DIAGENETIC AND METAMORPHIC EVOLUTION OF CAMBRO-ORDOVICIAN SILICICLASTIC SEDIMENTS IN NORTHWESTERN SPAIN[1]

GIJSBERT GIETELINK[2]

ABSTRACT

Gietelink, G. (1978). Diagenetic and metamorphić evolution of Cambro-Ordovician sediments in northwestern Spain. Geol. Mijnbouw, 57, p. 65-76.

The diagenetic and metamorphic evolutions of the Cambro-Ordovician siliciclastic sediments of the Luna area, which formed part of the Lower Palaeozoic stable shelf, and of the Sil area, which was situated in the adjacent geosyncline, were determined and compared.

The overburden in the Luna area reached 2500-3500 m, based on stratigraphic evidence. This is in agreement with the mineral association occurring in the sediments (authigenic illite with minor amounts of detrital kaolinite and authigenic muscovite and chlorite), indicating a depth of burial of 2000-3000 m. The sediments were subject to diagenetic changes, which occurred shortly after deposition before lithification (syndiagenetic phase), in the lithified rocks (anadiagenetic phase) and after uplifting and erosion of the area (epidiagenetic phase).

Based on stratigraphic evidence, the overburden of the sediments studied in the Sil area ranged from 5000 to 14000 m. The mineral associations observed in the sediments (illite-muscovite-chlorite in the upper part and muscovite-chlorite-biotite in the lower part of the succession) are not in agreement with the estimated depth of burial. The sediments of the Sil area underwent diagenetic alterations similar to those of the sediments in the Luna area, but the effects are obscured by metamorphic recrystallization, which occurred during the Hercynian orogeny. During this period the area was affected by regional metamorphism.

INTRODUCTION

The sedimentary petrography of Cambro-Ordovician silici-clastic sediments from two different areas in the northwestern part of Spain was studied in an attempt to determine and compare the diagenetic and metamorphic evolutions of the sediments of both areas.

The sediments studied constitute part of the Palaeozoic core of the Iberian Peninsula. This core was folded during the Hercynian orogeny. The Hercynian orogenic belt is subdivided into zones on the basis of stratigraphic, metamorphic, tectonic and igneous characteristics (MATTE, 1968; PARGA, 1970; BARD ET AL., 1971) (Fig. 1). In the Luna area the Oville Formation and the Barrios Formation were studied. These formations consist of shallow marine, tidal and littoral deposits, which range in thickness from 170 to 675 m, and represent a period from Middle Cambrian up to Arenig (Lower Ordovi-
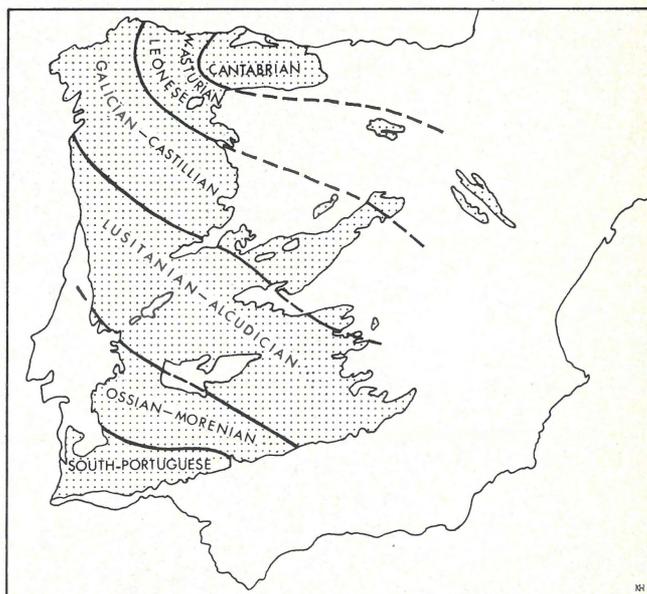

Fig. 1
The Iberian Hercynian orogenetic belt, with the zonation based on stratigraphic, metamorphic, tectonic and igneous caracteristics.

[2] Geological Institute, Garenmarkt 1b, LEIDEN, The Netherlands. Present address: Laan van Sevenstar 11, VOORBURG, The Netherlands.
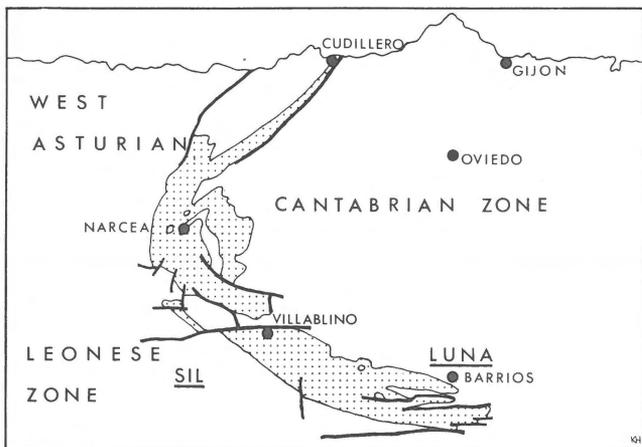
Fig. 2
The Sil area and the Luna area are situated in two different zones and separated by the Narcea Anticlinorium.

cian) inclusive (GIETELINK, 1973). The Luna area is situated near the southwestern border of the Cantabrian Zone, which formed a stable shelf during the Lower Palaeozoic. The Cabos Formation, which is stratigraphically the equivalent of the

above-mentioned formations, was studied in the Sil area. The Cabos Formation reaches a thickness of 8500 m and consists equally of shallow marine, tidal and littoral deposits (Gietelink, in prep.). The Sil area is situated near the northeastern border of the West Asturian-Leoneses Zone, which formed a geosyncline during most of the Lower Palaeozoic (Fig. 2).

Both the original mineralogical composition and the depositional environment were almost identical in both areas. The aim of the study was to compare the influence of stable shelf and geosynclinal conditions on the diagenetic and metamorphic evolution of the sediment.

MINERALOGICAL COMPOSITION

A picture of the variation of the mineralogical composition is given in table I.

