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REACTIVATION OF EARLY REVERSE FAULTS ASSOCIATED WITH OBLIQUE STRIKE-SLIP FAULTING: A MECHANISM FOR CRUSTAL SHORTENING (MACIZO DE NEVERA, SIERRA DE ALBARRACÍN, SPAIN)[1]

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ABSTRACT

Rondeel, H. E., R. Weijermars & H. G. van Dorssen 1984 Reactivation of early reverse faults associated with oblique strike-slip faulting: a mechanism for crustal shortening (Macizo de Nevera, Sierra de Albaraccín, Spain) – Geol. Mijnbouw 63: 387-398.

Detailed structural and lithological analysis in the Macizo de Nevera, Spain, has revealed that the structure of the Palaeozoic basement rocks is much more complex than hitherto recognized. Alternations of pelite and psammite, separated by distinct thin intervals of orthoquartzite, are folded around shallow north-plunging fold axes. Detailed mapping of these folds disclosed that the synforms are consistently excised by reverse faulting along strike. In the western part of the Macizo de Nevera, this essentially simple structure is seriously complicated by a NE-SW trending system of en-echelon arranged dextral strike-slip faults, that partly reactivated the earlier reverse faults. Due to reworking of the latter faults, the crustal section was further shortened in an E-W direction, the motion being mainly concentrated on faults on either side of the Vallejo Hondo Anticline. Neither of the fault generations discussed continue into the Mesozoic cover of Triassic and Jurassic sediments which surrounds the Macizo. In an appendix to the paper, the Palaeozoic rock units are described.

INTRODUCTION

The greenschist and lower grade Palaeozoic rocks of the Variscan Iberian massif in the NW-part of Spain, Cantabria, form a structural arc by a gradual change of over 90 degrees in the trend of the fold axes (Fig. 1). These rocks of the Variscan Leon-W.Asturian Zone (LOTZE, 1945), disappear eastwards beneath the undisturbed deposits of the Duero basin. They reappear further to the southeast as inliers amidst the Mesozoic rocks of the Keltiberian Chain. Here, the E-W trend of the Leon-W.Asturian Zone gradually changes back to a N-S direction (Fig. 1; MATTÉ & CAPDEVILA, 1973). The Zone thus forms a large, sweeping letter S, the origin of which is a matter of debate. The activity of strike-slip faults figures prominently in the explanations suggested (HEWARD & READING, 1980; ALVARO ET AL., 1979; CANEROT, 1979; ORTI CABO, 1981; VIALLARD, 1982).

The area reported here is one of the southernmost inliers, the Macizo de Nevera. It comprises part of the Sierra de Albarracín which includes two other inliers of Palaeozoic rocks. Regional surveys of the Sierra de Albarracín have been published by RIBA (1959), TRURNIT (1967) and HINKELBEIN (1969), Their maps and reports have been consulted during undergraduate mapping courses in the area by students from Amsterdam. The structure of the basement inliers was recognized to be much more complex than previously envisaged. A

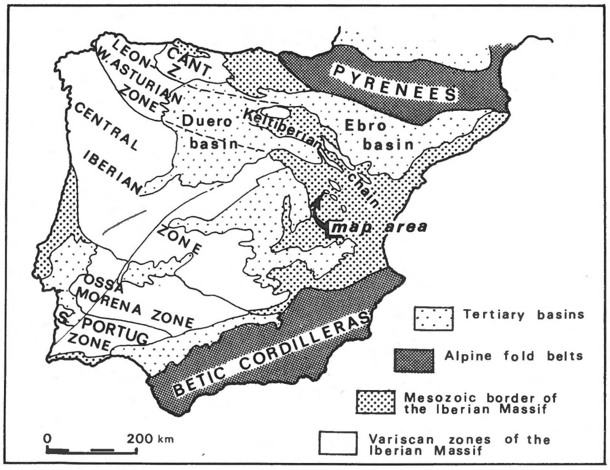

Fig. 1.
Schematical map showing the different tectonic provinces of Spain. The mapped area (Fig. 2) is located in the SE continuation of the Variscan Leon-W. Asturian Zone which reappears in the Keltiberian Chain as basement inliers amidst Mesozoic cover rocks of the Spanish Meseta.

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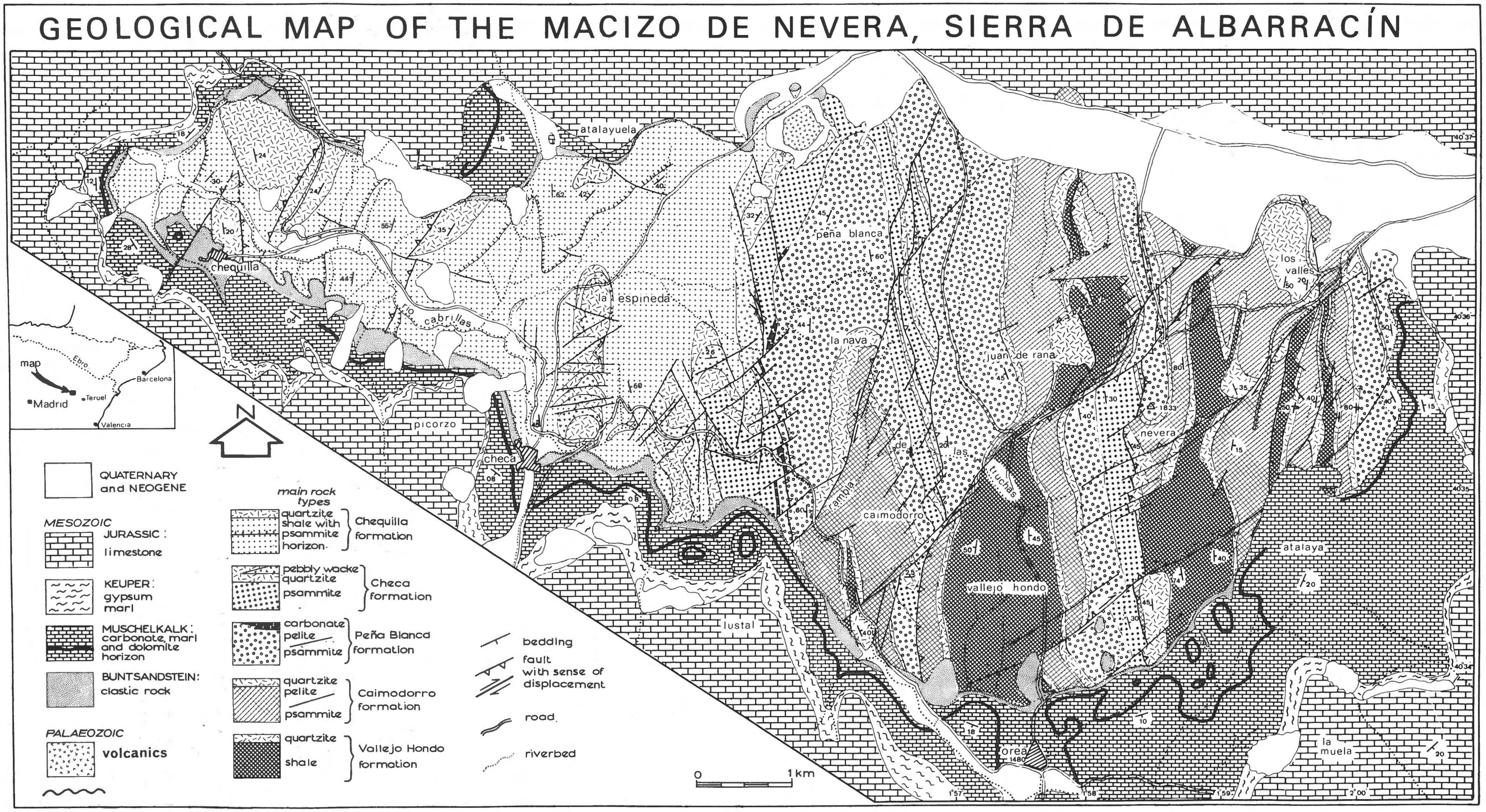


Fig. 2
Geological map of the Macizo de Nevera, reduced from a detailed reconnaissance map of 1:10.000 scale. Published with the courtesy of the Spanish Comisión Nacional de Geología, Madrid.

