

7. The clay minerals of the Yühuat'ai vermiculated clay at Kiukiang in the Lower Yangtze Valley, China

By

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Introduction

The sediments to be described were collected by E. NORIN in 1934 at the eastern outskirts of the town of Kiukiang during an excursion together with P. TEILHARD DE CHARDIN, G. B. BARBOUR, and J. S. LEE in the basin of the Yangtze River. The sediments are beautifully exposed in the steep front of the broad terrace upon which the town stands and which here flanks the Yangtze, the edge rising some 30 m above the river. This terrace which is known as the Yühuat'ai terrace, has been studied and described earlier by G. B. BARBOUR (1935) and by TEILHARD DE CHARDIN and C. C. YOUNG (1935).

"The Yuhuatai platform is unquestionably the best preserved and the most characteristic physiographical element observed in the lower Yangtze basin. All along the valley, between Nanking and Hankow, this 30-meter high built terrace is clearly observable, always keeping the same essential composition as in the type locality of Nanking: a very distinct basal horizon of conglomerate and sand, capped by a variable thickness of vivid red, generally vermiculated clay." (TEILHARD *et al.*, 1935 b, p. 167). A section at Anking was more closely investigated by BARBOUR (*op.cit.*, p. 69) who distinguished the following sequence of strata in descending order:

6. Overburden of soil, rewashed red loam, derived gravel, etc. 2-8 feet.
5. Brown manganiferous clay hardpan 2-3 feet.
4. Tough cellular laterite grading down into zone 3 2-6 feet.
3. Vermiculated red-clay, extending down into zone 2 to different depths depending on fineness of material.
2. Gravel with intercalated cross-bedded ferruginous sand 20 feet.
Unconformity.
1. Pukow sandstone of Lower Tertiary age.

BARBOUR states that "the structural relationship of the zones as well as general physico-chemical principles show that two climatic stages were invol-

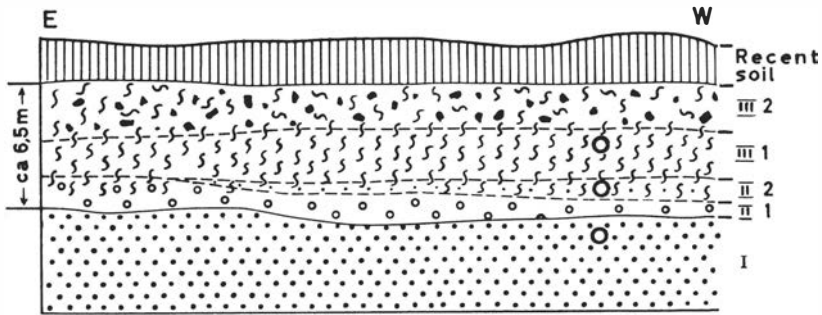


Fig. 1. Section of the Yühuat'ai terrace east of Kiukiang.

ved, the first producing the general penetrative alteration akin to lateritization, the recent one involving less humid conditions leading to concentration of the iron under capillary action". The sequences of humid red and vermiculated beds are, according to the authors cited, probably of Late Pliocene age, the following aridification of the climate eventually marking a climatic change at the beginning of the Pleistocene. "This lateritic period seems to have started approximately at the same time all over Central and Southern China, that is at the end of the great Late Pliocene erosional phase. But we must observe that its duration has possibly been different according to the latitude, due to a gradual shifting (or restriction) of the lateritization from North to South." (TEILHARD *et al.*, 1935 b, p. 188).

Concerning the section of the Yühuat'ai sediments at Kiukiang, E. NORIN has given the following information: "About 1 km west of the Lao-huang-tang pagoda, east of Kiukiang, the sequence of strata is as follows (Fig. 1) from the top downwards

Hor.III.2	Hard crust of reddish brown cellular, speckled rock	2.5 m.
Hor.III.1	Speckled vermicular clay	2.5 m.
Hor.II.2	Yellowish gray, vermicular sandy clay.	} 2 m.
Hor.II.1	Conglomerate of light gray chert and quartzite pebbles with yellowish gray matrix.	
	Disconformity	
Hor.I	Grayish white medium grained quartz sandstone with light coloured matrix	exposed about 1 m.

