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VARISCAN OLISTOSTROME DEPOSITION AND SYNSEDIMENTARY NAPPE EMPLACEMENT, VALDEON AREA, CANTABRIAN MOUNTAINS, SPAIN

BY

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ABSTRACT

In the area of Valdeón and adjacent western Liébana (along the provincial boundary between León and Santander, in northern Spain) Namurian and Westfalian flysch deposits are unconformably covered by a sequence of olistostrome and flysch units, alternating with nappes. This sequence, with a thickness of about 5 km, was built up during the Middle Cantabrian. It is covered in turn by the Picos de Europa nappes with intercalated flyschoid sediments of a slightly younger age. In this paper the discrete lithostratigraphic and structural units are described and analyzed. Special attention is given to the fusulinid biostratigraphy. The contact relations between nappes and olistostrome units and their complementary lithostratigraphy are explained as the result of synsedimentary emplacement of the nappes by gravitational sliding along the earth surface. The palaeogeologic setting prior to the deposition of the nappes and olistostromes is deduced from the stratigraphy and lithofacies of the nappes and the allochthonous elements constituting the olistostromes. The regional geologic setting thus obtained and its consequent sedimentological and structural development, are discussed. An attempt is made to fit these developments in the model of a continental strike-slip system.

CONTENTS

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CHAPTER I

INTRODUCTION

Aim and scope of the investigations

Investigations in the Upper Carboniferous flysch basin of Liébana by the first author resulted in a model of a small but deep asymmetric basin, probably fault-controlled, in which vertical movements and gravitational transport were the driving forces both for the sedimentary and the essentially contemporaneous structural development (Maas, 1974). Subsequently Reading (1975) introduced a model for fault-controlled sedimentation of the Upper Carboniferous of the Cantabrian Mountains: a number of asymmetric basins developed within a continental strike-slip system. The detailed application of that model to the Cantabrian zone was elaborated by Heward & Reading (1980).

When the first author had the opportunity to resume fieldwork in the Cantabrian Mountains as a staff member of the Department of Geology of the University of Leiden (fieldsummers 1980, 1981), he chose to investigate the area of western Liébana and Valdeón where chaotic sediments (olistostromes) are even more numerous than in the central part of Liébana. The idea was that stratigraphical analysis of the exotics in olistostromes and study of the internal structures within olistostromes and of the contacts between olistostromes and adjacent rock units, might link the history of deposition to fault activity in (a) provenance area(s). The five weeks allotted for fieldwork were not enough for a detailed mapping of the entire area. The map (Encl. I) offers no more than an outline of the geology. However, the better exposed parts of the olistostrome units were studied in detail and a careful stratigraphical analysis, especially by determination of the fusulinid assemblages in the limestone olistoliths - the second author's contribution -, has established a solid foundation of facts for the succession and the correlation of the discrete stratigraphic and structural units (Encl. IV). This paper reports the not quite expected results of the investigations.

Previous work

Kullmann (1960) did a stratigraphical reconnaissance in the Devonian and Lower Carboniferous rock units of the Valdeón area (the "Montó Schichten"). Kutterink (1966), one of de Sitter's students, mapped the Valdeón area and distinguished the various Devonian and Carboniferous lithostratigraphic units. He unravelled the complicated imbricated thrust- and fold structures of the Montó inlier. He also noted zones of intense distortion in the Upper Carboniferous rocks at the base of the Montó unit and along the road from Santa Maria de Valdeón to the Puerto de Pandatrave (Encl. I). Though he mentions the occurrence of slump structures, he essentially attributes the distortions to tectonic causes and does not extrapolate from narrow zones defined by a number of fortuitous outcrops. In his time the concept of olistostrome sedimentation was not yet applied to the geology of the Cantabrian Mountains. Van Adrichem Boogaert (1967) confirmed the stratigraphy of the Montó area, using the conodont assemblages from the limestone units for correlation. The Montó Devonian is a part of the Palentine Devonian facies area (Brouwer, 1964) which occurs between the Cardaño Line and the Picos de Europa (Fig. 1). Boschma (1968) incorporated the map of Kutterink in his geological map (1 : 100,000) of the Valdeón-Liébana-Polaciones area. Lobato (1974, published in 1977) offers an adapted and reinterpreted version of the geology of the Valdeón area. On his map, of the zones of distortion only the zone under the Montó unit is indicated; it is interpreted as a tectonic melange ("mezcla tectonica"). A slightly adapted version of the map of Kutterink was recently published on a 1 : 50,000 scale as a part of the Yuso map sheet of the Geological Map of the Southern Cantabrian Mountains (Savage & Boschma, 1980). The area adjacent to Valdeón, the westernmost part of Liébana, was mapped and published by the first author (Maas, 1974). The results of our recent investigations necessitate a drastic reinterpretation of the geology of the Valdeón and the westernmost Liébana area.

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CHAPTER II

DESCRIPTION AND ANALYSIS OF THE LITHOSTRATIGRAPHIC AND STRUCTURAL UNITS

GENERAL GEOLOGIC SETTING

During Middle and Late Devonian times, three distinct facies areas existed in the Cantabrian Mountains (van Adrichem Boogaert, 1967):

1) The Asturo-Leonese facies, a shallow marine, open shelf facies of alternating reefal limestones and shallow siliciclastics, occurs in the southern and western parts of the mountains.
2) The central to northeastern part of the mountains is characterized by a depositional hiatus representing middle Ordovician to late Famennian; this area is known as the "Asturian geanticline" and probably includes the Picos de Europa area (Fig. 1).

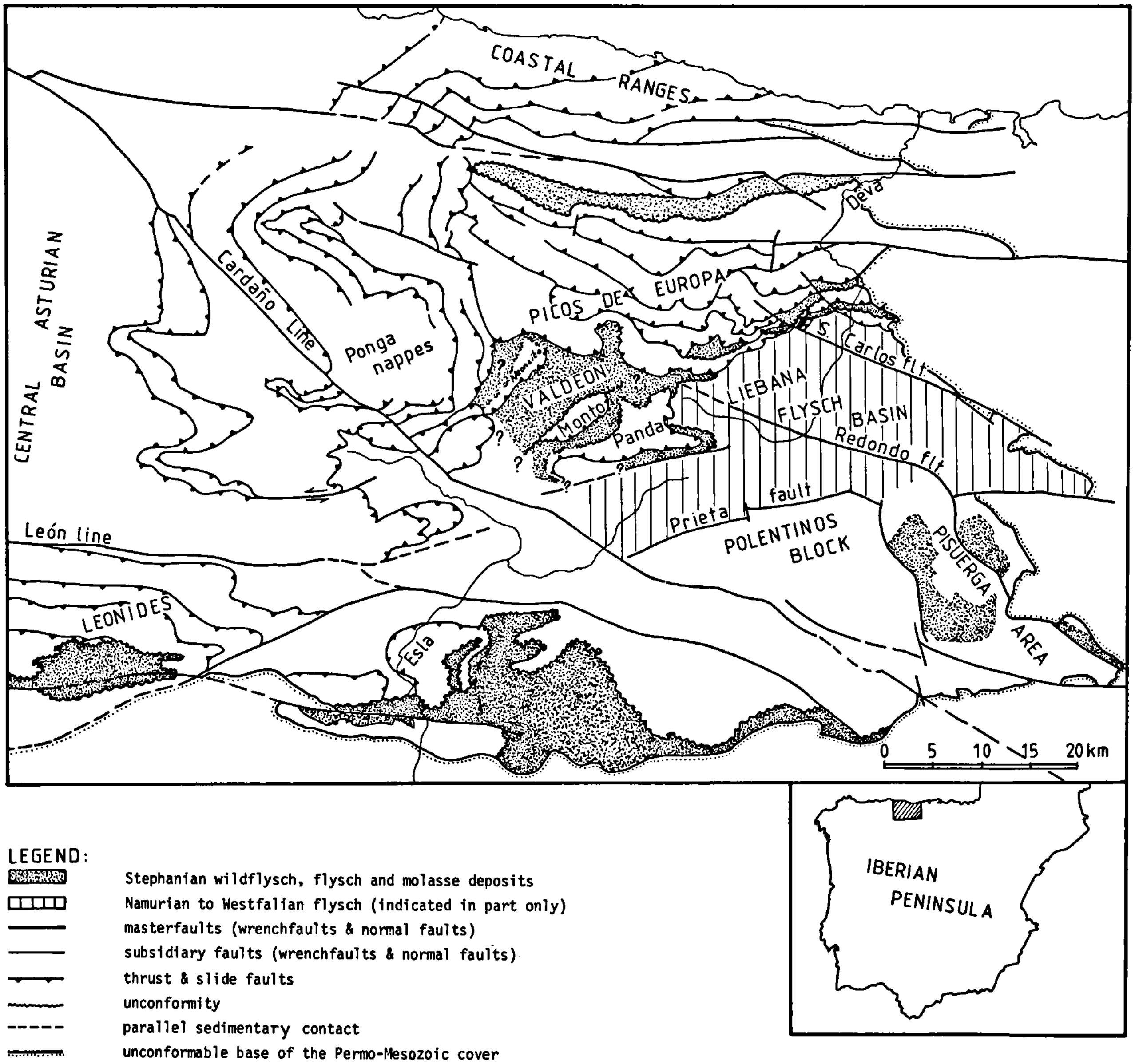

Fig. 1. Tectonic index map of the eastern Cantabrian Mountains (adapted from Savage, 1981).

3) The third facies area is known as the Palentine facies area (cf. Brouwer, 1964; van Veen, 1965), it occurs in the central eastern part of the mountains. Its Middle and Upper Devonian rocks represent a restricted basinal facies (cf. Raven, 1983), they consist of a condensed sequence of nodular limestones and shales, characterized by an impoverished pelagic fauna (the Eifelian-Givetian Gustalapiedra Fm., the Frasnian Cardaño Fm. and the Famennian Vidrieros Fm.). The condensed sequence consitutes the autochthonous basinal facies. This facies is interrupted by the influx of turbidites (mainly representing facies types B and C of Mutti & Ricci Lucchi, 1976) during early and middle Famennian times: the Murcia Fm. The autochthonous condensed sequence extends into the Lower Carboniferous (Vegamián and Villabellaco Fms., cf. Maas, 1974) for the entire Palentine facies area.

The Palentine succession forms the base for the larger part of the Upper Carboniferous flysch occurrences in the Cantabrian Mountains and underlies the Liébana and Valdeón basins, including the studied area. Exotics derived from Palentine facies rocks are frequently encountered in the olistostrome units of Liébana and Valdeón.

The Valdeón rock units rest upon the western extension of the Liébana flysch basin; both the Liébana and Valdeón rock units are covered in the north by the Picos de Europa nappes (Fig. 1). The Liébana flysch is a true flysch consisting largely of turbidite sequences (Maas, 1974; Rupke, 1977). This flysch has, however, some peculiar characteristics when compared to the facies types of turbidites and associated deposits as established by Mutti & Ricci Lucchi (M & RL, 1976) in the northern Apennines. The specific characteristics of the Liébana flysch and the Liébana basin have to be considered before we start with the description of the Valdeón rock units. Mutti & Ricci Lucchi (1976) distinguished a number of facies types of turbidites and associated deposits, facies types A-G, which they interpreted as submarine slope, deep-sea fan and deep-sea basin deposits. Maas (1974) distinguished the following facies types in the Liébana flysch:

1) Slump and olistostrome deposits which compare well with the chaotic facies (type F) of M & RL, which is mainly a submarine slope facies according to M & RL.
2) So-called high-density current deposits (i.e. debris flow and grain flow deposits) which compare well with the arenaceous-conglomeratic facies (types A & B) of M & RL, interpreted by them as the facies of deep-sea fan valleys.
3) Turbidites. These deposits need specification: The Liébana turbidites (cf. Maas, 1968; Lobato, 1977; Rupke, 1977) are marked by high sand/shale ratios (averaging 3 : 1) whereas the individual bed thicknesses are conspicuously low (usually well below 15 cm, averaging 5-10 cm). Individual turbidites hardly ever display the entire Bouma sequence (T_{a-e}), but are usually T_{ab}, T_{bc}, T_c or T_{cd}-turbidites. Their bottoms are always sharp but the tops too are more often sharp than gradual. The turbidite sands are medium- or fine-grained to very fine-grained lithic arenites to lithic graywackes which are poor in matrix. They contain 60-90 % subangular to subrounded quartz clasts, and a variety of very low-grade phyllite fragments, opaque grains, carbonate grains, chert grains and occasional grains of detrital colorless mica. Distinction between matrix and sheared phyllite fragments may prove difficult. Secondary carbonate- and limonite(?)-cement and carbonate veins are of common occurrence. The "shale" intervals between the turbidite beds are sandy mudstones.

