

The complementary phenomenon, i.e., the transport of salt under influence of a gradient of the electric potential, is well known from, e.g., the determination of transport numbers with the *Hittorf method*. The effects vanish for equal mobilities, e.g., in KCl solution. The magnitude of these effects in porous systems may be substantially larger than in aqueous solutions if the transport number of either one of the ion species is influenced substantially. This is almost always the case in soil systems where the mentioned electric field in the liquid layers against the solid matrix induces high or sometimes very high values for the transport number of the cation. The potential gradient then arising as a result of concentration gradients (even of KCl) has been discussed extensively in soil science literature in relation to potentiometric measurements in soils involving salt bridges and was then referred to as "liquid junction potential."

#### $L_{TV}$ and $L_{VT}$ , "thermofiltration" and thermoosmosis

The coupling between heat and mass fluxes is less obvious on first sight. In line with the preceding reasoning, coupling will arise only if the system exerts a selectivity with respect to "hot" and "cold" molecules ("hot" molecules indicating the molecules with a high value of  $(\bar{H} + \Psi)$  or total specific enthalpy, or specific heat content). In that case, an applied pressure gradient causes, e.g., preferred transport of the "hot" ones, thus increasing the heat content of the system at the low pressure end, which amounts to a heat flux in addition to the mass-flux ( $L_{TV} > 0$ ). In a liquid saturated soil system, not much can be expected of this effect, unless the liquid layers are extremely thin (which is difficult to effectuate in the saturated system). As follows from the observed heat of wetting of dry soil, the heat content is indeed decreased for the first few layers of water due to the presence of strong adsorption fields. These layers are presumably also less mobile than further layers. For thick water films, the difference in the "average" heat content of the mobile layers as compared to that of the entire film becomes vanishingly small. Alternatively, a temperature gradient applied to a water film subject to an adsorption field will induce a mass flux as a temperature rise will "lift" a larger part of the molecules from the adsorbed (immobile) layer to the mobile one. Again this effect should be expected to be quite small for the saturated soil system because of the width of pores, which is usually of concern.

For unsaturated and frozen soils, the phenomena of thermofiltration and thermoosmosis are quite important (cf. Kay and Groenevelt, 1974; and Groenevelt and Kay, 1974).

#### $L_{TD}$ and $L_{DT}$ and $L_{TE}$ and $L_{ET}$ , thermodiffusive and thermoelectric phenomena

In these phenomena coupling will arise if the selection mechanism operates on the salt molecules and the countercharge, distinguishing between "hot" and "cold" specimen. For the salts present in charged pores, the effect must be extremely small. The salt being repelled from the immobile layer close to the liquid-solid interface, only a tiny fraction of the salt (this time the fraction with high heat content) is situated in this layer. The difference in average heat content of the mobile part compared with that of all salt molecules will be vanishingly small. A locally applied temperature increase will likewise not contribute measurably to a decrease of mobile salt molecules. The coefficients  $L_{TD}$  and  $L_{DT}$  will thus be negative in principle and presumably zero in practice. In contrast the countercharge is accumulated in the immobile layer, so  $L_{TE}$  and  $L_{ET}$  should be positive (for a positive countercharge). As to the practical

significance of these coupling phenomena, it appears that  $L_{ET}$  might indeed be easily detectable. As was indicated earlier in the discussion of  $L_{EV}$ , this effect is usually met in the form of a gradient of the electric potential arising here when a temperature gradient is applied under the condition of zero current. In systems with low electrolyte concentration, the majority of the "relatively few" conducting ions is situated close to the solid surface, where the ionic mobility may be impaired considerably. A temperature increase could then increase substantially the number of actually conducting ions. A considerable gradient of the electric potential might thus be necessary to maintain a zero current when a temperature gradient is applied.

P. H. Groenevelt

#### Bibliography

- Baver, L.D., Gardner, W.H., and Gardner, W.R., 1972. *Soil Physics*. New York: Wiley, 498 pp.
- Bird, R.B., Stewart, W.E., and Lightfoot, E.N., 1960. *Transport Phenomena*. New York: Wiley, 780 pp.
- Carlslaw, H.S., and Jaeger, J.C., 1959. *Conduction of Heat in Solids*. Oxford: Clarendon Press, 510 pp.
- Childs, E.C., 1969. *The Physical Basis of Soil Water Phenomena*. New York: Wiley, 493 pp.
- Crank, J., 1957. *The Mathematics of Diffusion*. Oxford: Clarendon Press, 347 pp.
- De Wiest, R.J.M., 1969. *Flow Through Porous Media*. New York: Academic Press, 530 pp.
- Groenevelt, P.H., and Bolt, G.H., 1969. Nonequilibrium thermodynamics of the soil-water system, *J. Hydrol.*, **7**: 358–388.
- Groenevelt, P.H., and Kay, B.D., 1974. On the interaction of water and heat transport in frozen and unfrozen soils: II. The liquid phase. *Soil Sci. Soc. America Proc.*, **38**: 400–404.
- Hillel, D., 1971. *Soil and Water*. New York: Academic Press, 288 pp.
- Kay, B.D., and Groenevelt, P.H., 1974. On the interaction of water and heat transport in frozen and unfrozen soils: I. Basic theory: The vapor phase. *Soil Sci. Soc. America Proc.*, **38**: 395–400.
- Kirkham, D., and Powers, W.L., 1972. *Advanced Soil Physics*. New York: Wiley, 534 pp.
- Rose, C.W., 1966. *Agricultural Physics*. Oxford: Pergamon Press, 226 pp.
- de Vries, D.A., and Afgan, N.H., 1975. *Heat and Mass Transfer in the Biosphere*. New York: Wiley, 594 pp.

#### Cross-references

Conductivity, Electrical  
 Conductivity, Hydraulic  
 Conductivity, Thermal  
 Diffusion Processes  
 Electro-osmosis  
 Energy Balance  
 Flow Theory  
 Thermal Regime  
 Thermodynamics of Soil Water  
 Water Budget in Soil

---

#### TROPICAL SOILS

---

The term 'tropical soils' is not as geographically inclusive as might be inferred from the name. From a pedological point of view, the term traditionally does not include all soils occurring in tropical environments, soils of low-latitude deserts typically being excluded, as are some pedogenically young (poorly developed) soils. The inclusive and exclusive terminologies are somewhat inconsistent and their applications depend of

specific reference points, most commonly referring to extreme degrees of biochemical weathering for prolonged periods of time. Tropical soils are thus defined principally by climate and stage of genetic development with particular examples dependent upon parent material and topography. Identification of tropical soils is partially dependent on the classification system and its interpretation of the convergence of the factors of soil formation that come together to produce mantles of intensely weathered materials.

As interpreted by the FAO-UNESCO-ISIC (1990) system of soil classification (Nachtergaele et al., 2000), tropical soils cover about 30 million km<sup>2</sup> of the Earth's surface and are loosely defined as those weathering profiles that are characteristic of the non-desertic low latitudes. Tropical soils form under tropical climates, of which there are several main types. The tropical climates are roughly those where the mean annual temperature exceeds 20 °C and no monthly mean falls below 18 °C. Geographic boundaries, so constrained, are roughly the Tropic of Cancer and Tropic of Capricorn (23° N and S), from which some authorities would exclude the tropical equatorial climate, specifically the equatorial rainforest belt. With the latter proviso, the tropics are then distinguished by seasonal rainfall (part wet – part dry) but exclude hot desert regions with less than 25–40 mm annual precipitation, thus excluding much of the Sahara Desert (south of 23° N to 15° N), although they would include a specific tropical savanna-type climate with seasonal precipitation less than 1 000–1 400 mm (see Oliver and Fairbridge, 1987).

Tropical soils, disregarding those relating to specific geological histories such as recent volcanism, coastal plain eustatic fluctuations and principal river valleys, reflect extreme longevity, leaching, and (geo)chemical evolution (McFarlane, 1976; FitzPatrick, 1980; Eswaran et al., 1983; Nahon, 1991; Finkl, 1995). Soils formed in hot climates (subtropical, tropical, equatorial regions) where the effective precipitation produces environmental conditions that range from semi-arid to perhumid have common characteristics that are notably different from soils in other environments. Weathering of primary minerals is more complete than in temperate climates and it takes place to greater depths (e.g., Duchaufour, 1978; Büdel, 1982). Because organic matter is subject to rapid biodegradation and recycling and tends to remain near the ground surface, weathering favors geochemical processes (e.g., neutral or slightly acid hydrolysis) that result in higher concentrations of free oxides than is typical of temperate-region soils. Soil color in hot climates is generally much more intense (brighter) than in temperate climates because the freed iron and aluminum oxides remain in the profile. The ratio of free iron: total iron of a weathering horizon on granite, for example, never exceeds 50% in a temperate acid brown soil, but reaches 60–70% in a ferrallitic red soil and 100% in a ferrallitic soil (Duchaufour, 1977).

