Journal of Structural Geology, Vol. 8, Nos. 3/4, pp. 415 to 430, 1986
Printed in Great Britain

0191–8141/86 $03.00 + 0.00

Structural, metamorphic and magmatic history of the Mondoñedo nappe (Hercynian belt, NW Spain)

F. BASTIDA*, J. R. MARTINEZ-CATALAN† and J. A. PULGAR*

* Departamento de Geotectónica, Facultad de Geología, Universidad de Oviedo, Spain

† Departamento de Geotectónica, Facultad de Ciencias, Universidad de Salamanca, Spain

(*Received* 1 *July* 1984; *accepted in revised form* 26 *September* 1985)

Abstract—The Mondoñedo nappe is one of the main units in the internal zone of the Hercynian belt in the Iberian Peninsula. It consists of low to medium grade metasediments, late Precambrian and Palaeozoic in age, folded by large D_1 E-verging recumbent structures, and thrusted several tens of km towards the east during a D_2 deformation episode. At the base of the thrust sheet, a shear zone of up to 3 km thick was developed. Granitoid bodies, which were intruded after D_1 and before or during D_2, were deformed in this basal zone.

The geometry of D_1 and D_2 structures and that of the isograds are described, as well as the relationships between metamorphism and deformation. A structural evolution is inferred which shows an episode of generalized ductile deformation (D_1) followed by another in which the deformation tended to concentrate in a subhorizontal ductile shear zone (D_2) and culminated with the thrusting along an important fracture. The shear zone originated in a thermally softened level which, according to metamorphic data, was at temperatures in the range 550–600°C.

The structural, metamorphic and magmatic features of the Mondoñedo nappe suggest a relationship with a crustal scale overthrusting of the continental mass to which the nappe belongs. This could be a consequence of the collision against another continental mass to the west.

Résumé—La nappe de Mondoñedo est une des unités principales des zones internes de la chaîne hercynienne de la Péninsule Ibérique. Elle est composée par des metasédiments du Précambrien Supérieur et du Paléozoïque, affectés par des plis couchés (D_1). L'ensemble est chevauché vers l'Est sur plusieurs km lors d'une deuxième déformation (D_2). Dans la partie inferieure de la nappe, l'on trouve une zone de cisaillement de 3 km d'épaisseur. Des corps granitiques intrusifs dans cette zone ont été déformés pendant la deuxième phase.

On décrit la géométrie des structures et celle des isogrades, et on établit des relations entre le métamorphisme et la déformation. On peut déduire une évolution structurale qui aurait commencé par un épisode à comportement ductile généralisé (D_1), suivi par un autre à déformation concentrée dans une zone de cisaillement ductile subhorizontale (D_2) qui a fini par chevaucher au long d'une fracture importante. La zone de cisaillement basale a été développée, d'après les données sur le métamorphisme, à une temperature de l'ordre de 550–600°C.

Les caractéristiques tectoniques, métamorphiques et magmatiques de la nappe de Mondoñedo suggérent qu'elle est en relation avec un cisaillement profond intracontinental de la plaque a la quelle appartient la nappe qui peut bien être une conséquence de sa collision avec une autre masse continentale située a l'Ouest.

INTRODUCTION

THE HERCYNIAN belt of NW Spain comprises several superimposed allochthonous units, some of which have undergone displacements of tens or hundreds of kilometers. The town of Mondoñedo, to the north of the province of Lugo, gives its name (Marcos 1971) to one of the largest of these in the internal zone, a thrust sheet whose eastern limit is a thrust fault or imbricate stack extending over some 165 km along strike (Fig. 1). To the west, the unit is cut by the Vivero fault (Parga Pondal *et al.* 1967, Matte 1968), a N–S striking fault downthrown to the west. Its throw, comprising that of the fault itself and that of an associated ductile shear zone, is about 10 km, as may be inferred from stratigraphic, structural and petrological data (Martínez-Catalán 1981). The thrust sheet was folded by two systems of large open antiforms and synforms whose hinges and axial surfaces are more or less perpendicular, producing type 1 interference patterns (Ramsay 1967) to the west of the unit. Three principal domes were formed in this area. In the two northern ones, the base of the nappe and the footwall rocks (Gistral and Monte Carballosa tectonic windows) may be seen, while in the southernmost one, known as the Puertomarín dome, only the basal zone of the nappe crops out, the thrust fault not having been exposed by erosion.

Because of its dimensions and outcrop conditions, the Mondoñedo nappe offers particular facilities for the study of the evolution of a large allochthonous structure in the internal zone of an orogenic belt, including (i) its geometry, (ii) the temporal relationship between the different structures related to the nappe and thrust tectonics, (iii) the deformation mechanisms of the rocks, (iv) the relationships between the development of the structures, metamorphism and plutonism and (v) the link between the history of a unit of this kind and the general evolution of the orogen.

REGIONAL SETTING

The Hercynian belt of the NW Iberian Peninsula has been divided into three zones with different geological

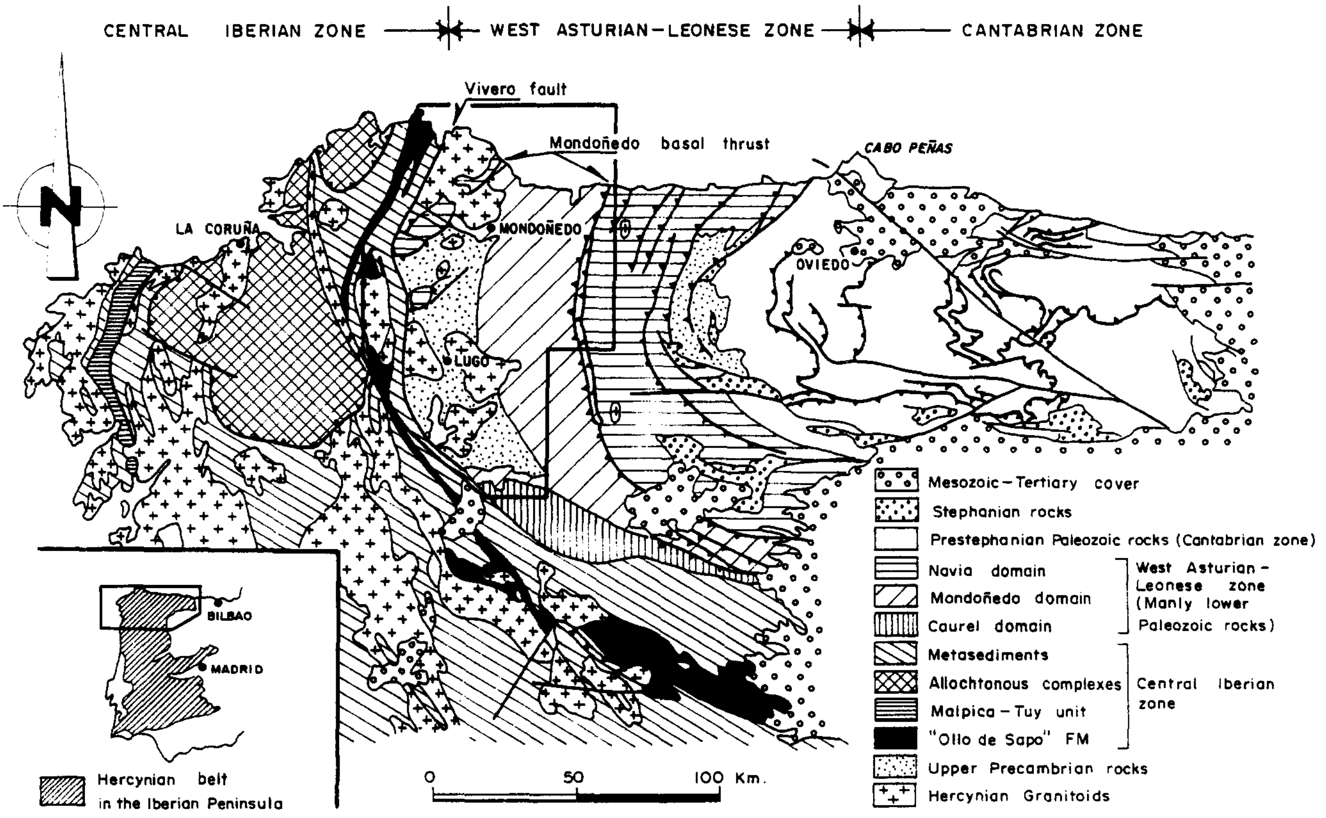

Fig. 1. Geological sketch-map of the NW Iberian Peninsula.

characteristics: the Cantabrian zone (CZ), the West Asturian–Leonese zone (WALZ) and the Central Iberian zone (CIZ) (Julivert *et al.* 1972). The three zones form concentric arcuate bands subparallel to the direction of the peculiar virgation of this sector of the Hercynian belt of Western Europe (Fig. 1).

The outermost of these, the CZ, is made up of an unmetamorphosed sedimentary succession in which two parts may be distinguished: (i) a pre-Carboniferous and pre-orogenic Palaeozoic sequence with clastic and carbonate sediments of shallow-water platform facies resting unconformably on turbidite facies deposits (the Narcea slates) attributed to the Upper Proterozoic and (ii) a Carboniferous synorogenic sequence, principally clastic in the Upper Carboniferous and mainly composed of paralic sediments (Julivert *et al.* 1980, Marcos & Pulgar 1982). The Hercynian orogeny gave rise to superficial thrust sheets with practically no internal deformation, and to two systems of parallel folds with steep axial surfaces which intersect to produce type 1 interference patterns (Julivert 1971, Julivert & Marcos 1973).

The metasedimentary materials making up the WALZ are pre-orogenic and consist of a thick sequence covering the whole of the Cambrian and Ordovician systems, a good deal of the Silurian and, locally, the Lower Devonian. They rest unconformably on Upper Proterozoic terrigenous sediments with turbidite facies, the Villalba Series, equivalent to those of the same age of the CZ. In general, the Palaeozoic sediments correspond to shallow-water facies with a noteworthy exception in the Upper Ordovician: the Agüeira formation, which exhibits turbidite facies (Marcos 1970, 1973, Crimes *et al.* 1974). According to the stratigraphic peculiarities, three domains may be differentiated in the WALZ: those of Navia, Mondoñedo and Caurel (Fig. 1). The stratigraphic successions corresponding to each of these are shown in Fig. 2.

Structurally, the WALZ was affected by three deformation phases. The first (D_1) gave rise to large recumbent folds (with greater amplitudes in the domains of Mondoñedo and Caurel than in that of Navia), and a generalized slaty cleavage or schistosity (S_1). The second (D_2) was responsible for the appearance of thrust-type structures and associated shear zones. The largest of the thrust sheets is the Mondoñedo nappe, which thrusted the domain of the same name over that of Navia towards the east. Minor folds, frequently non-cylindrical, sometimes sheath-like, were generated, as well as a new schistosity or a crenulation cleavage (S_2). The third (D_3) produced large open folds, approximately homoaxial with the first ones, and a crenulation cleavage (S_3). Locally, there is also a system of transverse folds which, superimposed on the earlier D_3 folds, produced type 1 interference patterns of Ramsay (1967). Metamorphism is of low grade in most of the zone, though in the Mondoñedo nappe it reached medium grade and, locally, attained high grade in the footwall rocks.

