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Regolith Exploration Geochemistry in Tropical and Subtropical Terrains

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Chapter I.4

SOIL FORMATION IN TROPICALLY WEATHERED TERRAINS

Y. LUCAS and A. CHAUVEL

INTRODUCTION

Soils are one of the principal sample media available for geochemical exploration and the physical and chemical processes involved in their formation are the same as those responsible for geochemical dispersion. Soils thus represent tangible evidence of these processes, hence an understanding of soil genesis is essential for the correct interpretation of geochemical exploration data.

Soil formation (pedogenesis) is the result of the transformation of bedrock, pre-existing saprolite or other regolith material at the interface with the atmosphere. The nature of the transformation depends on the balance between three main processes:

(1) alteration (weathering) of minerals;
(2) transport in solution or as solids;
(3) authigenesis of stable secondary minerals.

These processes result from the new thermodynamic equilibria controlled by the changed physico-chemical conditions that exist in the soil profile compared to the parent material. They act first at the base of profiles and lead to the formation of materials that have a new composition encompassing both residual and newly generated minerals and a new fabric determined by the spatial relationships of the secondary structures.

As the soil deepens, new physico-chemical conditions are imposed on the profile. Consequently, the processes of alteration, transportation and authigenesis interact with the initial soil to form another with a different composition and fabric. This pedogenetic differentiation produces layers of soil materials generally distributed as horizons that reflect the spatial variation of the soil environment. The horizons are the result of an equilibrium between soil composition, soil fabric and the physico-chemical conditions at each depth in the profile.

This chapter refers to the genesis of soil covers and their distribution within landscapes. As discussed in Chapters I.1 and I.3, the regolith is the product of processes that may have been active for millions of years: consequently, the pedogenetic processes that are active at present may be transforming a relic soil. In this chapter, a simplified model of soil development in tropical terrain is presented, followed by a discussion of the effects of climatic and other changes on the processes of pedogenesis.

DYNAMIC EQUILIBRIA IN SOILS

Physico-chemical conditions are generally more aggressive towards mineral components in the upper parts of a profile, principally because of the presence of biochemical compounds. Such compounds facilitate alteration by the action of water-soluble acids produced either directly by microorganisms or from the decomposition of organic matter. They also promote leaching of components released during weathering by the formation of soluble complexes. As pedogenesis proceeds, complete dissolution of some minerals in the upper horizons of the profile causes subsidence of the soil surface. This decrease in the profile thickness is compensated by weathering of further parent material at the base of the profile. The profile can thus be considered to represent a time sequence (Butt and Nickel, 1981), with each horizon having been derived from material similar to that now underlying it by the downward advance of a *transformation front* in which the processes of alteration, transport, authigenesis and fabric acquisition occur.

The differentiation and the evolution of a profile depend upon:

(1) The relative rates of the advance of the different transformation fronts (Fig. I.4-1). If the different horizons of a profile are maintained during profile development, they may be considered to be genetically continuous and the profile to be in *dynamic equilibrium*. However, if the fronts advance at different rates, and the vertical succession of horizons changes, then the horizons are genetically discontinuous and the profile is in *disequilibrium* (Millot, 1983).

(2) The geochemical balance of elements transported into or out of each horizon. At each front, elements may be mobile either in solution or as fine solids such as clays. Soluble components, released by weathering, may precipitate immediately as new minerals, or be transported to another horizon where physico-chemical conditions are more favourable for precipitation, or be leached from the profile entirely. Solid particles are released by the destabilization of soil fabric as fronts advance and may be transported by mechanical eluviation into underlying horizons (see p. 99).

FACTORS AND PROCESSES OF SOIL FORMATION IN THE TROPICAL ZONE

The nature and activity of pedogenetic processes depend on external factors such as climate, parent material and time, and on internal factors interdependent with the evolution of the soil such as topography, the biosphere and the hydrodynamics of the profile. In this section, the general processes of soil formation are described for the different climates of the tropical zone, then the specific effects of the others factors are discussed.

Climate

The climate of tropical environments and its influence on geochemical dispersion, including weathering, soil formation and landscape development, are con-

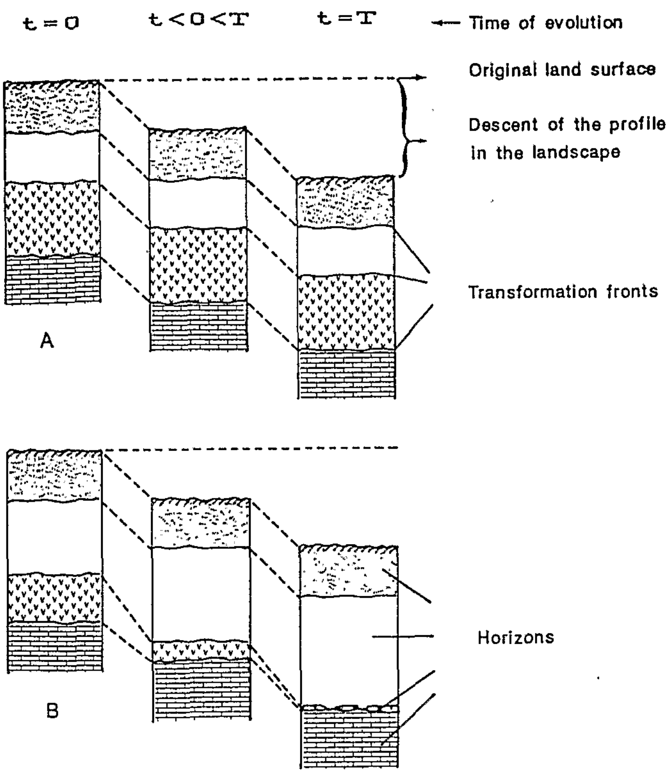

Fig. I.4-1. Dynamic equilibrium of a soil profile. (A) Profile in dynamic equilibrium. (B) Profile in disequilibrium.

sidered in Chapter I.1. The most important climatic elements are the distribution and annual averages of precipitation and temperature, summarized on the Köppen climatic map (Fig. I.1-1). These control, respectively, the amount of water percolating through the regolith in a given time, and consequently the concentration of the weathering solutions, and the rates of chemical reactions. Thus, the latitudinal variation of climatic conditions in the tropical zones north and south of the equator induces changes in the processes of soil formation.

In *humid equatorial climates* (rainfall > 2000 mm p.a., with a poorly defined dry season), the soil moisture regime is permanently humid and free draining. Soil horizons are characterized by fabrics having a good hydraulic conductivity, so that water-logging or gleying processes are only rarely developed in the landscape. The amount of water percolating through the regolith is considerable (more than 600 mm p.a.) and weathering reaches great depths.