Differences in soils from major world regions were highlighted in the last several decades by global soil mapping programs conducted under the aegis of the FAO and UNESCO (Deckers et al., 2002). The new soil maps of the world and development of modern national and international soil classification systems disclosed or unmasked the limitations of concepts, methods, and techniques based on experience and information acquired in mid-latitudes when applied to tropical soils (Steila, 1976; Lepsch and Buol, 1988; Finkl, 1995). Far too complex to fit readily into the framework of concepts based on mid-latitude soils, tropical soils are now regarded within modern soil paradigms as a distinctive climato-geographic

grouping (e.g., Duchaufour, 1978; Jenny, 1980) with unique properties and characteristics that require special care in land use and environmental protection. Another, more specific point of view, is offered by reference to soil temperature regimes. While recognizing the continual warmth of tropical (equatorial) conditions, soil temperature is regarded in some modern classification systems as an important parameter in soil development but it is not a definitive criterion for identifying tropical soils. According to the (U.S.) Soil Survey Staff (1992), for example, soil temperature classes are used as family differentiae in all orders. For soil families that have a difference of less than 5 °C between mean summer and mean winter soil temperatures at a depth of 50 cm from the soil surface, the temperature classes in warm climatic regions are defined as isothermic (10 to 22 °C) and isohyperthermic (22 °C or higher).

In the classical sense of tropical soil development, based largely on their experience in Indonesia, Mohr et al. (1972) helped to firmly establish that 'tropical soils' are products of long-term intense weathering under low-latitude conditions. However, readers should be aware that Mohr's model, Indonesia, is largely dominated by youthful volcanoes, and, with something like ten thousand islands, by littoral deposits, both of which are excluded from our global generalizations. Soil scientists traditionally recognize the importance of climate (*re* the climatic regime) among the five factors (topography or relief, biological activity, climate, parent material, and time) of soil formation, as described by Jenny (1941) in his functional-factorial approach to soil genesis. Historically known as soil-forming factors, Jenny (1980) regarded them as a group of variables that are ecosystem determinants or variables. Long-term cycles of precipitation and temperature, as well as seasonal variations in evapotranspiration, are regarded by pedologists as important factors in the development of soil properties that may, in turn, be used as indicators of past climates and changing environmental conditions. Although climatic regimes in the tropics range from deserts to perhumid conditions, weathering profiles of the humid tropics are commonly included in the concept of tropical soils (e.g., Ganssen and Hadrich, 1965; Duchaufour, 1978; Buol et al., 1980; Richards, 1996).

When viewed in detail, pathways of soil development in the humid tropics show great variation with climate being the primary and overriding controlling factor for many key soil properties in equatorial regions. These properties relate not only to pedogenic processes but also to soil as an essential natural resource (Jenny, 1980; Hassett and Banwart, 1992) and potential of the land to produce crops (Aubert and Tavernier, 1972; Foth and Schafer, 1980; Anderson and Ingram, 1993), particularly within the context of sustainable development as the human population of tropical regions continues to expand exponentially (Bouma et al., 1996).

Soils of the tropics are well known for their inherent low fertility, with the exception of some that are developed in alluvial or volcanic parent materials (Sanchez, 1994). The ability of tropical soils to support crops is limited by properties that are different from middle-latitude soils and that is why tropical soils require special management techniques to realize their yield potential (Steila, 1976; Foth and Schafer, 1980; Hassett and Banwart, 1992; Juo and Franzluebbers, 2003; van Wambeke and Nachtergaele, 2003).

#### Environments of formation for tropical soils

Tropical soils occur in regions that are humid enough to allow development of woody plants such as hygrophilic forest



(ferrallitic soils of humid climates), xerophytic forest (ferralsitic soils of semi-arid evergreen forests), or mixed savanna or bush (ferruginous tropical soils) (Duchaufour, 1977). When the climate becomes even dryer, xerophyllic steppe takes the place of forest and the soils become transitional to other kinds of weathering products. While climate remains the fundamental factor of pedogenesis in hot climates, its duration, seasonality, and long-term alteration or replacement by other climatic regimes during plate-tectonic migrations and ice ages (e.g., low-latitude cycles of aridity or pluviosity) becomes an important consideration in the development of profiles of deep chemical weathering on stable land surfaces. During the Quaternary ice ages there were severe climate fluctuations in the tropics (perhumid to arid and back) (Fairbridge, 1976; Fairbridge and Finkl, 1980; Retallack, 1990).

Characteristic of a different climatic zone, three basic phases of weathering (where organic matter is not significantly involved) are often recognized in hot climates. Pedologists have noted (e.g., Duchaufour, 1977; Goudie and Pye, 1983), however, that these phases of weathering can occur simultaneously in the same climatic zone and, when conditioned by site factors that may affect the relative duration or intensity of pedogenesis, they can also be considered as phases in the same overall weathering process. These three phases, following the nomenclature of Duchaufour (1977), are characterized by an increasing degree of weathering of primary minerals, an increasing loss of combined silica, and increasingly marked dominance of neoformed clays. In Phase 1 (ferralsitization), there is a dominance of 2 : 1 (expanding lattice) clays rich in silica, considerable amounts of free iron oxides that are more or less rubified (red or ochrous color), and the exchange complex is almost saturated by the upward movement of bases during the dry season. An argillic horizon often occurs as a result of illuviation and cheluviation. This phase is typical of subtropical climates with a dry season. Phase 2 (ferrugination) features stronger weathering (some primary minerals such as orthoclase or muscovite may persist), marked desilication, and more 1 : 1-type clays (kaolinite) than 2 : 1 transformed clays. Iron oxides may not be rubified, and base saturation is variable, depending upon the humidity of the climate and the importance of the dry season. Illuvial horizons are less well developed than in ferralsitic soils. Phase 2 weathering represents the final stage of pedogenesis in climates that are less hot (humid sub tropics) or marked by a dry season (dry tropics); these weathering products more or less correspond to ultisols (USA) and Acrisols (WRB). Phase 3 (ferrallitization) is characterized by complete weathering of primary minerals (except quartz) and the clays are all neoformed, consisting solely of kaolinite. Gibbsite occurs frequently but its presence is not essential. Although illuviation still occurs, there is no true argillic horizon. This phase in the tropical climofunction, which is generally reached in hot climates on those materials that are the oldest or where very rapid development can occur, produces ferrallitic soils, which more or less correspond to the oxisol order (USA) or the Ferralsols (WRB).

Humid tropical regions are characterized by a continually moist environment with a pattern of two precipitation maxima, a total rainfall of at least 2 000 mm or more, and a mean temperature greater than 25 °C with only slight daily variation. Within the tropics three basic climatic types, based on rainfall contrasts, are identified in the Köppen-Geiger climate classification system (Köppen and Geiger, 1930; see also Trewartha, 1954): the constantly wet or tropical rainforest climate (*Af*),

an Am monsoon rainforest climate in which unusually heavy rain compensates for a short dry season, and the *Aw*, tropical wet-and-dry (savanna) climate, with its summer or zenithal rains and low-sun dry season. In the savanna climates that occur towards the subtropics, there is only one rainy season corresponding with the high sun period; that season may be split in half, however, by the forward advance and retreat of the monsoon front (Smith, 1967). Rainfall varies considerably with location and altitude from 600 mm to 1500 mm per annum, and although the average temperatures are similar to those of the humid tropics, the range (20 °C) is much greater, frequently exceeding 40 °C near the end of the dry season and falling to 0 °C in nighttime lows near the winter solstice.