From a stratigraphic point of view, the CIZ in Galicia (NW Spain) and N Portugal is characterized by a lack of dated Cambrian sediments, by the unconformity which separates the Lower Ordovician from its substrate and by the existence of a very thick Silurian sequence which is mainly pelitic. In its eastern part may be found the Ollo de Sapo anticlinorium (Fig. 1). This is a complex structure within whose core the Ollo de Sapo Formation

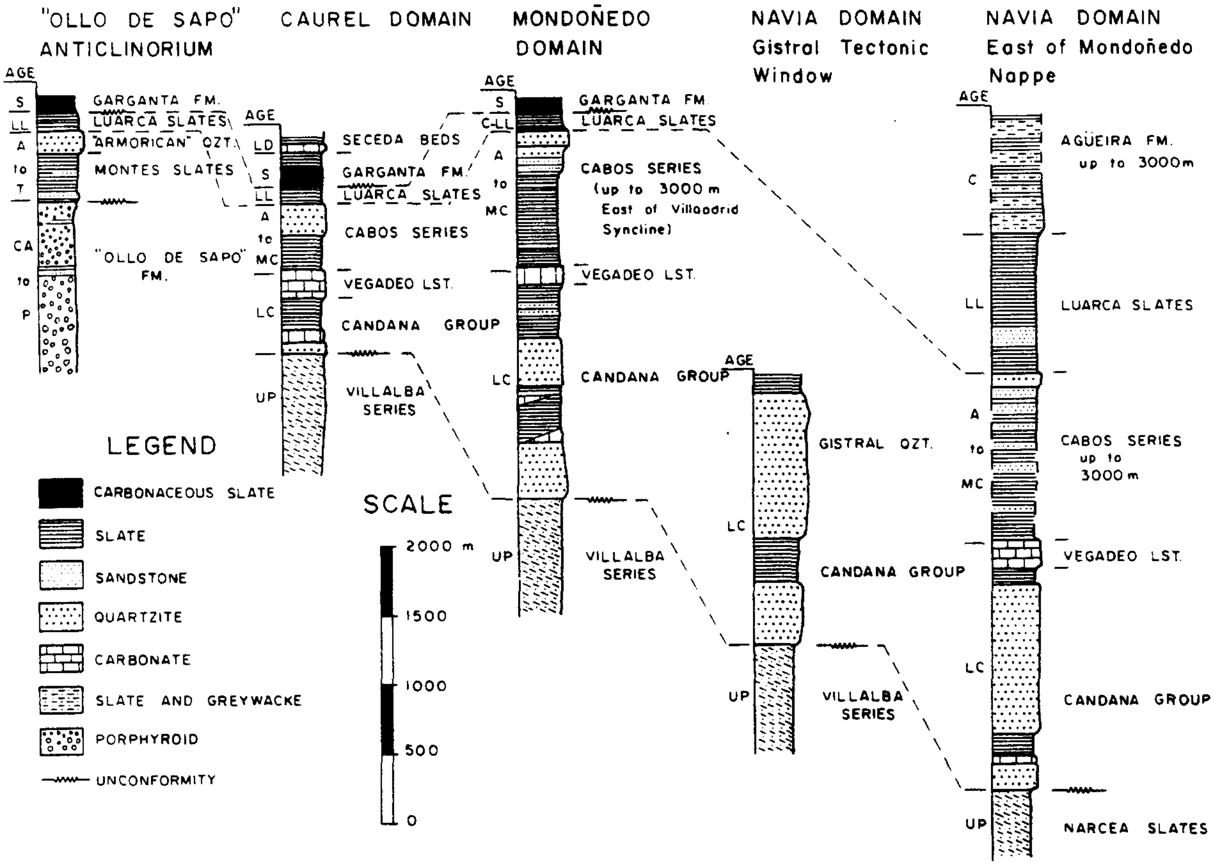

Fig. 2. Summary of the stratigraphy of the West Asturian–Leonese zone (WALZ) and the Ollo de Sapo anticlinorium (Central Iberian zone, CIZ). Ages: LD, Lower Devonian; S, Silurian; C, Caradocian; LL, Llandeilian and Llanvirnian; A, Arenigian; T, Tremadocian; CA, Cambrian; MC, Middle Cambrian; LC, Lower Cambrian; UP, Upper Proterozoic; P, Proterozoic.

(Parga Pondal *et al.* 1964), a porphyroid of supposed Proterozoic to Early Cambrian age (Fig. 2), crops out. This has been interpreted either as a sedimentary sequence, stemming from the erosion of an old granitic and gneissic basement (Parga Pondal *et al.* 1964) or a granitic and volcanic one (Bard *et al.* 1972), or else as a volcano-detrital complex with some sub-volcanic granitic bodies (Navidad 1978, González Lodeiro 1980). Even if the latter interpretation is correct, it is highly likely that the Ollo de Sapo Formation lies on a granitic or gneissic basement and even that its base forms part of this basement.

One peculiarity of the CIZ of the NW Iberian Peninsula is the existence of allochthonous complexes containing mafic and ultramafic rocks (Fig. 1), sometimes with high-grade eclogite or granulite facies metamorphism. Such complexes are composed of various stacked tectonic units, some of which are ophiolitic in nature (Bayer & Matte 1979, Iglesias *et al.* 1981, González Lodeiro *et al.* 1982, Bastida *et al.* 1983, Martínez-Catalán *et al.* in press).

The deformation phases in this part of the CIZ are comparable to those of the WALZ, though they exhibit peculiar characteristics. In general, metamorphism is of medium to high grade, but locally it may be of low grade.

Granitic rocks are abundant in NW Spain, especially in the CIZ west of the Ollo de Sapo anticlinorium, where they crop out over more than half of the area. They are also abundant in the Mondoñedo nappe (Fig. 1), but are scarcer in the rest of the WALZ and are practically non-existent in the CZ. They may be grouped into three large sets: synkinematic biotite granitoids, two-mica granites and postkinematic granitoids. The synkinematic biotite granitoids form part of the calcalkaline series described by Capdevila (1969) and Capdevila & Floor (1970). They are commonly intrusive granodiorites with no apparent relationship to the progressive regional metamorphism of the sector, which reflect crystallization in a relatively dry environment, and which originated as granitic melts at greater depths than those of the second group. The two-mica granites are granites in a strict sense, with evolutionary trends that generally agree with those of hydrated granitic magmas, and which are related to Hercynian metamorphism and regional migmatization (Barrera *et al.* 1982). Finally, the postkinematic granitoids, attributed by Capdevila (1969) and Capdevila & Floor (1970) to the calcalkaline series, are in fact a fairly heterogeneous group whose principal common characteristic is that their emplacement took place after the third Hercynian phase (D_3).

STRUCTURE

For the description and interpretation of the structure of the Mondoñedo nappe, we will consider the following

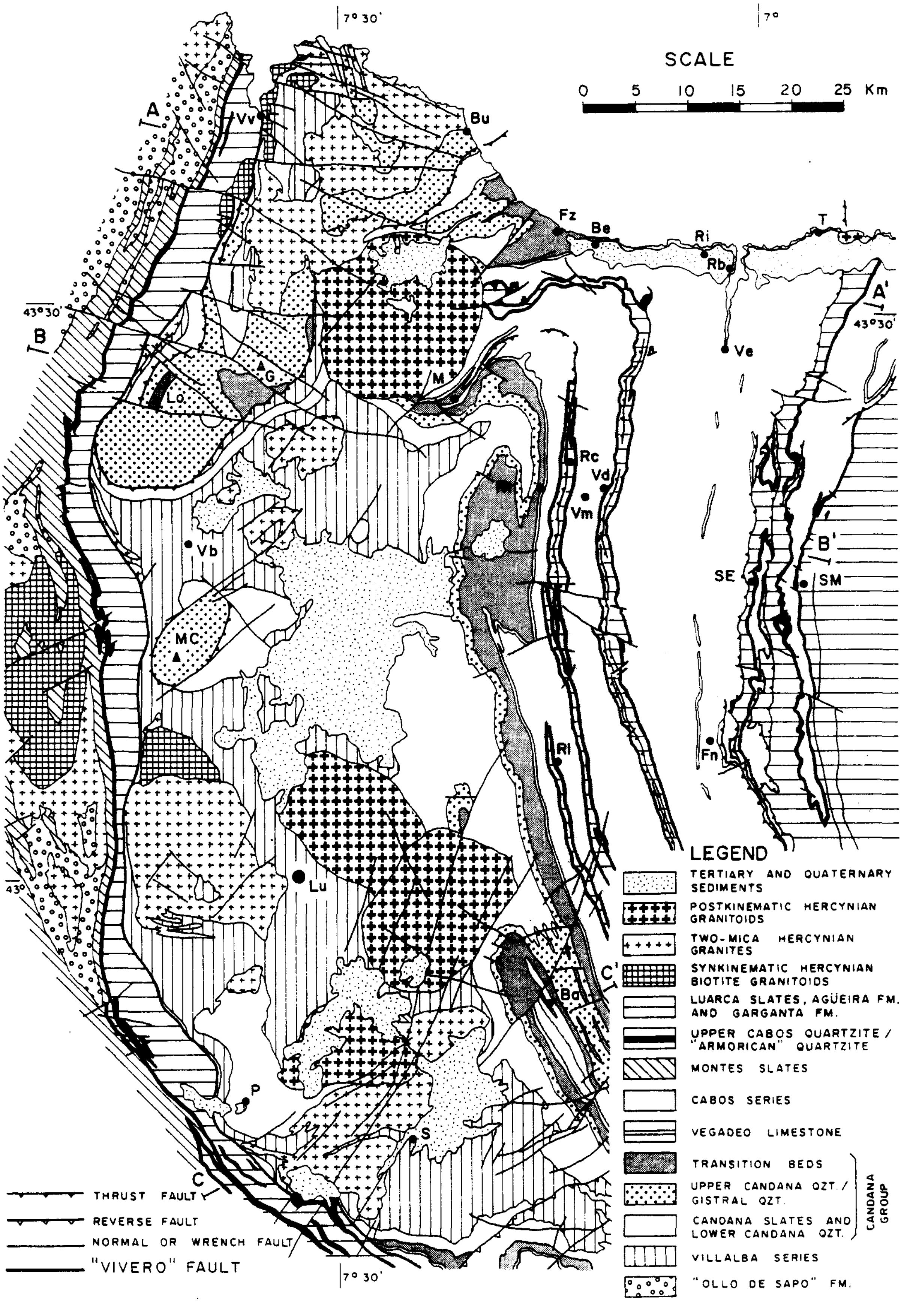

Fig. 3. Simplified geological map of the northern part of the Mondoñedo nappe and adjacent areas. Highs: G, Gistral; MC, Monte Carballosa. Localities: Ba, Baralla; Be, Benquerencia; Bu, Burela; Fn, Fonsagrada; Fz, Foz; Lo, Losada; Lu, Lugo; M, Mondoñedo; P, Puertomarín; Rb, Ribadeo; Rc, Rececende; Ri, Rio; Rl, Real; Rt, Riotorto; S, Sarria; SE, Santa Eulalia de Oscos; SM, San Martín de Oscos; T, Tapia de Casariego; Vb, Villalba; Vd, Villaodrid; Ve, Vegadeo; Vm, Villamea; Vv, Vivero.

