saturia, north of Escaldes in the Riu Valira del Nort and east of this village in the Riu Valira d'Orient.

Post-glacial deposition can be observed in the filling up of cirque lakes and oversteepened glacial valleys and in the occurrence of alluvial fans in these valleys.

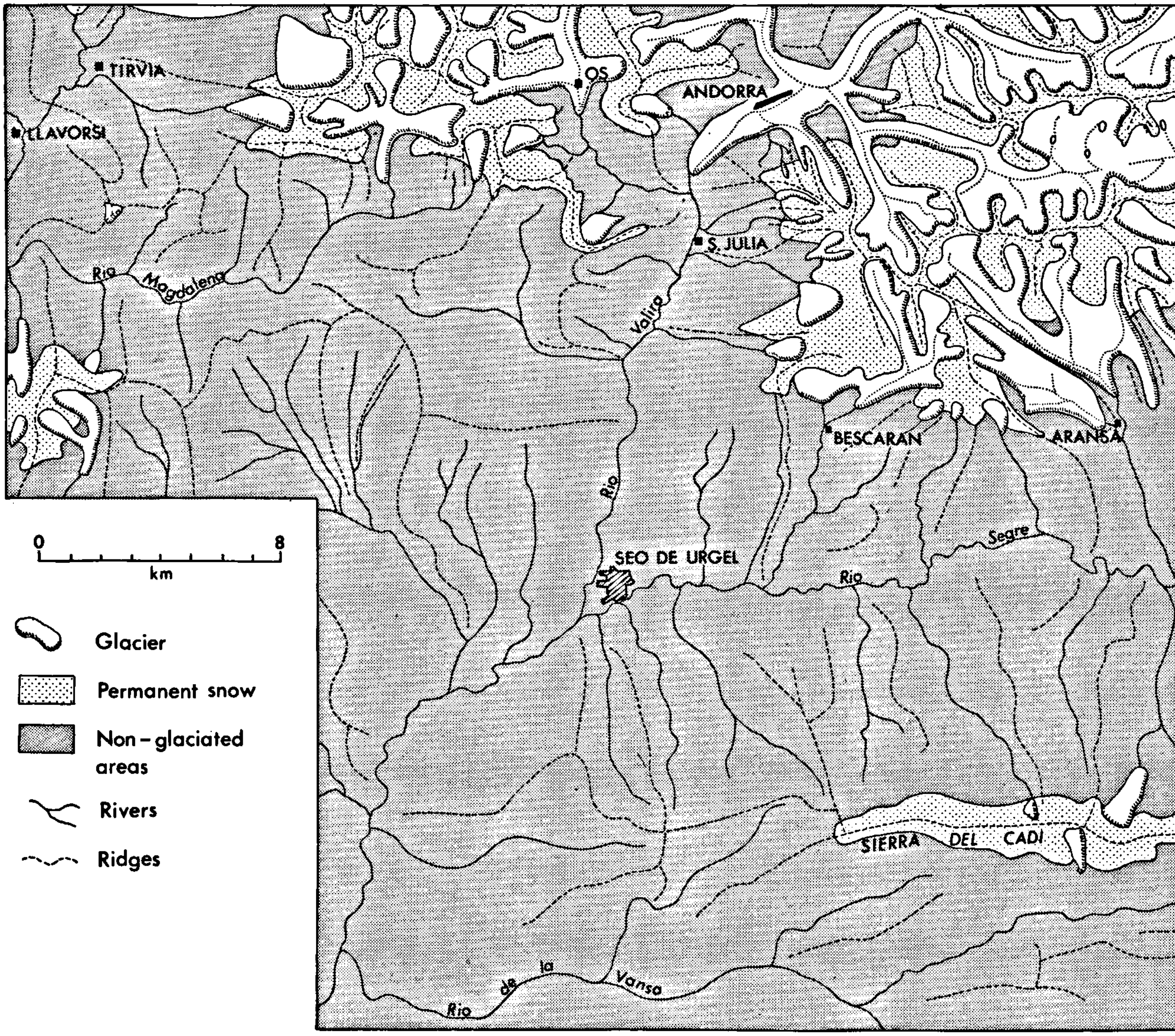


Fig. 67. Map of maximum extent of glaciations in the Segre and Valira area.

