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Identification of Ordovician block-tilting in the Hercynian fold belt of Central Brittany (France): field evidence and computer models

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Abstract—The significance of the Brioverian–Lower Ordovician unconformity in Central Brittany has been discussed by many authors. A new interpretation in terms of extensional tectonics is presented, based on structural, sedimentological and stratigraphical data. However, because sedimentary formations have been strongly folded by Hercynian deformation, extensional structures are obscured and difficult to recognize. To identify the normal faults, the geological map is compared with computer models of the folded tilted blocks. It is shown that major normal faults had a mean N70° trend, dipping toward the SE with an almost regular spacing of 7–8 km.

INTRODUCTION

THE Ordovician succession of Central Brittany lies unconformably upon an unfossiliferous and monotonous Brioverian group of pelites, sandstones and conglomerates. Until recently, the Brioverian supergroup was believed to be of Late Proterozoic age and the unconformity was interpreted either (i) as the result of a Cadomian phase of folding followed by erosion and then sedimentation of Ordovician strata (Chauvel & Phillipot 1957, Cogné 1962, 1972) or (ii) as the result of a synsedimentary deformation of Brioverian sediments by slumping prior to the deposition of an Ordovician succession (Darboux *et al.* 1975, Le Corre 1977). A new interpretation has recently been put forward by Ballard *et al.* (1986) who consider the unconformity to be the result of block faulting and tilting of the Brioverian with synchronous deposition of Lower Ordovician formations.

Both Brioverian and Ordovician rocks have undergone one single phase of folding during the Carboniferous (Le Corre 1978) associated with dextral wrenching on the crustal scale (Gapais & Le Corre 1980, Percevault & Cobbold 1981). As suggested by Ballard *et al.* (1986) geological map contours might reflect the interference resulting from the upright oblique folding of tilted blocks. We present here an analysis of geological map contours using computer models. A 20° obliquity is shown between the normal faults bounding tilted blocks and the fold axis trends. Field data supporting this interpretation are examined and regional geological implications are discussed.

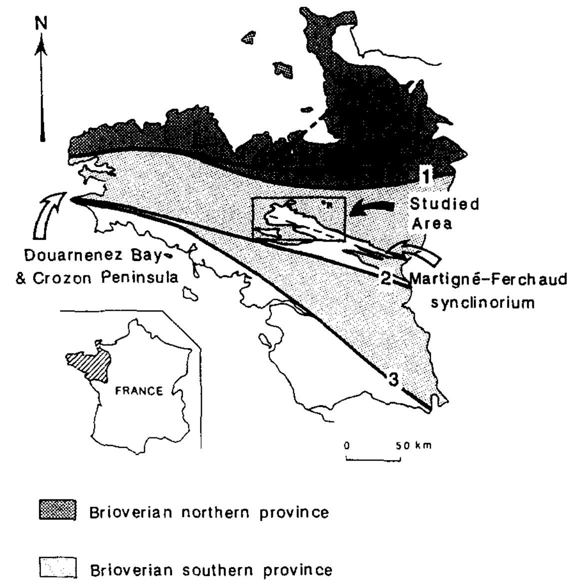

Fig. 1. Location of the study area in Central Brittany. 1, North Armorican shear zone; 2, northern branch of the South Armorican shear zone; 3, southern branch of the South Armorican shear zone.

GEOLOGICAL SETTING

Stratigraphy

The *Brioverian* of Central Brittany was originally defined by Barrois (1895, 1899); it is unfossiliferous and has always been considered as the main stratigraphic group of the Upper Proterozoic (Cogné 1962, 1972). Many authors have tried to establish an internal stratigraphy (Barrois & Pruvost 1929, 1931) but Le Corre (1977) has shown that all facies types can be found at a single outcrop locality. Although it has proven impossible to subdivide the succession, two distinct provinces are separated by the North Armorican shear zone (Fig. 1); a northern province without conglomerates and a southern province which contains conglomerates (Chantraine *et al.* 1982). The main sedimentary facies of the Brioverian in the southern province are, in order of abundance (Le Corre 1977): siltstones, sandstones characterized by immaturity of texture and composition, local lenses of conglomerates, whose pebbles are at 95% made of quartz.

The sedimentary environment was one of active erosion and rapid deposition. The nature and alternation of facies imply a rhythmic, shallow water sedimentation. Moreover, sedimentary environments locally involve temporary emergence with estuarine deposition (Chauvel & Phillipot 1957). The age of the strata is still debated as neither zone-fossils nor direct radiometric dating are available. A possible Cambrian age has been suggested from regional geology (Le Corre 1977) or structural arguments (Ballard *et al.* 1986).

South of Rennes, the *Palaeozoic succession* of Central Brittany ranges from the Arenig to the Silurian. For the purpose of the present paper, only the two lowest Ordovician formations are described (Fig. 2).

The Pont-Réan Formation (PRF). The basal unit consists of conglomerates, micro-conglomerates and sandstones that lie directly upon the Brioverian (Fig. 2a). The 'Montfort conglomerates' contain pebbly clasts up to 10 cm in size derived mostly from the Brioverian formations. The principal unit of the PRF is made of mica-rich red siltstones and sandstones. Picked out by a discontinuous outcrop pattern, the PRF varies from 0 to 800 m in thickness from place to place. In the upper part of the PRF, gritty interbeds become more numerous toward the top, defining a progressive transition into the Grès Armoricain Formation. The age of the PRF in Central Brittany is not known, but Bonjour *et al.* (1988) have obtained an Arenig age of 465 ± 1 Ma (U–Pb zircon method) from interbedded volcanics within a similar formation in the Crozon Peninsula (see Fig. 1). Sedimentological observations (texture, red colour) and biogenic structures (Scolithos, Cruziana) suggest a continental environment with marine influence.

The Grès Armoricain Formation (GAF) is generally considered to be Arenig in age (Paris 1981). It shows various facies from pure quartz sandstone to pelite indicating changes in palaeoenvironment ranging from coastal and upper continental shelf to a deeper water environment below wave base (Durand 1985). The GAF is associated with a general spread of the Arenig marine transgression over the whole Armorican area.