The horizons III.2 and III.1 are rather similar, but whereas III.1 is soft enough to be cut with a knife, III.2 is hard like concrete and of a darker chocolate brown colour. In the top part, Hor.III.1 is of a general red colour; deeper it becomes speckled with irregular spots and diffusely delimited streaks of brownish yellow material which gradually becomes predominant, the red material appearing as spots and veinlets with diffuse boundaries in the brownish yellow clayey sediment. The veinlets penetrate the horizon in all directions but

often extend more or less transversely to the ancient land surface. No trace of stratification is discernable. In the western part of the section, the red material almost disappears near the base of the bed, which here consists of grayish yellow sandy clay, Hor.II.2. Lower, it becomes gravelly, grading into the subjacent conglomerate, Hor.II.1, which contains a matrix of similar sandy clay. This conglomerate constitutes a well defined bed, resting on the subjacent grayish white quartz sandstone, Hor.I, with a perfectly sharp, undulating contact, due to erosion; thus, the thickness of the conglomerate varies between 1 and 2 m. In the eastern part of the section, on the other hand, the lower boundary of Hor.III.1 gradually descends to direct contact with Hor.II.1 and the light-coloured matrix of the conglomerate acquires the characteristic colour of the vermiculated clay, proving that the alteration process by which Hor.III.1 has obtained its colour and vermiculated structure is a phenomenon posterior to the deposition of the conglomerate.

The "quartz sandstone", Hor.I, which largely consists of grains of quartz and chert, like the conglomerate overlying it, is a residual sediment derived partly from Paleozoic quartzitic sandstones and chert-bearing formations, partly inherited from destroyed older red beds. Red material appears occasionally in the sandstone as solitary streaks and veinlets, and also as tubes coated with oxides of iron, possibly formed around roots of plants."

The location of the samples subjected to investigation appears from Fig. 1. The samples are marked KK III, KK II, and KK I, the Roman numerals corresponding to the respective horizon. In the text, the code signatures of the samples always refer to the clay fraction only (less than 2μ). These fractions have been isolated by sedimentation in very dilute ammonia solution and subsequently dried at 105°C .

Qualitative X-ray analysis

In Table I are given the X-ray diffraction data for the three untreated samples and for one sample specially treated to eliminate kaolinite and goethite. The values were obtained with a Philips 114.8 mm powder camera and filtered CuK_{α} -radiation. Diffraction data for some reference minerals are added in the table.

From Table I it is evident that all the samples contain a micaceous mineral, together with quartz and kaolinite, the last mineral culminating in the bottom horizon. Goethite reflexions are very easily identified in the sample from the upper horizon but are traced only with difficulty in those from the lower.

The absence of montmorillonite and other expanding clay minerals has been established by X-ray photograms of basis-oriented aggregates saturated with glycerol.

As to the nature of the micaceous mineral the diffraction data give no complete information. At any rate, it must be of a dioctahedral type judging by the

Table 1. Lattice spacings, in Å units, and estimated intensities in powder diagrams (b = broad line).