The characteristics of the predominant turbidite facies in Liébana do not compare well with those of the predominant turbidite facies types in the northern Apennines, i.e. facies types C and D of M & RL. Facies type C is the "proximal" turbidite facies of the inner and middle regions of a deep-sea fan; type D is the dominant "distal" facies of the outer fan parts and of the basinal plains. The Liébana turbidite facies compares most favourably with turbidite facies type E of M & RL, which is the turbidite facies of the lower slope region and of the overbank deposits in the inner parts of a deep-sea fan.

4) Very thin (0.5 cm or less) but very continuous streaks of graded very fine-grained sandstone to coarse-grained siltstone commonly alternate with the true turbidite beds. Such streaks also occur as the laminae in laminated mudstones which are often distorted by slumping. Maas (1974) interpreted the laminated mudstones as either a distal facies or a facies transitional between deep basinal and shelf sediments (Maas, 1974, cf. photograph 31). The streaked mudstone facies compares well with the hemi-pelagic slope facies (type G) of M & RL.

The comparatively small Liébana basin, although it contains small but distinct deep-sea fan associations (Rupke, 1977), has a flysch facies which in its general aspects is more characteristic for a deep-sea slope than for a deep-sea basin. *) This conclusion is supported by the relative abundance of slump and olistostrome deposits and by the ubiquitous occurrence of early tectonical gravitational deformations (collapse- and flap-folding, slip sheets etc.). The general direction of transport of these slump and collapse phenomena is north over south i.e. the transport direction of the deep-sea fan described by Rupke (1977).

We conclude that the Liébana basin was a small asymmetric unstable basin with a southward sloping bottom. A largely siliciclastic, intermediate shelf/slope area must have existed north of the Liébana basin: the provenance area of at least a part of the flysch sediments. That area in its turn must have been bounded further north by the Picos de Europa limestone shelf, characterized by algal biogenetic bank deposits and the associated bioclastic grain- and packstones of the taluds and the wacke- and mudstones of the lagoonal facies. This entire slope area from shelf to deep-sea is also the geological setting for the development of the structural-stratigraphic succession of the rock units in western Liébana and Valdeón.

*) A near-slope facies for the Liébana turbidites has been suggested to the first author by Dr. J.F. Savage, during a discussion in fall 1980.

CORISCAO OLISTOSTROME

Occurrence
West of the Puerto San Glorio the Panda Limestone Member (Maas, 1974) crops out in a synclinal structure with an E-W axis (Encl. I). The Coriscao Olistostrome (Savage, 1961; Kamerling, 1962; Maas, 1974) occurs in the north- and south-flank of that syncline, wedging out to NW in the north-flank and disappearing in an area without outcrops in the south-flank W of the road from the Puerto de Pandatrave. Its thickness increases to S, probably exceeding 500 m; faults cut out all but the northwestern part of the outcrop of the base which rests unconformably on siliciclastic turbidites of the upper Deva area.

Composition, transport and deposition, synsedimentary and other deformations
The wedge-shaped olistostrome unit displays a very crude layering which is best exposed in the mountain slope that faces the Pico Coriscao from the east (Photograph 1). This well-exposed part of the olistostrome is still disturbed by faults and folding. Enclosure III offers a sketch of a part of the succession as it must have been before the deformation, a sketch featuring the main varieties in lithology and internal structure. Photographs 1 to 12 demonstrate these varieties.

The olistostrome consists of angular to subrounded "clasts" varying from microscopic dimensions to rock slabs with thicknesses of more than fifty metres and hundreds of metres in width. In this description "clasts" with their largest diameter between 0.25 and 2.5 m are denoted as boulders or blocks and larger "clasts" as olistoliths. Cobbles and pebbles are distinguished conforming to standard usage.

The layering in the olistostrome results from vertical variations in the dimensions of boulders and olistoliths, from vertical variations in the clast/matrix ratio and from the near-horizontal deposition of slab-shaped olistoliths. Large parts of the olistostrome display such layering (Photographs 4, 5 & 6), but other parts are entirely chaotic (Photographs 1 & 7). The matrix of the olistostrome is usually a structureless marly mudstone to breccious wackestone, containing numerous angular to subrounded grains and small pebbles (Photograph 9), predominantly limestone fragments, which are often bioclastic. Subangular to subrounded pebbles, cobbles and boulders (limestone clasts and sporadic shale fragments) and olistoliths (limestone "clasts" and masses of turbiditic sandstone mostly) float in the matrix. The sediment has the character of a debris flow deposit.

The entire olistostrome consists of a large number of individual debris flow units, ranging from 0.5 m to more than 50 m in thickness. The thinner debris flow units are often synsedimentary distorted (Photograph 4). The larger units could easily carry the boulders and most olistoliths (cf. Photographs 6 & 7). However, during the waning of the transport, olistoliths and large boulders moved individually under their own weight, causing deformation of the sediment in front of them (Photograph 12, Encl. III). The largest olistoliths (Photographs 2 & 5, Encl. III), some of them slab-shaped masses of turbiditic sandstone/shale alternations, seem to have slid independently. The drag along the slide horizon, which tends to make a sharp angle with the internal bedding of the slab-shaped olistoliths, can cause small folds (Photographs 10 & 11, Encl. III) in the basal part of the olistolith or in the substrate. This folding conforms to the sedimentary transport direction i.e. N over S. Due to the drag small fragments of the bedded olistolith break away from its base and become embedded in the substrate (Photograph 10). Another type of deformation which is probably caused by independent olistolith transport, is the "flattening" of the olistolith by pull-apart along low-angle normal faults (Photograph 2, Encl. III, the inset sketch). Conditions for the development of such faults should be favourable near the end of the slide-transport. The imbrication resulting from the flattening is reverse (dipping "downstream") to the direction of the slide. Normal imbrication may occur when olistoliths pile up on one another. Fairly large folds may also develop in at least partly lithifield bedded olistoliths during transport (Photograph 5). Often such folds are not preserved intact, one finds isolated broken remnants embedded in the debris-flow matrix (Photographs 4 & 7). The average olistolith seems to have been lithified before transport, their outside forms often follow joint surfaces, they may have been faulted during transport. It is therefore remarkable that a large number of the limestone olistoliths, even the very large ones, are often subrounded and exhibit a fair sphericity. It is the more remarkable that the limestone olistoliths may have load- and flame-structures at their bottoms.

After deposition and subsequent burial the entire complex was folded and faulted in a complex synclinal structure, together with the formations on top. In a well-bedded slab-shaped turbiditic olistolith in the core of the syncline, regular buckle folds in sandstone beds are accompanied by beautifully developed divergent cleavage fans in the shale beds (Photograph 13). Irregular cleavages are developed throughout the mudstone matrix of the debris-flow units. The orientations of these cleavages seem to be dependent on the - random - orientation of both large and small clasts (Photographs 8 & 9). These cleavages impart to the mudstone a typical "scaly" appearance (cf. the Italian "argile scagliosi").

Palaeontology, lithostratigraphic origin and hypothetic provenance area of the exotics
The exotics of the Coriscao Olistostrome are either limestones, turbiditic sandstone/shale alternations or shallow-marine to continental sandstone/mudstone slabs. As far as has been observed, all limestones represent the algal biogenetic bank association which constitutes the dominant Upper Carboniferous limestone facies in the Cantabrian Mountains (van de Graaff,

1972). The turbiditic exotics represent a turbidite facies comparable to facies type E of Mutti & Ricci Lucchi (1976), the near-slope facies described above for the Liébana basin. The other sandstone/shale exotics are characterized by plant fossils, occurring on sandstone/shale interfaces or more dispersed in the mudstone which may also contain fossil root casts.

Fusulinid foraminifers have been obtained from a number of limestone exotics. Plant fossils were sampled on two localities which are probably both situated in the same huge faulted sandstone/mudstone olistolith just below the Panda Limestone in the eastern slope of the Pico Coriscao (Encl. I). All biostratigraphic data are summarized in Chapter III and in Encl. IV. The sample localities are given in Chapter III. A detailed treatise of the most important fusulinid biostratigraphic data is given in Chapter III. It appears that the Coriscao limestone exotics represent a number of limestone bodies deposited throughout the Westfalian, up to limestone bodies of lowermost Middle Cantabrian age. The continental sandstones seem but slightly older than the youngest limestones (Encl. IV). As usual no age could be obtained for the turbiditic sandstones. However, since all olistostrome material seems to be transported from the north (vide the synsedimentary deformations and the wedge-shape of the olistostrome), it seems reasonable to assume the same provenance area for all exotics. That provenance area must have had an Upper Carboniferous lithostratigraphy with vertical (and lateral) variations from a shallow shelf (limestones and continental sandstone) to a deep slope facies (turbidites). Such hypothetic succession is comparable to the Westfalian succession in the Redondo syncline in the Pisuerga area (van de Graaff, 1971).

The youngest exotics set a maximum age for the deposition of the Coriscao Olistostrome: an uppermost Myachkovian age (cf. limestone beds of the N_2-interval in the Donetz basin) which probably correlates with the earliest Middle Cantabrian (Encl. IV). Since the Panda Limestone, lying on top of the Coriscao Olistostrome, is unequivocally older (Chapter III of this paper), the contact at the base of the Panda Limestone must be a tectonic contact.

Stratigraphic classification

The Coriscao Olistostrome, formerly defined as a member of the Lechada Formation, with an approximate age of Westfalian C (Maas, 1974), appears to have the same age and represents a similar depositional facies as the near-by Remoña Olistostrome, a member of the Valdeón Formation (Maas, 1974). We propose that the Coriscao Olistostrome is redefined as the basal member of the Valdeón Formation.

PANDA NAPPE

The Panda nappe has its slide-horizon at the base of the Panda Limestone, extending from Casasuertes to the area of the Puerto Remoña, between Valdeón and Liébana (Encl. I). The nappe is covered unconformably by the Brañas Olistostrome. The nappe consists of only one formation, the Pandatrave Formation (Maas, 1974), with the lower Panda Limestone Member and an upper turbidite member. This description implies a redefinition: the Brañas Olistostrome, on top of the Pandatrave turbidites and formerly considered as the upper member of the Pandatrave Formation, is now classified as a member of the Valdeón Formation.

The Pandatrave Formation (and with it the Panda nappe) has an estimated maximum thickness near 1000 m: 100-200 m for the Panda Limestone and 800-1000 m for the turbidites (estimates after Kamerling, 1962; and Kutterink, 1966). The nappe tapers off to SW: near Casasuertes the Panda Limestone is laterally bounded and unconformably covered by an olistostrome unit very similar to the Coriscao Olistostrome: the coherent limestone member passes into a string of large olistoliths of Panda limestone mixed with and covered by other limestone olistoliths in a marly mudstone matrix. A similar lateral unconformable contact is exposed in the north, 200 m S of locality 81139. The Pandatrave turbidites in the nappe at that place are thick beds of limestone arenites, conglomerates and breccias. Large blocks and olistoliths of that material are floating in the adjacent Remoña Olistostrome; the contact is irregular but fairly steep and may have been accentuated by faulting. The lateral unconformity extends to NW, but to E (east of locality 81139), the Panda Limestone reappears, resting upon the Remoña Olistostrome. Apparently the olistostrome envelopes the northern margin of the nappe. West of locality 81095, the base of the nappe is visible as a lens of Upper Carboniferous yellow nodular limestone, apparently a product of shear along the nappe slide horizon (Photograph 14). A similar nodular limestone lens was found N of Casasuertes (locality 81176A). In other localities (e.g. near locality 81194) a tectonic breccia occurs at the base of the Panda Limestone.

The Panda Limestone represents the algal biogenetic bank facies. Thick (0.5-1 m) limestone-breccia turbidites and some restricted debris flow units (with large boulders to small olistoliths) rest on top of the Panda Limestone and constitute a transition to siliciclastic turbidites and laminated mudstones (facies types E & G, Mutti & Ricci Lucchi, 1976). In the upper half of the turbidite member thick-bedded calciclastic conglomerate and breccia turbidites, with an occasional olistolith-bearing debris flow, largely replace the siliciclastic turbidites. This well-bedded upper part of the Pandatrave Formation forms a striking contrast with the surrounding olistostrome deposits and facilitates the mapping.

The Panda nappe may well have the same provenance area as the exotics of the Coriscao Olistostrome. A similar shelf to slope facies area has to be inferred, and not only falls the age of the Panda Limestone well within the range of the Coriscao exotics but also the fusulinid assemblages indicate that many of the Coriscao exotics could be derived from the Panda Limestone.

THE OLISTOSTROME UNIT NEAR CASASUERTES

As mentioned above, near Casasuertes the Panda nappe is in lateral unconformable contact with an olistostrome unit consisting mainly of Upper Carboniferous limestone exotics. The only biostratigraphic determination (sample 81172, Chapter III) sets a maximum age identical to the maximum age of the Coriscao Olistostrome (Encl. IV). Owing to lack of time and of outcrops, the olistostrome has not been mapped to E and W.

