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Soil Science

Soil, Soil Classification, Problem Soils, Land Evaluation, Land Resource Information Systems

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Lecture notes on

SOIL

SOIL CLASSIFICATION

PROBLEM SOILS

LAND EVALUATION

LAND RESOURCE INFORMATION SYSTEMS

Prepared by

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PART 1

SOIL

By L.P. van Reeuwijk

CONTENTS

1. INTRODUCTION

To scientists of various disciplines "soil" has divergent meanings. Some geologists regard soil as 'dirt that obscures the rock'. Agronomists may look at soil as substratum that is cultivated to produce a crop. To civil engineers it may be material to build with or to build on. The place of soil (or pedosphere, pedon is Greek for soil) is illustrated in Fig. 1.

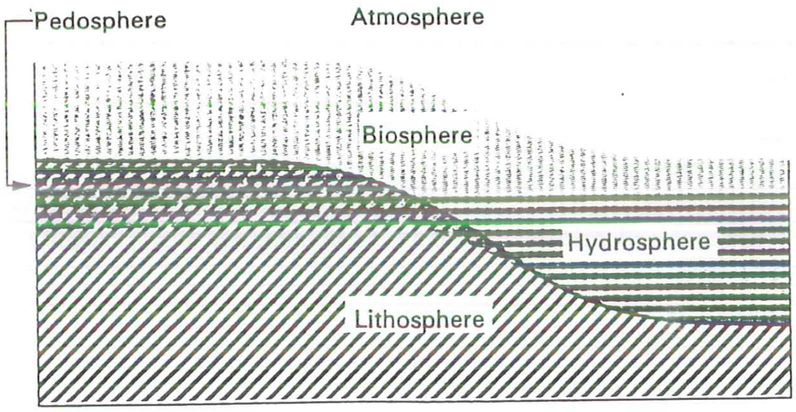


Fig. 1. The place of the pedosphere. (After Schroeder, 1984)

The soil scientist or pedologist examines the soil more closely than the other scientists. He (or she) does not only look at the material alone, but will dig a pit to observe on the profile the build-up of the different layers or horizons, the colours, structure, texture, porosity, consistence, and other relevant features. In addition, he/she considers the environment of the soil, e.g. the setting in the landscape, the climate, the vegetation as well as the relation with neighbouring soils. With the collected information, including data from laboratory analysis of samples taken from the profile, he/she will try to classify the soil, elucidate its genesis, and draw up a list of assets and constraints, to be stored these days as much as possible in digital form.

In this course several topics relating to the formation, properties, use and mis-use of soils will be discussed. This entails a treatise of the soil components and their organisation, the soil chemical and physical properties, particularly in relation to water, and the soil forming factors.
Proper use of soils is greatly helped by proper characterisation. The resulting data allow soil classification: a naming of the soil and implying a generalised indication of its qualities and constraints. Soil survey is aimed at determining the distribution of soils in a certain area.

The quality of soils and land as a production factor can be adversely effected by both natural and human actions. Such land *degradation* can occur as:

- loss of soil by wind or water erosion
- pollution
- accumulation of salts (in top and/or subsoil)
- deterioration of soil structure, e.g. sealing, compaction (physical degradation)
- loss of plant nutrients; acidification (chemical degradation)
- Rapid mineralisation (oxidation) of organic matter in topsoil. Deterioration of soil life (biological degradation)

Causes of soil and land degradation

- *Natural* causes are, for instance, floods, fires, earth quakes or earth slides.
- *Human* actions causing degradation are, amongst others, improper production practices caused by economic factors, poor management (neglect or ignorance), poor land use planning.

Whatever the causes are, good knowledge of the soil and its role as an essential part of our environment will help to avoid and/or to restore soil and land degradation. This implies that:

- we need basic knowledge of soils to understand specific problems and possible remedies;
- we must apply soil classification and survey, because it gives us an estimate of the magnitude of both problems and opportunities as well as their distribution;
- we must be aware and recognize soil degradation and problem soils,
- therefore, we must apply land evaluation and land use planning to avoid soil and land degradation as these are often the result of
    - wrong land use
    - wrong soil management

Thus, obviously, land use and soil management have to be adapted to the specific capacities and incapacities of an area of land.

If successful, such measures may lead to:

- increased *sustained production* on suitable land
- *protection* of less suitable land under cultivation
- *alternative forms* of land use

Collectively, these activities may be termed *SOIL QUALITY MANAGEMENT.*

2. COMPOSITION OF SOILS

2.1 DEFINITION

A soil can be defined as *"a natural body on which plants (can) grow"*. A more descriptive definition is: *"Soil is the transformation product of mineral and organic substances on the earth's surface under the influence of environmental factors operating over a long time and having defined organisation and morphology. It is the growing medium for plants and basis of life for animals and mankind. As a space-time system soil is four-dimensional.*

An "individual" soil type (see Fig. 2), also referred to as "pedon", is a three-dimensional body, a column in the pedosphere (i.e. from the parent rock to the atmosphere). Its surface is variable and ranges from 1-10 m^2 depending on the soil variability. Reversely. the pedosphere can be considered an assembly (continuum) of individual pedons.

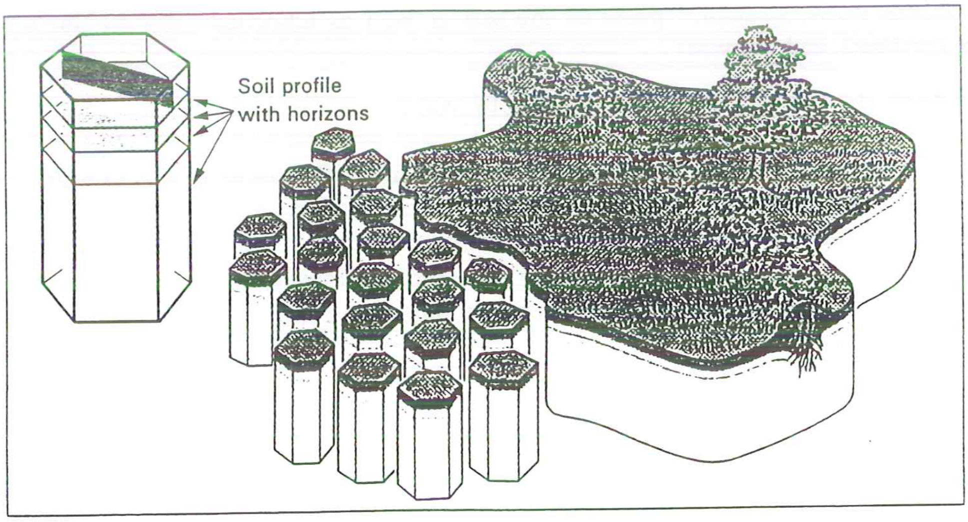


Fig. 2. Division of part of the pedosphere into pedons.

The soil profile is a 2-dimensional vertical section through the pedon or pedosphere (see Fig. 3). Usually in these profiles various layers or *soil horizons* are visible. They result from the various soil forming processes (see Section 5).

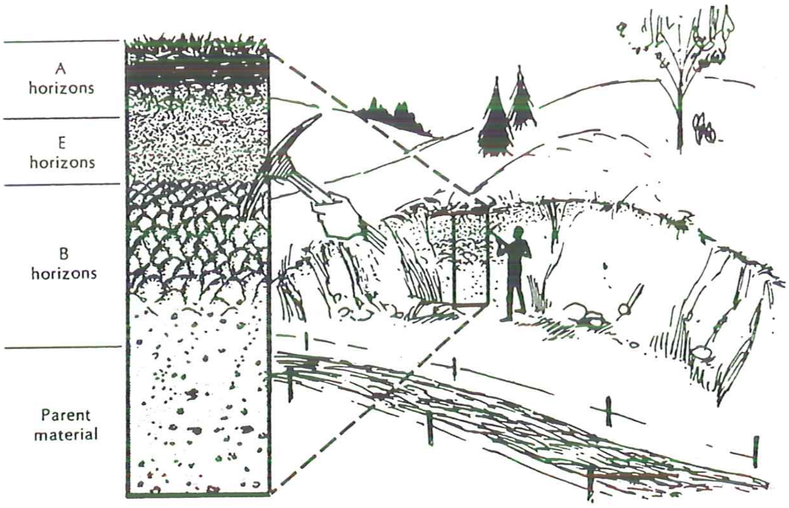


Fig. 3. Soil profile of a road cut. The A horizons are usually higher in organic matter and darker in colour. The B horizons often have distinctive structures.

2.2 COMPOSITION

Soil essentially consist of variable proportions of four major components:

a. Mineral matter
b. Organic matter
c. Water
d. Air

Ad a. Mineral matter consist of rock fragments and grains of individual minerals. Note: Minerals are usually crystalline, minor amounts of material occur in an amorphous state. Mineral matter constitutes the backbone of most soil materials and occurs in several size fractions e.g., gravel, sand, silt, and clay.

Ad b. Organic matter is a collective noun for the many forms in which organic material occurs in the soil: plant residues (roots, leafs), humus, organic acids, sugars, etc. Organic matter often acts as a cementing agent binding mineral grains and resulting in a spatial structure: the soil structure.

Ad c. Water occurs in the structural pores of the soil as well as adsorbed onto the surface of the solid particles (mineral and organic).

Ad d. Air will enter a soil when pores are vacated by water.

In most soils the solid fraction consists usually for more than 90% of mineral matter and, therefore, less than 10% of organic matter. Exceptions are soils very rich in organic matter such as peat soils. It is clear that water and air compete for the pore space. Thus, the proportion of the components may vary considerably. A rule of thumb is that for agronomic purposes, an "ideal" soil has a volume proportion of about 2:1:1 of solid:water:air (corresponding with 50% solid and 50% pore space on volume basis). Figure 4 illustrates this.

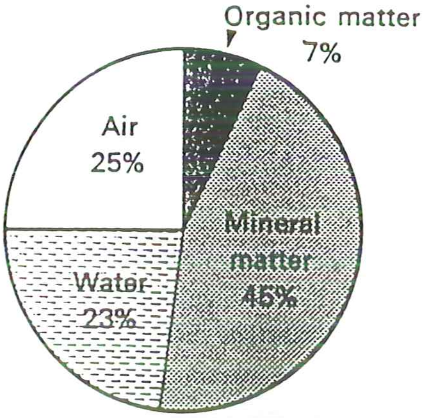


Fig. 4. Composition of a grassland soil (in vol.%). (After Schroeder, 1984)

2.2.1 Mineral Matter

2.2.1.1 Composition

The mineral composition (and consequently the chemical composition) of soils is to a large degree determined by the parent material in which and from which the soil is formed. This parent material, in turn, is inherited from parent rock, either in situ or after transport from elsewhere (or of mixed origin).

The composition of earth's crust (lithosphere) has the following levels of organization:

Earth's crust / Rocks / Minerals / Ions / Elements

Since rocks are composed of minerals, a brief discussion of the relevant minerals is given first.

By now, some 3000 minerals have been identified in the earth's crust. Fortunately, for the soil scientist only some tens of minerals are of importance as the bulk is made up of relatively few. Furthermore, the chemical composition of the bulk of the earth's crust is fairly simple as ten of the 90 natural elements make up more than 99% of the lithosphere.

Table 1. Abundance of elements in the earth's crust.