KK III	KK II	KK I	KK II (Kaolinite and goethite removed)	Muscovite	Illite, dioctah. (Gilead)	Quartz	Goethite	Kaolinite (only strong lines)
c. 10 m(b) 7.17 m 4.92 w 4.46 vst 4.15 st	c. 10 m(b) 7.18 m 4.96 vw 4.47 vst 4.27 vw 4.14 w	c. 10 m(b) 7.15 vst 4.97 vw 4.46 vst 4.13 vw	10.06 st 4.95 w 4.48 vst 4.26 m	9.98 st 5.0 st 4.49 st 3.73 w	9.98 st 4.97 w 4.47 st 4.11 vw 3.7 vw	4.21 st	4.96 mw 4.15 vst	7.15 vst + 3.566 vst +
3.53 m(b) 3.34 st	3.52 w(b) 3.34 vst 3.22 vw 2.99 vw 2.952 vw	3.34 m 2.99 vw 2.826 vw	3.52 m 3.34 vst 3.25 vw 2.99 vw 2.810 vw	3.5 m 3.33 vst 3.2 m 2.88 m 2.80 m	3.4 vw 3.31 m 3.2 vw 2.98 w 2.84 vw	3.35 vst	3.363 mw	
2.846 w 2.678 mw 2.564 st 2.508 vw	2.687 vw 2.571 st 2.461 vw	2.560 st 2.508 vw	2.574 st 2.467 w	2.57 vst 2.385 m	2.56 st 2.44 w 2.38 m	2.45 mw	2.674 mst 2.565 mw 2.471 mw 2.433 st	2.553 st 2.486 vst
2.433 mst 2.346 vw 2.281 vw 2.237 mw 2.171 w 2.127 w 1.986 w 1.895 vw 1.823 w 1.796 vw 1.716 }mw 1.653 }	2.373 vw 2.242 vw 2.128 vw 1.985 vw 1.815 w 1.704 vw 1.667 vw	2.339 mst 1.985 w 1.690 }mw 1.659 }	2.280 w 2.245 w 2.128 w 1.992 mw 1.891 vw 1.882 m	2.28 }w 2.19 2.134 st 1.995 vst	2.24 m 2.18 w 2.11 w 1.98 m	2.285 mw 2.235 vw 2.129 mw 1.981 w 1.814 st	2.237 m 2.175 m 1.997 vw 1.908 w	2.331 vst + 2.284 vst
1.549 mw 1.494 st 1.452 w 1.373 m	1.541 w 1.497 m 1.667 vw	1.539 vw 1.492 st	1.698 vw 1.665 vw 1.542 m 1.501 m 1.383 w 1.375 m	1.651 st(b) 1.50 st	1.65 w 1.64 m 1.50 st	1.667 mw 1.539 mst 1.447 vw 1.412 vw 1.376 st	1.788 w 1.709 mst 1.679 mw 1.554 m 1.500 mw 1.445 mw	1.659 st

spacing of the (060) reflexion at 1.50 Å. Eventually it could be mixed with smaller amounts of a trioctahedral mica, a matter impossible to settle because the characteristic line at ca 1.54 Å is masked by quartz. The intensity of the 10 Å line is difficult to estimate in the ordinary powder diagrams of the untreated samples, as the line is much broadened with a diffuse inner edge. From photograms taken with basis-oriented aggregates it is, however, possible to estimate the relative intensities of the 10 Å and 5 Å lines, in this case indicating a mica mineral rather rich in alumina ($I_{002}:I_{001} \sim 1:3$). On these photograms and on photograms of samples heated at 600°C the first-order basal reflexion of the micas appears quite sharp, excluding the existence of mixed-layer minerals.

The presence of kaolinite in all the samples is revealed by the 7.15 Å line, which persists after treatment with boiling 4 M HCl but disappears after heating at 600°C. When kaolinite is present in smaller amounts, the second-order basal reflexion can only be traced as a border of a broad line with an intensity peak at 3.52 Å, this peak being due to mica and anatase. The second-order basal reflexion appears, however, with normal development on photograms of samples rich in kaolinite, and always on photograms of oriented aggregates, even at low concentration.

Quantitative chemical analysis

Determination of goethite

There are many methods recommended for determining the amount of iron oxide hydrate present in a soil. They are almost exclusively based upon a "cooking-recipe" technique which can turn out well only under certain conditions. The principal difficulty lies in extracting the "free" iron without affecting the iron bound to the clay silicates. As the solubility of each of the different iron compounds varies within wide limits depending on their age, crystallinity, and crystal structure, no kind of patent prescription can cover every possible case.

B. OSTHAUS (1954, 1956) has solved some of the problems by studying the kinetics of the dissolution process, a method originally introduced in the clay mineralogy by G. W. BRINDLEY and R. F. YOUELL (1951). By digestion with hydrochloric acid for various periods of time and determining the amount of a certain ion extracted, a curve can be drawn showing the rate of solution of this ion by plotting the undissolved fraction on semilogarithmic paper against time. For a given type of ion occupying equivalent positions in the crystal structure, a straight line will be obtained, the pre-requisite condition being that the dissolution process is a first-order reaction. This cannot strictly be the case as the total areas of the particles diminish during the dissolution and consequently change the rate of solution. The above condition seems, however, to have only an approximate validity, the dissolution of an ion decreasing exponentially with time.