Tectonics

The Palaeozoic and Brioverian formations of Central Brittany were deformed during the Lower Carboniferous (Hercynian orogeny). The whole area was subject to dextral shearing (Gapais & Le Corre 1980) producing upright folds whose subhorizontal axes trend N90°E in the northwestern part of the studied area to N120°E in the southeastern part, forming a large syncline structure (Martigné-Ferchaud synclinorium, Fig. 1) with a steeply dipping axial-plane cleavage, and a subhorizontal stretching lineation parallel to fold axes (Le Théoff 1977, Le Corre 1978). The Carboniferous age of the deformation is bracketed by the radiometric dating (344 ± 8 Ma and 336 ± 13 Ma, whole-rock Rb–Sr) of syntectonic leucogranite emplacement along the northern branch of the South Armorican shear zone (Peucat *et al.* 1979). The syn-kinematic character of leucogranite emplacement is demonstrated by the zonation of cleavage types around plutons, the physical processes of slate and quartzite deformation (Le Corre 1977, Gapais & Le Corre 1981) and the pattern of cleavage trajectories with triple points in the vicinity of plutons (Hanmer & Vigneresse 1980, Hanmer *et al.* 1982, Vigneresse & Brun 1983).

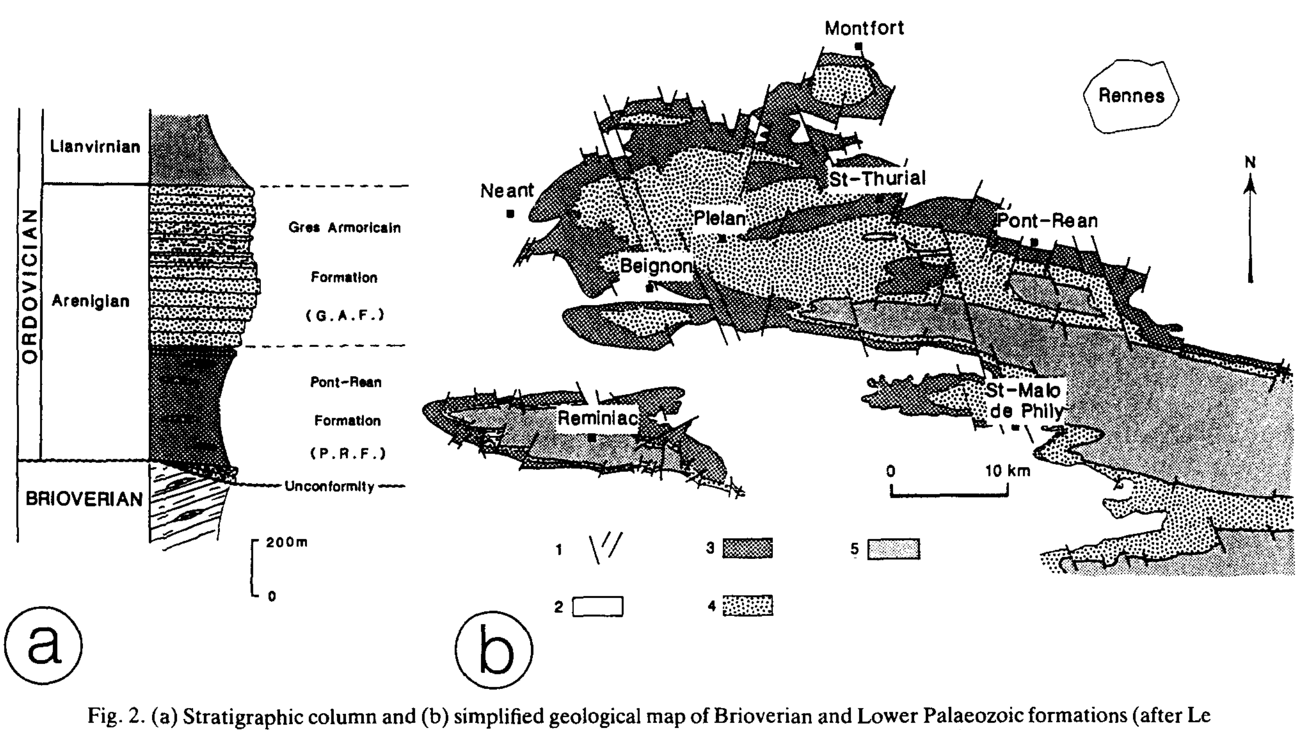

Fig. 2. (a) Stratigraphic column and (b) simplified geological map of Brioverian and Lower Palaeozoic formations (after Le Corre 1978). 1, Hercynian faults; 2, Brioverian formation; 3, Pont-Réan Formation (PRF); 4, Grès Armoricain and younger formations; 5, Upper Palaeozoic formations.