Element	Symbol		% by weight
Oxygen	O		46.6
Silicon	Si		27.7
Aluminium	Al		8.1
Iron	Fe		5.0
Calcium	Ca		3.6
Sodium	Na		2.8
Potassium	K		2.6
Magnesium	Mg		2.1
Titanium	Ti		0.6
Hydrogen	H		0.6
		Total	99.7

This does not mean, of course, that other elements are unimportant. Many of them are essential to life on earth e.g. carbon (0.05%), phosphorus (0.1%), and trace elements.

2.2.1.1 Minerals

Minerals, which can be defined as *"natural phases"*, are classified primarily according to their chemical composition. The main grouping is done according to the predominant anion, since minerals with the same anion usually have more properties in common than minerals with the same cation: carbonates are more similar than sodium minerals. This complies with the classification of inorganic compounds in general. The classification with some prominent examples is given in Table 1. Most minerals are crystalline by nature, only a few are amorphous e.g., opal (SiO_2)
As appears from Table 1, the silicates are by far the dominant group: more than 90% of the lithosphere consists of silicates, and this holds for soils as a whole too. In addition, all clay minerals, very reactive components of soils, are silicates. An important aspect of the silicates, as part of the parent material in which soils are formed, is their relative Si content. Silicates high in Si content (acidic of felsic silicates) are more stable with respect to weathering than those relatively low in Si (basic or mafic minerals).

A typical composition of the mineral part of soils would be more than 80% silicates (including clay minerals), the remainder consisting of calcite (lime), gypsum and salts (in the drier areas), and oxides and hydroxides of Fe en Al. In humid tropical areas the oxides are generally more prominent than in cooler climates. The brown and reddish colours of soil often result from thin coatings of Fe (hydr)oxides. Dark colours usually result from organic matter.

Primary minerals are those that are inherited from the primary parent rocks e.g., feldspars, micas, quartz, olivine. *Secondary minerals* are the result of transformation processes (chemical weathering, recrystallisation). Examples are the clay minerals, (hydr)oxides of Fe, Al and Mn, lime, gypsum, and salts.

TABLE 1. The classification of minerals

Group	Example
Elements	Sulphur, copper, silver, gold, graphite, diamond
Sulphides	Pyrite: FeS_2
Oxides	Hematite: Fe_2O_3; rutile: TiO_2; quartz: SiO_2
Hydroxides	gibbsite: $Al(OH)_3$; goethite: $FeOOH$
Halides	halite: $NaCl$; sylvite: KCl
Nitrates	soda niter: $NaNO_3$
Carbonates	calcite: $CaCO_3$; dolomite: $CaMg(CO_3)_2$
Sulphates	gypsum: $CaSO_4.2H_2O$
Phosphates	apatite: $Ca_3(PO_4)_3$
Silicates	Olivine: $(Mg,Fe)_2SiO_4$; Mica: e.g. $KAl_2(AlSi_3O_{10})(OH)_2$; feldspar: e.g. $KAlSi_3O_8$ Clay minerals: e.g. kaolinite, illite, vermiculite, chlorite, montmorillonite

2.2.1.2 Rocks

The concept rock can be defined as *"an association of minerals"*. Rocks are usually composed of several minerals but mono-mineralic species occur as well, e.g. marble (calcite) and quartzite (quartz). A rock is not necessarily consolidated: dune sand, for instance, is a rock in the strict sense.

Rocks are classified with respect to their

1. Origin
2. Structure
3. Chemical-mineralogical composition

Three main groups are distinguished with respect to *origin:*

1. *Igneous rocks*
   These form by solidification of molten magma, they make out 95% of the volume of the lithosphere (of 16 km thick), but only 25% of the surface of the earth (see Fig. 5).

2. *Sedimentary rocks* or *Sediments*
   These form by deposition of transported debris of other rocks (which may include other sedimentary rocks). To the sediments belong also deposits of organic or chemical origin. Because they are spread out relatively thinly, they make out 75% of the surface of the earth but only 5% of the volume of the lithosphere (Fig. 5).

3. *Metamorphic rocks*
   These are formed by alteration (metamorphism) of other rocks under the influence of heat and/or pressure. The alterations can be in structure or in mineral composition or both. As for their extension they are represented in above-mentioned figures according to their igneous or sedimentary origin. The mutual relationship of rocks is illustrated in Figure 6.

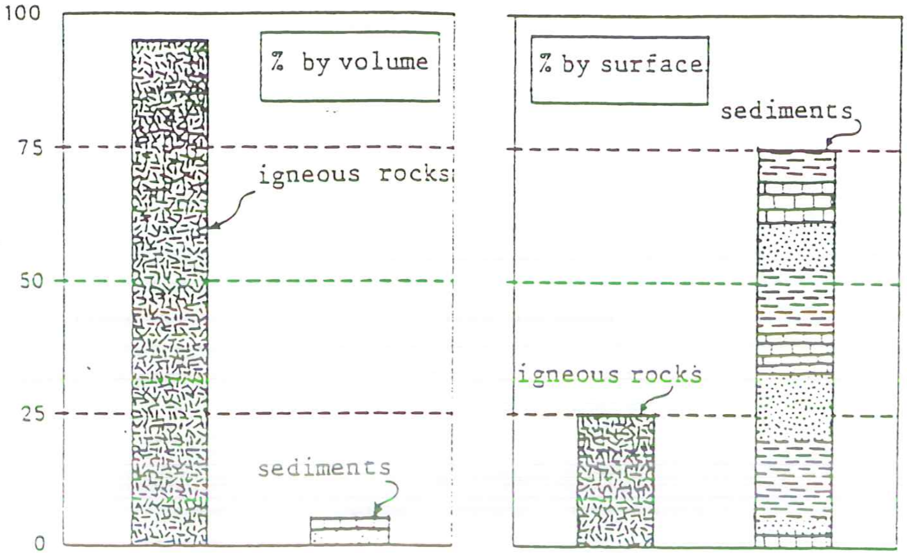


Fig. 5. Ratio of the occurrence of igneous and sedimentary rocks.

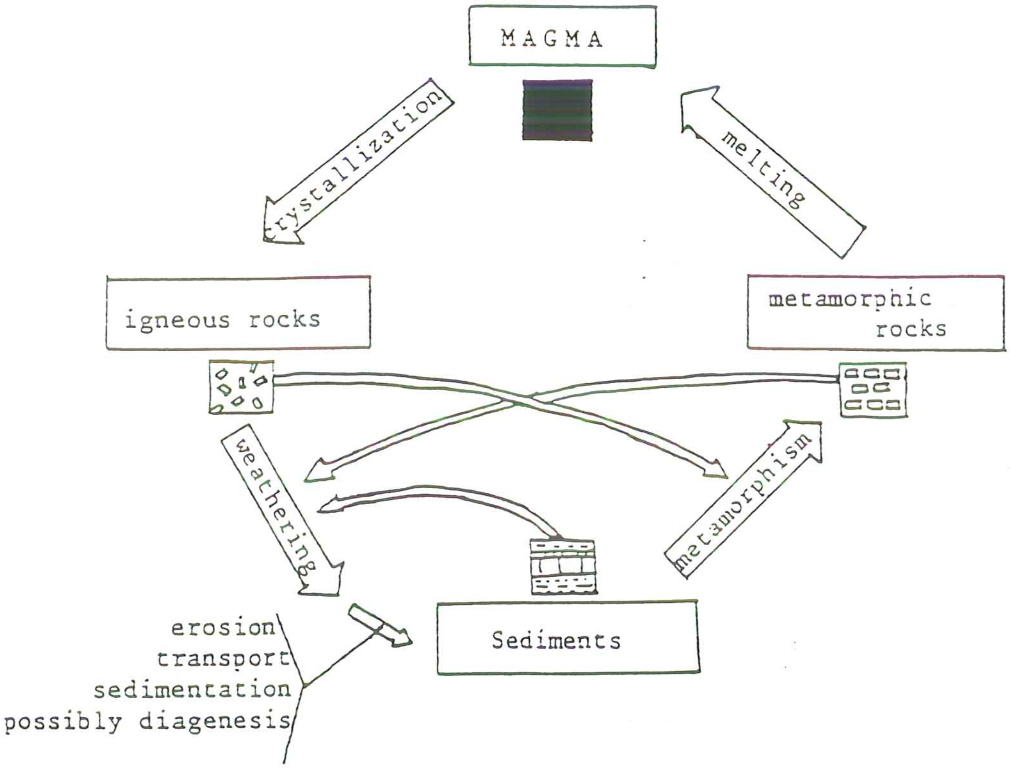


Fig. 6. The "cycle" of rocks.

1. Igneous rocks

Igneous rocks are, in turn, subdivided in three groups by origin: a. *plutonic* (formed at depth); b. *hypabyssal* or *dyke rocks* (formed in "pipes" to the surface of the earth); c. *volcanic* or *effusive rocks* (formed after outflow of lava over the surface).
The classification of the igneous rocks according to their origin is given in Table 2 and their mineralogical composition in Figure 7.

Table 2. The Classification of Igneous Rocks.

	Plutonic (at depth)	*Hypabyssal* or *Dike* (dikes and sills)	*Volcanic* or *Effusive* (lava flows)
acidic	Granite	Graniteporphyr, Quartzporphyr	Liparite (= rhyolite)
	Syenite	Syeniteporphyr, Orthophyre	Trachite
	(Granodiorite)		(Dacite)
	Diorite	Dioriteporphyr, Porphyrite	Andesite
basic	Gabbro	Gabbroporphyr, Dolerite, Diabase	Basalt
	Ultra basic		Basalt

An acid magma will produce granite as plutonic rock consisting of acid feldspars (albite, oligoclase, microcline), mica and much quartz. The corresponding dike rock (or hypabyssal rock) is called granite porphyr or quartz porphyr expressing its composition and structure. The effusive analogue (i.e. when acid magma flows out on the surface) is called liparite or rhyolite. A less acid magma will produce syenite as plutonic rock, the lower SiO_2 content results in the near absence of quartz. The feldspars are of a less acid type and the increase in mafic minerals results in a much darker coloured rock. Corresponding dike rocks are called syenite porphyr or orthophyre. The more basic the rocks are, the more difficult it becomes to identify them visually.

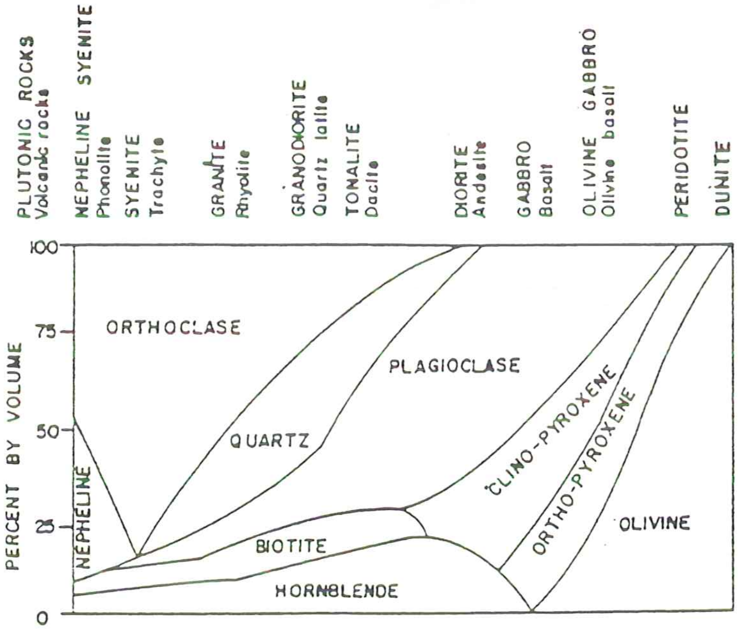


Fig. 7. Mineralogical composition of common igneous rocks. Volcanic rocks with names in lower-case letters consist of fine-grained minerals.

