Journal of Structural Geology, Vol. 5, No. 3/4, pp. 431 to 440. 1983
Printed in Great Britain

0191-8141/83 $3.00 + 0.00

Strain distribution within the Pardailhan Nappe (Montagne Noire, France) and structure of its basal thrust zone: implications for events associated with nappe emplacement

LYAL B. HARRIS

Centre Armoricain d'Etude Structurale des Socles (CNRS), Université de Rennes, 35042 Rennes Cédex, France

JEAN-PIERRE BURG and SOPHIE SAUNIAC

Laboratoire de Géologie Structurale Appliquée associé au CNRS N° 266, U.S.T.L., place E.-Bataillon, 34060 Montpellier Cédex, France

(*Received* 7 *July* 1982; *accepted in revised form* 14 *January* 1983)

Abstract—A strain study carried out within the overturned limb of the Pardailhan Nappe (Montagne Noire, France), using deformed trilobites, shows that a general agreement exists between the two-dimensional strain ellipse and the corresponding fabric ellipse determined from the measurement of the preferred orientation of phyllosilicates. The principal extension direction X parallels ENE-trending, first generation F_1 folds, contemporaneous with early stages of nappe emplacement and both are approximately perpendicular to the transport direction. Strain values are low ($X/Y = 1.3$–2.2) and strain gradients within the nappe are small.

Similar strain values are measured, using deformed fossils, along the nappe's basal thrust zone. In this zone F_1 folds are seen to have been flattened and passively rotated towards parallelism with the transport direction within the cleavage plane, by penetrative movement during thrusting. The deformation here is heterogeneous and concentrated within small, discontinuous shear zones. The different microstructures are described and their superposition is interpreted in terms of a progressive tectonic evolution associated with emplacement of the Pardailhan Nappe.

INTRODUCTION—GEOLOGICAL SETTING

THE SOUTHERN Montagne Noire (France) is formed structurally by a pile of Variscan recumbent fold nappes (Geze 1954) overturned to the south (Arthaud *et al.* 1966, Arthaud 1970). Three main units have been established which are (Figs. 1 and 2), from the bottom to the top, the Monts de Faugères, the Montpeyroux and the Pardailhan Nappes. These nappes are characterized by the widespread occurrence of inverted limbs (up to 20 km in length measured perpendicular to the fold hinge) and are composed of fossiliferous (Thoral 1935, Geze 1949), Palaeozoic non-metamorphic or low grade rocks which have undergone a process of continual, progressive deformation during nappe emplacement (Variscan), followed by a later compressional phase (Upper Visean–Lower Stephanian). The resulting structures have been classified into four phases by Arthaud (1970):

Phase 1 was the major phase characterized by an initial horizontal attitude of the S_1 cleavage subparallel to the axial planes of isoclinal F_1 folds. These folds are roughly ENE trending (when not rotated by later phases), with a constant vergence to the south.

Phase 2 was a local phenomenon of sliding on the early S_1 cleavage. The first phase microstructures were reoriented during this late stage of nappe emplacement (Arthaud 1969).

Phase 3 corresponds to local overturning of the nappes towards the south. This phase is the last episode in the sequence of deformation which can be related to the nappe emplacement.

Phase 4 was the second regional phase which produced open folds with steep axial planes parallel to a crenulation cleavage and with axes parallel to a N 070° crenulation lineation.

The geometry of the structures and their interpretation in terms of deformation phases are well known in this region, but the emplacement process and the deformation mechanisms associated with such structures, where large inverted limbs are preserved, are still poorly understood. This is true for other regions such as the Helvetic Nappes (Trumpy 1973, Ramsay 1981). The aim of this contribution is to provide preliminary answers to such problems by studying the strain distribution within the Pardailhan Nappe and the structures developed within its basal thrust zone.

THE PARDAILHAN NAPPE

Microstructures

Several kinds of first-phase microstructures can be observed within the Pardailhan Nappe, the most frequent of which is the ubiquitous and horizontal S_1 cleavage, parallel to the axial planes of F_1 microfolds. The intersection between S_1 and bedding S_0, produces an intersection lineation L_1 parallel to the F_1 fold axes (Fig. 3) within the nappe (Arthaud 1970).

In the nappe, cleavage development is often confined to pelitic layers between the sandstones, though a cleavage front is found in the southern frontal region of the nappes (Fig. 2), beyond which there is no clearly discer-

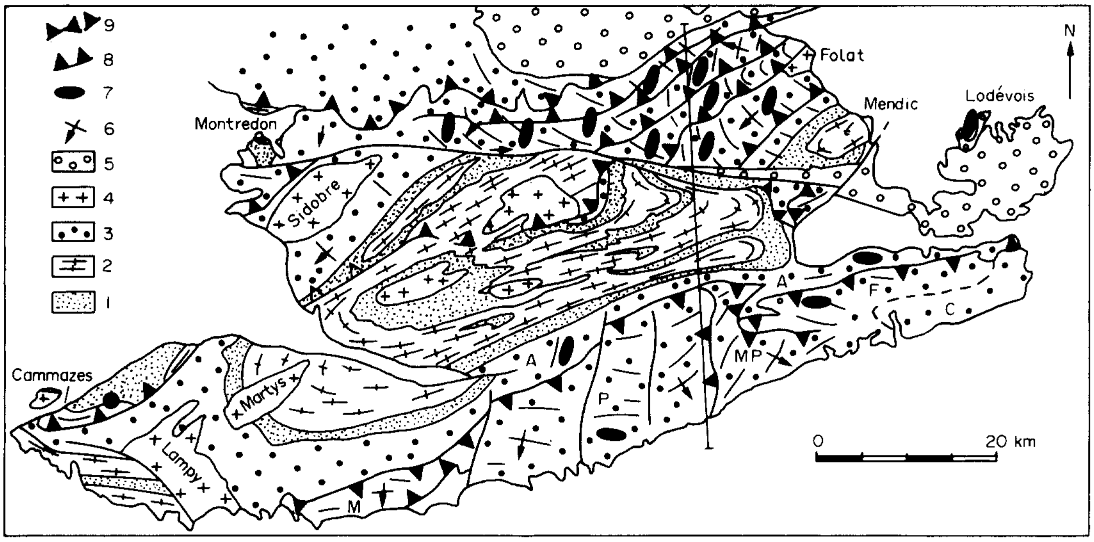

Fig. 1. Structural sketch map of the Montagne Noire. 1, Upper Precambrian pelitic series; 2, feldspathic augengneisses (mainly orthogneisses); 3, Lower to Middle Palaeozoic sediments; 4, Variscan granitoïdes; 5, unconformable Upper Carboniferous and Permian; 6, axial orientation and vergence of F_1 folds; 7, orientation of the finite strain ellipsoid; 8, thrust and base of nappes; 9, probable root zone of the nappes. A, autochthon–para-autochthon; C, Cabrières Nappe; F, Faugères Nappe; MP, Montpeyroux Nappe; M, Minervois Nappe; P, Pardailhan Nappe; —, NS cross-section of Fig. 2.