CHAPTER V

ECONOMIC GEOLOGY

Barite. - Barite is exploited from veins in Devonian limestones directly below the unconformity with the Bunter Formation southwest of Toloríu. The exploitation takes place on a small scale in open pits and some small drifts.

Bauxite. – Bauxite occurrences at the contact between the Bonanza Formation and the Adrahent or Bona Formation south of the Rio de la Vansa are not accessible for economic exploitation (Bataller, 1943).

Copper. – Chalcopyrite has temporarily been mined at Conflens and west of Os de Civis. The mineralizations, which are found in quartz veins in the Ansobell slates, also comprise siderite and pyrite.

Coal. – Anthracitic coal seams of the Malpas Formation have been mined in the past until economic exploitation was no longer feasible. The exploitation, mainly centred at Las Minas, has been irregular only meeting the demands of the thermo-electric power station at Adrall (Almela & Rios, 1947, p. 178).

Lignite. – Lignite seams of the Miocene Piedra Formation have, in the past, been exploited on a small scale near Seo de Urgel (Almela & Rios, 1947, p. 179).

Limestone. – The red nodular limestone of the Compte Formation is locally quarried on the Beneidó and north of Villech. The rock is ground and used for building purposes.

Slates. – Roof slates are quarried at Farga de Moles in the Ansobell Formation and south of Bixesarri in the Civis Formation.

Clay. – Neogene clays from the Piedra and Ballestá Formations in the Seo de Urgel basin are exploited for the local brick works at Seo de Urgel.

Sand and gravel. – Neogene sand and gravel are exploited in the Ballestá Formation at Montferrer and north of Adrall. Recent river deposits of the Rio Segre are also exploited for gravel.

Thermal sources. – At Baños de S. Vicente hot calcium, potassium, silica, sodium, and chlorine-bearing sulphurous springs occur with a temperature of 42.6°C. In the region of Baños de Sanilles (southwest of Traveseres) four hot springs have sodium, silica, and potassium-bearing water and one has ferruginous water, temperatures rangeing from 27° to 32 °C. Analyses of the waters of S. Vicente and Sanilles, which are mainly used for medical purposes are given by Solé Sabarís & Llopis Lladó (1946, p. 250). In Les Escaldes a hot sodium bearing sulphurous spring occurs with a temperature range of 38° to 66 °C (Zwart, 1965).

SAMENVATTING

De geologie van het Segre en Valira gebied werd gekarteerd en weergegeven op blad 10 van de serie geologische kaarten op schaal 1:50.000 van de Centrale Pyreneeën. De begeleidende beschrijving behandelt structurele, stratigrafische en morfologische aspecten van dit gebied. Het Segre-Valira gebied omvat een deel van de zuidelijke Axiale zone, de oostelijke uitloper van de Nogueras zone en een deel van de noordrand van het zuidelijke marginale bekken.

De gekarteerde formaties reiken van het Cambro-Ordovicium tot het Plioceen. Gedetailleerd lithostratigrafisch onderzoek van het Cambro-Ordovicium en Devoon heeft correlaties mogelijk gemaakt met andere gebieden van de Pyreneeën.

De Hercynische orogenese veroorzaakte structuren van verschillende vorm en oriëntatie die alle toegeschreven kunnen worden aan één pulserend N-S gericht krachtenveld dat een opeenvolging van deformaties veroorzaakte. Deze deformatiefasen kunnen ten dele gecorreleerd worden met deformatiefasen uit andere delen van de Axiale zone. De eerste fase vormde de grootste plooien met een golflengte van 20 km tot minder dan 1 km. De slaty cleavage is in het algemeen parallel aan de gelaagdheid en kan mogelijk worden geïnterpreteerd als een 'concentrische cleavage' behorend tot deze eerste fase. Tweede fase structuren, in het algemeen gekenmerkt door een assenvlaks crenulation cleavage, duiden op een verdere compressie. Tijdelijke vermindering van de hoofdspanning na de eerste en tweede fase resulteerde in de vorming van crossfolds en cascade plooien. Onstabiele structuren in Devonische gesteenten hebben gravitatieve afglijdingen vanaf de Rabassa dome naar de Arcabell syncline en het Segre gebied tot gevolg gehad die een dikke opeenhoping van Devonisch materiaal in deze depressies veroorzaakten. Sterke afwijkingen van de E-W richting van eerste en tweede fase structuren in het oostelijk deel van de Orri dome en in het Sègre gebied worden toegeschreven aan een veronderstelde infrastructuur dieper in de Orri dome.