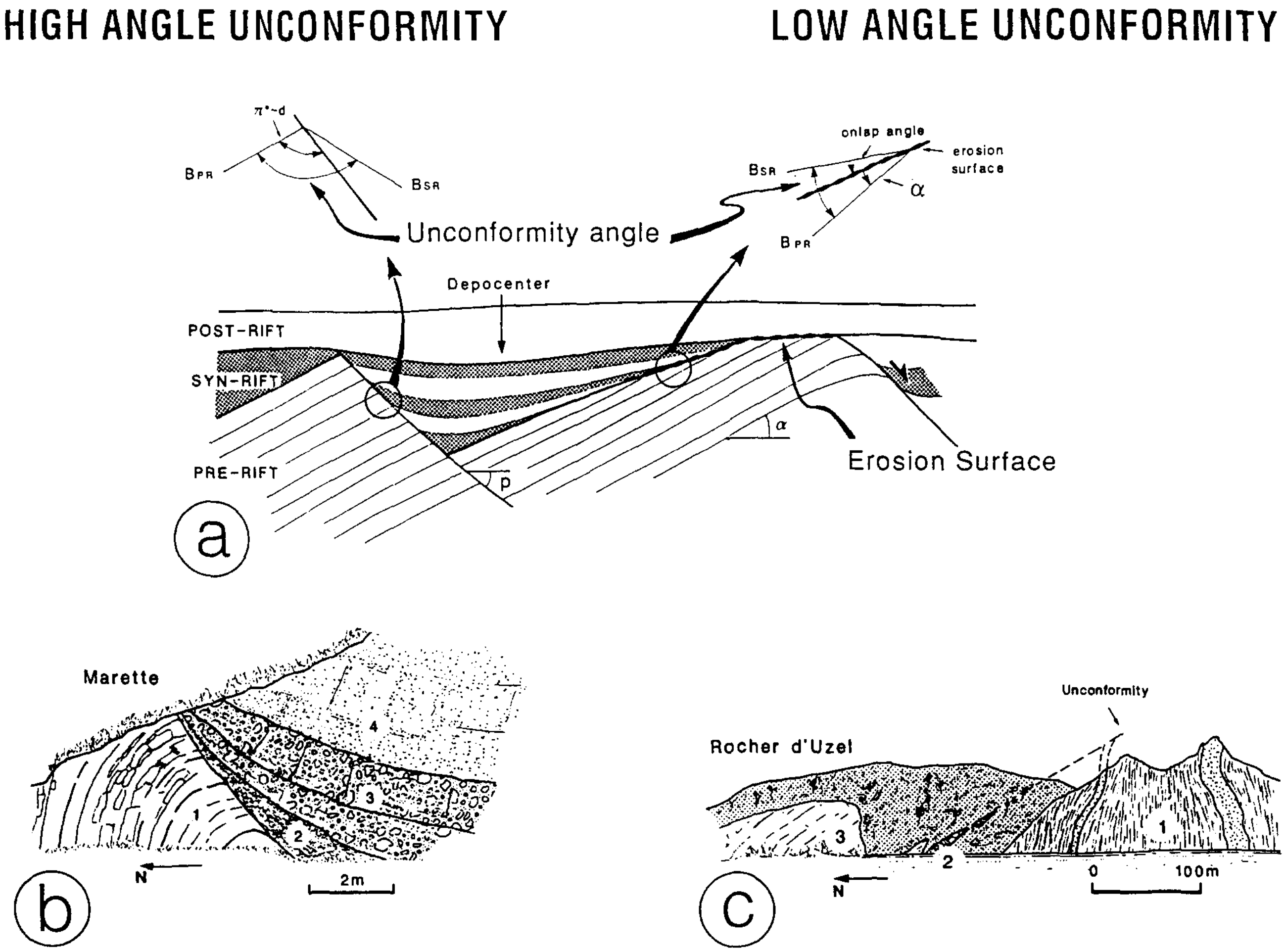

Fig. 3. (a) Geometrical model of a syn-sedimentary half-graben above tilted blocks showing high- and low-angle unconformities. B_{PR}, pre-rift bedding; B_{SR}, syn-rift bedding; α, tilt angle; p, normal fault dip. (b) High-angle unconformity at Marette (see location in Fig. 4a). 1, Brioverian; 2, basal microconglomerate; 3, conglomerate; 4, overlying Pont-Réan Formation (PRF). Note the wedging of basal conglomerates. (c) Low-angle unconformity at the Rocher d'Uzel (see location in Fig. 4a). The Pont-Réan Formation (2) is thin and only represented by conglomerates with quartz pebbles. The Grès Armoricain Formation (3) lies upon an erosional surface of the Brioverian (1).

Until the observation of field relationships between Brioverian and Palaeozoic rocks in the Crozon peninsula (Le Corre & Chauvel 1969), and extensive structural mapping in Central Brittany (Le Corre 1977, 1978), the deformation of the Brioverian was considered to be Cadomian (Late Proterozoic) in age (Cogné 1962). The argument given was the well-known unconformity between the Brioverian and Palaeozoic strata (Bolelli 1944, 1951). Le Corre (1978) interpreted the Brioverian–Palaeozoic unconformity as due to syn-sedimentary deformation (slumping-induced recumbent folds), as observed in the Baie de Douarnenez (see Fig. 1) (Darboux *et al.* 1975) prior to the deposition of the PRF. Thickness variations of this latter formation are supposed to be a consequence of residual Brioverian topography.

Ordovician tilted blocks

Reconsidering the available sedimentological and structural data, Ballard *et al.* (1986) have recently proposed a new interpretation for the Brioverian–Palaeozoic unconformity. The angle of unconformity is observed to vary with the thickness and nature of the Pont-Réan Formation (PRF). This angle is high (greater than 90°) where the PFR is thick (e.g. Marette, Fig. 3b) and small where it is thin (e.g. Rocher d'Uzel, Fig. 3c) or does not exist. This observation can be related to a tilted block pattern developed in the Brioverian terrane controlling the deposition of syn-rift sediments (PRF) (Fig. 3a). In the following sections we firstly review the geological data which support the hypothesis such as lithology, thickness variations or angle of unconformity. Secondly, we identify the location and attitude of normal faults at the map scale using computer models of folded tilted blocks.

THE PONT-REAN FORMATION AS A SYN-RIFT DEPOSIT

The characteristics that can be used as criteria of a syn-rift environment are the nature and location of conglomerates, the presence of olistoliths and the occurrence of interbedded volcanic sediments and flows as well as thickness variations and gaps (Ballard *et al.* 1986). None of these features alone is sufficient to argue for a syn-rift deposition for the PRF, nevertheless their combination and contemporaneity seem to be good indicators.

Conglomerates and olistoliths

Ordovician conglomerates occur throughout the studied area. A monogenetic microconglomerate with centimetric siltstone pebbles aligned parallel to the bedding is found only on the northern border of the area. A polygenetic conglomerate, whose pebbles include sandstone, quartz, black cherts and shale, is present throughout the synclinorium. Quartz pebbles are always well rounded and abundant in the southern area but less in the northern one. Strain analysis of the PRF conglomerates (Le Théoff 1977) shows that the long axis of computed strain ellipsoids is always eastward dipping. This may partly result from a pre-existing imbricate sedimentary fabric. It is important to note that PRF facies with quartz pebbles are always located in the vicinity of Brioverian conglomerate layers. This seems to indicate that quartz pebbles came from the erosion of poorly consolidated Brioverian conglomerates. More generally, the nature of pebbles in PRF conglomerates near the unconformity is directly dependent on the nature of underlying Brioverian sediments (Le Théoff 1977, Le Corre 1978, Bonjour 1988).

In the PRF, there are carbonate rocks similar to the carbonate facies as seen in the Brioverian (Les Rochelles quarry, see location on Fig. 4). As suggested by Le Corre (1978), these may be interpreted as sheets of an intermediate unit between the Brioverian and Palaeozoic formations. These blocks could also possibly be olistholiths.

Volcanism

Interbedded flows and volcaniclastic layers (tuffs, breccias and ash deposits) are observed everywhere in the PRF (Chauvel 1962, Bonjour *et al.* 1988, Herrouin in press). An especially well developed complex of acid volcanics is present at the eastern closure of the Reminiac syncline (Fig. 2) (Quété 1975). In this area, several successive volcanic episodes prior to the deposition of the Grès Armoricain have been revealed. The petrography and major element chemistry of these volcanic rocks suggest an extensional geodynamic environment (Quété 1975).

Thickness variations

An examination of map contours for the PRF shows that the Grès Armoricain Formation may lie directly upon the Brioverian (e.g. St Malo-de-Phily area, Fig. 2b). Thickness variations in the PRF can be rapid or progressive along contours (Fig. 4) and it is likely that this may reflect modification by Hercynian shortening. Nevertheless, we note that large variations (from 200 to 750 m) are observed along a single cylindrical fold closure even in the less deformed northern area.