Detrital grains

Quartz grains strongly dominate, generally more than 90% of the detrital grains are quartz grains, but sometimes this percentage is lower (80-85% at the minimum). Feldspars, mainly
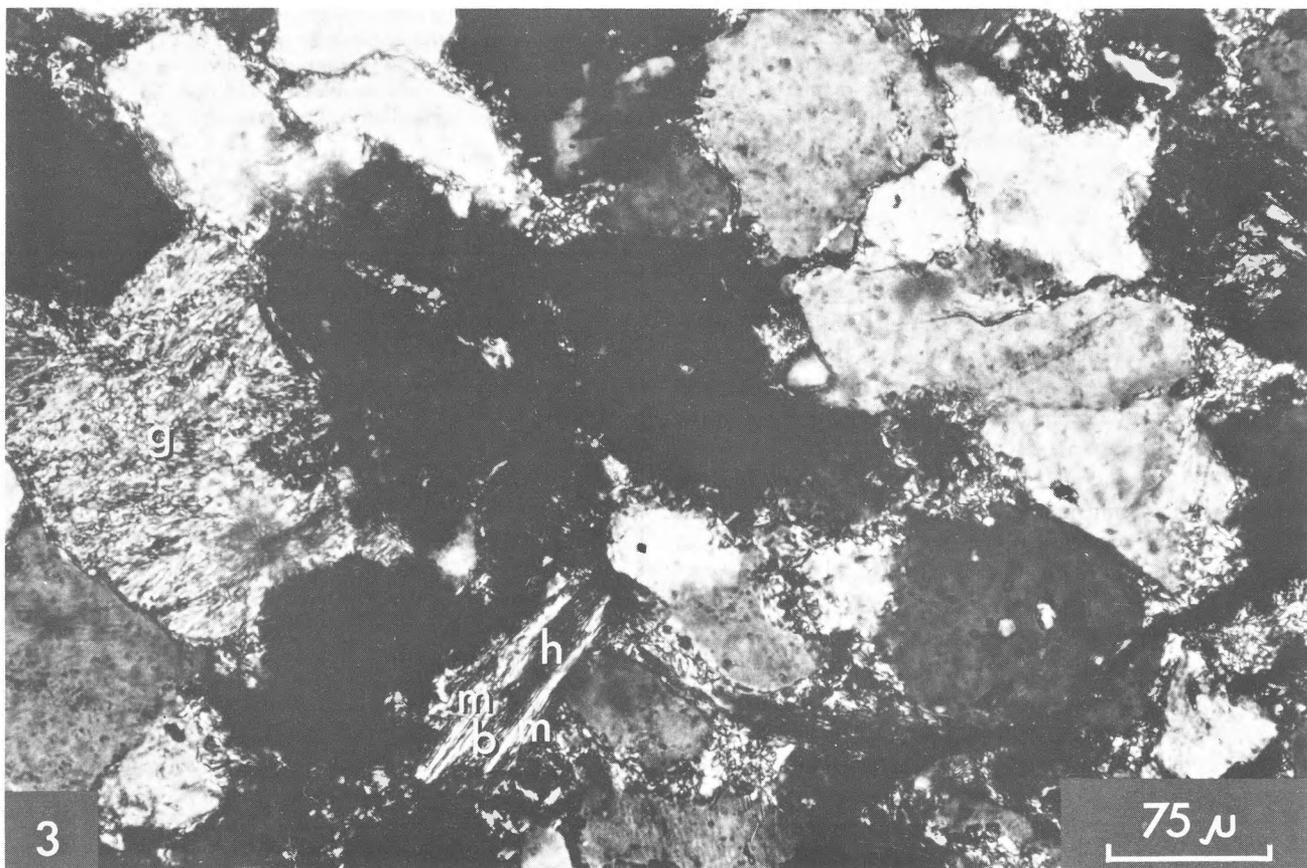

Fig. 3
Altered glauconite grain (g): only the greenish colour has been preserved. Also note the partly altered biotite (b) flake. m = muscovite; h = hematite.

	Luna area									Sil area													
GRAINS	72.2	65.2	64.8	64.0	54.2	54.0	40.0	22.8	16.0	73.6	71.2	69.0	65.2	68.8	73.6	62.0	57.4	50.2	45.0	41.2	29.2	17.0	4.2
Quartz	69.8	59.2	58.0	51.4	50.8	45.6	32.0	17.2	11.6	73.6)	70.2).	65.4)	63.4)	62.4)	63.2)	60.0)	53.0)	45.4)	42.8)	39.2)	28.0)	17.0	3.2
Microcline	–	–	–	5.2	–	2.4	–	0.8	0.4	–	tr.	0.4	1.0	4.0	9.6	0.8	1.6	–	0.6	0.4	–	tr.	–
Plagioclase	–	–	–	–	tr.	–	–	tr.	tr.	–	–	tr.	0.6	1.0	0.2	0.2	tr.	–	0.2	0.4	–	tr.	–
Muscovite	0.2	tr.	tr.	0.2	tr.	0.4	3.4	1.6	1.0	tr.	–	0.2	0.2	0.4	0.6	0.6	0.6	1.2	0.4	0.6	1.0	pr.	1.0
Biotite	–	–	–	tr.	–	0.2	0.8	2.0	1.6	–	–	0.2	–	0.2	–	–	–	0.4	0.2	0.4	–	pr.	–
Kaolinite	–	–	–	–	–	–	0.6	–	0.8	–	–	–	–	–	–	–	–	–	–	–	–	–	–
RF quartz	2.8	3.4	3.4	5.2	3.0	3.8	2.0	1.2	0.2	pr.	0.2+	1.4+	pr.	pr.	pr.	0.2+	1.8+	1.6	0.8+	pr.	tr.	tr.	–
RF shake	–	–	2.4	0.4	0.2	0.4	0.6	–	–	–	0.2	0.2	–	tr.	tr.	–	–	1.0	–	–	–	–	–
RF complex	–	0.2	0.6	1.0	0.6	–	–	–	–	–	0.2	tr.	–	0.2	–	–	–	0.4	–	–	–	–	–
RF vulcanic	–	tr.	tr.	tr.	tr.	–	tr.	–	0.2	–	tr.	–	–	0.2	–	–	–	tr.	tr.	–	–	–	–
Glauconite	–	2.2	–	0.4	–	0.2	0.2	–	–	–	–	1.0	–	0.2	–	–	0.4	–	–	–	–	–	–
Heavy minerals	0.4	0.2	0.4	0.2	0.2	0.4	0.4	tr.	0.2	tr.	0.4	0.2	tr.	0.2	tr.	0.2	tr.	0.2	tr.	0.2	0.2	tr.	tr.
MATRIX	0.4	0.8	8.8	3.6	6.0	12.2	42.6	68.4	63.0	0.2	1.2	1.6	tr.	3.6	2.0	0.8	7.6	6.6	5.8	tr.	11.4	1.8	4.2
AUTHI-GENETIC	26.4	34.0	26.4	32.4	39.8	33.8	17.4	8.8	21.0	23.2	27.6	29.4	34.8	27.6	24.4	37.2	35.0	43.2	49.2	58.8	59.4	81.2	91.6
(matrix alterations)	tr.	0.4	0.2	4.8	3.0	15.0	4.2	3.2	16.0	0.2	1.8	1.6	5.8	3.8	2.4	13.6	5.6	30.2	36.6	50.8	44.4	79.8	91.6
Muscovite	–	0.4	0.2	–	tr.	–	2.8	0.4	tr.	tr.	0.4	1.2	3.6	0.4	0.4	6.0	2.8	24.6	22.6	33.6	38.8	27.2	82.6
Clorite	–	–	tr.	tr.	–	0.4	tr.	0.4	2.8	–	1.4	–	0.4	0.2	–	6.6	0.2	1.8	1.4	5.8	tr.	–	0.8
Biotite	–	–	–	–	–	–	–	–	–	–	–	–	1.2	2.4	–	0.8	–	–	9.8	9.6	–	51.6	–
Iron minerals	tr.	–	tr.	4.8	3.0	14.6	1.4	2.4	13.2	0.2	tr.	0.4	0.6	0.8	2.0	0.2	2.6	3.8	2.8	1.8	5.6	1.0	8.2
(others)	26.4	33.6	26.2	27.6	36.8	18.8	13.2	5.6	5.0	23.0	25.8	27.8	29.0	23.8	22.0	23.6	29.4	13.0	12.6	8.0	15.0	1.4	tr.
Quartz	26.4	33.4	25.8	26.8	25.2	17.4	3.4	tr.	0.2	23.0)	24.0)	26.0)	29.0)	23.0)	22.0)	23.0)	25.0)	11.0)	10.0)	8.0)	2.4	1.4	tr.
Calcite	–	–	–	10.4	–	–	–	–	–	–	–	1.6	–	–	–	0.4	4.0	–	–	–	11.2	–	–
Muscovite	–	0.2	–	0.4	1.0	0.8	8.2	4.0	3.4	tr.	–	0.2	tr.	0.8	–	0.2	–	1.8	2.0	tr.	1.4	pr.	–
Chlorite	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–	tr.	–	0.2	0.6	tr.	–	–	–
Heavy minerals	–	–	0.4	0.4	0.2	0.6	1.6	1.6	1.4	–	1.8	–	–	tr.	–	tr.	0.4	tr.	tr.	tr.	tr.	tr.	tr.
QUARTZ TOTAL	99.0	96.0	87.4	83.8	79.0	66.8	37.4	18.4	12.0	99.6	94.4	92.8	92.4	85.4	85.2	83.2	79.8	58.0	53.6	47.2	30.4	18.4	3.2