Epirogenetische bewegingen langs de noordrand van het marginale bekken zijn belangrijk geweest, in het bijzonder gedurende het Perm en het Krijt. Het Devonische massief van de Monsech de Tost, gelegen te midden van post-Hercynische afzettingen in de Nogueras zone, is geïnterpreteerd als een gravitatieve afglijding van Hercynische ouderdom, opgeschoven ten tijde van de Pyreneese plooiings fase. De Pedraforca structuur is ook verklaard als een opgeschoven blok. Latere Alpiene bewegingen tijdens het Neogeen veroorzaakten slenken in het Sègre dal.

REFERENCES

Almela, A. & Rios, J. M., 1947. Explicacion al mapa geológico de la provincia de Lerida. 1:200.000. Rev. Inst. geol. minero España, 193 p.

Ashauer, H., 1934 (translation 1943). La terminación oriental de los Pirineos. Publ. alemanas Geol. Esp., 2, p. 201–342.

Astre, G., 1927. Le bassin néogène de Bellver. Bull. Soc. Hist. Nat. Toulouse, 56, p. 231–258.

Autran, A., Fonteilles, M. & Guitard, G., 1966. Discordance du Paléozoïque inférieur métamorphique sur un socle gneissique antéhercynien dans le massif des Albères (Pyrénées orientales). C. R. Acad. Sc. Paris, (D) 263, p. 317–320.

Bataller, J. R., 1943. Las Bauxitas del Pirineo de Lerida. Mem. Real. Acad. Cienc. Artes Barcelona, 27, no. 2, p. 41–93.

Bhattacharji, S., 1958. Theoretical and experimental investigations on crossfolding. Jour. Geol., 66, p. 625–667.

Birot, P., 1937. Recherches sur la morphologie des Pyrénées orientales franco-espagnoles. Thèse, Paris, 318 p.

Blagbrough, J. W. & Farkas, S. E., 1968. Rock glaciers in the San Mateo Mountains, south-central New Mexico. Am. Jour. Sc., 266/9, p. 812–823.

Bloemraad, J., 1969. Internal report Geol. Min. Inst. Leiden (Dept. Struct. Geol.).