Local estimates of thickness variation seem to define,

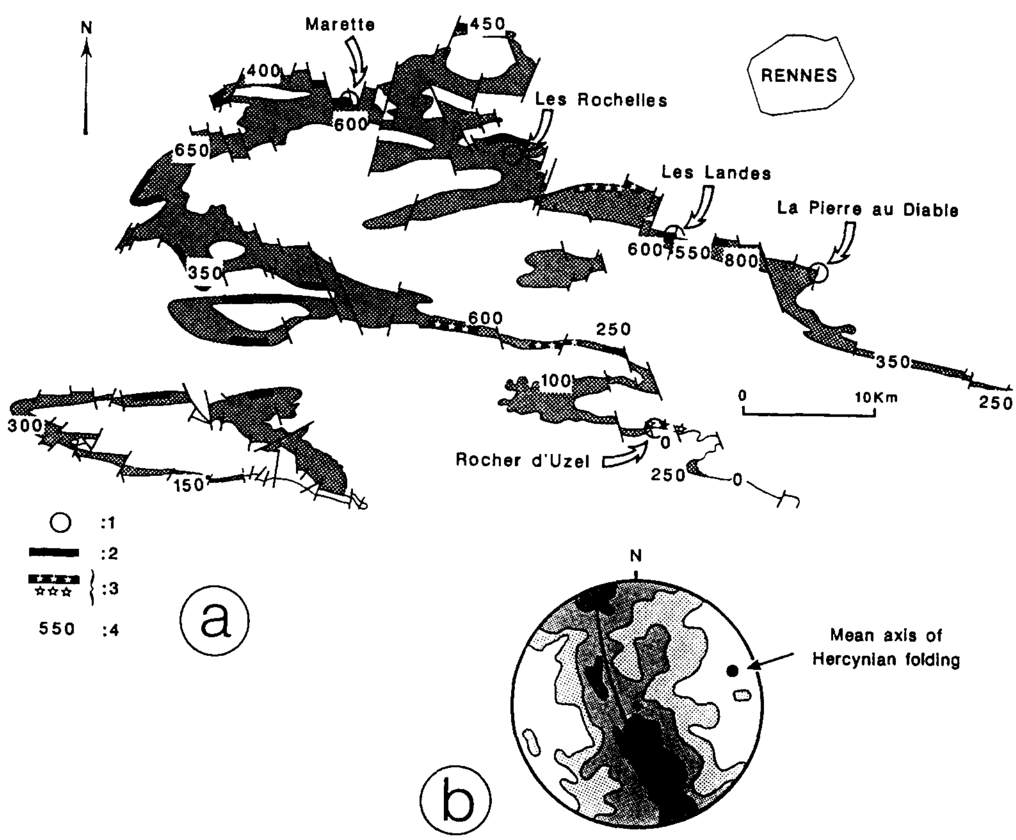

Fig. 4. (a) Characteristics of the Pont-Réan Formation in the Martigné-Ferchaud synclinorium. 1, Location of the well-known Brioverian Palaeozoic unconformities; 2, location of basal conglomerate; 3, conglomerates with quartz pebbles; 4, estimates of local thickness of the Pont-Réan Formation. (b) Poles to bedding planes within the Brioverian group. Lower-hemisphere stereographic projection. Density contours at: 0.5, 1, 3 and 4%; 250 data. (After Le Corre 1977.)

as in the northern part of the area, SW–NE oriented bands oblique to the E–W mean trend of Hercynian fold axes. We note that this trend is also observed in map contours of the Brioverian–PRF or Grès Armoricain unconformity or the PRF–Grès Armoricain contacts.

Palaeomagnetic data

A joint study of strain, magnetic susceptibility anisotropy and palaeomagnetism has been recently conducted in the PRF (Cogné 1988). The classical fold test and strain removal (Cogné & Perroud 1987) of palaeomagnetic directions gives two groups of data. Seven sites have inclinations higher than 70° and four sites lower than 60°. With reference to known Ordovician poles, the latter group appears too low and could be explained by an initial northwestward bedding dip between 10° and 20° (Cogné 1988), compatible with a half-graben style for the PRF bodies.

THE BRIOVERIAN–LOWER ORDOVICIAN UNCONFORMITY

The bedding of the pre-rift series (Brioverian) has a constant direction of dip toward the NW, compatible with a tilted block model on the regional scale. After folding, such tilting may be obscured. However, statistical analysis of Brioverian bedding attitudes gives a distribution pattern (Fig. 4b) which indicates two superposed deformations. The great circle girdle of poles to bedding and its ENE-trending axis are due to Hercynian folding. The maximum indicates an initial dip of bedding surfaces toward the NW. The NE–SW pre-existing structural trend in Brioverian sediments is also illustrated by recent geological mapping (Herrouin in press) in the Bain-de-Bretagne area (see later Fig. 7c).

Low-angle unconformities can be observed at the top of tilted blocks (Fig. 3a). The surface of unconformity can be an erosional surface. The pre-rift–syn-rift interbedding angle depends on the tilt angle and on the amount of syntectonic erosion. If no erosion occurs during tilting, the interbedding angle is low and merely reflects the onlap angle of the syn-rift deposits. If erosion occurs, the interbedding angle increases progressively and is the sum of the onlap angle and the tilt angle at the time of deposition. Due to progressive tilting, the interbedding angle increases from the depocentre of the basin toward the top of the block. The St Malo-de-Phily (Fig. 2) area is an example of an unconformity near the eroded top of a block. In this area, at the Rocher d'Uzel (Fig. 3c), the PRF is very thin and the GAF was locally deposited directly over the Brioverian sediments with a mean 30° obliquity angle. The whole structure was rotated northward in the southern limb of an Hercynian syncline which accounts for the present-day geometry.

High-angle unconformities can be observed near faults. Syn-rift sediments dipping toward the depocentre are due to fanning along the scarp and normal drag induced by fault slip. The surface of unconformity could locally be an eroded fault scarp. The pre-rift–syn-rift interbedding angle (Fig. 4a) is obtuse and depends on the fault dip (p), the tilt of pre-rift strata and the amount of displacement of syn-rift sediments along the fault. We interpret the Marette quarry unconformity (Fig. 3b) in this way. The basal layer of PRF conglomerates is seen to wedge toward the unconformity surface. The bending of bedding planes in the Brioverian, which was attributed to Hercynian folding, could be due to normal faulting. Some other quarries display examples of similar types of deposit over possibly eroded fault scarps (e.g. Les Landes, La Pierre au Diable, see locations on Fig. 4a).

Summary

A model of block tilting explains satisfactorily the sedimentology, stratigraphy and half-graben type geometry of Lower Ordovician sedimentary bodies in Central Brittany. It also provides a new explanation to account for the well known Brioverian–Ordovician unconformity. However, the quality of outcrop in Central Brittany is not good enough to map precisely the extensional fault pattern. Because geological contours mainly result from the upright folding of Ordovician tilted blocks their geometrical analysis can help to decipher the location and attitude of major normal faults. To assess the variety of geological contours which can be produced by oblique folding of tilted blocks a simple computer model has been used.

