

## THE BAUXITE OF JURUTI

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### Introduction

The extensive bauxitic terrains of the lower Amazon area exhibit common features, which have been described and discussed by a number of authors (Assad, 1973; Dennen & Norton, 1977; Grubb, 1979; Kronberg et al., 1979, 1982; Aleva, 1981; Kotschoubey & Truckenbrodt, 1981, 1984). Authors generally agree with the macroscopic description of the bauxitic profiles, which are characterized by six main horizons, from the top downwards:

- a kaolinitic loose yellow clay horizon, 1-14 m thick (also called "Belterra Clay"),
- a nodular horizon, 0,2-3 m thick, formed by gibbsitic and hematitic nodules embedded in a yellow clay material,
- indurated iron-rich horizons, 0.2-3 m thick,
- indurated bauxite, 1-9 m thick,
- kaolinitic mottled horizons, more than 2 m thick, white, and pale red in color,
- parent material, which is a quartzo-kaolinitic weathered clastic sediment.

However, authors differ significantly on the genetic interpretations of this profile. Some of them (Grubb, 1979; Aleva, 1981; Kotschoubey & Truckenbrodt, 1981, 1984) stress that the bauxitic horizons have been formed in the upper part of profiles which have suffered later erosion and re-deposition processes, considering, for example, the upper yellow clay horizon as a lake deposit named "Belterra Clay". Nevertheless, interpretations differ in the number of phases and the horizons which are considered as allochthonous. Others stress that the profile remained in situ since bauxitisation time, but differ in the processes involved. Dennen and Norton (1977) concluded that Al and Fe were leached from the upper part of the profile, transported downwards in solution and accumulated at depth. Kronberg et al. (1982) assume high flow rates of meteoric waters through highly permeable upper horizons, resulting in gibbsite formation by relative enrichment of Al beneath a less weathered zone. The possible occurrence of erosional-depositional processes during the profile formation is an important question in the history of the lower

## Amazon basin.

The purpose of this paper is to describe the alteration minerals and textures of sedimentary rocks in Central Amazonia, to infer the genetic sequence that has led to the present bauxites, and to discuss the pedogenetic and geochemical consequences of these results. We find that the entire profile is characteristic of an in situ geochemical differentiation. If erosional or depositional events have occurred in the past, sedimentological features and depositional facies have been completely erased or transformed by a subsequent geochemical evolution. Present-day evolution of the profile shows that gibbsite is currently precipitated in the zone of aeration, below a thick stable kaolinitic zone, which is at variance with most models of supergene weathering.

### I. The studied areas

The studied bauxitic formations are located in the lower part of the Amazon basin (Fig. III.1), between the Guyana and the Brazilian shields. The Juruti area, south of the Amazon river ( $2^{\circ}30'S$  and  $56^{\circ}14'E$ ), and the Trombetas area, north of the Amazon river ( $1^{\circ}27'S$  and  $56^{\circ}24'E$ )

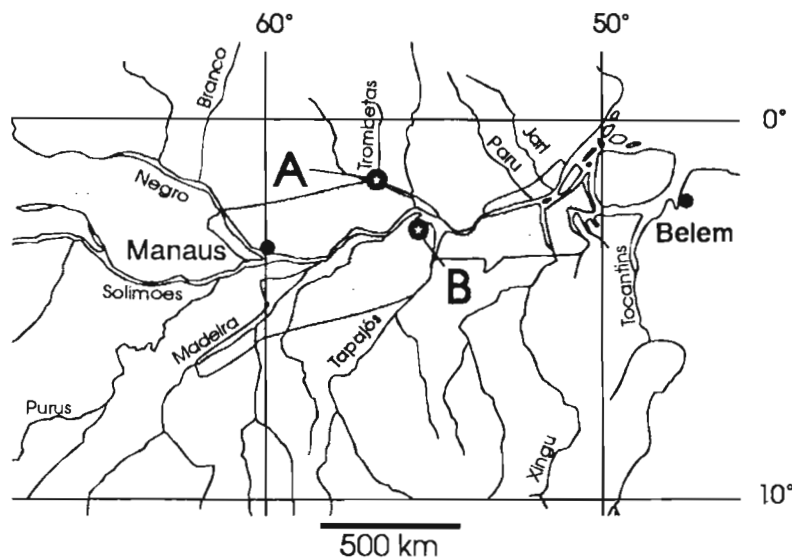


Figure III.1 - Location of the studied areas. A : Trombetas bauxitic area, B : Juruti bauxitic area. In gray, extent of the low plateaux landscape (data from Radam Brazil, 1978).

are amongst the major bauxite regions in the world (Bardossy, 1983).

The sediments beneath the bauxitic formation are composed of alternating clayey, silty and sandy layers of a weathered clastic sediment consisting of kaolinite, quartz and a small quantity of iron and titanium oxides. Although it lacks fossils, this sediment is related to the Cretaceous Alter-do-Chão Formation (Putzer, 1984).

The terrain forms gently undulating plateaus, 100 to 170 m above the level of the Amazon. This plateau landscape covers a great area north and south of the Amazon river (Fig. III.1), where most of the plateaus are not bauxitic, but covered by thick kaolinitic soils overlying the same Alter-do-Chão Formation (Chauvel et al., 1982; Lucas et al., 1986). No apparent geomorphologic features distinguish the bauxitic plateaus from the non-bauxitic ones. According to Radam Brasil (1978) or Klammer (1984), the plateaus are dissected remnants of a Plio-Pleistocene surface.

The present-day climate is equatorial, of Amazonian type, with a slight dry season and an average annual rainfall of 2100 mm. The average annual evapotranspiration is 1600 mm and the surface runoff on the plateaus is negligible (Franken & Leopoldo, 1988), so that only 500 mm percolates annually through the soil. The vegetation is a dense, humid, evergreen rain forest.