nible cleavage even within the pelitic layers. Locally the sandstones may have a weak cleavage, consisting of widely spaced pressure solution seams and best developed in the inner arcs of folds. The cleavage development varies with lithology (Fig. 4) and position within the folds. In addition to these local variations, there is an overall increase in the intensity of cleavage from S to N and from top to bottom of the nappe. This regional increase in intensity of cleavage development is accompanied by recrystallization and grain growth and by increase in metamorphic grade to produce chlorite slates.

Fracturing observed in the Acadian sequence within the nappe can be shown to have been developed during late stages of nappe emplacement: veins are seen to postdate S_1. Where bedding is at a steep angle to S_1, a transposed foliation has been produced which is cross-cut by later infilled fractures. Here, the slates have been highly folded and penetrate the limestone layers along a strongly developed fracture cleavage, which results in a transposed foliation along the cleavage plane and the eventual isolation of elongated limestone inclusions within the slate. Formed in this manner, the long axes of limestone inclusions are naturally parallel to the S_0/S_1 intersection direction, thus giving a false impression of extreme elongation parallel to F_1 fold axes.

Within the interbedded slate-limestone sequence, where S_1 is at a slight angle to S_0, no transposed foliation has been produced. Regularly spaced tension fractures, essentially limited to the slate layers, indicate extreme inhomogeneity in deformation. Differential E–W oriented sliding along pre-existing S_1 cleavage planes and, at times, along the slate-limestone interface, accompanied fracture opening during late stages of nappe emplacement (Burg & Harris 1982). Locally these veins were rotated and deformed with the S_1 cleavage within shear zones at the base of the nappe (Sauniac 1980).

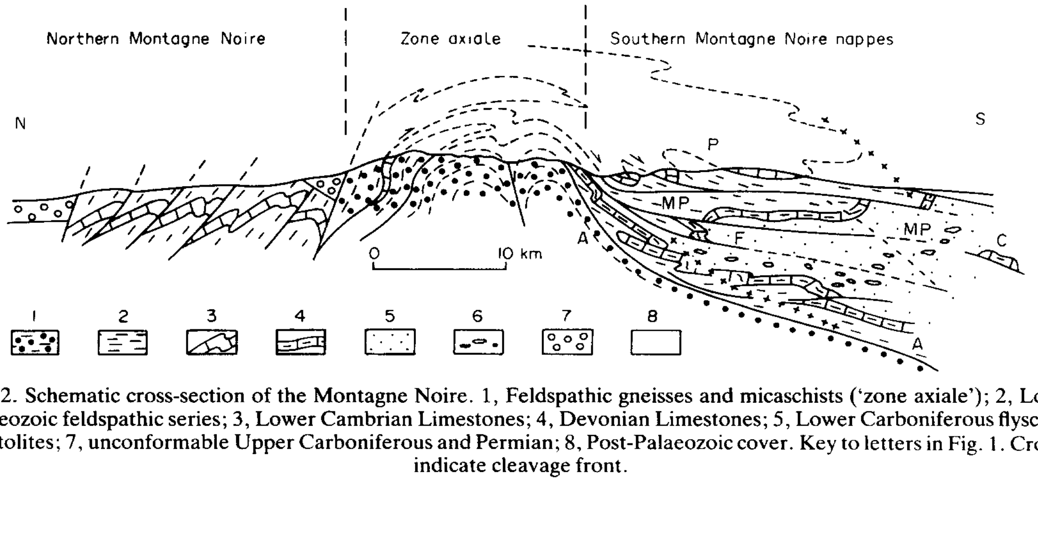

Fig. 2. Schematic cross-section of the Montagne Noire. 1, Feldspathic gneisses and micaschists ('zone axiale'); 2, Lower Palaeozoic feldspathic series; 3, Lower Cambrian Limestones; 4, Devonian Limestones; 5, Lower Carboniferous flysch; 6, Olistolites; 7, unconformable Upper Carboniferous and Permian; 8, Post-Palaeozoic cover. Key to letters in Fig. 1. Crosses indicate cleavage front.

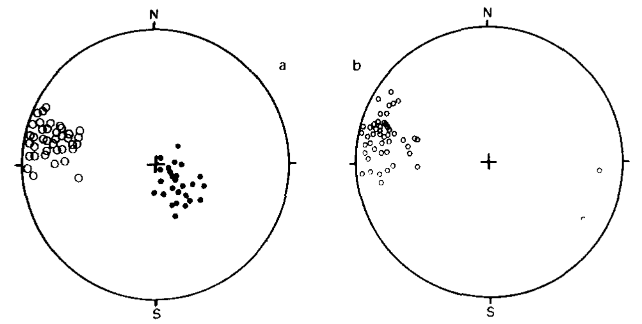

Fig. 3. Relationship between first phase fold axes and quartz-fibre directions in the Pardailhan Nappe: (a) points = 26 poles of S_1; circles = 39 intersection lineations (S_0/S_1) and F_1 fold axes. (b) 47 quartz fibre directions. Schmidt net lower hemisphere. Some data from Arthaud (1970) and Nicollet (1975).

STRAIN STUDY WITHIN THE PARDAILHAN NAPPE

Calculation of the two-dimensional strain ellipse X/Y *from deformed fossils*

Trilobite fossil imprints are abundant in the middle Cambrian shales of the Montagne Noire and have been used to establish a detailed biostratigraphy of the region (Courtessole 1975). These initially symmetrical markers have been deformed during the various stages of nappe emplacement and the methods of Ramsay (1967) can be used to measure the total finite two-dimensional strain in the bedding plane, S_0, for each fossil locality where S_0 is approximately parallel to cleavage S_1 (Harris 1980). We assume here that the cleavage developed perpendicular to the Z axis of the total strain ellipsoid ($X \geq Y \geq Z$), as has been previously shown by Oertel (1970), Cloos (1971), Wood (1973), Tullis & Wood (1975) and Wood & Oertel (1980). Measurements of deformed trilobites will therefore give the strain ellipse in the XY plane.

