Sedimentological studies on the central Wadi Kalabsha kaolin deposits, Southwest of Aswan, Egypt

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Kaolin deposits initially formed as a lateritic crust on a nearby basement complex, then were transported to the central Wadi Kalabsha and deposited as flood plain and river bar deposits. Repeated fluvial influx and drying out of the kaolin resulted in the following facies: a) Intraformational disorganized conglomerate over the flood plain and/or on the flat bar-tops; b) Intraformational partially organized and organized conglomerates on the slip faces of bars and river banks, and/or as lateral point bars in sinuous streams; and c) Pisolites of kaolinite, kaolinite-hematite, hematite-kaolinite and hematite composition, either in erosion hollows and/or on the slip faces at the downstream end of longitudinal and diagonal bars in braided rivers.

The surface of the fluvial kaolin deposits was subjected to a second phase of in situ lateritization which resulted in a partial or complete dissolution of quartz and the formation of pisolithic laterites containing traces of gibbsite, geothite, hematite and anatase. Isotopic analyses of δ¹⁸O and δD of separated kaolinites showed that the pisolithic kaolin deposits are sedimentary and weathering in origin. The original reworked sedimentary-lateritic pisolites may be deposited during the late Mesozoic while the pisolithic laterites may be formed later as a crust in the Tertiary.

Keywords: Kaolin deposits, Aswan (Egypt), Lateritization, Pisolithic laterite, Depositional environment, Diagenetic history

I. Introduction

This study deals with the sedimentological characteristics of the Lower Cretaceous kaolin deposits at central Wadi Kalabsha, southwest of Aswan, Egypt (Fig. 1). It aims to clarify the depositional environments and diagenetic history of the kaolin deposits in terms of their textures, structures and mineralogical compositions. The study area lies at longitude 32°42'E and latitude 23°32'N (Fig. 1). It is located about 100 km to the southwest of Aswan town and about 20 km from the coastal-line of lake Nasser in the southern Western Desert.

1. Methods of study

The central Wadi Kalabsha kaolin section was studied in the field and representative samples of the different kaolin facies were collected. Slabs and thin sections of the conglomeritic and pisolithic samples were studied. Some pisolitic samples were also studied with a Jeol Scanning Electron Microscope (SEM) JSM-35 with kevex Mx 7,000 system. The mineralogical composition of bulk samples of the different kaolin facies was identified using a Philips APD 1700 automated X-ray powder diffractometer, nickel-filtered Cu-Kα, radiation (λ = 1.54056, 1.54439 Å) generated at 45 kV and 60 mA. The Department of Geology, Indiana University, U.S.A. provided the δ¹⁸O and δD v.s. the Standard Mean Ocean Water (SMOW) data of the separated kaolinites of the fluvial reworked pisolites and pisolithic laterites. The results are expressed in the term of

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$\delta^{18}O$ and $\delta D \%$ relative to SMOW.

2. Geologic setting

The sandstones containing the kaolin deposits occur in the central part of the Nubian Sandstone. The Nubian Sandstone was studied by many authors. Among those authors, Hume (1911), Attia (1955), El-Naggar (1970), Zaghoul et al. (1983), Van Houten et al. (1984), Klitzsch and Wycisk (1987), Klitzsch and Squyres (1990) and Issawi and Osman (1993) are well known. Attia (1955) divided the Nubian Sandstone into three groups, while El-Naggar (1970) considered it as one group and subdivided it into five formations; three of which, the Abu Aggag, Timsah and Umm Barmil formations have been identified around Aswan overlying granitic rocks (pink granites or Aswan granites) which in part deeply weathered.

Klitzsch and Wycisk (1987) and Klitzsch and Squyres (1990) studied and divided the Paleozoic-Mesozoic sequence (strata broadly related to the Nubia Sandstones) in southern Egypt and northern Sudan into three cycles; the Paleozoic, Karroo and Nubian cycles (Fig. 2). However, Issawi and Osman (1993) classified the clastic section southwest of Aswan into seven Paleozoic-Mesozoic units (Fig. 2), of which the Sabaya Formation (Klitzsch and Wycisk, 1987) and the lower member of the Nubia Formation (Taref Sandstone Member) contain kaolin beds.