Boersma, K. Th., 1968a. Internal report Geol. Min. Inst. Leiden (Dept. Struct. Geol.).
—, 1968b. Internal report Geol. Min. Inst. Leiden (Dept. Strat. & Palaeont.)
Boissevain, H., 1934. Etude géologique et morphologique de la vallée de la Haute-Sègre. Bull. Soc. Hist. Nat. Toulouse, 66, p. 33–170. (also thesis, Utrecht).
Boschma, D., 1963. Successive Hercynian structures in some areas of the Central Pyrenees. Leidse Geol. Med., 28, p. 103–176.
Brouwer, Th. N., 1968a. Internal report Geol. Min. Inst. Leiden (Dept. Struct. Geol.).
—, 1968b. Internal report Geol. Min. Inst. Leiden (Dept. Sediment.).
Cavet, P., 1957. Le Paléozoique de la zone axiale des Pyrénées orientales françaises entre le Roussillon et l'Andorre. Bull. Serv. Carte geol. Fr., 55/254, p. 303-518.
—, 1964. Sur la stratigraphie du Paléozoïque du massif de l'Agly, aux environs d'Estagel. Mem. B.R.G.M., 33, p. 99–104.
Chevalier, M., 1906. Les glaciers pleistocènes dans les vallées d'Andorre et dans les hautes vallées espagnoles environnantes. Acad. Sc. Paris, 67, p. 662, 910.
—, 1909. Note sur la 'cuencita' de la Seo de Urgel. Bull. Soc. géol. Fr., (4) 9, p. 158–178.
—, 1910. Nouvelle note sur la 'cuencita' de la Seo de Urgel. Bull. Soc. géol. Fr., (4) 10, p. 9–10.
Clin, M., 1959. Etude géologique de la haute chaîne des Pyrénées Centrales entre le cirque de Troumouse et le cirque du Lys. Thèse Fac. Sc. Nancy, 379 p.
Dalloni, M., 1913. Stratigraphie et tectonique de la région des Nogueras (Pyrénées Centrales). Bull. Soc. Géol. Fr., (4) 13, p. 243–263.
—, 1930. Etude géologique des Pyrénées Catalanes. Ann. Fac. Sci. Marseille, 26, 373 p.
Dennis, J. G., 1967. International tectonic dictionary. Am. Assoc. Petrol. Geol., Mem. 7, 196 p.
Dessauvagie, Th. F. J., 1960. Internal report Geol. Min. Inst. Leiden (Dept. Struct. Geol.).
Destombes, J. P., 1953. Stratigraphie des terrains primaires de la Haute Garonne. C. R. 19e Congr. géol. internat., Alger (1952), sect. 2, fase 2, p. 107–129.
Diederix, D., 1963. Internal report Geol. Min. Inst. Leiden (Dept. Struct. Geol.)
Dieterich, J. H., 1969. Origin of cleavage in folded rocks. Am. Jour. Sci., 267/2, p. 155–165.
Embleton, C. & King, C. A. M., 1968. Glacial and Periglacial Geomorphology. Edward Arnold (Publishers) Ltd., 608 p.
Furtak, H., 1962. Die 'Brechung' der Schiefrigkeit. Geol. Mitt., 2/2, p. 177–196.
Glasenapp, W. von, 1967. Zur Geologie der mittleren Südpyrenäen im Osten des Rio Segre. Internal report Inst. Geol. München.
Grekoff, N., Guérin-Desjardins, B., Latreille, M., Lys, M., Sigal, J. & Siskind, B., 1961. Présence de niveaux marins du Néocomien et probablement du Malm dans les Pyrénées de Lérida (Espagne). C. R. Acad. Sc. Paris, (D) 252, p. 2262–2264.
Guérin-Desjardins, B. & Latreille, M., 1961. Etude géologique dans les Pyrénées espagnoles entre les rios Segre et Llobregat (prov. Lerida et Barcelone). Rev. Inst. Fr. Pétrole, 16/9, p. 922–940.
Hara, I., 1966. Movement picture in confined incompetent layers in flexural folding. Deformation of heterogeneously layered rocks in flexural folding (I). Jour. Geol. Soc. Jap., 72/8, p. 363–369.
—, 1967. A note on 'concentric' folding of multilayered rocks. Jour. Sci. Hiroshima Univ., (C) 5/3, p. 217–239.
Hartevelt, J. J. A., 1965. Internal report Geol. Inst. Leiden (Dept. Struct. Geol.).
—, 1970. Stratigraphic position of the limestones and conglomerates around the Marimaña granodiorite, Central Pyrenees, Spain. Geol. Mijnb. (in press).
Hartevelt, J. J. A. & Roger, Ph., 1968. Quelques aspects de la topographie permo-triasique dans le Haute-Sègre et la Haute-Pallaresa (Lerida, Espagne). C. R. Somm. Séances Soc. Géol. Fr., 6, p. 182–184.
Hellermann, E., 1965. Schiefrigkeit und Gebirgsbau im östlichen Sauerland. Geol. Mitt., 4/4, p. 333–396.
Hoorn, B. van, 1970. Sedimentology and paleogeography of a turbidite basin in Spain. Leidse Geol. Med., 45, p. 73–154.
Jacob, Ch., 1935. A propos du versant méridional des Pyrénées centrales. Bull. Soc. Géol. Fr., 6, p. 78–80.
Jacob, Ch., Fallot, P., Astre, G. & Ciry, R., 1926. Observations tectoniques sur le versant méridional des Pyrénées centrales et orientales. C.R. 14e Congr. Géol. Internat., Madrid, fasc. 2, p. 335–412.
Kleinsmiede, W. F. J., 1960. Geology of the Valle de Arán (Central Pyrenees). Leidse Geol. Med., 25, p. 131–241.
Knill, J. L., 1960. The tectonic pattern in the Dalradian of the Craignish-Kilmelfort district, Argyllshire. Quart. Jour. Geol. Soc. London, 115, p. 339–364.
Lahee, F. H., 1961. Field Geology. McGraw-Hill, New York, 926 p.
Langheinrich, G. & Plessmann, W., 1968. Zur Entstehungsweise von Schieferungsflächen in Kalksteinen (Turon-Kalke eines Salzauftriebs im Harz-Vorland). Geol. Mitt., 8/2, p. 111–142.
Llopis Lladó, N., 1965. Sur le Paléozoïque inferieur de l'Andorre. Bull. Soc. géol. Fr., (7) 7, p. 652–659.
—, 1966. Sobre la estratigrafía del Silúrico de Andorra y el limite Silúrico-Devónico. Pirineos, 81–82, p. 79–85.
—, 1969. Estratigrafía del Devónico de los Valles de Andorra. Mem. R. Acad. Cienc. Artes Barcelona, 39/7, p. 219–290.
—, 1970. Mapa geológico de Andorra, 1:25,000. Cuadernos geol. iberica, Madrid, 1, p. 347–348, sheets 1, 2, 4, 5 and 6.
Mey, P. H. W., 1967. The geology of the Upper Ribagorzana and Baliera valleys, Central Pyrenees, Spain. Leidse Geol. Med., 41, p. 153–220.
—, 1968. Geology of the Upper Ribagorzana and Tor valleys, Central Pyrenees, Spain. Leidse Geol. Med., 41, p. 229–292.
Mey, P. H. W., Nagtegaal, P. J. C., Roberti, K. J. & Hartevelt, J. J. A., 1968. Lithostratigraphic subdivision of post-Hercynian deposits in the South-Central Pyrenees, Spain. Leidse Geol. Med., 41, p. 221–228.
Mirouse, R., 1966. Recherches géologiques dans la partie occidentale de la zone primaire axiale des Pyrénées. Mem. expl. carte géol. dét. Fr., 451 p.
—, 1967. Le Devonien des Pyrénées occidentales et centrales (France). Internat. Symp. Devonian Syst., 1, p. 153–170.
Misch, P., 1934 (translation 1948). La estructura tectónica de la region central de los Pirineos meridionales. Publ. extranjeras Geol. Esp., 4, p. 5–178.
Morre, N. & Thiébaut, 1964. Constitution de quelques roches volcaniques Permiennes de la Sierra del Cadi (Pyrénées Catalanes). Bull. Soc. géol. Fr., (7) 6, p. 389–396.
Nagtegaal, P. J. C., 1962. Internal report Geol. Min. Inst. Leiden (Dept. Struct. Geol.).
—, 1966. Scour-and-fill structures from a fluvial environment. Geol. Mijnb., 45, p. 345–354.