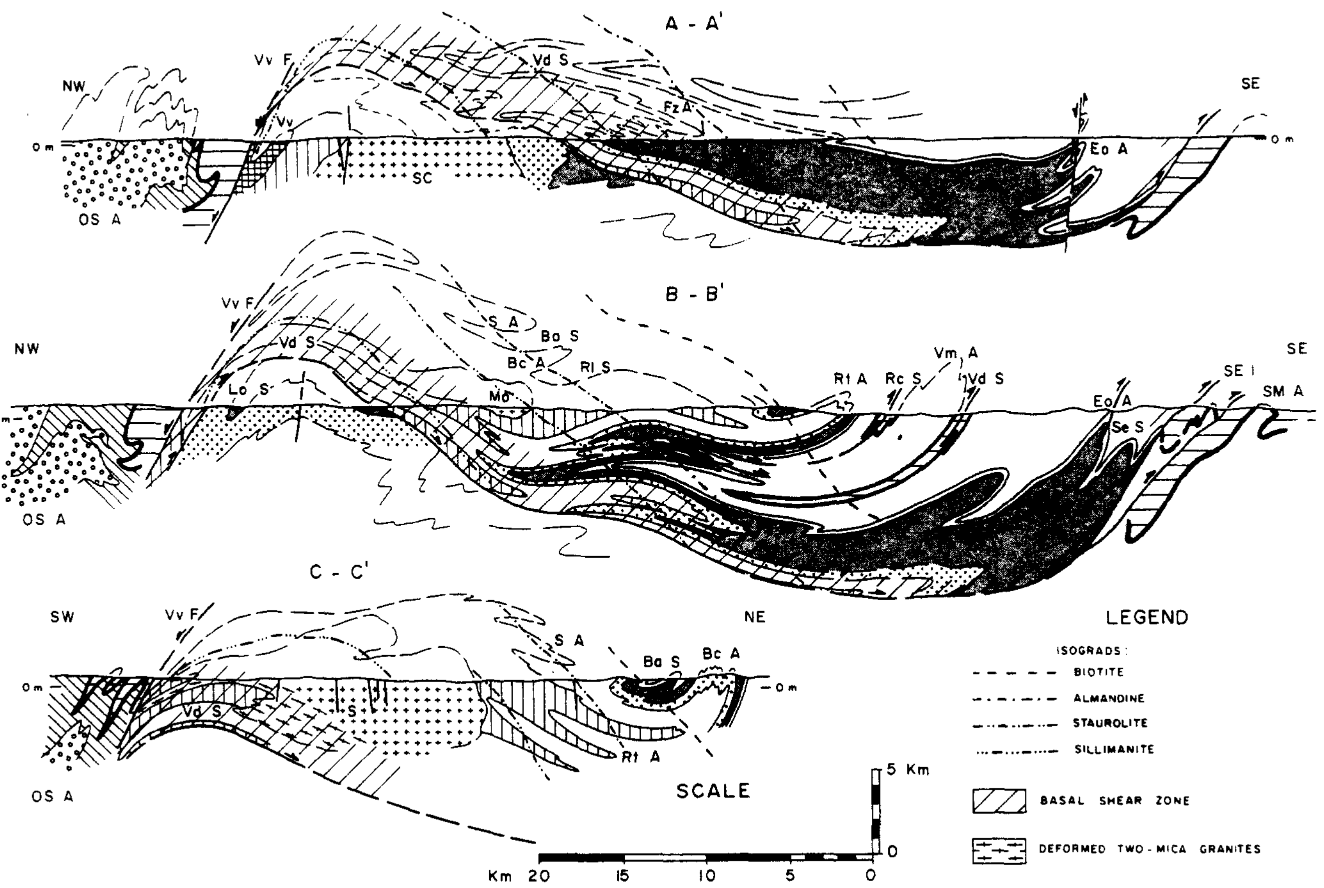

Fig. 4. Geological cross-sections through the Mondoñedo nappe. Structures: SE I, Santa Eulalia imbricate; Vv F, Vivero fault; OS A, Ollo de Sapo anticlinorium. Anticlines: S A, Sarria; Bc A, Becerreá; Rt A, Riotorto; Vm A, Villamea; Fz A, Foz; Eo A, Eo; SM A, San Martín. Synclines: Ba S, Baralla; Rl S, Real; Rc S, Rececende; Vd S, Villaodrid; Se S, Serantes; Lo S, Losada. Granitoid massifs: SC, San Ciprián; Vv, Vivero; Mo, Monseibán; S, Sarria. For frames other than those depicted in this figure and location of the sections, see Fig. 3.

basic elements: the thrust sheet, the footwall unit, and the thrust fault and related structures. In each of these, analysis will be made of the geometry and development of major and minor structures, paying special attention to minor structures, because they are accessible to direct observation and allow us to analyse the variation in deformation throughout the zone.

Structure of the thrust sheet

The thrust sheet, that is, the Mondoñedo nappe itself, is composed of large D_1 recumbent folds which were refolded by gentle D_3 folds, homoaxial with the former, and whose superimposition gave rise to type 3 interference patterns (Ramsay 1967). They are clearly visible on the map (Fig. 3) and the geological sections (Fig. 4). The large recumbent folds are associated with smaller folds of diverse order and to a generalized primary cleavage. At the base of the nappe, close to the thrust fault, there is a zone with very intense deformation in which small D_2 folds, S_2 cleavage and, in some cases, mylonites were developed; the description of these structures will be dealt with together with that of the thrust fault.

Major folds. These are recumbent, isoclinal or very tight folds with E vergence. Their outcrop on the map, appearance in profile section and names are shown in Figs. 3 and 4, respectively. The most dominant structures are two large folds: that described by Matte (1968) as the Mondoñedo–Lugo–Sarria recumbent anticline which includes our anticlines of Riotorto, Becerreá and Sarria, and the Villaodrid syncline (Walter 1966). The overturned limbs of both folds are more than 20 km long in profile section (Fig. 4). The axial traces of the folds in the northern part of the outcrop area show a curved form which is a consequence of the refolding by the D_3 folds. This aspect is particularly visible in the Villaodrid syncline. The isoclinal nature of many of these major folds results in the fact that on the map they show a narrow core (half-wavelength less than 2 km) which may extend in a N–S direction for distances up to 100 km. Thrust faults with small displacements were developed in the core of some of these great folds (Fig. 4), and these could be considered as D_2 structures.

Minor folds. In association with the major (first order) folds, various orders of smaller folds were developed. The second order folds are rare and have overturned limbs several hundred metres long. A particularly outstanding example is the fold developed on the normal limb of the Foz anticline which crops out along the beach of Benquerencia. A detailed cross-section of this outcrop is shown in Fig. 5.

The third order folds are the most abundant and their

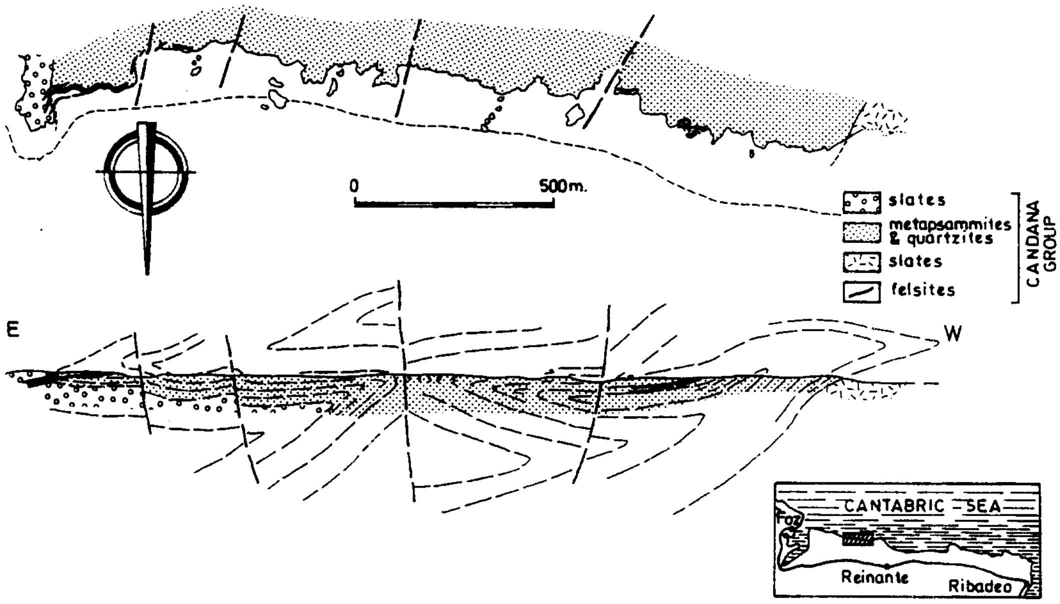

Fig. 5. Detailed geological cross-section of the Benquerencia area. Second order D_1 folds developed in the Transition Beds of the Cándana Group.

limbs rarely exceed 10 m in length. For their study, four sections have been analysed, three on the Cantabrian coast and the fourth to the south. These sections are (a) the Tapia de Casariego section, (b) the Rio section, (c) the Benquerencia section and (d) the Baralla section. The folds of sections (a), (c) and (d) are developed in sandstone, quartzite and slates or phyllites of the Cabos Series (a) and of the Cándana Group (c) and (d), while the folds of section (b) were developed in an alternation of slate, siltstone and sandstone of the Cabos Series, with a greater percentage of incompetent materials than are present in sections (a), (c) and (d).

The minor folds have subhorizontal hinges parallel to the direction of the major ones, approximately N 30°E in the first two sections of the coast, N 10°E in the Benquerencia section and N 160°E in the Baralla section (Fig. 6). The inclination of their axial surfaces depends on their position in the D_3 folds. They therefore dip between 60 and 80°W in Tapia de Casariego, are subhorizontal in the Rio and Benquerencia sections and are subvertical in Baralla (Fig. 4, sections A–A′ and C–C′, Fig. 6). The values of the interlimb angles are shown in Fig. 7; the most common values range between 10 and 50° except in the Baralla section where angles up to 100° are common.

In order to determine the geometry of the folded surfaces, Hudleston's (1973) classification has been applied in the sections (a), (c) and (d); the results are

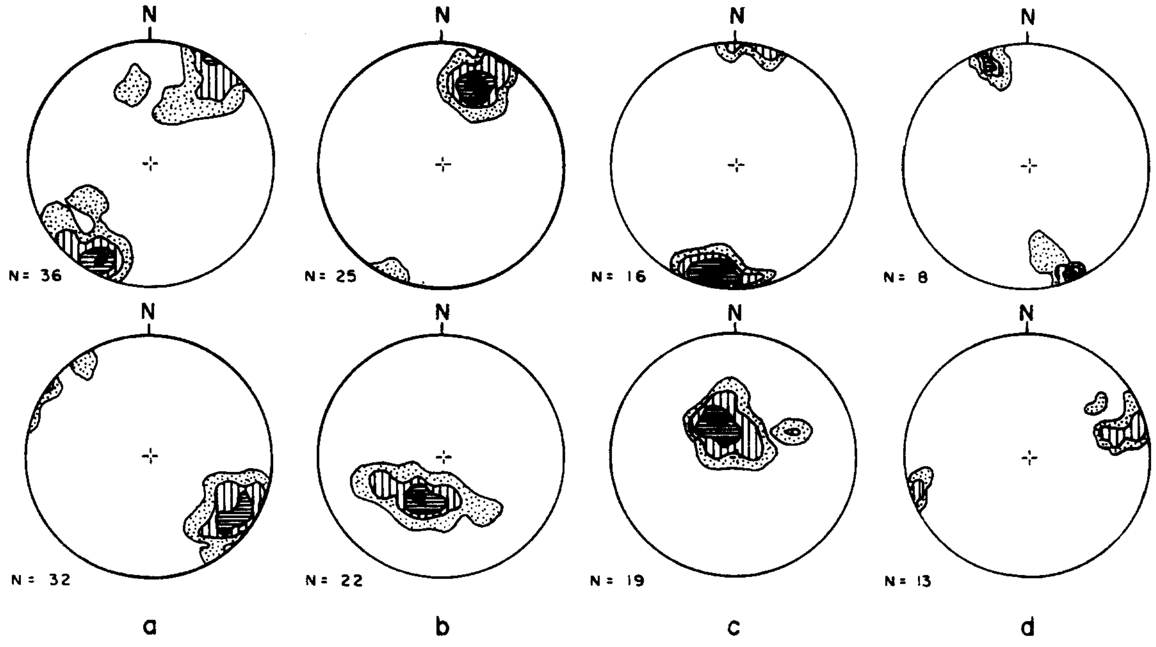

Fig. 6. Orientation of the fold axes (above) and poles to axial planes (below) of the minor D_1 folds in the four sections analysed in the thrust sheet. (a) Tapia de Casariego section. (b) Rio section. (c) Benquerencia section. (d) Baralla section. Contour interval is 5, 10, 15 and 30 per cent per 1 per cent area. Wulff net, lower hemisphere.