Calculation of the principal extension direction X from two or more fossils in the same sample (Ramsay 1967) and the alteration of initially circular markers (eyes and tubercles) on the trilobites into ellipses, shows that X is always parallel to L_1. When a single fossil is found in a sample containing a lineation, the method of Ramsay (1967, p. 234) can therefore be used.

Values of X/Y have been calculated from trilobites from the private collection of M. l'Abbé Courtessole (Carcassonne, France) and from samples collected in the field at localities within the Pardailhan Nappe, in the uppermost nappe where the effects of later deformational phases are minimal (Arthaud 1970) and F_1 folds trend E–W (Figs. 1 and 3).

Values of X/Y for each locality are presented in Table 1, whereas the average values for each region are shown on the map in Fig. 5. It can be seen from these data that two-dimensional strain is approximately homogeneous within the bedding planes: there is no two-dimensional bedding-parallel strain gradient either on a large scale within the nappe or on the outcrop scale. The average value of X/Y lies between 1.3 and 2.0. The internally consistent strain values determined at two neighbouring localities may show a large difference in magnitude (for example localities at Coulouma or Ferrals, Table 1). In some cases, there is a large difference in strain values determined from samples from the one outcrop; one or more samples may be at variance with otherwise consistent strain calculations (e.g. the Coulouma samples).

THE FABRIC ELLIPSOID FROM TEXTURE GONIOMETRY

The preferred orientation of phyllosilicates in a fine grained rock can be measured using an X-Ray Texture Goniometer in reflection mode following the methods described in detail by Baker *et al.* (1969), Means & Paterson (1966), Lipshie *et al.* (1976), Wood *et al.* (1976) and Gapais (1981). Assuming an initially random distribution of phyllosilicates, their preferred orientation in a deformed rock can be taken as a measure of the total strain the rock has undergone, following the March model for passive rotation of platelets (March 1932, Means & Paterson 1966). Using the above model, a

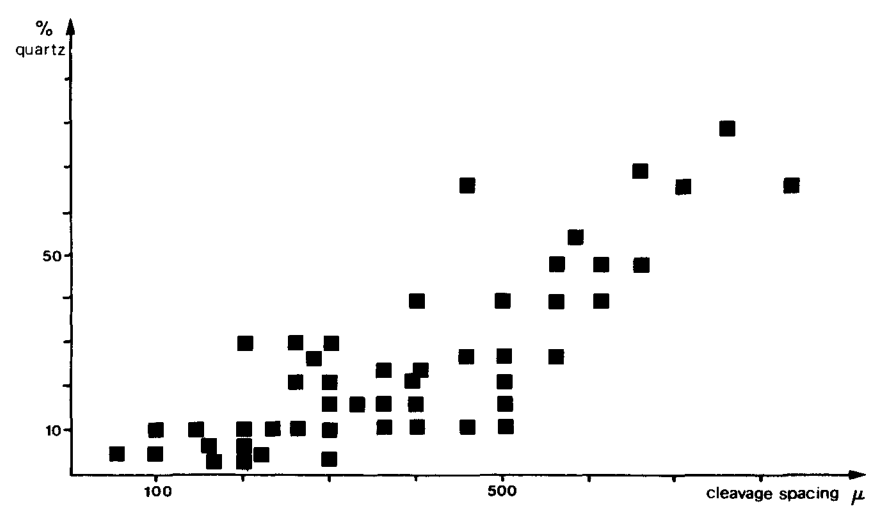

Fig. 4. Cleavage spacing vs quartz content.

Table 1. X/Y values for each fossil locality (second values for BTC include an extreme value of $X/Y = 2.8$)

Region	Fossil locality	Number of measures	Average X/Y
Coulouma	Banc de Touring Club (BTC)	3	1.9
		(4)	2.2
	1ère ravine	7	1.3
	2ème ravine	8	1.5
	Château	1	1.6
	Col de Treille	2	1.4
	Draille Roc Coupe (DRC)	3	2.2
	Vigne Cauquil	4	1.6
	Route à St Jean de Minervois	7	1.6
	Pt. 10	1	1.3
	Pt. 23	1	1.3
	Total (without 2 extremes)	39	1.5
	Total with BTC = 7.8 and DRC = 6.8	41	1.6
Ferrals-les-Montagnes	Route d'Autheze Point 5	6	1.7
	Route d'Autheze Point 6	5	1.8
	Route d'Autheze Point 8	2	1.7
	Route d'Autheze Point 13	1	1.4
	L'Eglise	7	1.7
	Total	21	1.7
Velieux		6	1.5
	Pt. 15	2	1.7
	11	1	1.1
	14	5	1.6
	16	2	1.7
	Total	16	1.6
Minerve	Pompage 1	2	2.0
	2	2	2.3
	3	3	1.6
	Total	7	1.9
Camprafraud		2	1.6
La Fraise		2	1.3
Cassagnoles		3	1.8
St Colombe		3	1.7
Merlaux		1	1.4
Autheze		2	1.6
Faveyroles	Point 19	5	1.5
Rieussec	18	4	2.1
Fallieres	8 road to Faveyroles	4	2.0
Bois du Vidal		4	1.3

Point numbers refer to localities of Courtessole (1975).

consistent relationship between the magnitude and direction of strains derived from texture goniometry and those determined from deformed natural markers has been found (Wood 1974, Oertel & Wood 1974, Tullis & Wood 1975, Wood *et al.* 1976, Wood & Oertel 1980).

Wood & Oertel (1980) have found that natural strains from texture goniometry are always slightly smaller than those given by deformed markers. In other cases, especially where cleavages are crenulate or where there has been an interaction between platelets, the March model has been shown to be invalid (Siddans 1976, Oertel 1970, Etheridge & Oertel 1979). The excellent trilobite strain markers from the Montagne Noire therefore provide another case study to test the reliability of the texture goniometer method.

Results

Texture goniometer calculations of the ratios of fabric ellipsoid axes X/Y and Y/Z along with the values of:

$$k = (X/Y - 1)/(Y/Z - 1) \qquad (1)$$

$$r = (X/Y) + (Y/Z) - 1 \qquad (2)$$

are presented in Figs. 6 (a) and (b).