## **II. Analytical procedures**

The macroscopic investigations were done in 5 to 25 m deep surveying shafts and road cuts (Juruti area) and on mine stope faces (Trombetas area). Color names are given from the Munsell Soil Color Chart (1954). Microscopic investigations were done by optical microscopy on thin section, and by Scanning Electron Microscopy (SEM) and Scanning Transmission Electron Microscopy (STEM) observations on suspensions and ultra-thin sections. Mineral constituents were identified and studied by X-rays diffractometry, thermogravimetric analysis and IR spectrometry on bulk samples or on microsamples collected by microdrilling of small volumes. Chemical analyses were done by atomic absorption spectrometry after strontium metaborate melting and by microprobe analyses on thin sections (Energy Dispersive Spectra). Porosimetry of the different facies was carried out by mercury pressure porosimetry (Fies, 1984). The pressure  $P$ , necessary to inject mercury in a given pore, is proportional to the inverse of the equivalent radius  $f(P)$  of the pore. Thus, volume of mercury injected in a sample between a  $P_1$

and P2 pressure indicates the volume of pores having an equivalent radius between  $f(P1)$  and  $f(P2)$ . The porosimeter which was used is a Carlo Erba 2000, permitting measurement of pores volume between 3.75 nm and 100  $\mu\text{m}$ .

### III. Profile description and petrographical characteristics

A typical plateau profile consists of four main horizons which are described from the base to the top as follows (Fig. III.2). A facies is a characteristic association of authigenic texture and minerals; the different facies are identified in each horizon.

#### 1. Mottled horizon

It consists of a white and pale red mottled kaolinitic clay, so called "mottled kaolinitic facies", organized into a polyhedral structure with platy structural voids. Transition from the underlying parent sediment is gradual: the sub-horizontal structure of the parent rock is cut by vertical fingers of the mottled kaolinitic facies, which upward anastomose with each other

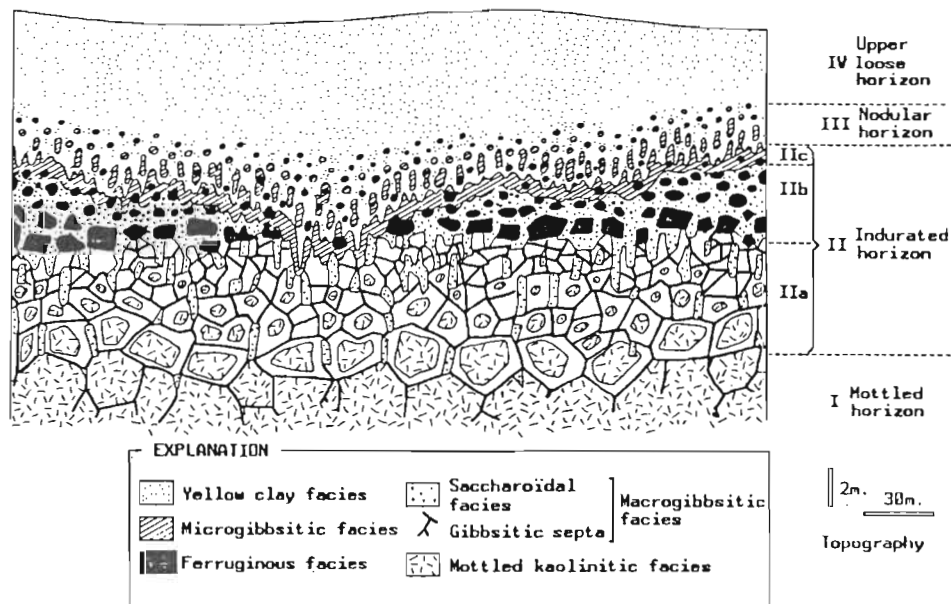


Figure III.2 - A typical plateau bauxitic profile of Juruti.



**2a** - With mottled kaolinitic and macro-gibbsitic facies. Ascending in the profile from horizon I, the gibbsitic septa become thicker and more abundant, giving a boxwork indurated structure. The mottled kaolinitic facies is reduced to rounded remnants in the core of the boxwork cells, the macroporosity (voids more than 0.1 mm) considerably increases. A "saccharoidal facies" appears and gradationally develops upwards. It is very indurated, white to pink colored, and forms centimetric to pluridecimetric vertical columns. It consists of a mosaic of coarse gibbsite crystals, whose fabric is dissected by tiny spherulae of kaolinite (Fig. III.3b). Both septa and saccharoidal facies consist, mainly, of coarse gibbsite crystals (1 to 20  $\mu\text{m}$ ), they are called "macro-gibbsitic facies". Some large vertical tubular voids, where water, now, percolates actively, are coated by unaltered coarse crystals of gibbsite.

**2b** - With macro-gibbsitic and ferruginous facies. In the middle part of the horizon II, the gibbsitic septa gradationally disappear as the saccharoidal facies becomes abundant and turns yellowish-red. A red-colored "ferruginous facies" is developed, forming numerous angular-shaped patches (average size 10-15 cm). In the upper part of the horizon II, the patches of ferruginous facies diminish in size to reach 1 to 3 cm of average diameter, and their shapes change, gradationally, from angular to rounded (Fig. III.4a). The ferruginous facies is, mainly, composed of hematite, kaolinite and gibbsite, with platy structural voids lined with coarse palisade hematite, and 0.1 to 3 mm cavities which exhibit an angular shape and internal hematitic septa. These cavities in the ferruginous facies are scarce in the lower part of the horizon and become numerous towards the top of the horizon. Their size, shape and septa are pseudomorphic after quartz grains and of their cracks, as those are observed in the sediment at depth. Many of these pseudomorphic cavities are secondarily filled by gibbsite macrocrystals (Fig. III.3d). The center of the larger pseudomorphic cavities is often occupied by a strongly corroded quartz grain. The association of both saccharoidal and ferruginous facies forms a very indurated breccia-like structure. On a microscopic scale, however, the saccharoidal facies clearly cuts across the structures of the ferruginous facies (Fig. III.3c).