REMONA OLISTOSTROME

The Remoña Olistostrome Member was defined as the basal member of the Valdeón Formation, resting unconformably on the Liébana turbidites of the upper Deva area (Maas, 1974). The present study has revealed that the olistostrome envelopes the northern part of the Panda nappe, its upper part is probably a lateral equivalent of the Brañas Olistostrome and it passes vertically into the turbidites of the upper part of the Valdeón Formation.

The Remoña Olistostrome is poorly exposed. The best outcrops are in the brook running E from Puerto Remoña, in the boundary mountain ridge between Liébana and Valdeón, along the road from Santa Maria de Valdeón to Puerto Pandatrave, and at the localities indicated on the map (Encl. I & II).

The Remoña Olistostrome differs from the Coriscao Olistostrome both with regard to its matrix and its exotics. Remobilized to severely slump-distorted laminated mudstones and "slope"-turbidites constitute the bulk of the predominantly siliciclastic matrix (cf. Photograph 15). The difference between entirely remobilized slope sediments and slope sediments that, though slump-deformed, retained their internal coherence, is apparently gradual (cf. the slump types in Liébana described by Maas, 1974). Conglomeratic and breccious debris flow units, as described for the Coriscao Olistostrome, are of subordinate occurrence, with exception of the zone marginal to the lateral unconformity with the Panda nappe. However, scaly cleavage in mudstone often bearing small angular clasts, is of common occurrence.

In the Remoña Olistostrome all kinds of exotics, in all shapes and sizes, do occur. Rock units of various ages ranging from Frasnian up to Upper Myachkovian (= Lower Cantabrian), are represented (cf. Encl. IV). The samples used to obtain conodont faunas represent at least the Middle Devonian to Lower Carboniferous limestone formations of the Palentine facies area (Cardaño, Vidrieros and Villabellaco Formations, cf. Maas, 1974). Rocks derived from the Famennian Murcia Quartzite, the Vegamián Shale (Tournaisian to Viséan) and the Namurian Caliza de Montaña Limestone were also observed as olistoliths. The Caliza de Montaña (e.g. W of locality 81095) is not developed as a finely laminated homogeneous limestone mud but as a banded calcarenite (cf. its development in the Montó unit). Olistoliths of Upper Carboniferous limestones and limestone conglomerates occur in numerous localities. Owing to the generally poor degree of exposure, slabs of turbiditic sandstone/shale alternations resting in a slump-distorted siliciclastic matrix, cannot be distinguished without doubt. Frequently, however, coherent but entirely overturned turbidite sequences, crop out between outcrops of slump-distorted siliciclastics.

The provenance area of the Remoña Olistostrome, with exception of its zone marginal to the Pandatrave turbidites in the Panda nappe, was at least situated in the Palentine facies area, and probably between the Picos de Europa shelf and the Liébana basin. In that slope area the submarine denudation providing the exotics apparently had progressed in comparison with the submarine denudation during the Coriscao Olistostrome deposition when only Upper Carboniferous material was transported.

BRANAS OLISTOSTROME

The Brañas Olistostrome corresponds to the Branas Member introduced by Maas (1974) as top member of the Pandatrave Formation. Our recent investigations have revealed that the Brañas Olistostrome has an unconformable basal contact with the top of the Panda nappe. The Brañas Olistostrome is consequently redefined as a member of the Valdeón Formation. The thickness of the Brañas Olistostrome below the Montó unit is 500-1000 m (estimate after Kutterink, 1966).

In its lithologic aspects the Brañas Olistostrome is intermediate between the Coriscao and the Remoña Olistostrome. Large parts of it have a matrix of breccious marly mudstone and bear Westfalian limestone olistoliths (e.g. the area near localities 81103 and 81109 - Encl. I, Encl. IV). However, somewhat to north (localities 81115 & 81116) the matrix consists of remobilized to slump-distorted siliciclastics as known from the Remoña Olistostrome, and there the largest single limestone olistoliths known in the mapped area (Photograph 16) represent the upper, fossiliferous part of the Namurian Caliza de Montaña Formation. Further to N we find an accumulation of olistoliths of a polymict limestone conglomerate, containing well-rounded limestone pebbles of a black laminated mudstone (the lower part of the Caliza de Montaña Formation), pebbles of Villabellaco Limestone, pebbles of the Upper Devonian nodular limestones, conglomeratic pebbles that miniaturize the conglomerate lithology, and quartzitic pebbles that may have been derived from the Murcia Formation. The mentioned formations are also represented in other homogeneous exotics in the neighbourhood.

In its upper reaches the Brañas Olistostrome probably has lateral unconformable contacts with the NE and SW flanks of the Montó unit; the olistoliths near the NE flank of the Montó unit are mainly large angular slabs of Murcia quartzite. The lateral contact between the Brañas and the Remoña Olistostromes is not exposed.

The provenance area of the Brañas Olistostrome includes a siliciclastic mud slope, a shallow shelf where well-rounded pebble conglomerates could accumulate, and possibly the southern margin of the Picos de Europa limestone shelf (e.g. the large chunks of fossiliferous Caliza de Montaña). The olistostrome may represent a slope to shelf to instable coastal area. intersected by submarine canyons and their contributing channels, slump scars etc. The geology of that area may have consisted of basinward dipping strata; the Lower Carboniferous and Upper Devonian formations cropping out in the coastal area were covered by Upper Carboniferous shelf/slope sediments toward the basin.

MONTÓ UNIT

The Montó unit, like the Panda nappe, is an apparently rootless structure resting upon - and in lateral contact with - the Brañas Olistostrome. It consists of a number of thrust sheets stacked upon one another with a normal imbrication with respect to a N-S transport. Frequently the sheets are intricately folded, intraformational folds and disharmonic folds are the dominant fold types. The structure was investigated and mapped in detail by Kutterink (1966).

The lithostratigraphy of the Montó unit represents the Palentine Middle Devonian to Lower Carboniferous topped by the unfossiliferous part of the Caliza de Montaña Formation, which is here developed as a well-bedded alternation of calcilutites, calcarenites, limestone breccias, limestone conglomerates and shales. The calcarenites resemble the olistolith found in the Remoña Olistostrome near locality 81095. The Montó unit probably derives from the area between the Liébana basin and the Picos de Europa shelf.

VALDEÓN FORMATION, UPPER PART

The upper part of the Valdeón Formation has a stratigraphic thickness of about 2000 m (Kutterink, 1966) and rests with an intraformational unconformity on top of the Montó unit. Its presumably gradual basal contacts with the Remoña and Brañas Olistostromes are not exposed.

The upper part of the Valdeón Formation is a turbidite sequence consisting mainly of facies types E and G (Mutti & Ricci Lucchi, 1976), the "slope"-facies, interfingering with lenses of the conglomeratic/arenitic facies. The conglomerate lenses increase in number and thickness to the top of the formation where, NW and outside the mapped area, the Camborisco Conglomerate (Kutterink, 1966) probably represents a small submarine fan of the type described from Liébana by Rupke (1977). Kutterink suggests that the sequence is asymmetric with a notable increase of its thickness from E to W. The sequence is mainly siliciclastic, limestone clasts occurring only in the conglomerates and associated coarse arenites.

THE NEONCITO STRUCTURAL UNIT AND ITS SURROUNDING SEDIMENTS

The Neoncito structural unit (Kutterink, 1966 and Gomez, 1974) is a nappe structure resting on top of the Valdeón turbidites, W of and outside the mapped area (Fig. 1). Its lithology represents the stratigraphy of the "Asturian geanticline": the Cambrian to Ordovician succession of Lancara, Oville and Barrios formations is covered, with a paraconformable contact, by the Viséan Alba Formation and the Namurian Caliza de Montaña Formation. The unit clearly derives from outside the Palentine facies area and according to its lithology it constitutes the easternmost nappe of the Ponga nappe province (Fig. 1). Still it seems to be a rootless structure; like the Panda and Montó structures it has a lateral unconformable flank contact: both Kutterink (op. cit.) and Gomez (op. cit.) map a lateral unconformity along the NE flank of the nappe, cutting through it from its top to its base. The unit seems to be embedded in the uppermost Valdeón turbidites, which according to Gomez continue on top of the Neoncito nappe.

THE PICOS DE EUROPA NAPPES AND THE LEBENA FORMATION

The Picos de Europa nappes (Fig. 1) are emplaced on top of both the Namurian-Westfalian flysch sediments of Liébana and on top of the Middle Cantabrian alternation of olistostromes, nappes and turbidites of the Valdeón area. The Liébana and Valdeón "basins" do not form a continuous conformable succession. The Liébana basin deepened in a general sense from W to E, its synsedimentary gravitational deformations and slump/olistostrome transports occurred mainly from NNE (Maas, 1974). The Valdeón succession rests unconformably on the western - i.e. the relatively shallow - part of the Liébana basin, and seems to have deepened from E to W (Kutterink, 1966). Its olistostrome and nappe transports occurred from N to NW, excluding the Neoncito nappe which seems to have been transported from WNW. The Picos de Europa nappes were transported from N, probably by a process of nappe for nappe gravitational gliding.

The Picos de Europa nappes all derive from the same limestone shelf area and have an identical lithostratigraphy: the Viséan to Westfalian D-Middle Cantabrian limestone succession of the Alba, Caliza de Montaña and Picos de Europa formations. The youngest fusulinid fauna obtained from near the top of the succession (Maas, 1974, locality 70271 - the Tabla de Lechugales fauna*), sampled again in 1980 - basal part of Protriticites Zone) is near the Moscovian/Kasimovian boundary which may be correlated with the Middle Cantabrian. A succession of shales containing conglomeratic to arenitic limestone turbidites and olistostrome levels with boulders and large olistoliths deriving from the Picos de Europa formation, lies on top of the eastern half of the first nappe, in part with a well-developed erosional basal conglomerate. The succession is covered in turn by the second nappe. Maas (1974) described this flysch succession as two separate formations. The Aliva Formation, without a basal conglomerate, was interpreted as a lateral equivalent of the Picos de Europa Limestone, a supposition that seemed warranted by the biostratigraphic data then known. The Lebeña Formation, resting with an angular unconformity on the partly eroded Picos de Europa Limestone, is obviously younger than that formation. Follow-up of the investigations of the fusulinid faunas of existing and of new samples, by the second author has now set an age of top Myachkovian to basal Kasimovian correlating with Middle Cantabrian (Protriticites Interval Zone cf. Donetz basin limestones N_2-N_3), for a number of limestone samples, both from the Aliva and from the Lebeña Formation. The two formations represent the same lithofacies of the same maximum age in the same area, their distinction is superfluous. We propose to denote the entire sequence as the Lebeña Formation (cf. Martínez García, 1981; Savage & Boschma, 1980), since the succession exposed in the road section near the village Lebeña is the more complete, including the basal conglomerate and having a thickness of at least 800 m.

The remaining four Picos de Europa nappes have no appreciable amounts of intervening sediments, the normal contact being Alba Formation on top of the Picos de Europa Formation of the preceding nappe. So the redefined Lebeña Formation is essentially confined to the first nappe, indicating a considerable subsidence of that nappe, but also a position at the earth surface, shortly after its transport.

PRELIMINARY CONCLUSIONS

Synsedimentary nappe formation

In the foregoing part of this chapter we have described a succession of olistostrome and flysch sediments (Coriscao Olistostrome, Remoña Olistostrome, Brañas Olistostrome and Valdeón turbidites) constituting the Valdeón Formation. This formation is overlain by the first (lowermost) Picos de Europa nappe which in turn is covered unconformably by the Lebeña Formation. The age difference between the base of the Valdeón Formation and the Lebeña Formation seems minute: the beginning of the Middle Cantabrian (i.e. Protriticites Zone; cf. Donetz basin, interval of N_2 limestones) versus a slightly younger Middle Cantabrian (i.e. Protriticites Zone; cf. Donetz basin, limestones N_2-N_3). Of course both ages are maximum ages derived from allochthonous materials. Reliable age determinations on autochthonous materials were established for the unconformable Puentellés Formation (Martínez García, 1981) on top of the uppermost (fifth) Picos de Europa nappe. The age of that formation is Upper Kasimovian (cf. limestone bed O_3 in the Donetz basin, and Triticites irregularis Zone of the USSR) which may correlate with upper Stephanian A (van Ginkel, 1971). So the maximum time span for the sedimentation of the Valdeón Formation is about the N_2-O_3 time span, which probably does not surpass five million years but the real time span may easily have been the N_2-N_3 time span which may be less than one million years.