COMPUTER MODELLING

Geometrical parameters

A description of a geometrical model of folded tilted blocks is given in the Appendix. For a particular application to Central Brittany model parameters fall into three categories.

First are those concerning fold geometries which are well constrained by geological data (Le Corre 1978). To fit natural examples, folds are taken with E–W-trending axes, vertical axial planes and regular final wavelengths (W_F). Wavelengths and amplitudes of folds are computed assuming a bulk horizontal stretch of $\lambda_{SF} = 0.85$ (see Le Corre 1978, fig. 62).

Second are those concerning suspected tilted blocks and normal faults. Because this is the unknown part of the problem a range of values is tested for each parameter starting from a reasonable estimate based on the analysis of field evidence, and geological contours. According to Ballard *et al.* (1986) and geological data reviewed in the previous section, a mean obliquity of $\beta = 20°$ between tilted blocks and fold axes can be assumed. From thickness variations of the PRF, fault spacing (W_B) can range between 5 and 15 km; i.e. roughly equal to the fold wavelength ($W_B \simeq W_F$). Models have then been computed for $\beta = 10°$, 20° and 30°, and $W_B/W_F = 1.0$ and 0.6.

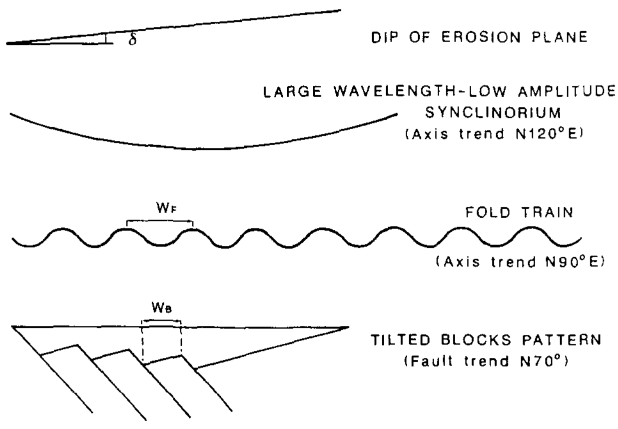

Fig. 5. Geometrical parameters used in the computation of synthetic geological maps. See text for details.

Third are the parameters concerning post-folding erosion and periclinal closures. An examination of the geological map (Fig. 2b) shows that the erosional level is structurally deeper in the northern part of the studied domain. Therefore models have been computed for an erosion surface inclined at an angle of $\delta = 5°$ and 10° with respect to the base of the post-rift sediments. To obtain periclinal closures similar to those observed on the map (Fig. 2b), a long wavelength low-amplitude syncline is superimposed with axis trending N120°E. Finally, this model is equivalent to the observed regional trend of the Martigné-Ferchaud synclinorium (Fig. 1). Figure 5 summarizes these geometrical parameters.

Results

Twelve models have been computed to cover the effects of parameters β, W_B/W_F and δ listed above. Those presented on Fig. 6 correspond to $\beta = 20°$. Figures 6(a) & (b) show the effect of a variation of 5° for δ. The degree of syncline isolation of post-rift strata in the northern part of the model is inversely proportional to the dip of the erosion surface. By comparison with the geological map (Fig. 4a), it is seen that a 5° dip (Fig. 6b) is more appropriate than a 10° dip (Fig. 6a).

Figures 6(b) & (d) show the effect of erosion depth for $\delta = 5°$. The deeper the erosion surface, the larger is the surface occupied by the syn-rift deposits and the smaller are the number of isolated synclines of post-rift strata.

Figures 6(c) & (d) show the effect of normal fault spacing W_B. The larger the W_B value, the more numerous are the periclinal closures between faults at the boundary between pre-rift and syn-rift deposits.

Identification of normal faults

It is important to note the first-order analogy between computed geometrical models (Fig. 6) and the geological map (Fig. 4a). A comparison helps to identify some of the possible pre-existing normal faults.

In the *north western border* of the synclinorium (Fig. 7) the countours of the Brioverian–PRF boundary present more lobes than the PRF–GAF boundary. This is observed in models with $\beta = 20°$ and $W_B/W_F = 0.6$ (Fig. 7b). Fault traces are undulated and have a long wavelength curvature. The detailed structural map presented in Fig. 6(c) shows a trend of bedding similar to that of the Brioverian and the PRF in the northern part, with a single smooth E-plunging syncline within the PRF passing westwards into two closures. It is especially relevant to note the contour–bedding obliquity in the northern

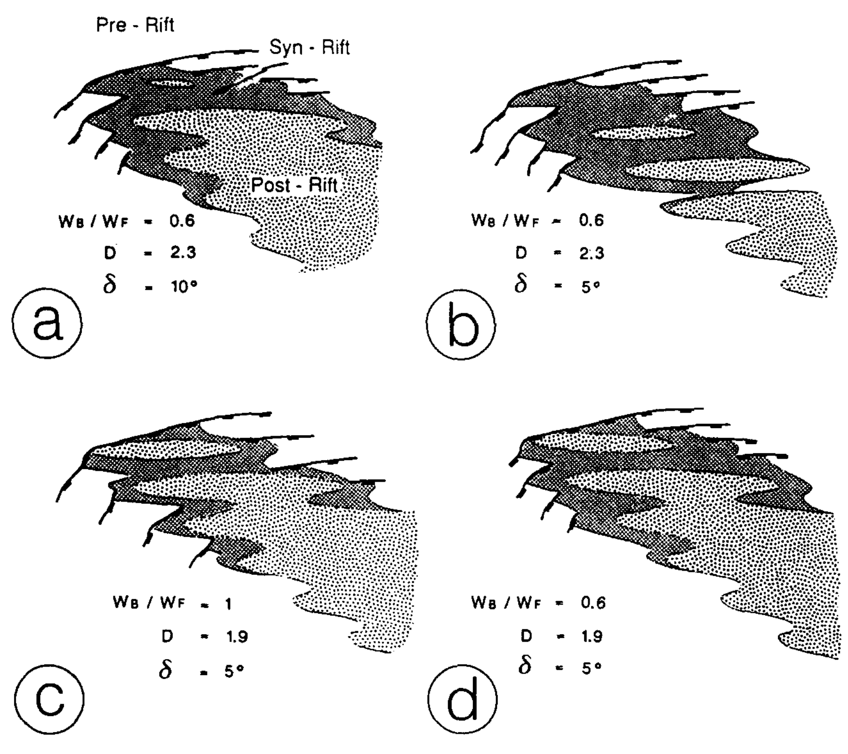

Fig. 6. Synthetic geological map of folded tilted blocks. W_B, initial normal fault spacing; W_F, fold wavelength, D, dimensionless depth of erosion surface near the first normal fault in the left part of each model; δ, dip of erosion plane. For all models $\beta = 20°$. See discussion in the text.