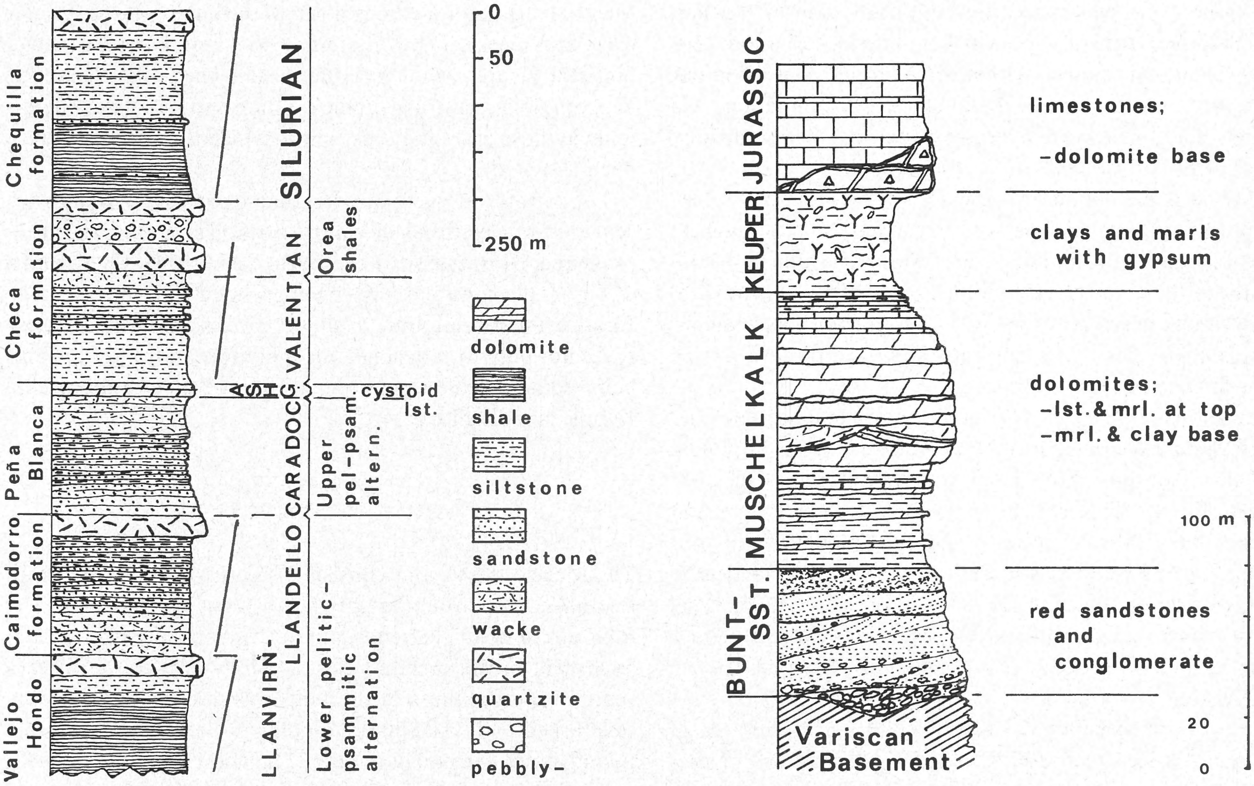


Fig. 3
Simplified stratigraphic columns of the Macizo de Nevera:
a – (left) the Palaeozoic sedimentary sequence constituting the basement of the Macizo. The correlation with related deposits elsewhere in the Keltiberian is discussed in the text of Appendix I.
b – (right) the Mesozoic cover, showing the tripartite division in the Triassic, and the Jurassic which is exposed along the margins of the Macizo.

more specified stratigraphic subdivision was required for the remapping of the Macizo de Nevera.

The bulk of the Palaeozoic rock pile consists of monotonous sequences of thick alternations of pelites and psammites, separated by relatively thin horizons of mainly (ortho-)quartzite. Due to the monotonous and repetitive character of the lithologies, it is not possible to map the area exclusively on the basis of lithological characteristics. Diagnostic lithologic criteria for the correlation of discrete stratigraphical levels are usually absent. Correlation could be certified and the lithostratigraphic framework completed by tracing the structure of individual lithologies with the help of the cleavage vergence and, to a lesser extent, fold vergence, both as defined by BELL (1981) and WEIJERMARS (1982). The result of this re-mapping of the Macizo de Nevera is shown in Figure 2. The interpretative character of the map cannot be overstressed, especially for those parts of the area which are poorly exposed and where only the resistent quartzite levels crop out. This is particularly so for the strongly forested northern slopes of the Macizo.

Below we first discuss the stratigraphy and thereafter, the structure of the Macizo de Nevera. In a final discussion our local data will be interpreted as suggesting pre-Mesozoic crustal shortening as a consequence of strike-slip movements.

STRATIGRAPHY

The Palaeozoic rocks of the Macizo de Nevera consist of five lithological units, each containing heterogeneous pelitic-psammitic series, displaying few diagnostic features. Four of the formations are composed of an overall upwards coarsening siliciclastic sequence that is topped by a relatively thin quartzite member; a fifth formation has a carbonate top level. The formations are in perfect stratigraphic continuity with each other judging from local observations. The stratigraphic column of Fig. 3 gives an overview of this more than 800 m. thick sedimentary pile that is thought to include parts of the Ordovician and Silurian.

A correlation of the lithostratigraphic units of the Macizo de Nevera with the facies-age subdivision of CARLS (1975) and HAMMANN (1976) is included in the column of Fig. 3. For further discussion see Appendix I. For a description of the formations the reader is referred to Appendix II.

The Ordovician-Silurian stratigraphy of the Iberian Chain lacks significant lateral facies changes notwithstanding local variations, so that the development of the rocks is quite uniform over large parts of the Iberian Chain (CARLS. 1975; HAMMANN. 1976). The siliciclastic sedimentary units of the Macizo de Nevera are mainly the products of a shallow shelf environment as demonstrated by the deposition of clay and silt with flaser structures and with common bioturbation. This depositional condition is further reflected in sea washed-out sands with much current bedding, cross-lamination, lag deposits, and this picture agrees with the facies described by CARLS (1975) for the eastern Keltiberian Chain.

As regards the metamorphic grade of the sediments, chlorite porphyroblasts are the only recognized mineralogical signs of metamorphism. However, conodonts from carbonate pebbles of the Checa Formation show a colour indicative of temperatures in excess of 300°C (R. J. ALDRIDGE, Nottingham, pers. comm.). The rocks can thus be judged to have passed the anchizone of metamorphism.

