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Relations between stages of diagenetic and metamorphic evolution and the development of a primary cleavage in the northwestern Moroccan Meseta

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Abstract—The metamorphic evolution of the studied region is determined by the illite crystallinity index and by the evolution of the other phyllosilicates, essentially chlorite and mixed-layer minerals. Several zones are described, at a regional scale, characterized respectively by a diagenetic, anchimetamorphic and epimetamorphic evolution. The estimation of total strain is approached by the type of the cleavage in greywackes. Mapping of all these parameters reveals, at the scale considered, a close relationship between metamorphism and strain, which is briefly discussed.

Résumé—L'évolution métamorphique de la région étudiée est déterminée par la valeur de l'indice de cristallinité de l'illite et par le stade d'évolution des autres phyllosilicates, en particulier la chlorite et divers interstratifiés. A l'échelle régionale, on décrit plusieurs zones, caractérisées respectivement par une évolution diagénétique, anchimétamorphique et épimétamorphique. L'estimation de la déformation totale est réalisée par la détermination du type de la schistosité rencontrée dans les grauwackes. On met ici en évidence une évolution parallèle du métamorphisme et de la déformation, montrant qu'à l'échelle de la région considérée, il existe une relation entre l'intensité du flux thermique synschisteux et celle de la déformation totale.

INTRODUCTION

THE NORTHWESTERN Moroccan Meseta (Fig. 1) is a part of the Hercynian (= Variscan) belt, where the Paleozoic rocks were folded during Carboniferous times (Piqué, 1979). A major regional shear zone, trending NNE–SSW, separates two less deformed domains: the Coastal Block to the west and the Inner Mesetan Zone to the east (Piqué *et al.* 1980). In this part of the Meseta, a complete transition between sediments without any recrystallization and rocks which have been slightly metamorphosed was studied. In addition, different stages of cleavage development in a monophase deformation were described. This region, therefore, contains suitable rocks for a study of the relationship between metamorphism and deformation. Three sections were examined in detail and compared with numerous sampling sites, in order to draw maps which showed the areal distribution of metamorphism and deformation. In this paper only one of these detailed sections will be described to support the general maps.

PHYLLOSILICATES

Illite crystallinity

Most of the region studied was formerly regarded as non metamorphic. Nevertheless, detailed microscopic observations revealed, in some zones, mineral assemblages characteristic of an obvious metamorphic evolution. Hence, it appeared necessary to describe the state of the diagenetic and metamorphic evolution of the whole domain. A now classical method to describe this evolution, in non- or weakly-metamorphosed series, is the evaluation of the crystallinity of illite (Weaver 1960, 1961, Kubler 1964, Dunoyer de Segonzac & Kubler 1966, Dunoyer de Segonzac 1969, Weber 1972).

The crystallinity of illite is represented by an index which is the width (mm) of the 10 Å peak at half height, in well-defined experimental conditions (Kubler 1964). The variation of this index reveals different chemical and crystallographic processes such as dehydration, fixation of K^+ and ionic changes inside the layers. All these processes are steps in the evolution of detrital illite toward true metamorphic white mica. This index helps to define the diagenetic domain (Dunoyer de Segonzac 1969), the anchizone (Kubler 1964), equivalent to the zone of very low grade metamorphism (Winkler 1976) and the epizone (Grubenmann 1904).

The crystallinity of illite is now commonly employed in regional studies (Weber 1972, Teichmüller *et al.* 1979, Frey *et al.* 1980, Kisch 1980). Some observations (Kubler 1968) should, however, be kept in mind in order to avoid possible misinterpretations. Firstly, it is a statistical method: even in the <2 μm fraction, the state of the illite evolution may vary in the sample from one crystal to another and the diffractogram integrates the various individual illite crystallinities. Secondly, illite crystallinity is dependent on many variables, principally the lithology of the sample and the palaeosedimentary environment: sandstones and other porous rocks frequently have higher illite crystallinity indices than contiguous shales or argillites, because they allow easier circulation of ionic solutions. A cold climate on the continent would inhibit or restrain mineral degradations so that the sediments accumulating in neighbouring parts of the basin could have a higher illite crystallinity than those evolving under tropical climates. So, in this study, in order to allow comparisons, the sampling was made initially in the same lithology and at the same stratigraphic level: shales and argillites of the Korifla River Formation, of early Viséan age (Piqué 1979). The section described here follows the valley of the Korifla River over a distance of 40 km (Fig. 1, A–A′). The regional structural trend is here roughly N–S.

The crystallinity indices of illite (Fig. 2) show an evolution from high values, characteristic of sedimentary minerals, to low values, corresponding to white micas. In the northern part of the section (A), inside the same zone, values of the crystallinity index present a large dispersion. This is due to the coexistence, here, of sediments coming from various environments: some have suffered a brief and essentially mechanical disaggregation and a short transport; consequently, the detrital illites are almost unaltered and they keep a relatively high crystallinity. Others, on the contrary, have been subjected for a long time to hydrolysis and the various processes of chemical alteration. They are deeply altered, and their illite crystallinity is low. Progressively, from the north to the south of the section, the dispersion of the index values inside the same region diminishes: the diagenetic and metamorphic aggradation (Lucas 1962, Dunoyer de Segonzac 1969) 'scars over' the degraded illite minerals whose crystallinity increases.