Intercalated between the olistostrome members of the Valdeón Formation are two nappe units: the Panda nappe and the Montó unit. The Neoncito nappe is possibly likewise intercalated within the uppermost part of the Valdeón Formation. The Panda nappe rests with a basal slide contact, locally exemplified by breccias or nodular limestone horizons, upon the Coriscao Olistostrome. The contact of the Panda nappe along its northern limit with the Remoña Olistostrome is a lateral unconformity. A similar lateral unconformity is exposed at the SW limit of the Panda nappe near Casasuertes. The Panda nappe is covered unconformably by the Brañas Olistostrome which is probably laterally equivalent to the upper part of the Remoña Olistostrome. The Montó unit rests with its basal slide surface upon the Brañas Olistostrome. The NE and SW lateral contacts between the Montó unit and the uppermost part of the Brañas Olistostrome, have to be interpreted as lateral unconformities. The Montó unit is unconformably covered by the partly conglomeratic base of the Valdeón turbidites.

The exotics in the olistostrome members exhibit an inverted stratigraphy: early Middle Cantabrian to Westfalian exotics in the Coriscao Olistostrome, early Middle Cantabrian to Upper Devonian exotics in the Remoña Olistostrome, and Westfalian to Upper Devonian exotics in the Brañas Olistostrome. The ages of the nappe materials do fit in this inverted stratigraphy: Upper Westfalian for the Panda nappe and Namurian to Givetian for the Montó unit (cf. Encl. IV). The Neoncito nappe would also fit, being highest in the succession it exposes the oldest rocks (i.e. the Cambrian Lancara Formation).

*) *Tabla de Lechugales*:
Ozawainella aff. *vozhgalica* Safonova, 1951; *Pseudostaffella sphaeroidea cuboides* Rauser-Chernoussova, 1951; *Schubertella* ex gr. *subkingi* Putrya, 1939; *Fusiella rawi* (Lee, 1927) (syn. *Fusiella* cf. *lancetiformis* Putrya (1939) of Loc. P 36, p. 106, van Ginkel (1965); *Fusulina* cf. *quasicylindrica brevis* Lee, 1927; *Obsoletes* ex gr. *grosdilovae* (Miklucho-Maklaj, 1949).
Zonation: Protriticites Zone; basal part.
Correlation: top of Myachkovian or base of Kasimovian; N_2 or N_3 of Donetz Basin.

To explain the lateral unconformable contacts of the nappes we must accept that the nappes are rootless structures enveloped by the olistostrome sediments. The Panda and Montó nappes fit within the inverted stratigraphy of the olistostrome exotics as if they were exotics too, albeit extremely large ones, deriving from the same area of provenance and resulting from the same progress of denudation. The Neoncito nappe, however, clearly derives from an area outside the Palentine basin, with a different lithostratigraphy. We propose that the nappes have been transported by sliding along the surface of a submarine slope and were emplaced in the olistostrome sediments of the Valdeón Formation while the sedimentation was going on.

The nappes and the olistostromes are derived from the same area of provenance. That provenance area must have had a lithostratigraphy consisting of Palentine Devonian, the normal Lower Carboniferous succession, a somewhat restricted sedimentation of Namurian clastic limestones and shales, and a Westfalian to Middle Cantabrian succession of alternating shallow shelf limestones (the biogenetic bank associations), siliciclastic flysch sandstones ("slope facies turbidites" - cf. Liébana basin) and calciclastic flysch arenites, conglomerates and breccias. This conjectural lithostratigraphic succession, constructed from the allochthonous elements of the olistostrome and nappe sequence, does not match the known lithostratigraphies of the areas to W, S and E of the studied area. There is a similarity with the overall stratigraphy of the Pisuerga area, but that succession occurs too far to SE, across the Prieta fault and the Polentinos block, and is preserved under a sequence of shallow-marine to terrestrial Cantabrian sediments. The remaining place for the provenance lithostratigraphy is N of the studied area below the Picos de Europa nappes. N or NW is also the direction that concurs with the wedge shape of the Coriscao Olistostrome, the internal synsedimentary deformations and with the general orientation and occurrence of the olistostrome and nappe units. Moreover, the constructed lithostratigraphy matches the hypothetic lithostratigraphy of the transitional area between the shallow-water limestones of the Picos de Europa shelf and the flysch sediments of the Liébana basin. We mentioned such a transitional area north of Liébana in our paragraph on the geologic setting as a hypothetic provenance area of submarine fan sediments (Rupke, 1977).

Alternative nappe formation

As a prima facie alternative for the formation of the nappe structures one may propose a process of tectonic imbrication: a Middle Devonian to Upper Westfalian rock sequence is in part eroded and subsequently covered unconformably by the olistostromes and turbidites of the Valdeón Formation; later tectonic imbrication effects the present arrangement of rock units. Such explanation implies that the basal thrust planes of the nappe units would have to be continued into the mixed lithology of the adjacent olistostromes ¬ rock mechanically an improbable suggestion. We would have a complicated process of structural imbrication along thrust planes that have no rock mechanical soundness, preceded by a fortuitous sculpting by erosion and subsequent juxtaposed sedimentation of flysch sands and olistostromes having no distinct relation with the overall configuration of the sedimentary basin. We would not be able to explain the existence of unconformable lateral contacts between the nappe units and the olistostromes; moreover the fit of the nappes in the inverted stratigraphy of the olistostromes would be difficult to explain.

Structural setting of the nappe formation

The entire sequence of olistostromes, intercalated nappes and the covering flysch beds, has a thickness of more than 5 km and is deposited in maybe about a million years. The deposition must have been compensated by adequate subsidence of the basin and uplift and denudation of the provenance area. Vertical movements of that order of magnitude and velocity are difficult to imagine without being fault-controlled. The downsliding of the nappes can also be explained as superficial movements subsidiary to near-vertical fault movements. The pertinent faults would constitute an E-W striking fault zone controlling the development of the transition area between the Picos de Europa shelf and the Liébana basin. These ideas are elaborated in Chapter IV. In that chapter we will try to link the results of our investigations to the ideas developed by Reading (1975) and Savage (1979), implying control by deep-seated fundamental faults of the sedimentary and structural development in the Cantabrian area during the Upper Carboniferous.

CHAPTER III

PALAEONTOLOGIC DATA

A DISCUSSION OF FUSULINID ASSEMBLAGES YOUNGER THAN THE ONES OF THE PANDA LIMESTONE AND OCCURRING IN STRATA LATERAL TO OR BELOW THIS LIMESTONE

The thesis of the present paper holds that the Pandatrave Formation with the Panda Limestone at its base should be envisaged as a sedimentary sequence of allochthonous nature surrounded by and resting upon younger strata. This idea of the Pandatrave Formation as a nappe structure is supported by the occurrence in some olistoliths lateral to or below the Panda Limestone of fusulinid assemblages younger than those of the Panda Limestone. These younger faunas were found near Casasuertes in an olistolith lateral to the Panda Limestone (locality 81172), in three olistoliths of the Coriscao Olistostrome below the Panda Limestone (localities 81196, 81197, and 80003), and in an olistolith of the Remoña Olistostrome below the Panda Limestone (locality 81098) (Encl. IV and the locality maps Encls. I and II). Correct age estimates with respect to these olistoliths are of crucial importance to the views put forward in this paper. For this reason the fusulinid species which define the age of these olistoliths between relatively narrow limits are discussed, and illustrated in the Plates I-III.

Locality 81098 (Remoña Olistostrome)

Species content: Staffellinae
Millerella sp.
Ozawainella ex gr. *pseudoangulata* et *mosquensis*
Pseudostaffella compacta Manukalova, 1950
Pseudostaffella sphaeroidea (Ehrenberg, 1842)
Schubertella cf. *subkingi* Putrya, 1939
Fusiella sp.
Fusulinella ex gr. *bocki* Möller, 1878

Zonation: Fusulinella Zone; Subzone B, subdivision B 3.
Correlation: Upper Moscovian; Upper Myachkovian; M_{10} or N_1 of Donetz Basin

Of the species listed only *Fusulinella* ex gr. *bocki* is a common constituent in the sample. It is this species which largely determines the zonal allocation and is therefore described below.

Fusulinella ex gr. *bocki* Möller, 1878
Plate I, Figs. 1-8

Material: 14 specimens including 7 axial sections.
Number of whorls: average 6½, range 5½-7½. Diameter proloculum: average 103 mc, range 84-115 mc. Diameter 6th to 7½ whorl: average 2400 mc, range 1680-2910 mc. Average diameter in outer whorls: 1640 mc (5th wh., N = 1), 2175 mc (6th wh., N = 5), 2400 mc (6½ wh., N = 3), 2560 mc (7th wh., N = 1), 2825 mc (7½ wh., N = 2). Diameter of 4th whorl (D_4): average 935 mc, range 725-1130 mc. Specimens with few whorls tend to have a large proloculum and a large D_4. The average D_4 in relation to number of whorls is as follows: 1125 mc (5 whs., N = 1), 1010 mc (6 whs., N = 3), 980 mc (6½ whs., N = 5), 925 mc (7 whs., N = 1), 730 mc (7½ whs., N = 2). Length/diameter ratio (L/D) in 6-7½ whorl: average 1.87, range 1.70-2.05. The test changes from oval/short fusiform to fusiform from 6-7½ whorl. Chomata extend to the poles in inner 2-4½ whorls; in subsequent whorls relatively narrower, symmetrical or asymmetrical chomata appear of quadrangular, triangular, or rounded semicircular shape; in inner 5-7 whorls they are high or moderately high; the slope at the tunnel side is usually steep. The tunnel forms a wide slit in outer whorls, which is about 1/3 as high as the corresponding chamber; its path is usually irregular. Septal folding is weak in general and is restricted to a narrow zone along the axis. Some specimens, however, show moderately intense folding at the poles of the 6-7½ whorl. Maximum thickness of wall including tectoria is attained in the 6-6½ whorl and amounts to 95 mc. Thickness of spirotheca in the outer whorl (6½-7½ wh.) usually varies between 65 and 80 mc, and that of the protheca between 17 and 30 mc. The wall is typically fusulinellid in inner 6 whorls; from thereon the upper tectorium disappears and mural pores can be observed in lower tectorium and diaphanotheca. This type of wall is either maintained up to the 7½ whorl or the distinction between lower tectorium and diaphanotheca becomes more and more indefinite until in the 7½ whorl both layers have merged and the wall consists of a single porous layer overlain by the tectum.

The present species is one of the largest of the genus. Its advanced wall structure shows that it is close to *Protriticites*. However, the late appearance of the Protriticites-type wall does not permit its inclusion in that taxon. *Fusulinella* ex gr. *bocki* described from the Cotarazo Limestone (Pisuerga Basin, locality P 58) is closely related and possibly ancestral to the present species (van Ginkel, 1965). This Cotarazo form is slightly older (Fusulinella Zone; Subzone B, Subdivision B 2) and has smaller dimensions. Also similar are *Fusulinella timanica* Rauser, 1951 and *Fusulinella pseudobocki* Lee & Chen, 1930. Both species as well as *Fusulinella* ex gr. *bocki* from the Cotarazo Limestone differ in their less evolved wall structure.

Locality 81197 (Coriscao Olistostrome)
Species content: *Millerella* sp.
Pseudostaffella sp.
Ozawainella ex gr. *pseudoangulata* et *mosquensis*
Ozawainella cf. *kumpani* Sosnina, 1951
Pseudotriticites ? spp.
Fusulinella cf. *bocki* Möller, 1878 (→ *Protriticites*)
Fusulinella pseudobocki Lee & Chen, 1930 (→ *Protriticites*)

Zonation: Fusulinella Zone; Subzone B, top of subdivision B 3.
Correlation: Upper Moscovian; Upper Myachkovian; N_1 or N_2 of Donetz Basin.

The following species are particularly useful for establishing the olistolith's age: *Fusulinella* cf. *bocki* Möller, 1878, *Fusulinella pseudobocki* Lee & Chen, 1930 and *Pseudotriticites*? spp. The wall structure of both species of *Fusulinella* resembles that of *Fusulinella* ex gr. *bocki* from locality 81098 (Remoña Olistostrome) yet in the present two species the wall is more advanced, i.e. closer to the Protriticites-type wall. The almost intermediate systematic position between *Fusulinella* and *Protriticites* is indicated informally as *Fusulinella* (→ *Protriticites*).

Fusulinella cf. *bocki* Möller, 1878 (→ *Protriticites*)
Plate II, Fig. 1

Material: 1 excentric axial section.
Number of whorls: 4½-5. Diameter (D), radius vector (Rv), and length/diameter ratio (L/D) for the last whorl are respectively: 1165 mc, 640 mc, and 1.67. The test changes from nautiloid (1st wh.), spherical or oval (2nd wh.), short subcylindrical (flattened spirotheca in median region) (3rd wh.), to short fusiform (4-5th wh.). Chomata in inner 2½ whorls low or moderately high, and high in the succeeding whorls asymmetrical, in the outer whorl symmetrical and rounded or subquadrangular. Septa in inner 4 whorls straight or only slightly twisted at the poles; septa are clearly folded in the ultimate whorl but only at the poles. Spirotheca is rather dark and a differentiation of it cannot be observed in inner whorls. The outer two whorls show very fine mural pores which are also observed in the chomata of the last whorl. The wall in outer whorls is 29-42 mc.