Said and Mansour (1971) divided the stratigraphic section containing kaolin deposits in the central Wadi Kalabsha into lower and upper sandstone members. The present author observed that these sandstones enclosing the kaolin deposits at central Wadi Kalabsha, are characterized by tabular and trough cross beds, horizontally stratified with lag deposits and scour and fill structures indicating deposition by low sinuosity or braided rivers. Lens shape kaolin deposits interfinger with sandstone and some mottled horizons in the sequence represent paleosols truncating the alluvial plain adjacent the stream channels. Similar facies and depositional models were described by Klitzsch and Wycisk (1987) and Hendriks et al. (1987) for the Sabaya and Abu Aggag formations respectively and by Germann et al. (1987) for the lower part of the Abu Aggag Formation at Wadi Abu Aggag area. Germann et al. (1987) showed that the Wadi Kalabsha kaolin deposits contain two kaolin beds; a lower nodular kaolin deposited by mass wasting and an upper pisolitic kaolin deposited as fluvial clastic deposits. EL-Askary (1992) concluded that the pisolitic kaolinitic deposit of Wadi Kalabsha was formed inorganically from detrital kaolinite-iron hydroxide colloid coagulates during early subaqueous diagenesis in agitated conditions.

Paleomagnetic reconstruction based on Late Cretaceous sandstone, volcanic rocks and hematitic iron ores of the Wadi Natash and Aswan area (El-Shazly and Krs, 1973; and Schult et al., 1978) as well as the tropical vegetation indicated by the flora of claystones accompanying the Aswan oolitic iron ores (Ger-
mann et al., 1987) may confirm the deposition of these rocks in the paleo-equatorial to sub-equatorial zone where warm and moist climate could produce kaolinite, iron oxides and bauxite in the source areas as a lateritic weathering crust over the nearby basement rocks which composed mostly of gneiss, migmatitic gneiss, granites and metavolcanic rocks.

Hendriks et al. (1987) concluded that the deposits of Abu Aggag Formation attain a thickness of 226 to 312 m at the Qena–Kom Ombo area and 15 to 80 m at Aswan–Abu Simble area. They also concluded that the pisolitic kaolin deposits at central Wadi Kalabsha occur within the Abu Aggag Formation of El-Naggar (1970) (Sabaya Formation of Klitzsch and Wycisk, 1987; and Issawi and Osman, 1993).

### II. Occurrence and Petrography

The kaolin deposits within the Sabaya Formation (Albian age) in the central Wadi Kalabsha, are characterized by greyish white, yellowish white, yellowish brown to brown, violet and red pisolitic kaolin lenses, 1 to 5 m thick and some 40 m in length, at and near the ground surface. This pisolitic facies is intercalated with thin (2 to 5 mm thick) kaolinite or hematite laminae which occur as wedge filling between pisoids. The exposed top parts of these lenses are sometimes affected by in situ lateritization processes. Laterally and below the pisolitic kaolin there are intraformational conglomerate lenses of about 50 cm thick of kaolinite and hematite clasts embedded in either kaolinite and/or hematite matrix (Fig. 3).
An exploitable mottled kaolin deposit, 3 to 7 m in thickness with red and violet oxides along joints and fractures, is commonly exposed on the ground surface as well as below the pisolitic kaolin. Intraformational conglomerate lenses are also observed below and within the mottled kaolin deposits. Lenses of sandy kaolin, and purple stained sandstone of about 50 cm thick are sometimes intercalated within the kaolin deposits.

The intraformational conglomerates are matrix supported and with either rounded to subrounded or angular to subangular clasts ranging in size from 2 cm to more than 5 cm. The rounded clasts are mostly elliptical and show prefered orientation of long axis with their flattened surfaces dipping in the same direction (Fig. 4A); whereas the angular clasts generally show an unorganized fabric (Fig. 4B).

The pisoids are either simple or compound (Fig. 4C). The simple pisoids consist of a central nucleus surrounded by an envelope of concentric laminae, whereas the compound pisoids consist of more than one pisoid surrounded by a common concentric envelope. The simple pisoid is considered as normal when the diameter of nucleus is less than the total thickness of envelope; and as superficial when the diameter of the nucleus is greater than the total thickness of the envelope. Pisoids are often associated with spherical or elliptical clasts, in the form of nodules of corresponding size range, which lake the concentric structure. These grains could be refered to as pseudo-pisoids.

The nucleus plays a major role in the shape of pisoid. Equant and platy nucleii resulting in spherical and elliptical pisoids respectively. The pisoids which have diameters between 1 to 7 mm are mostly spherical, but sometimes, they are elliptical, slightly deformed and are moderately to well sorted. The elliptical pisoids are sometimes oriented parallel to the general stratification of laminae. The laminae of the envelope are either continuous or discontinuous around the nucleus. The continuous laminae may be represented by one or many series of laminae of different thickness and/or composition (Fig. 4D). The discontinuous laminations can form either one complete series around the nucleus or are developed on one side as local overgrowths.