) = ratio primary to secondary quartz incertain
+ = probably higher value
tr. = trace
pr. = present but not distinguishable

Table I
Mineralogical composition in some characteristic thin sections of the various rock types, determined by point-counting (500 points).

microclines and a few plagioclases, occur locally in the Oville Formation of the Luna area (up to 6% of the sediment). In the Sil area feldspars are more evenly distributed over the succession and occur more frequently (up to 10%). Micas may form up to 7% of the sediment. At present mainly muscovite is found. Biotite, as well as kaolinite, has been subject to severe alteration during diagenesis. Kaolinite only occurs locally in very small amounts in the Luna area. Glauconite is found rather frequently (up to 2.5%) in the basal part of the succession in the Luna area, but occurs only locally in the sediments of the Sil area. Rock-fragments may form up to 7% of the sediment. Composite quartz grains (quartzite, polycrystalline and chert rock-fragments) and shaly rock-fragments occur rather frequently. Composite argillaceous rock-fragments containing silt-size quartz grains and/or large amounts of iron hydroxides were found less frequently (up to 0.5%). Rock fragments having a cryptocrystalline texture, which shows no preferred orientation, and that are partly isotropic were found locally. They are supposed to be of volcanic origin and may form up to 0.5% of the sediment. The most important detrital heavy minerals are, in decreasing order of frequency of occurrence: tourmaline, zircon and rutile. Opaque grains, alterites, cassiterite, epidote-group minerals, staurolite, olivine, augite and garnet occur in very small quantities.

Detrital matrix

The term matrix is used to indicate material of a grain size smaller than 20 μ. Röntgen diffractograms made of matrix material from the Luna area show that the bulk consists of well-crystallized illite. Kaolinite, quartz, muscovite, iron minerals and some chlorite form minor components. The ob-

served high degree of crystallinity of the illite and chlorite suggest that these minerals were formed by recrystallization of the original matrix clay minerals (OELE, 1964). In the Sil area well-crystallized illite, muscovite, chlorite and biotite, which are microscopically determinable, were formed during the diagenetic and metamorphic evolution from the original clay-minerals.

Authigenetic minerals and recrystallizations

Quartz cement, which was formed in the pore spaces by the secondary enlargement of quartz grains, occurs in all sandstones and siltstones from both the Luna and the Sil area. Calcite cement, replacing matrix material and secondary quartz overgrowths, is found occasionally in the Luna area (in two slides from the same bed) and occurs in a few beds of the Cabos Formation in the Sil area. Calcite cement occurs in patches and often contains large quantities of iron oxide and hydroxide inclusions (Fig. 4). Barite cement locally occurs in small quantities in the quartz sandstones of the Barrios Formation of the Luna area (Fig. 5). Authigenetic muscovite is very common in the deposits studied, especially in those from

the Sil area. Slides from the Sil area always show muscovites which were formed from the matrix clay-minerals by recrystallization (Fig. 6). In the Luna area authigenetic muscovite occurs in small amounts in the matrix. Muscovite was also formed during decomposition of biotite. Mica flakes, one lamella being muscovite and the adjacent lamella discoloured biotite, are frequently observed (Fig. 7 a, b). Muscovite is also one of the alteration products of glauconite (Fig. 7). Chlorite was formed by recrystallization of the matrix material. In the Sil area the majority of the succession shows this kind of chlorite, in the Luna area it occurs locally in small amounts (less than 3%) (Fig. 8 a, b). In the Sil area mica flakes composed of muscovite and chlorite lamellae are frequent, the original mica flake probably was a biotite flake. Chlorite was generated in small amounts in the alteration of glauconite. In the Sil area, especially in the lower part of the Cabos Formation, biotite is frequent. These biotites are a recrystallization product of matrix material and authigenetic chlorite (Fig. 9 a, b). Anatase is a very common accessory mineral and occurs scattered over the sediment or as inclusions in authigenetic muscovite (Fig. 7). Large parts of anatase crystals have been altered to leuxocene. Epidote group minerals occur in
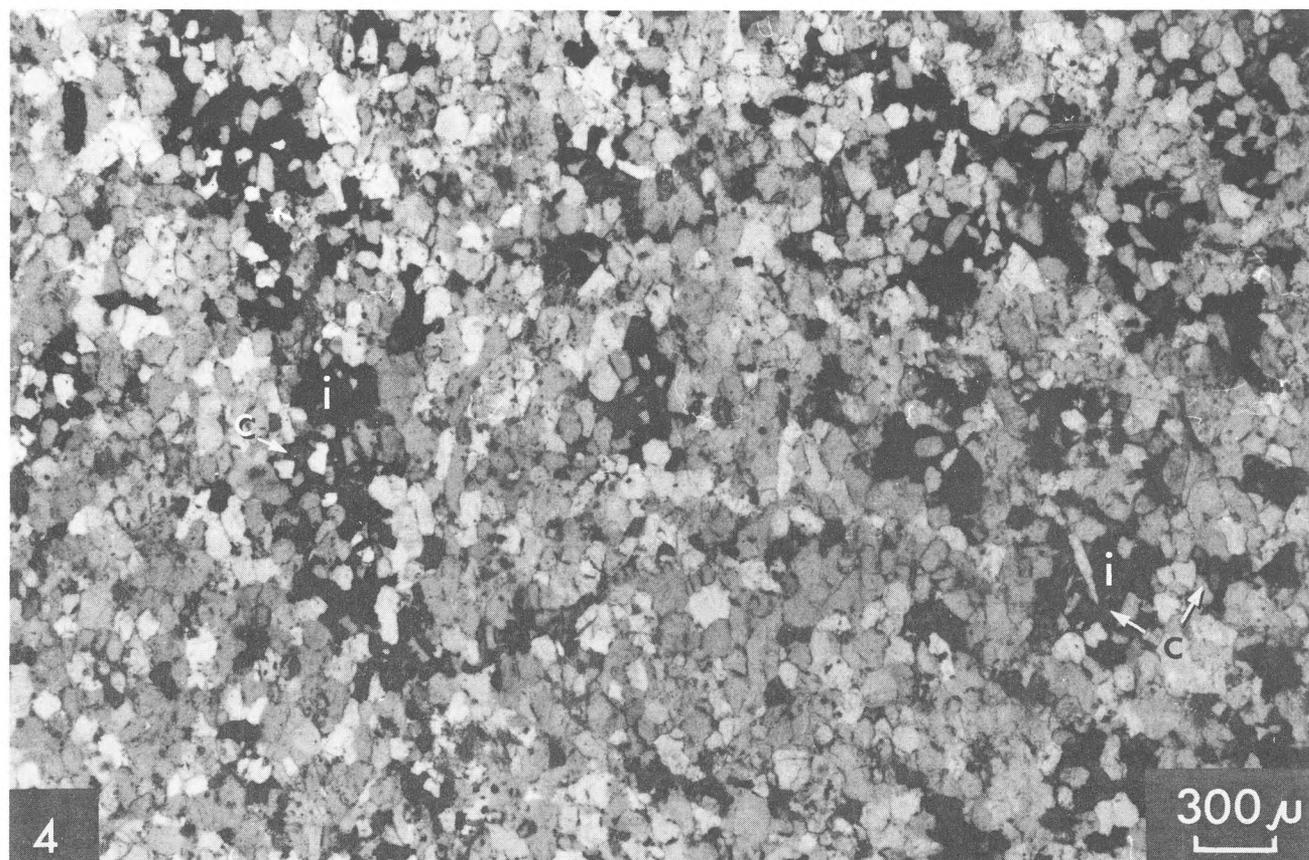