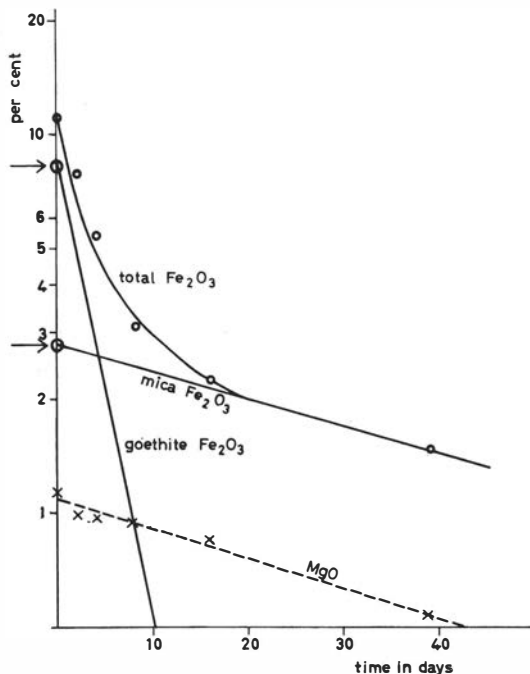


Fig. 2. Acid-dissolution curves of Sample KK III.

If there is a difference in the rate of solution between the same kind of ion in two different compounds or in two different positions in the same crystal structure, this is revealed by a change in the slope of the curve. If, after sufficient time, the contribution of the more easily soluble ion is negligible, the resulting curve will be a straight line. This can be extrapolated to zero time and the original percentage of the specific ion established.

In this way OSTHAUS has been able to determine the proportions in which iron is distributed between tetrahedral and octahedral positions in a nontronite sample as well as the amount of the extraneous iron. It is obvious that the more contrasted the different rates of solution are, the more exact results are obtained. In the present investigation this method has been adopted to separate goethite-iron from silica-bound iron.

The procedure used was to treat 200 mg of the sample with 100 ml 4 M hydrochloric acid for various periods of time. The experiments have been performed at constant temperature (20°C), continuous agitation of the suspensions being maintained by a magnetic stirrer.

For each digestion period, the filtrates containing the dissolved ions were analyzed for iron and magnesium. The dissolution curves of octahedral iron and of octahedral magnesium of the hydrous mica ought to be parallel. Knowing the rectilinear slope of the magnesian curve it is, then, possible to dissolve the complex iron curve into two componential straight lines.

Table II a. Chemical analyses of Sample KK III

	Total	Quartz	Kaolinite	Goethite	Anatase ?	Hydrous mica
SiO ₂ in %	43.66	4.38	7.05			32.23
TiO ₂	1.50				1.50	
Al ₂ O ₃	28.94		5.98			22.96
Fe ₂ O ₃	10.73			8.33		2.40
FeO	0.36					0.36
MgO	1.12					1.12
MnO	0.03					
CaO	0.22					
K ₂ O	2.20					2.20
Na ₂ O	0.16					
P ₂ O ₅	0.26					
H ₂ O ⁺	10.66		2.12	0.94		7.60
Sum:	99.84	4.38	15.15	9.27	1.50	68.87

Table II b. Chemical analyses of Sample KK II

	Total	Quartz	Kaolinite	Goethite	Anatase ?	Hydrous mica
SiO ₂ in %	46.95	7.04	8.08			31.83
TiO ₂	1.16				1.16	
Al ₂ O ₃	28.94		6.86			22.08
Fe ₂ O ₃	6.50			4.26		2.24
FeO	0.85					0.85
MgO	1.34					1.34
MnO	0.04					
CaO	0.28					
K ₂ O	2.56					2.56
Na ₂ O	0.13					
P ₂ O ₅	0.23					
H ₂ O ⁺	10.46		2.42	0.48		7.56
Sum:	99.44	7.04	17.36	4.74	1.16	70.46

Iron was determined by potentiometric titration with E.D.T.A. Magnesium was first isolated as a co-precipitate to nickel hydroxide, again brought into solution and titrated with E.D.T.A.