NORTH WESTERN BORDER

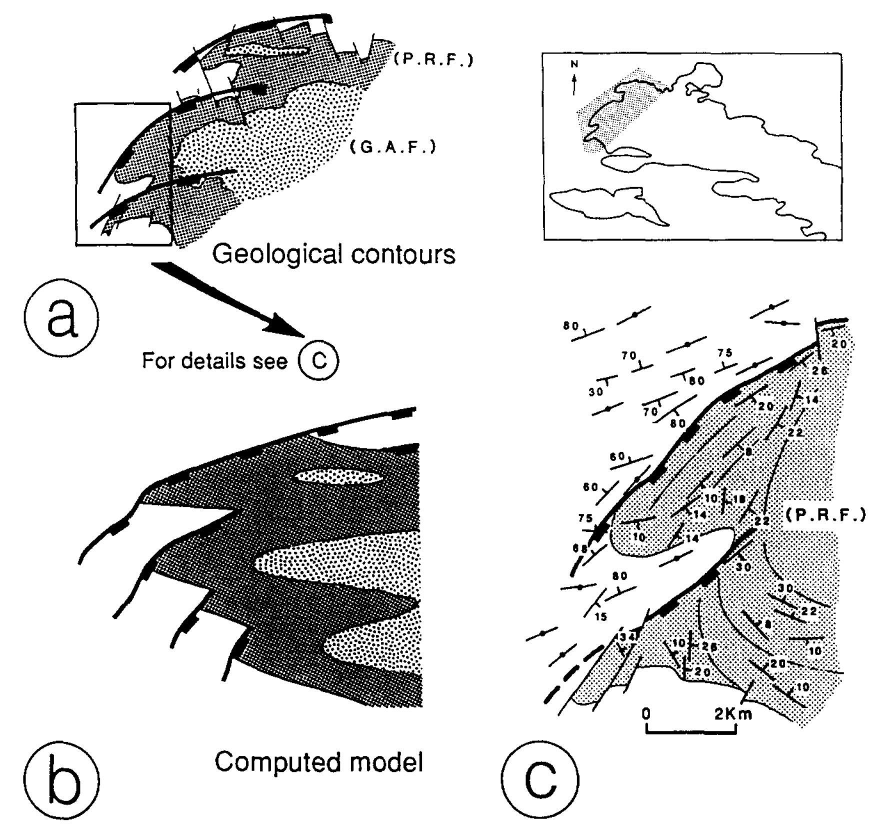

Fig. 7. Identification of major normal faults along the north western border of the Martigné-Ferchaud synclinorium. (a) Geological contours. (b) Computed model which can be compared to the observed configuration. (c) Detailed structural map with measurements of bedding planes within the pre-rift and syn-rift formations (see location on Fig. 7a).

NORTH EASTERN BORDER

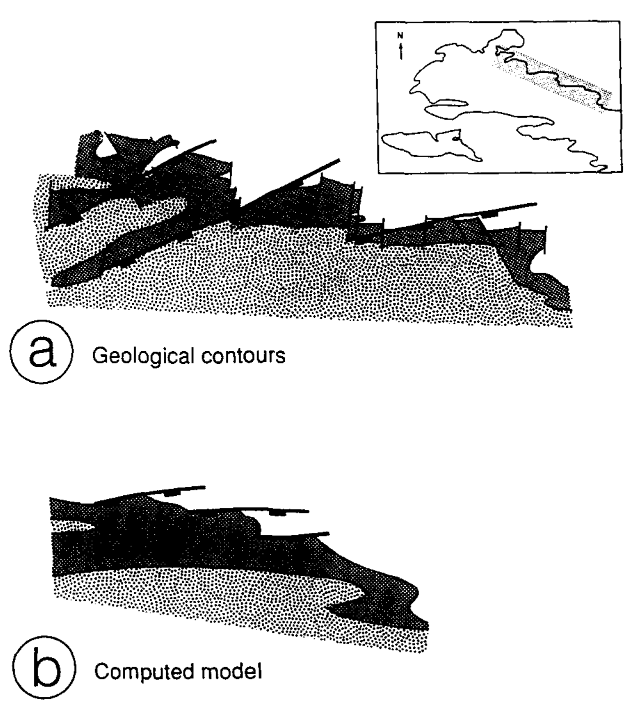

Fig. 8. Identification of major normal faults along the north eastern borders of the Martigné-Ferchaud synclinorium. Observed (a) and computed (b) geological contours.

closure which illustrates the pre-rift–syn-rift bedding relationship at the base of a half-graben (see Fig. 3a). The comparison of observed and computed contours leads us to draw at least three major curved normal faults in the north western border (Fig. 7a).

Along the *north eastern border* of the synclinorium, the Brioverian PRF boundary defines several regularly spaced asymmetric periclinal closures (Fig. 8). Only some of these closures have their equivalent at the Grès Armoricain–PRF (post-rift–syn-rift) boundary. The same feature is observed in computed models and results from the obliquity between normal faults and fold axes. The periclinal closures are observed where the Brioverian–PRF boundary corresponds to the top of a tilted block. Along these borders, normal faults are not strongly curved by folds. By analogy with the models, we locate normal fault traces along the long limb of periclinal closures.

At the *south eastern border* (Fig. 9) wedging of the PRF is clearly represented. It corresponds to thinning of the syn-rift series towards the top of a first-order tilted block. Folds in periclinal closures are very similar to those obtained in all computed models (Fig. 6). Note that a low-angle unconformity (the Rocher d'Uzel, Fig. 3c) is located in this area. The detailed geological map (Fig. 9c) shows the mean N70° trend of pre-rift (Brioverian) bedding and the wedging of the syn-rift (PRF)