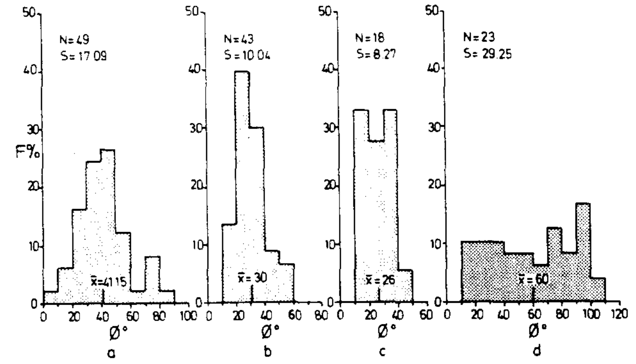

Fig. 7. Frequency histograms of the interlimb angle ϕ of the minor D_1 folds of the four sections analysed in the Mondoñedo nappe. (a) Tapia de Casariego section. (b) Rio section. (c) Benquerencia section. (d) Baralla section.

shown in Fig. 8. The folded layers have been classified after the method of Ramsay (1967); the competent layers generally belong to the 1C class, though on the coastal sections they are close to class 2.

The common occurrence of trains of periodic folds shows that these structures were developed in the early stages by a buckling process. The absence, on the other hand, of chevron forms in the folds of lower amplitudes, and the coexistence of chevron and rounded forms, even along the same axial surface, suggests that the chevron folds evolved from the rounded ones. The common appearance of chevron forms in the core of rounded folds suggests an evolutionary mechanism similar to that described by Johnson & Honea (1975). In advanced buckling stages and/or after them, the folds underwent a flattening process, as indicated by the geometry of the folded layers described above.

The fact that we are dealing with flattened parallel folds enables us to determine the mean overall shortening (Bastida 1981) in the different sections studied. This ranges between 36% for the Baralla section and 69% for the Benquerencia section. The Tapia de Casariego section immediately overlies the Mondoñedo thrust, those of Rio and Benquerencia at some 6 km above this fault and that of Baralla at least 10 km above (Figs. 3 and 4). The interlimb angles of the minor folds of the four sections (Fig. 7) indicate that the flattening decreases towards the east (coastal sections) and upwards (Baralla section). This conclusion may also be reached from the measurement of the mean overall shortening and may be extrapolated to the whole of the WALZ (Bastida 1981).

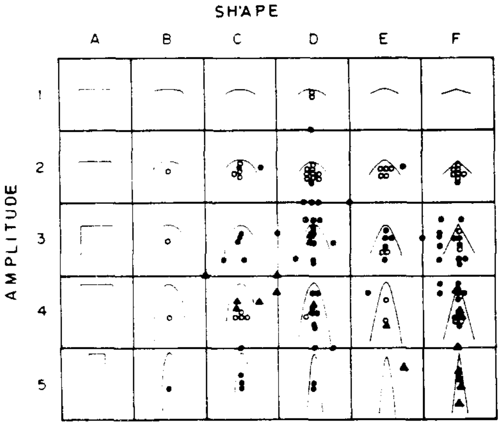

Fig. 8. Classification of the shape of the folded surfaces of the minor D_1 folds of the Mondoñedo nappe, according to the method of Hudleston (1973). Solid circles, Tapia de Casariego section; solid triangles, Benquerencia section; open circles, Baralla section.

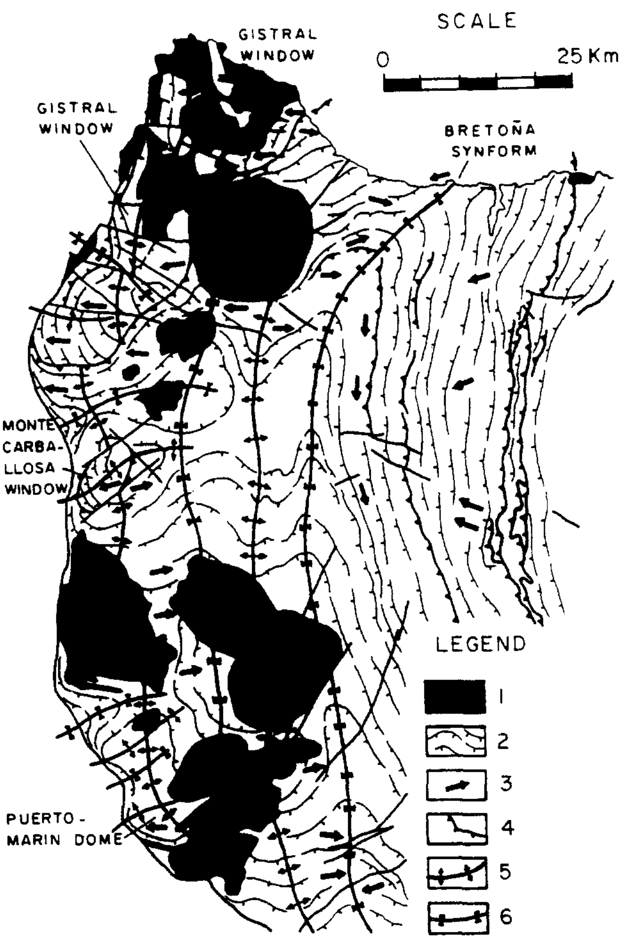

Fig. 9. Structural scheme of the Mondoñedo nappe and its autochthon. Note the dome and basin interference pattern in the western part and the different relationships between S_1 and the thrust fault in the thrust sheet and in the footwall unit in the south of the Gistral tectonic window. Legend: 1, Hercynian granitoids; 2, S_1 attitude showing dip; 3, mineral and stretching lineations; 4, thrust faults; 5, D_3 and transverse antiforms and 6, D_3 and transverse synforms.

The primary cleavage (S_1). A primary cleavage was developed in association with the folding; its attitude is shown in Fig. 9, where it may be seen to be slightly oblique to the basal thrust both in the tectonic windows and in the frontal part of the nappe. Its characteristics vary as a function of the lithology and of the intensity of deformation and metamorphism.

In the pelitic rocks, the cleavage varies from a fine-grained slaty cleavage in the eastern part of the nappe to a coarser schistosity, with grains often observable with the naked eye, in the western part. The fabric which develops in quartzite and sandstone shows a poorer preferred dimensional orientation and appears in the form of a rough cleavage except in the deeper zones of the nappe, to the west, where it appears as a penetrative fabric.

The most important mechanism in the formation of the slaty cleavage and of the schistosity in the pelitic rocks was crystallization and recrystallization of oriented phyllosilicates, while in the development of the rough cleavage, pressure-solution and intracrystalline deformation of quartz seem to have been more important.

On the schistosity surface, a mineral lineation or a stretching lineation is commonly developed, the orientation of which is shown in Fig. 9. In the upper levels of the thrust sheet, close to the axial trace of the Bretoña synform, the stretching lineation, deduced from deformed fossils, appears subparallel to the axes of the D_1 folds; but lower down the lineation has a transverse orientation which varies between N 70°E and N 110°E. The mineral lineations in the basal part of the nappe, around the tectonic windows and in the Puertomarín dome, must be interpreted as $D_1 + D_2$ structures.

Structure of the footwall unit

The footwall unit outcrops mainly to the east of the nappe, where it has been called the Navia unit (Marcos 1973). However, the footwall rocks also outcrop in two small areas in the western part of the nappe, the Gistral and Monte Carballosa tectonic windows (Martínez-Catalán 1980, 1981). The stratigraphic succession in these areas is different from that in the thrust sheet, and includes the Gistral Quartzite (Fig. 2), which is of particular interest. This is a thick formation within the Cándana Group comprising some 1000 m of massive quartzites which are clearly differentiable both from the Cándana Quartzites of the Mondoñedo nappe and from the platy quartzites in the lower part of the thrust sheet, to the west, surrounding the tectonic windows and in the Puertomarín dome. The deformational sequence is essentially the same as in the thrust sheet; nevertheless, the style of the structures is somewhat different, as described below.

Major folds. The axial surfaces of the major folds dip more steeply in the footwall unit than in the nappe. Though they were modified by the later D_3 refolding, their original inclination must have also been greater than that of the recumbent folds of the nappe. They are also smaller: in the footwall unit, the length of the overturned limbs ranges between 2 and 4 km. The interlimb angles are considerably greater in the footwall unit, commonly 50–60°, and isoclinal folds were not developed.

The attitude and geometry of the major folds controls the morphology of the D_1/D_3 interference patterns in the hangingwall and footwall units. Thus, the recumbent and isoclinal nature of the folds of the nappe favours the development of 'têtes plongeantes'. An example of such a situation is the synformal anticline of Foz (Fig. 4, section A–A'). However, in the footwall unit, the fact that the folds are more open and their limbs more inclined prevents the formation of 'têtes plongeantes' and favours the formation of peculiar interference patterns in which the reverse limb of the D_1 folds is not refolded, as in the San Martín anticline (Fig. 4, section B–B').

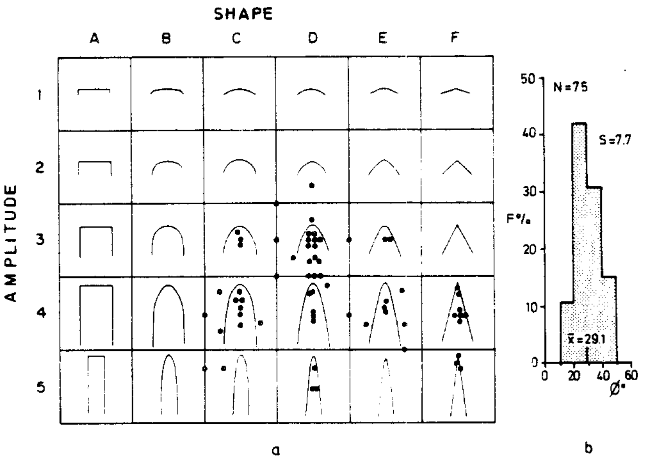

Fig. 10. (a) Classification of the folded surfaces of the minor D_1 folds of the footwall unit in the Burela area, according to the method of Hudleston (1973). (b) Frequency histogram of the interlimb angle ϕ of the same folds.

Minor folds. We distinguish those folds that are present in the footwall rocks of the western part of the nappe (Gistral and Monte Carballosa tectonic windows) and those to the east of its frontal zone.

Within the Gistral window, an exceptional site for the analysis of these folds is that of Burela. There a train of folds over 1 km in length is developed in the Upper Cándana Quartzite, close to the hinge zone in the normal limb of a major fold. The length of the limbs of the minor folds ranges between 3 and 12 m. The folds are asymmetric, the mean value of the long (normal) to short (reverse) limb-length ratio being 1.54. They are approximately cylindrical, with subhorizontal hinges trending approximately N 10°E and axial surfaces inclined 30–60° W. The frequency distribution of the interlimb angles (Fig. 10b) shows a modal interval of between 20 and 30°. The geometry of the folded layers is close to that of similar folds, though occasionally the ratio of orthogonal thickness to dip plots penetrate appreciably into the class 1C field (Ramsay 1967). The result of the classification of the folded surfaces according to Hudleston's method is shown in Fig. 10(a).