Comparing values of X/Y calculated from deformed trilobites with the X/Y ratio of fabric ellipsoid axes (Fig. 5a), it can be seen that there is good agreement between the two, except at Minerve and Rieussec, where strain calculations from fossils are greater. It is interesting to note that in these last two cases, texture goniometer measurements give an X/Y value approximating that of

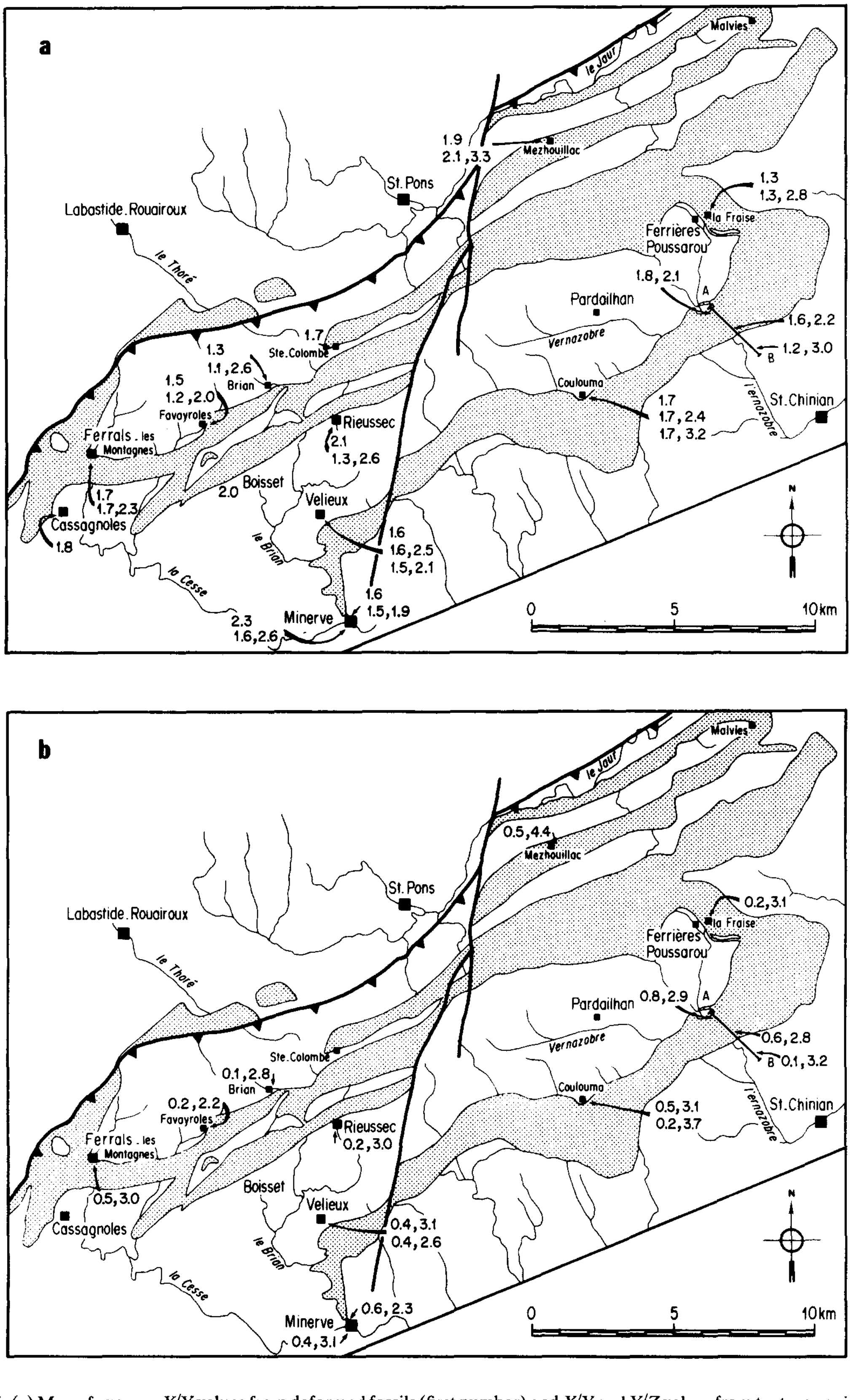

Fig. 5. (a) Map of average X/Y values from deformed fossils (first number) and X/Y and Y/Z values from texture goniometry (second data pairs) for each region of fossil localities (in some cases, two calculations of X/Y and Y/Z from different fossil localities are presented for a given region). (b) Map of k (first number in each data pair) and r (second number). Shaded area corresponds to Middle Cambrian outcrops in the Pardailhan Nappe.