Is it possible to enlarge this study from the section to the region? Although the illite crystallinity indices should be rigorously compared only inside the same lithology and at the same stratigraphical level, an attempt has been made, nevertheless, to compare the values obtained in several (more than 200) outcrops spread over a region centred on the previous section (Fig. 3). The lines of 'illite isocrystallinity', which join points of equal values of the index, cross stratigraphic boundaries. This fact indicates that the illite crystallinity, in the studied region, is not totally controlled by the lithology or other sedimentological conditions, and it warrants the enlargement of the study to the regional scale. By considering the isocrystallinity lines, a coherent pattern appears: a domain totally free of any metamorphic evolution, around point A of the previously studied section, is bordered by more or less concentric zones, where the diagenetic, anchimetamorphic and finally epimetamorphic stages are indicated by the illite crystallinity index values.

Other phyllosilicates of the clay fraction

The determination, by X-ray diffractometry, of the

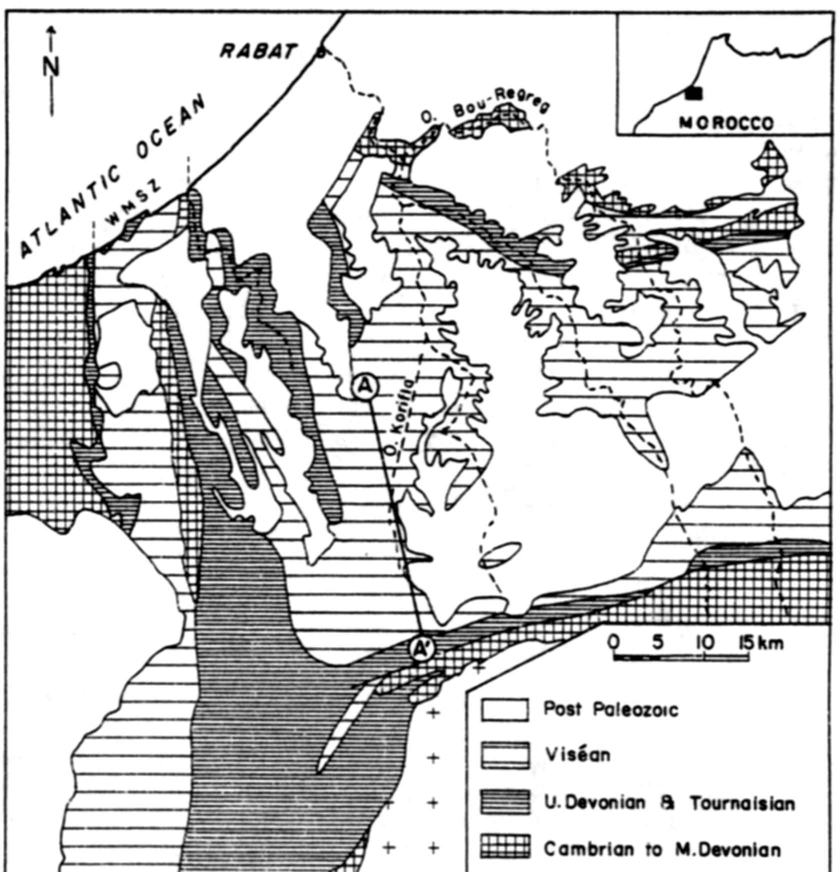

Fig. 1. Simplified geological map of the northwestern Meseta. WMSZ, Western Mesata Shear Zone. A–A′ is the line of section described.

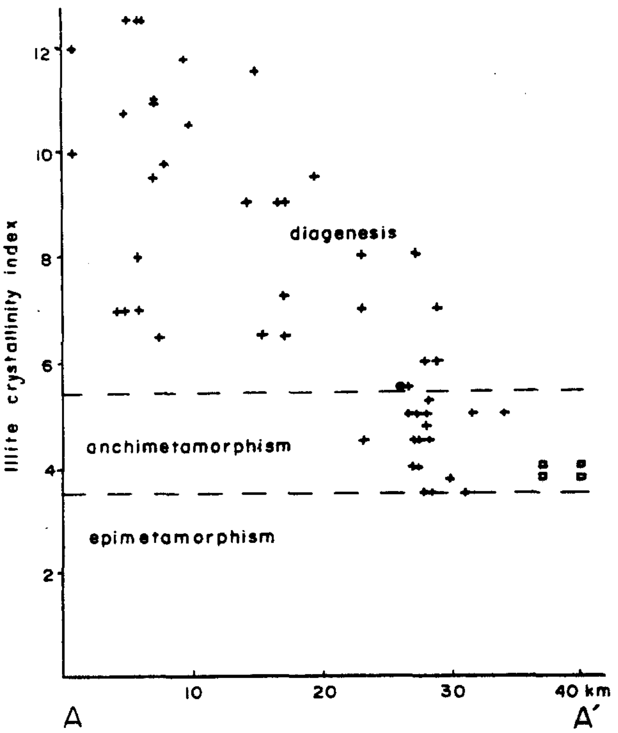

Fig. 2. Illite crystallinity indices of samples along the section A–A'; crosses, Lower Viséan; circles, Upper Viséan and squares, Tournaisian. Illite crystallinity decreases from the north (A: Fig. 1) toward the south (A').

other phyllosilicates was done in order to verify the results given by the illite crystallinity and to specify the general mineralogical evolution. In the studied section illite is the most abundant mineral; often it represents 50% of the clay fraction, whereas chlorite forms approximately 25% and the other clay minerals 25% of this fraction.