The existence in the zone of Burela of a long train of folds with a dominant wavelength suggests that the folds were originated by a buckling process, on which was superimposed, either simultaneously or later, an important and possibly rather oblique flattening, in the sense described by Hudleston (1973). The normal limbs characteristically exhibit a greater stretching than the reverse ones. The progressive extension of the limbs has commonly led to their rupture by boudinage. The mean overall shortening shown by these folds is in the order of 65%.

The minor folds to the east of the frontal part of the nappe are more open and less flattened than those in Burela and in most of the thrust sheet, with an overall shortening which ranges between 40 and 50%.

The S_1 cleavage in the footwall unit is similar to that in the thrust sheet, though in the former the intensity of development decreases progressively towards the east.

As may be seen in Fig. 9, especially to the S of the Gistral window, the S_1 schistosity is sharply truncated by the basal thrust. This is due to the relatively high angle formed by both surfaces, which is greater than that formed by the same surfaces in the thrust sheet (see axial surfaces in Fig. 4, sections A–A′ and B–B′). Generally, the S_1 is a slaty cleavage in the pelitic rocks and a rough cleavage in sandstone and quartzite. Evidence of crystallization and recrystallization of oriented phyllosilicates decreases towards the east, whereas evidence of the role of pressure-solution in the development of S_1 increases.

The thrust fault and related structures

The outcrop of the basal thrust fault of the Mondoñedo nappe is best displayed at the eastern limit of the unit. However, it also outcrops on the western border zone outlining the Gistral and Monte Carballosa tectonic windows (Fig. 3). An inspection of the trace of the Upper Cándana Quartzite along section B–B′ in Fig. 4 suggests a minimum displacement of the Mondoñedo nappe over its basal thrust in the order of 40 km. To draw this conclusion, it is necessary to assume that the Losada syncline was next to the Eo anticline. There is no direct evidence to demonstrate that these folds were in fact adjacent. Therefore, the reverse limb of the Eo anticline may have its prolongation in the footwall unit to the west of the Losada syncline, in which case the relative displacement of the two sides of the fault may be greater.

The characteristics of the zone of deformation associated with the thrust vary considerably from east to west, because of the different environmental conditions under which the deformation took place. The two areas will therefore be considered separately.

In the frontal zone of the nappe, the basal thrust may be followed for over 165 km, until it is buried below the Tertiary cover of the Iberian Plateau. In the northern branch of the WALZ, it strikes roughly N–S, while on the S branch, it curves progressively towards the E, parallel to the other structures that outline the Iberian Arc.

According to Marcos (1973), and as may be seen from the map and cross-sections, the thrust overlies the same formation for considerable distances, even though it cuts the D_1 folds, whose axial trace is slightly oblique to that of the thrust fault. The surface of the fault is in turn folded by the D_3 folds, such that its outcrop is often sinuous and its dip close to the vertical. The outline is further complicated by the appearance of related thrusts and associated imbricate stacks.

Associated with the basal fault is a zone of localized deformation, no more than 200 m in width, in which different kinds of structures were developed, preferentially in pelitic rocks (Marcos 1973). This zone exhibits small folds (limb length less than 1 m) which are very tight with curved hinges. Accompanying these folds is a well developed S_2 crenulation cleavage. Also common in the fault zone are brittle structures such as en-échelon tension-gashes and polished surfaces with fibrous quartz crystals (Marcos 1973). Bastida *et al.* (1979) suggested that the thrust was developed close to the leading edge of the nappe under conditions where mixed modes of failure could occur.

In the western part of the nappe a D_3 antiform and gentle E–W trending folds (Fig. 9) expose the basal thrust fault, which outlines the Gistral and Monte Carballosa tectonic windows. In this zone, the thrust occurs as a single fault that cuts the D_1 folds both of the footwall and hangingwall units. An important zone of intensified deformation some 3 km thick was developed around the fault, mainly in the hangingwall rocks. Within this zone, numerous minor folds have developed, as well as S_2 crenulation cleavage or schistosity and, in the area close to the fault, mylonitic and phyllonitic rocks.

The D_2 folds present in this zone are particularly easy to observe on the Cantabrian coast and are characterized by their consistent small size (inverted limbs with lengths less than 1 m), strong asymmetry and subhorizontal axial planes. They show always a vergence towards the outer zones of the orogen (E or SE). Their hinges are commonly curved, giving rise to a pronounced dispersion of axial directions (Fig. 11a). This curvature produces sheath folds with approximately elliptical sections normal to the movement direction, and sigmoidal en-échelon hinges, all reflecting the markedly heterogeneous nature of the deformation. The axial surfaces remain approximately plane and parallel, however. The folds are very tight, with interlimb angles between 10 and 30° (Fig. 11c). The geometry of the folded surfaces is shown in Fig. 11(b). That of the folded layers is of the classes 1C, 2 or 3 of Ramsay (1967), though they nearly always tend towards class 2 (subsimilar folds).

The D_2 folds also crop our abundantly in the Puertomarín region where they display identical characteristics. A Hansen (1971) diagram for this zone gives a separation line (Fig. 12) with a direction N 85°E, roughly parallel to the mineral lineation in this area.

The S_2 cleavage in this zone of ductile deformation is either a crenulation cleavage or a penetrative schistosity, and the phyllosilicates are visible with the naked eye. The S_2 crenulation cleavage usually exhibits tight microfolds, with interlimb angles of generally less than 50°. The geometry ranges between chevron and rounded forms, the former normally corresponding to a greater degree of amplitude and flattening. The S_2 schistosity may develop either from the S_1 cleavage by progressively greater flattening and concomitant increase in grain size, or from the prior development of S_2 crenulation cleavage, by a tightening of the microfolds, leading to the destruction of the hinge zones. In this latter case, isolated microfolds are usually preverved as relics.

At the base of the thrust sheet, the rocks show very intense deformation leading to the development of mylonitic fabrics in the quartzites through a process involving substantial grain-size reduction.

The zone which has just been described thus exhibits intense ductile deformation, though it is markedly heterogeneous, with a lack of cylindricity in the folds,

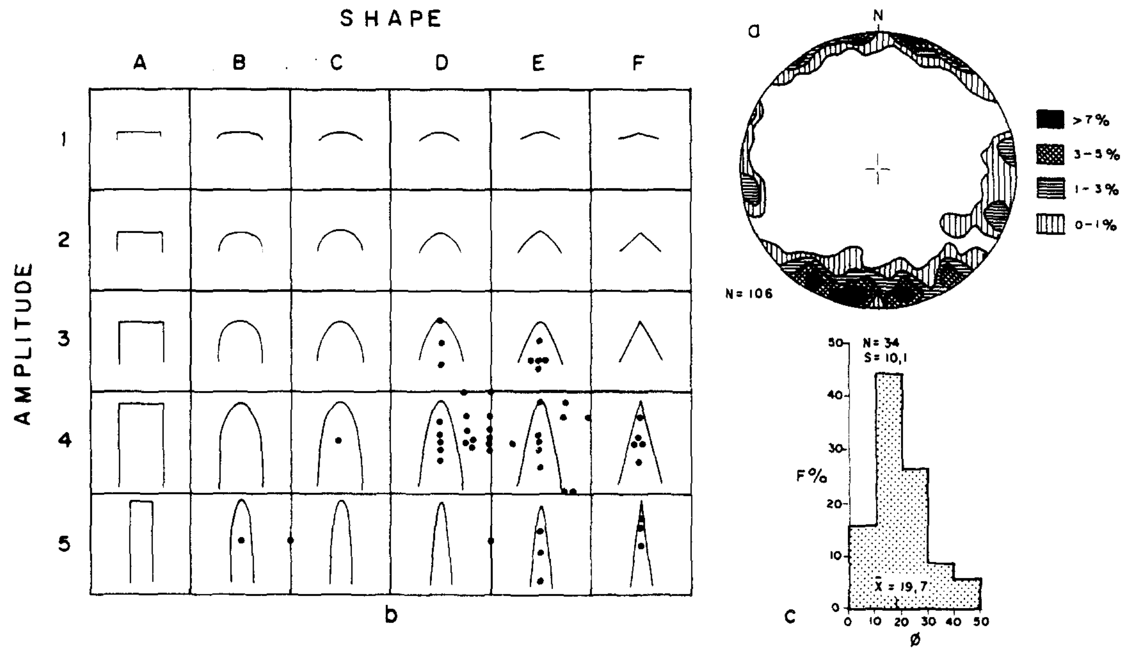

Fig. 11. Geometry of the D_2 folds generated in the ductile shear zone of the internal part of the Mondoñedo nappe in the section of the Cantabrian coast west of Foz. (a) Orientation of fold hinges. (b) Classification of the folded surfaces according to the method of Hudleston (1973). (c) Frequency histogram of the interlimb angle.

and in which it is possible to distinguish bands of different strain intensities. The folds, consistently small, are recumbent, show strong asymmetry and their vergence is always towards the external zones of the orogen. All these characteristics suggest an intense deformation by shearing, such that this zone has been interpreted as a ductile shear zone (Bastida & Pulgar 1978, Pulgar 1980, Martínez-Catalán 1979, 1980, 1981). In this context, the development of the thrust fault must have represented a late stage in the evolution of the shear zone.

The characteristics described for the thrust fault and the related structures confirm that the fault zone was produced under environmental conditions leading to mixed modes of failure in the eastern part, at higher levels and closer to the leading edge of the nappe; whereas at deeper crustal levels, now exposed to the west, this zone evolved as a wide ductile shear zone prior to the development of the thrust fault.

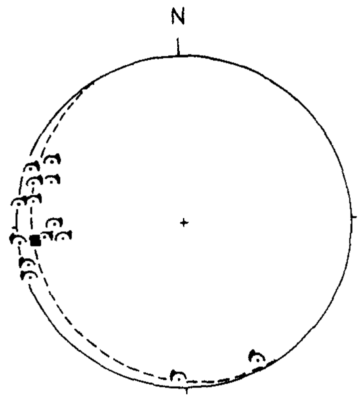

Fig. 12. Hansen (1971) diagram showing the asymmetry patterns of D_2 folds in the Puertomarín area. Wulff net, lower hemisphere. The square shows the separation line, roughly parallel to the shear direction. The plane containing it represents the mean of the axial surfaces of the folds.

METAMORPHISM

In the northern outcrop area of the Mondoñedo nappe, several metamorphic zones may be distinguished. The chlorite, biotite, almandine-rich garnet, staurolite and sillimanite zones are shown in Fig. 13,

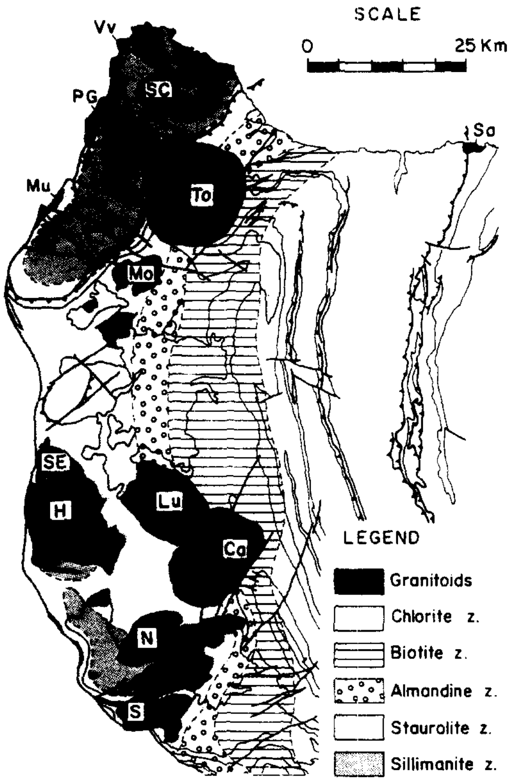

Fig. 13. Metamorphic zones in the Mondoñedo nappe and its autochthon based on the first appearance of index minerals in metapelites. Granitoid massifs: Ca, Castroverde; H, Hombreiro; Lu, Lugo; Mo, Monseibán; Mu, Muras; N, Neira; PG, Penedo Gordo; S, Sarria; SC, San Ciprián; SE, Santa Eulalia; To, La Tojiza; Vv, Vivero.

each one delimited by the earliest appearance of the respective mineral indices. An andalusite zone could also be distinguished, but in general this mineral appears near the beginning of the sillimanite zone, such that the isograds of both minerals are subparallel and very close. A kyanite zone exists in the Puertomarín zone, but this is overprinted by the sillimanite zone and has thus not been shown.