Climatic changes associated with Pleistocene glaciations in upper and middle latitudes also affected the margins of tropical areas (Büdel, 1982; Arnold et al., 1990). Much of tropical Africa, Australia, Brazil, and India underwent stages of ice-age hyperaridity (Fairbridge, 1972). Sand dunes from the Sahara reached as far south as the Congo (Zaire) Basin, as did Kalahari sands from the south. In South America, dune sands reached from Patagonia to the Amazon Basin; even in the littoral rainforest belt of Brazil, the geomorphology and sediments show cyclic evidence of extended seasonal aridity (Bigarella and de Andrade, 1965; Fairbridge and Finkl, 1980). Thus, in the case of the tropical arid Sahara (*BWh* climate) and the subtropical Sahelian region savanna (*BSh* climate), i.e., the belt that runs from Senegal and Sierra Leone to Nigeria and the southern Sudan, when subjected to arid conditions during these dry phases, the restricted vegetative growth led to more rapid natural soil erosion. In the moist or pluvial phases, corresponding to interglacial climates, semi-arid boundaries (with monsoonal summer rains) shifted poleward and the vegetative balance was permitting more growth of savanna-type trees, which resulted in greater soil stability. During the approximately 2 million years of the Quaternary Period, there were at least 20 cycles of alternating humidity and aridity (interglacial-glacial global regimes). These changes in climatic regime affected the distribution and density of vegetation cover which in turn resulted in stabilized-destabilized phases on land surfaces that profoundly affected the areal extent and degree of weathering (saprolitization), formation of duricrusts, and soil development as well as the loss of soil materials from hillslopes (e.g., Finkl and Churchward, 1976; Williams, 1985; Finkl, 1988). Strong climatic fluctuations were not limited to the Quaternary ice age, but can be traced back in geological time. Ancient soil surfaces, notably with formation of duricrusts (e.g., laterite) (Beckmann, 1983), can be identified and dated 50, 100 or even 200 million years (Retallack, 1990). Examples of laterite deep weathering profiles developed in crystalline rocks are found on the tropical/subtropical shield areas of Australia, Africa, and South America. Figure T13 shows typical profile development in the granitic terrain of the Darling Ranges in southwestern Western Australia. A similar deep weathering profile on the Brazilian craton is shown in Figure T14. Both profiles are developed in situ with a truncated surface (eroded surface horizons) and extensive mottled (oxidized) zone. The Brazilian example (Figure T14) shows more extensive development of pallid clays near the base of the road cut. Remains of some of these ancient paleosols still provide the framework of some modern tropical soils (Finkl, 1988; Nahon, 1991). Pedologic mantles of tropical humid (equatorial) regions, which often reach thicknesses of 100 m or more of chemically weathered or altered materials (alterites), often



**Figure T13** Chemical alterite of the lateritic type near Toodyay (85 km east of Perth) in the Darling Ranges, escarpment of the Darling Fault line. The results of deep chemical weathering shown in this railway cut shows oxidation of granite parent rocks to produce thick zones of chemically altered materials. Quartz veins comprise less weathered residuum that features small core stones and provides proof of in situ development. The cut shown here is about 20 m deep, but these deeply weathered soil mantles typically extend to great depths that are often in excess of 50 to 100 m. The mottled zone shown here overlies pallid kaolinitic clays at depth.

contain several differentiated zones that are superposed vertically and which are strongly lithodependent in the lowermost portions. As a consequence of climatic change and continental drift due to shifting of tectonic plates, deep weathering profiles with or without protective duricrusts now often occur as relict soil formations in many subtropical and temperate zones (Fairbridge, and Finkl, 1980; Retallack, 1990). In contrast to their great longevity in the landscape, duricrusts formed during the Holocene (the last 10 000 yr or so) were mostly established within a century or two (Goudie, 1973; Fairbridge, 1976).

#### Tropical geomorphic environments and soils

Many inter-tropical areas have been dry land for long periods of geological history (in places 200 million yr or more), and the deposits on them are often deeply weathered terrestrial materials dating back at least to mid-Cenozoic times (Bridges,



**Figure T14** Laterite developed on shield rocks in southeastern Santa Catarina State, Brazil. The oxidized mottled zone (*upper part of the profile*) and kaolinitic pallid zone (*lower part of profile*) are typical of chemical laterites that are differentiated into zones based on fluctuating water tables and oxidation. Much of the mottled zone has been eroded away at this location leaving a thin oxidized layer that transitions in pallid kaolinitic clays that may extend for many tens of meters below the surface.

1978; Bremer, 1981). Within any particular region, geomorphologists commonly recognize several surfaces associated with cycles of erosion of different ages (King, 1962; Twidale and Campbell, 2005). Although different kinds of subdued relief states (including soils and weathering mantles) associated with near end products of climatomorphogenic planation are recognized in tropical and peritropical regions (e.g., peneplains, pediplains), the widespread occurrence of etchplains is becoming quite generally appreciated. As Büdel (1982) stresses, the existence of etchplains as a dominant relief type is proof that rivers cannot cut down faster than planation and cannot carve valleys even over very long periods of time. Inextricably intertwined in the etchplain concept is the prolonged and intense chemical weathering (alteration) of parent rocks on stable cratons where the basal surface of weathering (weathering front) chemically decomposes at depth, while in the rainy season finely worked material is correspondingly removed from the ground-surface by highly effective sheet wash. Büdel (1957) proposed that this “mechanism of double planation surfaces” (*Mechanismus der doppelten Einebnungsflächen*) is responsible for creating etchplains over long periods of geologic time. Other workers have confirmed the deduction as a general rule and emphasized its importance to the development of tropical landscapes in South America, Africa, and Australia (e.g., Finkl and Churchward, 1973; Thomas, 1974; Finkl, 1979; Bremer, 1981; Spath, 1981; Twidale and Campbell, 1993, 2005). As explained by Büdel (1982), Bremer (1981) and Spath (1981), this pedo-geomorphological concept is of fundamental importance to understanding tropical landscapes because in the entire ectropics, and in arid and humid tropical mountains, mass wasting increases as the slope becomes steeper. In the humid tropics and peritropical zones, the situation is completely reversed – only where soil is present does erosion occur, continuing the development of flat surfaces. Where bedrock is exposed, as in inselbergs, shield inselbergs, or tors, the rock surfaces are edaphically arid and physical weathering, consisting of small-scale exfoliation and grus weathering, works far more slowly than chemical weathering (Bremer, 1975; Twidale and Campbell, 2005). The rock surface itself is frequently “armor-plated” by superficial concentration of silica, iron and magnesium oxides (see Fairbridge, 1968, pp. 552, 556, 1 104).



These planation surfaces have soils of different ages upon them with different profiles and properties. On tropical cratons (stable platforms or continental crustal areas with a nucleus of Archean age), Fairbridge and Finkl (1980) recognize a "cratonic regime" that is characterized by long-term (on the order of  $10^7$ – $10^8$  yr) alternations between distinct high and low relief states. Dependent upon these two conditions are the vegetation, weathering, and soils. The stable biostatic state is characterized by high sea levels, when land areas are reduced in size, and maritime climates associated with the thalassocratic condition. In the unstable rheixistatic state when land areas are expanded, relief is amplified, and continental climates spread out under the epeirocratic condition, vegetative cover is reduced and soils become eroded. The extensive redistribution of materials which thus takes place on geomorphic surfaces of great antiquity results in complicated patterns of soils, which can be understood only if their mode of origin is first deciphered.

Concepts of erosional and depositional phases in soils, as developed in Australia (e.g., Butler, 1959), have greatly assisted in the elucidation of soil patterns found on these old continental blocks. Also, in many parts of the tropics and sub tropics there are various types of duricrusts (e.g., ferricrete, silcrete, calcrete) (Fairbridge, 1968; Goudie, 1973; McFarlane, 1976; Goudie and Pye, 1983) that have a striking impact on the landscape, a fact appreciated by many of the great explorers and travelers of the nineteenth century. Perhaps their strongest visual impact is on landforms, but their effect on vegetation associations and protection of underlying soils is considerable, as is their impact on human activities and the cultural landscape. The importance of an indurated laterite crust in the landscape is emphasized by recognition of distinct environments commonly referred to as "lakere" and "bowe" in West Africa where they stretch over considerable portions of east and southeast Senegal; southern Mali, Chad, and Sudan; the Fouta Plateaux of Guinea; and the north parts of the Central African Republic. Bowe may develop extremely quickly, within one season of vegetation clearance for plantation crops or grazing land, as seen in India and Amazonia when cleared tracts became virtually pavements of rock in five years (McNeil, 1964). Modern examples of the bovalization process occur in many parts of Amazonia where tropical rainforests are cut down for grasslands that feed beef stock. This anthropogenic duricrust formation appears to be related to the dehydration of silica and ferruginous gels. These develop in poorly drained rainforest soils, evaporation initiates a rapid loss of soil moisture and cement-hard cryptocrystalline mineralization occurs (creating ferricrete and silcrete).

At low and intermediate elevations within the tropics, temperatures are high throughout the year without substantial seasonal variations. In these high isothermic regimes, soil-forming processes (see *Weathering systems in soil science*) occur faster than in temperate regions (Bridges, 1978; Nahon, 1991), particularly the processes leading to advanced weathering stages of the parent materials. High temperatures also accelerate the biological turnover of organic matter in tropical soils (by bacteria, earthworms, and termites), a process that builds up humic complexes (Martius et al., 2001). The advanced stages of weathering, which occur in the humid tropics, are partly due to rapid rates of decomposition for long periods.