Chlorite. It does not occur in the northern part of the section, but it appears in the anchimetamorphic domain. The I(002)/I(001) ratio (Brown 1955, Brindley & Gillery 1956), of between 0.16 and 0.45, indicates an Fe-rich chlorite. No significant variation of this ratio can be found over the length of the section. Regionally, the chlorite distribution is parallel to the pattern of the illite crystallinity (Fig. 4). True chlorite is absent in the centre of the region and around the northern end of the studied section and it appears in the samples for a higher value of the diagenetic evolution, not far from the anchizone limit.

Kaolinite. Its localization is roughly symmetric about the chlorite occurrences: it is present in the north of the section and absent in the south.

Smectite and pyrophyllite. These minerals are almost absent. Pyrophyllite is restricted to occurrences outside the diagenetic domain.

Mixed-layer minerals. Other phyllosilicates are frequently found in the diagenetic domain. These minerals are constituted by alternating layers of different compositions. For instance, the alternation of 10 Å (mica)–layers and 14 Å (chlorite)–layers inside the

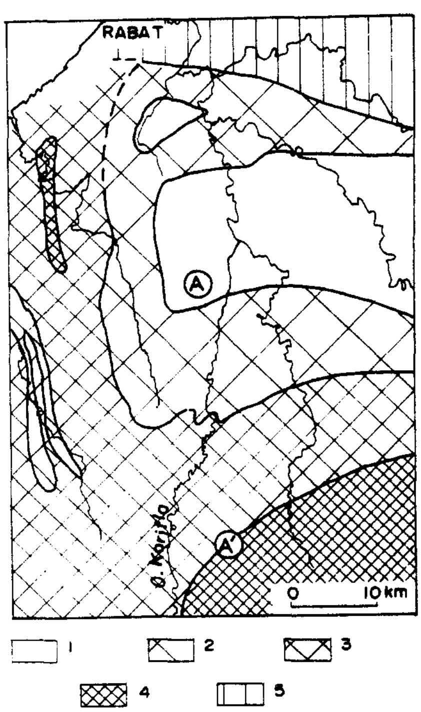

Fig. 3. Regional distribution of illite crystallinity. 1 and 2, diagenetic zone (1 illite crystallinity index <10; 2, index between 10 and 5.5); 3, anchimetamorphic zone; 4, epimetamorphic zone and 5, pre-Hercynian metamorphism. Main rivers marked. Geographic frame —see Fig. 1. A–A': cross-section (Fig. 2).

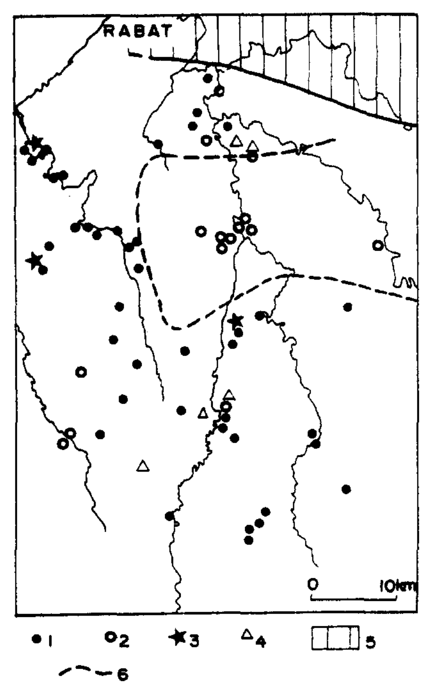

Fig. 4. Regional distribution of some clay minerals. 1, occurrence of chlorite; 2, occurrence of kaolinite; 3, occurrence of pyrophyllite; 4, occurrence of smectite; 5, pre-Hecynian metamorphism and 6, limit of the chlorite distribution. Chlorite is absent in the centre of the region, where the illite crystallinity is low (Fig. 3), but it is a common mineral inside the anchimetamorphic and epimetamorphic zones. Kaolinite is present mostly in the centre of the region and in western zones, where the crystallinity of illite is low. Pyrophyllite and smectite are rare, and thus their distribution is not significant.