Under these conditions, intense alteration at the weathering front at the base of the regolith affects all the weatherable minerals simultaneously. Even quartz, usually considered a stable mineral, can be highly corroded. Transport of mobilized elements is significant and includes the loss of all the bases and some of the silica. Authigenic minerals are abundant and form most of the soil material; they include kaolinite (monosiallitization (Pedro, 1966); see Chapter I.2), oxyhydroxides of aluminium (allitization) and iron. These processes lead to the formation of saprolite in which bedrock fabrics are preserved as pseudomorphs by secondary minerals. For example, gibbsite or goethite may form septa along boundaries and cleavages of primary minerals before and during their dissolution, ultimately replacing them (Delvigne et al., 1979; Embrecht and Stoops, 1982). The saprolite is characterized by relative accumulation of Fe and Al, although absolute accumulation due to transport and deposition of material from the upper horizons may be significant (Bocquier et al., 1984). For example, granitic saprolite in the Ivory Coast may register a 250% increase in the Fe content (Boulangé, 1984).

As discussed above, the profile may be considered to be developed by the advance of different transformation fronts, named for the principal process that occurs. The *pedoplasmation front* corresponds to the loss of much of the lithic fabric in the upper part of the profile. It is often associated with a *cementation front*, which corresponds to the accumulation of oxyhydroxides of iron or aluminium mainly derived by leaching from the upper horizons (Bocquier et al., 1983; Lucas et al., 1986). Iron, mainly as goethite and hematite, and aluminium, mainly as gibbsite, accumulate in pores in the saprolite, giving cemented patches that preserve the lithic fabric as relics in more or less hardened nodules or cuirasse (Eswaran et al., 1977; Chauvel et al., 1983; Bocquier et al., 1984; Didier et al., 1985). Outside these patches, solution and authigenesis of minerals and mechanical processes (such as shrink and swell) in the clay fraction cause the lithic fabric to be replaced by a new, homogeneous, plasmic fabric (pedoplasmation, Flach et al., 1968). This new matrix is also characterized by the establishment and maintenance of ferric hydrate-clay bonds, and by high-chroma colours. However, oxyhydroxide nodules may also form in this kaolinitic matrix, so that relicts of both lithic and pedoplasmic fabrics may coexist in the same horizon. In summary, the processes at the secondary transformation fronts may result in generally complex successions of mineral parageneses and facies and lead to the formation of horizons ranging from a soft kaolinite matrix without nodules to hardened cuirasse.

The *microaggregation front* corresponds to the appearance of a microaggregated fabric in the kaolinitic clay matrix. This fabric is produced by inorganic processes in an acid ferro-kaolinitic environment that is freely drained and constantly humid (Pedro et al., 1976; Chauvel, 1977). Where microaggregation occurs, the cemented facies start to dissolve and either disappear or remain as rounded nodules or blocks. These coarse elements, together with resistant quartz pebbles, can form a residual stone line at the base of the microaggregated horizons (see Chapter III.2).

At the top of the profile, an increase in concentration of resistant minerals, particularly quartz, corresponds to a *surface degradation front*. Under certain conditions, for example over quartz-rich parent materials, the hydrodynamics are such that secondary podzolization processes occur, resulting in the loss of leached elements from the soil and an impoverishment in clay (Righi and Chauvel, 1987).

In the *wet savannas* (rainfall 1200–2000 mm p.a., dry season 3–6 months), alteration at the weathering front is less intense than in humid equatorial climates, so that some weatherable primary minerals can remain in the soil and quartz is only weakly corroded. Loss of material from the profiles is less significant and some of the alkaline cations are retained. Leaching of silica is only minor, and gibbsite is less stable than clay minerals; the authigenic minerals in the saprolite are dominantly kaolinite and goethite. Water-logging and gleying processes are more common in lower parts of the landscape.

The seasonal climate is responsible for contrasts in the soil moisture regime. The *pedoplasmation* and *cementation fronts*, in the upper part of the saprolite, are in a moist environment that is alternately saturated and unsaturated due to fluctuations of the water-table. The corresponding alternation in conditions from reducing to oxidizing allow the reduction and migration of iron to form enriched and depleted zones, giving a pseudo-gley facies (Vizier, 1983) or a mottled clay facies (Nahon, 1986). The loss of iron causes destabilization and eluviation of clay particles from the depleted zone, so that the horizon as a whole is relatively enriched in ferruginous components in the form of nodules or massive accumulations such as cuirasse. These processes, coupled with only relatively minor generation of gibbsite, result in the formation of major accumulations of Fe oxides in the form of laterites (Chapter I.3).

The upper soil horizons are subjected to desiccation in the dry season, corresponding to an *ultra-desiccation front* (Chauvel, 1977). One consequence is that alternately humid and dry conditions affect even the very fine pores (voids less than 0.2 μm), so that the microfabric of the soil particles becomes unstable. In addition, under dry conditions, residual water molecules in thin films on the surfaces of soil particles are protonated, giving local strong acidity (Mortland, 1968; Pedro and Melfi, 1983). Clays and iron oxyhydroxides become dissociated (Brinkman, 1970; Chaussidon and Pedro, 1979) and the clays may even be dissolved. These processes promote the mechanical eluviation of clays which, together with dissolution, results in the formation of surface soil horizons impoverished in clay.

In *dry savannas and semiarid to arid regions* (rainfall less than 1200 mm p.a., dry season longer than 6 months), erosional processes increase, but weathering and pedogenetic loss of components from the profile are minor. The weathering solutions concentrate alkaline cations, the saprolite is thin and smectites are the dominant authigenic clay minerals. Because of the low rainfall, there is only weak redistribution of iron into nodules (Boulet, 1974). The fabric of the pedoplasmic material reflects the swelling and shrinking properties of the smectites under alternately humid and dry conditions, such as in vertisols. The upper horizons are subjected to strong contrasts in moisture content during the year, with saturation prevailing in the rainy season due to water-tables perched on the compact smectitic horizons, but with ultra-desiccation in the dry season. Clay impoverishment due to eluviation and dissolution leads to the formation of sandy surface horizons, especially over coarse textured lithologies. However, this process is weak when rainfall is < 700 mm p.a. Under sufficiently arid conditions (< 600 mm p.a.), the presence of weathering solutions rich in calcium and bicarbonate ions causes, in wet periods, the dissolution of quartz and clay minerals and, in the dry periods, the precipitation of calcite at a Ca-cementation front in a variety of forms of calcretes, ranging from coatings and nodules to duricrusts (Nahon, 1976; Millot et al., 1977; Halitim et al., 1983, Milnes and Hutton, 1983). The formation and exploration significance of calcretes are considered in Chapter III.3. Gypsum and other salts (Na, Mg, Cl) may also accumulate in the profile or as superficial crusts.