Not shown in Fig. 3 is a sixth, non-metamorphic and undeformed Palaeozoic formation which comprises rhyolitic volcanics exposed in the N-central-part of the Macizo de Nevera (Fig. 2). These occur as flat-lying, banded flows that postdate the Variscan folding and that are overlain on a regional scale by red clastic deposits of the Buntsandstein. Pebbles of similar volcanic rocks occur in the overlying basal Triassic along the margin of the Variscan inlier directly south of the Macizo de Nevera. It is concluded therefore that the volcanics are pre-Triassic.

A variety of volcanites in a similar setting in the area of Atienza, 120 km further NW, occurs in formations of presumed Stephano-Autunian age. Here, one of the lowermost andesites has been dated radiometrically at 287±12Ma (HERNANDO, 1980). A Stephano-Autunian age for the volcanics of the Macizo de Nevera is therefore suggested (see also NAVIDAD, 1983).

The Palaeozoic formations are unconformably overlain by Triassic and Jurassic sediments. The Triassic occurs in Germanic facies in which the tripartite subdivision Buntsandstein-Muschelkalk-Keuper is exceptionally clear (Fig. 3b). The Buntsandstein overlies a paleorelief of at least 60 metres in the Variscan basement. It normally consists of a red basal conglomeratic unit covered by red sandstone. The thickness of the unit is variable and amounts to a maximum of 60 metres. The Buntsandstein is absent in the eastern part of the Macizo de Nevera (Fig. 2), so that higher units transgressively overlie Variscan rocks. The presence of Permian in the basal conglomerates seems unlikely in view of its lithologic characteristics.

The clastic Buntsandstein is conformably covered by the marly carbonate series of the Muschelkalk facies that has most recently been the subject of studies by HINKELBEIN (1969) in the Sierra de Albarracín. It consists mainly of dolomitic beds that alternate with ductile clay-marl-gypsum levels (see Fig. 3b). HINKELBEIN has determined a non-variable thickness of about 130 m. for the entire Muschelkalk. In the area of Fig. 2 it reaches a maximum of about 110 m; reduced thicknesses may largely be due to later decollement excision. Since the Muschelkalk occurs beneath a clay- and gypsum-rich Keuper the dolomite beds of the Muschelkalk are thus sandwiched between plastic levels which later gave rise to strongly disharmonic structural behaviour during folding and thrusting of the Muschelkalk dolomite beds. Flow of the highly plastic Keuper clays and marks makes its initial thickness indeterminable. Since the Keuper lithology is identical to parts of the Muschelkalk lithology, its stratigraphic position and association determine in these cases the rock unit to which the sediments have been assigned.

The gypsiferous Keuper is overlain by an irregular, massive sequence of cavernous dolomits (carniolas) of unknown (Suprakeuper? Infraliassic?) age (ORTI CABO, 1974). On the map of Fig. 2, the carniolas have been grouped with the overlying Jurassic limestones into a single cartographic unit. These rocks not only form a single morphologic unit, but they also behaved as a single rigid superstratum relative to the ductile Keuper underneath.

VARISCAN STRUCTURE

The Macizo de Nevera consists of Variscan basement rocks that are folded around gently north-plunging fold axes. The folds are of open character and have steep axial planes (Fig. 5), except in the extreme east. Directly east of the Nevera summit, the tightness of the folds increases as they become overturned with axial planes dipping to the west. Small-scale folding is not extremely frequent. It is best developed in rock units with a well-bedded, rapidly alternating lithology of contrasting competence, as e.g. in the Caimodorro Formation.

An axial plane cleavage accompanies the folds. It normally occurs in a convergent cleavage fan that is well-developed in silty and clayey beds. The cleavage usually has an irregular and wavy morphology, a result of the heterogeneity of the lithology (flasers, burrowing) and the low temperature conditions under which it has been formed. In photomicrographs it appears as an anastomosing spaced cleavage (Fig. 6e).

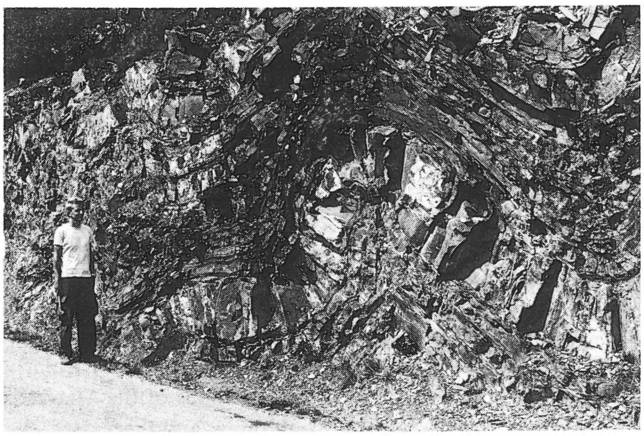


Fig. 4
The rhythmic sandstone-shale alternation of the Checa Formation as exposed along the road between Orea and Checa (km. 3), looking north. Mesoscopic parasitic buckle folds on the scale shown are typical for these rocks.

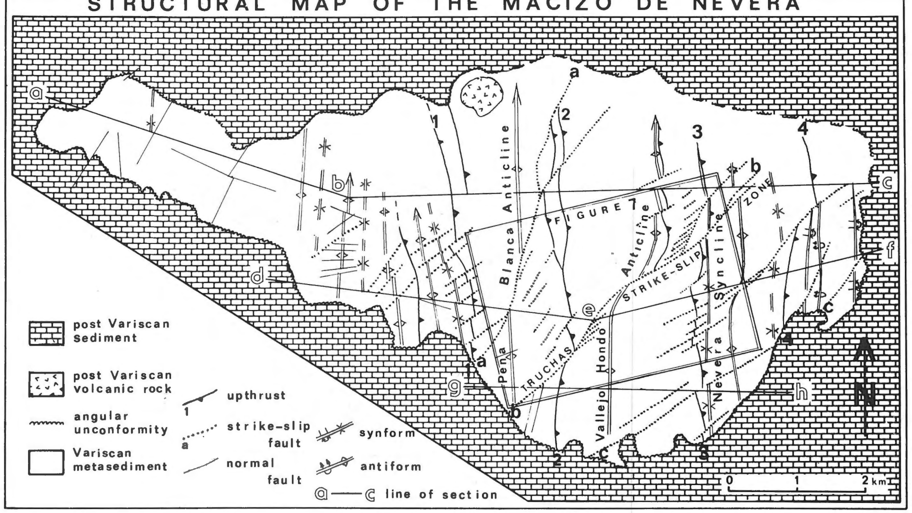
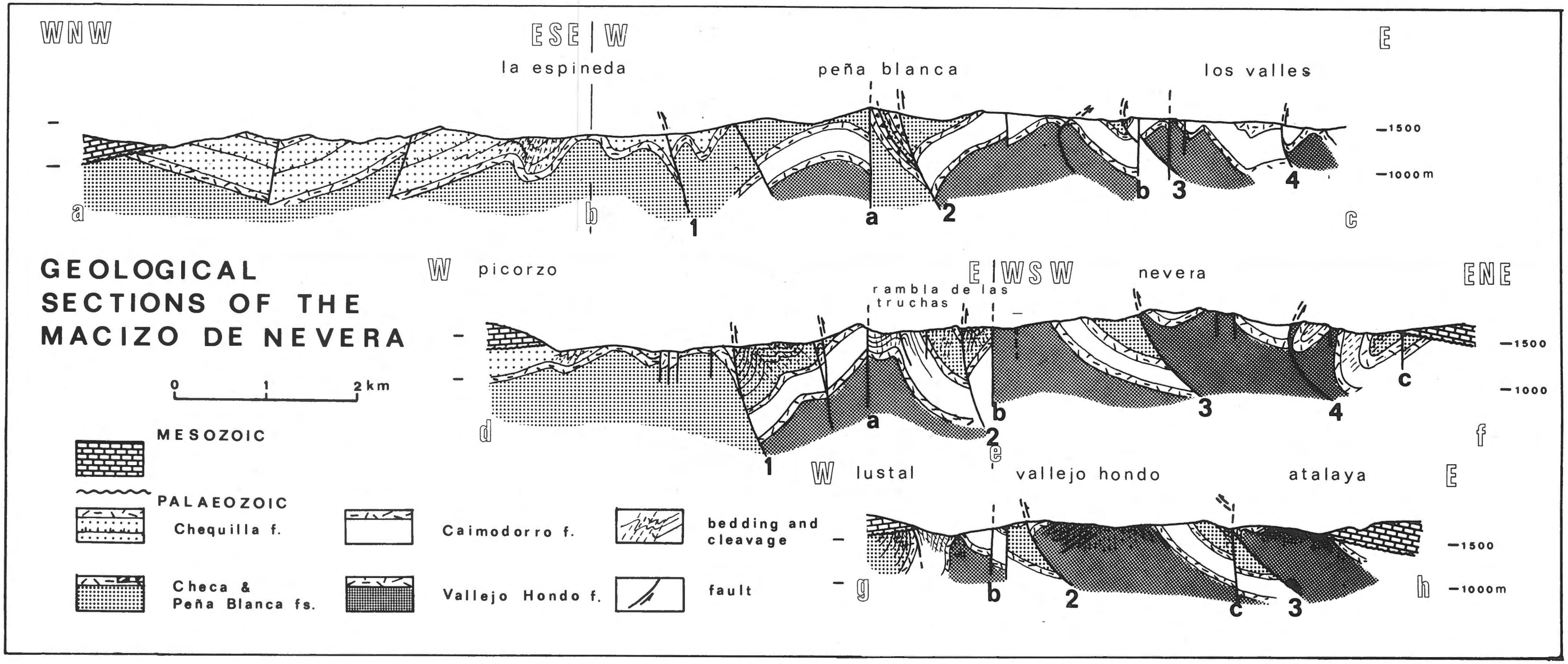