2. Sediments

The weathering of rocks and minerals results in loose masses of crushed, crumbled and mellowed materials as well as material in solution. These materials are being eroded constantly, washed into the rivers, carried into lakes, seas and oceans and, by sedimentation or precipitation forming sedimentary deposits. Crushed materials can also be taken up by winds thus giving rise to eolian sediments. In this way it became possible that sediments cover about 75% of the surface of the earth (making out only 5% of the mass of the lithosphere, see Fig. 5).

By *origin* the sediments can be subdivided into three main groups: 1. *clastic,* 2. *organic,* and 3. *chemical* sediments.

Clastic, detrital, or fragmental sediments

By accumulation of mineral or rock fragments (gravel, sandstone, claystone, shale). Some examples are:

Alluvial sediments
Alluvial or by water deposited sediments can occur in many forms. The most important are the gravel, sand and clay deposits of marine (sea), fluvial (river) or lacustrine (lake) origin. The detrital material can be of a widely varying origin, they can be residues of all types of rocks including other sediments as well as eroded material from soils.
Consolidation of sediments often occurs. Shale is formed by consolidation and compression of clay deposits. Sandy deposits may be consolidated and cemented by calcareous, silicious or iron compounds to form sandstones. Conglomerates and breccia are consolidated rocks of rounded-off gravel and angular, hardly transported gravel respectively.

Glacial sediments
Glaciers are responsible for particular types of clastic sediments. They pick up all kinds of material in their forward move. Upon melting and retreat toward the arctic regions or toward higher altitudes of mountains, glaciers drop their loads, forming sediments varying in size from very fine particles of clay and silt to gravel, stones and boulders. Collectively, these sediments are known as *glacial till* or *glacial* drift. The area over which this debris is deposited is known as the *terminal* or *end moraine.* The deposits show hardly any stratification.

Eolian sediments
Air-borne or eolian sediments are loess, cover sands and dunes. Particles smaller than 50 µm are taken in air suspension and move nearly as fast as the wind and are deposited in vegetated areas. As a consequence, loess is very porous. Particles between 2 mm and 50 µm move by *saltation* (jumping) over rather short distances: the cover sands are the result of this process. Particles of 2-4 mm are being rolled over the ground and their displacement is limited. These processes are very important in areas without vegetation (desert, glacial periods). The result of these processes is that the air-borne sediments have a very uniform grain size and are far less stratified than river deposits.

Colluvial sediments
Colluvium is a type of sedimentation common in mountainous and hilly areas. The material is removed by gravity and deposited at the bottom of slopes. This process is very pronounced if the deeper layers are frozen: *solifluction.*

Pyroclastic sediments
Volcanoes may produce *pyroclastic sediments* (or *tephra)* in the form of volcanic ash, sand, tuff, etc. which, after the material was blown into the air, were deposited on land or in water.

Organic sediments

Organic sediments can be produced by the residues of *plants* and of crustaceous *animals* (skeletons). *Carbonate* rocks or limestones originate from corals, foraminifers and crinoids whereas *siliceous* sediments are formed by radiolaria and diatoms.
Carboniferous sediments result from accumulation of plant residues, e.g. peat, coal.

Chemical sediments

These result from the precipitation of compounds from solution caused by changes in physicochemical environment. For instance in lagoons with a continuous influx of sea water together with high evaporation, the solubility product of several salts may be exceeded so that precipation occurs (evaporites), e.g. calcite (limestone), halite, gypsum, dolomite. In volcanic thermal areas, near geysers, precipitates of calcite, sulphur and limonite may occur.

Of all the sediments, it appears that 99% is made out by *shale, sandstone* and *limestone* and that the other sediments are quantitatively of minor importance. This is illustrated in Figure 8, which also shows the ratio of these three major sediments.

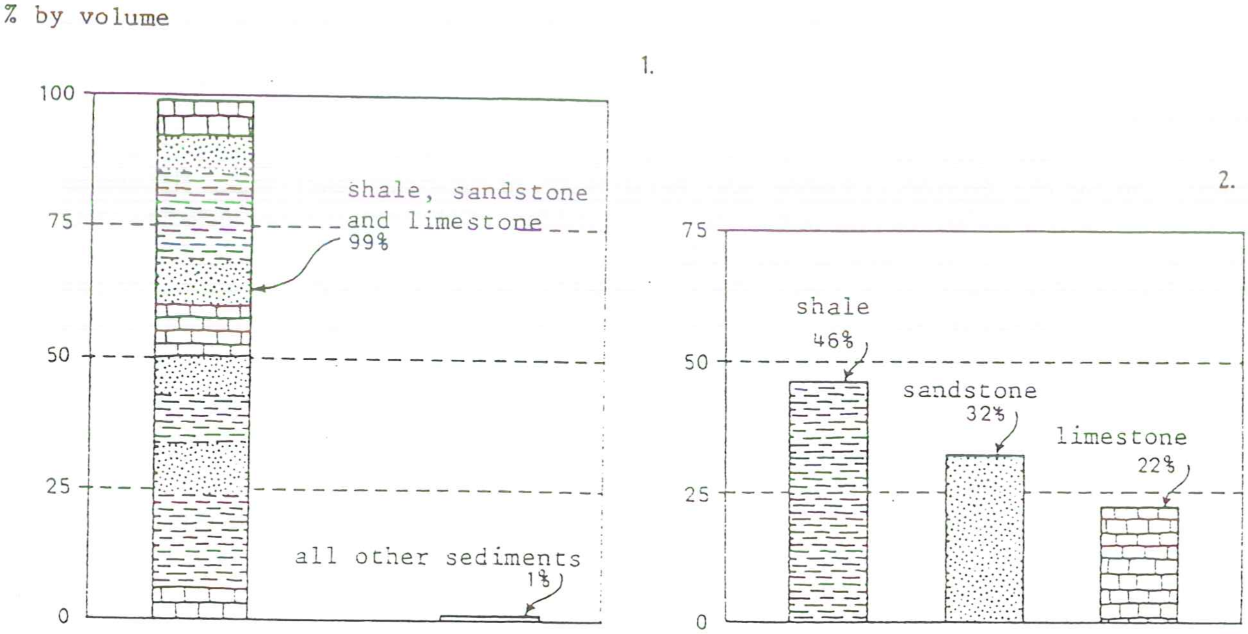


Fig. 8. Block diagram showing the dominance of shale, sandstone and limestone in sediments (1), and the ratio of these three rocks (2).

3. *Metamorphic rocks*

When igneous or sedimentary rocks are subjected to heat or pressure (whereby usually heat is produced), they may undergo changes in structure and/or in mineralogical composition. This process is called *metamorphism.* In its weakest form, the minerals may become somewhat viscous and accommodate to the pressure resulting in elongated crystals and banding of the rock. In this way gneiss may be formed from granite. Often, however, the metamorphism is so complete that the original rock may not even by recognized.

When the changes in sedimentary rocks are restricted to dehydration, compaction, and cementation these processes are not considered metamorphism but are called *diagenesis.* Some examples of metamorphism are given below.

Sandstone	→	Quartzite	
Granite	→	Gneiss	
Arkose	→	Gneiss	(*arkose* = feldspar-rich sand)
Ultra basic	→	Serpentine	
Limestone	→	Marble	

Clay → Claystone → Shale → Slate → Phyllite → Schist
(diagenesis) →|← (metamorphism)

2.2.1.3 Particle-size distribution

Not only the chemical/mineralogical composition but also the physical composition of soils is very important. Many physical properties are strongly influenced by the particle size distribution of the mineral grains *(texture)* and their kind and degree of aggregation *(structure)*. The size of soil particles ranges widely from boulders and stones to minute clay particles. Although already more than a century ago C.G. Hopkins advocated the standardisation of grain sizes in texture analysis, unfortunately at this moment several systems of particle size classes are still in use (see Fig. 9). Some more detail is found in Figure 10.

Because the soil texture is a parameter of soil quality, classification schemes have been set up to give names to texture classes. The most widely used is the USDA triangle given in Fig. 7. With some experience, a "guesstimate" of the textural class can be made by "feeling" the soil between the fingers, or, when very finely textured, between the teeth. Often, however, one is fooled by micro-aggregation of particles (pseudo-silt!) leading to too "coarse" estimates. Therefore, always have your guestimates checked by laboratory analysis!.
Soil texture is not only used for agronomic purpose but also for civil engineering.
Some important properties of the main soil particle separates (i.e. not mixed) are listed as follows:

Sand (0.05 - 2 mm)

- not sticky when wet
- low water holding capacity
- easily drained and aerated
- little or no nutrient adsorption

Silt (0.002 - 0.05 mm)

- slightly sticky when wet
- fair water holding capacity
- liable to compaction
- low nutrient adsorption

Clay (< 0.002 mm)

- sticky and plastic when wet, hard when dry
- high water holding capacity + retention
- high nutrient adsorption
- colloidal properties (i.e. can be dispersed in water)

Remark: It is important to distinguish the terms "*clay*" and "*clay minerals*". The former is a *size* fraction (all particles < 2 μm) in which all sorts of minerals can occur. The latter is a specific group of secondary (phyllo)silicates, usually (but not always) of such a small particle size that they occur in the clay fraction.

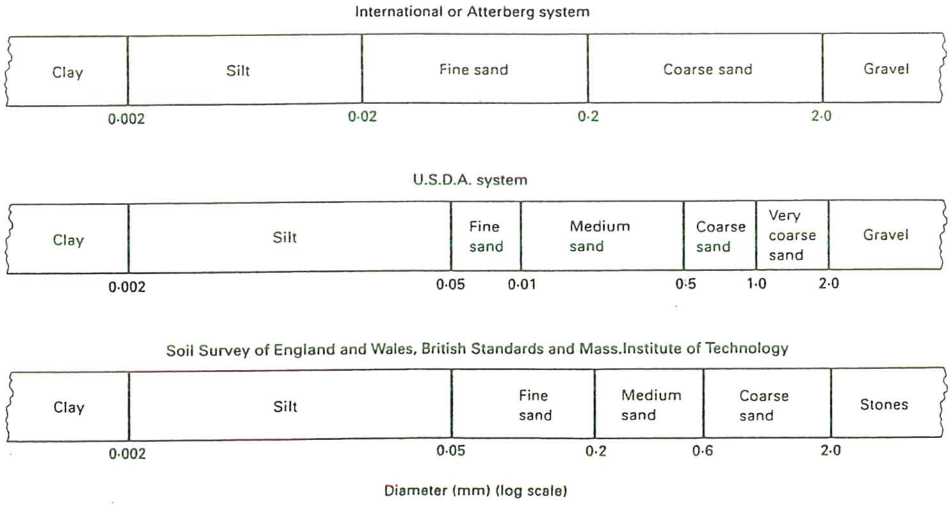
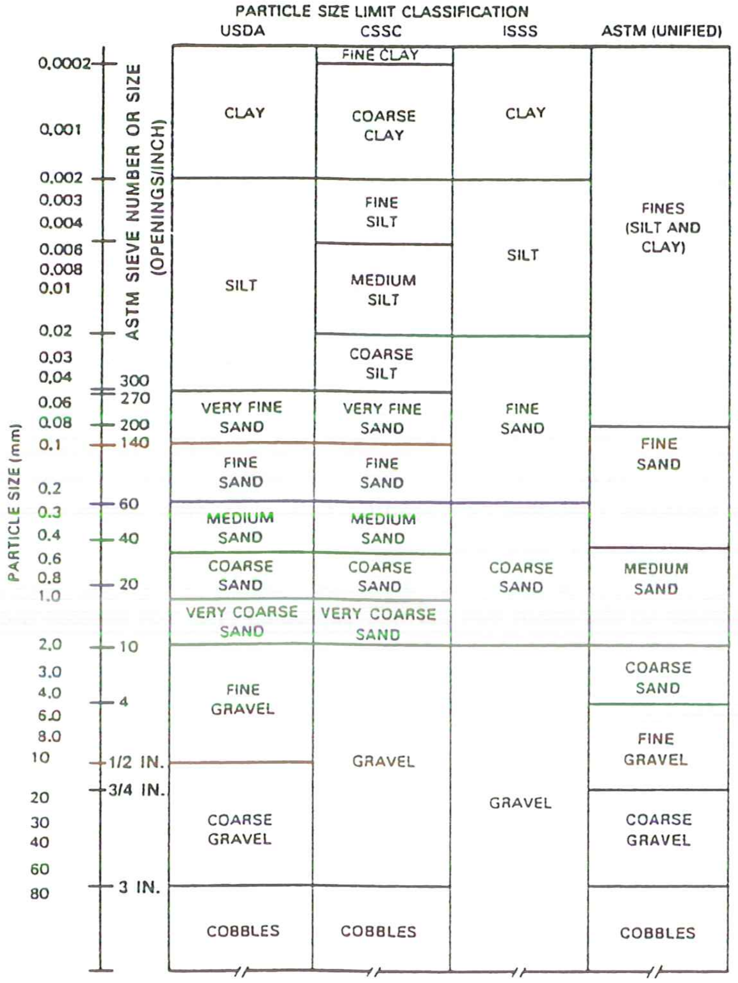