In summary, changes evident in the climatogenic vertical soil profile between humid equatorial, seasonally humid and arid climates are as follows:

(1) the *saprolite* becomes thinner, with the disappearance of gibbsite and the replacement of kaolinite with smectite;

(2) in the *oxyhydroxide* accumulation layer, the Fe-Al ratio increases. In sufficiently arid climates, the oxyhydroxides do not accumulate and this cemented facies disappears; however, in semiarid regions, carbonate and salt accumulations appear;

(3) *microaggregation* disappears and the upper horizons are characterized by more massive microfabrics. The amount of smectite increases progressively, giving polyhedric fabrics. Clay impoverishment in the topsoil increases, thus producing sandy horizons where quartz is abundant.

(4) physical *erosional* and *depositional* processes increase.

As described in Chapter I.1, at any latitude, high relief gives vertical climatic, vegetational and soil zoning. The increase in precipitation and the diminution of evapotranspiration result in continuously humid soil regimes, more intense leaching and accumulation of organic matter. On volcanic ash, which occurs frequently in high relief tropics, allophane accumulates, giving andosols (Tan, 1984). In the higher level of this vertical zonation, podzolization is the main pedogenetic process (Zebrowski, 1975).

Topography

Topography is both an internal and external factor in pedogenesis, either influencing or being a consequence of soil evolution. Different geochemical conditions exist upslope or downslope in the same landform unit, depending upon the influence of the topography on the drainage and the hydrology of the soil cover. For example, some Si-cementation fronts may be related to certain low topographic situations in which silica transported for several kilometres has been deposited (Milnes and Twidale, 1983). The processes of lateral surface movement of particles by water or wind action, and by mass flow on slopes, which range from thin reworking of the topsoil to continental sedimentological processes, also depend mainly on topography and climate (see Chapter I.6). Under the dense forests of humid climates, physical erosion is less significant than leaching, even on steep slopes, but in drier areas, the steepness of slopes has a considerable bearing on the processes of erosion. Thus, an instability due to tectonism or climatic change may induce erosion of an initial regolith, so that newly developing soils will form from different parent materials according to the depth of truncation in different parts of the landscape (Stace et al., 1968). In semiarid Niger (rainfall 400–600 mm p.a.), Gavaud (1977) described an initial regolith that consists of a 30–40 m thick smectitic saprolite overlain by 40–70 m thick kaolinitic and cuirasse horizons. As a result of the truncation of this initial cover, newly formed soils have developed upslope on ferruginous and kaolinitic materials and downslope on smectitic materials. The truncated profiles may also be overlapped by new sediments, giving complex polygenetic profiles (Milnes et al., 1985). Such variations in soil parent materials have a profound effect on the surface geochemical expression of mineralization and is discussed in Part III of this volume. The interdependence of topography and soil evolution is considered further below.

Soil parent material

The foregoing description of variations in pedogenetic processes due to climate refer particularly to the weathering of common crystalline rocks such as granite, but other variations can be due to differences in lithology. Even minor differences may be significant, especially under intermediate climates. The nature and abundance of resistant minerals that become relatively enriched may influence soil fabric, drainage conditions and geochemical processes. For example, soils on a coarse-grained granite may develop a less strongly cemented facies than soils on fine-grained granite, the difference being attributed to freer drainage conditions on the former due to large residual quartz grains. The chemical composition of the bedrock can control the composition of the weathering solutions and hence of the authigenic minerals. The effect is greater when the amount of percolating water is low. On Al-poor rocks, such as dunites (and limestones), silica released at depth by weathering of primary minerals, or transferred from the degradation of topsoil, precipitates as quartz, chalcedony or opal in the absence of aluminium (Lelong et al., 1976; Nahon, 1976; Nahon et al., 1979; Leprun, 1979; Butt and Nickel, 1981). Weathering of Fe-rich parent material such as basic rocks forms, under sufficiently oxidizing conditions, a skeletal framework of Fe oxides in the saprolite that is free draining and thus prone to more intense eluviation and oxidation at depth compared to saprolite formed on acid rocks. As an example, in Niger, the saprolite on granite is 30–40 m thick and overlain by thick kaolinitic horizons, whereas on Birrimian schists the saprolite is 150–200 m thick and immediately overlain by a cuirasse (Gavaud, 1977).

Biotic factors

Biotic factors, such as vegetation, microorganisms or animals are both internal and external factors in pedogenesis, or can be a consequence of soil evolution. Organisms, whose existence and activity depend on the prevailing environment, are strongly involved in soil formation (Dommergues and Mangenot, 1970; Alexander, 1977). The presence of organic matter and the processes of biotic reworking affect the soil fabric and hence influence the soil moisture regime. Organic acids or chelating compounds produced during the decomposition of organic matter or as metabolic products of soil biota may be active in the transport of the released elements (Huang, 1986). Vegetation influences the type of soil organic matter and its presence or absence determines the effect of physical erosion processes.

Time

Time is an important factor in soil development in tropical regions, where weathering mantles have been evolving since the Tertiary or earlier. The time required to form a profile with an Fe oxide cemented horizon has been estimated by Nahon and Lappartient (1977) to be a few million years and that for intense bauxite formation to be from 1.4 to over 20 million years (Bardossy, 1983) (see Chapter I.3). Soil formation has thus operated over very long time intervals, and any changes in the factors involved will influence the nature of the product.

The *external factors* of soil formation may have changed. Various authors have estimated the rate of descent of the weathering front to be between 30 and 290 cm per 100,000 years (Leneuf, 1959; Hervieu, 1968; Gac and Pinta, 1973; Trescases, 1975; Boulangé, 1984; see Table I.3-2). These rates are high compared to the probable length of soil evolution in much of the tropics. Accordingly, *the soil may have descended through several lithologies*. Thus, some soil characteristics (such as trace element or heavy mineral abundances or assemblages) may not be related to the underlying saprolite, but be derived from a different parent material which has since disappeared. Climate and the chemical characteristics of the atmosphere have changed throughout pedological (and geological) time (Rognon, 1976; Frakes, 1979; Tardy, 1986) resulting in different weathering conditions. For example, Berner et al. (1983) suggest that the average temperature and CO_2 content of the atmosphere have declined from 26 to 15°C and from 2.5 to 0.055 10^{18} mole respectively during the last 100 million years. There are many examples of the effects of such changes on soil evolution. Thus, in Burkina Faso, Leprun (1979) demonstrated the replacement of a kaolinitic weathering front by a smectitic front, and in the Ivory Coast, Boulangé (1984) described the replacement of a gibbsitic front by a kaolinitic front. In each case, the successions are related to a change to less humid climates. In French Guiana, Boulet (1978) related the presence of a leaching front at depth in a lateritic profile to lowering of the water-table due to tectonic uplift. Numerous examples in Australia show successive weathering and continental sedimentological processes interacting with soil formation (Mabbutt, 1980; Butt and Nickel, 1981; Milnes and Hutton, 1983; Milnes et al., 1985; Butt, 1985). All these examples indicate that the present soil represents an integration of successive variations of the external factors, as discussed later in the chapter.