**2c** - With micro-gibbsitic facies. The top of the horizon II is very irregular and interfingering with horizon III. It is sublimed by an indurated micro-horizon (1 to 30 cm thick) which consists of a hard, compact, porcelaneous, yellow to pink facies. This facies is made up of a mosaic of small crystals of gibbsite (<1  $\mu\text{m}$ ), so it is called "micro-gibbsitic facies" (it is sometimes called "porcelaneous facies" in literature). The micro-gibbsitic facies obviously cuts across the structure of the macro-gibbsitic

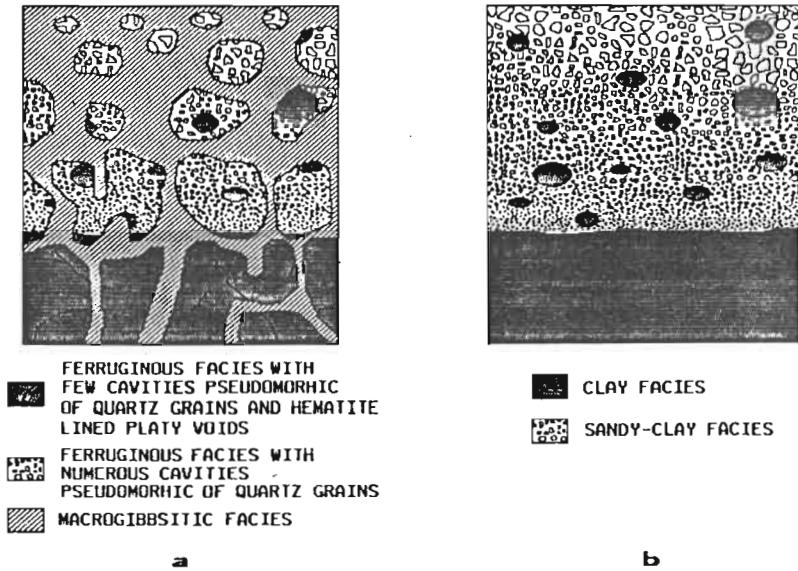


Figure III.4 - a: sketch of the vertical sequence of ferruginous facies in the upper part of the indurated horizon; b: reconstitution of the initial sediment structure.

facies as well to that of the ferruginous facies (Fig. III.3e).

### 3. Nodular horizon

The upper limit of the micro-gibbsitic facies horizon is irregularly penetrated by tongues of a yellow loose clay facies, so-called "yellow clay facies". This facies develops upward at the expense of the microgibbsitic facies; isolated vertically lengthened and contorted blocks and nodules of gibbsitic facies of gradationally smaller size are observed upwards through the horizon. Rounded small spots of ferruginous facies are observed in these nodules and blocks.

### 4. Upper loose kaolinitic horizon

This horizon, apparently greatly homogeneous, is composed mainly of the yellow clay facies, and by sparse micronodules (diameter < 1 mm) of ferruginous or microgibbsitic facies. The structure is finely polyhedral, with rounded micro aggregates ( $\varnothing < 0.1$  mm) among the elementary polyhedrons. The yellow clay facies consists of kaolinite, closely associated with small gibbsite and goethite crystals (size < 1  $\mu$ m). On a microscopic scale, the transition with the microgibbsitic facies is deeply

late

1928), in which the very progressive size corresponds to a progressive characteristics, as measured from infra

The above description re-  
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- \* The second group cuts across the former succession:
  - the microgibbsitic facies, as a micro-horizon (sub-horizon IIc) at the top of the indurated horizon, and as blocks and nodules in the nodular horizon (horizon III);
  - the yellow clay facies which constitutes the upper loose horizon (horizon IV)

#### **IV. Water circulation in the profile**

Plateau profiles are entirely situated high above the water-table level, in the zone of aeration. Field observation and measurement of moisture content in the upper loose horizon (Chauvel & Lucas, 1988) reveal that water percolates slowly and steadily in the upper loose horizon. The roots uptake in the top meter of the profile is sufficient to buffer the free water (water with pressure potential > atmospheric pressure) which appears in the topsoil after the rains, even in very rainy season. Deeper, water percolates mainly as capillary water (water with pressure potential < atmospheric pressure). Rosanski et al. (1991) give 2 m. within 139 days as the downward movement of the soil moisture in the same upper loose horizons from Manaus area. Extrapolation of this result gives more than 1,5 year for water percolating through an average 8 meters thick upper loose horizon. Free water reappears at the base of the upper loose horizon, even at the end of the dry season, then percolates downwards as laminar films in the coarse voids of the indurated horizon.

#### **V. Chemical and mineralogical composition**

Chemical and mineralogical compositions of the main facies are given in Table III.1. Absolute values of the composition may change from a profile to another, but relative abundance of the different elements and minerals varies in the same way along all the profiles. Compositions of the mottled kaolinitic facies and the microgibbsitic facies are slightly variable; compositions of the macrogibbsitic facies and the ferruginous facies are highly variable, according to their microscopic heterogeneity; the yellow clay facies exhibits a very progressive diminution of gibbsite content from 10-12% in depth to less than 5% near the topsoil. The sediment now observed at depth in Juruti area were studied from a drilling penetrating 20 m below the bauxitic profile. Analysis of 20 samples (a sample each meter) gives an average composition of the

Table III.1 - Chemical and calculated mineralogical composition of the different facies. Normative values in % weight. avg. = average, std = standard deviation.

Type of facies		Chemical composition				Mineralogical composition					Aluminium substitution in Fe oxides (molar %)	
		SiO <sub>2</sub>	Al <sub>2</sub> O <sub>3</sub>	Fe <sub>2</sub> O <sub>3</sub>	TiO <sub>2</sub>	Kaolinite	Gibbsite	Iron as Fe <sub>2</sub> O <sub>3</sub>	Quartz	Anatase		
Yellow clay (1 m. above indurated facies, 10 analyses)	avg	47,5	41,9	7,7	2,9	78,0	8,5	6,2	4,8	2,5	Goethite Hematite	30 9
	std	0,8	1,1	0,2	0,4							
Microgibbsitic (22 analyses)	avg	5,6	86,5	5,1	2,8	8,4	86,2	3,5	0	1,9		
	std	3,0	3,8	1,9	2,2							
Ferruginous (examples)		3,5	4,1	90,5	1,8	2,2	0,4	93,1	2,4	1,8	Hematite	3
		10,5	26,4	62,3	0,8	20,3	22,9	56,1	0	0,7		
		18,1	25,6	54,7	1,7	35,6	10,9	51,9	0	1,5	Hematite	5
		27,9	24,2	46,8	1,1	55,5	0,3	43,2	0	1,0		
Macrogibbsitic (examples)		3,1	89,6	5,9	1,5	4,5	90,5	4,0	0	1,0	Hematite	4
		15,6	65,0	16,4	3,0	25,5	59,9	12,4	0	2,2		
		26,6	70,1	1,9	1,3	43,1	54,9	1,1	0	1,0		
White kaolinite (13 analyses)	avg	51,4	44,5	1,7	2,3	95,5	1,9	0,6	0	2,0	Hematite	1
	std	0,6	0,9	0,4	1							
Pale red kaolinite (13 analyses)	avg	47,7	43,6	5,4	3,4	88,5	4,7	3,9	0	2,9		
	std	2,4	1,9	2,5	0,7							

sediment at 51% kaolinite, 43% quartz sand and 6% others, but the composition of the sediment is quite variable from one sample to the other, the kaolinite content ranging from 20 to 76%, and the heavy minerals content from 0.2 to 5%. The heavy minerals observed along the profile are mainly zircon and Fe-Ti oxides (ilmenite, rutile, anatase). A weathering sequence of the Fe-Ti oxides grains is observed from the base to the top of the profile, characterized by the progressive diminution of the grain size, the increase of the grain porosity, the diminution of their ilmenite content and the increase of their rutile and anatase content.