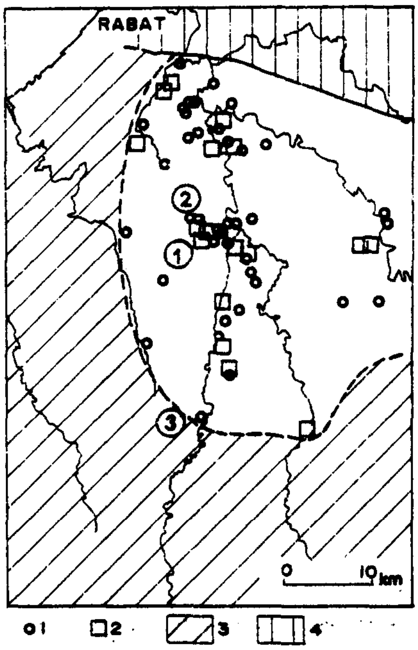

Fig. 5. Regional distribution of some mixed-layers minerals. 1, occurrence of (10–14c) (illite–chlorite); 2, occurrence of (10–14s) (illite–smectite); 3, zone free of (10–14c) and (10–14s) minerals and 4, pre-Hercynian metamorphism. These mixed-layers minerals are distributed in the central part of the map, where the crystallinity of the illites is low and chlorite is absent (see Figs. 3 and 4). 1, 2 and 3 circled on the map are positions of the samples plotted in Fig. 6.

same crystal constitutes the 10–14c (illite–chlorite) mixed-layer mineral. This mineral is regular when it is constituted by an equal amount of 10 and 14 Å layers. It is then characterized by a 12 Å peak. In the studied region 10–14c (illite–chlorite) and 10–14s (illite–smectite) are rarely regular minerals. More often, they are rather traces, giving a discrete and broad reflection between 10 and 14 Å. They disappear completely in the anchimetamorphic and epimetamorphic zones as shown by the cleaning up of the diffractograms (Figs 5 and 6).

The 14c–14v (chlorite–vermiculite) mixed-layer mineral is common in the section and in the whole region. When regular, it distinguishes itself from chlorite by its 12 Å peak after heating. In the studied section and in the surrounding region, its reflection patterns are variable (Fig. 7) and two general features can be described for this reflection: a flat level, broad and weak reflection between 10 and 14 Å, and a sharp peak, well individualized, at 12 Å, which often coexists with the 14 Å of the true chlorite. The first type is localized in the diagenetic zone (Fig. 8), whereas the regular minerals of the second type are found in the anchimetamorphic and epimetamorphic domains: there is a progressive and complete evolution from an irregular stage, near the vermiculite pole, toward true chlorite,

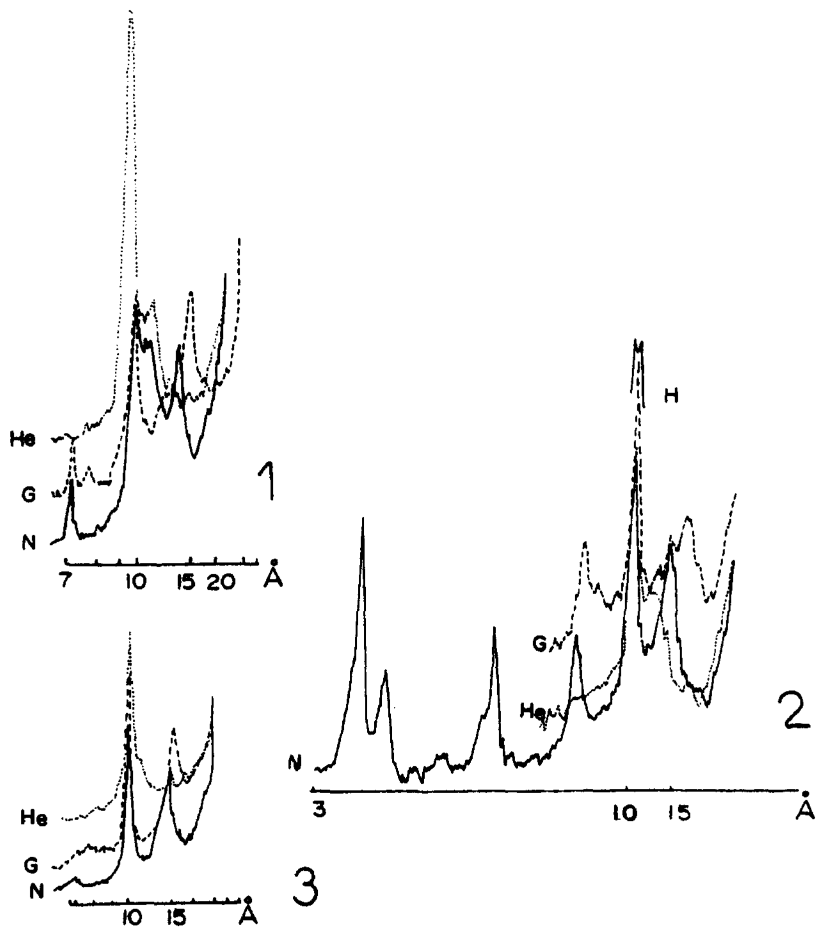

Fig. 6. Phyllosilicates assemblages (<2 μm fraction). 1, point A = illite, (14c–14v), traces (10–14s); 2, 5 km north of point A = illite, kaolinite, (14c–14v), smectite and 3, 22 km south of point A: illite, chlorite. N, normal diffraction. Special treatments = H, hydrazine; He, heating, G, glycol. Values in Å. Samples localized in Fig. 5.