—, 1969. Sedimentology, Paleoclimatology and Diagenesis of post-Hercynian continental deposits in the south-central Pyrenees, Spain. Leidse Geol. Med., 42, p. 143–238.

Nussbaum, F., 1946. Orographische und morphologische Untersuchungen in den östlichen Pyrenäen. Jahresb. Geogr. Ges. Bern, 35–36, 247 p.

Oele, E., Sluiter, W. J. & Pannekoek, A. J., 1963. Tertiary and Quaternary sedimentation in the Conflent, an intramontane rift-valley in the eastern Pyrenees. Leidse Geol. Med., 28, p. 297–320.

Oele, J. A., 1966. The structural history of the Vall Ferrera area, the transition zone between the Aston massif and the Salat-Pallaresa anticlinorium (Central Pyrenees, France, Spain). Leidse Geol. Med., 38, p. 129–164.

Oliver, P. G., 1967. Graptolite evidence for rocks of Llandoverian (Silurian) age in south-western Andorra. Geol. Mag., 104/4, p. 390–392.

Pannekoek, A. J., 1937. Die jungtertiäre morphologisch-tektonische Entwicklungsgeschichte der östlichen Pyrenäen. Ass. Et. Géol. Médit. Occ. (Géol. Pays Catal.), III, no. 4, partie I, 25 p.