The isograds were distorted by the third deformation phase and, as may be seen from Fig. 13, those of biotite, almandine and staurolite cut across the folds of the first deformation phase. The outcrop configuration of these isograds suggests that they are strongly inclined (Fig. 4). The biotite isograd, and a considerable part of almandine isograd, lie in the hinge zone of the D_3 synform, implying that they have undergone very little modification during the third phase. Thus both isograds were, in this zone, strongly inclined by the time the displacement of the thrust sheet had finished.

The sillimanite isograd reflects the interference produced by the D_3 and the transverse folds (Fig. 13), because the sillimanite zone outcrop coincides with the Gistral and Puertomarín domes. However, in the dome responsible for the appearance of the Monte Carballosa tectonic window, neither sillimanite nor andalusite appear (see also Fig. 14). This implies that in a N–S cross-section both isograds cut across the D_1 and D_2 structures and that they are not parallel to the rest of the isograds, suggesting a relationship with deeper seated and localized thermal upwarps.

In the zones with medium grade metamorphism of the thrust sheet, the metamorphism is plurifacial (Bastida & Pulgar 1978). A first paragenesis exists with almandine, staurolite and, locally, kyanite, indicative of a Barrovian-type metamorphism (M_1) with a medium-pressure geothermal gradient. A second paragenesis with andalusite and sillimanite is indicative of a metamorphism (M_2) with a low-pressure geothermal gradient, in all probability linked to an increase in crustal heat flow. The andalusite commonly replaces the staurolite porphyroblasts of the first association. A third metamorphic episode (M_3), retrograde in character, may be identified in the basal shear zone, mainly where this crops out around the tectonic windows. In this episode garnet was replaced by pseudomorphic aggregates of chlorite or chlorite and biotite, andalusite by mica, and phyllonites formed in the schists, with crystallization of chlorite and biotite in microshears. The relationship between metamorphic episodes and deformation phases is shown in Fig. 14.

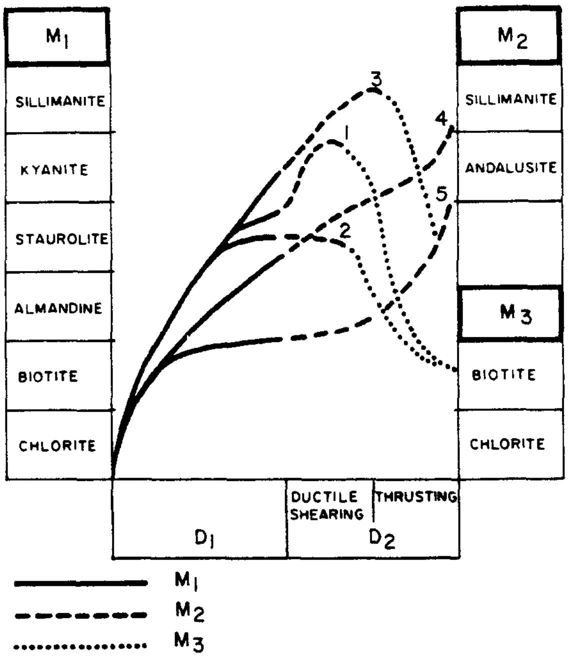

Fig. 14. Scheme of timing and grade of the different metamorphic episodes with respect to the D_1 and D_2 phases of deformation in three areas of the thrust sheet (1–3) and two of the footwall unit (4 and 5). 1, To the east of the Gistral window (west of Foz). 2, east of the Monte Carballosa window. 3, Puertomarín dome. 4, SW of the Gistral window (Losada). 5, NE of the Gistral window (Burela). The index minerals of the three metamorphic episodes have been used as indicators of the degree of metamorphism. M_1 is a Barrovian-type metamorphism, pre-D_2, common to the thrust sheet and the footwall unit. M_2 is a low-pressure metamorphism which shows different relationships with D_2 in the thrust sheet and in its autochthon. M_3 is a retrograde metamorphism which is only appreciable in the allochthonous unit.

The first paragenesis reached its climax at the end of D_1 or between D_1 and D_2. The second did the same at the beginning of D_2 in the part of the basal zone of the Mondoñedo nappe which surrounds the Gistral tectonic window. The andalusite porphyroblasts appear in this zone with a markedly preferred orientation (N 110° E on the coast between Burela and Foz) owing to the effect of the deformation in the shear zone. In the Puertomarín area, the climax of M_2 was reached somewhat later, though within the D_2 phase, and produced the growth of sillimanite and the local development of injection migmatites coeval with the intrusion of the Sarria granite. In the sense of Winkler (1974), the metamorphism did not reach high grade. The retrograde metamorphism coincided with the final stages of the displacement of the thrust sheet.

The metamorphic evolution in the footwall unit may be appreciated only in the Gistral tectonic window, because in the Monte Carballosa tectonic window only massive Gistral Quartzites crop out (without granitic injections) and to the east of the frontal thrust the metasediments were not subjected to conditions above those of the chlorite zone. In the Gistral window, high grade metamorphism was reached and migmatites of both in-situ and injection type developed. However, in two areas close to the thrust fault, in Burela (to the east of the window) and in Losada (to the southwest, see Fig. 3), unaffected by the migmatization, the pre-thrusting metamorphism appears to have been similar to that of the M_1 episode in the thrust sheet, with almandine and staurolite in Losada and, at most, biotite in Burela (Fig. 14). A second progressive metamorphic episode (M_2) characterized by the appearance of andalusite and sillimanite and by the development of migmatites, was initiated during the second phase of deformation, reaching its climax somewhat after, between D_2 and D_3 or at the beginning of D_3. No evidence of retrograde metamorphism has been found in the footwall unit.

GRANITE EMPLACEMENT

The two-mica granites and synkinematic biotite granitoids of the Mondoñedo nappe show evidence of having been affected by the second deformation phase. The most spectacular case is that of the Sarria massif (Fig. 15b). This is an intrusive two-mica granite whose bottom cuts the overturned limb of the Villaodrid syncline and extends upwards to the normal limb of the Sarria anticline (Fig. 4, section C–C′). Most of the massif is undeformed, though an *L–S* shape fabric is present in a band about 3 km thick at its base. This fabric must have been nearly horizontal before D_3 folding, with the mineral lineation trending approximately E–W. Dykes and veins of granite, aplite and pegmatite intruded into the adjacent schists show the same fabric and are boudinaged and affected by folds of the same geometrical characteristics as the D_2 folds in the metasediments of the basal shear zone. The deformed rocks show, in thin section, a recrystallization (sometimes total) and a subparallel arrangement of the inequigranular grains, which define a schistosity (Martínez-Catalán 1983).

The Santa Eulalia granodiorite (Fig. 15a), which belongs to the group of synkinematic biotite granitoids, commonly exhibits *L–S* shape fabrics and, locally, has undergone phyllonitization, with the development of subhorizontal microshears and an oblique schistosity arranged sigmoidally between the microshears, indicating a W to E overthrusting sense of movement. Where deformed, recrystallization is quite apparent, accompanied by a reduction in grain size. Moreover, decimeter-thick mylonite bands have been identified. The eastern part of the massif, which remains outside the basal shear zone of the Mondoñedo nappe, has not been deformed.

The two-mica granite of Hombreiro (Fig. 13) also shows evidence of deformation on its southern and western margins. That on the western margin is probably related, at least partially, to the Vivero fault, a post-D_2 structure. However, that of the southern border is clearly attributable to the second phase, since it occupies a structural position very similar to that of the deformed area of the Sarria granite, except that it is to the north instead of to the south of the Puertomarín dome. However, deformation is less intense than in the adjacent Santa Eulalia massif, which could be due to a later emplacement, during the second deformation phase.

The Muras massif (Fig. 13) is also deformed near the Vivero fault. The deformation is essentially ductile and the sense of shear, deduced from microshears, is towards the west, coherent with the movement of the above-mentioned fault. However, this granite may be the prolongation towards the west of the Monseibán massif (Fig. 13) which is only weakly deformed, on its western margin closest to the thrust fault (Fig. 4, section B–B′). In this case the upper part of the massif would have been moved several km to the east with respect to its root zone as a consequence of the ductile deformation in the shear zone.

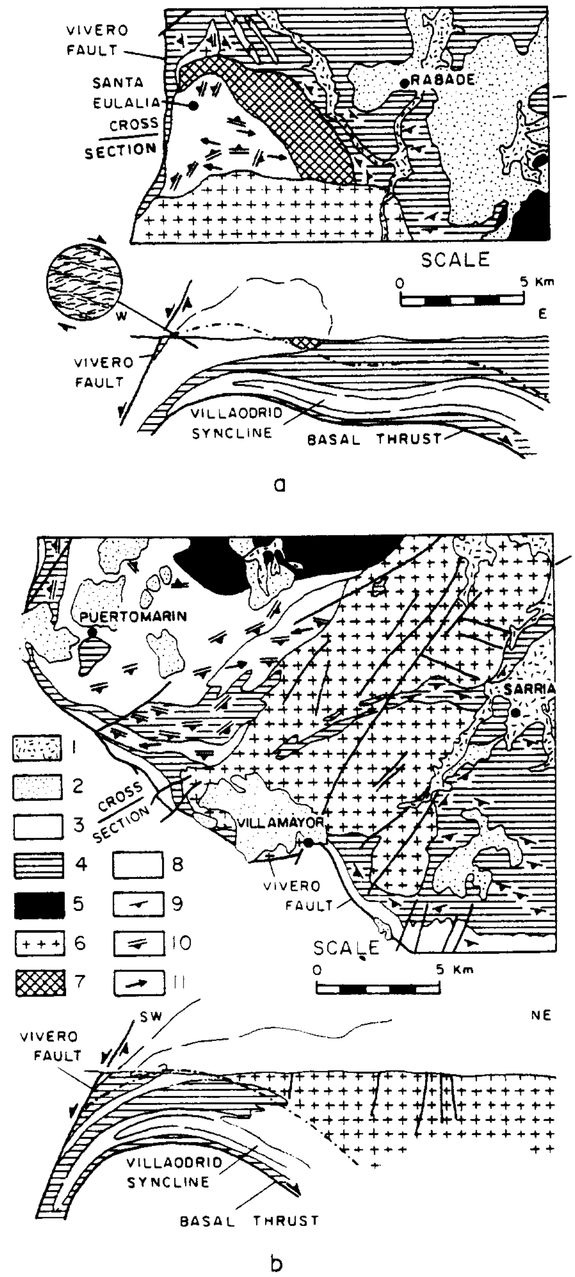

Fig. 15. Geological maps and cross-sections of the Santa Eulalia (a) and Sarria (b) massifs, showing their relationship with both the basal thrust and the basal shear zone of the Mondoñedo nappe. The insert in (a) represents microshears and sigmoidal schistosity between them, indicating the sense of shear. Legend: 1, Quaternary; 2, Tertiary; 3, Lower Cambrian (Cándana Group); 4, Upper Proterozoic (Villalba Series); 5, postkinematic granitoids; 6, two-mica granites; 7, synkinematic biotite granitoids; 8, parts of Santa Eulalia and Sarria massifs deformed by D_2; 9, S_1 schistosity; 10, S_2 schistosity; 11, mineral lineation in granitoids. The dash–dot line represents the upper limit of the basal shear zone.