Fig. 9. Various particle-size classes



USDA—U.S. DEPARTMENT OF AGRICULTURE, (SOIL SURVEY STAFF, 1975)
CSSC—CANADA SOIL SURVEY COMMITTEE, (McKEAGUE, 1978)
ISSS—INTERNATIONAL SOIL SCI. SOC. (YONG AND WARKENTIN, 1966)
ASTM (UNIFIED)—AMERICAN SOCIETY FOR TESTING & MATERIALS (ASTM, D-2487, 1985a)

Fig. 10. Particle-size limits according to several classification schemes.

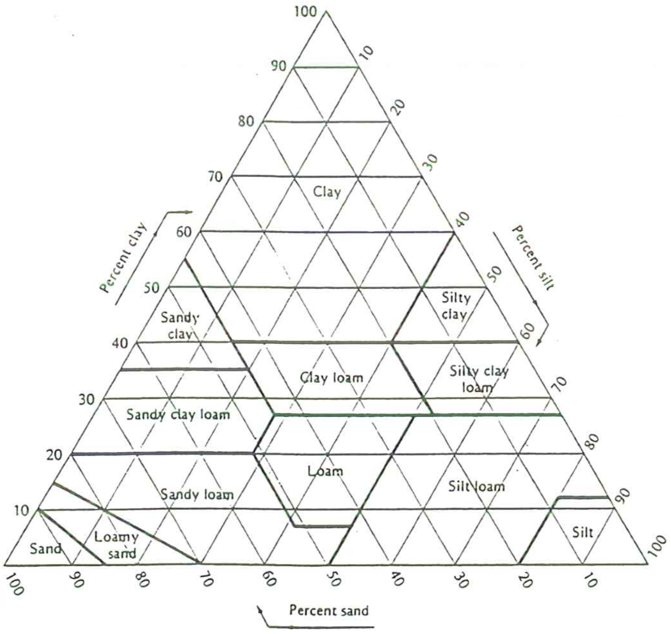


Fig. 11. Major soil textural classes according to the contents of sand, silt, and clay (USDA system).

Surface area

The size of the soil particles is not only important for the physical behaviour of soils but also for the chemical reactivity. Most reactions in the soil are liquid-surface reactions: adsorption and desorption of ions, water and gas, weathering reactions, etc. Physical behaviour such as swelling result from combined *physicochemical* reactions (wetting, desorption, etc.). Therefore, the surface area determines to a large extent the intensity of these reactions. The general rule is that the smaller the particle, the larger its surface area (on weight basis) is. Thus, clay has a (much) higher surface area than sand. This relationship is shown in Figure 12.

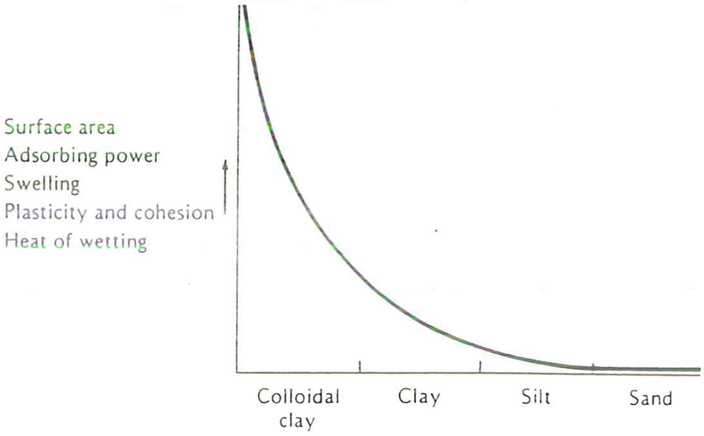


Fig. 12. Relationship between particle size and surface area.

2.2.2 Organic Matter

The content of organic matter in soils is generally relatively low: typically 1-6% by weight. However the (usually positive) influence on soil physical and agronomic properties is very large. Organic matter occurs in soils as:

- Living organisms (soil flora and fauna)
- Accumulation of (partly) decayed plant residues
- Accumulation of "stable" products of decomposition: humus.

Some of the most important roles of organic matter are:

- Cementing mineral particles into aggregates (improvement of soil structure)
- Source of plant nutrients (N, P, S)
- Adsorbent of nutrients (humus has a high cation exchange capacity)
- Increasing water holding capacity
- Source of energy (food) for micro-organisms
- Plays a role in soil formation (e.g. complexation reactions)

The small living organisms (e.g. worms, ants, termites, plant roots) play a very important role in the soil in that they promote porosity and homogenisation of the soil. Some species representing the soil fauna are given in Figure 13.

Organic matter is a rather dynamic component of the soil because of the relatively rapid decomposition and transformation processes, particularly at higher temperatures: in the cool climates accumulation is generally much stronger than in the warmer climates where soil organic matter contents tend to be low. Decomposition is effected by micro-organisms (bacteria, fungi) that use organic matter as food. Thus decomposition are usually oxidation reactions. Even under waterlogged conditions such oxidation can take place: anaerobic bacteria use the soil oxides as source of oxygen (e.g. Fe_2O_3, NO_3, SO_4).

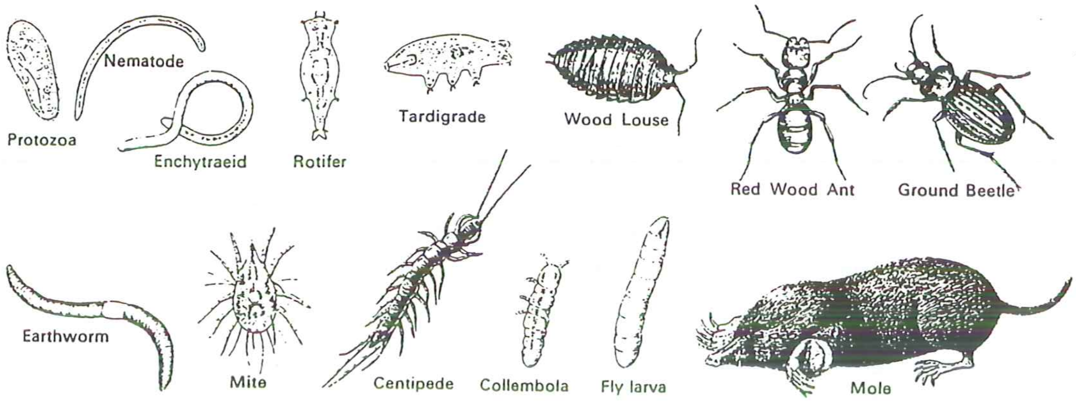


Fig. 13. Representatives of soil fauna. (After Muckenhausen).

The generalized reaction is as follows

$$\underset{\text{organic matter}}{(CH_2O)} + O_2 = CO_2 + H_2O + \text{energy} \tag{1}$$

Under extremely anaerobic conditions reduction to methane (CH_4) may occur.

Note: The reverse of Reaction (1) is the synthesis of organic matter, This occurs in plants under the input of energy from sunlight, an essential process for the existence of life: *photosynthesis* (see also Fig. 15).

Because of the many reactions taking place, the components of organic matter are usually large molecules of a complex nature. Well-known components are fulvic acids, humic acids, humins, hydroxybenzoic acid, but also more simple acids such as citric acid and oxalic acid do occur. Some components of humic substances and reactive groups are given in Figure 14.

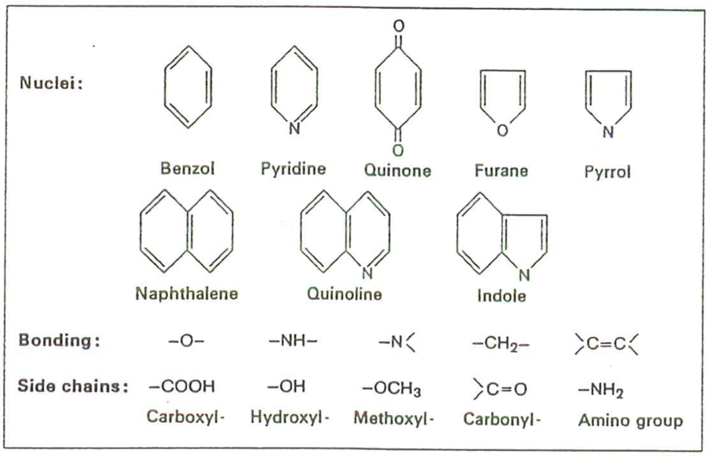


Fig. 14. Main constituents of soil humic substances and reactive groups. (From Schroeder, 1984).

It is interesting to follow the cycle of carbon and the role of the soil in this. This is illustrated in Figure 15. This drawing also clearly indicates the role of soil and plants to store carbon thereby reducing the CO_2 content of the atmosphere and consequently reducing the *greenhouse effect* of the atmosphere leading to *global warming*.