Fig. 4
Calcite cement (c) and authigenetic iron minerals (i) are found in patches in a quartz-cemented sandstone. They are late diagenetic new-formings, which preferentially replace matrix material.

the basal part of the Cabos Formation in sediments that are rich in newly-formed biotite. The iron minerals are mainly hematite, limonite and goethite. Iron minerals generally occur finely dispersed over the matrix or as irregularly-shaped aggregates.

LUNA AREA

The combined presence of authigenetic illite with minor amounts of detrital kaolinite and authigenetic chlorite and muscovite is generally believed to give a measure of the depth of burial of the sediments. Comparisons with data from other authors (BURST, 1959; WEAVER, 1959; KOSSOVSKAYA & SHUTOV, 1963; SCHERP, 1963; DUNOYER DE SEGONZAC, 1965) suggest a probable depth of burial of about 2000-3000 m. This is in agreement with evidence from the stratigraphic sequence and structural reconstructions. The sedimentary succession that probably once covered the studied formations amounts to 2500-3500 m, in the area studied (VAN DEN BOSCH, 1969).

Biotite appears to have been unstable in the sediments of the Luna area. Detrital biotite flakes occur in all stages of decomposition. The alteration products are muscovite, hematite, anatase, clay-minerals (illite) and in a few cases chlorite. We already mentioned the frequent occurence of mica flakes with muscovite and discoloured biotite lamellae (Fig. 7 a, b). Hematite and anatase, formed from components that could not be accommodated in the newly-formed muscovite (DAPPLES, 1967; MEISL, 1970), occur in the form of inclusions in the authigenetic muscovite, or scattered and in aggregates in the sediment surrounding the altered biotite flakes. Hematization of biotite flakes, a feature described by LEWIS (1971), was occasionally observed in the sediments studied.

The decomposition of biotite flakes began early in the diagenetic history of the sediments (syndiagenetic phase of FAIRBRIDGE, 1967), because dust-rings around quartz grains with secondary overgrowths are often formed partly by small crystals of authigenetic anatase and hematite. This means that the authigenesis of anatase and hematite and therefore the decomposition of biotite predated the enlargement of quartz grains. Another indication for an early beginning of biotite decomposition is the fact that authigenesis of anatase occurs preferentially at low temperatures and low pressures (MILNER, 1962; SCHUILING & VINK, 1967) and is favoured by high porosi-
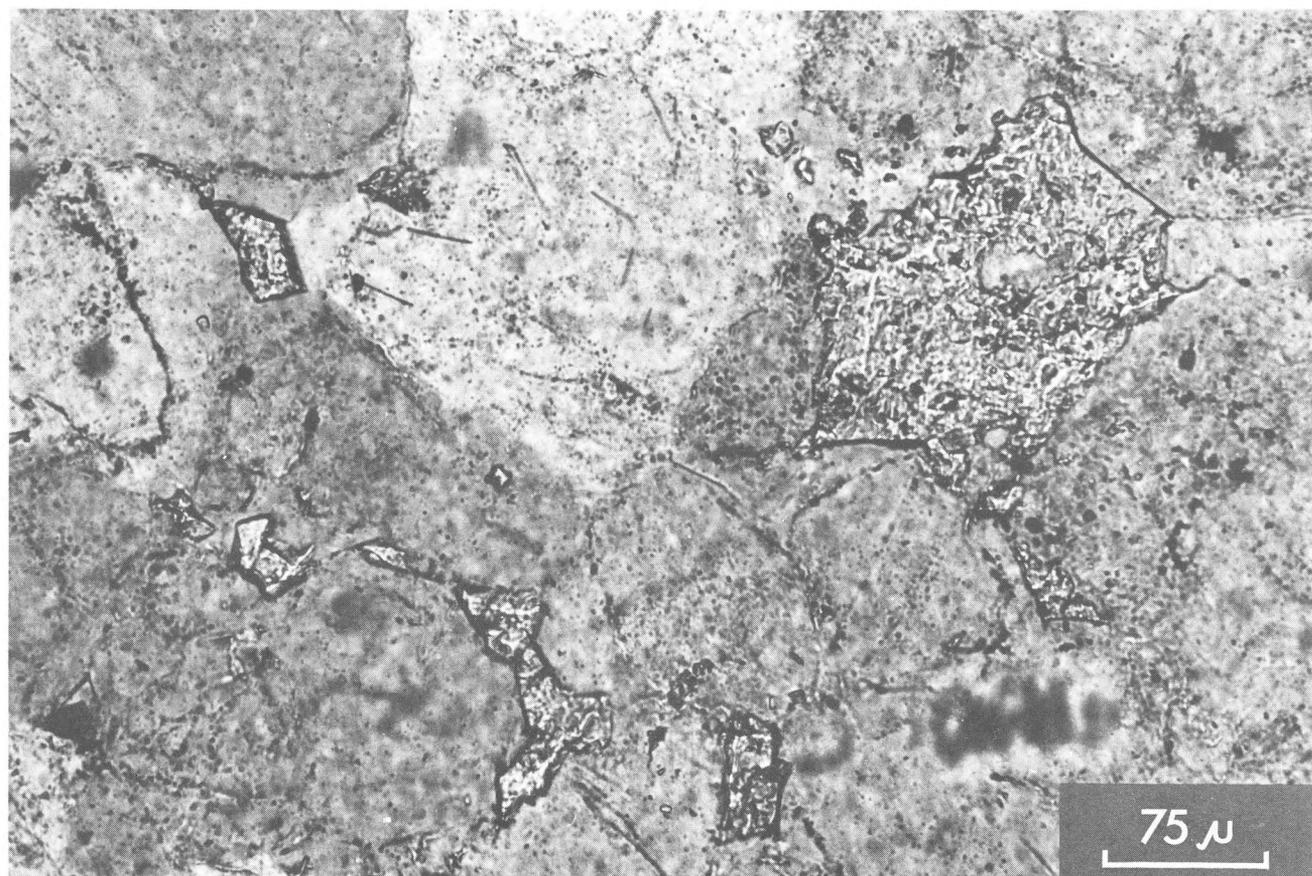

Fig. 5
Barite cement replacing secondary quartz overgrowths in an orthoquartzite.

ties of the sediment (MILNER, 1962).

The occasionally observed alteration of biotite into muscovite, chlorite and hematite indicates that biotite decomposition still occurred at a late stage in the diagenetic evolution.

The sediments usually have a quartz cement, that was formed by secondary enlargement of the detrital quartz grains. Pressure solution was observed in several slides and clearly postdates the enlargement of the quartz grains. According to many authors there is a general relationship between the degree of pressure solution and the depth of burial. Sutured contacts begin to appear at a depth of burial of about 1500-2000 m (FÜCHTBAUER, 1967). The fact that pressure solution is not well-developed in the studied sediments is in agreement with the overburden of 2500-3500 m, derived from geological evidence.