Peybernès, B., 1968. Précisions stratigraphiques sur le Jurassique terminal et le Crétacé inférieur aux abords de la vallée du Sègre (province de Lerida, Espagne). C. R. somm. Soc. géol. Fr., 1, p. 15–16.

Plessmann, W., 1964. Gesteinslösung, ein Hauptfactor beim Schieferungsprozess. Geol. Mitt., 4/1, p. 69–82.

—, 1966. Lösung, Verformung, Transport und Gefüge (Beiträge zur Gesteinsverformung im nordöstlichen Rheinischen Schiefergebirge). Zeitschr. deutsch. geol. Ges., 115, p. 650–663.

Ramsay, J. G., 1967. Folding and fracturing of rocks. Mc Graw-Hill, New York, 568 p.

Rat, P., 1966. Sur les facies du Crétacé inférieur dans l'Est du domaine pyrénéen. Pirineos, 81–82, p. 117–127.

Richter, D., 1965. Verkürzung von Fossilien und Entstehung von Flaser- und Knollenkalken durch Lösungsvorgänge in geschieferten kalkigen Gesteinen. Geol. Mitt. 4/3, p. 235–248.

Rickard, M. J., 1961. A note on cleavages in crenulated rocks. Geol. Mag., 98/4, p. 324–332.

Rios, J. M., 1951. Analysis estratigráfico y tectónico de una parte del Valle del Segre en la Provincia de Lerida. Bol. Inst. geol. min. Esp., 63, p. 561–656.

—, 1959. El sistema cretaceo en los Pirineos de Espana. Congr. Geol. Internacional, 20th Sess., México, 1956, p. 321–403.

Roger, Ph., 1965. Etude stratigraphique et structurale de la zone des Nogueras entre l'Esera et l'Isabena (Huesca-Espagne). Act. Soc. Linnéenne Bordeaux, 102, p. 3–27.

Roggeveen, P. M., 1929. Geologisch-petrographische onderzoekingen in het granietmassief van Llés-Aristot in de oostelijke spaanse Pyreneeën. Thesis, Utrecht, 115 p.

Rijnsburger, A., 1967. Internal report Geol. Min. Inst. Leiden (Dept. Struct. Geol.).

Savage, J. F., 1961. Internal report University of London.

—, 1967. Tectonic analysis of Lechada and Curavacas synclines, Yuso basin, León, NW Spain. Leidse Geol. Med., 39, p. 193–247.

Schmidt, H., 1931. Das Paläozoikum der spanischen Pyrenäen. Abh. Ges. Wiss. Göttingen math.-phys. Kl. 3, Folge, H. 5, no. 8, p. 1–85.

Seguret, M., 1964. Sur le style en tête plongéante des structures pyrénéennes de la Zone de Nogueras (versant Sud des Pyrénées centrales). C.R. Acad. Sci., 259, p. 2895–2898.

—, 1969. La nappe de Pedraforca: nouvelle unité allochtone du versant sud des Pyrénées. C.R. Acad. Sci. Paris, (D) 269, p. 552–555.

Sitter, L. U. de, 1954. Schistosity and shear in micro- and macrofolds. Geol. Mijnb., 16, p. 429–439.

—, 1956. A cross section through the Central Pyrenees. Geol. Rundschau, 45, p. 214–233.

—, 1959. The structure of the axial zone of the Pyrenees in the province of Lerida. Est. Geol., 15, p. 349–360.

—, 1964. Structural Geology 2nd ed., McGraw-Hill, New York, 551 p.

—, 1965. Hercynian and Alpine orogenies in northern Spain. Geol. Mijnb., 44, p. 373–383.

Sitter, L. U. de & Zwart, H. J., 1962. Geological map of the Paleozoic of the Central Pyrenees, sheet 1 Garonne and sheet 2 Salat, France. Leidse Geol. Med., 27, p. 191–236.

Solé Sabarís, L. & Llopis Lladó, N., 1946. Explicacion de la hoga no. 216 Bellver. Mapa geológico de España. Inst. geol. min. Esp., 109 p.

—, 1947. Mapa geológico de Andorra, 1:50,000, Lerida, Inst. Est. Ilerdenses.