## VI. Discussion

As described above, the bauxitic profile is formed by four main horizons, each one being characterized by the nature and fabric of various interpenetrated facies. Cross-cutting relations between the facies, geometry of the facies boundaries and vertical variation of facies textures give information about the facies genesis and the profile formation.

### 1. Genetic relationships between facies

Deeply contorted and interpenetrated boundaries between facies are inconsistent with allogenic hypotheses, and indicate an authigenic origin of the facies, which have geochemically replaced each other in

situ. The direction of the replacement is given by the cross-cutting relationships on a microscopic scale. The following relationships are thus assessed from petrographical data:

- \* the macrogibbsitic facies has replaced in situ the ferruginous facies;
- \* the microgibbsitic facies has replaced in situ the macrogibbsitic facies and the ferruginous facies;
- \* the yellow clay facies has replaced in situ the microgibbsitic facies.

Regarding the ferruginous facies, the progressive vertical sequence of size, shape and texture of the ferruginous patches included in the gibbsitic indurated material is another evidence against a clastic origin of the patches (Fig. III.4a). Thus, we infer that they are the relic of a former continuous ferruginous horizon, which have been partially dissolved and replaced in situ by the macrogibbsitic and microgibbsitic facies. Reconstitution of the vertical sequence of quartz repartition in the parent material before ferruginisation and gibbsitisation is sketched on Figure III.4b. Such a repartition of quartz grains is frequently observed in the Alter-do-Chão sediment. This suggests that the vertical variation in the ferruginous facies texture is originated from an heterogeneous parent material, which was formed by a clay material buried by a sandier layer. The hematitic impregnation of the clay material has given a ferruginous facies with hematite-lined platy voids, which are relics of the structural clefs in the clay. The hematitic impregnation of the sandier layer has lined the quartz grains by hematite; the later dissolution of quartz grains has left pseudomorphic cavities, a few of which containing a corroded relictual quartz grain.

Regarding the macrogibbsitic facies, the network of septa and the palisade crystalline fabric of the septa suggests a genesis by progressive growth in the structural clefs of a clay material. All of the septa and saccharoidal facies at depth, however, exhibit microscopic features of dissolution and replacement by a kaolinitic material; thus, they are relics of a former process. Septa and vertical columns of saccharoidal facies have been formed at the expense of a clayey parent material, and afterwards have suffered dissolution processes. Only scarce gibbsite macrocrystals without features of corrosion are observed at depth, coating large voids where water now percolates. This means that genesis of gibbsite macrocrystals occurs nowadays at depth, but is limited to active water percolation zones.

These genetic relationships give the following temporal succession of the facies: (1) ferruginous facies; (2) macrogibbsitic facies; (3) micro-gibbsitic facies; (4) yellow clay facies. This temporal succession is in accordance with the cutting relationships observed at the plateau scale

(Fig. III.2). The consequences of these relationships on the profile genesis will now be examined.

## 2. Profile Genesis

The genesis of old, deeply weathered profiles may combine erosional processes, depositional processes by water or wind, and in situ geochemical processes.

Most models of in situ profile genesis are based on the premise of downward advance of transformation fronts, each horizon having been derived from material similar to that now underlying it (Nahon, 1976; Millot, 1983); the vertical succession of horizons or facies can, thus, be considered to represent a time sequence (Butt & Nickel, 1981).

In the present profile, three main transitions have a fingered, deeply serrated geometry, which is characteristic of geochemical transformation fronts and discards the hypothesis of a sedimentological unconformity.

- At depth, the transition between the sediment and the mottled horizon (I) corresponds to the dissolution of quartz grains in a kaolinitic matrix. This is in accordance with results from Irion (1984), Lucas et al. (1984) or Lucas (1989), who studied genesis of kaolinitic soils formed from the same sediment in the same geographical area.

- At the upper part of the indurated horizon, the transition between the microgibbsitic horizon (IIc) and the underlying indurated horizon indicates the replacement of a highly heterogeneous material by an homogeneous, gibbsitic material.

- Immediately above, the transitions between the microgibbsitic horizon (IIc), the nodular horizon (III) and the upper loose horizon (IV) indicates the replacement of the indurated microgibbsitic facies by the loose yellow clay facies. More, the vertically lengthening of the blocks and nodules is also inconsistent with a depositional origin: the nodular horizon does not correspond to a detrital stone-line, but to a chemically formed stone-line (Lucas et al., 1990). Block and nodules in the nodular horizon are relics progressively dissolved as the front proceeds down.

Regarding the indurated horizon (II), it is assessed above that its constitutive facies have formed in situ by authigenic geochemical processes; allochthonous processes during their differentiation must be discarded. There is, however, a discrepancy between the vertical succession of facies (or sub-horizons), and the temporal succession of facies assessed from microscopic data. More, the structure of the ferruginous facies in horizon IIb is inherited from the sediment structure, when this

sediment structure has disappeared in the underlying horizon. Thus, IIb horizon cannot derive from the horizon now underlying it, and the ferruginous facies is a relic of a former paragenesis. In the same way, it is assessed above that the macrogibbsitic facies is now being replaced by a kaolinitic material, and is a relic of a former paragenesis. The conclusion is that the vertical succession of facies in the indurated horizon cannot be considered as a time sequence; the limits between I, IIa and IIb horizons are not presently active transformation fronts, but are relics of past processes.