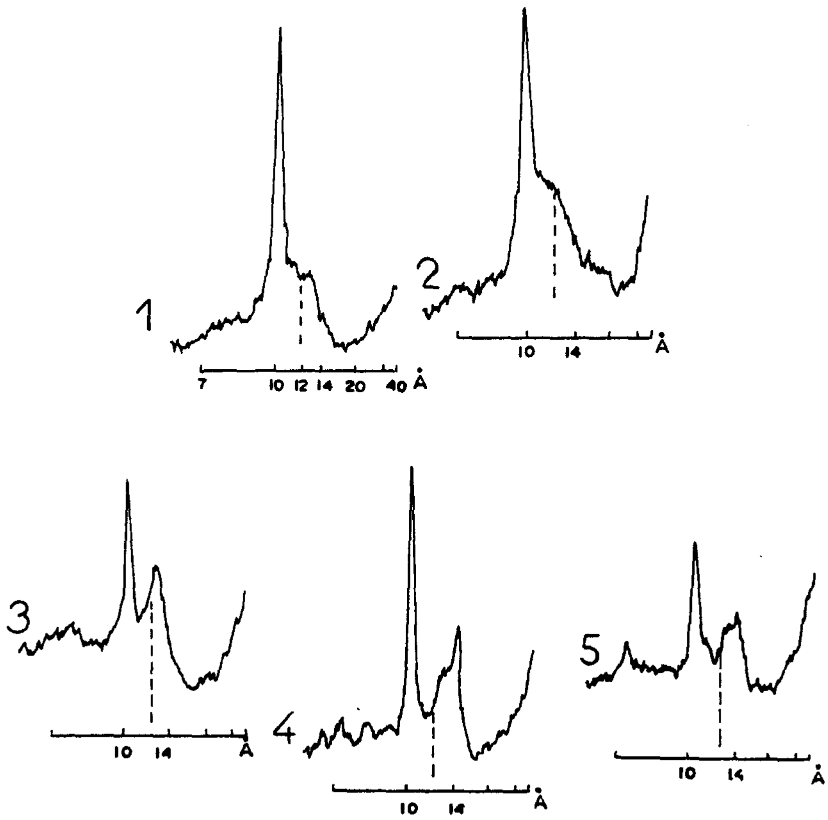

Fig. 7. (14 c–14 v). X-ray diffractograms, after heating. 1 and 2, weak and broad reflection between 10 and 14 Å: irregular stages. 3, 4 and 5, peak centered on 12 Å or moved to 14 Å: regular stage, transition to chlorite.

parallel with the evolution of the other phyllosilicates.

The regional evolution

The centre of the studied region (Fig. 3) is characterized by the lack of any metamorphic evolution. The sedimentary assemblages are still preserved, with a range of detrital argillaceous minerals: illite, kaolinite and various mixed-layer minerals. From the centre of this domain towards its boundaries, the mineralogical evolution becomes obvious: illite gains a better crystallinity whereas kaolinite is destroyed and mixed-layer minerals evolve towards higher temperature minerals such as chlorite, by the way of mineralogical changes which are not yet completely understood.

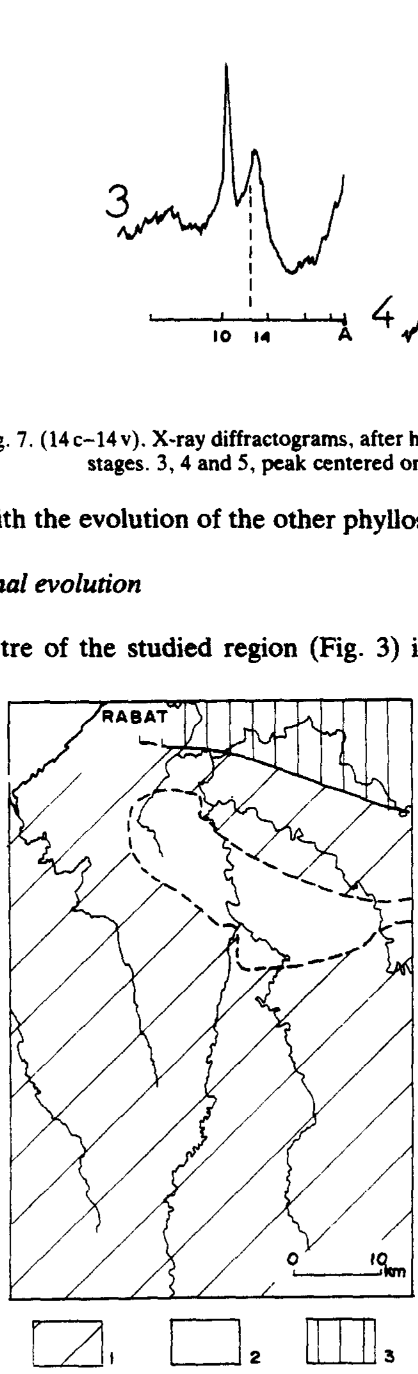

Fig. 8. (14 c–14 v) mineral. 1, regular stage (see Fig. 7); 2, irregular stages and 3, pre-Hercynian metamorphism. The (14 c–14 v) mineral is represented by irregular stages in the central area and it is a regular mixed-layers mineral in the zones where the diagenetic and metamorphic evolution is important (compare with Fig. 3).