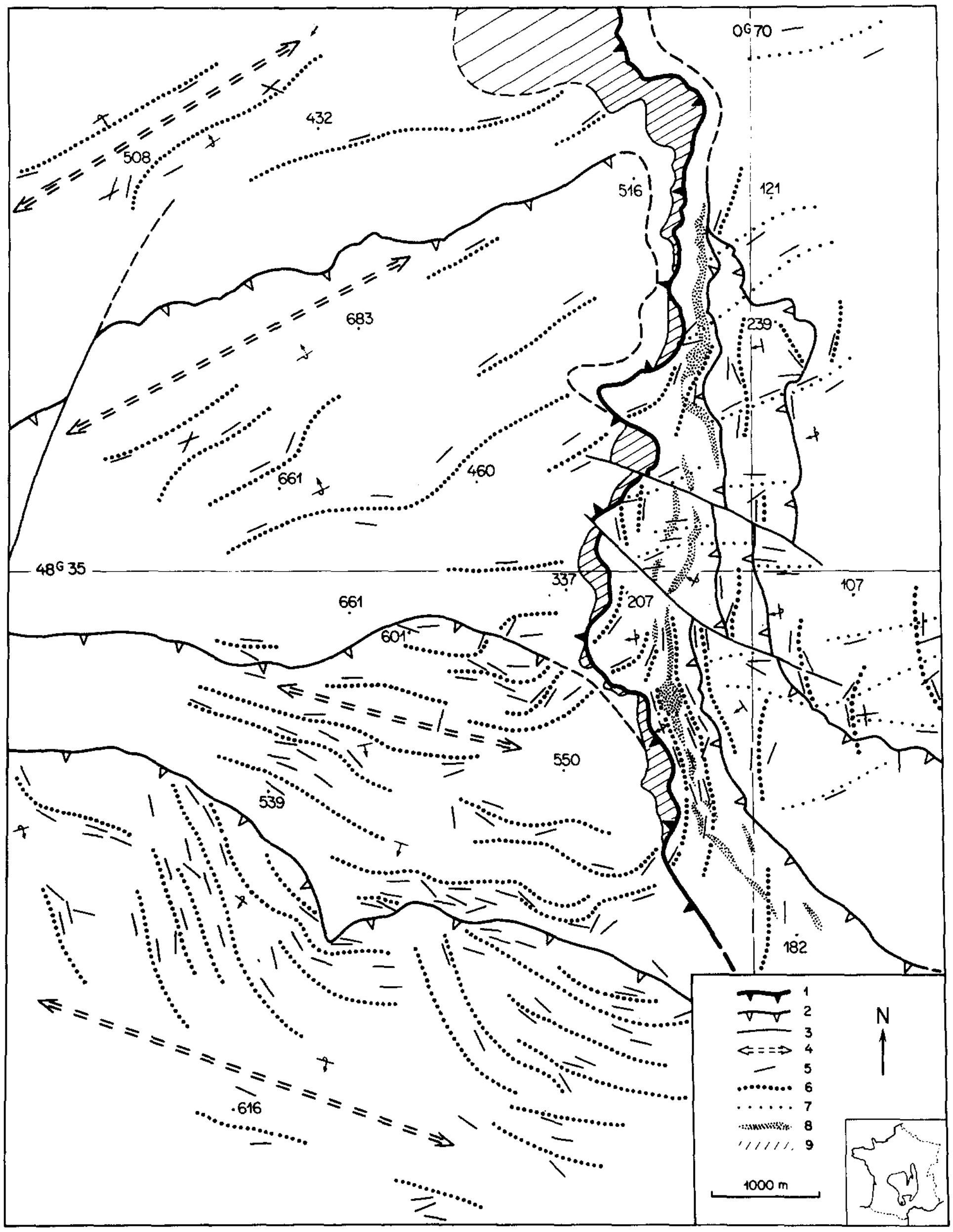

Fig. 6. Structural map of studied area of the basal thrust zone. 1, sole of Pardailhan Nappe; 2, internal thrusts in Pardailhan Nappe and para-autochthon; 3, late faults; 4, hinges of anticlinal recumbent folds; 5, direction of lineations; 6, trajectories of first phase lineations; 7, trajectories of second phase lineation; 8, Ordovician quartzite; 9, Devonian limestones.

neighbouring localities and do not record the abnormally large strain calculated from trilobites.

At first sight the increased deformation of a fossil compared with the matrix of the host rock is unrealistic if the deformation is homogeneous at all scales. We infer therefore that strain may vary locally due to slight differences in lithology (especially in quartz content) or due to the presence of discrete planes of slip.

The values of Y/Z (averaging 2.6) do not vary greatly between localities and no systematic regional gradient is observed. Again where two similar samples are taken a short distance from within the same general area (as at Velieux and Minerve), a slight difference exists between Y/Z values, showing an inhomogeneity of strain at this scale.

A similar situation is seen for values of k (which show that the rocks are in the flattening field of deformation) and r (Fig. 5b) as for Y/Z.

Summary of strain data

Strain values are seemingly small for such a structural setting, bearing in mind that they come from the overturned limb of the fold nappe. The X direction is however E–W oriented parallel to F_1 fold axes and roughly perpendicular to the direction of nappe emplacement (Arthaud *et al.* 1966). If this regular E–W extension is due to extension along a frontal zone of the nappe (Arthaud 1969, 1970, Matte 1968), we should observe a gradient of deformation with increasing E–W elongation in this zone. This is not observed and other reasons must be found to explain this phenomenon.

GEOMETRICAL ANALYSIS IN THE BASAL THRUST ZONE

This study has been carried out along the sole of the Pardailhan nappe, for which the thrust strikes approximately NS and dips about 20° towards the west in a transverse flexure (Fig. 6). The country rocks above and below the thrust are Ordovician quartzo-pelitic series from a thick monotonous and fossiliferous flyschoid sequence (Thoral 1935, Geze 1949). Shreds of Devonian limestones indicate precisely the location of the thrust in the Lower Ordovician pile (Bedarieux 1/80 000, and St Chinian 1/50 000 map sheets). Below the nappe, a boudinaged quartzite runs parallel to the basal contact for over 5 km, indicating parallelism of the regional stratigraphic bedding, the thrust plane and the regional attitude of the S_1 cleavage.

The thrust is a zone of deformation some tens of metres in thickness, which consists either of a crenulation cleavage zone (S_3 of Arthaud 1970) or of numerous, discontinuous shear zones 10 cm to 1 m in width, anastomosing around approximately lozenge-shaped blocks of country rocks which contain 'first phase' deformation features. The thrust zone affects a wider slice of rocks in the underlying series that in the nappe itself. Within these shear zones, the S_1 planes are bent. Slickensides and quartz fibres at a small angle to the shear zone indicate that slip occurs on the pre-existing S_1 cleavage. In the homogeneous parts of these shear zones, a second cleavage S_2 along which no movement has occurred developed oblique to S_1. Both S_1 and slip planes define asymmetric lens-shaped microstructures and movements along these surfaces have led to the formation of gashes infilled with quartz. The quartz veins may be folded around tight to isoclinal parasitic folds with axial planes parallel to S_2. The hinges of these folds are E–W oriented and parallel to an intersection lineation L_2 (Fig. 6) which is developed in the thrust zone. The different microtectonic observations suggest that the slip planes and S_2 are synchronous. Elongate chlorites and quartz grains locally define a mineral lineation within S_2, parallel to L_2.

Using several qualitative criteria, Sauniac (1980) has described a rough strain zonation across this thrust zone. She has shown that the quartz vein density increases with the intensity of shearing and that vein shapes and orientations can be used to determine the shear sense, showing that the upper tectonic unit was displaced towards the south.

ROTATION OF FIRST PHASE MICROSTRUCTURES

Regional fold axial trends

A systematic variation in the F_1 axial trends is recorded at the base of the Pardailhan Nappe (Figs. 6 and 7). With increasing proximity to the thrust zone, most of the observed folds have curving hinge lines and widely dispersed trends (Fig. 6). The F_1 axial trends within the nappe appear to rotate in an anticlockwise sense (Fig. 6) until the axes plunge at a low angle to the north or south (Fig. 7). Tension gashes comparable to those higher in the nappe (see above) and the fibres they contain are not generally folded but seem to be passively rotated with the fold axes and the intersection lineation L_1 on the S_1 cleavage. This secondary rotation of the first phase microstructures and the progressive curvature correspond to Phase 2 of Arthaud (1970).