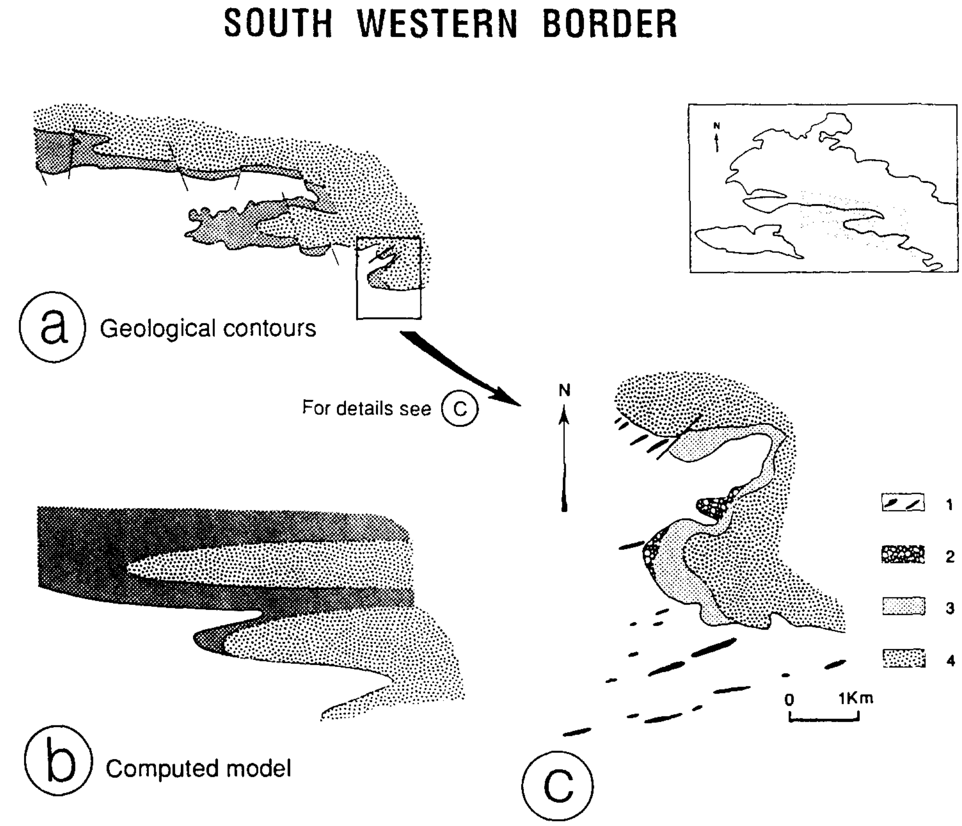

Fig. 9. Geometry of the periclinal closures without major normal faults along the south western border of the Martigné-Ferchaud synclinorium. (a) Geological contours. (b) Computed model. (c) Detailed geological map of the Bain de Bretagne anticline (after Herrouin in press) showing the N70° trend of conglomerate units (1) within the pre-rift formations (Brioverian), lithological units of the PRF syn-rift formations (2), coarse-grained sandstones (3), pelitic sandstones and the post-rift formations (4).

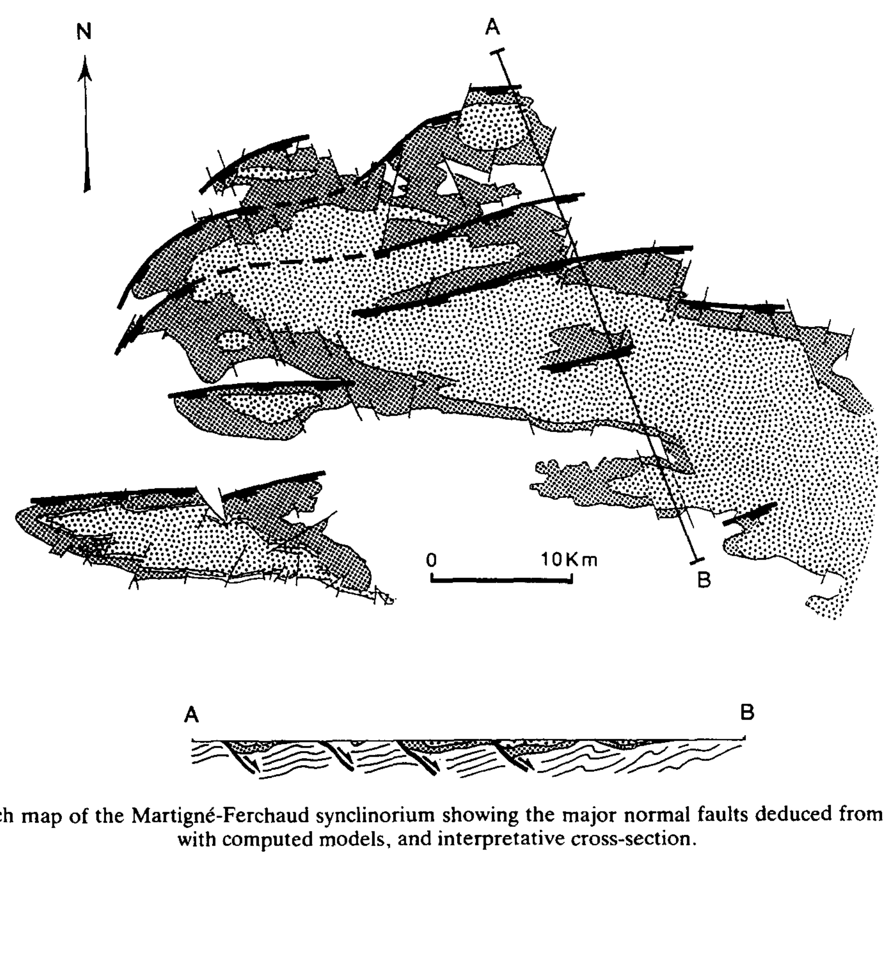

Fig. 10. Sketch map of the Martigné-Ferchaud synclinorium showing the major normal faults deduced from comparison with computed models, and interpretative cross-section.

sediments, in the vicinity of the southern periclinal closure (see location Fig. 9a).

Deduced fault traces are located on Figs. 7(a), 8(a) and 9(a) and summarized on Fig. 10. These faults could correspond to fault sets on a small scale, but outcrop conditions do not allow a better precision. In the northern part of the domain, faults have sinuous traces due to Carboniferous folding and display a mean N70°E orientation and a 7–8 km spacing. Toward the southern part of the domain, fault orientation is seen to rotate toward a N90°E orientation. This is probably due to the increase of Carboniferous strain toward the south (i.e. toward the northern branch of the South Armorican shear zone). The regional Carboniferous strain pattern is interpreted as the result of a large-scale dextral and heterogeneous simple shear (Gapais & Le Corre 1978), with a mean value of $\gamma = 1.2$ (Percevault & Cobbold 1981). So the present-day orientation of normal fault traces has most probably suffered a right-hand (clockwise) rotation. Because at least some of these faults have been re-activated, and not just passively rotated during Carboniferous deformation, it is difficult to determine their original direction. Nevertheless, the persistance of an obliquity between faults and folds indicates that the direction of extension was initially oblique to the latter fold belt trend. A mean NW–SE extension direction can be assumed.

CONCLUSIONS

In Central Brittany, where Palaeozoic sediments were folded during the Hercynian orogeny, the existence of extensional structures of Lower Ordovician age is argued. A simple model of block faulting and tilting during Lower Ordovician times explains satisfactorily the relationships between Brioverian and Palaeozoic sediments and the sedimentological characters of the Pont Réan Formation as a syn-rift deposit. However, the extensional structures are strongly obscured by superimposed Hercynian upright folding.