The evolution of the soil cover can itself induce changes in the *internal factors of soil formation*. Landscape evolution due to pedogenesis may alter the conditions of groundwater movement. The development of horizons with different fabrics involves changes in permeability and in the rates at which minerals dissolve and are formed. These variations in the factors affecting pedogenesis induce disequilibrium in the soil; this in turn is expressed as transformation fronts that result in alteration to a profile form in equilibrium with the prevailing conditions. These fronts progress vertically, as described above, or laterally through the landscape, giving three-dimensional transformation systems (Boulet et al., 1984). The evolution of such systems is discussed and illustrated in the following section.

TRANSFORMATION SYSTEMS IN THE TROPICAL ZONE

Transformation systems represent the replacement of an initial regolith with one that is different in its mineral composition or fabric. The initial soils may either be locally preserved, or have disappeared after complete transformation. The dynamics of the systems depend on the causative factors, the location at which the transformation begins and the mode in which the fronts advance. Commencement may depend on external factors such as climatic change, base level change, or on internal factors related to the evolution of the soil itself. In both cases, all factors that control the movement of water in the soil cover interfere. When initiated, the advance of the transformation is facilitated by its own effects, as a feedback mechanism. For example, in eluvial–illuvial systems, the leaching front creates a porous soil fabric which favours transportation of material and hence further leaching; conversely, an accumulation front creates an impervious fabric which favours deposition and mineral authigenesis.

The apparent lateral advance of a transformation front (Fig. I.4-2) is the result of (a) vertical subsidence of the soil profile, and (b) lateral displacement of the transformation front. Vertical subsidence is more rapid in humid environments compared to dry environments, so that the soil material is partly renewed by advancement of the weathering front as the transformation proceeds.

The age and rates of development of systems can only be estimated indirectly. The age of the climatic change gives an upper limit to that of related transformation systems. In humid areas, assuming an average rate of descent of 1 m per 100,000 years, weathering may have advanced by 350 m since the mid-Tertiary, hence the transformation systems on the Precambrian shields may have been

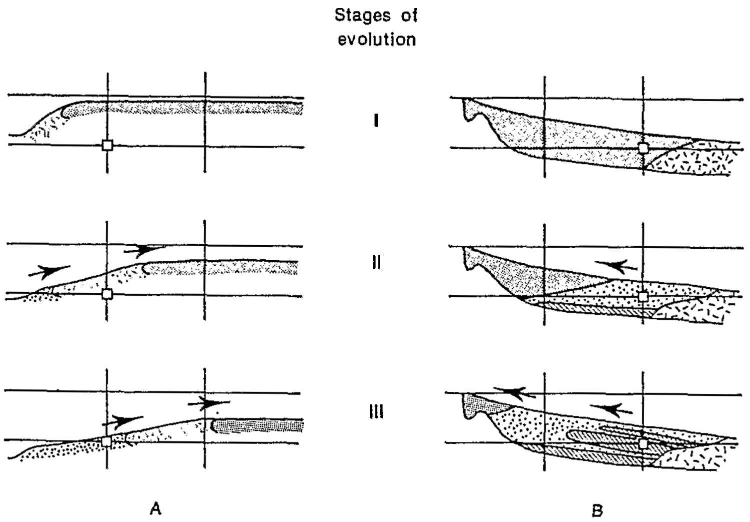

Fig. I.4-2. Dynamics of transformation systems: apparent lateral displacement of the transformation fronts. Arrows give direction of the displacement: (A) Vertical subsidence rapid compared to lateral advance of transformation fronts; (B) Vertical subsidence negligible compared to lateral advance of transformation fronts.

evolving throughout this period. However, systems developed on Pleistocene sediments are obviously much younger.

Transformation systems under humid tropical climates

All transformation systems so far recognized in humid areas demonstrate massive leaching of an initial lateritic cover, with the leached material lost to the drainage network.

Systems due to internally imposed changes in drainage conditions

In central Amazonia, the lateritic cover developed on Tertiary sandy-loam sediments is being replaced by podzols (Lucas et al., 1984, 1987). The present annual rainfall is about 2000–2500 mm, with a weak dry season. The initial soil on the plateaux is a thick clay having a microaggregated horizon consisting of more than 90% kaolinite overlying a nodular horizon consisting of a kaolinitic matrix with gibbsite and hematite nodules. On the slopes, the soils are more sandy, with giant podzols developed on the lower part of long slopes. There is a range of intermediate types between the clay-rich lateritic soils of the plateaux and the sandy podzols downslope, and most pedological features vary progressively between these two extremes. This soil differentiation is explained as follows (Fig. I.4-3). On the plateaux, elements released by the surface degradation front are transported deep into the profile. Aluminium is trapped as kaolinite

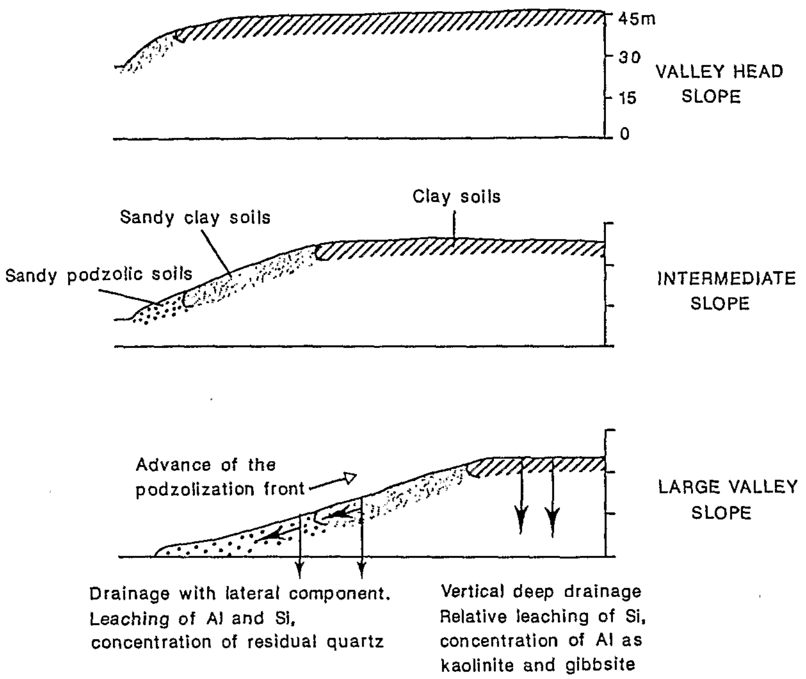

Fig. I.4-3. Evolution of soil system on Barreiras sediments, central Amazonia.

and gibbsite, and there is a net loss of SiO_2. This results in the formation of kaolinitic clay horizons as the soil descends through the sediment. On the slopes, lateral waterflow leaches more Al from the system so that, firstly, more material is removed than from the plateaux, emphasizing the slope and, secondly, there is a net loss of Al, giving a relative enrichment of SiO_2 as quartz. The soils are thus sandier, and podzolisation occurs on the lower slopes. As the soil descends through the sediment, these differences in vertical evolution result in the lateral advance of a leaching and podzolizing front that progresses upslope as the slope develops.