Fusulinella cf. *pseudobocki* Lee & Chen, 1930 (→ *Protriticites*)
Plate II, Figs. 2-4

Material: 1 axial section, 3 oblique sections. The following data refer to the axial section. Number of whorls: 5. Diameter proloculum (D_o): 88 mc. Diameter (D), radius vector (Rv), and length/diameter ratio (L/D) for the 5th whorl are respectively: 1310 mc, 690 mc, and 2.18. $D_{max.}$: 1435 mc (oblique section). The test changes from thickly nautiloid (1st wh.), oval/short fusiform (2nd wh.), to fusiform (3-5th wh.). Chomata in inner 4 whorls are of medium height and wide; they have steep (~90°) slopes at the tunnel side. Septal folding starts at the poles of the 4th whorl but is conspicuous only at the poles of the 5th whorl. The wall is neither typically fusulinellid nor characteristic for *Protriticites*. A very conspicuous feature is the dark colour. A differentiation of the wall is often indistinct. Mural pores can be observed in the outer two whorls, sometimes only in the last whorl. In the last whorl the upper tectorium disappears and the porous wall shows in addition to the tectum, either a diaphanotheca and lower tectorium or only a single layer. The maximum thickness of the wall is attained in the penultimate whorl and amounts to 50-55 mc. In one specimen a thickness of 80 mc was measured. The thickness in the ultimate whorl is 17-35 mc. A conspicuous difference between the two species of *Fusulinella* from this locality and *Fusulinella* ex gr. *bocki* from locality 81098 (Remoña Olistostrome) is the dark-coloured wall of the present form. These dark walls were also observed in other species of this area which should be considered as true *Protriticites* (see below).

Pseudotriticites ? spp.
Plate II, Figs. 5-10

Material: 4 Specimens (excentric axial, transversal and oblique sections).
There are two species present, both of which probably descended from the genus *Beedeina*. Their systematic position is about intermediate between Eurasian-type *Beedeina* and typical *Pseudotriticites*. One species (sp. A) is very close to species of the group of *Beedeina samarica* (Rauser & Beljaev, 1937); the other (sp. B) resembles *Pseudotriticites donbassicus* (Putrya, 1939) the genotype of *Pseudotriticites*, as well as species of the group of *Beedeina elegans* (Rauser & Beljaev, 1937), notably *Beedeina elegans devexa* (Rauser, 1951).
Species A (3 specimens).
Number of whorls: 4½-6. Diameter proloculum: 107-118 mc. Diameters of three specimens with 4½, 5½ and 6 whorls are respectively: 1080 mc, 1525 mc, and 1755 mc. The radius vector in this order is: 625 mc, 835 mc, and 1000 mc. Diameter 4th whorl (D_4): 790-870 mc. L/D ratio: 1.67-1.75. The shell is subrhomboidal or short fusiform in outer whorls. Chomata are high and wide in inner 2-3 whorls; they are replaced by pseudochomata in the 3-4th whorl and are absent in the 5-6th whorl. Septal folding is very high and rather regular. Septal pores have been observed occasionally. Mural pores are distinct from the 4-5th whorl; in earlier volutions wall

structure is rather obscure although locally one may observe a diaphanotheca in the 2½-3½ whorl. The wall is 25-45 mc thick in outer whorls.
Species B (1 specimen).
Number of whorls: 4. Diameter proloculum: 50 mc. D_4: 900 mc. Rv_4: 460 mc. L/D: 2.45. The test changes from short fusiform (1-2nd wh.), fusiform or elongate subrhomboidal (3rd wh.), to fusiform (4th wh.). Chomata are low and wide in inner whorls and moderately high and relatively narrow in the 4th whorl. Septal folding is high and regular. Mural pores are quite distinct in the 4th whorl; wall thickness: 31-38 mc.

Locality 81196 (Coriscao Olistostrome)
Species content: *Pseudoendothyra* sp.
Pseudoendothyra umbonata Rauser-Chernoussova, 1951
Ozawainella ex gr. *pseudoangulata* et *mosquensis*
Ozawainella cf. *pseudotingi* Putrya, 1956
Ozawainella cf. *nikitovkensis* Brazhnikova, 1939
Pseudostaffella cf. *paraozawai* Manukalova, 1951
Pseudostaffella sp.
Fusiella sp.
Taitzehoella ex gr. *librovitchi* (Dutkevich, 1934)
Hemifusulina? sp.
Fusulina cylindrica Fischer, 1830
Pseudotriticites? sp.
Fusulinella cf. *asiatica* Igo, 1957
Fusulinella sp.
Protriticites sp.

Zonation: Protriticites Zone; near base of the zone.
Correlation: Upper Moscovian; top of Myachkovian; N_2 of Donetz Basin

The present locality as well as the localities 80003 and 81172 described hereafter yielded specimens which can be referred to primitive species-groups of the genus *Protriticites*. Their presence is ample proof that the olistoliths which contain these forams are of younger age (i.e. Protriticites Zone) than the overlying Panda Limestone, the upper part of which yielded fusulinids of the Fusulinella B Subzone, B 2 subdivision (p. 356). In addition to *Protriticites* sp., locality 81196 yields *Fusulina cylindrica* Fischer, 1830 and *Pseudotriticites*? sp. both providing additional evidence for the young age of the olistolith in comparison with the Panda Limestone.

Pseudotriticites ? sp.
Plate II, Figs. 14-15

Material: 1 axial section.
Number of whorls: 4½. Diameter proloculum: 109 mc, $D_{4½}$: 1625 mc, $Rv_{4½}$: 918 mc, D_4: 1230 mc, L/D: 2.25. The shell is subrhomboidal up to the 2½ whorl and elongate subrhomboidal thereafter; lateral sides are straight, or slightly concave at the poles; periphery broadly arched. The fairly high chomata extend to the poles in the inner 2 whorls; they are still wide in the 2½ whorl and absent in the last whorl. The tunnel is very narrow in the inner 3 whorls and about as high as wide. Septal folding reaches the tunnel in the 3rd whorl and extends from pole to pole in subsequent whorls. Septal pores are observed at the poles of outer whorls. The wall may be homogeneous in inner whorls; distinct pores appear in the 3rd whorl and here the wall shows tectum, diaphanotheca, and lower tectorium. In the outer whorl the lower tectorium and the diaphanotheca locally merge which results in a wall having a single porous layer only.

The single specimen present may be conspecific with *Pseudotriticites* ? sp. A from locality 81197 in spite of the larger L/D ratio of the present form. The larger L/D ratio, the fewer whorls and the Pseudotriticites-type wall distinguish our form from the similar *Beedeina samarica* (Rauser & Beljaev, 1937).

Fusulina cylindrica Fischer, 1829
Plate II, Figs. 11-13

Material: 5 specimens including an axial and a sagittal section.
Number of whorls: 4½. Diameter proloculum (D_0): 160-238 mc (N = 2). $D_{4½}$: 1150-1510 mc (N = 3). $Rv_{4½}$: 690-780 mc (N = 2). D_4: 1025 mc (N = 1). Length/diameter ratio (L/D) in the last whorl: 3.30 (N = 1). The test in outer whorls is elongate fusiform. Narrow chomata of medium height in the inner 1½-3 whorls are replaced by pseudochomata in the succeeding whorls; these deposits can be absent in the last whorl. Tunnel narrow and high in the inner whorls, but wide and relatively low in the last whorl. Inner 1½-2 whorls show straight septa in the median region, i.e. above the tunnel and the chomata; in subsequent whorls folding may extend from pole to pole, but is rather weak in the median region. Axial filling is absent or weakly developed. The wall is rather thin throughout growth; its thickness amounts to 20-40 mc in outer whorls; up to the 3-3½ whorl one can distinguish a lower tectorium, a diaphanotheca and locally an upper tectorium although in the inner 1-2 whorls the wall is often indistinctly differentiated. Distinct mural pores appear as early as the 3rd whorl; the last whorl has only a single - porous - layer below the tectum. With respect to wall structure the Spanish specimens slightly differ from *F. cylindrica*. According to Thompson (1945, p. 446), who studied topo-

types of the species, mural pores are poorly developed and distinct in the last (= 5th) whorl only.

The present find of *Fusulina cylindrica* Fischer establishes the occurrence in NW Spain of this familiar species. However, the species has been mentioned by others who claimed its presence. Verneuil & Collomb (1852, pp. 124, 125) have been the first to report on the occurrence of fusulinid foraminifera in NW Spain. They mention the presence of *Fusulina cylindrica*. The precise locality as well as illustrations of the form were not given but the information suggests that they found this species in Asturias amongst the rich Lower Moscovian fusulinid faunas of the Lena Formation. Should this be true then the fauna cannot have included *Fusulina cylindrica* in its present-day strict definition since this species does not occur below the Upper Moscovian. From the same region a subspecies of *Fusulina cylindrica* was described by Gübler in 1943 at approximately the boundary between the Lena and Sama Formations. This subspecies (= *Fusulina cylindrica* var. *hispanica* Gübler, 1943) has been referred to *Hemifusulina* (see: *Hemifusulina hispanica* (Gübler, 1943); van Ginkel, 1973).

Protriticites sp. indet.
Pl. II, Figs. 16-17

Material: 4 specimens; 1 sagittal, 1 transversal and 2 tangential sections.
There are only a few and ill-oriented sections available whose description does not seem very rewarding. The advanced wall structure permits their allocation to the genus *Protriticites.*

Locality 80003 (Coriscao Olistostrome)
Species content: *Beedeina* sp.
Fusulinella pseudobocki ovoides Rauser-Chernoussova, 1951
Fusulinella ex gr. *pseudobocki* Lee & Chen, 1930
Protriticites ex gr. *pseudomontiparus* Putrya, 1948

Zonation: Protriticites Zone; near base of the zone.
Correlation: Upper Moscovian; top of Myachkovian; N_2 of Donetz Basin.

The species described below determines the allocation in the Protriticites Interval Zone of this assemblage.

Protriticites ex gr. *pseudomontiparus* Putrya, 1948
Plate III, Figs. 1-4

Material: 3 specimens; 1 sagittal, 1 oblique and 1 near-axial section.
Number of whorls: 4½-5. Diameter proloculum: 120 mc (N = 1). Diameter of outer whorl ($D_{4½-5}$): 1195-1475 mc. Diameter of fourth whorl (D_4): 1065 mc (N = 1). Radius vector of fifth whorl (Rv_5): 870 mc (N = 1). Length/diameter ratio (L/D): 1.75 (N = 1). The shell in outer whorls is probably oval or short fusiform. In inner whorls the spirotheca in the median area is flattened which has been also observed in the probably related *Fusulinella* cf. *bocki* (→ *Protriticites*) (locality 81197) and *Fusulinella* cf. *asiatica* (locality 81196). Chomata are fairly low and very wide in inner whorls; in outer whorls they are moderately high or high and often get a rounded or quadrangular, symmetrical outline. The wall is rather dark and appears to be homogeneous in the 1st whorl; from the 1½-2½ whorl mural pores can be observed, upper tectorial deposits are absent, and there are one or two layers below the tectum. Traces of a more primitive - i.e. fusulinellid-type - wall are sometimes present up to the 4½ whorl. The single-layered wall (not counting the tectum) has a thickness of about 30 mc in the outer whorl. The maximum thickness of 40-55 mc is usually attained in the two-layered wall of the penultimate whorl.

Locality 81172 (Casasuertes)
Species content: *Ozawainella* cf. *pseudoangulata* Putrya, 1939
Ozawainella krasnokamski kirovi Dalmatskaja, 1961
Ozawainella angulata (Colani, 1924)
Pseudostaffella ex gr. *parasphaeroidea* (Lee & Chen, 1930)
Pseudostaffella ex gr. *rostovzevi* Rauser-Chernoussova, 1951
Beedeina sp.
Fusulina cf. *megaspherica* Sheng, 1958
Protriticites ex gr. *nakahatensis* Ishizaki, 1963

Zonation: Protriticites Zone; near base of the zone.
Correlation: Upper Moscovian; top of Myachkovian; N_2 of Donetz Basin.

The young age of this olistolith as compared to the Panda Limestone is indicated by the presence of *Fusulina* cf. *megaspherica* Sheng, 1958 and *Protriticites* ex gr. *nakahatensis* Ishizaki, 1963.

Fusulina cf. *megaspherica* Sheng, 1958
Plate III, Figs. 5-8

Material: 3 specimens (1 axial, 1 exc. axial and 1 sagittal section)
Number of whorls: 3½-4. Diameter proloculum (D_O): 310-395 mc. Diameter adult ($D_{3½-4}$): 1475-1840 mc. Radius vector of 3½-4th whorl: 836-1165 mc. L/D: 3.12. The test is subrhomboidal up

to the 2½ whorl and elongate fusiform in the succeeding whorls. The low to moderately high and narrow chomata of inner 2½ whorls are replaced by pseudochomata in the 3rd whorl. These secondary deposits have not been observed in the 3½-4th whorl. Axial filling is weak or absent. Septal folding is high and somewhat irregular. The wall is four-layered and has a relatively thin upper tectorium; distinct mural pores appear in the 1½-2nd whorl. The maximum thickness of the wall is generally attained in the penultimate whorl where the tectoria are relatively well developed and amounts to 45-60 mc. The last whorl is usually thinner and may have a thickness of only 17-21 mc.