Fig. 5

a – (left) Structural map of the Macizo de Nevera, showing the principal structural elements. The location of the Figs 5b & 7 have been indicated.

b – (below) Geological sections of the Macizo de Nevera, the location of which is indicated in Fig. 5a.



The most prominent planar anisotropy in the Chequilla shales is a strongly developed bedding fissility, enhanced by the high content of organic matter. There is some evidence of diagenetic thinning and attenuation of this bedding fissility. Detrital mica flakes are deflected around the remains of micro-organisms and around detrital quartz grains (Fig. 6a, b, c). Locally, a weak crenulation cleavage overprints the bedding fissility as a primary axial plane fabric, mostly restricted to areas of small-scale folding. It is best developed in the hinge areas of such folds (Fig. 6d). In the core of microfolds, mica flakes are mechanically fragmented; there is evidence for neoformation of phyllosilicates parallel to the axial planes of these folds.

Detailed mapping of the megascopic fold structure revealed that the synforms are consistently excised by longitudinal reverse faulting (Figs. 2 & 5). The most important of these longitudinal faults are upthrusts to the west that bring anticlines on top of synclinal structures, or that cut away parts of the synclines. Rocks of the Chequilla Formation are preserved in restricted occurrences in the footwall of the thrusts. The longitudinal fault (No. 4 in Fig. 5) in the extreme east of the Macizo separates an eastern subarea with west-dipping axial planes from a western subarea with upright folds. An explanation for the western inclination of both axial plane fabric and fold structure is sought in a backward rotation of the eastern subarea as a consequence of westward thrusting along the listric thrust-plane 4 (Fig. 5). Westward thrusting is further held responsible for the deformation of a number of anticlines, resulting in overturning of beds in a westerly direction, and in a relatively flat-lying east-dipping cleavage as observable in the core of the Peña Blanca Anticline along the Río Cabrillas (Fig. 5b, section g-h).

The longitudinal faults are locally marked in the field by up to 3 m. thick, white quartz veins that stand in relief within the surrounding terrain. Although these faults are held responsible for the rotation and deformation of the axial plane fabric in various locations, they are thought to relate in age to the cleavage formation and thence with the folding. Not only do the faults fit in with the geometry of the fold structure, but the quartz veins at the locus of the faults seem to indicate

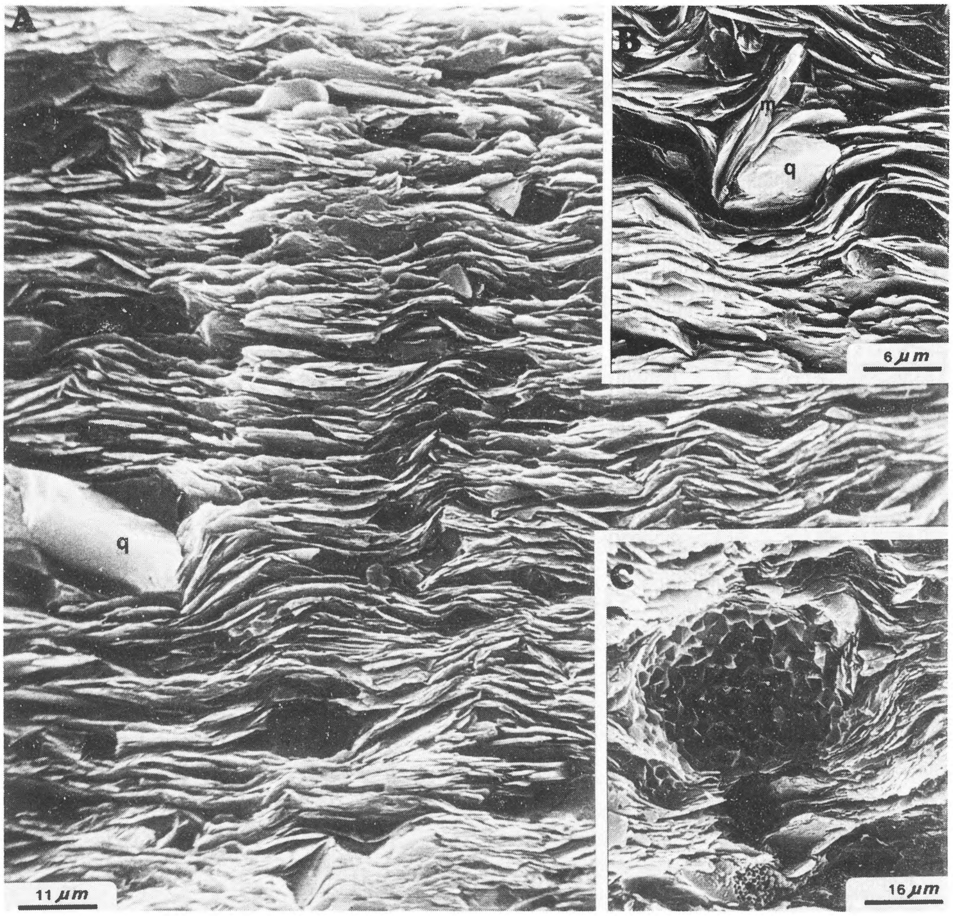


Fig. 6
Photo-micrographs of the foliations in the Palaeozoic rocks q = quartz, m = colourless mica, cl = chlorite:
A, B, C – Wavy bedding fissility in Chequilla black shales. The fissility wraps around framboidal quartz grains (B) and detrital quartz fragments (C), in the immediate vicinity of which detrital micaflakes have transverse orientation.