The cause of this transformation is internal, related to the flow percolating water. The transformation starts at the edges of the plateau landforms and advances towards the centre, contributing to the development of slopes.

In French Guiana, kaolinitic soils have developed over Archaean migmatites (Boulet et al., 1984). The present annual rainfall is 2500–3000 mm, with a weak dry season. The initial soil consists of the vertical superposition of a thick saprolite and an upper microaggregated sandy-clay horizon, 3–8 m thick. A thin quartz pebble stone line is commonly present between these two horizons. The quartz content increases progressively in the top metre. This cover has resulted from the partial solution of quartz and the relative accumulation of kaolinite as

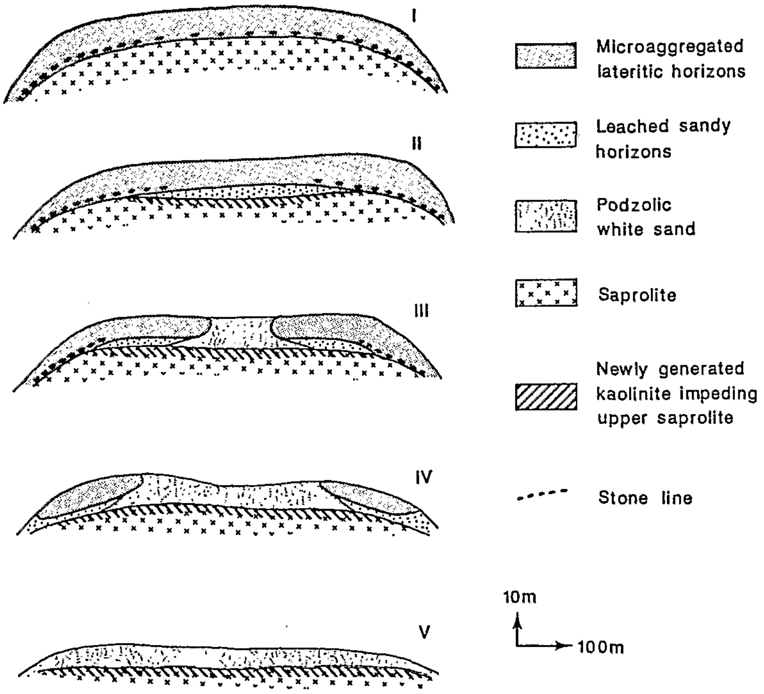

Fig. I.4-4. Evolution of the laterite–podzol system developed on migmatites, French Guiana.

the soil formed from the saprolite (Fig. I.4-4). As in the previous example, this initial cover is now being transformed into giant podzols, and the transition exhibits the same features. However, here the podzols develop at the centres of the plateaux and advance towards their margins; the kaolinitic soils, including nodules with lithic fabrics, remain on the slopes. The difference is due to the movement of percolating water, which forms a deep drainage network perched on the upper part of the saprolite, the permeability of which is impeded by newly-formed kaolinite. This transformation begins at the centre of the plateaux, where the surficial drainage is minor and the groundwater is closest to the top of the saprolite. The deep drainage allows lateral leaching of some Al from the system, so that soils become sandy and the landforms evolve towards plateaux having central depressions occupied by podzols. The more mature plateaux are entirely podzolized.

In these examples, podzolization is a secondary process that occurs after transformation has caused the initial clay or sandy-clay soil to become sandier. However, if the initial soil is already a sandy loam, podzolization may itself cause the transformation. This has been observed on Quaternary sediments in French Guiana (Fig. I.4-5), where a landscape of low flat hills with an initial cover of lateritic sandy loam, 1–3 m thick, has been transformed into a podzol with a

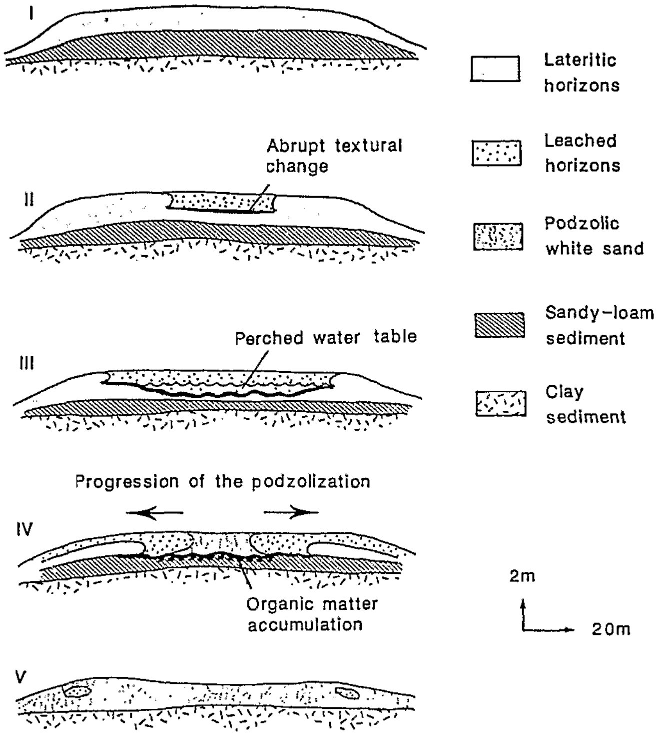

Fig. I.4-5. Evolution of the laterite–podzol system developed on coastal sediments, French Guiana. The size of landscape units is small compared to the system of Fig. I.4-3.

50–150 cm thick white sand horizon overlying a humic iron pan (Turenne, 1977; Boulet et al., 1984). In the centre of the landscape units, the external drainage is minor and the water-table is close to the surface in the rainy season. The process commences by topsoil degradation, the development of an abrupt textural change and the establishment of a temporary, shallow, perched water-table. Thereafter, alternately saturated and unsaturated conditions favour leaching of kaolinite, so that podzolization progresses both vertically and laterally. Settling of the soil material causes slight subsidence of the podzolized zones.

Systems due to external changes

Transformation may be initiated by a change in drainage conditions caused by external factors. Systems of this type are developed on Precambrian schists in French Guiana, where the water-tables have become lower following tectonic uplift during Quaternary (Boulet, 1978; Fritsch, 1984). The present climate is characterized by an annual rainfall of 2000–3000 mm and a short dry season. The landscape consists of rounded hills having an initial soil cover consisting of:

(1) a microaggregated soil horizon, 2–4 m thick;

(2) a red pedoplasmic clay horizon, 1–2 m thick, with a few ferruginous nodules with lithic fabrics;

(3) saprolite with relic muscovites.