Changes in F_1 *fold fabrics*

Geometrical analysis (Ramsay 1967) shows that minor folds in the nappe (Fig. 8a) belong to classes 1B and 1C for the sandstones and to classes 1A, B and C for slates, depending on bed thickness and control of folding by sandstone layers. By comparison, in the thrust zone (Fig. 8b), the rotated non-cylindrical folds belong to classes 1B and 1C for the sandstones and to classes 1C and 2 for the slates. These geometrical changes are accompanied by changes in the interlimb angles which range from 150 to 50° in the nappe to between 100 and 35° in the thrust zone. All these features indicate that,

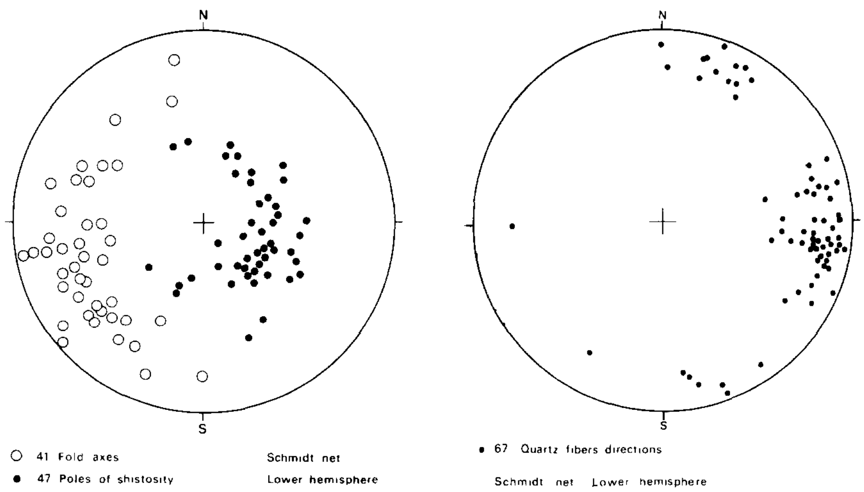

Fig. 7. (a) Orientation pattern of F_1 folds and S_1 schistosity in the thrust zone. (b) Quartz fibre directions in the thrust zone.

SG 5:3/4-N

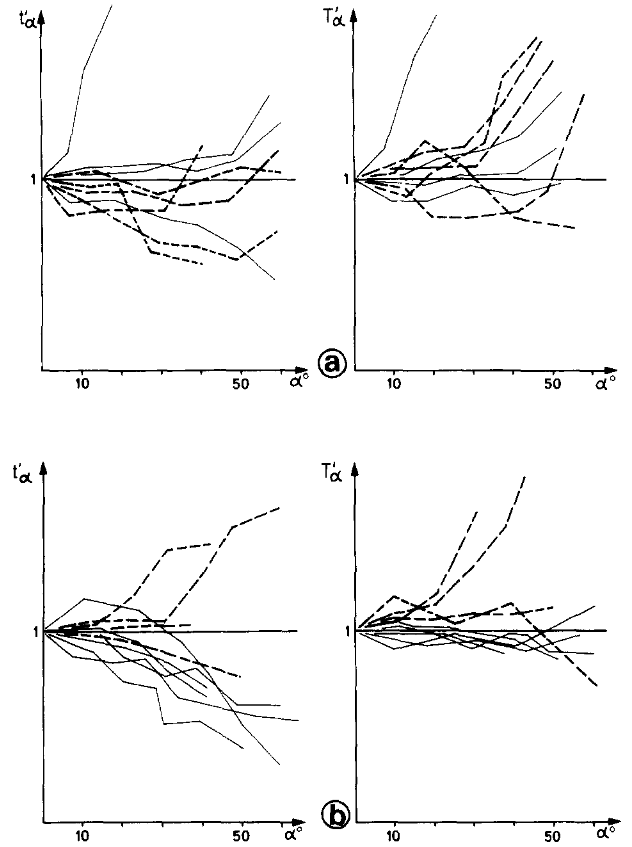

Fig. 8. Plots of t'_α and T'_α (Ramsay 1967, p. 361) of F_1 folds (a) in Pardailhan Nappe; (b) in the thrust zone. Dashed lines: folded sandstones; thin lines: folded schists. Some data from Darimont (1974).

during progressive deformation and northward rotation, the folds tightened. Similar variations in fold geometry associated with their reorientation have been described by Berthé & Brun (1980).

MICROSTRUCTURES IN THE THRUST ZONE

The S_2 *cleavage*

The average spacing of the S_2 cleavage decreases towards the most deformed parts of the thrust zone with a discrete crenulation cleavage evolving into a slaty cleavage (Gray 1977, Powell 1979). Discrete crenulation cleavage occurs in the more pelitic layers with a dip greater than S_0 and S_1, whereas it is only weakly developed in the more silty layers. From the margin to the centre of the anastomosing shear zones, the cleavage domains occupy a larger and larger proportion of the rock. In the centre, the S_2 cleavage is almost parallel to S_0 and S_1 and overprints those microstructures. Along the S_2 cleavage, there is an increase in relative content of phyllosilicates and metal oxides. In the quartzites, the development of the S_2 cleavage involves a change in grain shape fabric. Within shear zones, length-width ratios of quartz grains in the S_2 cleavage zone are larger than those outside the shear zone. In the cleavage zone, the flattening index (Krumbein & Graybill 1965) equals 2.20 whereas, in the bulk of the rock, it equals 1.51.

Slip planes

One remarkable feature throughout the basal region of the Pardailhan Nappe is the presence of many discontinuous shear zones. The first phase cleavage is bent and, in the centre of the shear zone, slickenside striations trending N020–N030° show that S_1 has been reactivated as a slip plane. Thin sections show finely crushed and dark gouges of essentially chloritic material, resulting from brittle failure, repeated fracture and rotation of rock particles. Chlorites, formed within cracks between fractured grains, optically resemble those in the S_1 cleavage within the adjacent rock. From our observations, it is obvious that brittle deformation along slip planes is important during the development of these shear zones.