Carbonate cement (calcite) was observed in a few slides (Fig. 4). The calcite is clearly younger than the quartz overgrowth, replacing secondary quartz. Generally calcite was deposited simultaneously with iron minerals, which might indicate that authigenesis of calcite occurred at a moment at which iron was still being supplied by decomposing biotite or clay minerals.

A remarkable fact is the high calcite content of the sedi-

ments of the Oville Formation of the Esla area further to the east (OELE, 1964), in contrast to the sediments studied. Close to the Esla area the formations studied were eroded and subaerially exposed in Famennian (Upper Devonian) times, whereas they were deeply buried in the Luna area, where the Famennian unconformity is absent (VAN DEN BOSCH, 1969). Replacement of quartz by calcite is favoured by the presence of formation water with a high pH value (DAPPLES, 1962; SHARMA, 1965). The subaerial exposure of the formations may have caused migration of formation waters. The migration must have been difficult and slow, because the waters had to migrate through already cemented sediments. In the Esla area, near the subaerial exposures, waters with high pH values could apparently still be introduced (from the surface or from the underlying limestone formation?), whereas in the more distant Luna area formation water migration was probably of less importance and little calcite was formed.

Glauconite grains with a microcrystalline texture (Fig. 3) occur in sediments, which were interpreted as shallow-marine deposits on the basis of sedimentological evidence. This indicates that glauconite was formed in the area during deposition. The glauconite grains are rounded, which probably means that transportation occurred after the glauconite was formed.

Glauconite grains are almost completely altered into illite,

Fig. 6
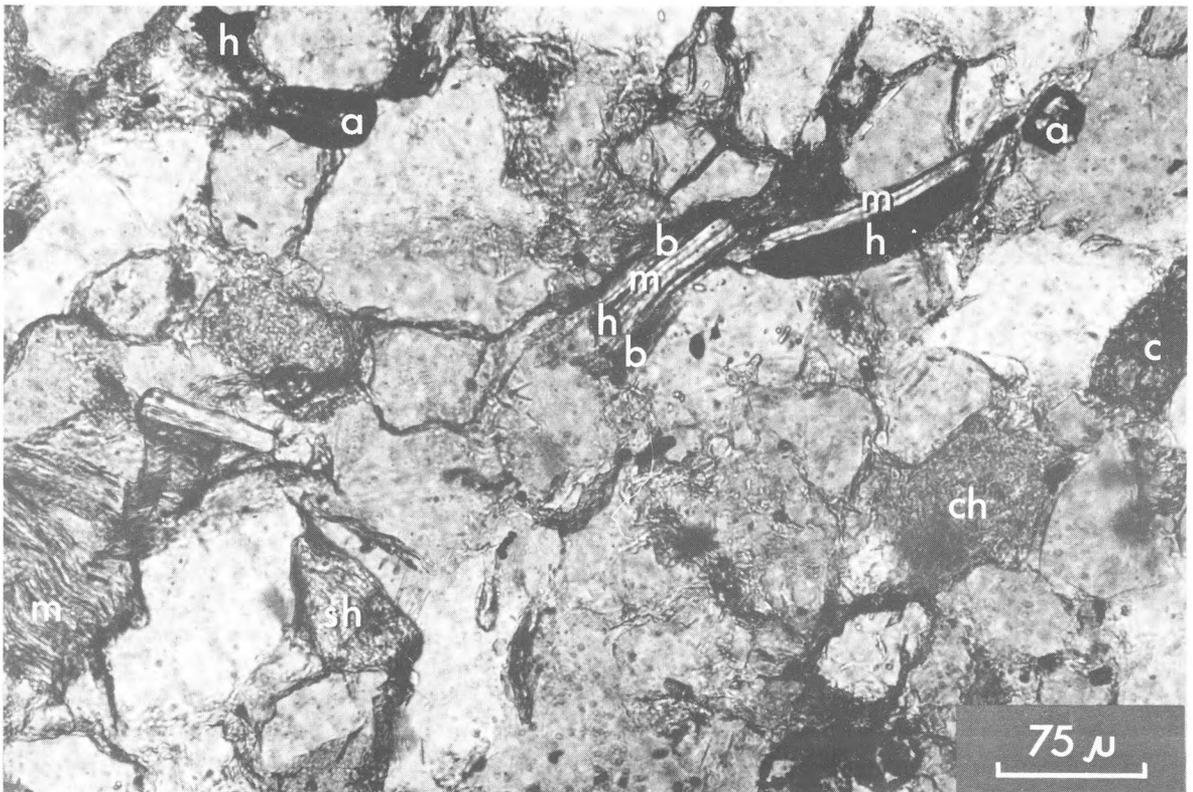Authigenic muscovite formed from the clay minerals in the matrix.

Fig. 7 (a and b)
Partly and entirely decomposed biotite flakes (b) combined with authigenic muscovite (m), hematite (h) and anatase (a). ch = chlorite; c = chert fragment.
Fig. 7 a (above): crossed nicols; Fig. 7 b (below): plane light.

muscovite, quartz and occasionally chlorite; only the characteristic greenish color was preserved in most cases. The decomposition of glauconite probably postdates the secondary enlargement of quartz grains, because glauconite is altered to a lesser degree in well-cemented sandstones than in argillaceous sediments. DAPPLES (1967) observed that glauconite decomposition was severe on the crest of folds, whereas glauconites on the flanks of the same folds were hardly altered at all. Glauconite appears to be unstable under conditions of well-developed folding. As the Luna area was folded during the Hercynian orogeny, the glauconites probably became unstable at that time.

Large parts of anatase aggregates are often opaque and have been clearly altered. The opaque mineral is brightly white in incident light and is therefore probably leucoxene. Anatase crystals in dust-rings around the quartz grains with secondary overgrowths never show any sign of alteration, which means that the alteration of anatase into leucoxene postdates the enlargement of quartz grains.

Barite cement occurs locally in small quantities in the quartz sandstones of the Barrios Formation. Barite cement occurs in patches and replaces secondary quartz overgrowths (Fig. 5). According to MEISL (1970) barite cement is a late diagenetic authigenesis (epi-diagenetic stage of FAIRBRIDGE, 1967). The barite cement must have been formed as the rocks were uplifted towards the surface, because deep waters do not contain the SO_4^{2-} required for the formation of barite. The occurrence of barite in patches, replacing secondary quartz probably rules out the hydrothermal introduction of barite.

At present weathering processes cause renewed migration and redeposition of iron minerals along fractures in the rocks.

The conclusions drawn above are systematically compiled in table II.

SIL AREA

The petrographic development of the sediments of this area can be subdivided into a diagenetic and a metamorphic story. The metamorphic recrystallizations have obscured much of the earlier diagenetic changes, but there are many indications that the diagenetic evolution in the Sil area has been similar to that of the Luna area.

Remnants of the diagenetic evolution

Part of the matrix material is still present in the form of well-crystallized illite (Fig. 6), which means that the original clay minerals were recrystallized to illite. Illitic matrix occurs relatively more frequently in well-cemented sandstones than in argilleceous sediments and is more frequent in the upper part of the succession than in the lower part.

The scarcity of detrital biotites in the sediments studied, combined with the presence of mica flakes composed of alternating muscovite and chlorite lamellae showing hematite inclusions, and the presence of authigenetic anatase in the surrounding matrix probably indicate that biotite was unstable during at least part of the diagenetic evolution.