Fusulina megaspherica differs mainly by its larger L/D ratio. Similar are some species of the group of *Fusulina kamensis* (cf. *F. agujasensis*, *F. kamensis*) which differ in having much shorter inner whorls. These latter species as well as *Fusulina megaspherica* do not show the mural pores of the present form so distinctly and in such an early stage of growth.

Protriticites ex gr. *nakahatensis* Ishizaki, 1963
Plate III, Figs. 9-11

Material: 6 specimens including 2 axial sections.
Number of whorls: average 5, range 4-6 (N = 4). Diameter proloculum (D_o): average 117 mc, range 90-130 mc (N = 4). Diameter of outer (4-6th) whorls: average 1545 mc, range 790-2000 mc (N = 6). Diameter of 4th whorl (D_4): average 1000 mc, range 790-1230 mc (N = 4). Radius vector outer (= 4-5½) whorls: average 740 mc, range 418-985 mc (N = 4). Length/diameter ratio (L/D): 2.4 (5th wh.) (N = 1). The test changes from spherical (-1½ wh.), oval/short fusiform (2-4th wh.), to fusiform (4-5th wh.). In the last whorl the poles can be slightly extended. The first half whorl can be obliquely coiled. Chomata are wide and high or moderately high in inner whorls; in outer whorls they are relatively narrower and of medium height or low. Slope at the tunnel side is steep (-90°). Septal folding starts at the poles of the 4th whorl and remains weak and restricted to the poles in subsequent whorls; more rarely the folding becomes moderately intense. In inner 2-4 whorls the wall is either of Fusulinella-type or differentiation is obscure. The lower tectorium is thin or absent in inner whorls but usually appears in the 2½ whorl; reversely the upper tectorium is relatively better developed in inner whorls but disappears beyond the 2½-3rd whorl. At this stage, i.e. in the 2½-3rd whorl, mural pores are clearly observed. The wall in the adult stage contains two layers with pores, homologous to the diaphanotheca and the lower tectorium. The ultimate whorl usually contains a single porous layer which is the result of merging of the two layers. The maximum thickness of the wall is attained in the 4-5th whorl and amounts to 55-65 mc. The thickness of the protheca increases from 15 mc (4th wh.) to 25-30 mc (4½-6th wh.).

The species belongs to a primitive group of *Protriticites* which may have descended from *Fusulinella pseudobocki*. A related species but with a more primitive wall has been found in the Coriscao Olistostrome (see: *Fusulinella pseudobocki* ⟶ *Protriticites* of locality 81197).

In conclusion we briefly state the main characteristics of the fauna described above (localities 81098, 81197, 81196, 80003, and 81172).

1. The species of *Fusulinella* belong to *F.* ex gr. *bocki* Möller, 1878 and *F.* ex gr. *pseudobocki* Lee & Chen, 1930. They differ in their advanced wall structure from other species of these groups and in this respect resemble *Protriticites*. The wall in inner whorls consists of four layers and is typically fusulinellid; in the outer 1-2 whorls mural pores become distinctly visible and the outer tectorium has disappeared.
2. The species of *Protriticites* belong to primitive species-groups of the genus and in all probability descended from *Fusulinella* ex gr. *bocki* Möller, 1878 and *F.* ex gr. *pseudobocki* Lee & Chen, 1930. They have a rather dark wall in thin section.
3. The species of *Pseudotriticites*? obviously descended from *Beedeina* and are about intermediate between typical *Pseudotriticites* and *Beedeina* of the groups of *B. elegans* (Rauser & Beljaev, 1937) and *B. samarica* (Rauser & Beljaev, 1937).

Faunas with characteristics as enumerated above are well known from strata near the top of the Myachkovian in the U.S.S.R. and from corresponding strata in China and Japan. The listed fauna of the five olistoliths includes species of the genera *Pseudoendothyra*, *Millerella* (very rare), *Schubertella*, *Fusiella*, *Taitzehoella*, *Ozawainella*, *Pseudostaffella* and *Beedeina*. The stratigraphic range of these species does not seriously interfere with the conclusions regarding the age of the olistoliths: these species or related forms can occur in Myachkovian strata. *Ozawainella nikitovkensis* is perhaps more common in the Kasimovian. *Pseudostaffella paraozawai* and *Pseudostaffella compacta* which are common in Kashirian and Podolskian strata may be absent in the Myachkovian and the same may hold for *Ozawainella pseudotingi* and *Ozawainella pseudoangulata*. However, species related to *Ps. compacta* and *Ps. paraozawai* may occur near the Myachkovian/Kasimovian boundary, as follows from the occurrence of *Pseudostaffella ozawai* in the N_1 interval of the Donetz Basin (Putrya, 1940, p. 14). *Ozawainella pseudotingi* is close to *Ozawainella tingi* Lee, 1937, which latter species was found in "transition beds between the Middle and Upper Carboniferous; N_3 zone" of the Donetz Basin (Lee, 1937, p. 78). *Ozawainella pseudoangulata* belongs to the group of *O. pseudoangulata* et *mosquensis* like *Ozawainella* cf. *kumpani* Rauser, 1951, which latter species has been reported by Pasini (1963) from the base of the Auernig sequence (near the boundary of the Myachkovian and the Kasimovian) of the Carnic Alps.

THE FUSULINID ASSEMBLAGE OF THE PANDA LIMESTONE

We went into some detail to prove that the olistoliths may contain younger faunas than the overlying Panda Limestone. However, information on the age of the Panda Limestone in the previous part of this chapter has been brief and the supporting evidence was not described in detail. In the following lines this subject is treated more fully. Fusulinid foraminifera are very common in the Panda Limestone. They have been described of three localities yielding the following species (van Ginkel, 1965: appendix 1):

Locality L 426
Fusulinella pandae van Ginkel, 1965 n. sp.
Hemifusulina ex gr. *moelleri* Rauser-Chernoussova, 1951

Locality L 21 (top of Coriscao Mountain)
Fusulina agujasensis van Ginkel, 1965 n. sp.
Fusulinella ex gr. *bocki* Möller, 1878

Locality L 408
Fusulina rossoschanica Putrya, 1947 subsp. *kamerlingi* van Ginkel, 1965 subsp. nov.

This fauna indicates the Fusulinella Zone, Subzone B, Subdivision B 1 and correlates with the Podolskian (L 426), or Middle to Upper Podolskian (L 21, L 408) (van Ginkel, 1965: appendix 2). The established Podolskian age of the Panda Limestone has been confirmed by recent additional sampling (localities 80039, 80039A, 80042, 80043, 81153A, 81176A, 81194; Encl. I, II, and IV) and subsequent investigation of the fusulinid fauna by the present authors.

Siliciclastic rocks immediately below the base of the Panda Limestone and within a few meters of locality L 21 yielded a macroflora to Dr. J.F. Savage. This flora has been reported to have "affinities to higher Westfalian as well as to Stephanian floras" (Dr. F. Stockmans written comm. in van Ginkel, 1965, p. 211). Recently the first author found a macroflora in two localities (localities 80017, 80031, Encl. I, II, and IV) both situated near Savage's fossil plant locality and L 21. The floras are of "uppermost Westfalian D or Lower Cantabrian age" according to Dr. R.H. Wagner (written comm., 1983). This result supports and is a further refinement of the quoted opinion of Stockmans. This age determination by means of plant fossils does not accord with the Podolskian age for the Panda Limestone. According to existing correlation charts the flora should have been older or reversely the fauna younger. The discrepancy between flora and fauna has been explained at the time by assuming different evolutionary rates for the two fossil groups (van Ginkel, 1965, p. 214). Although this may hold for other cases, in this particular case there is a more appropriate solution to the problem. This solution follows from the discussion in the previous chapters: tectonic movements caused the Podolskian Pandatrave Formation with the Panda Limestone at its base to slide down over the Uppermost Myachkovian-Lower Kasimovian Coriscao Olistostrome. The siliciclastic slab which contains the flora is an olistolith of the Coriscao Olistostrome close to the base of the overlying Panda Limestone.

Some of the earlier conclusions with respect to the age of the Panda Limestone (van Ginkel, 1965) have to be reconsidered. The investigation of the recently collected fusulinid-containing samples from various levels of the Panda Limestone shows that faunas from basal strata indicate the upper part of the B 1 subdivision (= Middle to Upper Podolskian) whereas the middle and upper part of the limestone contain faunas of the lower B 2 subdivision (= Upper Podolskian). Previously it was assumed that the limestone contains only faunas of the B 1 subdivision (van Ginkel, 1965: appendix 2, localities L 408, L 426, L 21). The faunas of L 21 and L 426 are from the base of the limestone and indicate the B 1 subdivision all right, but the fauna of L 408 which contains *Fusulina rossoschanica kamerlingi* is from a higher level and is considered to be in the B 2 subdivision. This accords with the occurrence of *F. rossoschanica* Putrya in the U.S.S.R. Of interest is the fauna of Ge 133 close to the base of the limestone near L 21 (see Encl. I, II, and IV). This fauna contains *Pseudoendothyra* ex gr. *composita* (Dutkevich, 1951) in profusion. The genus *Pseudoendothyra* is very common in Spain in the Profusulinella Zone, Subzone B. The impression is gained that a second optimum in specimen numbers occurs near the top of the B 1 subdivision of the Fusulinella Zone. The Pendueles section at sampling locality A 12-2, probably also in the upper part of the B 1 subdivision, shows that there as well the genus is a common constituent of the fauna (van Ginkel, 1983).

Although the main part of the Panda Limestone is slightly younger than we formerly held it to be, this has no bearing on our proposition that the Panda Limestone rests on strata of younger age, i.e. on the Coriscao Olistostrome. Relevant to our thesis is the age-span of the Panda Limestone as now derived from the fusulinid faunas of 11 localities, representing the entire vertical and lateral range of the limestone member. Fusulinid foraminifera representative of these strata are fully described as well as illustrated from the localities L 21, L 408 and L 426 (van Ginkel, 1965). These faunas should be compared with the faunas derived from olistoliths below or lateral to the Panda Limestone, which are described and illustrated in the present paper. Our conclusion is that the latter faunas are of younger age. This is indeed the main evidence in support of our thesis that the Pandatrave Formation is a nappe and not a part of a regular succession.

LIST OF FOSSIL SAMPLES WITH THE RESULTS OF BIOSTRATIGRAPHIC DETERMINATIONS

Sample localities, rock samples and derived materials (i.e. slides, conodont collections, fossil plant fragments) are identically coded. The code is according to year (1980, 1981) and field station number in the fieldbook. Samples with a different coding are cited from literature or internal reports. The localities are given in Encl. I & II.

Sample material pertaining to fusulinid faunas is stored in the collection of the second author (Rijks Geologisch Museum, Leiden, The Netherlands). Conodont faunas are also stored in the collection of the Rijks Geologisch Museum. The fossil floras are stored in the collection of Dr. R.H. Wagner (Geology department of the University of Sheffield, England).