DEFORMATION

Cleavage

In the studied section (A–A′, Fig. 1) it is possible to distinguish in the field two zones, a zone where cleavage is obvious in the samples and another zone where no evidence of cleavage can be seen. The junction of these zones is called the 'cleavage front'. For a more accurate description, a systematic study was made, using hand specimens and thin sections. In order to allow comparisons from one sample to another, this study has been restricted to greywacke beds presenting the same lithology: approximately 30% quartz, 30% detrital micas and 40% clay matrix. The study led to the recognition of four fundamental stages in the cleavage initiation and development (Piqué 1981).

Type A, non planar and spaced cleavage. This type of cleavage is not perceptible in hand specimen; it appears only at the microscopic level of observation. At that scale, the cleavage is outlined by brown seams, opaque-rich domains several micrometers wide. These seams, which include detrial opaques and Fe–Mg rich phyllosilicates (Gray 1978, White & Johnston 1981), are discontinuous and up to 1 mm long. They appear in the argillaceous matrix and never penetrate the detrital grains of quartz or mica, which are embedded in it. On the contrary, they wrap around them, giving the rock its discontinuous and non planar pattern (Fig. 9a). Often, they are gathered in narrow cleavage stripes, separated by wider zones where no cleavage is apparent. In terms of a morphological classification of cleavage (Powell 1979), it is an anastomosing cleavage where the cleavage domains are broad (0.5 mm to 1 cm) and the degree of microlithon – microfabric realignment, random.

Type B, planar and spaced cleavage. The brown seams are here longer (usually several millimeters) and more regular (Fig. 9b), although they still wrap around the largest detrital grains. The cleavage planes are always separated; between adjacent planes are microlithons, where the initial fabric is preserved and sometimes microfolded. In the centre of the microlithons, the detrital elements are at a high angle to the trace of the cleavage plane and within them quartz grains keep their sedimentary shape. Toward the microlithon boundaries the mean volume of quartz grains decreases, their shapes become more flattened and they are, like the detrital micas, more or less completely reoriented in the cleavage planes. Thus, this B type represents a rough cleavage where the cleavage domains are 0.3–1 mm broad, and the degree of microlithon–microfabric realignment is weak to moderate.

Type C, planar and penetrative cleavage. In the argillaceous beds (Fig. 9c), cleavage planes constitute the main anisotropy of the rock. The cleavage is almost smooth (Powell 1979), its traces are regular and the thickness of the microlithons is about 30–50 μm. Between two adjacent cleavage planes, the realignment of the sedimentary fabric is complete. Most of the detrital components are reoriented in the cleavage planes; only the largest detrital quartz and mica grains are still at a high angle to the trace of the cleavage planes. All the detrital quartz grains have lost their sedimentary shape and they are characterized by well-developed sharp ends located upon the extremities of the grains. Tiny white micas (10–20 μm) develop parallel to the cleavage planes, sometimes as 'mica beards' (Williams 1972). In the sandy beds, cleavage planes are irregular and an unchanged sedimentary fabric persists in the centre of the microlithons.

Type D, continuous cleavage. The cleavage is perfectly penetrative at the scale of observation given by the optical microscope and no trace of the sedimentary fabric remains (Fig. 9d). This fabric affects the sandy beds as well as the pelitic beds, and all the minerals are disposed parallel to the trace of the cleavage, except the lozenge-shaped chlorites (Weber 1976, Roy 1978). In other words, the tectonic fabric is completely realized at that scale, and the cleavage is continuous (Powell 1979).

The regional evolution

It is suggested that a rock which, for example, presents a C-type of cleavage has been characterized, during the incipient stages of the deformation, successively by the A and B types. So, the spatial relationship between the cleavage types represents a developmental sequence, and the cleavage types described here are stages in the progressive development of the cleavage.

On the 40 km long section previously described, the distribution of the cleavage types shows a regular pattern (Fig. 10): at the northern end of the section, cleavage is absent or represented at most by the incipient stages (A above). Toward the south, the cleavage becomes more

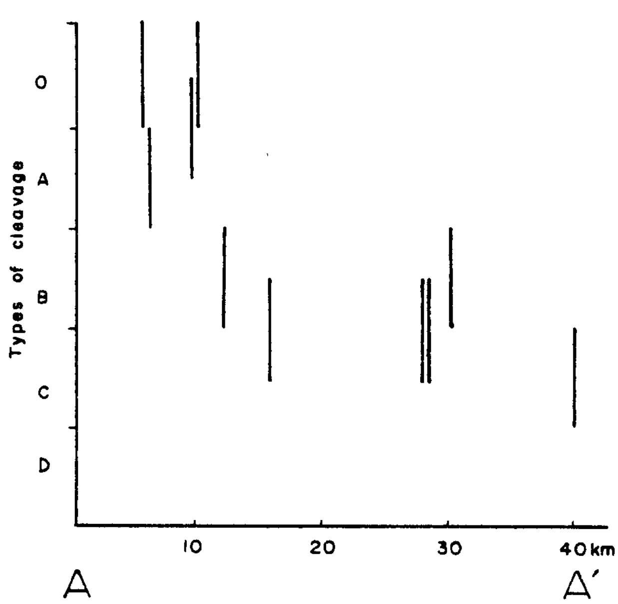

Fig. 10. Evolution of the cleavage type along the A–A′ section. From the north (point A) toward the south (point A′) the cleavage becomes more and more penetrative.