Water percolates vertically to the water-table deep in the saprolite. The soil cover is being modified by an ascending transformation which commenced downslope (Fig. I.4-6). Morphologically, the microaggregated layer becomes thinner and is replaced by a sandy-loam horizon. At the same time, the fabric of the upper part of the pedoplasmic horizon changes and becomes nearly impervious, so that lateral waterflow occurs in the surficial sandy-loam horizon. Soil creep and mechanical erosion are almost negligible. The transformation front thus corresponds to the development of a horizon permitting lateral drainage. The large volume of water percolating along the base of this horizon increases the rate at which clay dissolves and hence the quantity of soluble material lost from the system. This transformation front advances both upslope and to depth, at rates greater than that of the descent of the weathering front and is thus discordant across the initial vertical sequence of horizons, and the mean slope increases with its advance.

Cemented facies, in the form of ferruginous or gibbsitic nodules, are common in soils of the humid tropics. They are present in all stages of their genesis throughout the regolith, so that they mostly seem to be related to active cementation processes. However, in massive cemented facies such as the bauxitic formations of central Amazonia, complex successions of Fe and Al facies indicate that significant changes in weathering processes have taken place during their development (Grubb, 1979; Lucas et al., 1988). Under present conditions, they are being dissolved and replaced by a kaolinitic plasma.

Transformation systems under seasonal (savanna) climates

Eluvial–illuvial systems

In savanna climates, most of the known transformations are eluvial–illuvial systems, in which some elements are leached from the upper parts of landform units and transferred to the lower parts. The soils and upper horizons of upslope units tend to be acid and rich in kaolinite, iron oxyhydroxides and residual quartz sand, whereas deeper horizons and soils of downslope units tend to be rich in 2/1 clays and alkaline cations. As the transformation progresses, the latter become more abundant and extend upslope at the expense of the original kaolinite. Most of these systems, in Africa, South America and Australia, are attributed to the transformation of a thick, relic, lateritic cover which, because of a drier climate, is no longer in equilibrium with present pedogenetic processes (Fig. I.4-7). This is particularly obvious in the dry savannas (annual rainfall 600–1000 mm). In wet savannas (rainfall > 1000 mm p.a.), the distinction between relic soil materials and those formed under present conditions is less evident (Brabant and Gavaud, 1985). The present pedoclimate upslope is open, well drained and in equilibrium with lateritic materials, whereas downslope it is

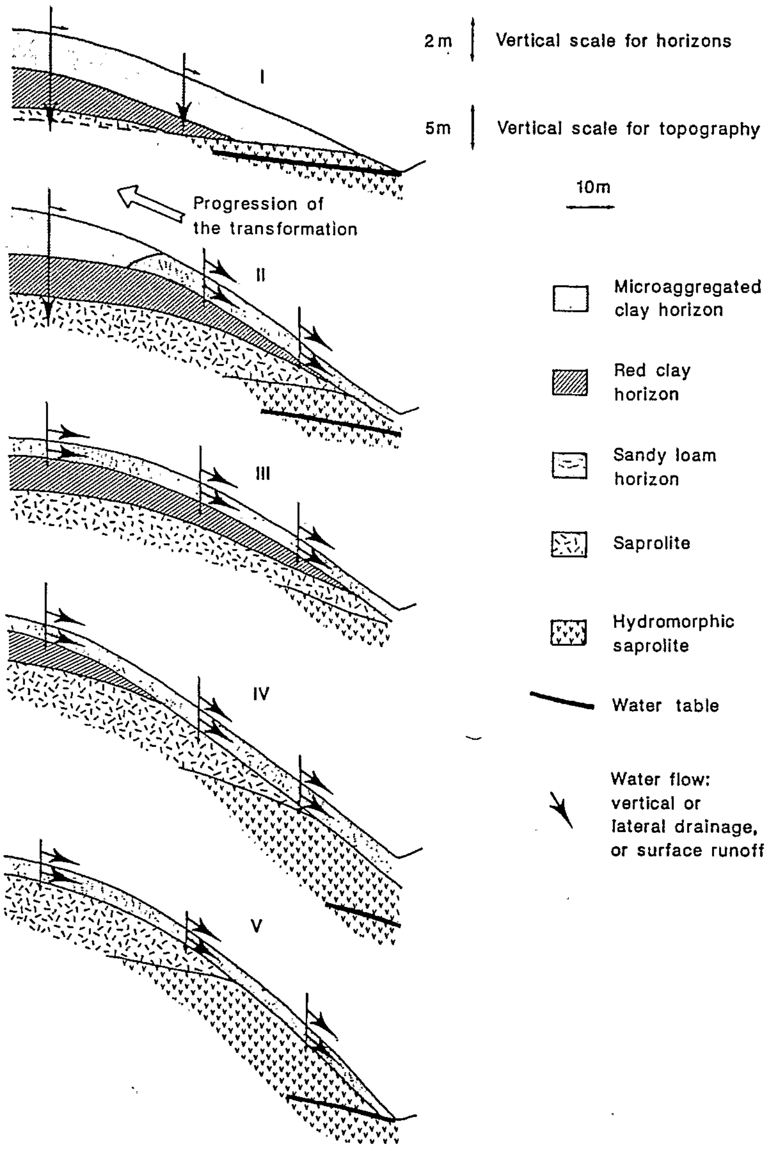

Fig. I.4-6. Evolution of a soil system developed on schists, French Guiana. Derived from Boulet (1978).

more confined and enriched with elements leached from upslope. The soils that are formed are thus organized as catenas, that is a sequence of soils.

Systems related to a climate change are developed at the base of granitic inselbergs in southern Burkina Faso (Fig. I.4-8) where the annual rainfall is