Upper Carboniferous limestone samples containing fusulinids

Coriscao Olistostrome

Ge 100	Fusulinella Zone; top of Subzone A or base of Subdivision B 1
70515	Fusulinella Zone; Subzone A or base of Subdivision B 1
80003	Protriticites Zone; near base of the zone
80006	Fusulinella Zone; upper part of Subdivision B 1
80009	Fusulinella Zone; Subzone B 1 or B 2
80012	Fusulinella Zone; upper part of Subdivision B 1 or lower part of Subdivision B 2
80013	Fusulinella Zone; upper part of Subzone A or lower part of Subdivision B 1
80014	Fusulinella Zone; base of Subdivision B 1
80015	Fusulinella Zone; base of Subdivision B 1
80016	Fusulinella Zone; top op Subdivision B 1
80025A	Fusulinella Zone; upper part of Subdivision B 1
80025B	Fusulinella Zone; upper part of Subdivision B 1
80025C	Fusulinella Zone; upper part of Subdivision B 1
80026	Fusulinella Zone; upper part of Subdivision B 1
81180	Profusulinella Zone; upper part of Subzone B
81181	Profusulinella Zone; upper part of Subzone B
81182	Profusulinella Zone; upper part of Subzone B
81187	Profusulinella Zone; upper part of Subzone B
81189	Profusulinella Zone; upper part of Subzone B
81196	Protriticites Zone; near base of the zone
81197	Fusulinella Zone; top of Subdivision B 3

Panda Limestone

L 21	Fusulinella Zone; top of Subdivision B 1
L 408	Fusulinella Zone; lower part of Subdivision B 2
L 426	Fusulinella Zone; upper part of Subdivision B 1
Ge 133	Fusulinella Zone; upper part of Subdivision B 1
80039	Fusulinella Zone; lower part of Subdivision B 2
80039A	Fusulinella Zone; lower part of Subdivision B 2
80042	Fusulinella Zone; lower part of Subdivision B 2
80043	Fusulinella Zone; lower part of Subdivision B 2
81153A	Fusulinella Zone; lower part of Subdivision B 2
81176A	Fusulinella Zone; top of Subdivision B 1
81194	Fusulinella Zone; top of Subdivision B 1

Remoña Olistostrome

81095	Fusulinella Zone; upper part of Subdivision B 1
81097	Fusulinella Zone; upper part of Subzone A
81098	Fusulinella Zone; Subdivision B 3
81139	Fusulinella Zone; Subzone A

Olistostrome occurrence S of Casasuertes

81172	Protriticites Zone; near base of the zone

Brañas Olistostrome

II 51	Profusulinella Zone; Subzone A (van Ginkel in Kutterink, 1966)
81103	Fusulinella Zone; upper part of Subdivision B 1 or Subdivision B 2
81104	Fusulinella Zone; upper part of Subdivision B 1
81105	Fusulinella Zone; Subdivision B 1 or B 2
81106A,B,C	Fusulinella Zone; lower part of Subdivision B 2
81107	Fusulinella Zone; upper part of Subzone A or lower part of Subdivision B 1
81108	Fusulinella Zone; upper part of Subdivision B 1
81109A,B	Fusulinella Zone; top of Subdivision B 1
81115	Millerella Zone; Subzone of Pseudostaffella antiqua
81116	Millerella Zone; Subzone of Pseudostaffella antiqua
81121	Millerella Zone; Subzone of Pseudostaffella antiqua

Devonian and Lower Carboniferous nodular limestone samples containing conodonts
Determination by Mr. J.G.M. Raven and Dr. M. van den Boogaard, unless mentioned otherwise:

Remoña Olistostrome

70C172	Frasnian, asymmetrica Zone (Mr. K. Boersma in Maas, 1974)
7130	Upper Famennian (Mr. K. Boersma in Maas, 1974)
81092	Upper Famennian, costatus Zone (Vidrieros Limestone)
81092B	Upper Famennian to Tournaisian, costatus Zone to crenulata Zone (probably Vidrieros Formation)
81100	latest Viséan (Villabellaco Limestone)
81129B	Givetian to Frasnian
81138	middle to upper part of Viséan, possibly earliest Namurian (Villabellaco Limestone)
81160	Famennian, velifer Zone to costatus Zone (Vidrieros Limestone)

Brañas Olistostrome

81123	uppermost Viséan (Villabellaco Limestone)

Upper Carboniferous sandstone/shale samples containing plant fossils
Determination by Dr. R.H. Wagner:

80017	floral remains occurring mainly on the sand-shale interfaces in microripple laminated sandstone in a very large coherent rock slab in the upper part of the Coriscao Olistostrome: *Neuropteris* sp. indet.; *Linopteris* sp. indet.; *Calamites suckowi*; *Licophytes* sp. (?). Age indication: no distinct age difference results from a comparison with the flora of sample 80031.
80031	floral remains occurring in a mudstone sample from a very large rock slab in the upper part of the Coriscao Olistostrome just below the Panda Limestone on the summit of the Pico Coriscao; the olistolith is probably the same as the one containing 80017, but divided by the fault that passes the summit of P. Coriscao: *Sphenophyllum* cf. *emarginatum* Brongniart; *Lobatopteris vestita* (Lesquereux) Wagner (i.e. vestita Zone); *Linopteris palentina* Wagner; *Pecopteris* sp. indet. Age indication: uppermost Westfalian D to Lower Cantabrian (A flora from this locality was collected by Dr. J.F. Savage and determined by Dr. F. Stockmans as having "affinities to higher Westfalian as well as to Stephanian floras", Stockmans cited in van Ginkel (1965) and in Maas (1974).

Nodular limestone sample containing goniatites
Collected by Mr. J.A. Kutterink, determination by Dr. J. Kullmann:

II 105	upper part of Viséan (Kutterink, 1966)

CHAPTER IV

SYNTHESIS

PALAEOGEOLOGIC RECONSTRUCTIONS

At the end of our preliminary conclusions (Chapter II), we suggested a system of faults controlling the northern basin slope of the developing Valdeón basin. Such fault control can be argued both for the Valdeón and for the Liébana basin (cf. Maas, 1974). In Liébana the prolific occurrence of olistostrome and slump deposits and notably the formation of large gravitational flap folds, have been explained by basement faulting (Maas, op. cit.). In the Valdeón area the even more profuse olistostrome sedimentation, the downsliding of synsedimentary nappes, and the velocity and magnitude of the relative basin subsidence (more than five kilometres in maybe no more than one million years), favour a fault-control of the basin slope.

We endeavoured to extend our hypothesis of basin filling by fault-controlled resedimentation and shallow gravitational slide and collapse tectonics, by including in it the formation of the Picos de Europa nappes. The extended hypothesis is visualized in Encl. Va, b, and summarized in Fig. 2. The controlling fault system proposed, is rendered as a series of normal faults dipping toward the basin; fault activitiy is progressively stepping back from the basin toward the shelf area. The early Westfalian fault movements (Encl. Va) caused resedimentation and flapfolding, mainly placing contemporaneous basin sediment upon basin sediment (Liébana basin, cf. Maas, 1974: fig. 4). The late Westfalian fault movements (Encl. Va) caused gravitational collapse folds, olistostrome sedimentation and the deposition of coarse-grained submarine fan sediments (Porrera and Viorna conglomerate members - Maas, op. cit.; Rupke, 1977). The early Middle Cantabrian fault movements (Encl. Vb) caused large scale resedimentation and the downsliding of the Panda and Montó nappes (Chapter II of this paper), bringing early

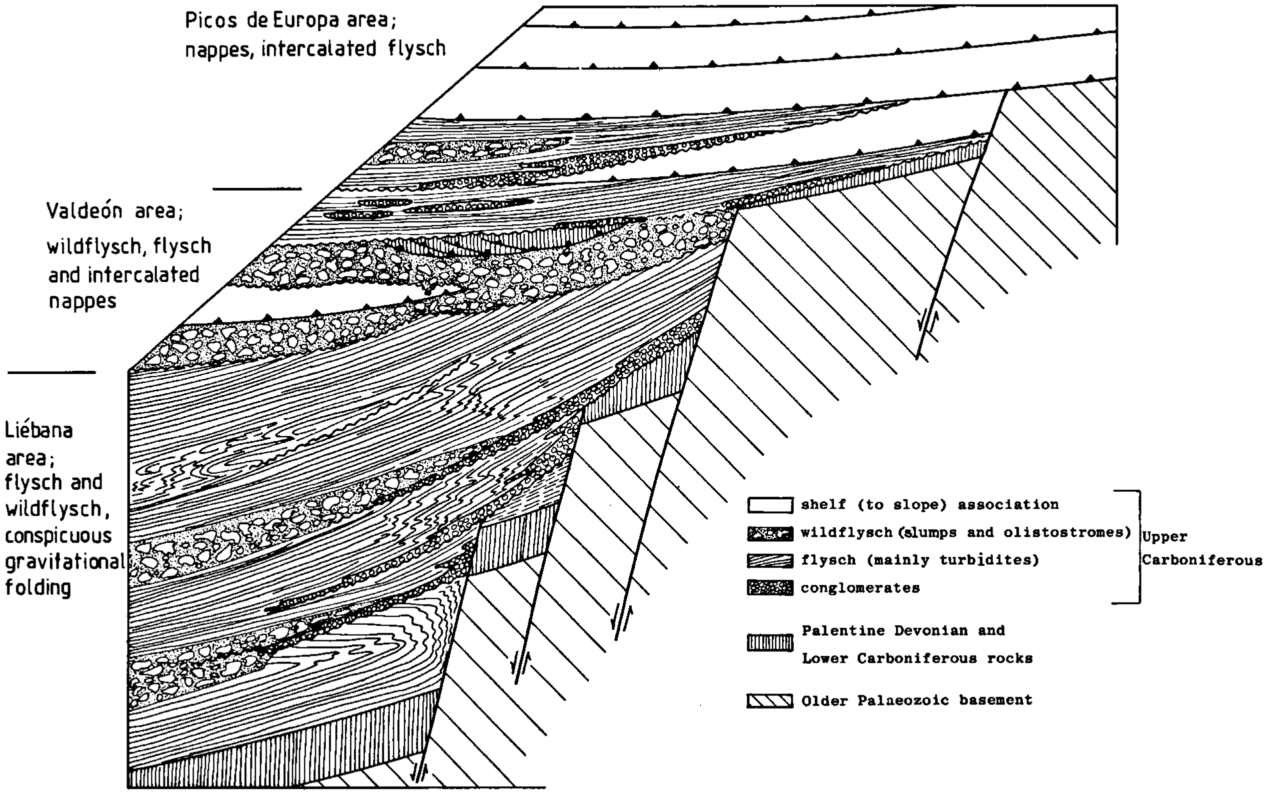

Fig. 2. Simplified geology of Liébana, Valdeón and the Picos de Europa

Middle Cantabrian to Middle Devonian slope and shelf rocks into the basin. The late Middle Cantabrian fault movements caused the sliding into the basin of a large part of the shelf rocks: the Picos de Europa nappes. The limestone sequence of the first Picos de Europa nappe was - before or subsequent to the downsliding - in part eroded and locally covered by a residual conglomerate. After, or simultaneously with, the downsliding it subsided to a depth that permitted the deposition of about 800 m of mainly calciclastic flysch (the Lebeña Formation). The following four Picos de Europa nappes lie straight on top of one another, leaving no direct indications for their depths of emplacement. The hypothetic shallow transport along the earth surface of these latter nappes can only be argued from circumstantial evidence.

STRUCTURAL GEOLOGIC CONSIDERATIONS

The faults drawn in Encl. V and in Fig. 2 are normal faults characterized by a shallow dip of the originally horizontal strata toward the downthrow. Such a configuration is possible but the common normal fault causes the strata to dip away from the downthrow. H. Cloos (1930) did several normal fault experiments with a clay cake on a rubber base. The "common" normal fault was produced by inflating air under the rubber base; the resulting updoming caused a tension in the arched clay cake producing the configuration of normal faults and outward dipping strata. In another experiment the rubber base was not arched but stretched in the horizontal plane, this time the resulting normal faults were accompanied by a shallow dip of the strata, toward the downthrow, i.e. the type of normal fault depicted in our figures. In our case the initiation of the nappe sliding would be favoured by a dip of the strata toward the basin. With a reverse dip the nappe formation would be impossible without a lateral tectonic force.

The type of faults depicted in our figures are consistent with a tensional stress that did not develop subsidiary to regional uplift. Large scale regional uplift is usually a thermal updoming concomittant with either magmatism, metamorphism or both. The sedimentary fill of the resulting fault-bounded basins, in an ensialic environment, is continental or shallow marine. Considering our case, the rate of basin subsidence, the deep submarine facies of the sediments, and the near-absence of extensive metamorphism and magmatism in the Variscan orogeny of the Cantabrian Mountains (cf. Savage, 1979), are all facts consistent with the fault type depicted in our figures.

REGIONAL GEOLOGIC SETTING

The visualization of our hypothesis given in Encl. V and in Fig. 2 is, of course, a simplification. It is important to realize that no stable basin configurations existed during the Upper Carboniferous in the Cantabrian area. In Namurian to early Westfalian times, the Liébana basin, probably much larger than the part exposed at present, seems to have deepened in a general sense to E (cf. Maas, 1974). The basin configuration during late Westfalian time is problematic, thick sediment wedges terminate abruptly against faults, localized compressional (?) uplift occurred along a narrow and faulted anticlinal structure (the Mid-Liébana ridge - Maas, op. cit.). In Middle Cantabrian times the Valdeón basin developed unconformably on the western, relatively shallow part of the Liébana basin. The structures depicted in Encl. Vb below the basal unconformity of the Valdeón Formation, are fictitious, but similar structures do occur, and probably already existed at that time, in the Mid-Liébana ridge, not too far to E of the Valdeón area. The centre of subsidence was shifted to W, as indicated e.g. by the westward thickening of the Valdeón turbidite sequence. During the downsliding of the Picos de Europa nappes the centre of subsidence had shifted again to E, as indicated by the eastward thickening of the Lebeña Formation. The Picos de Europa nappes rest "unconformably" on both the Valdeón and the Liébana deposits.