Age is a significant variable that determines many attributes of soils in tropical environments and distinguishes them from younger soils in temperate regions. The largest land areas of the tropics belong to tectonically stable continental shields (cratons) and tablelands, which have not been subjected to

geologically recent folding. Rather, they have been subjected to gentle upwarping into continental swells, described by King (1962) as cymatogens, and broad downwarping into large basins. This tectonic framework is particularly true for much of Africa, Australia, the Brazilian and Guinean shields in South America, and the Gondwana part of the Indian peninsula. Soil erosion has not been strong enough to remove the products of weathering on these relatively flat-lying cratons of low relief and consequently vast waste mantles have accumulated from deeply weathered materials (e.g., Finkl, 1988). The weathered materials comprising these waste mantles possess a very low mineral reserve for supplying nutrients to plants, are largely dominated by kaolinite in the clay fraction, and have retained a high concentration of free iron oxides in the parent material. Low cation absorption capacities have these soils particularly susceptible to leaching, although most tropical soils contain some free aluminum oxides. In general, they have a deep acid solum that is poor in both major nutrients and micronutrients. Generally, they are well drained and their structure provides good aeration. However, as a result of their evolution during millions of years under varying environmental conditions, they often show an altered upper layer in which gravels and even stones play an important role. These deep, well-drained soils of the humid tropics have been called by such names as latosols, sols ferrallitiques, ferrisols, kaolisols, oxisols, etc. in different countries (cf. Figures T13 and T14 for examples in Australia and Brazil).

Not all soils within inter-tropical environments are developed on old landscapes and some highly productive soils may occur on young constructive geomorphic surfaces in equatorial regions. Recent mountain building in alpine orogenic belts has exposed fresh rock to erosion and provided a source of mineral nutrients to crops in the foothills of the Andes, Himalayas, and Southeast Asia. Depressions and valleys in tropical Asia, for example, are filled with detrital materials that react to management differently from the soils on continental shields. A greater variability occurs within the profile characteristics of these soils and no longer is the low mineral supply a major limiting factor for plant growth. During a long dry season in areas with *Aw* and *BS<sub>h</sub>* climates (or humid subtropical climatic regimes), however, these soils may develop a structure unfavorable to rain-water penetration and root development, which increases the erosion hazard. Soil scientists have variously referred to these soils as: sols ferrugineux tropicaux, sols ferrallitiques lessives, red yellow podzolics, ferrisols (in part), ultisols, ferruginous soils, krasnozems (red earths), zheltosols (yellow earths), or gray podzolics (Aubert and Tavernier, 1972). The younger or rejuvenated soils developed from various parent materials have been called sols bruns, tropepts, ferrisols, or alluvial soils; those from basic volcanic rocks such as basalts, reddish brown lateritic or terra roxa estruturada; and those from volcanic ash have been referred to as Andosols or andisols. The most productive soils are those among Nitrosols (cf. nitosols) (Spaargaren, 1994).

### Distribution of major tropical soils

Due to the complexity of soil geography in equatorial environments and the desire to perceive an overall view of tropical soil distributions that can be reasonably grasped, simplification of data is necessary. Because there are a large number of soil classification systems in use with regional specializations on different continents and within separate counties (see, for example, discussions in Buol et al., 1980; FitzPatrick, 1980; Finkl, 1982a,b; Eswaran et al., 1983; FAO-UNESCO-ISRIC, 1990;



Van Wambeke et al., 1990; Soil Survey Staff, 1992; Spaargaren, 1994; Deckers et al., 2002), it is advantageous to consider a global regionalization of tropical soils in terms of a single comprehensive international system. It is also seen from geographical applications of modern soil classification systems, summarized on a worldwide basis by Buol et al. (1980) and Finkl (1982a,b), among others, that tropical soils comprise a major part of many systems of classification. The complexity of nomenclature has, to a large extent, been rationalized in the FAO legend for the Soil Map of the World (FAO-UNESCO, 1974, and subsequently updated with the release of additional map sheets). For the sake of clarity, the common language of the FAO-UNESCO Soil Map of the World (FAO-UNESCO-ISRIC, 1990) is used as a central reference system in the following discussion of tropical soils. In their 30 million km<sup>2</sup> extents, they are typical of the ancient cratons, specifically the Precambrian shields of South America, Africa, India, and Australia. As described below, the main categories of tropical soils are Acrisols (1 000 million ha), Ferralsols (750 million ha), Fluvisols (350 million ha), Lixisols (435 million ha), Luvisols (650 ha), Nitisols (200 million ha), sesquisols (60 million ha), and Vertisols (335 million ha). Synonymous terms or previous widely used terminologies are indicated where appropriate in reference to the FAO terminology (Deckers et al., 2002).

### Soils of humid tropical environments

From a pedo-geographic perspective, soils of the tropical soil-forming environment are restricted to lowlands in the equatorial domain. Tropical soils are known as those with the greatest depth and intensity of chemical weathering. Because there are almost no weatherable minerals left in the plant-rooting zone, many tropical soils have a low nutrient status and require special land care procedures to maintain natural productivity.

### Sesquisols

Sesquisols (from *E. sesquioxide*; connotative of iron and aluminum; soils containing a petroplinthite or plinthite layer) are soils either containing at shallow depth a layer indurated by iron (petroplinthite), or at some depth mottled material that irreversibly hardens after repeated drying and wetting (plinthite). These soils mainly occur in the tropics but examples also are found in subtropical and temperate regions, such as the rafia surface of central Spain or the Great Plateau of Western Australia. Those with a shallow petroplinthic horizon were also known as high-level laterites, ironstone soils, or sols ferrugineux tropicaux a cuirasse (Soil Survey Staff, 1992). They have widespread occurrence in western Africa, especially in the Sudano-Sahelian region where they cap structural tablelands; central-southern India; the upper Mekong River catchment area, parts of northern and western Australia, and the eastern part of the Amazon region in Brazil and on shield areas of Surinam and the Guianas.

The sesquisols with plinthite are known as Plinthosols (FAO-UNESCO-ISRIC, 1990), groundwater laterite soils, low-level laterite, lateritas hidromorficas, sols gris lateritiques, or plinthaquox (Soil Survey Staff, 1992). They are found in extensively flat terrains with poor exorheic (external) drainage, such as the late Pleistocene or early Holocene sedimentary plains of eastern and central Amazonia, the central Congo basin, lowlands of Indonesia, and older Mesozoic surfaces in Western Australia. In many regions, the subsurface petroplinthite layer becomes exposed at the surface by erosion to form a ferricrete

caprock or duricrust. Geomorphically, the soils may occur on gentle rectilinear slopes with an impermeable substratum and in footslopes positions along concave slopes in rolling or tableland landscapes.

Well-drained soils with loose ironstone concretions (pisolitic materials) are frequent nearly everywhere in the tropics and subtropics, in many landscape positions. The gravelly material is the result of former plinthite formation, subsequent hardening, and transport or re-weathering. The soils concerned are geomorphically related to sesquisols, but pedomorphometrically belong to other soil classes. The global extent of soils with plinthite is estimated at about 60 million ha (FAO-UNESCO-ISRIC, 1991). Those soils with a shallow petroplinthic horizon have poorer, stunted natural vegetation than consociated soils without such a hardpan. Where the horizon is less shallow, less dense, broken up, or transported downslope into sesquiskeletal accumulations (e.g., colluvial pisolitic gravel deposits commingled with freshly weathered rock fragments), the vegetation may be even more luxurious than its non-stony counterparts because of the presence of less weathered pockets within the plinthic material. Arable cropping and tree planting is problematic because of the stoniness of the soils, but the latter feature is often welcomed by construction engineers who use the materials in road construction and for building purposes.

Imperfectly drained soils with a plinthite horizon are more sparsely vegetated by phanerophytes than geographically associated well-drained soils, for instance tree savanna or grassy savanna instead of closed-canopy high forest. Also, the land use on such soils is often restricted to extensive grazing because arable crops would suffer from poor rooting conditions. Artificial drainage of the soils would entail a serious hazard of irreversible hardening of the plinthic material. The hardening liability is, however, an asset for non-agricultural uses, including mining (for iron ore, manganese, bauxite) and building material (brick making, road building, terracing). Weathered alluvial materials, which sometimes contain placer deposits (e.g., gold, diamonds, and tin), are mined in deeply weathered landscapes.

### Ferralsols

Ferralsols (from *L. ferrum*, iron; connotative of a high content of sesquioxides; soils with a ferralic, oxic B horizon) are deeply weathered, iron-rich soils that are often associated with a hard horizon generally known as laterite. These soils include most of the soils previously called laterite, groundwater laterite, and latosols (Soil Survey Staff, 1992; Spaargaren, 1994). In the soil continuum, soils with ferralic properties represent the weathering extreme of the spectrum. These soils are formed by the progressive hydrolysis and complete transformation of the parent rock into clay minerals, oxides, oxyhydroxides, concentrations of the resistant residue, and loss in the drainage of much material, particularly basic cations and silica (Buol et al., 1980; Nahon, 1991). The central concept is represented by a soil formed on a stable landscape having been subjected to weathering and soil formation for a long time. Profile development follows a number of different pathways as determined by the nature of the parent material and drainage characteristics. The upper parts of these soils that have free drainage may become saturated with water during the rainy season but this condition does not persist because there is no evidence of reduction. In the lower part of the soil, the mottled clay that seems to derive its color pattern from prolonged periods of

wetness in an anerobic environment contributes to ferruginous nodular differentiation (FitzPatrick, 1980; Nahon, 1991).