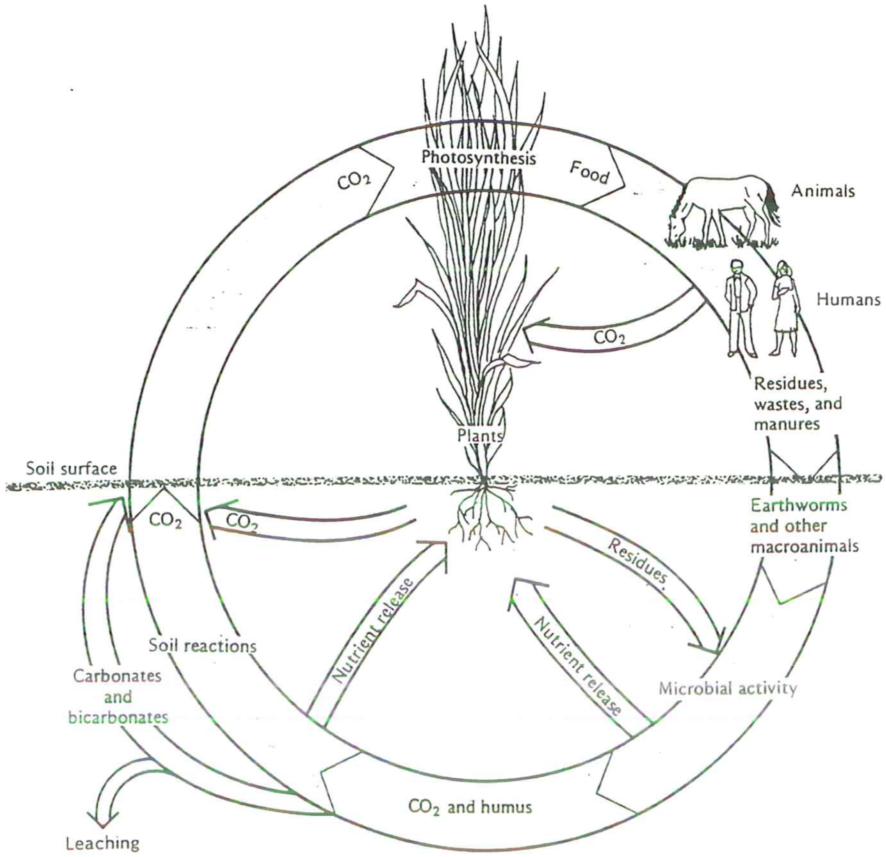


FIGURE 10.3
Transformation of carbon commonly spoken of as the *carbon cycle*, the *biocycle*, or *life cycle*. Plants assimilate CO_2 from the atmosphere into organic compounds using energy from the sun. Man and other higher animals obtain energy and body tissue from plant products and return wastes and residues to the soil. Macro- and microorganisms digest these organic materials, releasing nutrients for plants and leaving CO_2 and humus as relatively stable products. Carbonates and bicarbonates of Ca, Mg, K, etc., are removed in leaching, but eventually the carbon returns to the cycle in the form of CO_2. The total CO_2 is released to the atmosphere where it is again available for plant assimilation. This biocycling illustrates how carbon is the focal point of *energy* transformations and makes possible the continuity of life on Earth.

Fig. 15. The carbon cycle. (From Brady, 1990).

2.2.3 Soil water

Without water, little happens in the soil. The main functions of water are:

- essential component of the various substances and organisms in the soil
- takes part in almost all chemical reactions in the soil
- transporting agent (nutrients, weathering products, salts, small particles, gasses).

It is clear that in a dry soil most processes and all life comes to a standstill (compare the situation in the desert). As soon as water is added to a soil many processes are started up and flora and fauna will resume their activities, unless they died by the drought: then a next generation will have to take their place.

2.2.4 Soil air

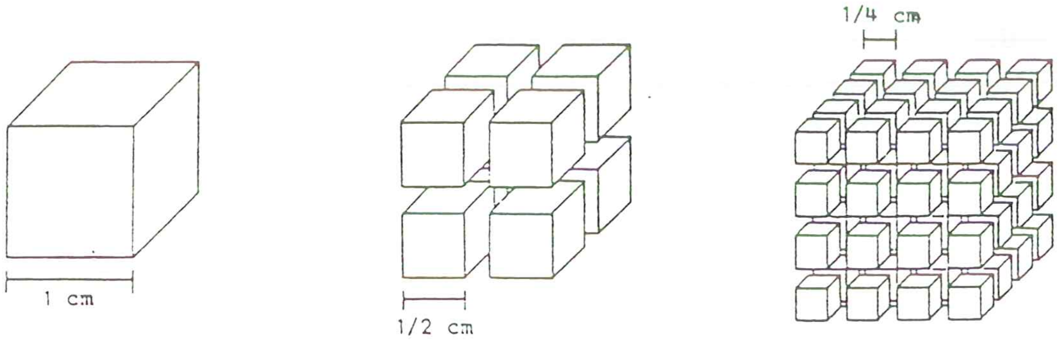
Pore volume in the soil not occupied by water will be filled with air. The composition of soil air usually differs considerably from that in the atmosphere. This is mainly due to the respiratory activity of flora and fauna. During the day, plant roots excrete CO_2 leading to a concentration of CO_2 in soil air that may be a factor 10 to 100 higher than that in air (in air: $P_{CO_2} = 10^{-3.5}$ bar, or about 0.03%). This goes particularly for topsoils where root activity is highest. With increasing depth the concentration carbon dioxide decreases.
For the same reason, the oxygen concentration is usually lower than in the atmosphere: typically about half (in atmosphere: about 20%). Under reducing conditions (waterlogging!) the concentration may drop to zero due to the respiration of micro-organisms.

3. WEATHERING OF ROCKS AND MINERALS

To better understand soil formation (Section 5) it is useful to discuss a fundamental aspect first, namely the weathering of rocks and minerals from which the parent material of soils result.

Weathering is the process of disintegration and decomposition of rocks and minerals under atmospheric conditions. It transforms the original rock (parent rock) into the parent material (the French synonyms are: *roche mère* and *roche fille).* Two natural forces, *physical* and *chemical,* operate in this process.

Physical weathering is the process whereby rocks and minerals are cracked, crumbled, crushed and mellowed. The result is a material that varies in size from boulders to very fine particles. The chemical composition of the rocks and minerals is not changed, however. Hence, physical weathering is a mechanical disintegration of the solid mass; it causes a change in shape, form and size of the rock and mineral debris, without changing its chemical composition. However, the surface area of the material is increased (see Fig. 16) which means that the subsequent chemical weathering (surface reactions!) is intensified.



cube with edge = 1 cm
volume = 1 cm^3
total surface = 6 cm^2

8 cubes with edge = ½ cm
volume = 1 cm^3
total surface = 12 cm^2

64 cubes with edge = ¼ cm
volume = 1 cm^3
total surface = 24 cm^2

Fig. 16. Relation between volume, particle size and total surface area. It is clearly shown that decrease in particle size by fragmentation leads to an increase in surface area and consequently to an increase in intensity of the chemical weathering.

Chemical weathering is the decomposition of the complex substances of the rocks and minerals into compounds with a lower free energy content. It should be realized that the important primary minerals (e.g. feldspars, amphiboles, pyroxenes, micas, quartz) have all been formed at much higher temperatures and pressures than those prevailing at the surface of the earth. Therefore, they are to a greater or lesser extent unstable under atmospheric conditions. Some of the decomposition products dissolve and may be leached out, others evolve as gases (CO_2). The bulk of them, however, remains in place.

It is evident that weathering is a destructive process. However, parallel with this *destruction,* a building-up process or *synthesis* may take place. Often certain constituents, set free during weathering, react to give rise to secondary minerals. Such minerals are, for instance, calcite, gypsum, phosphates, oxides, and a very important group of silicates: the clay minerals.

Biological weathering is the physical and chemical action of organisms (plants and animals).

3.1 PHYSICAL WEATHERING

The agents of physical weathering are: temperature, water, wind, glaciers and crystallization.

3.1.1 Temperature (fluctuations)

Expansion upon heating and shrinkage upon cooling is very active in heterogeneously composed rocks. In this case not all parts of the rocks expand and contract at equal rates, and unequal stresses are formed in the rock. The result is a disintegration of the rocks. This process is possible because of the fact, that rocks are poor conductors of heat. The temperature at the surface can consequently rise to very high values (55°C), especially in deserts and high mountains, while they cool considerably during the night. The temperature differences between the surface of the rock and a few centimetres within may cause a crevice to form between two zones of temperature.
As a consequence of this temperature fluctuation, on the northern hemisphere the southern exposures of mountains have more rock fragments than the northern exposures. On the southern hemisphere this is just the opposite.

Ice formation and thawing is very active in the disintegration of rocks in the climate of northern areas. If water solidifies, it expands.

3.1.2 Water

Water is a very active and powerful agent in the mechanical disintegration of rocks. The transporting power of a stream varies as the sixth power of its velocity:

$$p = f(v^6)$$

This means that when a stream is going to move twice as fast, its transporting power is increased 64 times! An idea about the carrying capacity of water is given by the following data: A stream moving 15 cm/sec will carry fine sand; one moving 30 cm/sec will carry gravel; 120 cm/sec: stones of about 1 kg; 250 cm/sec: stones of 60 kg; 900 cm/sec: blocks of 320 tons.

The abrasive power of rock loads carried in water is very impressive in mountain streams during the seasons with high rainfall. Large boulders are smashed into small fragments by the very frequent collisions. The pieces become increasingly rounded and smooth. It is evident that the various rocks (being different in hardness, cohesion and structure) behave differently. As an example one can say that a beach with a slope of 0.2% (~20 cm per 100 m) needs a transport way of 11 km to disintegrate granite blocks of 20 cm diameter into particles of 2 cm diameter; in the case of gneiss 5-6 km is needed to do this; in the case of sandstone only 1.5 km.
The result of this action of water is that disintegrated rock mass is accumulated at the base and build up slopes of loose fragments. This is known as *talus.* The largest blocks accumulate at the foot of the talus slope. With further disintegration and decomposition, masses of loose rock are moved by water and wind to expose the bedrock to the forces of weathering. This is the so-called *normal geologic erosion.* The process is gradual and needs a long time. Nevertheless, geologic erosion levels the rough contours of the landscape. The final state is a condition of low relief or nearly a level plain; this process is called *peneplanation* (formation of peneplains).

3.1.3 Wind

The action of wind, as an agent of weathering, is similar to that of water. The abrasive power of dust, sand and gravel carried by winds is well-known. It grinds away the surface of rocks and removes cleavage materials from talus slopes.

3.1.4 Glaciers

Glaciers vary strongly in length, thickness and width. They move as solid masses of ice and do not require a definite stream channel, as rivers do. The glaciers advance over hills and valleys and across rives and lakes. It is quite clear that in this way much material can be taken up and ground. As climate changed, the ice receded, and much of the load remained behind. These deposits are called *moraines.* The material, thus being products of physical weathering, is very heterogeneous.

3.1.5 Crystallization

A very peculiar agent in physical weathering is the *crystallization* of saturated salt solutions. In all places, where super-saturated solutions can be formed, the volume of which are smaller than the sum of the volumes of the saturated solution and the separated crystals. Crystallization can produce a considerable pressure on the rocks, so that they can be split into smaller units.
This process can be found specifically in arid areas, where high salt concentrations are common and where super-saturated solutions can be formed on cooling or evaporation of water.

3.2 CHEMICAL WEATHERING

The driving force behind weathering reactions is the energy (usually heat) that is released during the reactions. The unstable mineral has a higher free energy content as compared with its weathering products and it will, therefore, decompose.

Yet, unstable minerals may persist for considerable time as a result of a high *threshold energy* or *activation energy* which is the energy that is required to trigger off a reaction (see Fig. 17). An increase in temperature by the sun can often supply this energy. Such minerals are said to be *resistant* to weathering, e.g. quartz.