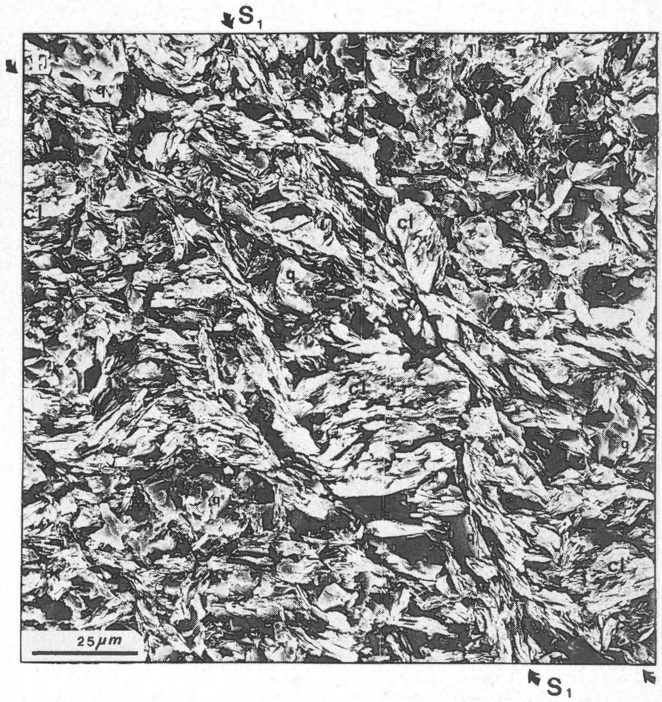


Fig. 6 continued
D – Microcrenulation of bedding fissility in a mesoscopic fold core within Chequilla black shales
E – Anastomosing spaced cleavage lamellae (S1) of fine-grained colourless mica in a coarser grained sediment of chlorite, mica and quartz grains.

conditions of large-scale component migration, like during the development of axial plane fabrics (WILLIAMS, 1972; STEPHENS ET AL., 1979; GLEN, 1982). Diffusion processes are demonstrable where the longitudinal faults cut the uppermost quartzite bed of the Checa Formation at very low angle. Here, the quartzite exhibits a subtle protomylonitic banding with fluid inclusion trails in bedding-parallel orientation across the grain boundaries. Zones of fluid inclusions are the cause of the grey colour of the hand specimen that contrasts with the normally lighter colours of the quartzites.

The fold structure is dissected by a system of NE-SW running en-echelon strike-slip faults (Figs. 2 & 5). They displace both individual folds and longitudinal faults in a consistent dextral manner and do not continue into the unconformably overlying Buntsandstein. The cleavage and bedding near these faults are deflected towards the slip direction and the strike-slip faults become true shear zones in the incompetent rock units. The Truchas strike-slip zone with its 6 km length is the most important of these structures (Figs. 2 & 5). In its central part, a rough secondary cleavage is superposed on the primary axial plane fabric. This secondary cleavage morphologically resembles the extensional crenulation cleavage that has been previously described from major shear zones in the Vanoise Massif of the French Alps (PLATT & VISSERS, 1980). The map pattern (Figs. 2 & 5a) reveals a variable displacement of markers on a single strike-slip fault.

Careful examination of the map (Fig. 2) further shows that on the longitudinal thrusts labelled (2) and (3) in Fig. 5a a different dip separation occurs on either side of the Truchas strike-slip zone. The abrupt change in amount of dip separation on these thrusts at the locus of the Truchas Fault, sufficiently demonstrates that the dip slip varies accordingly. This relationship led us to the conclusion that the variable separation as observed along the Truchas Fault is due to dissipation of its horizontal component of motion in upthrusting along already existing longitudinal thrusts. It is not clear whether this varying deformation on opposite sides is related to or a direct consequence of the undulating shape of the Truchas movement zone.

The block diagrams in Figs 7 & 8 clarify the final structure of the faulted area. They show the differing displacements on two of the longitudinal thrusts as the consequence of offset and remobilisation during strike-slip faulting. The figures also show the consequence of the non-planar character of the Truchas fault surface, leading to a shattering of the leading front limb of the Vallejo Hondo Anticline by faulting while the trailing limb is strongly elongated in a more homogeneous, ductile manner. The distortion of the anticline during strike-slip movement might also have led to the thrusting in the anticlinal core of the NW fault-block. The net result of the strike-slip motion concentrated in the Truchas fault zone and of the terminal deflection of the motion into the earlier reversed faults is an E-W crustal shortening of approximately 1 km magnitude as can be inferred from the map pattern in Fig. 2.

DISCUSSIONS AND CONCLUSIONS

The Macizo de Nevera is an inlier of Variscan basement located about 200 km east of Madrid, Spain, comprising rocks of Lower Palaeozoic age of low metamorphic grade. The rocks have been correlated with similar deposits described from elsewhere in the eastern Keltiberian. The result is shown in Fig. 3a. The entire Lower Palaeozoic sequence reflects deposition under shallow conditions on a gradually subsiding shelf. The Lower Palaeozoic alternation has been folded around uniformly north-plunging axes, the synforms being excised by longitudinal reverse faults. Both the axial planes and the reverse faults are displaced by a consistent system of en-echelon arranged dextral strike-slip faults (Fig. 2).

The Truchas strike-slip zone, the most important of these strike-slip zones, is 6 km long and over most of its length confined to a relatively narrow, often poorly exposed zone about 10 to 20 m wide. The horizontal displacement is greatest in the central portion of the fault and diminishes along the fault by upthrusting along the earlier reverse faults to either side (Fig. 7). Within the horizontal displacement zone, kink bands with subvertical axes and non-disrupted kink planes, slickensides, drag folds with a rough secondary cleavage and shear band foliation have all been observed.

The crustal shortening of about 1 km associated with the strike-slip motion on the Truchas Fault zone and the reworking of the earlier thrusts has been schematically reconstructed in Fig. 8. The mechanism discussed might provide a model for crustal shortening by wrenching in other regions.

Many questions remain to be solved, mainly concerning the cause of the fault motion. What, for example, is the implication of the consistent westward upthrusting along the early reverse faults? Is it coupled to the large scale fold structure of the Macizo, consisting of a western synclinorium and an eastern anticlinorium as a suprastructure above the rigid 500 m thick Armorican quartzite? In this view, decoupling from this rigid level could have caused thrust faults to displace parts of the anticlinorium towards the plastic Chequilla Formation in the core of the synclinorium.