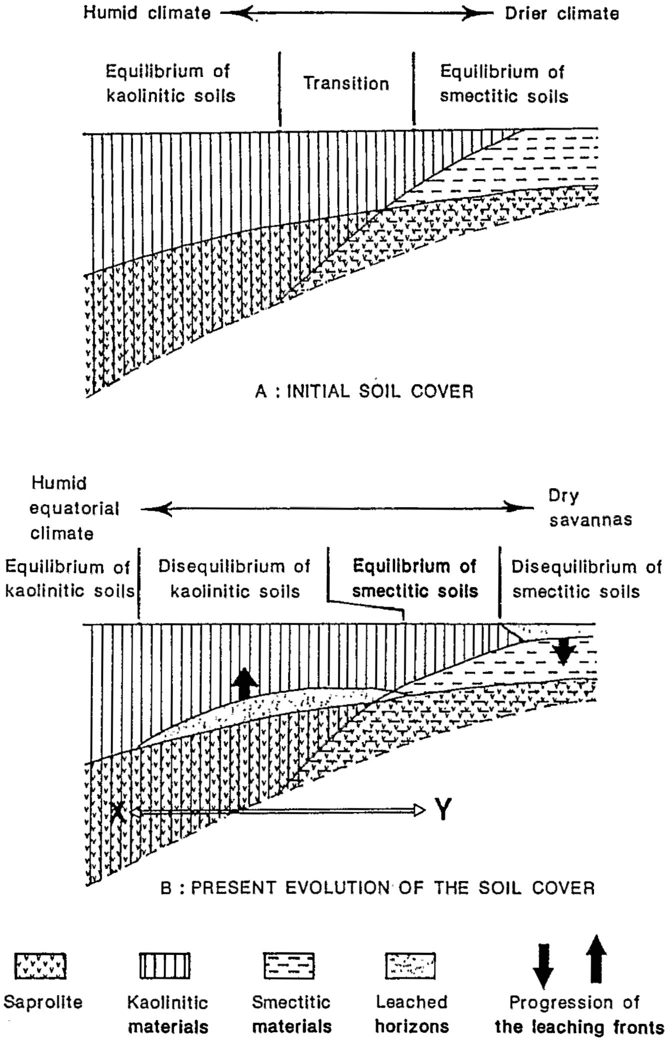

Fig. I.4-7. Disequilibrium in soil cover south of Sahara. Derived from R. Boulet, unpublished. The transformations indicated by the arrow X–Y are illustrated in Fig. I.4-8.

900 mm with a long dry season (Boulet, 1974). In the upper part of the catena, kaolinitic horizons are developed from the pre-existing saprolite, which represents the initial cover. At the centre of the catena, there is a sandy soil with features characteristic of leaching processes, such as relic kaolinitic material. This indicates that a leaching front is progressing upslope at the expense of the kaolinitic cover. The lower part of the catena has a smectitic clay soil with features characteristic of illuvial processes. Relics of leaching features at the

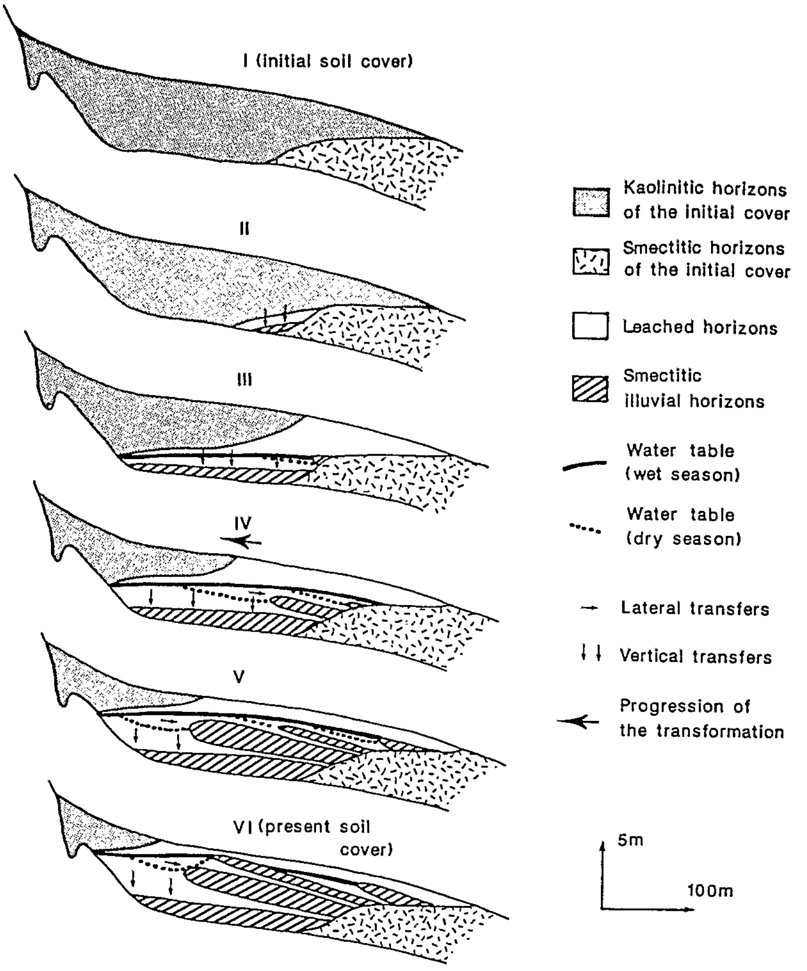

Fig. I.4-8. Soil system in Burkina Faso. Derived from Boulet (1974).

transition with the sandy soil indicate that an illuviation front is progressing upslope at the expense of the sandy soils. At the leaching front, the contrasting soil moisture regime due to the present seasonal climate facilitates the destabilization of clay particles. The released elements are transported as solids and in solution and accumulate downslope in the zone of water-table fluctuation, where the generation of massive smectite units occur. In this particular example, the presence of an inselberg increases the water input and hence the intensity of the eluvial processes during the rainy season. Similar transformation systems are present where there are no inselbergs, but where coarse granites have given rise to soils with a porous, sandy texture that facilitates water percolation.

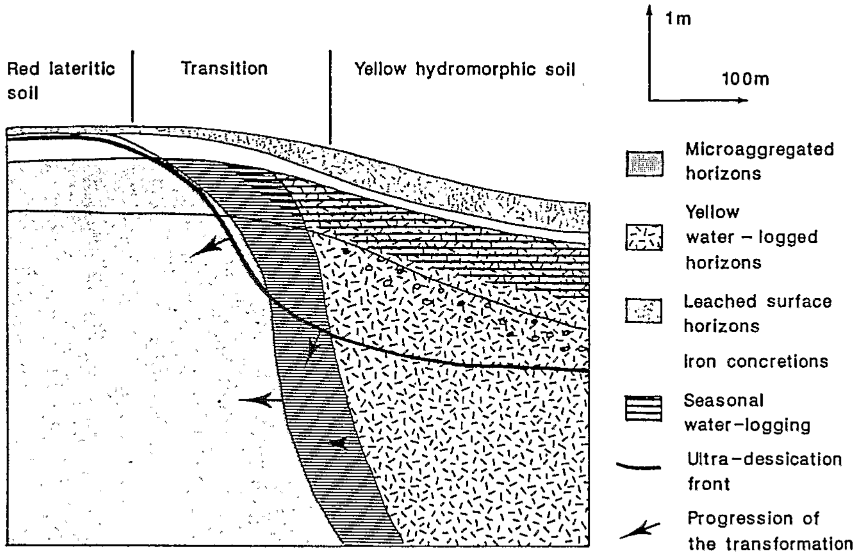

Fig. I.4-9. Sketch of the transition between red microaggregated soil and yellow hydromorphic soil, Casamance, Senegal. Derived from Chauvel (1977).