Regarding the upper loose horizon, the idea of a clastic clay deposition on top of an old bauxitic profile, which gave rise to the "Belterra Clay" stratigraphic name, was supported in the literature by the following arguments: (1) the ferruginous patches and the nodules found on top of the cemented horizons indicate a detrital stone-line; (2) the upper loose horizon, greatly homogeneous, overlies bauxitic as well as non-bauxitic horizons on a large geographic extent, which indicates a widespread sedimentary process.

The first argument may be discarded in view of the present petrographic data discussed above. Regarding the second argument, the homogeneity of the upper loose horizon is only apparent. On the contrary, its chemical and mineralogical composition change vertically very gradationally. This has also been observed for the upper loose horizon of non-bauxitic plateau soils (Lucas et al., 1984, 1986). Furthermore, studies from Irion (1984) establish, in various sites, the lithodependence of the "Belterra Clay" with the underlying parent material. The conclusions are (1) that the possibility of an old depositional event is restricted to the upper loose horizon; (2) if erosional or depositional events have occurred in the past, sedimentological features and depositional facies have been completely erased or transformed by subsequent geochemical evolution.

These conclusions imply necessary events during the profile genesis, which are explained on Figure III.6. These events are necessary in view of the petrographic data, but other events may have occurred which cannot be deduced from the petrographic analysis.

(1) The first event identified is a ferruginous impregnation of the parent sediment, giving a ferricrete. The ferruginous facies is a relic of this ferricrete. The ferruginous textural sequence (Fig. III.4) is likely due to an initial heterogeneity of the sediment. According to the available data, it is not possible to conclude whether this ferruginisation have occurred in depth or near topsoil.

(2) The second event identified is the formation of macrogibbsitic

facies, which has partly replaced the ferruginous facies. This process has been more intense in the upper part of the indurated horizon, leading to massive bauxite. In the lower part of the indurated horizon, septa were likely formed in the structural clefts of a clayey material. Most of the quartz grains were dissolved, quartz/kaolinite ratio of the parent material is, however, undetermined. Under present-day climate, the strong dissolution of quartz grains in saprolite is broadly observed in the area (Lucas et al., 1984). The problem of the origin of the upper loose horizon remains. Whether the profile has formed entirely in situ, and bauxite formation has occurred at depth, leaving in the topsoil a stock of silica sufficient to form, afterwards, the upper loose horizon; or depositional processes have occurred since the bauxitisation, being, afterwards, morphologically erased, but contributing to the presence of a silicic horizon above bauxitic horizons.

(3) The third event is the current evolution of the profile. On the top of the indurated horizon, intense resilicification is due to the downward geochemical progression of the yellow clay, preceded by a transient microgibbsitic facies. Gibbsite and hematitic nodules in the upper yellow clay are relicts left behind as the yellow clay formation front progresses downwards. In the middle part of the indurated horizon, dissolution of the macrogibbsites gives large voids which cut the facies structure. At the base of the indurated horizon, moderate resilicification is brought about by the replacement of the macro-crystalline gibbsite by a kaolinitic matrix. Locally, the precipitation of macrogibbsite continues as coatings on large active voids.

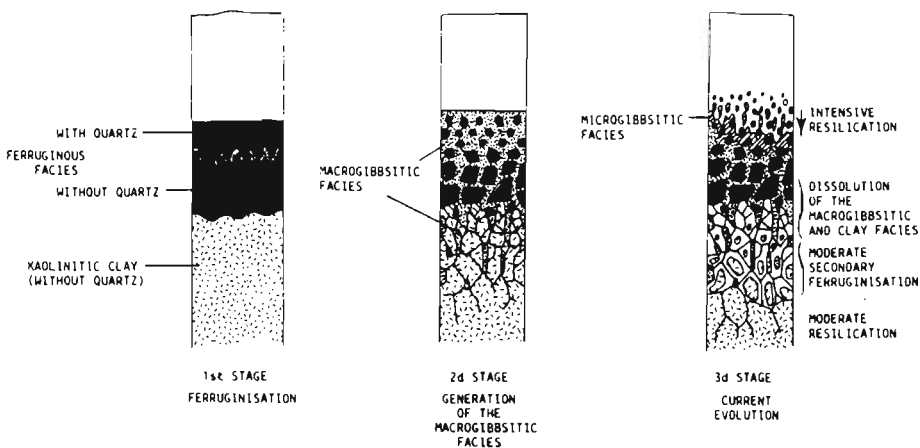


Figure III.6 - Successive stages of the bauxite profile genesis.

3. Geochemical outcomes of the profile current evolution

The current evolution of the profile is sketched on Figure III.7. Gibbsite is currently precipitated as microgibbsitic facies at the base of the upper loose horizon, and as macrocrystalline coatings at depth. The detailed observation of the base of the upper loose horizon shows two successive fronts which follow each other closely: a front of microgibbsitic facies generation, and a front of yellow kaolinitic facies generation. This means that gibbsite is currently precipitated some centimeters beneath an horizon, where kaolinite is currently precipitated: the new profile which develops at the expense of a bauxitic paleoprofile is characterized by the formation of a gibbsitic horizon beneath a thick, stable kaolinitic horizon. In other words, the profile structure characterized by a kaolinitic horizon overlying a gibbsitic horizon is in equilibrium in the climatic zone.

These facts have broad geochemical and pedogenetic consequences, because this general structure of profile is widely observed in the climatic zone, on very ancient soils as well as on younger soils (Leneuf, 1959; Delvigne, 1965; Siefferman, 1969; Boulet et al., 1978; Chauvel et al., 1982; Muller, 1988; Tardy & Novikoff, 1988; Bitom, 1988). Most soils not having undergone hydromorphic and podzolic processes are