Even more intriguing is the causality responsible for the crustal shortening in the narrow zone transected by the Truchas Fault. Does the fault continue into underlying levels, or is it restricted to the domain above the Armorican quart-

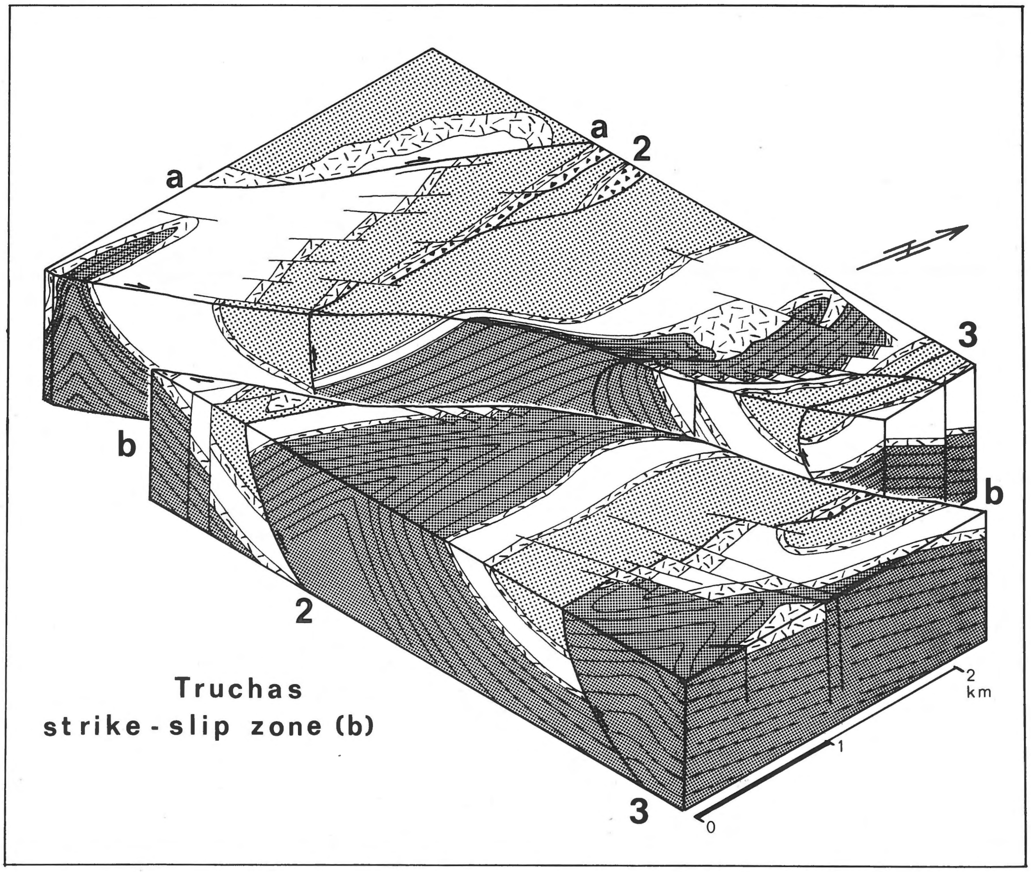


Fig. 7
Block diagram illustrating the structure of the area transected by the Truchas Fault. The location of the block diagram is indicated in Fig. 5a.

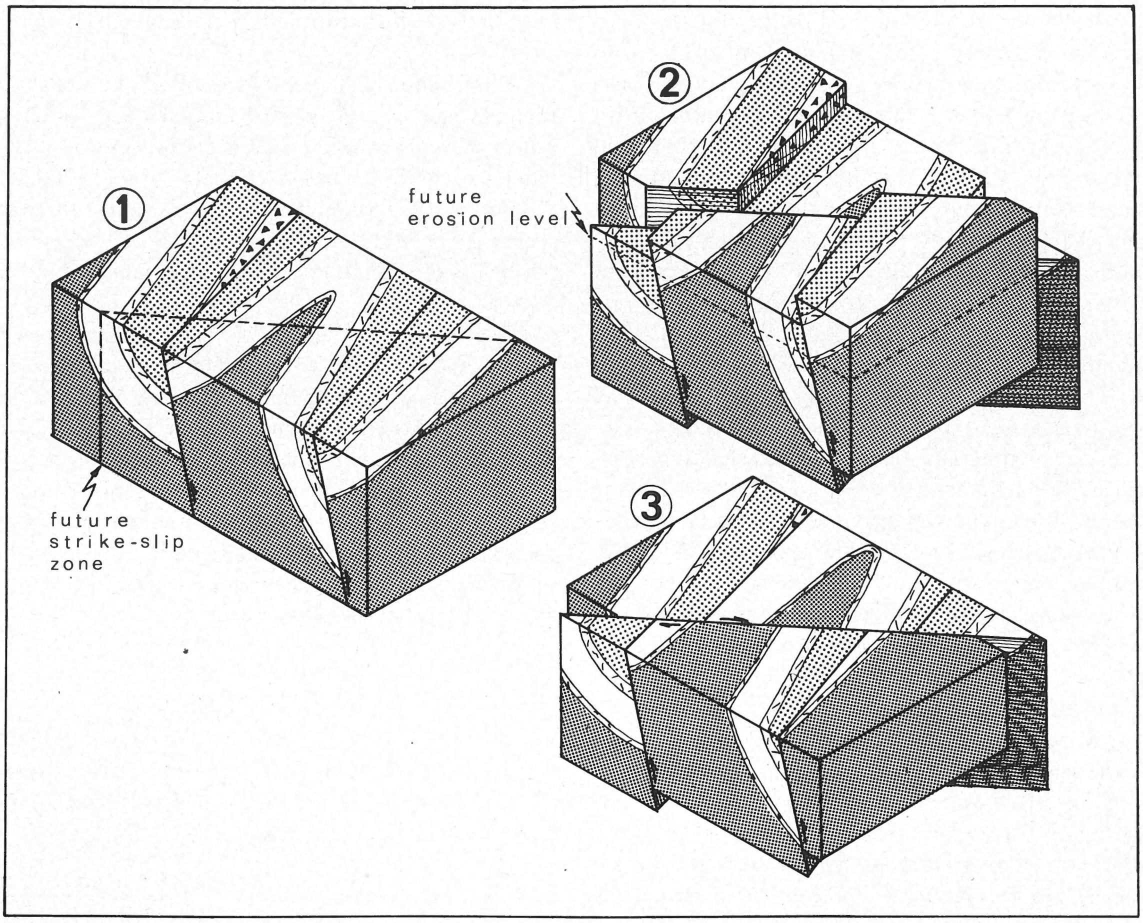


Fig. 8
Block diagrams showing reconstructions of the composite fault motion associated with the strike-slip movement along the Truchas Fault.
1 – situation prior to strike-slip movement
2 – situation after strike-slip movement, showing the increased displacement on the upthrusts on the farther side of the Truchas Fault, and a diminished total displacement on this side of the Fault.
3 – situation as in 2; eroded top surface

zite? Is the wrench fault just a means to accomplish a horizontal step in a major secondary reverse fault motion? Or is the secondary faulting in turn an attribute of the Truchas Fault to accomplish termination of its strike-slip motion? If indeed strike-slip movement figures dominantly in the explanations for the large sweeping S-shape of the Leon-W.Asturian zone, could the Truchas Fault be part of that system? Causality and effect need still sorting out. We expect to find some answers to these questions from surveys of nearby inliers.

ACKNOWLEDGEMENTS

It is with pleasure that we here record the permission for publication obtained from the Spanish Comisión Nacional de Geología. This paper is the first of a number which present detailed maps of the basement inliers of the Sierra de Albarracín. The principal aim of the work is to obtain a better insight in the complex fault history of Variscan basement rocks by mapping a composite surface which has an area five times larger than that covered in this paper.

R. J. Aldridge (Nottingham) is thanked for the determination of some conodont species and W. Hammann (Würzberg) for the determination of the trilobites. The contribution of J. Akkerman in the micrographs of the cleavage morphology is highly appreciated. An earlier version of this paper was considerably improved after comments of C. J. Talbot.

APPENDIX I

The regional facies-age correlation presented in Fig. 3a is partly based on the trilobite fauna of the Caimodorro Formation, which indicates an Upper Llanvirn to Lower Llandeilo age (see formation description in Appendix II).