Based on the kaolinite/haematite ratio of the pisoids, Szamatek et al. (1993) distinguished four types of pisoids in the Wadi Kalabsha pisolitic kaolin. These types of pisoids are: 1) pure kaolinitic; 2) hematitic-kaolinitic; 3) kaolinitic-hematitic; and 4) hematitic. However, in the present work, the author differentiated the pisoids at central Wadi Kalabsha, according to the mineral composition of the nucleii and envelopes, into three types.
The first is the kaolinitic pisoids in which the nucleus and envelope are mostly composed of kaolinite. These pisoids are usually lighter in colour than the matrix. They are greyish white in colour as compared to yellowish, earthy white colour of the matrix. The second is the hematitic pisoids in which both the nucleus and envelope are composed mainly of hematite. They range in colour from violet, pale red to red. These pisoids exist in either yellowish white kaolinitic matrix or in violet, yellowish red and red hematitic matrix. The third is the mixed kaolinitic-hematitic and hematitic-kaolinitic pisoids in which a kaolinite or hematite nucleus is surrounded by an envelope of alternating kaolinite and hematite laminae. They also range in colour from violet, pale red to red and exist in either yellowish white kaolinitic matrix or in violet, yellowish red and red hematitic matrix. The nuclei and envelopes of pisoids are characterized by thin irregular cracks. In hematitic
envelope and nucleus, cracks are filled with kaolinite (Figs. 4E and 4F), while cracks in kaolinitic nucleus are filled with hematite. This may indicate repeated dry and humid seasons during the deposition of these pisoids in the fluvial environment (depositional or sedimentary pisolites). This type of pisolites are sometimes encrusted by yellowish white and reddish brown kaolinitic-hematitic and hematitic pisolitic laterite formed by an in situ lateritization process (Fig. 4G). The pisolitic laterite is characterized by the presence of traces of gibbsite, goethite and anatase.

III. Depositional environmets

The intraformational conglomerates are composed of either rounded to subrounded (organized) or angular to subangular (disorganized) clasts. Rounding of the clasts in the organized conglomerates suggests that they have moved as free individuals independent of one another on the surface of the bed and that they have been later incorporated into the bed. On the other hand, the disorganized fabric of the second type of conglomerate suggests that the clasts were either deposited in situ by fragmentation and redeposition with a rapid rate of accumulation; or were deposited from a viscous medium (mass wasting) with a rapid rate of accumulation. An intermediate stage of poorly sorted relatively organized fabric also exists. This may represent deposition of clasts from relatively high viscosity, high density flows in which the clasts were transported with the matrix and were forced by intergranular collision into the position of least resistance to surrounding flow.

The contacts between the organized, relatively organized and disorganized conglomerates are generally distinct, and are marked by variations in grain size and position relative to the pisolitic kaolin. The organized conglomerate is characterized by finer clasts which occur in contact with the pisolitic kaolin, whereas the disorganized conglomerate is characterized by coarser clasts that occur far from the pisolitic kaolin. These distinct facies with marked contacts are the product of a series of discrete depositional events which denote changes in energy or subenvironment. This may represent the change from river channel to flood plain deposits in an area which is inundated only during high flood.

Drying out of an exposed muddy sediment on a river flood plain, causes contraction, giving an isotropic horizontal, tensional stress field which resulted in the formation of desiccation mud cracks (polygons) which range in diameter from millimeters to centimeters. The mud cracked layer was rolled-up, and the roll-up clasts were subsequently reworked by the high fluvial influx over the flood plain. Angular to subangular disorganized clasts were deposited in situ over the flood plaine, and/or on the flat bar tops, whereas, the rounded to subrounded, spherical and elliptical organized clasts, arising through constant abrasion and reworking by fluvial action, were deposited either on the slip faces of bars and river banks and/or as lateral point bar deposits of sinuous streams.

Interbedded kaolinite laminae as well as the coating of some clasts with thin kaolinite laminae forming superficial pisoids may represent periods of protracted setting from suspension during more quiet sedimentary regimes. However, the continuous laminae of the envelope which resemble those of ooids, suggest continuous movement under agitated conditions. On the other hand the discontinuous laminae that form a complete series around the nucleus, suggest intermittent agitated and quiet conditions, while the discontinuity due to local growth of laminae reflect periods of non-agitation where the particles were static. The pisolitic facies shows horizontal and inclined laminae and sometimes tabular cross
stratification, which may represent either a fill of erosion hollows and/or the deposition on the slip faces at the downstream end of longitudinal and diagonal bars in braided rivers.