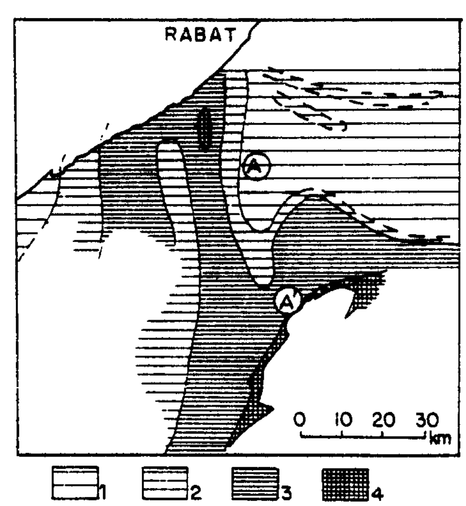

Fig. 11. Regional distribution of cleavage types. 1, no cleavage or very incipient cleavage (type A); 2, type B; 3, type C and 4, type D. A weakly-deformed (and unmetamorphosed, see Figs. 3, 4 and 8) domain extends at the SE of Rabat. It is separated from another one, WSW of Rabat, by a more deformed and complex domain. A–A′, studied section.

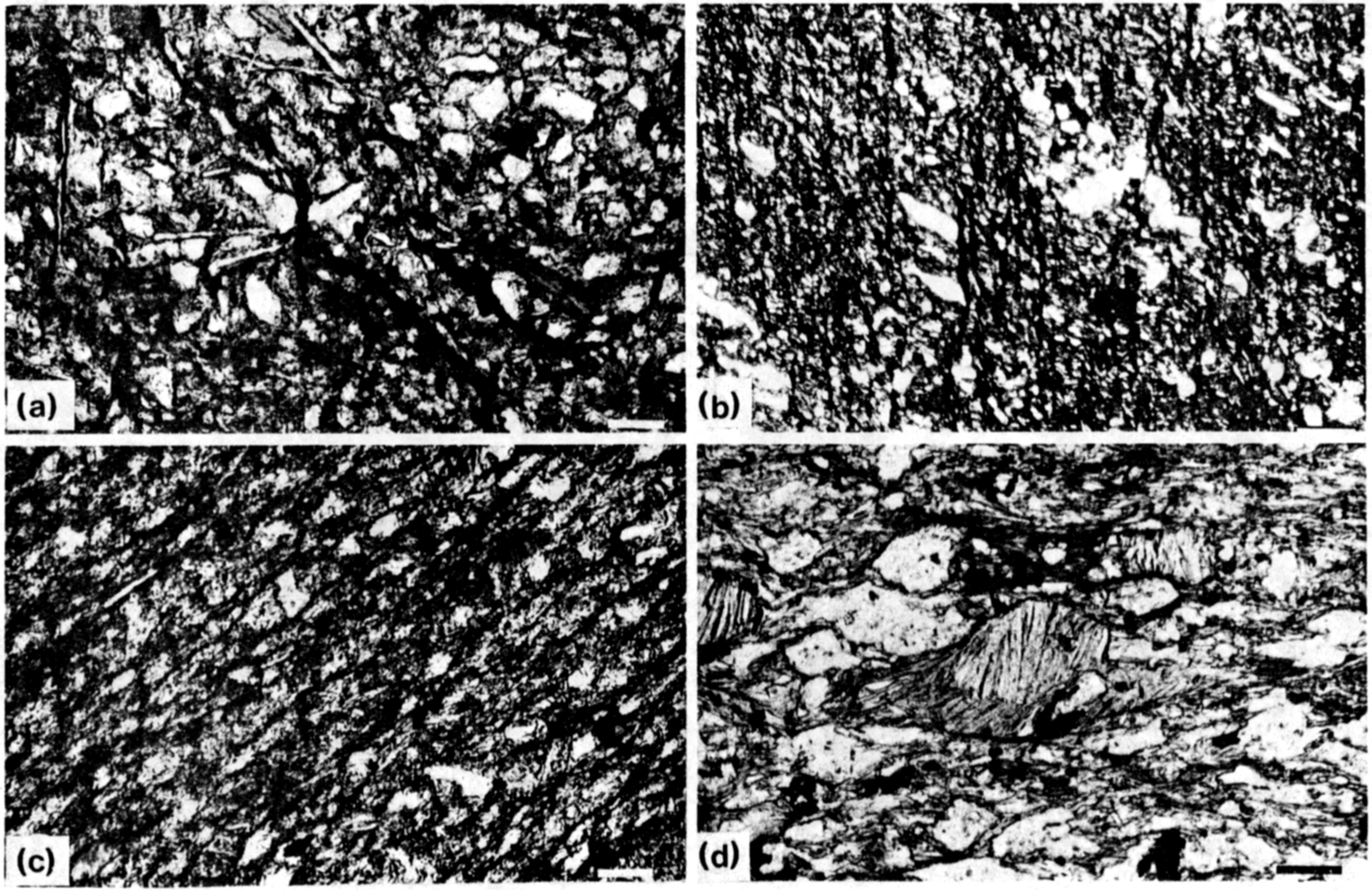

Fig. 9. Cleavage types. (a) Type A. The incipient cleavage is represented by irregular and discontinuous brown seams. Throughout the picture, the sedimentary fabric is conspicuous. Bedding trace is parallel to the length of the picture, cleavage trace is oblique. Plane polarized light. Scale bar, 0.2 mm. (b) Type B. The brown seams, regular and continuous, separate microlithons where the sedimentary fabric is preserved. Cleavage trace is parallel to the width of the picture. Plane polarized light. Scale bar, 0.5 mm. (c) Type C. The sedimentary fabric is almost destroyed inside the microlithons. Except for the biggest ones, the detrital minerals are parallel to the cleavage planes. Plane polarized light. Scale bar, 0.2 mm. (d) Type D. The tectonic fabric is perfectly realized. Detrital minerals as well as newly formed phyllosilicates are parallel to the cleavage planes. Note the presence of lozenge-shaped chlorites. Plane polarized light. Scale bar, 0.2 mm.

and more accentuated until it becomes penetrative (type D above). At the regional scale, in the domain extending around the northern end of the studied section, the cleavage is generally absent and the beds are bent in open chevrons or sometimes almost unfolded (Fig. 11). This domain is bordered by zones where the cleavage is conspicuous.

In the studied greywackes as in other lithologies, the cleavage development is related to several mechanisms where solution transfer plays an important role (Gray 1978, Beutner 1978, Alvarez *et al.* 1978). Therefore, the cleavage type may be an approximate estimate of the shortening suffered by the rock. Indeed, preliminary results along the A–A′ studied section suggests a concurrent development of cleavage and total strain measured here by the homogeneous flattening in concentric folds (Piqué 1979).

CONCLUSIONS

A common observation in the field is that the cleavage generally develops as the metamorphic grade increases and that the 'cleavage front' is roughly localized within the anchizone. The present study confirms this result on a regional scale: there is here a concurrent progressive development of cleavage and metamorphism. The increase in metamorphism in the studied A–A′ section occurs within the same stratigraphic level and at the same depth. If we assume that the tectonic stress did not vary at the same time throughout the region, the variation of the cleavage type could be due, at constant lithostatic pressure, to the variable temperature to which the rocks have been subjected. The results of this study lead, as in other examples (Brazier *et al.* 1979), to the recognition and reconstitution of the palaeo-heat flow.