None of the massifs mentioned have been dated radiometrically and the only information available comes from syntectonic massifs of similar composition intruded in the Ollo de Sapo anticlinorium, west of the Vivero fault. Some of these massifs have been dated, giving Viséan–Namurian ages (between 315 and 323 ± 10 Ma after Capdevila & Vialette 1970 and Ries 1979).

A special case is that of the San Ciprián and Vivero massifs. The former outcrops in the migmatitic area occupying the core of the Gistral tectonic window. It is a two-mica granite, commonly inhomogeneous, which intruded into the Mondoñedo nappe, cutting the thrust fault in a zone between Vivero and Mondoñedo (Fig. 3). The second is a biotite granitoid cropping out to the northwest in the same tectonic window. The deformation in both is intense only in areas close to the Vivero fault, where a mineral lineation trending N 130°E, and microshears with sense of movement towards the NW, are developed. The deformation must therefore be related to this structure and not to the second phase (D_2). In any case, the San Ciprián massif postdates the emplacement of the Mondoñedo nappe, whose basal thrust it cuts in the above-mentioned zone, where it is undeformed.

It may be concluded that both the synkinematic biotite granitoids and the two-mica granites forming part of the Mondoñedo nappe were affected by the second deformation phase though not by the first, since outside the shear zone these granitoids cut across the large overturned folds and are not deformed. Only the San Ciprián granite intruded into the Mondoñedo nappe once its displacement had been completed. All the massifs mentioned were affected by the third deformation phase.

The intrusion of the granitic massifs and the concomitant thermal domes must have exerted some influence on the metamorphism of nearby rocks. This would explain why the metamorphic climax corresponding to the greater geothermal gradient (M_2) was reached at a relatively early stage (Fig. 14), within the D_2 phase where granites did not intrude (as in the area around the Gistral window) and later where such intrusion took place (as in the Puertomarín dome and in the Gistral window).

DISCUSSION OF THE EVOLUTION OF THE MONDOÑEDO NAPPE AND ITS RELATIONSHIP WITH A POSSIBLE INTRACONTINENTAL MEGASHEAR

Two stages may be distinguished in the structural evolution of the metasedimentary cover of the WALZ and of the Mondoñedo nappe in particular; these are of clearly different tectonic significance. The first is characterized by important horizontal displacements, leading to the appearance of E-vergent folds and overthrusts. The second consisted of a general refolding of the previous structures and gave rise to gentle folds with subvertical axial planes, representative of the last episodes in the shortening of the belt.

During the first stage, deformation was particularly important in the western part of the WALZ where large D_1 recumbent isoclinal folds were developed, implying a period of intense generalized ductile deformation. As the orogeny progressed, the deformation tended to concentrate in a subhorizontal ductile shear zone (D_2) oblique to the D_1 folds, and culminated in thrusting along an important fracture, the Mondoñedo thrust fault, which is also considered to be a D_2 structure. The evolution during this stage is represented schematically in Fig. 16.

The minimal translation due to the basal fault is about 40 km. If we add to this the translation in the ductile shear zone and that related to the generation of the D_1 folds, the total displacement of the metasedimentary cover may be more than 80 km. Such a displacement must pass downwards in some way into the basement and, in our opinion, it is highly probable that in the prolongation of the basal shear of the Mondoñedo nappe there exists an intracontinental megashear, similar to those that have been proposed, according to seismic profiles, in the Alps (Hsü 1979), the Scottish Caledonides (Soper & Barber 1982), the Scandinavian Caledonides (Cuthbert *et al.* 1983) and, according to other criteria, in the Himalaya (Le Fort 1975, Mattauer 1975, Pêcher 1977, Valdiya 1980). Such a megashear would have been developed as a result of a continental collision whose suture would be situated to the W of the present coast of NW Spain (Dewey & Burke 1973, Bard *et al.* 1980.

The intraplate convergence zone would have been active during the first deformation phase, which would explain the exceptional magnitude of the D_1 folds in the Mondoñedo nappe. It has already been pointed out that the magnitude of the folds decreases immediately to the east, and the same is likely to occur to the west of the Ollo de Sapo anticlinorium, where no large D_1 structures have been identified. Furthermore, from the analysis of the D_1 minor folds it may be inferred that the strain increases towards the west in the Mondoñedo nappe as well as throughout the whole WALZ for a given tectonic level. It is highly probable that in the prolongation of the core of the Riotorto–Becerreá–Sarria anticlinorium to the W of the Vivero fault, there exists a basement wedge responsible for the generation of this huge fold.

Paleogeographically, the Ollo de Sapo zone is characterized by being a positive area in which the thick detrital deposits of Upper Proterozoic age (Villalba Series) are lacking, and the same is in general true also for the thick platform facies sediments of Cambrian age (Fig. 2). The decrease in thickness observed in these Cambrian sediments in the WALZ towards the Ollo de Sapo anticlinorium (Fig. 2, Caurel domain) suggests that to a large extent their absence in the Ollo de Sapo zone is due to lack of sedimentation. If the Ollo de Sapo zone represents a non-subsiding area to the west of a sinking furrow (the WALZ), this is probably due to it having had a thicker continental crust than that of the adjacent furrow (Fig. 16a). The boundary between the two may therefore have been a fractured zone, which could have localized the megathrust in the basement.

Basement may lie immediately below the Ollo de Sapo Formation, or form its basal part. The degree of metamorphism in the Ollo de Sapo anticlinorium is lower than that in the internal parts of the Mondoñedo nappe. This suggests that it forms a wedge separated from the nappe by another shear zone which, at depth, may coalesce with the first (Fig. 16). In a situation involving intracontinental duplication, there is a reasonable possibility of the basement cropping out on the overthrust side. The remarkable continuity of the Ollo

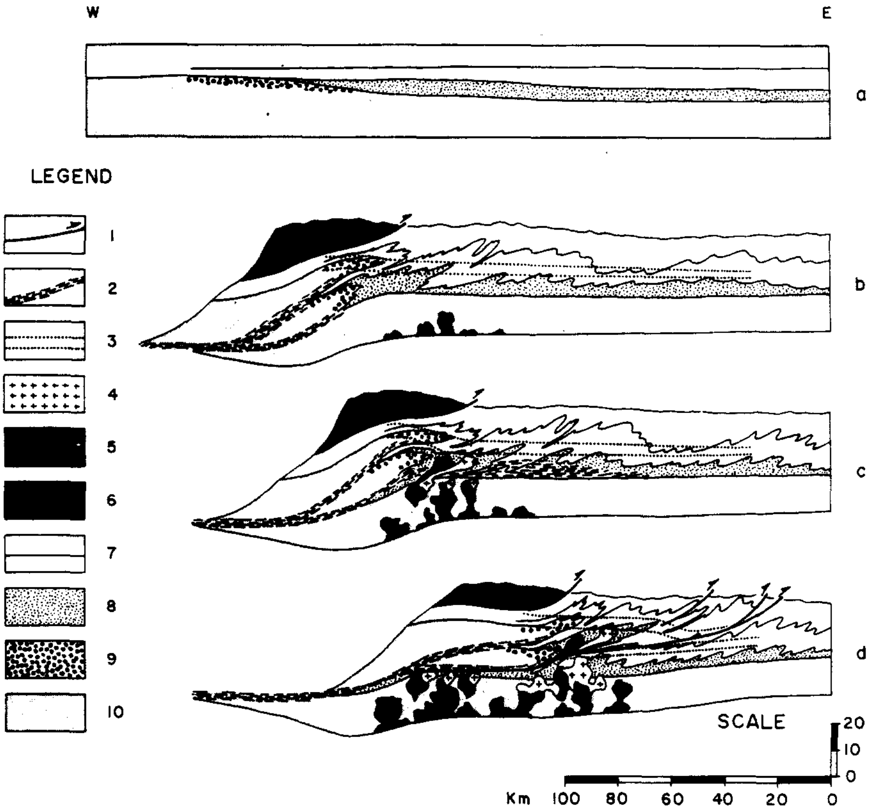

Fig. 16. Evolutionary sequence proposed for the first stage of deformation in the Mondoñedo nappe and surrounding areas. (a) Aspect of the crust before deformation. (b) Towards the end of the D_1 episode. The synkynematic biotite granitoids are being generated in the lower crust. (c) Ductile behaviour during the first part of the D_2 episode. Generation of two-mica granites and creation of a shear zone in the lower parts of the metasedimentary pile. (d) Thrust tectonics during the D_2 episode. The displacements are tentative. The allochthonous complexes with mafic and ultramafic rocks were thrust over the Ollo de Sapo anticlinorium and then passively translated. Legend: 1, thrust fault; 2, shear zone and 3, M_1 isograds of biotite (above) and almandine (below). The isograds have been deduced from the first parageneses and, in (c) and (d), do not represent the disposition of the metamorphic zones at that moment, especially in the footwall unit, where they were overprinted by others of higher degree. 4, two-mica granites; 5, synkinematic biotite granitoids; 6, allochthonous complexes with mafic and ultramafic rocks of NW Spain; 7, Palaeozoic sediments. The continuous line represents the top of the Arenigian quartzite; 8, Upper Proterozoic Villalba Series; 9, Ollo de Sapo Formation; 10, basement.

de Sapo anticlinorium in the Iberian Arc (almost 600 km) further supports the argument in favour of the existence of a continental-scale megashear in relation to this structure.

Prograde metamorphism took place during the first deformation stage. This was epi- to mesozonal and, at a given moment, between the D_1 and D_2 phases, changed from a medium-pressure geothermal gradient (M_1) to a low-pressure one (M_2). During M_2, two kinds of granitic massifs were intruded into the Mondoñedo nappe.

The synkinematic biotite granitoids, supposedly Viséan–Namurian in age, originated from deep zones of the crust. In view of the general progression of the deformation towards the outer zone of the orogen, it is unlikely that when they were produced there were other intracontinental shear zones further to the east than those proposed above. This would preclude an origin for the granitoids by heating of the lower crust caused by crustal duplication or by shear-generated heating (Fig. 16b). The latter mechanism may furthermore be discarded because at the depth at which the shearing would have occurred, the temperature would have favoured dislocation creep mechanisms in crustal rocks. This would have hindered a rise in temperature large enough for fusion to occur (Toksöz & Bird 1977). As these authors have proposed, a possible source of heat to form these granitoids lies in the deeper mobile rocks of the asthenosphere, if they came in contact with the base of the crust. The granitoids which were unaffected by the D_2 episode but which were deformed by the D_3 phase are abundant to the west of the nappe (though they have not been depicted in Fig. 16) as well as in the Gistral tectonic window. These, together with the postkinematic granitoids may, in contrast, have been generated as a consequence of the continental duplication, which could explain why they were unaffected by the thrust tectonics.

The synkinematic biotite granitoids in turn may have heated the upper parts of the basement and the overlying metasediments. This would be reflected in localized fusion of the rocks, giving rise to the two-mica granites (the two types are commonly spatially related); and in a change in the geothermal gradient. The temperature reached at that time by the internal zones of the Mondoñedo nappe may be estimated, on the basis of the mineral associations, at some 550–600°C. The increase in temperature may have given rise to a thermally softened zone in which the shear zone would have been located as a detachment level immediately above the basement (Fig. 16c), according to the model proposed by Armstrong & Dick (1974).