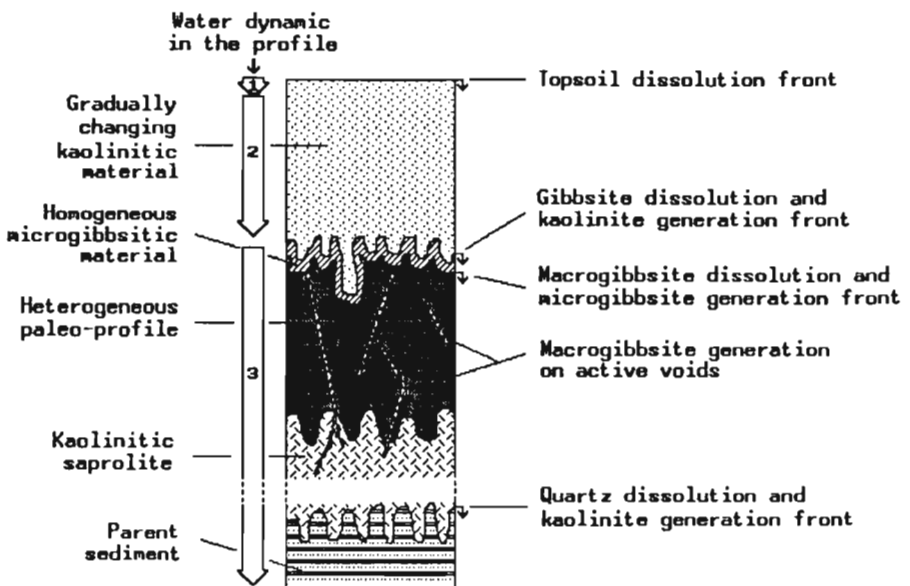


Figure III.7 - Sketch of the current evolution of the profile.

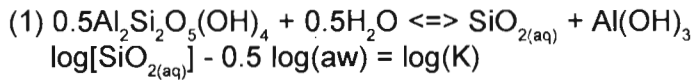
characterized by the accumulation of gibbsite beneath a stable upper kaolinitic horizon. Gibbsite accumulates in the saprolite, or in an intermediate horizon between the saprolite and the upper kaolinitic horizon. Because the gibbsite accumulation in the intermediate horizon often occurs as gibbsitic or gibbsite-hematitic nodules, these horizons have often been considered as detrital stone-lines buried by kaolinitic material itself also interpreted by others as allochthonous. This interpretation was widely accepted in the 50-60s. Since this time, numerous studies have pointed out that most of the upper loose horizon are strictly lithodependent and that most of the stone-lines were formed in-situ (Laporte, 1962; Collinet, 1969; Colin & Lecomte, 1986; Lecomte, 1988; Lucas et al., 1990), bringing up the problem of their genesis. The results from the present study establish that this type of profile is a steady state geochemically in equilibrium in the climatic zone, which questions most of geochemical models for weathering and soil formation: how can gibbsite precipitates beneath a thick, stable kaolinitic horizon, and why is the upper kaolinitic horizon stable?

In models based on relative rates of water flow and reaction (Brown & Garrels, 1980), as proposed by Kronberg et al. (1982), water flows through the upper kaolinitic horizon at a rate much faster than the rate of mineral-water reactions, then is focused in an aquiferous gibbsitic zone. This hypothesis cannot be applied to the present profile, because (1) gibbsite precipitation occurs far above the groundwater level; (2) kaolinite precipitation occurs above the gibbsite precipitation zone; (3) water percolates slowly and steadily in the upper loose horizon (Fig. III.7).

Most of the models of soil formation are based on equilibrium thermodynamics of mineral assemblages (Fritz & Tardy, 1973; Fritz, 1975; Lasaga, 1984; Ambrosi, 1990): the percolating water reaches an equilibrium with mineral species at each level of the profile. In the studied soil, the solution which has percolated the upper kaolinitic horizon are in equilibrium with kaolinite, which is assessed by kaolinite generation at the base of this horizon. To generate gibbsite immediately underneath, some parameters of the equilibrium have to change. Tardy et al. (1988) explain the presence of gibbsite beneath kaolinitic horizon by seasonal fluctuation of the water activity ( $a_w$ ) along the profile, due to water uptake by roots. In the studied profile, there is actually a change of the physical state of the water at the front between the upper loose horizon and the microgibbsitic horizon. Above the front, water percolates as a capillary water which is at a negative pressure potential (positive suction), which means  $a_w < 1$  (Bourrie et al., 1983). Beneath the front, water percolates as free water, at a zero pressure potential, which means



$a_w=1$ . Values of  $a_w$  in the upper loose horizon are, however, always close to 1. Even considering that, during exceptionally dry periods, soil suction can reach 15 bars, which is the limit for roots water uptake, maximum change of  $a_w$  at the transformation front would be from 0.989 to 1.



K: equilibrium constant of the reaction

$a_w = [\text{H}_2\text{O}]$ : water activity

Considering the equation (1), such change of  $a_w$  corresponds to a change of +0.0024 for  $\log[\text{SiO}_{2(\text{aq})}]$ , i.e. a +0.55% shift of  $[\text{SiO}_{2(\text{aq})}]$  at the equilibrium gibbsite-kaolinite. Though the direction of the resulting equilibrium shift agrees with the petrographic observations, this change seems actually very small.

Studying the soil solution in weathering ferrallitic systems from the same climatic zone (French Guyana), Grimaldi (1987, 1988) has established a differential behavior for Al and Si in the soil solution: Al concentration is the higher in the topsoil, and progressively decreases with increasing water time residence in the soil, that is as the water percolates in the soil; but Si concentration increases with time residence. This means that Al and Si are released in the topsoil, and that Al is controlled in depth by a mineral likely to be gibbsite, as Si is preferentially leached out of the system, and questions the presence of organic  $\text{Al}^{3+}$  chelates migrating through the upper kaolinitic horizon. In agreement with this results, Chauvel et al. (1989) show that significant quantities of Al, Si are released by dissolution of the soil material in the topsoil, due to microbiological activity under native Amazonian forest.

The conclusion is that the present geochemical dynamics of the weathering cover is in agreement with results from petrographical observations: the profiles characterized by a kaolinitic upper horizon above more gibbsitic horizons are in equilibrium with present climate of the equatorial zone. Geochemical modeling of these soils has to take into consideration biological activity in the topsoil and to identify the chemical state of the mobile Al.

#### 4. The Problem of the Upper Loose Kaolinitic Horizon

From the results exposed above, the presence of the upper loose

horizon can be explained only by geochemical processes, without the need of an allochthonous deposit of siliceous material on top of the profile. In this hypothesis, the source of silica for current resilicification is due to a stock of silica which has ever existed in the upper part of the profile.

Although the upper loose horizon has petrographical features indicating that it is mainly the result of a geochemical evolution, and although no trace of sediment deposits was identified by the methods used in this study, it is impossible to completely discard past allochthonous events. Such events would have contributed to the maintenance of the silica stock over the bauxitic horizons. Amongst the possible events, deposition of aeolian dust is one of the more acceptable because of the broad extent of the thick upper kaolinitic horizon in lower Amazonia and because the nature of the deposit allows its progressive incorporation in the upper horizon, by biological processes (bioturbation) as by geochemical processes (dissolution of quartz grains).