Pressure solution phenomena are abundant in the sediments of the Sil area. Pressure solution is generally weak to moderate in the top of the succession and increases in intensity towards the bottom. In most slides it is possible to see that pressure solution postdated the enlargement of quartz grains.

Calcite cement occurs in patches in part of the sediment and appears to be a late diagenetic authigenesis, replacing quartz overgrowths. Possibly the calcite cement is also related to the Famennian epeirogenic movements.

Most of the anatase crystals show alterations to leucoxene. The alterations probably occurred at the same diagenetic stage as they did in the Luna area.

Authigenesis of iron minerals occurred during decomposition of biotite and alteration of the clay minerals of the matrix.

Metamorphism

In the upper part of the succession studied the original matrix-clay minerals have been recrystallized to an illite-muscovite-chlorite association (Fig. 8 a, b). In the lower part a muscovite-chlorite-biotite association is generally found (Fig. 9 a, b).

The thickness of the Cabos Formation amounts to approximately 8500 m in the Sil area. The thickness of the overlying Palaeozoic deposits amounts to 3000 m (MATTE, 1968), but Silurian and Devonian deposits do not occur in the Sil area, though it is generally accepted that these have been present and have been eroded. Based on stratigraphic evidence from surrounding areas the total thickness of the overlying Palaeozoic deposits is estimated at 5000-6000 m. This means that the overburden of the sediments in the Sil area ranged from approximately 5000 to 14000 m.

The metamorphic mineral associations mentioned above are not in agreement with the depth of burial. The critical transition temperature between diagenesis and metamorphism is 300 °C and is almost independent of pressure (WINKLER, 1965). In burial metamorphism at normal geothermic gradients (30 °C/km) this temperature is reached at depths of approximately 10000 m. The observed mineral associations are thought to be the result of burial metamorphism due to the large overburden, combined with regional metamorphism caused by the Hercynian orogeny during which period additional thermal energy was produced, so that the critical temperature occurred at much shallower depths.