As an illustration of the method, the dissolution curves of the KK III sample are given in Fig. 2. The arrows indicate the extrapolated values of the original content of the two different iron ions (expressed as Fe₂O₃).

Determination of free quartz

For this purpose the phosphoric acid method of H. HIRSCH and W. DAWIHL (1932), as modified by E. NORIN (1958), was used.

Table II c. Chemical analyses of Sample KK I

	Total	Quartz	Kaolinite	Goethite	Anatase ?	Hydrous mica
SiO ₂ in %	47.11	3.05	14.73			29.33
TiO ₂	0.61				0.61	
Al ₂ O ₃	33.63		12.50			21.13
Fe ₂ O ₃	2.73			0.94		1.79
FeO	0.36					0.36
MgO	1.07					1.07
MnO	0.02					
CaO	0.28					
K ₂ O	2.26					2.26
Na ₂ O	0.09					
P ₂ O ₅	0.13					
H ₂ O ⁺	11.68		4.42	0.11		7.15
Sum:	99.97	3.05	31.65	1.05	0.61	63.09

Determination of kaolinite

Kaolinite was determined by comparison between the solubilities in hydrochloric acid of the unheated sample and the sample heated to 600°C, as recommended by K. A. VESTERBERG (1924).

The total chemical analyses of the different samples (executed at the Central analytical laboratory, head: F. Nydahl) and the results of the partial mineral analyses are given in Table II a–c. The amount of hydrous mica is computed as difference.

The chemical nature of the hydrous mica

It is well known that the main chemical difference between an ordinary, well-crystallized mica and a clay mica mineral consists in deficiency of potassium and excess of water in the latter. The alteration into a clay mica mineral may be effected by a simple exchange of hydronium ions for potassium ions without disturbance of the silica skeleton. In many cases, however, another difference is also in evidence, namely a smaller substitution of aluminium for silicon in tetrahedral coordination, tantamount to a higher silica-to-alumina molecular ratio of the clay micas. This relation has been clearly demonstrated by R. GRIM, who used it to characterize the illites.

From diffraction data it is evident that the mica minerals in question are based upon the same structural features. Nevertheless, there is some difficulty in illustrating the compositional differences in a structural formula. Many of the methods in use do not take the water surplus into consideration, which necessarily must give false results if the above exchange reaction of hydronium for potassium in the interlayer positions is valid. As an example we can take the

method of calculating formulas for mica-like minerals devised by S. B. HENDRICKS and C. S. ROSS (1941). They compensate the charge deficit due to the reduced amount of alkali ions by transferring aluminium from tetrahedral to octahedral coordination, resulting in an increase of the octahedral population compared to the ideal muscovite structure. By a trial and error procedure they finally arrive at a formula with balanced charge distributions.

In this paper the formula has been derived in the following way. After transforming the oxide values of the chemical analyses to molal amounts of the ions, the sum of the valences of tetrahedral and octahedral ions has been calculated. It is assumed that this sum, call it n , should correspond to that found in the idealized muscovite formula, that is 21 valences per half the structure unit ($= \text{KAl}_2(\text{AlSi}_3)\text{O}_{10}(\text{OH})_2$). By using the conversion factor $21/n$ the content of the different ions in the formula is obtained. Finally hydronium is added to the extent needed to fill the interlayer positions completely. The same principal line of reasoning has already been utilized by G. BROWN and K. NORRISH (1952).