To the moment, no data from paleoclimate studies allows to identify such a dust source on the Amazonian or peri-amazonian area. Paleoclimates reconstitution from the tertiary and most of the quaternary are mainly speculative in the Amazonian zone (Irion, 1989). It is assessed that climatic variations in the past 60000 years have lead to local replacement of forest by savannah (Absy et al., 1991), but these data do not permit to conclude about the presence of a dust source, as it presently exists in Australia (Brimhall et al., 1987) or Africa (Orange et al., 1990). Saharan dust, having crossed the Atlantic, is a very likely source in view of recent studies about aerosols in the Amazon basin (Artaxo et al., 1990; Talbot et al., 1990; Swap et al., 1991). It is assessed that Saharan dust reach the Amazonian atmosphere, but estimation from these studies gives a low Si and Al input with regard to chemical erosion.

The stability of the kaolinite in the upper loose horizon is better explained by biological activity. Quantitative measurement of chemical elements in the litterfall of an Amazonian ecosystem indicates that the forest cycles a significant amount of elements, particularly silicon. As a result fluids that percolate through the upper part of the soil already contain dissolved silicon. This process keeps silica from being leached down, and may account for the stability of kaolinite in the upper loose layer (Lucas et al., 1993, 1996).

##### *5. Simulation of the Profile Evolution*

Most of the plateaus from the lower Amazon area, situated in the

same geomorphological and climatic area and developed over the same sedimentary formation, are not bauxitic but covered by thick kaolinitic soils. The determinism of this difference in evolution may be suggested by simulating the profile evolution.

The Alter-do-Chão sediment has little mineral species variability, but the relative abundance of the different minerals may greatly change from a layer to the other: the kaolinite content ranges from 10 to 95%, and the heavy mineral content from 0.2 to more than 5%. The consecutive uncertainty on the parent material composition with regard to a chemically immobile element (as Zr or Ti) turns impossible the determination of strain and translocation of the constitutive elements in the profile by the well-known and widely used mass-balance calculations (Millot et Bonifas, 1955; Leneuf, 1959; Lelong, 1967; Hervieu, 1968; Trescases, 1975; Nahon & Lappartient 1977; Brimhall et al., 1985). It is however possible to calculate, in the hypothesis of an in-situ formation and for various composition of the parent material, minimum values for the thickness of parent material which has been weathered to form the profile. In the hypothesis of weathering conditions identical to current ones, it is possible to calculate the thickness of sediment weathered, the general strain and the time of genesis to form a profile having the same Al/Si ratio than the studied profile. Volume composition of the different horizons of the average present profiles were calculated from bulk densities and ponderal composition. The minimal thickness, which have been weathered to produce the average present profiles, was calculated comparing the Al, Si, Fe and Ti volume contents of the profiles with those of the parent material. The current rates of the geochemical evolution were calculated from the volume and the composition of the waters presently leached out of the system. From these rates, there were calculated the time of genesis, the thickness of parent material and the lowering of the soil surface required for forming, under same weathering conditions, profiles having the same Al/Si ratio than the present profiles. Such calculations do not present any difficulties, they are described with values of parameters in the Appendix.

Calculations were done considering various average compositions of the parent material, from 20 to 70% kaolinite, for an average bauxitic profile and an average kaolinitic profile from non-bauxitic plateaus situated at the north of Manaus. These soils consist of an upper loose kaolinitic yellow clay horizon, a middle horizon with gibbsite and hematite nodules embedded in a kaolinitic matrix, and a lower kaolinitic saprolite (Lucas et al., 1986).

Results are given in Table III.2. The given time of genesis do not

Table III.2 - Rates of evolution of average profiles: results in sublimed characters for an average bauxitic profile, in normal character for an average bauxitic profile.

Mineralogical composition of the protolith	Kaolinite %	20	30	40	50	60	70
	Quartz %	74	64	54	44	34	24
	Others % (Fe <sub>2</sub> O <sub>3</sub> -TiO <sub>2</sub> )	6	6	6	6	6	6
Whatever the conditions	Minimal thickness of sediment weathered (m)	58	40	31	25	22	19
		<u>139</u>	<u>95</u>	<u>73</u>	<u>60</u>	<u>52</u>	<u>46</u>
Under current weathering conditions	Thickness of sediment weathered (m)	169	63	40	30	24	20
		<u>419</u>	<u>157</u>	<u>99</u>	<u>73</u>	<u>59</u>	<u>50</u>
	Lowering of the soil surface (m)	154	48	25	15	9	5
		<u>393</u>	<u>131</u>	<u>73</u>	<u>47</u>	<u>33</u>	<u>24</u>
	Time of genesis (million years)	96	30	15	8	5	2
		<u>243</u>	<u>79</u>	<u>42</u>	<u>26</u>	<u>17</u>	<u>11</u>

pretend to determine the age of bauxitic or kaolinitic soils: weathering conditions may have been greatly different during profile formation, due to climate changes (Berner et al., 1983) or to change in water dynamics in the profile. This approach is a simulation with the aim to compare bauxitic and kaolinitic soils. The rates of the profiles evolution vary greatly, in a 1/3 or 1/4 ratio, as the initial composition of the parent material varies from 30 to 60% kaolinite. The time of genesis and the surface lowering required to form a kaolinitic profile, from a 30 to 40% kaolinite sediment, vary respectively from 15 to 30 m.y. and from 25 to 48 m. These values are close to the values required to form a bauxitic profile from a 50 to 60% kaolinite sediment, which are respectively 17 to 26 m.y. and 33 to 47 m. These results support the hypothesis by which kaolinitic as well as bauxitic plateaus have suffered the same weathering conditions, the difference in evolution being due to an initial difference in kaolinitic content of the parent material. The average kaolinite content, presently observed, is around 51% beneath the Juruti bauxite (analysis of 20 m of sediment), but such value has little meaning, because the present profiles have formed at the expense of a parent material which may have been different.