A progressive transition from systems with abundant kaolinitic soils in upper landscape units in the highlands to more transformed systems in the lowlands in which this cover has disappeared or remains only as relics have been recognized in north Cameroun (Brabant and Gavaud, 1985). Similar systems, in which the initial kaolinitic domain has already disappeared, have been described from Tchad (Bocquier, 1971).

The transformation and disintegration of cuirasse by leaching under the present climate and their replacement by kaolinitic horizons have been described from Senegal and Burkina Faso by Nahon (1976) and Leprun (1977, 1979). As the different transformation fronts have progressed upslope, Fe has been leached from both the base and the top of the crust and transported in solution to form secondary ferricretes downslope. Such processes are closely linked to the morphogenesis of the landscape.

Related systems

Changes to drier savanna climates may initiate the transformation of only the fabric or mineralogy of soil, without transport of material. Soil fabric changes are reported from Casamance, Senegal (Chauvel, 1977) where the annual rainfall ranges between 800 and 1200 mm, with a long dry season (Fig. I.4-9). The landscape is one of low plateaux having an initial lateritic cover developed on Tertiary sandstones. The profile consists of saprolite, a loamy sand horizon with soft ferruginous nodules and a red microaggregated horizon 2 m thick. The latter remains humid during the dry season because of the "mulching" effect of the microaggregated fabric. However, with the advance of an ultra-desiccation front, the red microaggregated horizon is being replaced by a yellow horizon with a compact microfabric. This horizon has a contrasted moisture regime, being highly desiccated in the dry season and temporarily saturated in the rainy season. The initiation of this transformation is related to the onset of drier climatic conditions over the last 20,000 years, during which time the microaggregated materials have been out of equilibrium with the prevailing climate. The process commenced in the topsoil with the disintegration of the microaggregates by hydric constraints in the dry season. The transformation is more intense in the centres of the plateaux because shrinkage of the soil material causes slight subsidence. Temporary water saturation of the surface horizons in the wet season accelerates the collapse of microaggregates and the consequent modification of porosity permits even greater desiccation in the dry season.

The temporary waterlogging also results in the replacement of hematite by goethite and hence the soil colour changes from red to yellow. The ultra-desiccation front progresses both downwards and laterally and the compact yellow soil thus progressively replaces the red soil. This explains the abundance of yellow soils in the landscape. Red microaggregated soils are only observed where the annual rainfall exceeds 1200 mm. In drier climates, yellow soils appear at the centres of the plateaux. At 1000 mm annual rainfall, they occupy the most of the surface, with the red soils being restricted to the edge of the plateaux. Where rainfall is less than 1000 mm p.a., the processes have been working for a longer time and the yellow soils occupy the whole land surface. Examples of this system have been described from Brazil by Volkoff (1985), who found that an increase in aridity correlates with a *decrease* in the hematite/goethite ratio in soil material. However, in many other examples (Taylor and Graley, 1967; Lamouroux, 1972; Kampf and Schwertmann, 1983), an increase in aridity correlates with an *increase* in the hematite/goethite ratio. This ratio, which determines the yellow or red colour of the soil, is thus not only a function of climate but also of other factors such as soil texture, which controls the moisture regime, and the weathering history.

Transformation systems under arid climates

In arid areas, soil catenas in equilibrium with the present environment can be present, the accumulation of soluble salts in depressions being an example. Most of these are not transformation systems, however, because they are not progressively replacing an earlier soil cover. However, the disruptive replacement of cuirasse, saprolite or saprock by calcrete (Millot et al., 1977), known from arid and very arid climates in West Africa and Australia, can be recognized as transformation systems. In Senegal and Mauritania, for example, the pre-existing lateritic cover, which consists of saprolite overlain by a cuirasse, has been formed under past humid climates and is now in disequilibrium (Nahon, 1976).

As the climate became drier, the original kaolinitic weathering front has been replaced by a smectitic front, so that smectitic saprolite now underlies the kaolinitic saprolite. The iron cementation front has proceeded downwards at the expense of the kaolinitic saprolite and in places may directly overlie the smectitic saprolite. In a third stage of development, under the present arid climate, a calcite cementation front, corresponding to the in-situ formation of calcrete, develops downslope at the expense of the smectitic saprolite and, upslope, into the cuirasse. Consequently, the cuirasse is now progressively confined to the upper part of the landform units.

CLASSIFICATION OF TROPICAL SOILS

Consideration of the classification of tropically weathered soils is beyond the scope of this chapter. A number of classification systems have been developed

TABLE I.4-1

Main soil taxonomic units in use in the studied areas; French (CPCS, 1967) and FAO (FAO-UNESCO, 1974)

Climate	Humid	Wet savanas	Dry savanas	Arid	Humid mountains
French	Sols ferrallitiques (commonly leached or indurated)		Sols fersiallitiques	Sols minéraux bruts	Andosols
	Podzols (humic-ferruginous)	Sols ferrugineux tropicaux (commonly leached)		Sols peu évolués xériques	Sols ferrallitiques (commonly humiferous
		Vertisols		Sols gypseux sodiques (solonetz, solods)	Podzols
		Sols hydromorphes			
FAO	Ferralsols (Haplic) (commonly Plinthic)		Cambisols (Chromic, Vertic, Calcic)	Arenosols	Andosols
				Regosols	
		Luvisols (Plinthic, Ferric)		Yermosols	Ferralsols
	Acrisols (Ochric, Plinthic, Rhodic)		Nitosols (Dystric, Eutric)	Xerosols	Podzols
				Solonetz	
	Podzols (Ferric, Humic)	Vertisols		Planosols	
		Gleysols and Gleyic Groups			

and for the main concepts behind them, reference should be made to the principal publications: FAO classification: FAO–UNESCO, 1974; French: CPCS, 1967; Australian: Stace et al., 1968; USA: Soil Taxonomy, Soil Survey Staff, 1975). The FAO and the French classifications, which have the most comprehensive coverage of tropical soils, are summarized in Table I.4-1. Several of the classifications are genetic and reflect the climatic zonality of the principal processes involved in soil formation, as described in this chapter. However, they do not adequately account for the weathering history of an area; for example, as discussed above, the presence of ferrallitic soils (or ferralsols) in dry savannas and arid areas commonly relates to old kaolinitic saprolites that are no longer developing under present conditions.

All existing systems of soil classification tend to refer to soil bodies in terms of vertical successions of horizons in units such as profiles, pedons and elementary soil areas. The genetic relationships between the different horizons in the classifications are, accordingly, only considered as vertical processes. However, as seen in the descriptions of transformation systems, most regoliths in tropically weathered terrains are only understandable if considered in three dimensions which, at the scale of the systems involved, is generally at least that of the landform (see Fig. I.1-4). Applying the existing soil classifications to such systems tends to subdivide the continuum and to focus attention either on typifying concepts, or on the arbitrary boundaries between types. Thus, the use of these classification schemes alone in soil surveys is inadequate for describing and understanding the geography and genesis of soils.