This tropical formation of soils, ferralsolization, characterized by a colloidal fraction dominated by low activity (kaolinitic) clays and sesquioxides produces Ferralsols (Eswaran et al., 1983). These soils have very low amounts of weatherable minerals that have the potential to release nutrient cations on weathering, a uniform profile characterized by its lack of horizonization, a reddish color, weak expression of pedal structure, and few marks of other soil forming processes such as clay accumulation through translocation (Paramanathan and Eswaran, 1981). Weathering has been sufficiently advanced that rock fragments with weatherable minerals are absent. Secondary accumulation of stable minerals (e.g., gibbsite or iron oxyhydrates) may be present in concretionary (pisolitic) forms or as part of the fine earth fraction of the soils (e.g., Fairbridge and Finkl, 1984). Typical soils are situated on geomorphologically old surfaces, which have been formed through erosion and deposition (Van Wambeke et al., 1980). Many such soils are formed in transported and reworked materials and may have little relationship to the underlying geological strata. Figure T15 shows a soil profile in Western Australia that contains transported laterite gravels (pisoliths) that have moved

down slope under the influence of gravity. The pisoliths and finer-grained soil materials have no genetic (parent material) relationship with the underlying granite.

#### Environment of laterite (ironstone)

The term *laterite* was reportedly first used in India (cf. Buchanan, 1807; Babbington, 1821) in reference to the vesicular mottled, red and cream, clay which was dug out of the ground, shaped into bricks, and allowed to dry in the sun. Because the early workers did not describe the morphological properties of the material in great detail nor provide chemical analyses, the term has been used rather loosely for ferruginous clayey material that was hard or would harden on exposure. In addition, the term was often stretched and widely applied to many weathered materials so that red, strongly weathered, tropical soils became known as lateritic soils and, consequently, the term was often applied whether or not soils contained true laterite. In spite of these difficulties, modern researchers now apply the term laterite (plinthite, petroplinthite) to a soil horizon that is hard or will harden on exposure and is composed mainly of the oxides, oxyhydroxides of iron and/or aluminum with varying amounts of kaolinite and quartz and sometimes oxides of manganese.

This broad definition of laterite includes quite a range of material, because horizons that are hard or harden, can be vesicular, concretionary, massive, or mixtures of these types (FitzPatrick, 1980). Although vesicular laterite is very common, the greatest variability seems to be in the number and type of concretions (glaebules) in anyone formation. Some workers divide the concretions into two main types – pisolitic and nodular. Here there is also confusion, because some rounded fragments of detrital vesicular laterite are termed pisoliths. There are a large number of profile and landscape positions in which laterite will form. In the majority of cases, laterite seems to be a specific soil horizon formed above or within the mottled clay. Such a type of laterite occurs dominantly on flat or gently undulating landscapes and is attributed to a fluctuating water table. Because laterite hardens on exposure, in many (now semiarid) landscapes it forms a surface capping following uplift and aeration of the topsoil, so that the landscape has a number of mesas (or buttes) and escarpments (called breakaways in Australia) below which are long pediment (or parapedimented) slopes with a cover of pedisements. With time, the laterite weathers, breaks off into fragments of various sizes and forms relatively thin (<1 m) colluvial covers on the slope but with much thicker accumulations on foot slopes and in valleys (McFarlane, 1976). In some circumstances, this colluvium becomes cemented to give a secondary type of laterite.

Although the morphology, mineralogy, and genesis of laterites is extremely variable, several general types are recognized, as summarized by FitzPatrick (1980): (1) mottled red and cream material with weak vesicular structure composed of kaolinite, iron oxides, gibbsite with partially preserved rock structure; will harden on exposure. This type is derived from the characteristic mottled clay of the tropics, (2) vesicular, composed of a hard continuous phase of dark brown or black, iron impregnated material surrounding cream kaolinitic or gibbsitic clay, (3) very dense massive material dominated by subspherical black iron oxide concretions embedded in a reddish-brown matrix, (4) reddish-brown massive with abundant black concretions and thin veins of well crystalline kaolinite between the concretions, (5) yellowish-brown to reddish-brown, hard, scoriaceous and



**Figure T15** Transported lateritic detritus moving down slope on the West Australian craton near Narrogin (east of Perth). Derived from lateritic materials farther upslope, these colluvial deposits move down slope to a point where they become stabilized and form new parent material for another phase of soil formation. The materials here are unconformably overlying granitic boulders. Surface horizons commonly contain pisoliths (ironstone gravels). Indurated ferricrete boulders are often associated with these transported or secondary laterites.

composed dominantly of gibbsite, (6) yellowish-brown, massive, dominated by gibbsite concretionary material, (7) black or very dark brown massive containing much quartz sand, gravel and small rock fragments on lower slope positions, (8) black or very dark brown, fused nodular concretions, very high bulk density, composed mainly of iron oxide and manganese dioxide in wet lower slope and depression situations, (9) yellowish-brown to red cemented black concretions, very high bulk density in a high porosity mass of fused concretions that have been concentrated by differential erosion and then cemented, (10) massive, dark-brown to black, containing abundant black spherical units and rounded rock fragments forming a cemented colluvial deposit. Sometimes two different layers of laterite occur one above the other in the same profile. Since most laterites are very old, they are often influenced by later pedogenic processes so that some have accumulations of calcite, gypsum and may even be silicified.

Ferralsols cover an estimated worldwide area of about 750 million ha with roughly 60% found in South and Central America, and the rest in Africa (FAO, 1991) and Australia. These soils are geographically associated with Cambisols in areas with rock outcrops or where rock comes near to the surface. On the stable surface they occur together with Acrisols, which seem often to be related to the presence of more acidic parent materials (e.g., gneiss). On more basic rocks they occur associated with Nitisols, with no apparent relation to underlying rocks or topographic positions. Near valleys, Ferralsols merge into Gleysols, Planosols or even Arenosols (the southern African dambo system) and sequisols (in the eastern Amazon basin).

The low nutrient (base) status and low organic matter content give these soils very low natural fertility but they often support high forest. The relationship between Ferralsols and their natural forest vegetation is a good example of the delicate balance of nature in which nutrients are constantly recycled to maintain the forest community. When the forest cover is removed by land clearing and agriculture is practiced, fertility is quickly exhausted and crop failure normally results. This closed forest-soil nutrient recycling system accounts for the practice of shifting cultivation so highly developed in parts of equatorial Africa, Brazil, Malaysia, Borneo (Kalimantan) and Sumatra, and Papua New Guinea (Steila, 1976; Anderson and Ingram, 1993). Modern practices, including liming and application of fertilizers, provide a more stable agricultural system but no completely satisfactory system of land use has been developed for many of these soils. In some cases, the greatest success is achieved with tree crops such as cocoa, oil palm, tea, coffee, and rubber (Spaargaren, 1994). Very hardy plantation crops such as sugar cane are successful. Because the profit per hectare for carbohydrate crops is small, these soils seldom sustain a high standard of living except when their utilization is on an extensive scale.

#### Nitisols

Nitisols (from *L. nitidus*, shiny; connotative of shiny ped surfaces) are very common in tropical and subtropical areas and were first given a separate designation in the Australian soil classification system. These soils cover more than 200 million ha globally, of which almost half is found in eastern Africa (FAO, 1991). Other regions with Nitisols are found in southern Brazil, Central America, Java, and the Philippines. These well-drained soils contain dusky red to dark brown clays with a strongly developed fine blocky (polyhedral) structure and shiny ped

faces. Initially named nitosols (FAO-UNESCO, 1974), Nitisols soils have a high aggregate stability, friable consistence, high porosity, fair to good soil moisture storage capacity (5–15%) per unit volume, and an easy rooting. They may contain variable amounts of organic matter and be acid or neutral in reaction. As strongly weathered kaolinitic soils, their characteristic feature is a steady increase in clay content with depth to a maximum in the middle part of the profile, thereafter remaining uniform for some depth. Then there is a steady decrease in clay-sized particles with depth as the saprolite (alterite) becomes progressively less weathered (FitzPatrick, 1980). In many cases, intrinsic properties of the individual layers are similar to Ferralsols.