In the muscovite-chlorite-biotite association chlorite appears to have been unstable. Mica flakes consisting of lamellae of chlorite and biotite are frequent and clearly indicate that chlorite has recrystallized to biotite. All stages of recrystallization occur. According to KOSSOVSKAYA & SHUTOV (1970) the mass appearance of biotite occurs during regional metamorphism due to the interaction of muscovite, chlorite and titanium minerals.
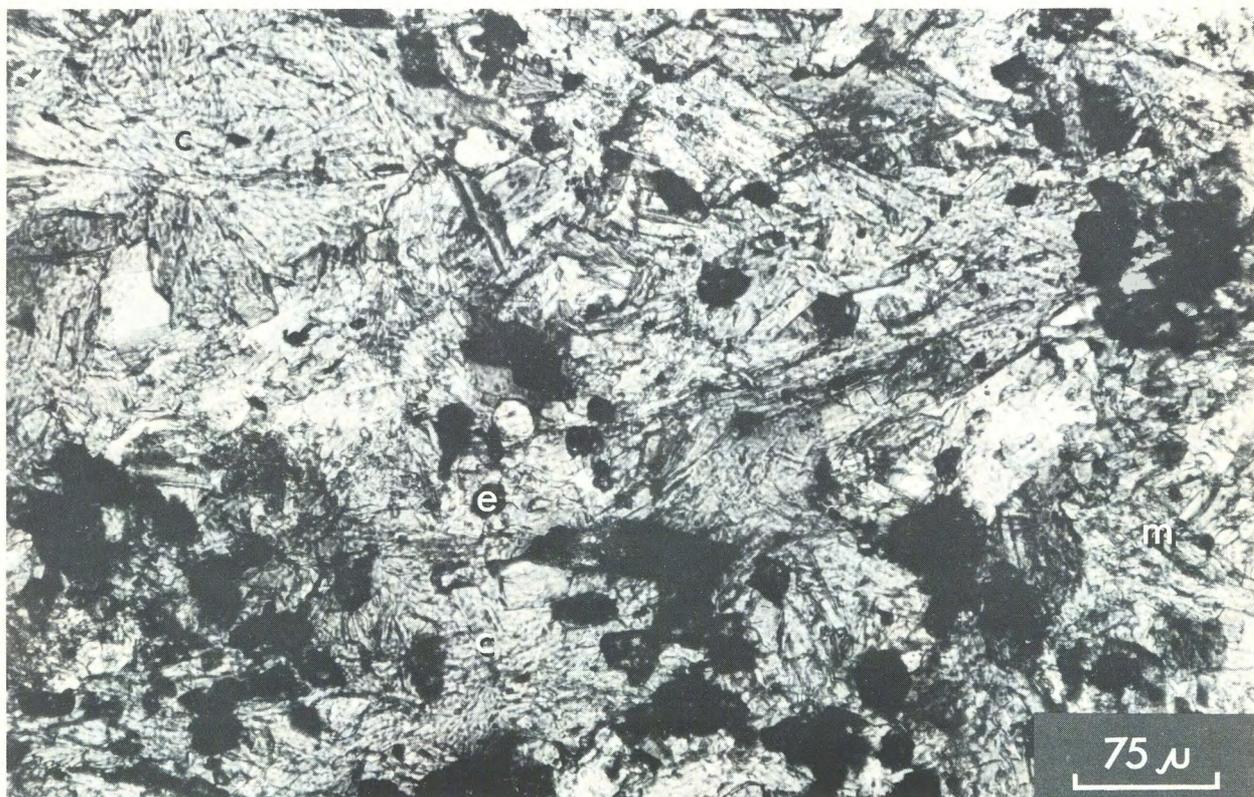
Fig. 8 (a and b)
Authigenetic chlorite formed by decomposition of the pre-existing matrix material (mainly illitic clay minerals).
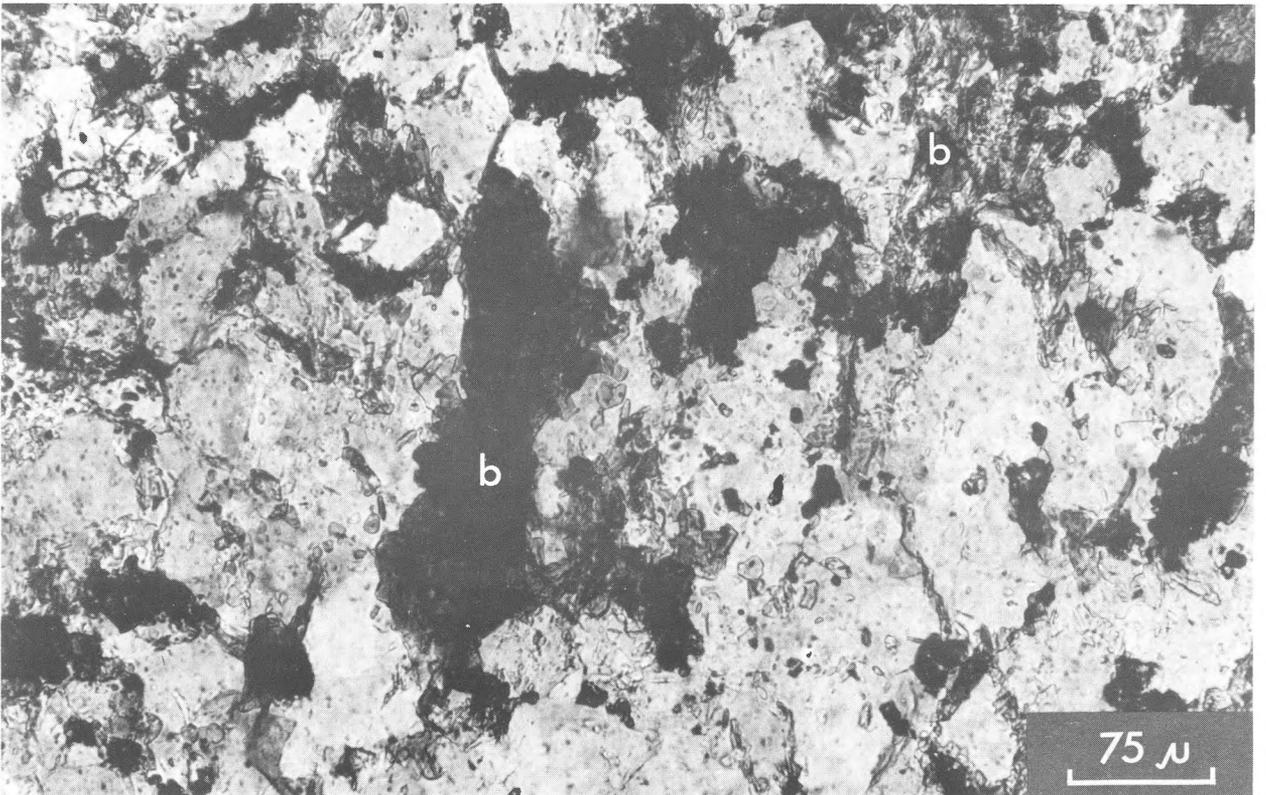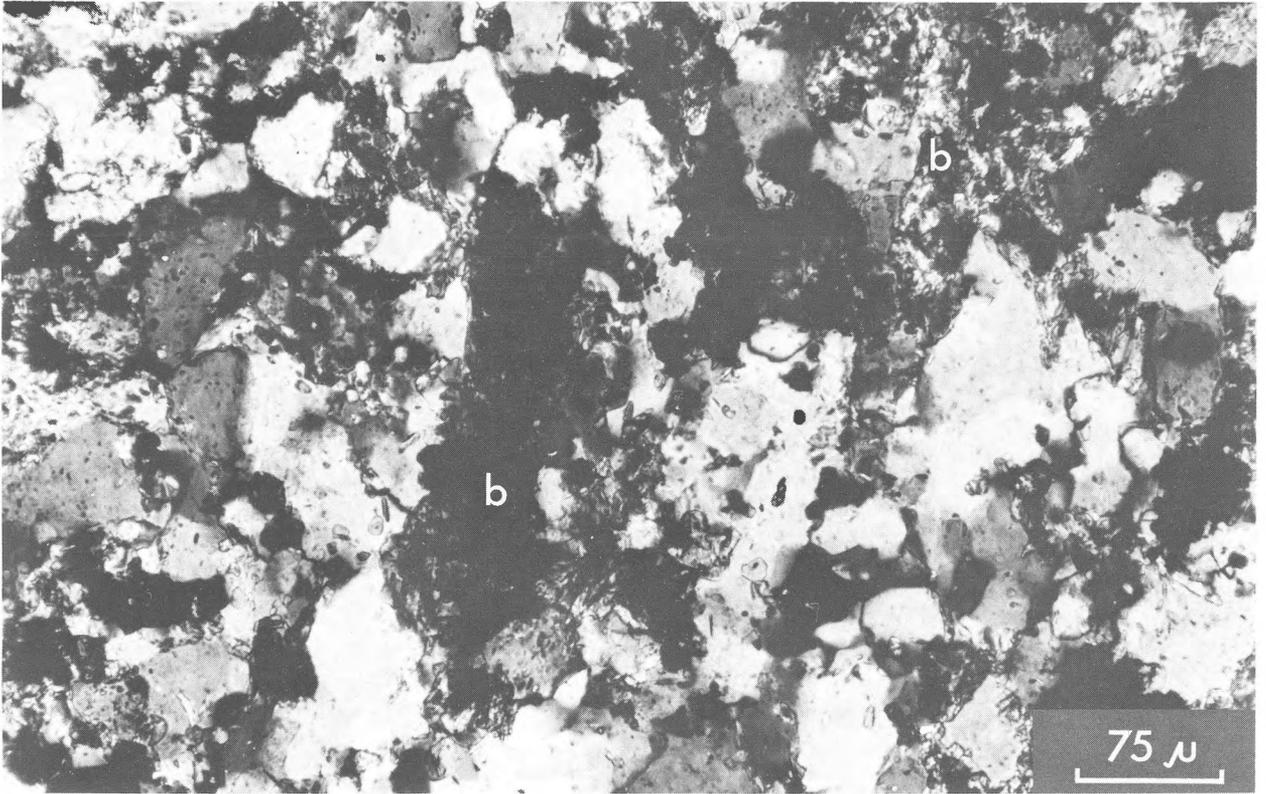Fig. 8 a (above): crossed nicols; Fig. 8 b (below): plane light.
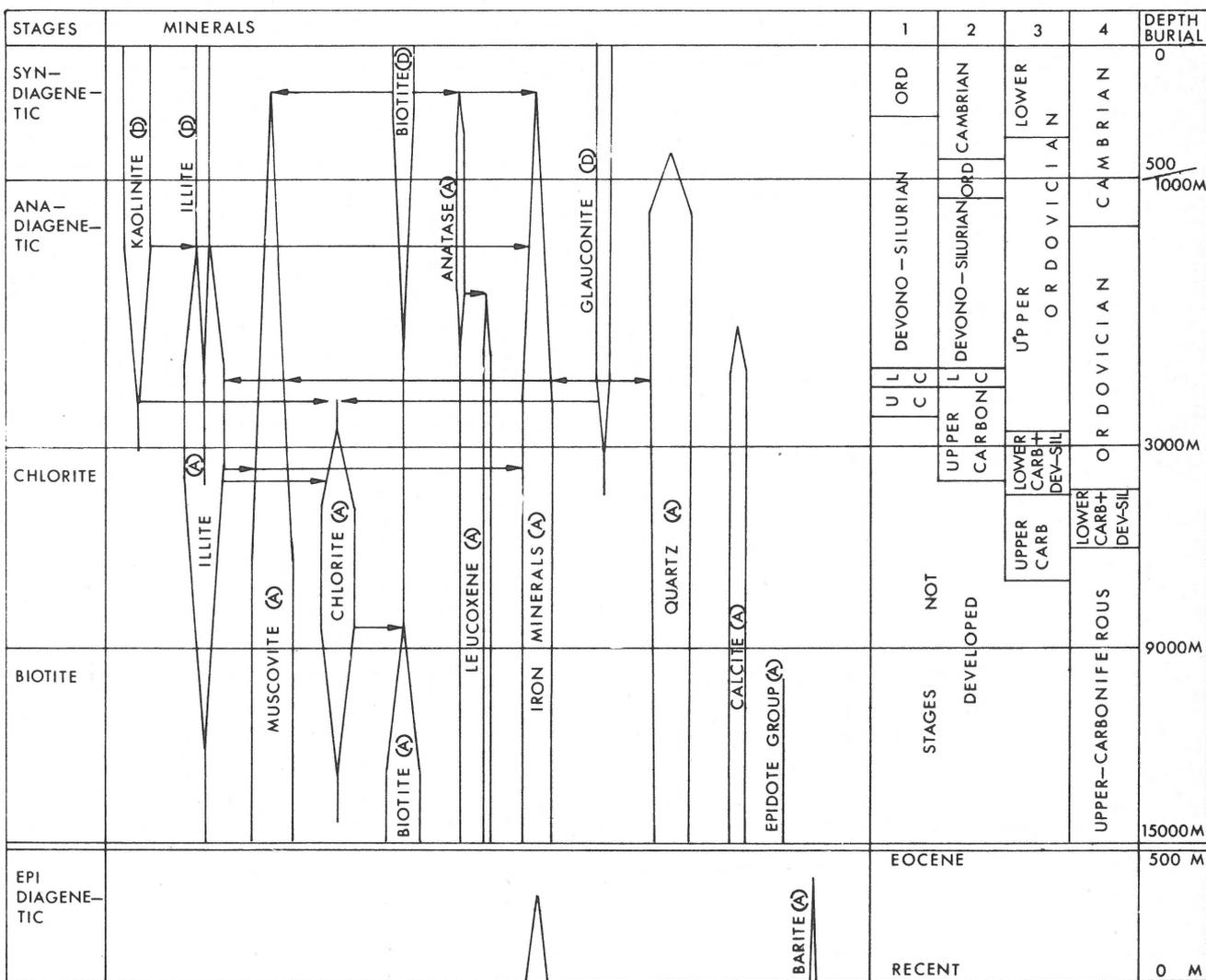
74