Quartz veins

In addition to the tension fractures previously described, there are veins associated with the second phase and lying within the S_1 cleavages. Two main fibre directions in these veins have been measured; one is roughly N–S parallel to the striae in the gouges, the other is N060 to N110° (Fig. 7b). No microtectonic criteria have been found which indicate time relationships between these two sets. In sections normal to both S_1 and S_2 and parallel to the fibres, these veins have asymmetric lens shapes and may be refolded. They range from a few centimetres up to 20 centimetres in length, and up to several centimetres in width. In sections orthogonal to the fibre direction, veins appear to be long (up to 1 m) and symmetric.

The deformation in the shear zones is non-coaxial, in a direction parallel to the slickenside striations on S_1 and to the N–S fibres of the second phase veins. Each site of crystallization is clearly associated with an area where sliding on S_1 has created a potential sink. The asymmetric lens shape of the veins can be used to determine a sense and a direction of displacement towards the south (Sauniac 1980). Pressure solution of the country rock during deformation has provided the material for vein infill.

Numerous secondary fluid inclusions within vein quartz have been used to obtain the minimum fluid pressure during their formation. Results indicate epizonal conditions of deformation (Guiraud *et al.* 1981).

Finite strain in two dimensions

Sixty three fossils (*Bellerophon oehlerti*, a gastropod) have been collected in a pelitic bed 100 m below the shear contact of the Pardailhan Nappe. Undeformed fossils have logarithmic spirals and bilateral symmetry. During the deformation, their umbilicals have been twisted and Blake's method (Blake 1878) has been used to give an approximate value of $X/Y = 1.3$.

Along the same contact, measurements of fifty-three ooids in the Devonian limestones give $X/Y = 1.35$ and $Y/Z = 1.88$. The principal direction (X) parallels L_1 which has been rotated in this area to an approximately

N–S orientation (Fig. 3). These finite strain values are not fundamentally different from those in upper levels of the nappe, in spite of the different lithologies involved. Presumably gliding on discrete S_1 surfaces was more important during late stages of deformation than penetrative deformation.

DISCUSSION AND CONCLUSION

From our study of the Pardailhan Nappe, we may consider the following stages of nappe emplacement.

(a) Isoclinal folding about horizontal axial surfaces with slight extension parallel to the axes of the folds. The inverted limbs must be formed at these stages as the late phases of deformation affect already inverted strata.

(b) Nappe displacement along a thrust zone in which the first phase linear elements are rotated towards parallelism with the N–S transport direction.

We know that the main problem with all the nappes of the Montagne Noire is the existence of extensive inverted limbs. We have seen that in the Pardailhan Nappe, finite strain values associated with the first stages of deformation are small and homogeneous. Furthermore, in the nappe itself, only one cleavage of constant attitude occurs. None of these data could therefore indicate rolling of the major recumbent fold hinge, possessing an axial plane cleavage, as being reponsible for the development of long overturned limbs.

Even though a change in intensity of cleavage is observed, there is no corresponding change in strain magnitude. This therefore implies that only a change in deformation mechanisms and not intensity of deformation takes place from S to N. As no late deformation affects the higher part of the nappe, E–W extension parallel to axes of the F_1 folds is a feature of the initial formation of recumbent folding.

The analysis of the microstructures in the nappe sole shows that during the waning stages of the emplacement processes the deformation is fundamentally heterogeneous and concentrated within shear zones. The fold hinges, showing a divergent trend from the associated host fold on regional scale, can be accounted for in the continuing strain subsequent to the initiation of the fold hinge. The minor folds parallel to the direction of transport are formed by tightening, flattening and passive rotation of earlier more open folds, originally perpendicular to the direction of transport (see also Berthé & Brun 1980). This rotation takes place within the S_1 cleavage by movement during thrusting.

Several authors have described lineations parallel to nappe transport directions (Kvale 1953, Bryant & Reed 1969, Sanderson 1973, Escher & Waterson 1974, Mattauer 1975) where the rotation of fold axes towards parallelism with the transport directions is thought to be consistent with a simple shear hypothesis. This is also true for the low grade metamorphic nappes in the Montagne Noire. In the proximity of a thrust plane, axial trend parallel to the transverse extension lineation implies an increase in the finite shear component of the deformation. The systematic anticlockwise sense of rotation of fold axes could be interpreted in terms of another shear component acting at an oblique angle to the original mean direction of Y, possibly related to the existence of a lateral boundary controlling the emplacement of nappes (Brun 1977).

At the final stage of the nappe emplacement, we infer an E–W trend of movement from the E–W trend of slickenside striations and fibres. This is also in agreement with the late E–W translation along S_1 seen from tension fractures within the nappe (Burg & Harris 1982), but does not agree with the regional N–S displacement demonstrated by analysing the dispersion of fold axes (Arthaud 1969).

Acknowledgements—F. Arthaud, J. P. Brun and P. Choukroune are thanked for constructive criticisms of the various drafts of this paper. P. Matte is especially thanked for his help in the initiation of the strain study. We are grateful to R. Courtessole for kindly providing access to his private collection of trilobite fossils from the Montagne Noire.

REFERENCES

Arthaud, F. 1969. Un exemple de relations entre l'étirement dans B, la dispersion des linéations et la courbure des axes de plis: la virgation des plis couchés hercyniens du versant sud de la Montagne Noire. *Rev. Géol. dyn. Géogr. phys.* **11**, 523–532.

Arthaud, F. 1970. Etude tectonique et microtectonique comparée de deux domaines hercyniens: les nappes de la Montagne Noire et l'anticlinorium de l'Iglesiente. Pub. USTELA Montpellier. *Ser. Géologie structurale* **1**, 175.

Arthaud, F., Mattauer, M. & Proust, F. 1966. La structure et la microtectonique des nappes hercyniennes de la Montagne Noire. Coll. Etages tectoniques, Neuchâtel, 231–243.

Baker, D. W., Wenk, H. R. & Christie, J. M. 1969. X-ray analysis of preferred orientation in fine grained quartz aggregates. *J. Geol.* **77**, 144–172.

Berthé, D. & Brun, J. P. 1980. Folds formed by progressive shear in the South Armorican Shear Zone. *J. Struct. Geol.* **2**, 127–133.

Blake, J. F. 1878. On the measurements of the curves formed by cephalopods and other mollusks. *Phil. Mag.* **5**, 241–262.

Brun, J. P. 1977. Cumulative strain and boundary effects in the gravity flow of a viscous slab. *Tectonophysics* **41**, T7–714.