In more elevated zones, one or several fractures would have been created, which would have coalesced downwards into a single basal thrust. This would probably have a geometry not very different from that of similar structures in foreland thrust belts (Dahlstrom 1970, Elliott 1978). The existence of frontal imbricates has already been mentioned and it is also obvious that, in large scale terms, the fault cuts up-section. Furthermore, the peculiar geometry of the M_1 isograds, markedly oblique to the basal thrust (Fig. 4), exhibit relationships with it which are similar to those of the beds with the thrusts in areas with ramps in the outer zones of orogenic belts. If these isograds were roughly planar and horizontal after M_1, which is reasonable in geological terms, their present disposition suggests the possible existence of a ramp to the west of the nappe (Fig. 16d).

The continued movement of the intracontinental megashear would have finally moved the basal ductile shear zone onto the thrust, superimposing the lower parts of the Mondoñedo nappe over the more superficial and less intensely deformed rock of the autochthon. During this last part of the first stage, retrograde metamorphism would have taken place in the basal zone of the thrust sheet (M_3) while, apparently, the overridden rocks of the Gistral tectonic window were subjected to a prograde metamorphism.

Acknowledgements—We should like to thank Nick Skinner for invaluable help in the English version of the manuscript. We should also like to extend our thanks to the two anonymous referees whose comments and suggestions helped to clarify and improve it.

REFERENCES

Armstrong, R. L. & Dick, H. J. B. 1974. A model for the development of thin overthrust sheets of crystalline rock. *Geology* **2**, 35–40.

Bard, J. P., Burg, J. P., Matte, Ph. & Ribeiro, A. 1980. La chaîne hercynienne d'Europe occidentale en termes de tectonique des plaques. In: *Géologie de l'Europe du Précambrien aux bassins sédimentaires post-hercyniens* (26 Congrés Géologique International, Colloque C6). *Mém. Bur. Rech. geol. min. Fr.* **108**, 233–246.

Bard, J. P., Capdevila, R., Matte, Ph. & Ribeiro, A. 1972. Le Précambrien de la Meseta Ibérique. *Not. Mem. Serv. Mines Carte géol. Maroc* **236**, 315–335.

Barrera, J. L., Bellido, F., Pablo Maciá, J. G. de & Arps, C. E. S. 1982. Evolución petrológico geoquímica de los granitoides hercínicos del NO gallego. *Cuad. Lab. Xeol. de Laxe*, **3**, 21–52.

Bastida, F. 1981. Medida de la deformación a partir de pliegues paralelos aplastados. *Trab. Geol.* University of Oviedo, **11**, 15–33.

Bastida, F., Marcos A., Marquínez, J., Martínez-Catalán, J. R., Pérez-Estaun, A. & Pulgar, J. A. 1983. La Coruña (1). *Mapa Geológico de España*. E. 1:200,000. I.G.M.E. Madrid.

Bastida, F., Marcos, A., Pérez-Estaun, A. & Pulgar, J. A. 1979. Aproximación a la mecánica de los cabalgamientos en el contexto general de la deformación en el NW de la Península Ibérica. *Acta geol. hispanica* **14**, 135–142.

Bastida, F. & Pulgar, J. A. 1978. La estructura del manto de Mondoñedo entre Burela y Tapia de Casariego (Costa Cantábrica, NW de España). *Trab. Geol.* University of Oviedo, **10**, 75–124.

Bayer, R. & Matte, Ph. 1979. Is the mafic/ultramafic massif of Cabo Ortegal (northwest Spain) a nappe emplaced during a Variscan obduction? A new gravity interpretation. *Tectonophysics* **11**, T9–18.

Capdevila, R. 1969. Le métamorphisme régional progressif et les granites dans le segment hercynien de Galice nord oriental (NW de l'Espagne). Thesis, University of Montpellier.

Capdevila, R. & Floor, P. 1970. Les differents types de granites hercyniens et leur distribution dans le Nord-Ouest de l'Espagne. *Bol. Geol. Min. Inst. Geol. Min. España* **81**, 215–225.

Capdevila, R. & Vialette, Y. 1970. Estimation radiométrique de l'âge de la deuxième phase tectonique hercynienne en Galice moyenne (Nord-Ouest de l'Espagne). *C. r. hebd. Séanc. Acad. Sci. Paris*, **270**, 2527–2530.

Crimes, T. P., Marcos, A. & Pérez-Estaun, A. 1974. Upper Ordovician turbidites in Western Asturias: a facies analysis with particular reference to vertical and lateral variations. *Palaeogeogr., Palaeoclimat., Palaeoecol.* **15**, 169–184.

Cuthbert, S. J., Harvey, M. A. & Carswell, D. A. 1983. A tectonic model for the metamorphic evolution of the Basal Gneiss Complex, Western South Norway. *J. metamor. Geol.* **1**, 63–90.

Dahlstrom, C. D. A. 1970. Structural Geology in the Eastern margin of the Canadian Rocky Mountains. *Bull. Can. Petrol. Geol.* **18**, 332–406.

Dewey, J. F. & Burke, K. C. A. 1973. Tibetan, Variscan and Precambrian basement reactivation: products of continental collision. *J. Geol.* **81**, 683–692.

Elliott, D. 1978. Lectures on the structural geometry and evolution of thrusts. Course at the University of Lausanne.

González Lodeiro, F. 1980. Estudio geológico estructural de la terminación oriental de la Sierra del Guadarrama (Sistema Central Español). Unpublished thesis, University of Salamanca.

González Lodeiro, F., Hernandez Urroz, J., Klein, E., Martínez-Catalán, J. R. & Pablo Maciá, J. G. de 1982. Lugo (8). *Mapa Geológico de España*. E. 1:200,000. I.G.M.E. Madrid.

Hansen, E. 1971. *Strain Facies*. Springer, Berlin.

Hsü, K. J. 1979. Thin-skinned plate tectonics during Neo-Alpine orogenesis. *Am. J. Sci.* **279**, 353–366.

Hudleston, P. J. 1973. Fold morphology and some geometrical implications of theories of fold development. *Tectonophysics* **16**, 1–46.

Iglesias, M., Matte, Ph., Ribeiro, A. & Ribeiro, L. 1981. Mise en place par obduction des nappes varisques du Nord-Ouest ibérique. *Terra Cognita* **1**, 70.

Johnson, A. M. & Honea, E. 1975. A theory of concentric, kink, and sinusoidal folding and of monoclinal flexuring of compressible, elastic multilayers — III. Transition from sinusoidal to concentric-like to chevron folds. *Tectonophysics* **27**, 1–38.

Julivert, M. 1971. Decollement tectonics in the Hercynian Cordillera of Northwest Spain. *Am. J. Sci.* **270**, 1–29.

Julivert, M., Fontboté, J. M., Ribeiro, A. & Conde, L. 1972. Mapa tectónico de la Península Ibérica y Baleares. E. 1:1,000,000. I.G.M.E. Madrid.

Julivert, M. & Marcos, A. 1973. Superimposed folding under flexural conditions in the Cantabrian zone (Hercynian Cordillera, NW Spain). *Am. J. Sci.* **273**, 353–375.

Julivert, M., Martínez, F. J. & Ribeiro, A. 1980. The Iberian segment of the European Hercynian foldbelt. In: *Géologie de l'Europe du Précambrien aux bassins sédimentaires post-hercyniens* (26 Congrés Géologique International, Colloque C6). *Mém. Bull. Rech. geol. min. Fr.* **108**, 132–158.

Le Fort, P. 1975. Himalayas, the collided range; present knowledge of the continental arc. *Am. J. Sci.* **275**, 1–44.

Marcos, A. 1970. Sobre la presencia de un flysch Ordovícico Superior en el occidente de Asturias (NW de España). *Breviora geol. Astúr.* **14**, 13–28.

Marcos, A. 1971. Cabalgamientos y estructuras menores asociadas originados en el transcurso de una nueva fase herciniana de deformación en el occidente de Asturias (NW de España). *Breviora geol. astúr.* **15**, 59–64.

Marcos, A. 1973. Las Series del Paleozoico Inferior y la estructura herciniana del occidente de Asturias (NW de España). *Trab. Geol.* University of Oviedo **6**, 1–113.

Marcos, A. & Pulgar, J. A. 1982. An approach to the tectonostratigraphic evolution of the Cantabrian Foreland thrust and fold belt, Hercynian Cordillera of NW Spain. *Neues Jb. Geol. Paläont. Abh.* **163**, 256–260.

Martínez–Catalán, J. R. 1979. La prolongación del Manto de Mondoñedo en la zona occidental del domo de Lugo (Galicia, España). *Brevior. geol. astúr.* **23**, 17–22.

Martínez–Catalán, J. R. 1980. L'apparition du chevauchement basal de la nappe de Mondoñedo dans le dôme de Lugo (Galice, Espagne). *C. r. hebd. Séanc. Acad. Sci. Paris* **290**, 179–182.

Martínez–Catalán, J. R. 1981. Estratigrafía y estructura del domo de Lugo (Sector Oeste de la zona Asturoccidental-leonesa). Unpublished thesis, University of Salamanca.

Martínez–Catalán, J. R. 1983. Deformación heterogénea en los macizos graníticos de Sarria y Santa Eulalia de Pena (provincia de Lugo). *Studia geologica salmanticensia* **18**, 39–63.

Martínez–Catalán, J. R., Klein, E., de Pablo Maciá, J. G. & González Lodeiro, F. in press. El Complejo de Ordenes: Subdivisión, descripción y discusión sobre su origen. *Cuad. Lab. Xeol. de Laxe*.

Mattauer, M. 1975. Sur le mécanisme de formation de la schistosité dans l'Himalaya. *Earth Planet. Sci. Lett.* **28**, 144–154.

Matte, Ph. 1968. La structure de la virgation hercynienne de Galice (Espagne). *Revue Géol. alp.* **44**, 1–128.

Navidad, M. 1978. Las series glandulares "Ollo de Sapo" en los sectores Nord-Occidental y Centro Oriental del Macizo Ibérico. *Estudios Geol.* **34**, 511–528.

Parga Pondal, I., Matte, Ph. & Capdevila, R. 1964. Introduction a la géologie de "l'Ollo de Sapo", formation porphyroide antésilurienne du Nord Ouest de l'Espagne. *Not. y Com. del I. G. M. E.* **76**, 119–154.

Parga Pondal, I., Matte, Ph., Capdevila, R., Parga, J. R., Teixeira, C. & Floor, P. 1967. Carte géologique du Nord-Ouest de la Peninsule Ibérique. E. 1:500,000. Servicos Geologicos de Portugal, Lisboa.

Pêcher, A. 1977. Geology of the Nepal Himalaya: deformation and petrography in the MCT zone. In: *Himalaya, Sciences Terre*, C.N.R.S., 301–318.

Pulgar, J. A. 1980. Análisis e interpretación de las estructuras originadas durante las fases de replegamiento en la zona Asturoccidental-leonesa (Cordillera Herciniana, NW de España). Unpublished thesis, University of Oviedo.

Ramsay, J. G. 1967. *Folding and Fracturing of Rocks*. McGraw-Hill, New York.

Ries, A. C. 1979. Variscan metamorphism and K–Ar dates in the Variscan fold belt of S Brittany and NW Spain. *J. Geol. Soc. Lond.* **136**, 89–103.

Soper, N. J. & Barber, A. J. 1982. A model for the deep structure of the Moine thrust zone. *J. Geol. Soc. Lond.* **139**, 127–138.

Toksöz, M. N. & Bird, P. 1977. Modelling of temperatures in continental convergence zones. *Tectonophysics* **41**, 181–193.

Valdiya, K. S. 1980. The two intracrustal boundary thrusts of the Himalaya. *Tectonophysics* **66**, 323–348.

Walter, R. 1966. Resultados de investigaciones geológicas en el Noreste de la provincia de Lugo (NO-España). *Not. y Com. del I. G. M. E.* **89**, 7–16.

Winkler, H. G. F. 1974. *Petrogenesis of Metamorphic Rocks*. Springer, Berlin.