A comparison between structural formulae derived (a) in this way and (b) by the HENDRICKS-ROSS method, has been given in Table III, showing only the cation distribution in half the structure unit. Five of the most reliable analyses of dioctahedral hydrous micas have been chosen. The deviations are relatively small but significant. As HENDRICKS and ROSS assume that the interlayer positions are not completely filled, they must give a higher charge to the tetrahedral-octahedral unit, effected by lesser substitution of Al for Si and an increase of octahedral ions. The most striking difference, however, can be studied in the bottom column, showing the weight percentage of "adsorbed" water at 105°C, that is water not accounted for in the structural formula. The lower values are in agreement with dehydration investigations as carried out by E. MAEGDEFRAU and U. HOFMANN (1937) and with hydrogen-deuterium exchange experiments by J. MOUM and I. TH. ROSENQVIST (1958). The corresponding values obtained by the HENDRICKS-ROSS method must be considered as abnormally high.

The structural scheme of the hydrous micas from the present profile is shown in Table IV. The figures are derived from the chemical analyses in Table II. The calcium and sodium ions have been omitted; in any case they would only make small contributions. Titanium, too, has been neglected under the assumption that all or most of it occurs as oxide. Analyses for titanium in the residues obtained after the hydrochloric acid treatment at the determination of goethite (see above) have shown that titanium is dissolved at a much slower rate compared to silica-bound, octahedral iron and magnesium. This points to the fact that titanium is not incorporated in the mica, for if it were, it would most likely occupy octahedral positions.

The most noticeable feature of this material is the exceptionally low potassium value. In fact, no mica mineral lacking more than half of the alkali ions has yet been reported in the literature, whereas the present mineral contains only

Table III. Cation distribution in dioctahedral hydrous micas (per half the structure unit) calculated (a) with interlayer hydronium ions and (b) without interlayer hydronium

I. Fithian, Illinois (Grim *et al.*, 1937); II. Gilhead, Illinois (Grim *et al.*, 1937); III. Ballater, Scotland (Mackenzie *et al.*, 1949); IV. Sárosptak, Hungary (Maegdefrau and Hofmann, 1937); V. Ogofau, Wales (Brammall *et al.*, 1937).

Structural positions	I a	I b	II a	II b	III a	III b	IV a	IV b	V a	V b
tetrahedral	Si ⁴⁺ Al ³⁺	3.45 0.55	3.34 0.66	3.39 0.61	3.34 0.66	3.38 0.62	3.27 0.73	3.32 0.68	3.07 0.93	3.10 0.90
octahedral	Al ³⁺ Fe ³⁺ Fe ²⁺ Mg ²⁺	1.40 0.23 0.09 0.28	1.51 0.23 0.10 0.29	1.31 0.26 0.08 0.39	1.39 0.26 0.09 0.40	1.66 0.12 0.03 0.13	1.78 — — 0.19	1.86 — — 0.19	1.90 0.04 0.02 0.05	1.95 0.04 0.02 0.05
interlayer	Ca ²⁺ Na ⁺ K ⁺ H ₃ O ⁺	0.01 0.02 0.51 0.45	0.02 0.01 0.59 0.36	0.03 0.01 0.60 —	0.05 0.02 0.65 0.23	0.05 0.02 0.65 0.03	0.04 0.07 0.56 0.29	0.04 0.07 0.56 —	0.01 0.05 0.68 0.25	0.02 0.05 0.68 —
“adsorbed” water at 105° in weight %	—0.42	2.69	0.29	2.75	0.03	1.65	0.28	2.43	0.12	1.81

Table IV. Cation distribution in hydrous mica from Kiukiang (per half the structure unit)

Structural positions		KK III	KK II	KK I
tetrahedral	Si ⁴⁺	3.08	3.10	3.08
	Al ³⁺	0.92	0.90	0.92
octahedral	Al ³⁺	1.67	1.63	1.70
	Fe ³⁺	0.17	0.16	0.14
	Fe ²⁺	0.03	0.07	0.03
	Mg ²⁺	0.16	0.19	0.17
interlayer	K ⁺	0.27	0.32	0.30
	H ₃ O ⁺	0.73	0.68	0.70

one-third of the potassium ions of mica proper, indicating a very pronounced leaching. The existence of such a mineral has generally been denied for reasons of stability; thus I. TH. ROSENQVIST (1955) assumed that a mica of this composition should be transformed into a mixed-layer mineral. In spite of this remarkable potassium deficit, the close resemblance to muscovite is well seen in the number of octahedral ions and the degree of tetrahedral substitution of aluminium for silicon. If the calculations are made according to HENDRICKS and ROSS, the discrepancies will grow in a conspicuous way (the figure of tetrahedral Al will become only 0.80 and the sum of the octahedral cations greater than 2.20).