Nitisols are frequently derived from weathering products of basic rocks. Intense rapid weathering of these parent rocks results in deep profiles with a high clay content, a high amount of total and active iron, a low silt content, and a clay fraction dominated by kandites by minor amounts of gibbsite and other accessory minerals. Lateral (facies) relations observed in the landscape are controlled by topographic-hydrologic position, age of the landscape elements, and the degree of admixture with airborne materials, especially volcanic ash (Spaargaren, 1994). In undulating landscapes on basic and ultrabasic (mafic) rocks, Nitisols tend to occupy upper and middle slopes, merging into soils with Vertisols or vertic units of other major soil groups on the lower slopes and in valleys. This is the classic Lateritic-Margalitic sequence of Edelman, as described by Mohr et al. (1972). On volcanic landscape, Andosols occupy the upper slopes while Nitisols occur on the lower slopes. The Kikuyu Red Loam Nitisols of the Kenyan highlands are an example (Wakatsuki and Kyuma, 1988). On uplifted and remodeled plateau landscapes on old land surfaces, Nitisols occupy the slope positions whereas Ferralsols occur on flatter plateau parts. A classic example is the pattern of terra roxa estruturada (a Nitisol) and the terra roxa litigima or latossolo roxo (a Rhodic Ferralsol) of the basaltic plateaus of southern Brazil (Lepsch and Buol, 1988). On landscapes formed on limestone, Nitisols may occur as pockets and frequently in juxtaposition to shallower reddish soils (e.g., Luvisols, Chromic Cambisols). Examples are found in the Mediterranean region, for instance in Italy.

#### Acrisols

Acrisols (from *L. acris*, very acid; connotative of low base status) are characterized by a subsurface accumulation of low activity clays, a distinct clay increase with depth, and a base saturation (by 1 M  $\text{NH}_4\text{Oac}$ ) of less than 50%. These soils have been named red-yellow podzolic soils, podzólicos vermelho-amarelo distróficos a argila de atividade baixa, sols ferrallitiques fortement ou moyennement désaturés (CPCS, 1967), red and yellow Earths, latosols, and oxic subgroups of alfisols and ultisols. The latter have recently been redefined as kand- and kanhapl- great groups in the USDA's Soil Taxonomy (Soil Survey Staff, 1992).

Acrisols are common in tropical, subtropical and warm temperate regions, on Pleistocene and older surfaces where arid and humid periods have alternated. Acrisols are estimated to cover almost one billion ha worldwide (FAO, 1991), where about one-third is found in South and Central America and about 25% in southern and southeastern Asia.

On ancient shield landscapes in tropical regions, Ferralsols and Acrisols are a dominant consociation. The former soils, present on the flatter parts of the landscape or where sediments derived from weathered soils on uplands have been deposited,



are little affected by erosion. Acrisols, however, are often found in these landscapes on surfaces subject to erosion. For example, they are found on low hills covered by quartz and ironstone gravel, surrounded by pediments with Ferralsols, or on lower surfaces cutting into stable uplands with Ferralsols. Old alluvial fans in tropical regions often have Acrisols, with sesquisols in associated depressions.

Acrisols border a number of other kinds of soils with which they have linkages. These are mainly Ferralsols, Cambisols, and Arenosols. Acrisols are distinguished from Ferralsols by having a larger range in cation exchange capacity of the clay (24 versus 16 cmol(+) kg<sup>-1</sup>). Clay increase in Cambisols is not uncommon (Spaargaren, 1994). However, to qualify for a cambic horizon, the clay increase should not exceed the amounts set for the agric horizon. Thus, a continuum exists between Cambisols dominated by intermediate activity clays and Acrisols. Where the agric horizon is overlain by deep coarse textured horizons, linkages exist with Arenosols. By definition, the diagnostic agric horizon should occur within 2 m of the soil surface; below this depth these soils become Arenosols.

Most Acrisols in tropical regions are still under forest vegetation, which ranges from high canopy dense rain forest to open woodland or savanna. With the bulk of the nutrients concentrated in the vegetation, various forms of "slash-and-burn" have developed to cultivate these soils under traditional agriculture. Shifting cultivation (also known by such names as milpa, swidden, ladang, caingin, roza) therefore is the most common use of Acrisols. When the fallow period is sufficiently long to allow regeneration of the vegetation, this practice is probably the most sustainable form of agriculture on Acrisols. Continuous cultivation requires recurrent high input in terms of fertilizers and lime as well as other costly land management practices such as occasional ripping and deep plowing. Removal of the surface organic layer inevitably leads to significant yield decrease as the acid and aluminum toxic subsoil layers are exposed at the surface (Spaargaren, 1994).

Perennial crops like coffee, oil palm, rubber, cashew, mango and plantation of *Pinus caribaea* are well adapted to these soils. In South America, Acrisols are also common under savanna vegetation with a strong dry season. Some of these soils are placed under rainfed and irrigated agriculture after liming and fertilization. Rotation with annual crops in improved pastures should be recommended to maintain or improve the organic matter content.

#### Soils of wet dry (sub)tropical and warm temperate environments

Soils of savanna, steppe, and subtropical forest regions with dry periods are mainly characterized by profile varieties with cambic or argillic B horizons. Soils of the warm-temperate east coast margin climate (*Caf* in the Köppen-Geiger system) suggest the effects of strong leaching (podzolization) and ferralitization.

#### Lixisols

Lixisols (from *L. lix*, ly; connotative of high base saturation) are characterized by subsurface accumulations of low activity clays (cation exchange capacity of the clay is less than 24 cmol(+) kg<sup>-1</sup>) and moderate to high base saturation. They show a distinct clay increase with depth in the B horizon (agric horizon). The agric horizon in Lixisols often lacks clear illuviation features and most Lixisols are therefore characterized by a sharp clay increase occurring over a short distance. Root

penetration is usually good as there are no chemical barriers as in Acrisols. The absolute amount of exchangeable bases is generally not more than 2 cmol(+) kg<sup>-1</sup> fine earth due to the low cation exchange capacity. Many surface horizons of Lixisols are thin with a low amount of organic matter, especially in regions with pronounced dry seasons. Only under fairly humid conditions and/or low temperatures, as occur in tropical highlands, is there considerable accumulation of organic matter.

These soils have been named red-yellow podzolics, podzólicos vermelho-amarelo eutróficos a argila de atividade baixa, sols ferrugineux tropicaux lessives, and sols ferrallitiques faiblement desaturés appauvris (CPCS, 1967), red and yellow earths, latosols, and oxic subgroups of alfisols. These soils are found mainly in seasonally dry tropical, subtropical, and warm temperate regions and in areas with frequent additions of airborne dust, on Pleistocene and older surfaces. Lixisols cover an estimated area of about 435 million ha, of which more than half is found in Africa and one-quarter in South and Central America (FAO, 1991).

Lixisols are differentiated from the Nitisols by lacking a nitic horizon with its moderate to strong straight edged blocky structures and shiny ped faces that cannot be associated with illuviation cutans in thin sections. Lixisols may merge into Nitisols where the clay content of the soils is fairly high (more than 30%) and where the nitic horizon is located deeper in the soil. Lixisols are distinguished from Ferralsols by having a larger range in cation exchange capacity of the clay (24 cmol(+) kg<sup>-1</sup> versus 16 in the ferralic horizon). Where the agric horizon occurs under deep coarse-textured over-lying horizons, linkages exist with Arenosols. By definition, the diagnostic agric horizon should occur in these cases within 2 m of the soil surface. Below this depth, these soils become Arenosols.

On ancient shield landscapes in the tropics, Lixisols are found in association with Ferralsols. Lixisols tend to occur on slopes and surfaces subject to erosion. Old alluvial fans in tropical regions often have Lixisols, with sesquisols in associated depressions.

The natural vegetation of most Lixisols in the tropical and subtropical regions is savanna (see *Savanna*) and open woodland. Such areas with Lixisols are often used for extensive grazing. Because Lixisols are relatively well supplied with nutrients, they are frequently brought into cultivation. The low absolute levels of nutrients require maintenance of soil fertility on a regular basis and the low cation exchange capacity often dictates split-level applications to prevent fertilizer loss.

Continuous cultivation is possible but requires recurrent fertilization and/or liming, and occasional ripping and deep plowing. Destruction of the surface organic layer degrades soil structure that in turn leads to subsequent sealing and crusting which inhibits infiltration of surface water (Juo and Franzluebbers, 2003; van Wambeke and Nachtergaele, 2003). Significant yield decrease due to adverse surface soil characteristics is regularly recorded on these kinds of soils. Rotation of annual crops with improved pastures should be recommended to maintain or improve the organic matter content. Perennial crops like cashew, mango, citrus and other fruit trees are well adapted to these soils, although some supplementary irrigation may be required in the drier parts of the tropics and subtropics.