HAMMANN (1976) has distinguished a Lower pelitic-psammitic alternation that is considered of similar Llanvirn-Llandeilo age. In the eastern Keltiberian chain, this Lower alternation is separated form an Upper pelitic-psammitic alternation of Caradocian age by a distinct 10 m. thick unit of oolite and bryozoan marl (CARLS, 1975). This unit has not been found in the Macizo de Nevera, and correlation therefore rests on the next higher characteristic level: the Ashgillian cystoid limestone that overlies the Upper alternation. The clastic part of the Peña Blanca Formation can be correlated with this Upper alternation in view of the lithological equivalence of its top carbonate member with these Ashgillian limestones, the age of which is well established on the basis of macro- and microfauna (CARLS, 1975; HAFENRICHTER, 1979; 1980).

In the eastern Keltiberian chain, the Orea shales of Llandoverian age rest on top of the Cystoid limestone without any transition; the top of the limestone being rough at some locations in the manner of a karst topography (ROBARDET, 1981). The Orea shales and the Cystoid limestone are considered to be separated by a fossil-documented erosional unconformity comprising the Hirnantian (HAFENRICHTER, 1979; 1980) and the complete lack of Cystoid limestone in some areas is ascribed to erosion prior to deposition of the Orea shales (ROBARDET, 1981). In the Macizo de Nevera this unconformable character has not been observed. However, on the basis of the strongly discontinuous character of the Peña Blanca carbonate toplevel, the regional unconformity at the base of the Silurian deposits is here thought to occur at the top of the Peña Blanca Formation. This unconformity most likely relates to a glaciation of continental extent near the Ordovician-Silurian boundary as indicated by data from various disciplines (HAMBREY & HARLAND, 1981). In the nearby Moroccan Anti-Atlas, tillites are seen to cover a precisely dated intra-Hirnantian unconformity (DESTOMBES, 1981) and in the area of Checa, the pebbly wackes of the Checa Formation have been convincingly argued to be of glaciomarine origin (FORTUIN, 1984). The coarse clastics in the pebbly wackes should represent dropstones that were still being deposited in shallow water judging from the invervening sandstones.

According to TRURNIT (1967), grapholites in the quartzites of the higher part of the Checa Formation indicate a Valentian age, thus also dating the interfingering Orea shales. The early Silurian age of these Orea shales is generally accepted, notwithstanding few diagnostic criteria (CARLS, 1975; HAMMANN, 1976; HAFENRICHTER, 1980). TRURNIT (1967), however, has suggested an Ashgillian age for the basal part of the Checa Formation. The present assumption of an unconformity at the top of the Peña Blanca Formation that marks the Hirnantian, dates the entire Checa Formation most likely as Valentian.

Judged from the conclusions drawn by CARLS (1975) in the Eastern Keltiberian chain, and on the basis of observations in the Macizo de Nevera, the marly and dolomitic rocks in the top of the Checa Formation are the only sediments that demonstrate a complete interruption in the supply of terrigenous detritus. They have been deposited under shallow water conditions, as demonstrated by their faunal contents (CARLS, 1975).

The turbiditic sequence of the Checa Formation and the rhythmic deposits in the Peña Blanca Formation seem to reflect increasing water depth. However, in view of the shallow origin of the other sediments, it is likely that shallow conditions also pertained during deposition of these rhythmic sequences. The sandstones could then be interpreted as the result of stormgenerated turbidity currents, as indicated by FORTUIN (1984).

The youngest pre-Variscan Chequilla Formation is of Silurian age, although the top part might reach into the Devonian. The shales and siltstones that form its lower part are correlatable with the Bádenas Formation from the eastern Keltibirian chain in which the late Silurian has been demonstrated (CARLS, 1975; TRURNIT, 1967). These rocks most likely represent a continuous, fine clastic basin deposit under restricted euxinic conditions, judging from the dark colour of the sediment, its organic content, the richness in graptolits, the alum content and the abundant pyrite concretions.

APPENDIX II

The area of the Macizo de Nevera lends itself extremely well for instructive mapping courses, and the descriptions below could be used as a field guide.

Vallejo Hondo Formation
The base of this lowermost rock unit is unknown from the Macizo de Nevera where it occupies the core of several large anticlines. Best sections can be found in the southern parts of these structures (Fig. 2).

The formation comprises a uniform pelitic rock sequence – largely dark shales – that upwards becomes increasingly coarse-clastic, grading into a siltstone member, followed by a quartzite member. In the basal part, slaty cleavage is always well developed and bedding is difficult to discern. Bedding can only be determined by the occurrence of grey silt laminae amidst the shales and by mesoscopic cleavage refraction. The poorly exposed siltstone member in the top of the formation contains some sandstone intercalations and the thickly bedded quartzite at the top has some siltstone intercalations.

The formation reaches a thickness of at least 120 m., the siltstone and the quartzite members in the top accounting for 20 m. each.

Caimodorro Formation
Good sections are found on the east limb of the Vallejo Hondo Anticline, and in the core of the Peña Blanca Anticline, particularly on the little disturbed east limb of this structure as exposed along the Rambla de las Truchas (Fig. 2).

The lower part of the formation comprises mica-rich and strongly bioturbated wackes with irregular, dark shaley streaks in the bedding planes. It grades into an extremely heterogeneous wacke-siltstones-shale alternation that locally contains large numbers of tectonically deformed trilobite remains, among which *Neseuretus sp.*, *Colpocoryphe sp.* and *Placoparia sp.* figure prominently. The fauna, elaborated by HAMMANN (1976), does not permit a more precise dating than Upper Llanvirn to Lower Llandeilo. The amount of coarse clastics gradually increases upwards and the top part of this heterogenous alternation is formed by distinct sandstone beds, in some places finely laminated with interbedded siltstones. The sediments are rich in white mica flakes, specifically on the planes of lamination in cross-bedded units.

The above heterogenous clastic assemblage of about 100 to 150 metres is topped by a 20 m thick quartzite level.

Lateral variations in thickness and lithological development of the formation are noticeable. The main problem is to decide whether these variations are of a sedimentary origin or are tectonic. In a number of cases it can be established that longitudinal faults cut out parts of the formation.

Peña Blanca Formation

The formation can best be observed on the western limb of the Peña Blanca Anticline as exposed on either side of the Rio Cabrillas between Orea and Checa (Fig. 2), even though longitudinal faults and minor folds consistenly complicate detailed stratigraphic observation.

The basal sandstone part of more than 30 m thick is in perfect continuity with the underlying quartzites of the Caimodorro Formation. It is highly characteristic in the field as weathering along joints and bedding produces dm-sized cubicles covering large slopes.

The sandy base is followed by a sequence of irregularly alternating wackes, pelites and psammites, grading upwards into a well-bedded rhythmic alternation that is locally disturbed by bioturbation. The amount of shale increases upwards.

The highest part of the formation is a bioturbated wacke of about 40 m thickness that continues into a carbonate member of about 12 m at the top. Greenish marly shales with bryozoan and crinoid remains from the lower part of the member are covered by coarse-crystalline dolomite beds of discontinuous character. The dolomite succession thickens upward.

Checa Formation

The best exposures are on both valley sides of the Rio Cabrillas, about 2 km east of Checa, where mesoscopic minor folds make thickness estimates inaccurate.

The formation comprises an upwards coarsening sequence that commences with a characteristic homogeneous black siltstone of about 50 m thickness. It is followed by about 50 m of a rhythmic sandstone-shale alternation with turbiditic features (Fig. 4). This turbiditic sequence grades into the overlying wacke member of about 15 m.