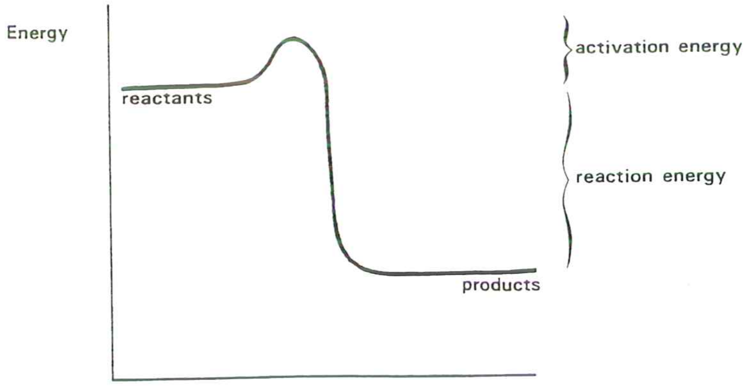


Fig. 17. Diagrammatic representation of energies involved in weathering reactions.

In almost all weathering reactions water plays an important role. In fact, very little would happen at the surface of the earth if water were absent (compare the moon).
The most prominent processes, active in chemical weathering are: *hydration, hydrolysis, (dis)solution, oxidation, reduction, carbonation* and *chelation.* These will be discussed in the ensuing sections.

3.2.1 Hydration

The combination of solids with water to form hydrates is called *hydration.* In soils, oxides of iron and aluminium are the main compounds undergoing hydration e.g:

$$\underset{\text{hematite}}{Fe_2O_3} + H_2O = \underset{\text{goethite}}{2\ FeOOH} \qquad (2)$$

Gibbsite is a hydrated mineral of Al_2O_3. Muscovites easily convert into hydro-muscovites. Generally the volume increases upon hydration. This volume increase ranges from a very few percents up to

160%, as in the alternation of Al_2O_3 to gibbsite. Commonly the increase is smaller than 50%. In arid areas where temperature may be very high, and humidity very low in the dry summer season, *dehydration* can occur. Consequently iron oxides, which are normally hydrated and then have a yellowish brown or red colour, become dehydrated and turn the colour into brown.

3.2.2 Hydrolysis

Perhaps the most important weathering action of water is through hydrolysis. The hydrolysis is the result of the dissociation of water into H^+ and OH^--ions and consists of the separation of salts, containing the anion of a weak acid and/or the cation of a weak base, into the corresponding base and acid. Aqueous solutions of salts of a strong acid and weak base (e.g. $AlCl_3$) react consequently acid and solutions of salts of a weak acid and strong base (e.g. silicates) react alkaline.

Many silicates are difficultly soluble combinations of the various weak silicic acids or aluminosilicic acids and the strong bases of the alkali and alkaline earth metals (Na, K, Ca, Mg). Their hydrolysis differs from that of soluble salts: in case of silicates the hydrolysis proceeds at the surface of the solid substances. The metal ions of the surface are exchanged by H-ions of the water. The metal ion occupies an OH-ion and gives the solution an alkaline reaction.

Because the feldspars are very important in the characterization of soils, the *hydrolytic weathering* will be exemplified by the following equations:

$$\underset{\text{orthoclase}}{KAlSi_3O_8} + HOH = HAlSi_3O_8 + KOH \qquad (3)$$

The formation of KOH can be demonstrated by an aqueous suspension of orthoclase powder which reacts alkaline (pH > 10). The aluminosilicate formed is unstable, is hydrolysed further and changes either into a clay mineral, kaolinite (Reaction 4), or into silica and aluminium hydroxide (Reaction 5).

$$2\ HAlSi_3O_8 + HOH = \underset{\text{kaolinite}}{Al_2Si_2O_5(OH)_4} + 4\ SiO_2 \qquad (4)$$

$$HAlSi_3O_8 + HOH = \underset{\text{gibbsite}}{Al(OH)_3} + 3\ SiO_2 \qquad (5)$$

Note: More accurately, silicic acid is (temporarily) formed: $H_4SiO_4 = SiO_2 + H_2O$.

Ca-feldspars weather also into kaolinite, as demonstrated in the reaction schemes (6) for anorthite.

$$\underset{\text{anorthite}}{CaAl_2Si_2O_8} + 3\ HOH = \underset{\text{kaolinite}}{Al_2Si_2O_5(OH)_4} + Ca(OH)_2 \qquad (6)$$

The bases, formed in several of the above mentioned reactions can react with the CO_2 dissolved in the water and form carbonates of potassium, sodium (for Na-feldspars) or calcium, of which the latter is insoluble and an important source of calcium carbonate (lime) in soils:

$$Ca(OH)_2 + CO_2 = CaCO_3\downarrow + H_2O \qquad (7)$$

It can be stated that the above given reaction schemes can be applied under conditions of very permeable materials and good drainage.

The hydrolysis equilibrium, established at the surface of the decomposing minerals, is in the soil continually disturbed by several processes:

(1) neutralisation of the OH-ions by H-ions of carbonic acid and other acids
(2) formation of insoluble substances, e.g $CaCO_3$, Fe, and Al oxides, silica, clay minerals
(3) transport of soluble components, including organo-metallic complexes

Consequently, the hydrolysis attacks ever renewing surfaces. On the other hand, the reaction is impeded by the formation of coatings that may precipitate on the surface of the weathering

3.2.3 Dissolution

Dissolving a solid in water changes solid materials into separate ions that distribute themselves evenly. The solvent action of water on minerals is usually slow but it can be increased considerably when certain acids or salts are present. It becomes then one of the most powerful factors in the decomposition of rocks and minerals. Calculated throughout the entire globe 40 tons of material per km^2 is removed in solution annually. The 40 tons are composed of: 21 tons $CaCO_3$; 8 tons $CaSO_4$; 3.5 tons NaCl; 3 tons SiO_2; 2.5 tons of alkali carbonates and sulphates; 1.5 tons $MgCO_3$; 0.5 ton Fe_2O_3.

The dependence of the solubility of some substances on the pH is given in Figure 18.

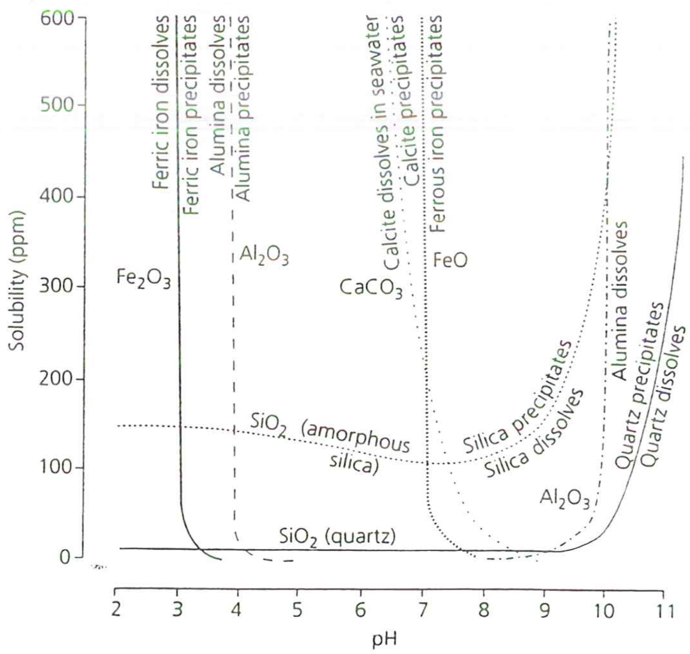


Fig. 18. Solubility diagram of a number of compounds in soils.

3.2.4 Carbonation

One of the most important acids, that increases the solvent action of water considerably is *carbonic acid* (H_2CO_3) The concentration of carbonic acid depends on the CO_2-gas pressure of the atmosphere. The higher the CO_2 content, the higher its partial pressure, and the more CO_2 dissolves in water to form carbonic acid. In dense tropical rain forests the CO_2 content of the atmosphere can be much higher than in the air above bare surfaces. In the neighbourhood of volcanoes the CO_2 content can also be higher. Rain water contains a 15 to 40 times higher concentration of CO_2 than air. Certain natural mineral waters contain large quantities of CO_2.

The action of CO_2 is especially effective on limestones. Water containing carbonic acid dissolves $CaCO_3$ readily in the form of calcium bicarbonate:

$$CO_2 + H_2O = H_2CO_3 \tag{8}$$

$$CaCO_3 + H_2CO_3 = Ca(HCO_3)_2 \tag{9}$$

The effect of the partial CO_2 pressure of the atmosphere on the dissolution of the limestone is shown in Table 3. The table shows that the solubility of $CaCO_3$ increases strongly with increasing CO_2-content of the atmosphere.

Table 3. Solubility of $CaCO_3$ in water of 25°C at various CO_2-contents of the air.

% CO_2 in the air	0.031*	0.33**	1.6**	4.3**	10	100
$CaCO_3$ (mg/L)	52	117	201	287	390	900
pH of solution	8.3	7.7	7.2	6.9	6.7	6.1
pH without $CaCO_3$	5.7	5.2	4.9	4.7	4.5	4.0

* our atmosphere ** in soil air

3.2.5 Oxidation and Reduction

For convenience these two processes are discussed together since the one is the reverse of the other:

- *Oxidation* is the loss of electrons by an ion or element (to be accepted by another which is thus reduced)
- *Reduction* is the acceptance of electrons by an ion or element (donated by another which is thus oxidized).

Oxidation

Because oxygen is the most usual acceptor of electrons the process is called *oxidation.* An example illustrating the process is the widely occurring oxidation of *ferrous* iron to *ferric* iron:

$$Fe^{2+} = Fe^{3+} + e^- \qquad (10)$$

this electron to be accepted by oxygen:

$$O_2 + 4\ e^- = 2\ O^{2-} \qquad (11)$$

this O^{2-} radical will immediately combine with H^+ from water:

$$O^{2-} + HOH = 2\ OH^- \qquad (12)$$

In reaction (10) Fe^{3+} is a very insoluble ion which will precipitate by hydrolysis:

$$Fe^{3+} + 3\ HOH = Fe(OH)_3 + 3\ H^+ \qquad (13)$$

By simple dehydration this ferric hydroxide may transform to goethite:

$$Fe(OH)_3 - H_2O = \underset{\text{goethite}}{FeOOH} \qquad (14)$$

or further to hematite:

$$2\ \underset{\text{goethite}}{FeOOH} - H_2O = \underset{\text{hematite}}{Fe_2O_3} \qquad (15)$$

These reactions show that oxidation of ferrous iron produces more protons (H^+) than hydroxyls (OH^- and is, therefore, an acid reaction. The overall reaction can be condensed to:

$$2\ Fe^{2+} + \tfrac{1}{2}\ O_2 + 5\ H_2O = 2\ Fe(OH)_3 + 4\ H^+ \qquad (16)$$

which means that for each oxidized ferrous ion *two protons* are produced. If the soil does not contain *buffering* compounds such as $CaCO_3$ or other weatherable minerals then the pH will become lower.

Oxidation affects especially the Fe(II)- and Mn(II)-containing minerals. Ferrous minerals will decompose and form iron hydroxides, such as goethite. Manganese (II) oxidizes into brownish-black coloured substances, such as MnOOH and MnO_2. Since many minerals contain ferrous and manganous ions, the weathering is usually accompanied by a tanning of the material.