IV. Origin of kaolin deposits

The kaolin originates in at least three widely different geochemical environments: by weathering, by hydrothermal alteration and by crystallization from colloids having the chemical composition of kaolinite (Keller, 1982). The studied kaolin is compact, non-fissile, resists slacking (non-slacking) and has almost no natural plasticity when immersed in water. Such type of kaolin (flint clay) is not a first-cycle direct-weathering product of feldspar as is free-slacking kaolin (Keller, 1982). In situ crystallization from colloids having the chemical composition of kaolinite is supported by the tightly interlocking, randomly oriented, compact packets and sheaves of kaolin crystals exhibit mutual-boundary texture. The envelope of the kaolinitic pisoid shows a concentric arrangement of these compact packets and sheaves of kaolin crystals (Fig. 5A). However, the matrix between pisoids is mostly composed of poorly arranged detrital platy kaolinite crystals that represent primary products of weathering (Fig. 5B). The matrix is generally porous, while the envelope is non-porous. This may be due to the effect of in situ crystallization and rolling on the arrangement and packing of the kaolinite crystals within the envelope and the loose aggregation of randomly oriented detrital kaolinite flakes within the matrix. A rose like (rosette) of authigenic hematite crystals exist in the pore spaces within the matrix of the depositional pisolite (Fig. 5C). On the other hand radial aggregates of goethite (goethite after hematite) are usually observed in the hematite matrix of the pisolitic laterite (Fig. 5D). Quartz grains in the pisolitic laterite are generally etched and embayed by hematite crystals (Fig. 5E). This facies also includes sharp angular voids with remnants of small etched quartz grains in some of these voids denoting that quartz has been dissolved by diagenetic action. However, the fresh sharp and angular quartz grains in the depositional pisolite and the mottled kaolin deposits may show that the in situ lateritization process is restricted to the exposed crustal layer.

Kaolinite, quartz, feldspar and hematite are present throughout the section with kaolinite being by far the dominant mineral. In the crustal lateritic zone traces of gibbsite, goethite and anatase are present. This observation as well as the absence of the in situ lateritization products within the kaoline section may indicate that the crustal layer of the kaolin deposits was subjected to tropical weathering conditions after deposition, where kaolinite was altered to gibbsite, goethite, hematite and anatase. In the southern part of Wadi Kalabsha, Fisher and Germann (1987), identified a higher percentage of gibbsite and anatase and concluded that the kaolinite deposits were subjected to an in situ lateritization process, leading even to a boehmite-bearing latosol profile.

V. Hydrogen and oxygen isotopic compositions

Since the hydrogen isotope fractionation factor between kaolin and water is not sensitive to the temperature in the range below 250°C (Marumo et al., 1980), the hydrogen isotopic composition of kaolin minerals is expected to depend mainly on that of water with which the respective kaolin mineral was in equilibrium (Marumo et al., 1981). However, the oxygen isotope fractionation factor varies considerably with temperature and therefore suggests that the oxygen isotopic composition of kaolin minerals provides a useful clue for distinguish-
Fig. 5. A: SEM photograph showing higher porosity in the matrix between pisoids (M) than the envelope (E) of kaolinitic pisoids.
B: Hypidiotopic to idiotopic kaolinite crystals in the matrix of kaolinitic pisoids. (Photograph courtesy of Dr. Szamatek).
C: Rose-like hematite crystals (rosette) filling pore spaces in the hematite matrix of the depositional pisoids.
D: Radial aggregates of goethite after hematite crystals in the hematitic matrix of the pisolitic laterite.
E: Corrosion of quartz grain by hematite crystals in the pisolitic laterite facies.

ing whether they are hydrothermal or weathering in origin (Marumo et al., 1981). The isotopic composition can be used as a paleoclimatic indicator in paleosols of weathered clay (Savin, 1989).