Solé Sugrañes, L. & Santanach Prat, P., 1970. Nota sobre la escama de corrimiento del Monsech de Tost en el Prepirineo español (pro. de Lérida). Acta Geol. Hispánica, 5, p. 24–28.

Souquet, P., 1967. Le Crétacé supérieur sud-pyrénéen en Catalogne, Aragon et Navarre. Thèse, Fac. Sci. Toulouse, 529 p.

Taillefer, F., 1951. Le Piémont des Pyrénées francaises. Privat, Toulouse, 383 p.

Trouw, R. A. J., 1969. Internal report Geol. Min. Inst. Leiden (Dept. Struct. Geol.).

Viers, G., 1963. Les moraines externes de la Cerdagne et du Capcir (Pyrenees-orientales, France) et leurs rapports avec les terrasses alluviales. Rep. 6th. Internat. Cong. Quaternary, Warsaw 1961, 3, p. 385–393.

Virgili, C., 1958. El Triassico de los Catalánides. Bol. Inst. Geol. Min. Esp., 69, 856 p.

Wahrhaftig, C. & Cox, A., 1959. Rock glaciers in the Alaska Range. Geol. Soc. Am. Bull., 70, p. 383–436.

Waterlot, M., 1969. Contribution à l'étude géologique du Carbonifère anté-stéphanien des Pyrénées centrales espagnoles. Mem. Inst. Geol. Min. Esp., 70, p. 3–259.

Wees, H. van, 1970. Internal report Geol. Min. Inst. Leiden (Dept. Struct. Geol.).

Weiss, L. E., 1968. Flexural-slip folding of foliated model materials. Proc. conf. res. tectonics (Kink bands and brittle deformation), Geol. Surv. Canada paper, 68–52, p. 294–362.

Zandvliet, J., 1960. The geology of the Upper Salat and Pallaresa valleys, Central Pyrenees, France/Spain. Leidse Geol. Med., 25, p. 1–127.

Zwart, H. J., 1954. La géologie du massif du Saint-Barthélemy, Pyrénées, France. Leidse Geol. Med., 18, p. 1–228.

—, 1963. The structural evolution of the Paleozoic of the Pyrenees. Geol. Rundschau, 53, p. 170–205.

—, 1965. Geological map of the Paleozoic of the Central Pyrenees, Sheet 6, Aston, France, Andorra, Spain. 1:50.000. Leidse Geol. Med., 33. p. 191–254.

GEOGRAPHICAL COORDINATES

Topographic names of sheet 10 used in the text.
(longitudes east of Madrid)

Adrahent	42°16½′-5°11′
Adrall	42°19½′-5°05′
Alás	42°21′ -5°11½′
Aguila, Peña del	42°15′ -5°14½′
Aguila, Roch del	42°22½′-5°21′
Andorra la Vella	42°30½′-5°13′
Ansobell	42°19½′-5°16½′
Arcabell	42°26′ -5°10′
Arcabell, Sierra de	42°25′ -5°12′
Argolell	42°26′ -5°08′
Aristot	42°23′ -5°19′
Ars	42°27′ -5°05′
Arseguell	42°21′ -5°16′
Asnurri	42°27′ -5°06½′
Ballestá	42°22′ -5°06½′
Baños de Sanilles	42°22½′-5°22′
Baños de San Vicente	42°22′ -5°17′
Bar	42°21½′-5°19½′
Barceloneta	42°14½′-5°08½′
Beixech	42°21′ -5°21′
Beneidó	42°22½′-5°15′
Bescarán	42°24′ -5°14′
Bescarán, T^te^. de	42°22′ -5°13′ 42°26′ -5°17′
Bestanis	42°19′ -5°22½′
Bixesarri	42°29′ -5°09′
Bona, Rio de	42°15′ -5°09′ 42°17′ -5°13′
Bordas del Ras	42°26½′-5°03½′
Burch	42°30′ -4°57½′
Cadi, Rio de	42°21′ -5°16′ 42°18′ -5°19′
Cadi, Sierra del	42°16′ -5°11′ 42°17′ -5°22′
Calvinya	42°23′ -5°09′
Can Franch	42°18′ -5°07′
Capiscot, T^te^.	42°21′ -5°22′
Castellá	42°16′ -5°04′
Castellbó	42°22½′-5°03′
Castellnou de Carcolsé	42°23′ -5°16′
Caumer	42°28′ -5°08′
Cavá	42°19½′-5°19½′
Civis	42°28′ -5°07½′
Coll d'Arnat	42°15′ -5°06′
Coll de Ser	42°21′ -5°18′
Conflens, Bordas de	42°30′ -5°03½′
Cotaróns	42°14′ -5°19′
Corves, B. de	42°28½′-5°04′
Encamp	42°32′ -5°16′
Escaldes, les	42°30½′-5°14′
Estana	42°19′ -5°21′
Estimaríu	42°22½′-5°13′
Farga de Moles	42°26′ -5°09′
Farrera	42°30′ -4°57½′
Firal	42°21½′-5°09′
Fontaneda	42°27½′-4°09′
Fornols	42°15′ -5°12′
Ges	42°19′ -5°11′
Gosól	42°14′ -5°21′
Grau	42°20′ -5°16′
Hostalets	42°17′ -5°03′
Josa del Cadi	42°15½′-5°18½′