The normal-fault system activating the basinward sedimentary and shallow tectonic transport, had a roughly W-E strike for the entire interval. However, the general strike seems to have shifted from WNW-ESE during the filling of the Liébana basin, to WSW-ENE during the filling of the Valdeón basin, and back to W-E during the downsliding of the Picos de Europa nappes. Not too far to SE, in the Pisuerga area, adjacent to Liébana, the entire sedimentary sequence of Namurian to Upper Cantabrian is preserved as a continuous sequence with at most a number of (sub)parallel nonconformities, representing time intervals of non-deposition. The synsedimentary tectonical activity described above apparently was of a local nature, being restricted to the Valdeón-Liébana region without affecting the regions nearby.

The general development of the Liébana-Valdeón basin-to-slope area is characterized by: a deviant near-slope turbidite facies (Chapter II of this paper); prolific slump and olistostrome deposits; rapid to very rapid deposition in a fault-controlled tensional basin; submarine denudation of the slope area; contemporaneous local uplift along a narrow compressional (?) structure making an oblique angle with the general trend of the basin; shifting transport directions for the mass-transported sediments and the downsliding rock masses; and unconformities and gravitational tectonics of only local, or at most regional extent. A notable negative characteristic is the absence of any extensive magmatism or metamorphism.

Summarized as above the Upper Carboniferous development of the Liébana-Valdeón area does not fit the classic orogenetic cycle. It does, however, agree with basin development in an oblique-slip mobile zone (cf. Ballance & Reading, 1980). The characteristics of basin formation in that structural setting are described by Reading (1980) and can be summarized as follows: Sedimentary basins governed by a strike-slip fault system are elongate parallel to the strike-slip system (1); they are rather small, 10 x 20 km to 30 x 100 km (2), often asymmetric in cross section (3) and usually very deep in relation to their width (4). There is a rapid accumulation of thick (several km) piles of sediments (5), sedimentation centres shift laterally in time (6), facies units are thick and of restricted lateral extent (7). In marine basins with an important continental source the facies is (hemi-)terrigenous, coarser clastics are transported by subaquous mass-gravity processes (8). There is evidence of a nearby active source area (9). There is evidence of rapid vertical movements indicative of tensional faulting (10). There are localized zones of compression (11). Uplift and erosion expressed in strong angular unconformities are contemporaneous with sedimentation nearby (12). Regional metamorphism is normally absent or low-grade (13), igneous rocks are generally absent or sparse (14). And for the general geologic setting: the classical geosynclinal cycle, governed by a subduction process, does not apply (15). The summarized particulars of the Liébana-Valdeón area correspond with the large majority of the criteria enumerated. As for the fifteenth criterion, Savage stated that for the Cantabrian area the concept of a geosyncline is better replaced by "a mosaic of loosely connected depositional basins, each possessing its own source area" (Savage, 1979, citing Davis & Ehrlich, 1974, and applying their model of the Appalachians to the Cantabrian area). The strike-slip model was proposed for the Cantabrian area at large by Reading (1975) and applied in detail by Heward & Reading (1980). In the next chapter we will discuss the possible implications of the strike-slip model for the studied area and the surrounding regions.

CHAPTER V

DISCUSSION

Savage (1979) stated that some of the Upper Carboniferous wildflysch deposits (i.e. slumps and olistostromes) in the Cantabrian area, can be related to the advance of larger and smaller nappe sheets and that the smaller nappes were almost certainly gliding under gravity. He suspected that this would prove to be so for the larger nappes too. Connecting nappe transport with wildflysch deposition, a transport of the nappes virtually at the surface of the earthcrust was tacitly assumed. The study reported in the present paper demonstrates that assumption to be correct, as far as our field data permit such demonstration. Of the two most evident surficial slide-structures, the Montó and Panda nappes, the Panda nappe has about the size of the Esla nappe (Rupke, 1965), that is about 10 x 20 km. The Picos de Europa nappes are much larger, their surficial transport has not been established beyond doubt. However, the contemporaneous sedimentation of the Lebeña Formation, the fact that the Picos de Europa nappes developed as flat sheets, not related to any folding (Savage, op. cit.), and the emplacement of these nappes in or upon a basin up to then subsiding, are all facts that favour a surficial sliding of these nappes too.

Now as for the large-scale tectonic setting of these surficial slide-structures, we have already mentioned the concepts of Heward & Reading (1980). Heward & Reading, citing Arthaud & Matte (1977), pointed out that the Hercynian orogeny of the Cantabrian Mountains, during Late Carboniferous times, developed within an enormous dextral shear zone, separating the American-European plate from the African plate; the entire Iberian peninsula was then situated within that shear zone (Fig. 3). Within that continental shear zone, a number of dextral en-echelon strike-slip faults (Riedel shears, cf. Tchalenko, 1970), fundamental faults in their own right, can be distinguished. One of them is the Biscayan-N Pyrenean fault. The León line (Fig. 1) can be interpreted as a subsidiary to that fault. In that setting, the Cardaño line (Fig. 1), according to its orientation, should be interpreted as a conjugate Riedel shear, with a sinistral offset. The large WNW-ESE faults in Liébana (San Carlos fault and Redondo fault - Fig. 1) can be interpreted as subsidiaries to the Cardaño line.

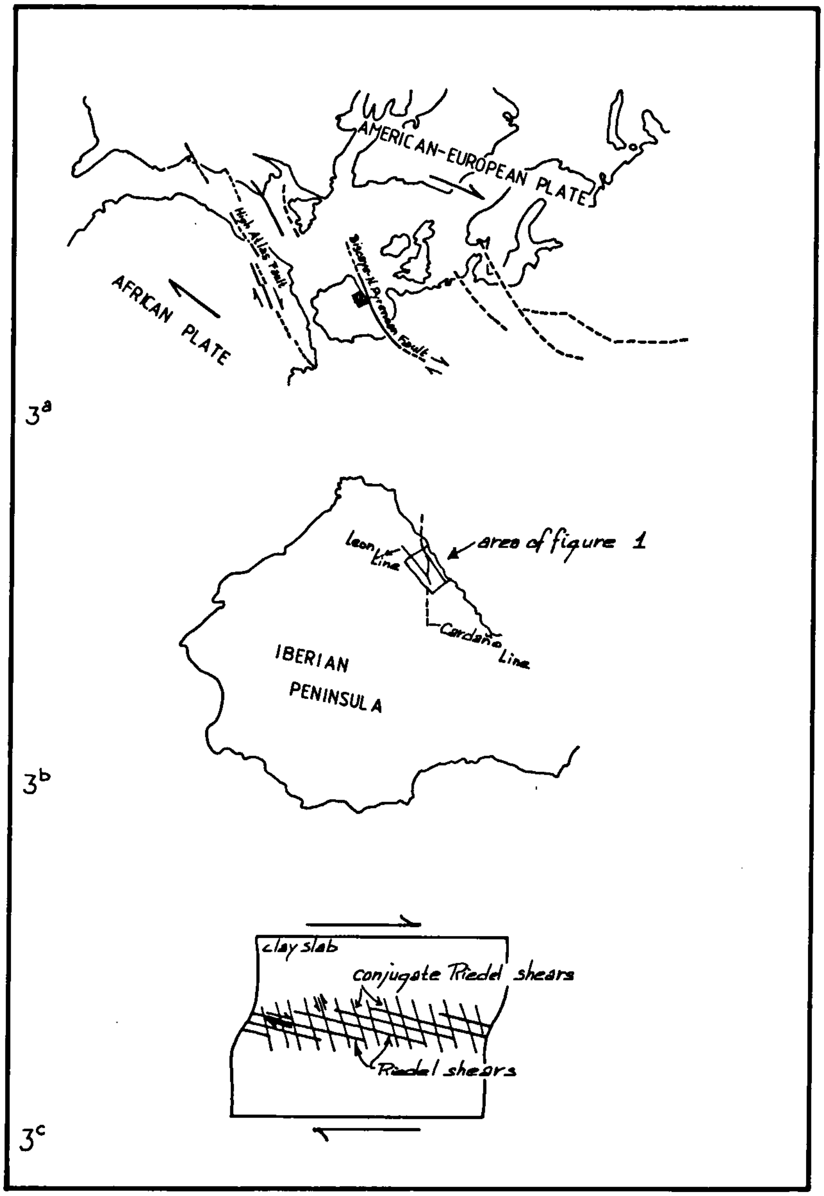

Fig. 3. Late Hercynian palaeotectonic position of the Iberian peninsula in an intercontinental dextral shear zone (3^a, after Arthaud & Matte, 1977). The orientation of the León and Cardaño line in that shear zone (3^b). The orientation of the Riedel shears and conjugate Riedel shears in a shear experiment (3^c, after Tchalenko, 1970).

The model proposed by Heward & Reading (op. cit.) can be criticized. Savage (oral comm.) draws attention to the fact that the León line and Cardaño line are deep fundamental faults whereas Riedel shears are faults of a shallow nature, having developed above a strike-slip fault. However, it should be possible that fundamental faults, when temporarily emplaced within a plate-boundary shear zone, temporarily act as Riedel shears if they have approximately fitting orientations. Another debatable aspect of the Heward & Reading model is the fact that the off-sets one expects along the León and Cardaño lines acting as Riedel shears, i.e. a dextral off-set along the León line (the Riedel shear) and a sinistral off-set along the Cardaño line (the conjugate Riedel shear), have never been established. Along the León line no horizontal slip has been measured up to now; the horizontal slip measured along the Cardaño line is a dextral slip (Sjerp, 1967). We will, however, accept the model offered by Heward & Reading (op. cit.) as it is and see how the Upper Carboniferous development of Liébana and Valdeón fits in it.

During the Namurian-Westfalian development of the Liébana basin, a stepped series of (trans-?)tensional faults (Harland, 1971), oriented parallel to the Cardaño line, probably controlled the northern basin slope. Although the Liébana basin, especially during the Namurian, may have been fairly large (40 x 100 km seems a realistic estimate), the typical near-slope facies of the turbidites, the frequent occurrence of slumps and olistostromes, the shallow gravitational collapse structures and the number of unconformities, mostly of no more than local importance (Maas, 1974), all suggest syntectonic sedimentation in an areally restricted basin. The Cardaño line was probably the active southern boundary of that basin (cf. Reading, 1975). The Iberian peninsula probably had not yet reached its position in the geotectonic setting described above (cf. Scotese et al., 1979), but the Cardaño line is an old fundamental feature, active at least since the Middle Devonian.

During the development of the Valdeón basin, a much smaller basin (about 20 x 40 km), the Iberian peninsula was situated in the dextral continental shear zone described above. Sinistral transport along the Cardaño line, the southern boundary of the basin was probably transpressional in a W over E sense. In the existing stress field WSW-ENE striking normal faults developed, delimiting the Valdeón basin to the north and making a sharp angle with the Cardaño line. Tensional movements along these faults triggered the mass transport of the olistostromes and the downsliding of rock masses which retained their internal coherence: the Panda and Montó nappes. West of the Cardaño line the Asturian basin area (Fig. 1) was compressed W over E and uplifted; W over E low-angle thrust structures developed terminating listrically in a decollement surface, the Cambrian Lancara Formation (cf. Julivert, 1971). A part of the easternmost thrust structure crossed the Cardaño line and slid into the Valdeón basin, forming the Neoncito nappe.

The tectonic transport from N and W was maintained during the final stages of the Hercynian continental shearing. The Picos de Europa nappes slid into the basin from the north, possibly triggered by transtensional dextral fault movements along the fault system now separating the Picos de Europa from the Coastal Ranges (Fig. 1). Later, the thrust structures developed in the Asturian basin area, the Ponga nappes (Fig. 1), transgressed the Cardaño line from the W. Their transport probably resulted from the same W over E stress field that caused the sinistral W over E transpression along the Cardaño line. The folding along E-W axes of both sets of nappes (Julivert & Marcos, 1973; Maas, 1974) is a late Stephanian or even younger event, occurring in an entirely different tectonic setting. The dextral transport along the Cardaño line (about 5 km, Sjerp, 1967) may have occurred at that time, the dextral slip fits a N-S compression.

The development described above is the hypothetic geologic development obtained when the facts known from the studied area and from the surrounding regions are fitted in the oblique-slip model proposed by Heward & Reading. All things considered, their model offers an acceptable explanation for the salient characteristics of the geology of the eastern half of the Cantabrian Mountains. The importance of decollement processes in the structural history of the Cantabrian area, resulting in an allochthonous position for most structural rock units, and the fundamental role of the large faults (cf. de Sitter, 1965; Savage, 1979) is stressed by the results of our investigations. If our hypotheses are correct we must assume that very large tracts of Upper Carboniferous sedimentary rocks in the Cantabrian Mountains are now buried under nappe structures. It seems beyond doubt that a large part of the Upper Carboniferous shelf/slope sediments, originally situated between the Picos de Europa shelf and the Liébana basin, are now covered by the Picos de Europa nappes. The possibility of accumulation of fossil hydrocarbons in the Cantabrian area in Upper Carboniferous rock sequences now buried under nappes, should not be excluded.

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