A prominent reaction occurs in reclaimed marine sediments containing pyrite (FeS_2). Upon lowering the water table air (O_2) enters the soil leading to the oxidation of pyrite into goethite and sulphuric acid:

$$\underset{\text{pyrite}}{4\,\mathrm{FeS_2}} + 15\,\mathrm{O_2} + 10\,\mathrm{H_2O} = \underset{\text{goethite}}{4\,\mathrm{FeOOH}} + 8\,\mathrm{H_2SO_4} \qquad (17)$$

This overall reaction can be divided into three sub-reactions (18), (19), and (20). It is possible that the reaction series stops after (17), (18) or (19) depending on the conditions under which the weathering takes place.

$$2\,\mathrm{FeS_2} + 7\,\mathrm{O_2} + 2\,\mathrm{H_2O} = 2\,\mathrm{FeSO_4} + 2\,\mathrm{H_2SO_4} \qquad (18)$$

$$4\,\mathrm{FeSO_4} + \mathrm{O_2} + 2\,\mathrm{H_2SO_4} = 2\,\mathrm{Fe_2(SO_4)_3} + 2\,\mathrm{H_2O} \qquad (19)$$

$$2\,\mathrm{Fe_2(SO_4)_3} + 8\,\mathrm{H_2O} = 4\,\mathrm{FeOOH} + 6\,\mathrm{H_2SO_4} \qquad (20)$$

It is evident that reaction (20) is not an oxidation but a hydrolysis. It is most common in nature that these two reaction types, oxidation and hydrolysis, occur together.

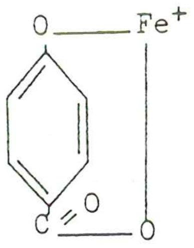
Reduction

Reduction in soils occurs under anaerobic conditions (water-logging!) when microbes can no longer use atmospheric oxygen for their metabolism which implies the oxidation of organic matter (digestion of food). They will then resort to compounds that can be reduced under the production of oxygen (accept electrons). Under such conditions, ferric compounds may be reduced to ferrous compounds according to the reverse of reactions (10) and (16). Because the reduction of ferric compounds consumes protons, this reaction is basically an alkaline reaction (if not buffered by acids such as carbonic acid).

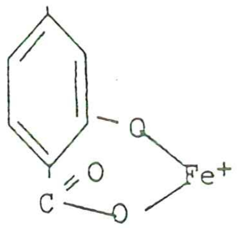
Oxidation and reduction processes are prominent in *gley* in alternatingly dry and wet soils.

3.2.6 Complexation

Many organic compounds in the soil, formed by degradation of plant remains, have the capacity to form strong bondings with metal ions. Such acids are for example the fulvic and humic acids (containing hydroxyl and carboxyl groups) but also acids of smaller molecular size such as oxalic and citric acid. The formed complexes may be insoluble and precipitate or soluble and be transported in solution. As an example, the complexes of Fe with para- (A) and ortho- (B) hydroxybenzoic acid can be represented as follows:



(A)



(B)

Such complexes in which the metal ions are taken up in a structure like by a pair of pincers are called *chelates.* The complexing capacity greatly enhances chemical weathering since on the one hand the organic compounds may remove metal ions (Ca, Mg, Al, Fe, Mn) directly from primary minerals, and on the other may remove (hydr)oxides from the soil system as soluble compounds. Together with their activity as acid (protons), organic acids thus have a dual function in soils.

3.2.7 Weathering stability

The weatherability of the various minerals is very different. Some are very stable, e.g. quartz, others offer only slight resistance, e.g. calcite. By distinguishing classes of weatherability, it is possible to arrange the minerals in the following order:

very unstable : gypsum, calcite, dolomite
unstable : olivine, anorthite (Ca-feldspar)
moderately stable : plagioclase (Na,Ca-feldspars), albite (Na-feldspar), biotite: Fe(II)-mica
stable : orthoclase (K-feldspar), muscovite (mica)
very stable : quartz, (hydr)oxides, titanite (TiO_2), clay minerals

The weatherability depends on several factors. The silicates weather *more readily* according to the following general rules (among others):

- the weaker or less soluble the base of the corresponding cation ($Fe^{3+} < Al < Mg < Ca < K < Na$)
- the higher the content of Fe(II)
- the lower the Si content of the silicates

Because of the unequal weatherability of the minerals, the weatherability of the various rocks is also different:

Igneous rocks

The *acidic igneous rocks,* high in SiO_2 (65-80%) and formed from magma containing an excess of SiO_2 over bases, weather to sand and sand-like constituents. These acid rocks contain an abundance of quartz and feldspars, minerals belonging to the most stable classes. Examples are *granite* and *gneiss,* which produce sands on weathering. Fragments of feldspars, especially orthoclase, are easily detected in the mass of the weathered material. Because of the low iron content, the weathered product is light in colour.

Intermediate or *neutral igneous rocks* with an average composition of 50-70% SiO_2, without free quartz. The diorite-andesite and syenite-trachyte rock families are typical representatives of this class of rocks. They contain the more basic feldspars, predominantly plagioclase. The weathered material is similar to that of the acidic rocks, except for being less sandy in nature.

Basic igneous rocks have 30-65% SiO_2, without free quartz. They are rich in ferro-magnesian silicates, some olivine and plagioclases. Consequently, the basic rocks, (gabbro, diabase, basalt) weather easily and give rise to clayey materials. In the arid regions, carbonates of Ca and Mg are found in considerable quantities. The colour of the weathered material is generally dark brown to red. This is due to the presence of coatings of iron hydroxide on the grains.

Sedimentary rocks

The *clastic sedimentary rocks* are more complex than igneous rocks, because they may consist of all sorts of rocks and their weathering products. Consequently, they are enriched with the stable end products of weathering, being quartz, (hydr)oxides and clays. In addition, fragments of the original rocks may be present. Some examples:

Clay rocks, as an example of sedimentary rocks, consist of clay with some grains of quartz, feldspar, mica, and secondary materials, such as silica, (hydr)oxides, calcium carbonate, gypsum, pyrite, and organic matter. Such rocks weather only slightly, because the material has already weathered before.

However, the secondary materials weather for the first time, amongst which the *clay minerals.* Although the clay minerals belong to the very stable minerals, they remain stable only under the conditions of their formation. As soon as these conditions are changed (e.g. because they are eroded and deposited in quite distant areas with a different environment) they may become unstable and decompose. Nevertheless, the changes in clay rocks by weathering are far less drastic than those in primary rocks.

Shale is a laminated clay rock. It may be grey, black, or red in colour, depending on the conditions under which the mud, from which the shale originated, was formed. If it was mixed with iron oxides, the colour is red; if mixed with organic mater, the colour is black and if it was formed from only clay, it is grey. Sandy shales contain much sand, calcareous shales some lime. The rocks disintegrate into platy material and soils, formed from such material, often have a platy structure in some part of the profile.

Sandstones consist of sand grains cemented with sesquioxides, amorphous silica, lime (calcite), or some clay. The sand grains may be pure quartz or grains of rock or of any other material. Siliceous sandstone (containing predominantly quartz) and calcareous sandstone weather to pure quartz sands. Micaceous sandstone (containing micas) weather into quartz-rich materials with some clay. Ferruginous sandstone does not change very much upon weathering, because it consists predominantly of the end-products of weathering (SiO_2 and Fe_2O_3).

The weatherability of *conglomerates* and *breccias* depends on their composition, because they consist of consolidated sand, gravel, pebbles and boulders (conglomerate), or of consolidated angular fragments, formed from talus which has been transported over a short distance (breccia).

Chemical sediments

The weathering materials of *limestones, marls* or *dolomites* depend on the impurities of the rocks. The end-products of weathering are generally clayey in nature. Very often the material is red coloured, as a consequence of iron-containing admixtures.

3.3 BIOLOGICAL WEATHERING

Biological weathering can be defined as physical and chemical weathering by biological agents.

The *physical* part is especially exerted by the plant roots, which penetrate into the crevices and cracks of rocks. By expansion, the growing roots may crack and split enormous boulders.
The *chemical* action is caused by the excretion products of plant roots or of micro-organisms such as lichens, mosses, bacteria, that grow on the surface of rocks and penetrate into minute cracks in the rocks. Some micro-organisms excrete acids as metabolic products. These acids are able to form complexes with certain constituents of the rocks, e.g. some bacteria produce 2-ketogluconic acid, which appeared to chelate Ca strongly and can therefore be a strong weathering agent in basic rocks.

In addition, the already mentioned excretion of CO_2 is a potent agent in chemical weathering, by intensifying the hydrolytic and dissolving action of water. Also other acids may be important occasionally, e.g. citric acid, oxalic acid and malic acid.

The place of weathering in the formation of soils is depicted in Figure 19.

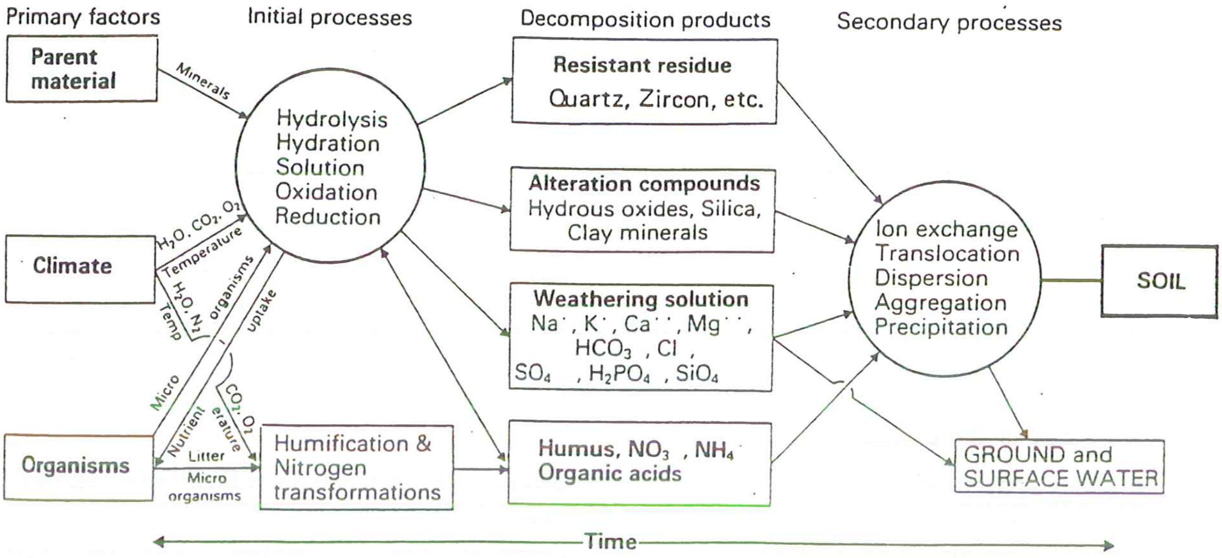


Fig. 19. Aspects of weathering in soil formation (adapted after Yaalon).

4. CLAY MINERALS

In this section, the structure of clay minerals as well as their formation and possible transformation will be discussed.

With the exception of a small group (the hormites, see Section 4.1.4), all clay minerals belong to the subclass of the phyllosilicates (i.e. sheet-silicates). The basic structural elements are the silicon tetrahedron and the Al or Mg-octahedron (Fig. 20). Sheets of both have similar dimensions and can be stacked upon each other with covalent bonds. The formation of a clay mineral should not be visualized as a completed sheet of octahedra meeting a completed sheet of tetrahedra, but rather as a growing together, unit by unit. By definition, clay minerals* are *secondary* minerals, resulting from weathering reactions, as distinct from primary minerals (such as micas) in rocks.

When in the octahedral sheet all octahedral positions are occupied the clay mineral is referred to as *tri-octahedral,* e.g with Mg (brucite-sheet), and when only two out of three positions are occupied the mineral is *di-octahedral:* with Al (gibbsite sheet). Isomorphous substitution may occur in both the tetrahedral layer (mainly Al for Si, sometimes Fe^{3+}) and the octahedral layer (e.g. Fe^{2+} for Mg, Fe^{3+} for Al, Mg for Al) usually giving rise to negative charges. Electric charge also arises from edge-hydroxyls and broken bonds which protonize or deprotonize (associate or dissociate H^+) depending on the prevailing pH of the (soil) solution.