#### Luvisols

Luvisols (from *L. luo*, to wash; connotative of illuvial accumulation of clay) cover some 650 million ha worldwide (FAO, 1991) for the greater part in the tropical wet-dry (*Aw* climates merging

to *BSh* and *BWh*), humid subtropical (*Ca*, *Cs* climates), to subhumid temperate regions of central and Western Europe, the USA, Mediterranean regions, and Southern Australia. The dominant characteristic of Luvisols is the textural differentiation in the profile showing a surface horizon depleted in clay and an accumulation of clay in a subsurface argillic B horizon. These soils are further characterized by moderate to high activity clays and a low aluminum saturation. Luvisols form under aerobic conditions where there is free movement of water through the upper and middle parts of the soil. A distinct dry season is required for the soil to develop (FitzPatrick, 1980). The differentiation of most argillic B horizons seems to have taken place during the Holocene but some of these soils may be Pleistocene in age, especially those having formed under cooler climates.

The potential of the soils varies from moderate to good. Those soils with thick A horizons are included among the world's most productive soils. Because they occur under moist conditions, they are frequently used for mixed farming, dairying, or horticulture but wheat, oats, and maize can also be grown. Fertility is maintained by liming and fertilizer application. Erosion is a common feature and rigorous control methods must be maintained.

### Vertisols

Vertisols (from *L. verto*, to turn; connotative of turnover of surface soil) occur in large areas in tropical and subtropical regions with pronounced uni- or bimodal rainfall regimes. These clayey soils, dominated by the expanding smectite (montmorillonitic) clay minerals, develop wide, deep cracks during the dry season. During the rainy season, the cracks disappear when the land becomes fairly inaccessible due to a slippery surface. The soils are very difficult to work, being hard when dry and sticky when wet. They tend to be dark colored but have a low organic matter content. The apparent shrinking and swelling of the soil mass often results in small mounds and depressions at the surface and hence the name grumosol (from *L. grumus*, small earth mound). These soils are known by a variety of names from different regions viz. regurs (India), gilgai (Australia), adobe (USA, Philippines), badobes (Spain), tirs (Morocco, northern Africa), and margalite (Indonesia).

Vertisols generally have a high cation exchange capacity (CEC) on the order of 30 to 80 cmol(+) kg<sup>-1</sup> soil. The pH (H<sub>2</sub>O) is neutral or slightly alkaline in most cases. Base saturation (by NH<sub>4</sub>Oac) is usually high, also because many Vertisols show accumulation of lime in some form or another. Dominant cations are Ca and Mg, while in place Na also plays an important role. In coastal regions, Vertisols occur with high amounts of soluble salts and/or sulfides or sulfates present. Vertisols are set apart from other soils by the combination of having a vertic horizon, a high clay content throughout, and deep, wide cracks upon drying. Vertisols normally occupy the lower parts of the landscape, comprising nearly level to gently undulating piedmont, flood and coastal plains (Spaargaren, 1994). Associated vertic intergrades occur in relatively higher positions, comprising strongly sloping to moderately steep plateau, mesa and piedmont surfaces.

Land use in Vertisol areas ranges from very extensive (rough grazing, firewood production, charcoal burning) through small-holder post-rainy season crop production (millet, sorghum, cotton) to small-scale (rice) and large-scale irrigated crop production (cotton, wheat, sorghum). Several management practices are deployed to improve the water dynamics. Beds, ridges and furrows protect crops from waterlogging in the rooting zone

whereas contour cultivation and bunding improve infiltration. Vertisols are usually N-deficient due to the general low amount of organic matter. Other nutrients which may need correction are phosphorus and occasionally sulfur and zinc.

Charles W. Finkl

### Bibliography

- Anderson, J.M., and Ingram, J.S.I., 1993. *Tropical Soil Biology and Fertility*. Oxford: Oxford University Press, 240 pp.
- Arnold, R.W., Szabolcs, and Targulian, V.O., 1990. *Global Soil Change*. Laxenburg, Austria: International Institute for Applied Systems Analysis, 110 pp.
- Aubert, G., and Tavernier, R., 1972. Soil survey. In *Committee on Tropical Soils, Soils of the Humid Tropics*. Washington, DC: National Academy of Sciences, pp. 17–44.
- Babbington, B., 1821. Remarks on the geology of the country between Tellichery and Madra. *Trans. Geol. Soc. Lond.*, 5: 328–329.
- Beckmann, G.G., 1983. Development of old landscapes and soils. In *Soils: An Australian Viewpoint*. Melbourne, Victoria: CSIRO/Academic Press, pp. 51–72.
- Bigarella, J.J., and de Andrade, G.O., 1965. Contribution to the study of the Brazilian Quaternary. *Geol. Soc. Am. Spec. Pap.*, 84: 433–451.
- Bouma, J., Kuyvenhoven, A., Bouman, B.A.M., Luyten, J.C., and Zandstra, H.C., 1996. *Eco-Regional Approaches for Sustainable Land Use and Food Production*. Dordrecht, Netherlands: Kluwer, 505 pp.
- Bremer, H., 1975. Intermontane Ebenen, Prozesse der Flächenbildung. *Z. Geomorphol., Supplement-Band* 23: pp. 26–48.
- Bremer, H., 1981. Reliefformen und reliefbildende Prozesse in Sri Lanka. In Bremer, H., Schnutgen, A., and Spath, H., eds., *Zur Morphogenese in den feuchten Tropen*. Berlin: Gebrüder Borntraeger, pp. 7–183.
- Bridges, E.M., 1978. *World Soils*. London: Cambridge University Press, 128 pp.
- Buchanan, F., 1807. *A Journey from Madras Through the Countries of Mysore, Canara and Malabar*. London: East India Company, 3 Vols. See Vol. 2, pp. 426–437, pp. 440–441, p. 460, p. 559; Vol. 3, p. 66, p. 89, p. 251, p. 258.
- Büdel, J., 1957. Die "Doppelten Einebnungsflächen" in den feuchten Tropen. *Z. Geomorphol., NF*, 1: 201–228.
- Büdel, J., 1982. *Climatic Geomorphology*. Princeton, NJ: Princeton University Press, 443 pp. (translated by Fischer, L., and Busche, D.).
- Buol, S.W., Hole, F.D., and McCracken, R.J., 1980. *Soil Genesis and Classification*. Ames, IA: Iowa State University Press, 404 pp.
- Butler, B.E., 1959. *Periodic Phenomena in Landscape as a Basis for Soil Studies*. Canberra, ACT: CSIRO (Australia) Soil Publication No. 14, 20 pp.
- CPCS, 1967. *Classification des sols*. Grignon, France: ENSA, 87 pp.
- Deckers, J., Driessen, P., Nachtergaele, F., and Spaargaren, O., 2002. World reference base for soil resources. In Lal, R., ed., *Encyclopedia of Soil Science*. New York: Marcel Dekker, pp. 1446–1451.
- Dent, D., 1986. *Acid Sulphate Soils: A Baseline for Research and Development*. Wageningen, Netherlands: International Institute for Land Reclamation and Improvement, 204 pp.
- Duchaufour, P., 1977. *Pedology: Pedogenesis and Classification*. London: Allen and Unwin, 448 pp.
- Duchaufour, P., 1978. *Ecological Atlas of Soils of the World*. Paris: Masson, 178 pp.
- Eswaran, H., Ikawa, H., and Kimble, J., 1983. Oxisols of the world. *Proceedings of the Symposium on Red Soils. Beijing, China*.
- Fairbridge, R.W., ed., 1968. *The Encyclopedia of Geomorphology*. New York: Van Nostrand Reinhold, 1295 pp.
- Fairbridge, R.W., 1972. Climatology of a glacial cycle. *Quatern. Res.*, 2: 283–302.
- Fairbridge, R.W., 1976. Effects of Holocene climate change on some tropical geomorphic processes. *Quatern. Res.*, 6: 529–556.
- Fairbridge, R.W., and Finkl, C.W., 1980. Cratonic erosional unconformities and peneplains. *J. Geol.*, 8: 69–86.
- FAO, 1991. *World Soil Resources. An Explanatory Note on the FAO World Soil Resources Map at 1:25000000 Scale*. Rome, Italy: Food and Agriculture Organization of the United Nations, World Soil Resources Report No. 66, 58 pp.