Lladorre, Rio de	42°31′ -4°55′
Llavorsi	42°30′ -4°54′
Llés	42°23½′-5°22½′
Minas, las	42°18′ -5°06½′
Monsech de Tost	42°15′ -5°07′
Montant	42°14′ -5°04½′
Montescladó	42°30′ -4°56′
Montferrer	42°20½′-5°07′
Mosbé, Serrat de	42°20′ -5°21′
Musa	42°23′ -5°21′
Novés de Segre	42°17½′-5°02′
Orri	42°24½′-4°54′
Ortedó	42°20′ -5°12′
Os de Civis	42°31′ -5°08′
Os, Rio de	42°30′ -5°08′ 42°29′ -5°10′
Pallaresa, Rio	42°29′ -4°54′
Palleróls	42°18′ -5°02′
Pedraforca	42°15′ -5°23′
Piedra, B^co^. de	42°21′ -5°09′ 42°18′ -5°10′
Pla de Llés	42°24½′-5°19′
Pla de San Tirs	42°19′ -5°04′
Prada, Sierra de	42°15′ -5°00′
Purredón	42°15′ -5°05′
Quer, B^co^. de	42°22′ -5°21′ 42°18′ -5°20′
Querforadat	42°19½′-5°19′
Rabassa, Pic de la	42°26′ -5°13′
Rabassa, Riu de la	42°27′ -5°12′
Sabaña	42°17′ -5°08′
Saloria	42°31′ -5°04′
Santa Coloma	42°29½′-5°11′
Sant Joan Fumat	42°26′ -5°06½′
Sant Julia	42°28′ -5°11′
Segre, Rio	42°14′ -5°02′ 42°42′ -5°22′
Seo de Urgel	42°21½′-5°09′
Serch	42°20½′-5°11′
Sisque	42°15′ -5°09½′
Syspony	42°32′ -5°12′
Tirvia	42°31′ -4°56′
Toloríu	42°22′ -5°19′
Tora	42°16½′-5°06′
Tora de Tost, Rio	42°17′ -5°03′ 42°17′ -5°09′
Torres	42°21½′-5°13′
Tossa Plana	42°28′ -5°21′
Tost	42°16½′-5°04′
Traveseres	42°22½′-5°22½′
Tressot, B^co^. de	42°28′ -4°59′
Turo de Call Pubill	42°19½′-5°19′
Tuta. Sierra de la	42°23½′-5°17′
Tuxent	42°14′ -5°15′
Valira, Rio	42°21′ -5°08′ 42°30½′-5°13′
Valira del Nort, Riu	42°31′ -5°13′
Valira d'Orient, Riu	42°31′ -5°14′
Vansa, la	42°14′ -5°10′
Vansa, Rio de la	42°15′ -5°02′ 42°14′ -5°15′
Vilamitjana	42°21′ -5°04′
Vilanova de Benat	42°20½′-5°14½′
Villech	42°20′ -5°22′
Villech, Tossal de	42°20′ -5°22′