Regional studies, which associate maps of intensity of deformation and maps of metamorphism (Le Corre 1978, this study), may help to answer the question of the control of deformation by metamorphism (Jones 1981). In the past, numerous workers have distinguished two independent phenoma: deformation, often estimated by the presence of the cleavage, and metamorphism. For them, the planar fabric results only from the reorientation of detrital minerals under a non-hydrostatic stress. This has been proposed for the cleavages found in some synsedimentary folds (Williams *et al.* 1969) and for some slates (Maxwell 1962, Clark 1970), where the cleavage initiated in a soft and water-rich sediment. It has been claimed that this could be the general origin for slates; the fluidizing agent, permitting the rotation of the detrital minerals, being either pore water (Braddock 1970), or the adjacent crystals themselves (Tullis 1976). After this mechanical reorientation, a subsequent metamorphism can produce a mimetic recrystallization of the clay-size minerals. Following these ideas, deformation and metamorphism are not causally related.

Here, cleavage initiation and development is described as a more complex and progressive mechanism, which includes mechanical rotation and shape changes of detrital minerals (mostly by pressure solution in the conditions of very low-grade metamorphism) and also syntectonic recrystallization (Weber 1981, Knipe 1981). This is confirmed by numerous observations in slates (Wood 1974, White & Knipe 1978, Beutner 1978, Roy 1978, Siddans 1980) as well as in rough cleavage (Gray 1978, Piqué 1981) and in the laboratory experiments (Williams *et al.* 1977). Indeed, the cleavage problem, with its structural and metamorphic aspects may be studied by several routes depending upon the scale of the object studied. At the individual mineral scale (Knipe 1979, 1981, White & Johnston 1981) and at the scale of a small fold (Knipe & White 1977), it is shown that the mineralogy of the crystals depends upon their structural site. At the regional scale, sections described in a continuous series show a parallel evolution of the amount of strain, measured for instance by the flattening of reduction spots (Siddans 1980) and metamorphism, indicated in particular by illite crystallinity (Siddans 1977, 1980).

The present study illustrates, over a larger region, the parallel development of metamorphism and cleavage and it suggests that these phenomena are related in the region studied. Both scales of investigation are necessary and useful. It is hoped that they will converge to a more complete and accurate understanding of the processes involved in synmetamorphic deformation.

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