Computer modelling has been found suitable to compute the interference patterns which result from folding of a tilted block pattern. Synthetic maps have been produced which can be compared with geological maps and which allow the normal faults to be recognized from geological contours and then draw upon the map scale. A mean NW–SE stretching direction can be assumed for the Lower Ordovician extensional event.

Finally, the geodynamic significance of such an extensional event 460 Ma ago remains to be determined. It might correspond to the opening of an oceanic domain. But only 40 Ma appears to separate this event from the earliest evidence of ocean closure (blueschist metamorphism in southern Brittany; Peucat 1986). It could more probably correspond to the opening of a back-arc basin at the onset of subduction.

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APPENDIX

FOLDED TILTED BLOCKS. A GEOMETRICAL MODEL

A tilted block geometry is defined by the three sets of planes bounding the syn-rift strata (Fig. 4), which are (Fig. A1a).

P1: fault planes,
P2: top of faulted blocks and
P3: base of post-rift series.

In an X,Y,Z co-ordinate system, the long axis of blocks is taken parallel to Y. P3 is horizontal, i.e. parallel to the XY plane. Initial fault dip p_0 has been taken equal to 60°. Final fault dip p and block tilt α depend on the amount of stretching, λ_B according to the following equations (Brun & Choukroune 1983):

$$\lambda_B = \sin p_0/\sin p \quad \text{and} \tag{A1}$$

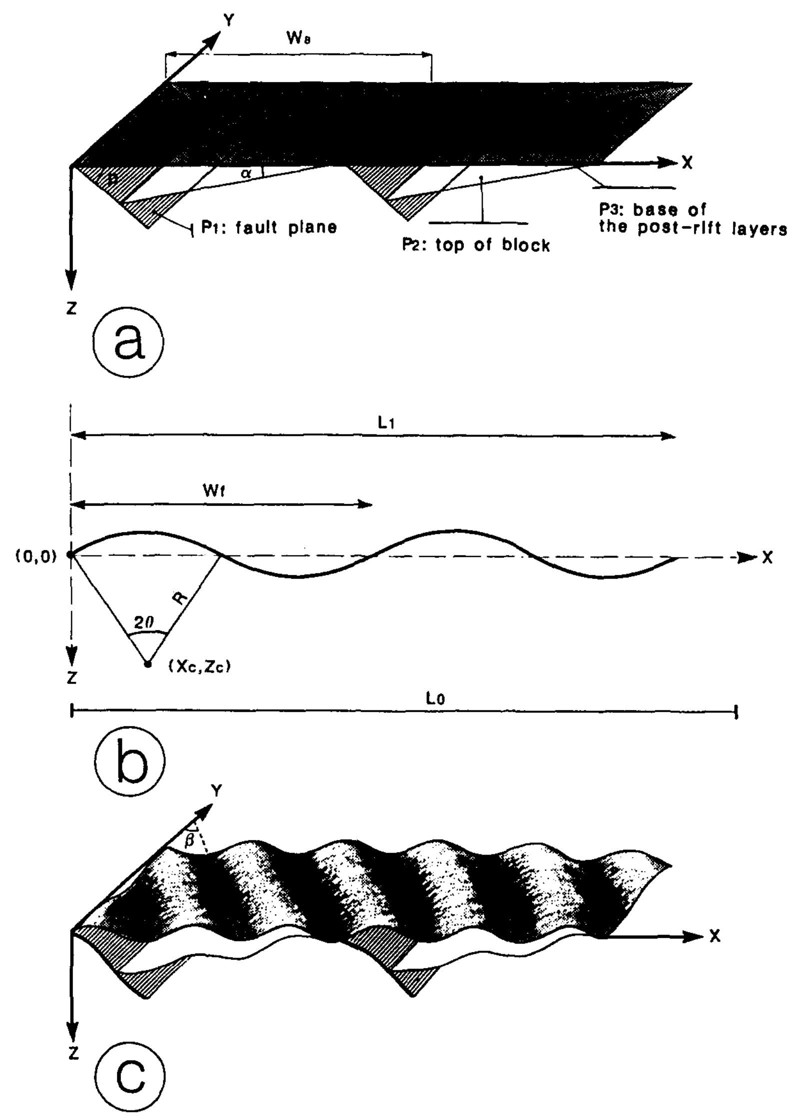

Fig. A1. Geometrical model of (a) tilted blocks, (b) folds and (c) folded tilted blocks showing the parameters used for the computations. See detailed description in the Appendix.

$$\alpha = p_o - p. \quad (A2)$$

Folds are chosen upright and cylindrical with a regular wavelength (W_F) (Fig. A1b). The shape of synclines and anticlines are assumed as arc circles. The relations between wavelength and fold shape is given by

$$W_F/2 = 2R \sin\theta, \quad (A3)$$

where R is the radius of curvature and θ the half angle of arc curvature (Fig. A1b).

The initial length, L_o, of a horizontal straight line becomes, after folding, L_1 such as L_1 can be divided into an integer number U of half wavelengths.

Therefore:

$$L_1 = 2U \sin\theta \quad \text{and} \quad (A4)$$

$$L_o = 2UR\theta. \quad (A5)$$

On the other hand, we have strain

$$e_F = (L_1 - L_o)/L_o. \quad (A6)$$

Introducing (A4) and (A5) into (A6) we obtain:

$$e_F = (\sin\theta/\theta) - 1. \quad (A7)$$

For any value of stretch λ_F (with $\lambda_F = 1 + e_F$) the value of θ is found from (A7) using a method of numerical approximation. A limiting value is attained for $e_F = -0.33$, the value at which the fold shape is a half circle and $Z_c = 0$.

The co-ordinates (X_c, Z_c) of circle centres are given by:

$$X_c = R \sin\theta + (tW_F/4), \quad (A8)$$

$$Z_c = \pm R \cos\theta, \quad (A9)$$

where t is the number of fold which varies from 0 to $(U - 1)$. Z_c is positive for anticlines and negative for synclines. The equation for the corresponding circles is:

$$(X - X_c)^2 + (Z - Z_c)^2 = R^2. \quad (A10)$$

The obliquity β of fold axes on Y (Fig. A1c) is introduced in equation (A8):

$$X_c = R \sin\theta + (tW_F/4) - Y \tan\beta. \quad (A11)$$

The position after folding of any point (X,Y,Z) belonging to the planes P1,P2,P3 which define a tilted block, is found using equations (A9)–(A11).

Models presented in the paper have been computed for a planar erosion plane either horizontal at a constant depth Z_M or inclined $Z_M = f(X)$.