- FAO-UNESCO, 1974. *Soil Map of the World*. Volume I. Legend. Paris, France: Unesco, 59 pp (updated with subsequent releases of different map sheets).
- FAO-UNESCO-ISRIC, 1990. *Soil Map of the World. Revised Legend*. Reprinted with corrections. World Soil Resources Report No. 60. Rome, Italy: FAO, 119 pp.
- Finkl, C.W., 1979. Stripped (etched) land surfaces in southern Western Australia. *Aust. Geogr. Stud.*, **17**(1), 33–52.
- Finkl, C.W., ed., 1982a. *Soil Classification*. Stroudsburg, PA: Hutchinson Ross, 391 pp.
- Finkl, C.W., 1982b. The geography of soil classification. *Quaestiones Geographicae*, **8**: 55–59.
- Finkl, C.W., 1988. Saprolite, regolith and soil. In Finkl, C.W., ed., *The Encyclopedia of Field and General Geology*. New York: Van Nostrand Reinhold, pp. 726–737.
- Finkl, C.W., 1995. New interpretations of chemical alterites on paleosurfaces. In Menon, J., ed., *Trends in Chemical Geology*, Vol. 1, Trivandrum, India: CSIR Research Trends, pp. 187–198.
- Finkl, C.W., and Churchward, H.M., 1973. The etched land surfaces of south-western Australia. *J. Geol. Soc. Aust.*, **20**(3): 295–307.
- Finkl, C.W., and Churchward, H.M., 1976. Soil stratigraphy in a deeply weathered shield landscape in south-western Australia. *Aust. J. Soil Res.*, **14**: 109–120.
- FitzPatrick, E.A., 1980. *Soils: Their Formation, Classification and Distribution*. London: Longman, 353 pp.
- Foth, H.D., and Schafer, J.W., 1980. *Soil Geography and Land Use*. New York: Wiley, 484 pp.
- Ganssen, R., and Hiidrich, F., 1965. *Atlas zur Bodenkunde*. Mannheim, Germany: Bibliographisches Institut AG, 85 pp.
- Goudie, A., 1973. *Duricrusts in Tropical and Subtropical Landscapes*. Oxford: Clarendon, 174 pp.
- Goudie, A.S., and Pye, K., 1983. *Chemical Sediments and Geomorphology*. London: Academic, 439 pp.
- Hassett, J.J., and Banwart, W.L., 1992. *Soils and Their Environment*. Englewood Cliffs, NJ: Prentice-Hall, 424 pp.
- Jenny, H., 1941. *Factors of Soil Formation*. New York: McGraw-Hill, 281 pp.
- Jenny, H., 1980. *The Soil Resource*. New York: Springer, 377 pp.
- Juo, A.S.R., and Franzluebbers, K., 2003. *Tropical Soils – Properties and Management for Sustainable Agriculture*. Oxford University Press, Oxford: 281 pp.
- King, L.C., 1962. *The Morphology of the Earth*. Edinburgh: Oliver & Boyd, 699 pp.
- Köppen, W., and Geiger, R., 1930. *Handbuch der Klimatologie*. Berlin, Germany: Gebrüder Borntraeger, 5 Vols.
- Lepsch, I.F., and Buol, S.W., 1988. Oxisol-landscape relationships in Brazil. In Beinroth, F.H., Camargo, M.N., and Eswaran, H., eds., *Proceedings of the Eighth International Soil Classification Workshop. Classification, Characterization and Utilization of Oxisols* (Rio de Janeiro, Brazil). Part I: Papers, pp. 174–189.
- McFarlane, M.J., 1976. *Laterite and Landscape*. London: Academic Press, 151 pp.
- McNeil, M., 1964. Lateritic soils. *Sci. Am.*, **211**(5), 96–102.
- Martius, C., Tiessen, H., and Vlek, P.L.G., 2001. The management of organic matter in tropical soils: what are the priorities? *Nutr. Cycl. Agroecosyst.*, **61**: 1–6.
- Mohr, E.C.J., van Baren, F.A., and van Schuylenborgh, J., 1972. *Tropical Soils, A Comprehensive Study of Their Genesis*. The Hague: Mouton-Ichtiar Baru-van Hoeve. [1959 edition, The Hague: Uitgeverij W. Van Hoeve, p. 498].
- Nachtergaele, F.O., Spaargaren, O., Deckers, J.A., and Ahrens, B., 2000. New developments in soil classification: world Reference Base for soil resources. *Geoderma*, **96**: 345–357.
- Nahon, D.B., 1991. *Introduction to the Petrology of Soils and Chemical Weathering*. New York: Wiley, 313 pp.
- Oliver, J.E., and Fairbridge, R.W. eds., 1987. *The Encyclopedia of Climatology*. New York: Van Nostrand Reinhold, 986 pp.
- Oosterbaan, R.J., 1982. Natural and social constraints to polder development in Guinea-Bissau. In *Polders of the World*. Vol. 1, Wageningen, Netherlands: ILRI, pp. 141–160.
- Paramanathan, S., and Eswaran, H., 1981. Morphology of Oxisols. In Theng, B.K.G., ed., *Proceedings of the Symposium on Soils with Valuable Charge, New Zealand*.
- Pons, L.J., and van Breeman, N., 1982. Factors influencing the formation of potential acidity in tidal swamps. In Dost, H., and van Breeman, N., eds., *Proceedings of the Bangkok Symposium on Acid Sulphate Soils*. Wageningen, Netherlands: ILRI Publication No. 31, pp. 37–51.
- Retallack, G.J., 1990. *Soils of the Past*. London: HarperCollins Academic, 520 pp.
- Richards, P.W., 1996. *The Tropical Rain Forest*. New York: Cambridge University Press, 600 pp.
- Sanchez, P.A., 1994. Tropical soil fertility research: towards the second paradigm. *Trans. 15th World Congr. Soil Sci. (Acapulco, Mexico)*, **1**: 65–88.
- Sanchez, P.A., and Logan, T.J., 1992. Myths and science about the chemistry and fertility of soils in the tropics. In Lal, R., and Sanchez, P.A., eds., *Myths and Science of Soils of the Tropics*. Madison, WI: Soil Science Society of America, SSSA Special Publication 29, pp. 35–46.
- Smith, D.E., 1967. Equatorial and tropical climates. In Fairbridge, R.W., ed., *The Encyclopedia of Atmospheric Sciences and Astrogeology*. New York: Reinhold, pp. 364–367.
- Spaargaren, O.C., ed., 1994. *World Reference Base for Soil Resources*. Rome, Italy: FAO, 161 pp.
- Spath, H., 1981. Bodenbildung und Reliefentwicklung in Sri Lanka. In Bremer, H., Schnutgen, A., and Spath, H., eds., *Zur Morphogenese in den feuchten Tropen*. Berlin: Gebrüder Borntraeger, pp. 185–238.
- Soil Survey Staff, 1992. *Soil Taxonomy*. Washington, DC: U.S. Government Printer.
- Steila, D., 1976. *The Geography of Soils*. Englewood Cliffs, NJ: Prentice-Hall, 222 pp.
- Thomas, M.F., 1974. *Tropical Geomorphology*. New York: Wiley, 332 pp.
- Trewartha, G.T., 1954. *An Introduction to Climate*. New York: McGraw-Hill, 402 pp.
- Twidale, C.R., and Campbell, E.M., 1993. *Australian Landforms*. Adelaide: Gleneagles, 560 pp.
- Twidale, C.R., and Campbell, E.M., 2005. *Australian Landforms: Understanding a Low, Flat, Arid and Old Landscape*. New South Wales, Australia: Rosenberg, Dural Delivery Centre. 336 pp.
- Wakatsuki, T., and Kyuma, K., 1988. Chemical characteristics of Kenyan Nitisols. In Hirose, S., ed., *Agriculture and Soil in Kenya, A Case Study of Farming Systems in the Embu District and Characterization of Volcanogenic Soils*. Tokyo: Department of Land Development, Nihon University, pp. 130–133.
- van Wambeke, A., and Nachtergaele, F., 2003. *Properties and Management of Soils of the Tropics*. Rome, Italy: FAO Land and Water Digital Media Series, CD ROM.
- Van Wambeke, A., Eswaran, H., Herbillon, A., and Comerma, J., 1990. Oxisols. In Wilding, L.P., Smeck, N.E., and Hall, G.F., eds., *Pedogenesis and Soil Taxonomy. Part II. The Soil Orders*. Amsterdam: Elsevier, pp. 325–350.
- Williams, M.A.J., 1985. Pleistocene aridity in tropical Africa, Australia and Asia. In Douglas, I., and Spencer, T., eds., *Environmental Change and Tropical Geomorphology*. London: Allen and Unwin, pp. 219–233.

## Cross-references

Alisols  
Biomes and their Soils  
Ferralsols  
Geography of Soils  
Nitisols  
Thionic or Sulfidic Soils  
Weathering Systems in Soil Science

## TRUNCATED SOIL

A soil that has had part or all of its solum removed by erosion, or by human activity. In the Northern Hemisphere soils (technically paleosols) truncated by Wisconsinian ice are found





<http://www.springer.com/978-1-4020-3994-2>

Encyclopedia of Soil Science  
Chesworth, W. (Ed.)  
2008, XXVI, 902 p., Hardcover  
ISBN: 978-1-4020-3994-2