As a consequence of this near relationship to muscovite there is some reason to call the mineral a hydromuscovite. This mineral term has been used at least since the beginning of this century but often in a more or less loose sense. The one well-established analysis of a hydromuscovite (of hydrothermal origin) has been presented by A. BRAMMALL, J. G. C. LEECH and F. A. BANNISTER (1937) and is illustrated in Table III. column V. The principal difference from the present hydromuscovite is the lesser degree of depotassification. To make a distinction between the minerals, the latter one might be called a degraded hydromuscovite analogous to a degraded illite, an expression used by R. E. GRIM to describe an illite with abnormally low potassium content.

The result of a determination of the cation-exchange capacity made in a preliminary stage of the investigation may be of interest in this connection. The cation-exchange capacity of the KK II sample, consisting of a mixture of hydrous mica, kaolinite, and goethite, was determined to 68 meq per 100 g. In view of the silicates present in the sample, this value was considered as unwarrantedly high, wherefore the conclusion was drawn that the presence of goethite created some complications. However, there might be another explanation, assuming that the cation-exchange capacity is related to the intensity of replacement of hydronium for potassium; the more interlayer hydronium the greater exchange capacity.

Finally, some aspects of the partly controversial nomenclature of the mica minerals should be added. The concept of illite, as introduced by R. E. GRIM and coworkers (1937) comprises a rather expansive group name, the non-specific character of which GRIM has later emphasized. In his textbook (1953), however, he attributes to the illite minerals the structural characteristics of distinctly smaller substitution of aluminium for silicon in the tetrahedral layer compared to normal mica. This statement would consequently narrow the limits of the illite group, the existence of normal mica derivatives — with unchanged tetrahedral structure — being confirmed in clay material.

G. BROWN (1951) made the proposal that only “those clay micas which show no significant change of the 10 kX reflection when subjected to mild thermal or chemical treatments” should be called illites, independent of the chemical composition. This has become a widely accepted definition, which also incorporates hydromuscovite among the illites.

The unsettled conditions of this question are reflected in the nomenclature used by K. JASMUND in his textbook (1955). He prefers to distinguish between hydromuscovite and illite according to the grain size, illite representing the smaller sizes of the colloidal material.

Future contributions to this subject will possibly show that it is more convenient to divide the (dioctahedral) hydrous micas into two groups according to the replacement relations to normal mica; the one group with interlayer replacements only, the other with changed tetrahedral Al–Si configuration. The former group is represented by “true” hydromuscovite. As far as is known, the minerals belonging to the latter group show no gradations to the normal micas. Therefore, a specific name might be assigned to these minerals and it is quite natural to reserve the term illite for that purpose.

From purely theoretical considerations there could be some speculation about the genesis of the clay micas in question. The amount of energy required to bring about the changes in the silica layers must be of quite a different order than that responsible for simple exchange reactions between the layers. If the hydromuscovites are looked upon as degradation products of muscovite, the formation of the illites (in the present sense) could be presumed to indicate a more complete decomposition of the detrital material, followed by a reorganization into new minerals. The authigenous character of the standard illites from Illinois has been demonstrated by GRIM (1935), who declares that “these minerals developed from the original debris during or soon after deposition”.

Microscopical examination

Microscopical studies have been executed on the isolated coarser grains down to 30 microns, and on thin sections of methylmetacrylate-impregnated samples.