Fig. 9 (a and b)
Biotite formed at the expense of pre-existing authigenic chlorite and other matrix material.
Fig. 9 a (above): crossed nicols; Fig. 9 b (below): plane light.

Table II
Attempt to summarize the conclusions of this paper. The horizontal arrows indicate the relations between the decomposition of one mineral and the authigenesis of others. The diagenetic/metamorphic stages during which the various alterations occurred are indicated on the left. The depth of burial related to the diagenetic/metamorphic stages is shown on the right. The approximate period in geological time during which certain diagenetic or metamorphic changes occurred in the sediments, or at which the sediments were in the various diagenetic/metamorphic stages, is given for 4 cases:
(1) the Barrios Formation, the upper part of the succession in the Luna area;
(2) the Oville Formation, the lower part in the same area;
(3) the upper part of the Cabos Formation in the Sil area;
(4) the lower part of the Cabos Formation in the same area.
The top line of the table represents the sedimentary surface.
D = detrital; A = authigenetic.

Newly formed epidote-group minerals occur in sediments in which recrystallization to biotite is abundant. These epidote-group minerals were formed during the recrystallization of chlorite to biotite from components that could not be accommodated in the biotites.

The conclusions drawn above are compiled systematically in table II.

CONCLUSIONS

The diagenetic evolution of the sediments of both areas was similar. This indicates that the higher rate of deposition under the geosynclinal conditions as compared to the stable shelf conditions hardly influenced the diagenetic evolution of the sediments studied.

During the Hercynian orogeny additional thermal energy was added to the system in the geosynclinal region, which disturbed the normally expected depth of burial/mineral association relationship. On the stable shelf this relationship still exists, indicating that the addition of thermal energy by the orogenetic event was of minor importance.

REFERENCES

Bard, J. P., R. Capdevila & P. Matte 1971 La structure de la chaîne Hercynienne de la Meseta Iberique: comparison avec les segments voisins – In: Histoire structurale du Golfe de Gascogne – Publ. Inst. français Pétrol. (colln. Colloques et Séminaires) 22, Technip (Paris): 994 pp.

Burst, J. G. 1959 Postdiagenetic clay-mineral environmental relationships in the Gulf Coast Eocene – In: Clays and clay minerals – Proc. 6th Nat. Conf. (Berkeley, 1957), Pergamon Press (London/New York/Los Angeles): 327-341.

Dapples, E. C. 1962 Stages of diagenesis in the development of sandstones – Amer. Ass. Petr. Geol. Bull. 73: 913-933.

—— 1967 Diagenesis of sandstones – In: G. Larsen & G. V. Chilingar (eds.): Diagenesis in sediments (Dev. Sediment. 8) – Elsevier Publ. Cy. (Amsterdam/New York/London): 91-126.

Dunoyer de Segonzac, G. 1964 Les argiles du Crétacé supérieur dans le bassin de Dovala (Caméroun): problèmes de diagenèse – Bull. Serv. Carte Géol. Alsace-Lorraine 17: 287-310.

Fairbridge, R. W. 1967 Phases of diagenesis and authigenesis – In: G. Larsen & G. V. Chilingar (eds.): Diagenesis in sediments (Dev. Sediment. 8) – Elsevier Publ. Cy. (Amsterdam/New York/London): 19-90.

Gietelink, G. 1973 Sedimentology of a linear prograding coastline followed by three high-destructive delta-complexes (Cambro-Ordovician, Cantabrian Mountains, NW Spain) – Leidse geol. Meded. 49: 125-144.

—— 1974 Sedimentology of the 8,500 m thick shallow marine Cabos Formation in the Sil area (Cambro-Ordovician of NW Spain) – in prep.

Kossovskaya, A. G. & V. D. Shutov 1963 The correlation of zones of regional epigenesis and metagenesis in terrigeneous and volcanic rocks – Translation Amer. Geol. Inst. 1963: 732-736.

—— 1970 Main aspects of the epigenesis problem – Sedimentology 15: 11-40.

Lewis, D. W. 1971 Qualitative petrographic interpretation of Potsdam Sandstone (Cambrian, Southwestern Quebec – Can. J. Earth Sci. 8: 353-382.

Matte, P. 1968 La structure de la virgation hercynienne de Galice (Espagne) – Thesis, Extr. Trav. Lab. Géol. Fac. Sci. Grenoble 44: 157-280.

Meisl, S. 1970 Petrologische Studien im Grenzbereich Diagenese-Metamorphose – Abh. Hessischen Landesamt Bodenf. 57: 93 pp.

Milner, H. B. 1962 Sedimentary petrography II: principles and applications – G. Allen and Unwin (London) 4th ed.: 715 pp.

Oele, E. 1964 Sedimentological aspects of four Lower Paleozoic formations in the northern part of the province of Léon (Spain) – Leidse geol. Meded. 30: 1-99.

Parga, J. R. 1970 Evolucion del Macizo Hesperico en los tiempos ante-Mesozoicos y sus relaciones con otras areas europeas – Bol. Inst. Geol. Min. España 81: 115-143.

Scherp, A. 1963 Die Petrographie der paläozoischen Sandsteine in der Bohrung Münsterland 1 und ihre Diagenese in Abhängigkeit der Tiefe – Fortschr. Geol. Rheinland-Westfalen 11: 251-282.

Schuiling, R. D. & D. W. Vink 1967 Stability relations of some titanium minerals (sphene, perowskite, rutile and anatase) – Geochim. Cosmochim. Acta 31: 2399-2411.

Sharma, G. D. 1965 Formation of silica cement and its replacement by carbonates – J. Sed. Petr. 35: 733-745.

Van den Bosch, W. J. 1969 Geology of the Luna-Sil region, Cantabrian Mountains (NW Spain) – Leidse geol. Meded. 44: 137-225.

Weaver, C. E. 1959 The clay petrology of sediments – In: Clays and clay minerals – Proc. 6th Nat. Conf. (Berkeley, 1957), Pergamon Press (London/New York/Los Angeles): 154-187.

Winkler, H. G. F. 1965 Die Genese der metamorphen Gesteine – Springer (Berlin/Heidelberg/New York): 218 pp.