## Conclusions

The entire profile of the studied Amazonian bauxitic formations is characteristic of an in situ geochemical differentiation. Erosional or depositional events may have occurred in the past, but sedimentological fea-

tures and depositional facies have been completely erased or transformed by a subsequent geochemical evolution: the upper loose kaolinitic horizon ("Belterra Clay") is a pedogenic material.

The succession in time of the different parageneses observed on the profile indicates three main weathering episodes.

(1) The oldest is a ferruginisation of the sediment, giving a lithorelictual ferricrete. Relicts of such a lithorelictual ferricrete were broadly observed in soils from the Amazonian area (Melfi et al., 1988; Tardy et al., 1988; Nahon et al., 1989), suggesting that this episode probably belongs to an old, generalized, intense lateritic episode, which has transformed weakly weathered parent materials.

(2) The second episode is a bauxitisation by precipitation of a macrocrystalline gibbsite facies, which has partly replaced the ferricrete and has formed massive bauxite horizons. Simulation of the profile evolution under current weathering conditions suggests that the differential evolution between bauxitic and non-bauxitic plateaus from the same geomorphologic area is due to an initial difference in kaolinite content of the parent material: bauxitic soils has been formed where the kaolinite content of the sediment was over 40-50%; elsewhere kaolinitic soils has been formed.

(3) The last episode is the current evolution of the profile. A new profile develops at the expense of the old bauxitic profile. On the top of the indurated horizon, intense resilicification is due to the downward geochemical progression of a kaolinitic clay. Gibbsitic and hematitic nodules forming a nodular horizon are relicts left behind as the kaolinitic clay formation front progresses downwards. Beneath the upper kaolinitic clay horizon, a thin gibbsitic horizon is characterized by the precipitation of a microgibbsitic facies. In the middle part of the indurated horizon, dissolution of the bauxite gives large voids which cut the facies structure. At the base of the indurated horizon, moderate resilicification is brought about by the replacement of the macrocrystalline gibbsite by a kaolinitic matrix. Locally, the precipitation of macrogibbsite continues as coatings on large active voids.

The present-day evolution of the profile demonstrates that gibbsite is currently precipitating in the unsaturated zone, beneath horizons where kaolinite is currently precipitating and stable. Consequences lie far beyond bauxite genesis, because such a striking spatial relation was identified in many soil profiles developed in humid tropical areas, in which gibbsite accumulates in the saprolite or as nodules at the limit between the saprolite and a stable upper kaolinitic horizon. This prevalent structure of the profile is in dynamic equilibrium with the climate, a conclusion

opposite to most models of soil genesis. Geochemical modeling of these soils has to take into consideration biological activity in the topsoil, Si and Al migration through the upper kaolinitic horizon, and to identify the chemical state of the mobile Al.

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## Appendix

### Simulation of the Profile Evolution

#### Parameters

Weathering profile:

Lp Thickness

Pa  $\text{Al}_2\text{O}_3$  content/surface unit °

Ps  $\text{SiO}_2$  content/surface unit °

Pf  $\text{Fe}_2\text{O}_3$  content/surface unit °

Pt  $\text{TiO}_2$  content/surface unit °

Calculated from each horizon bulk density and chemical composition

Sediment:

Sa  $\text{Al}_2\text{O}_3$  content/volume unit

Ss  $\text{SiO}_2$  content/volume unit

Sf  $\text{Fe}_2\text{O}_3$  content/volume unit

St  $\text{TiO}_2$  content/volume unit

Water leached out off the system:

Vw Volume of water/surface unit/year

Ca  $\text{Al}_2\text{O}_3$  concentration

Cs  $\text{SiO}_2$  concentration

#### Equations

Local parameters:

Wa = Vw.Ca Weight of  $\text{Al}_2\text{O}_3$  annually leached / surface unit

Ws = Vw.Cs Weight of  $\text{SiO}_2$  annually leached / surface unit

\* Minimal thickness of sediment necessarily weathered to form the present profile, whatever the weathering conditions (Tmin):

$$T_{\min} = \max(Pa/Sa, Ps/Ss, Pf/Sf, Pt/St)$$

\* Under present weathering conditions, rates of evolution to form a profile with the same Al/Si ratio:

- Thickness of annually weathered sediment (Ts):

$(Ts.Sa-Wa)/(Ts.Ss-Ws) = Pa/Ps$ ; then:

$$Ts = (Wa.Ps-Pa.Ws)/(Sa.Ps-Pa.Ss)$$

- Thickness of profile annually formed ( $T_p$ ):

$T_p.Pa=Ts.Sa-Wa$  then  $T_p = (Ts.Sa-Wa)/Pa$ ; and replacing  $T_s$ :

$$T_p = (Wa.Ss-Sa.Ws)/(Sa.Ps-Pa.Ss)$$

- Time for the profile genesis ( $T_{im}$ ):

$$T_{im} = L_p/T_p$$

- Thickness of sediment weathered to form the profile ( $T_{sed}$ ):

$$T_{sed} = T_{im}.T_s$$

- Lowering of the soil surface during the profile genesis ( $L_{ss}$ ):

$$L_{ss} = T_{im}(T_s-T_p)$$

**Parameters used for calculation:**

\* Average bauxitic profile (calculated from 16 profile):

	Thickness (m)	Bulk density	Chemical composition (%)			
			Al <sub>2</sub> O <sub>3</sub>	SiO <sub>2</sub>	Fe <sub>2</sub> O <sub>3</sub>	SiO <sub>2</sub>
Upper loose horizon	7,0	1,4	42,4	46,2	8,3	3,1
Indurated horizon	3,9	1,6	55,9	55,9	23,9	2,8
Lower kaolinitic horizon	15,0	1,4	49,6	49,6	3,6	2,9
Oxyde content (t/m <sup>2</sup> ) (Oxyde weight for a column of 1m <sup>2</sup> section)						
			18,0	14,9	3,0	1,1

\* Average nodular lateritic profile (from Lucas, 1989):

Oxide content: 7,6 t/m<sup>2</sup> Al<sub>2</sub>O<sub>3</sub>; 11,1 t/m<sup>2</sup> SiO<sub>2</sub>; thickness of the profile 15m.

\* Composition of the sediment:

SiO<sub>2</sub> and Al<sub>2</sub>O<sub>3</sub> vary with the kaolinitic content of the sediment.

Average Fe<sub>2</sub>O<sub>3</sub> content: 4,3%

Average TiO<sub>2</sub> content: 1,8%