Bryant, B. & Reed, J. C. 1969. Significance of lineation and minor folds near major thrust faults in the Southern Appalachian and the British and Norwegian Caledonides. *Geol. Mag.* **106**, 412–429.

Burg, J. P. & Harris, L. B. 1982. Tension fractures and boudinage oblique to the maximum extension direction: an analogy with Lüders Bands. *Tectonophysics* **83**, 347–363.

Cloos, E. 1971. Microtectonics along the western edge of the Blue Ridge, Maryland & Virginia. Johns Hopkins Press, Baltimore.

Courtessole, R. 1975. Le Cambrien Moyen de la Montagne Noire. Imprimerie d'Oc, Toulouse.

Darimont, A. 1974. Analyse de la géométrie des plis de la phase 1 dans l'Ordovicien de la Nappe de Pardailhan (Montagne Noire). Unpublished D.E.A. thesis, Montpellier.

Escher, A. & Watterson, J. 1974. Stretching fabrics, folds and crustal shortening. *Tectonophysics* **22**, 223–231.

Etheridge, M. A. & Oertel, G. 1979. Strain measurements from phyllosilicate preferred orientation, a precautionary note. *Tectonophysics* **60**, 107–120.

Gapais, D. 1979. Orientations préferentielles, de réseau et déformations naturelles. Applications aux quartzites (segment hercynien de Bretagne centrale) et aux amphibolites (couverture des dômes gneissiques de Kuopio, Finlande). Thèse 3e Cycle, Rennes.

Geze, B. 1949. Etude géologique de la Montagne Noire et les Cévennes méridionales. *Mém. Soc. Géol. Fr. Paris* XXIX (62), 215.

Geze, B. 1954. A propos des nappes du versant méridional de la Montagne Noire (Hérault). *C. r. somm. Soc. Géol. Fr*, 242–245.

Gray, D. R. 1977. Morphologic classification of crenulation cleavage. *J. Geol.* **85**, 229–235.

Guiraud, M., Sauniac, S. & Burg, J. P. 1981. Précisions sur les conditions pression-température lors de la mise en place de la Nappe de Pardailhan (Montagne Noire) par la détermination des inclusions fluides. *C. r. hebd. Séanc. Acad. Sci. Paris* Ser. II **292**, 229–232.

Harris, L. B. 1980. Etude de la déformation finie et des fentes syntectoniques dans les schistes du Cambrien moyen de la nappe de Pardailhan, Montagne Noire, France. Unpublished D.E.A. thesis Montpellier.

Krumbein, W. C. & Graybill, F. A. 1965. *An Introduction to Statistical Models in Geology*. McGraw-Hill, New York.

Kvale, A. 1953. Linear structures and their relation to movement in the Caledonides of Scandinavia and Scotland. *Q. Jl geol. Soc. London* **109**, 51–74.

Lipshie, S. R., Oertel, G. & Christie, J. M. 1976. Measurements of preferred orientation of phyllosilicates in schists. *Tectonophysics* **34**, 91–99.

March, A. 1932. Mathematische Theorie der Ragelung nach der Korngestalt bei Affiner Deformation. *Z. Krist.* **81**, 285–295.

Mattauer, M. 1975. Sur le mécanisme de formation de la schistosité dans l'Himalaya. *E.P.S.L.* **28**, 144–154.

Matte, P. 1968. La structure de la virgation hercynienne de Galice (Espagne). Extrait des travaux du laboratoire de géologie de la Faculté des Sciences de Grenoble. Tome 44, 128p.

Means, W. & Patterson, M. S. 1966. Experiments on preferred orientation of platy minerals. *Contr. Miner. Petrol.* **13**, 108–133.

Nicollet, C. 1975. Fentes de tension, cristallisations syncinématiques et déformation: deux exemples pris dans les structures hercyniennes de la Montagne Noire. Unpublished DEA thesis, Montpellier.

Oertel, G. 1970. Deformation of a slaty, lapillar tuft in the Lake District, England. *Bull. geol. Soc. Am.* **8**, 1173–1188.

Oertel, G. & Woods, D. S. 1974. Finite strain measurement: a comparison of methods. *Trans. Am. Geophys. Union* (abst.) **55** (7), 695.

Powell, C. McA. 1979. A morphological classification of rock cleavage. *Tectonophysics* **58**, 21–34.

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

Ramsay, J. G. 1981. Tectonics of the Helvetic Nappes. In: *Thrust and Nappe Tectonics* (edited by K. R. MacClay & N. J. Price). *Spec. Publ. Geol. Soc. Lond.* **9**, 293–309.

Sanderson, D. J. 1973. The development of fold axes oblique to the regional trend. *Tectonophysics* **16**, 55–70.

Sauniac, S. 1980. Utilisation des exsudats de quartz comme critères de reconnaissance d'un régime cisaillant: exemple de la base de la nappe de Pardailhan (versant sud de la Montagne Noire). *Rev. Géol. Dyn. Geog. Phys.* **22**, 177–186.

Siddans, A. W. B. 1976. Deformed rocks and their textures. *Phil. Trans. R. Soc. Lond.* **A283**, 43–54.

Thoral, H. 1935. Contribution à l'étude géologique des Monts de Lacaune et des terrains cambriens et ordoviciens de la Montagne Noire. *Bull. serv. Carte Géol. Fr.* (192) **38**, 319–637.

Trumpy, R. 1973. The timing of orogenic events in the Central Alps. In: *Gravity and Tectonics* (edited by De Jong, K. A. & Scholten, R.). Wiley, New York, 229–251.

Tullis, T. E. & Wood, D. S. 1975. Correlation of finite strain from both reduction bodies and preferred orientation of mica in a slate from Wales. *Bull. geol. Soc. Am.* **86**, 632–638.

Wood, D. S. 1973. Patterns and magnitude of natural strain in rocks. *Phil. Trans. R. Soc. Lond.* **A274**, 373–382.

Wood, D. S. 1974. Current views on the development of slaty cleavage. *A. Rev. Earth Planet. Sci.* **2**, 369–401.

Wood, D. S. & Oertel, G. 1980. Deformation in the Cambrian Slate Belt of Wales. *J. Geol.* **88**, 285–308.

Wood, D. S., Oertel, G., Singh, J. & Bennett, H. F. 1976. Strain and anisotropy in rocks. *Phil. Trans. R. Soc. Lond.* **A283**, 27–42.