The particles of the coarser fractions consist almost exclusively of quartz

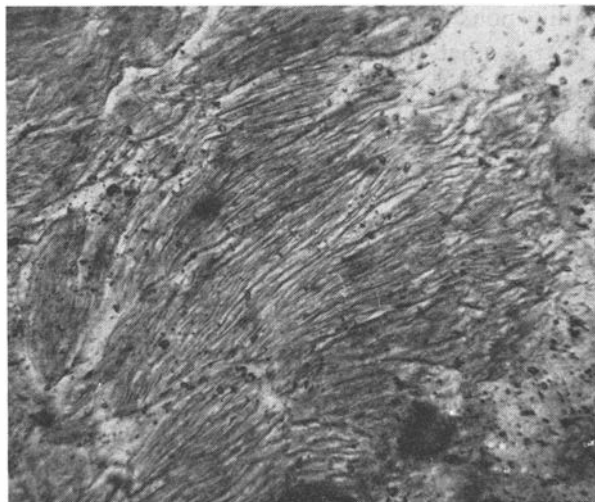


Fig. 3. Matrix of fibrous chalcedony in sandstone from Hor. I. 80 x, 1 nic.

(often coated with goethite) and of microcrystalline chert. Muscovite and biotite, and some other persistent minerals, notably tourmaline and opaque minerals, occur in very subordinate amounts. The samples are totally void of feldspars.

The matrix studied in the thin sections exhibited strong dissimilarities. The constituents of the very fine-grained matrix of the upper horizons (Hor. III and II) are difficult to identify, all the more so, as the presence of goethite gives rise to a strongly red pigmentation. The matrix of the quartz-sandstone (Hor. I) is dominated by fibrous aggregates of chalcedony (Fig. 3), draping itself around the detrital grains and evidently being of authigenous origin.

The weathering character as reflected by the mineral composition

As this investigation has been concentrated on the clay minerals — the small samples available not permitting a complete study of the entire sediments — it is not possible to get a clear picture of the processes involved in the development of the sediments. At any rate, the type of clay minerals present and their distribution in the section might give some information about the weathering reactions.

The iron-rich, hard crust underlain by a mottled, clayey horizon most certainly indicates a lateritic character of the profile. Considering the mineralogical composition of the clay material, however, the lateritic character is not so obvious. As far as is known, the association of kaolinite (or halloysite) and free sesquioxides or their hydrates is generally accepted as a distinctive feature of lateritic material. In the present case, kaolinite is not the dominating clay mineral, being subordinate to hydrous mica. Besides, most of the kaolinite is

probably not authigenous but transported from adjacent, extensive kaolinite occurrences. It is true, that the absence of feldspar suggests a kaolinisation *in situ* but this process cannot have been very advanced since the micas have survived.

The unexpected predominance of hydrous mica may be explained if the original source of rock is regarded. Judging by the literature available, the laterites and lateritic soils previously subjected to investigation, have, with few exceptions, developed on igneous rocks and pyroclastic sediments. In this case, however, the parent rock is a clastic sedimentary rock, rich in quartz and mica, and likely to furnish lateritic weathering products of a different composition.

As already stated, the nature of the potassium-deficient clay mica indicates a very pronounced leaching, not only confined to the upper horizons. As well known, high rainfall and a great permeability allowing the water to soak the sediments are fundamental conditions for the evolution of a lateritic profile.

The eluviation brought about by the leaching process must subsequently have been followed by an illuvial accumulation of goethite, which gradually increases upward to culminate in the upper crust. The old conception that the iron concentration is due to capillary rise during dry seasons is denied by L. D. BAVER (1956), who declares that this mechanism is over-estimated. Referring to the statement by BARBOUR, cited in the introduction to this paper, there is an alternative and radically different interpretation namely that the final accumulation has developed during a long-continued arid climate. It seems, however, to be rather unwarranted to assign the weathering decomposition and the goethite deposition to two different phases, widely separated in time.

Besides the clay minerals, the abundance of chalcedony in the matrix of the "quartz-sandstone" is of importance from the genetical point of view. The association of free silica with laterite has not been given much attention in the recent literature and the silica liberated during the lateritisation is in general assumed to be transported away and withdrawn from the weathering profile. The authigenous chalcedony at the bottom of the present profile may represent this released silica, which has descended during the downward leaching. The position of the ground-water level has probably some determining influence upon the movements and flocculation of colloidal silica.

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