Marumo et al. (1981) showed that kaolins of hydrothermal origin have values of \( \delta^{18}O \) ranging from +1 to +9\(^{\circ}\), whereas kaolins of weathering and sedimentary in origin have values of \( \delta^{18}O \) ranging from +9 to +19\(^{\circ}\). Murray and Janssen (1984) showed that kaolins formed by hydrothermal alteration have values of \( \delta^{18}O \) ranging from +2 to +14\(^{\circ}\), whereas kaolin formed by surficial weathering have values of \( \delta^{18}O \) ranging from +15 to +19\(^{\circ}\) and that is of sedimentary kaolin ranging from +19 to +23\(^{\circ}\). The pisolitic kaolin at central Wadi Kalabsha, southwest of Aswan, Egypt, have values of \( \delta^{18}O \) ranging from +16.3 to +18.2\(^{\circ}\). In accordance with Marumo et al. (1981) and Murray and Janssen (1984), the pisolitic kaolins were formed by surficial weathering. This is in agreement with Szamatek et al. (1993) conclusion. The plots of \( \delta D \) and \( \delta^{18}O \) values (Fig. 6) for the studied kaolins on the \( \delta D-\delta^{18}O \) diagram showed that they are near the kaolinite line of Savin and Epstein (1970) and are of weathering and sedimentary origin. Oxygen isotope data obtained by Clauer et al. (1989) for kaolinite existing in the lateritic profile indicate the small but significant variations of \( \delta^{18}O \). They showed that towards the top \( \delta^{18}O \) changes
from +17.3%\(_o\) (saprolite zone) to +17.4%\(_o\) (topsoil zone). A similar observation is clearly seen in the studied kaolinite profile, where there is a significant variations of \(\delta^{18}O\) values from +16.3%\(_o\) (depositional pisolite or saprolite zone) to +17.9%\(_o\) and +18.2%\(_o\) (pisolitic laterite or topsoil zone).

Bird et al. (1989) determined \(\delta^{18}O\) %\(_o\) in kaolinitic weathered profiles of different ages. They concluded that Permian clays have values of \(\delta^{18}O\) ranging from +6 to +9%\(_o\), the pre-Late Mesozoic clays have values of \(\delta^{18}O\) ranging from +10 to +15%\(_o\), the Late-Mesozoic/Early Tertiary clays have values of \(\delta^{18}O\) ranging from +15 to +17.5%\(_o\), and the post-Mid Tertiary clays have values of \(\delta^{18}O\) ranging from +17.5 to +21.4%\(_o\). Based on Bird et al. (1989) conclusion, the depositional pisolites studied in the present work which have \(\delta^{18}O=+16.3%\(_o\)\), may indicate that they are of Late-Mesozoic age, whereas the pisolitic laterite, which have \(\delta^{18}O=+17.9%\(_o\)\) to +18.2%\(_o\), indicate that they were formed later (probably during the Tertiary) as an in situ lateritic topsoil zone. Because the \(\delta D\) values of meteoric water largely depends on the latitude of sample locality (Sheppard et al., 1969) the difference in \(\delta D\) values of the studied kaolinite (from −70.7%\(_o\) for the depositional pisolites to −59.5%\(_o\) and −63.0%\(_o\) for the pisolitic laterite) may suggest a change in the latitude of the sample locality and the age of their formation.

VI. Conclusions

The sedimentary structures and textures as well as the mineralogical characteristic of the kaolin deposits in central Wadi Kalabsha, southwest of Aswan, Egypt, suggest the following depositional and diagenetic models.

1) Under humid, temperate climatic conditions, during the Mesozoic, the exposed basement complex nearby Wadi Kalabsha area was subjected to intensive weathering forming kaolinite and iron oxides as a thin cover of lateritic crust. This crust was scoured off by fluvial action and deposited in the study area as flood plain and river bar deposits.

2) Drying out of the deposited muddy sediments resulted in the formation of desiccation mud cracks. The mud cracks were subsequently reworked by a new high fluvial regime, where the river channel has switched or migrated to areas of the pre-existing river flood plain. This resulted in the formation of a) intraformational disorganized conglomerate over the flood plain and/or on the flat bar top; b) intraformational organized conglomerate either on the slip faces of bars and river banks and/or as lateral point bar deposits in sinuous streams; and c) pisolitic facies, composed of kaolinite, kaolinite-hematite, hematite-kaolinite and hematite and are characterized by horizontal and inclined laminae as well as tabular cross stratification which formed either in erosion hollows and/or on the slip faces at the downstream ends of longitudinal and diagonal bars of braided rivers.

3) The in situ lateritization of the depositional kaolinite sediments resulted in the formation of kaolinitic-hematitic and hematitic pisoids in the crustal layer. In this lateritic
zone, it appears that some of the detrital kaolinite crystals and most of the quartz grains have been dissolved. Traces of gibbsite, goethite and anatase are recorded in the lateritic zone.

4) The δD-δ18O diagram as well as the δ18O values (+16.3‰ to +18.2‰ SMOW) of the pisolitic kaolin indicate that they are sedimentary and weathering in origin. The δ18O values show a significant variation from +16.3‰ to +17.9‰ SMOW in the depositional pisolite and +18.2‰ SMOW in the crustal pisolitic laterite. These values may show that the depositional pisolite is of Late Mesozoic age, while the pisolitic laterite is of younger age probably Tertiary age.

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