The top part of the formation is a white quartzite (quartz sandstone) with an interfingering pebbly-wacke member, together attaining a thickness of roughly 80 m. The pebbly-wacke member separates a lower quartzite level from an upper one. The lower quartzite is frequently laminated and cross-bedded. It has a thickness of 30 to 40 m and is normally covered by uniformly developed, essentially unstratified pebbly-wackes of variable thickness, locally with quartzite lenses that reach several metres in thickness. The pebbly elements in the rock are mainly limestone and dolomite clasts that are supposedly derived from equivalents of the Peña Blanca Formation. The pebbles range from cm-scale to several decimetres. They have yielded a conodont fauna of typical Upper Ordovician aspect (R. J. ALDRIDGE, Nottingham, pers. comm.). The wackes are topped by an upper quartzite level, varying in thickness between 20 cm and 15 m. The stratigraphic thinning makes this upper quartzite level locally not mappable on the scale of our map (Fig. 2), as for example on the west limb of the southern Espineda Anticline.

Chequilla Formation

This highest Variscan clastic sequence is coarsening upward. It comprises three units, the lower of which is usually extremely well-exposed. It consists of black shales with alum and abundant graptolites, mostly *Monograptus, Rastrites* and *Spirograptus.* The rock shows an extremely well-developed bedding fissility, brought about by a strongly preferred orientation of its principal mineral constituents, colourless mica of 20-40 µm. The shales locally contain disc-shaped carbonate-rich, mineralized concretions (Cabezos de Moros) of variable size. Thin sandstone beds are intercalated in this basal part of the formation, the thickness of which is estimated at about 80 m.

The middle part of the formation is about 100 m thick and comprises a uniformly developed siltstone series. It rests on several metres of rapidly alternating fine sandstones and shales that frequently stand in relief and that has been used as a marker level (Fig. 2). The siltstone member increasingly receives sandstone intercalations in its higher part. The sandstone beds have an increasing thickness and become more abundant upwards in the sequence. They are laterally discontinuous and they feature slump structures and loading phenomena.

The higher part of the formation consists of 20 m of white quartzite with some shale and siltstone intercalations in the lower levels. The rocks show current-bedding and ripple-marks, with trace fossils on the bedding planes.

REFERENCES

Alvaro, M., R. Capote & R. Vegas 1979 Un modelo de evolucíon geotectónica para la Cadena Celtibérica – Acta Geol. Hisp. 14: 172-177.

Bell, A. M. 1981 Vergence: An evaluation – J. Struct. Geol. 3: 197-202.

Canerot, J. 1979 Les Ibérides: essai de synthèse structurale – Acta Geol. Hisp. 14: 167-171.

Carls, P. 1975 The Ordovician of the Eastern Iberian Chains near

Fombuena and Luesma (Prov. Zaragoza, Spain) – Neues Jahrb. Geol. Paläont. Abh. 150: 127-146.
Destombes, J. 1981 Hirnantian (Upper Ordovician) tillites on the north flank of the Tinbdouf Basin, Anti-Atlas, Morocco. In: Hambrey, M. J. & W. B. Harland (eds): Earth's pre-Pleistocene glacial record – Cambridge University Press: 84-96.
Fortuin, A. R. 1984 Late Ordovician glaciomarine deposits (Orea shales) in the Sierra de Albarracín, Spain – Palaeogeogr. Palaeoclim. Palaeoecol. 48.
Glen, R. A. 1982 Component migration pattern during the deformation of a metamorphic layering, Mount Franks area, Willyama Complex, N.S.W., Australia – J. Struct. Geol. 4: 457-467.
Hafenrichter, M. 1979 Paläontologisch-ökologische und lithofazielle Untersuchungen des 'Ashgill-Kalkes' (Jungordovizium) in Spanien – Arb. Paläont. Inst. Würzburg 3: 139 pp.
—— 1980 The lower and upper boundary of the Ordovician System of some selected regions (Celtiberia, Eastern Sierra Morena) in Spain. II: the Ordovician/Silurian boundary in Spain – Neues Jahrb. Geol. Paläont. Abh. 160: 138-148.
Hambrey, M. J. & W. B, Harland 1981 Earth's pre-Pleistocene glacial record – Cambridge University Press: 1004 pp.
Hammann, W. 1976 The Ordovician of the Iberian Peninsula – a review. In: Bassett, M. G. (ed.): The Ordovician system: Proc. Pal. Ass. Symp. (Birmingham, 1974): 387-409.
Hernando, S. 1980 Mapa geólogico del Permico y Triasico de la region Ayllon-Atienza – Cuad. Geol. Ibérica 6: 21-54.
Heward, A. P. & H. G. Reading 1980 Deposits associated with a Hercynian to late Hercynian continental strike-slip system. Cantabrian Mountains, northern Spain. In: Ballance, P. F. & H. H. Reading (eds): Sedimentation in oblique-slip mobile zones – Int. Assoc. Sediment. Spec. Publ. 4: 105-125.
Hinkelbein, K. 1969 El Triásico y el Jurásico de los alrededores de Albarracín – Teruel 41: 35-75.
Lotze, F. 1945 Zur Gliederung der Varisziden der Iberischen Meseta – Geotekt. Forsch. 6: 78-90.
Matté, P. & R. Capdevila 1973 La structure de la chaine hercynienne dans l'est de la Péninsule Ibérique – Acad. Sci. C.R. Ser. D. 276: 2867-2870.
Navidad, M. 1983 El vulcanismo permocarbonifero de la peninsula iberica. In: Martinez Diaz, C. (ed.) Carbonifero y Permico de Espana – Congreso Intern. Estrat. Geol. del Carbonifero (Madrid, 1983): 473-482.
Orti Cabo, F. 1974 El Keuper del Levante espanol – Estud. Geol. 30: 7-46.
—— 1981 Diapirismo de materiales triasicos y estructuras de zocalo, en el sector central Valenciano (Espana) – Estud. Geol. 37: 245-256.
Platt, J. P. & R. L. M. Vissers, 1980 Extensional structures in anisotropic rocks – J. Struct. Geol. 2: 397-410.
Riba, O. 1959 Estudio geológico de la Sierra de Albarracín – Inst. 'Lucas Mallada', Cons. Sup. Inv. Cient. Monogr. 16: 283 pp.
Robardet, M. 1981 Late Ordovician tillites in the Iberian Peninsula. In: Hambrey, M. J. & W. B. Harland (eds): Earth's pre-Pleistocene glacial record – Cambridge University Press: 585-589.
Stephens, M. B., M. J. Glasson & R. R. Keays 1979 Structural and chemical apsects of metamorphic layering development in metasediments from Clunes, Australia – Am. J. Sci. 279: 129-160.
Turnit, P. 1967 Paläozoikum und mesozoischer Rahmen im SE-Teil der Sierra de Tremedal (Hesperische Ketten/Spanien) – Neues Jahrb. Geol. Paläont. Abh. 127: 349-369.
Viallard, P. 1982 Les virgations de la Chaine Ibérique: déformation rotationelle en régime de cisaillement par décrochement. Acad. Sci. C.R. Ser. II, 295: 765-768.
Weijermars, R. 1982 Definition of vergence – J. Struct. Geol. 4: 505.
Williams, P. F. 1972 Development of metamorphic layering and cleavage in low grade metamorphic rocks at Bermagui, Australia – Am. J. Sci. 272: 1-47.