Classification of clay minerals is based on the manner in which the tetrahedral and octahedral sheets are stacked, i.e. 1:1, 1:2 or 2:2. Table 4 gives schematic representations of structures as well as some properties of clay minerals.

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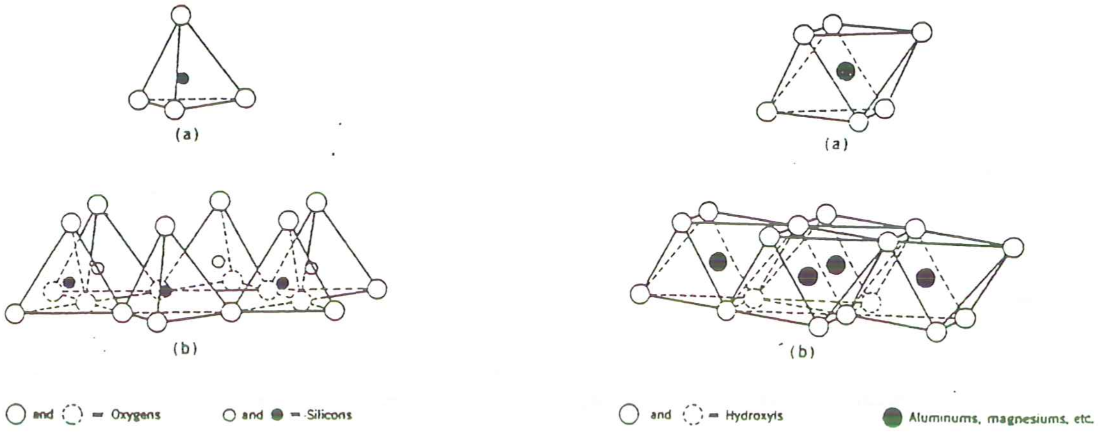
* The term *clay mineral* should not be confused with *clay fraction.* The latter refers to a particle-size fraction (usually < 2 µm) in which not only clay minerals but also other minerals such as quartz, feldspars, micas and calcite may occur.



Diagrammatic sketch showing (a) a single silica tetrahedron and (b) the sheet structure of silica tetrahedrons arranged in a hexagonal network.

Diagrammatic sketch showing (a) a single octahedral unit and (b) the sheet structure of the octahedral units.

Fig. 20. The basic building elements of clay minerals.

4.1 CLASSIFICATION AND STRUCTURE

4.1.1 Kaolinite family, Kandites, or 1:1 clay minerals

The most simple clay mineral is kaolinite. It consists of stacking of one octahedral layer of Al on one tetrahedral layer of Si. This 1:1 unit is in turn stacked upon other units (Fig. 21) with a repetition distance or *basal spacing d* of 0.7 nm (= 7 Å). Halloysite is closely related, in this mineral half of the silicon tetrahedra have their apical oxygen pointing outward, leading to a larger basal spacing whereby, as a result, water molecules can enter this *interlayer* space (*d* = 1.03 nm or 10.3 Å). This water can be removed by dehydration and the *d*-spacing becomes about 0.73 nm (i.e. somewhat larger than that of kaolinite).

Kaolinite has Al in the octahedral layer and is, therefore, dioctahedral. Related tri-octahedral 1:1 minerals, having Mg in the octahedral layer are the *septe-chlorites.*

Minerals of the kaolinite family have no isomorphous substitution, their charge is due to broken bonds and edge hydroxyls only.

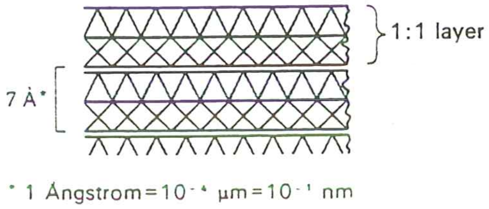


Fig. 21. Schematic structure of kaolinite.

4.1.2 2:1 Clay minerals

The three-sheet or 2:1 clay minerals consist of units of an octahedral sheet between two silica sheets. Several types can be distinguished:

1 nm (10 Å) Minerals

Due to isomorphous substitution, e.g. Al^{3+} in the octahedral layer is replaced[*] by Mg^{2+} (during growth of the crystal) or when Si^{4+} in the tetrahedral layer is replaced by Al^{3+}, there is a charge deficit which is compensated by cations in the interlayer. K^+ is a very suitable cation for this. It fits

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[*] Isomorphous substitution takes place during formation of the mineral and is, therefore, a permanent construction feature. Such "replacement" should not be confused with simple *ion exchange* which takes place on the adsorption complex of clays.

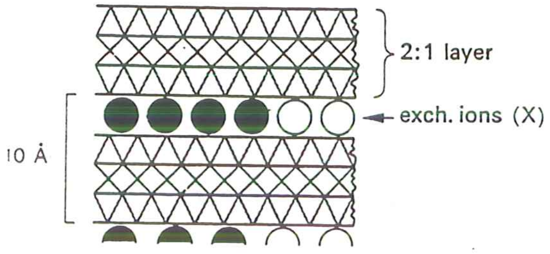


Fig. 22. Schematic structure of illite.

exactly in the 6-ring holes of the tetrahedral sheets so that there can be a close stacking of the 2:1 units with K$^+$ in between (Fig. 22). This is the case in (primary) *micas,* e.g. the di-octahedral muscovite, and tri-octahedral biotite). The basal spacing is about 1 nm (10 Å). *Illite* is a clay mineral related to muscovite with the difference that illite is not completely filled with K$^+$.

Because the potassium ions fit so well between the stacking units they are not replaceable by simple cation exchange *(potassium fixation).*

1.4 nm (14 Å) Minerals

When 2:1 units like those of mica and illite are not held closely together by K$^+$ but by other cations such as Mg^{2+}, Ca^{2+}, and Na$^+$, then they have to be further apart at a basal spacing of 1.4 nm (14 Å). *Vermiculite* and *montmorillonite* belong to this group (Fig. 23). They differ from each other in that vermiculite has a higher interlayer charge (more isomorphous substitution in the tetrahedral layer and therefore a charge closer to the interlayer) and can "collapse" to 1 nm after taking up K$^+$.

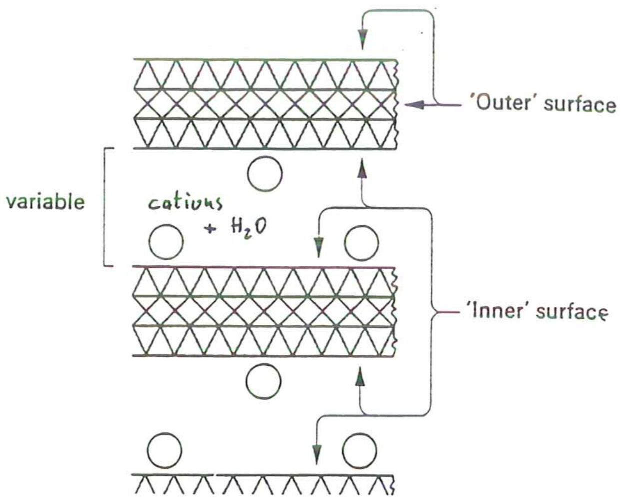


Fig. 23. Schematic structure of smectite.

Montmorillonite, with its lower interlayer charge (predominantly octahedral isomorphous substitution) cannot collapse to 1 nm with K$^+$, but can expand to 1.8 nm upon taking up water (this is helped by organic substances such as glycol or glycerol). Vermiculite cannot expand to 1.8 nm and this difference is a diagnostic feature in X-ray diffraction analysis.

Like kaolinite is a member of a family (the kandites), is montmorillonite an individual species of a large family of expanding clay minerals, collectively called the *smectites.* Vermiculites are a group by themselves.

It is only natural that intermediate types occur, i.e. types that can both collapse to 1 nm with K$^+$ and expand to 1.8 nm with glycerol. Such a clay mineral is *beidellite.*

4.1.3 2:2 Clay minerals (or 2:1:1)

These minerals are essentially 2:1 type minerals in which the interlayer is filled with an octahedral layer of Mg or Al (Fig. 24). These minerals are called *chlorites.* The primary (rock) types have a tri-octahedral brucite (Mg) layer whereas the secondary chlorites (soil chlorites or pedogenic chlorites) usually have a di-octahedral gibbsite (Al) layer. Such a mineral can easily form when adsorbed Al^{3+} cations polymerize in the interlayer.

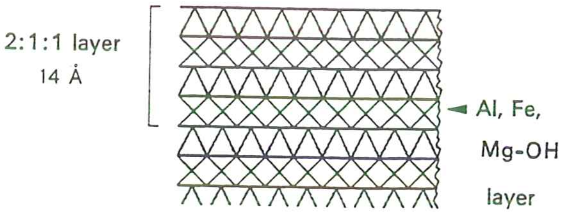


Fig. 20. Schematic structure of chlorite.

In nature, especially in the soil, things are not always ideally organized. Therefore, with clay minerals mixed types frequently occur, e.g. a stacking of vermiculite and chlorite units: or vermiculite with mica or montmorillonite with chlorite etc.. Such minerals are referred to as *intergrades* or *mixed-layered* or *interlayered* minerals.

4.1.4 Hormites

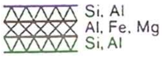
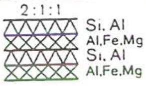
A small group of clay minerals are not phyllosilicates but inosilicates (chain silicates), viz. the *hormites* to which belong *palygorskite* (= attapulgite) and *sepiolite.* These occur mainly in the arid and semi-arid regions.

4.1.5 Allophane and Imogolite

Allophane is a collective noun for X-ray amorphous (non-crystalline) aluminosilicates. They occur notoriously in soils developed in volcanic ash and related materials as well as in podzols. In most soils some allophane is present as an intermediate product in mineral transformations. It has been found to occur as hollow spheres of about 5 nm diameter.

Imogolite is a para-crystalline clay mineral, genetically related to allophane, and consisting of long tubes of about 2 nm diameter and up to several μm long with an inner wall of silica tetrahedra and an outer wall with a gibbsite structure.

Table 4. Summary of structure and properties of clay minerals

	Kaolinite	Illite	Transitional minerals	Vermiculite	Montmorillonite	Chlorite
Type Layer structure	1:1 	2:1 				2:1:1 
Isomorphous substitution	–	mainly in tetrahedra			mainly in octahedra	in tetra- and octahedra
Layer charge per formula unit	~0	0.60–0.90			0.25–0.60	~0
Expansion	–	–	×	××	×××	–
Accessibility of inner surfaces	–	×	××	×××	×××	–
Water holding capacity	×	×	××	×××	××××	×
Adsorption capacity	×	××	×××	××××	×××	×
Contraction on K entry	–	–	×××	×××	×	–
Plasticity, cohesion	×	×	××	×××	×××	×

4.2 CLAY MINERAL FORMATION AND TRANSFORMATION

Clay minerals are formed by weathering of other minerals. This may occur in basically two ways:

1. *Diagenesis.* Direct formation by simple reactions whereby the basic crystal lattice is preserved such as the loss of K^+ by micas forming vermiculite.
2. *Neogenesis.* The clay mineral is newly formed from decay products in soil solution. For instance, feldspars, being tectosilicates, have to go through a state of solution to be transformed to clay minerals since these are almost exclusively phyllosilicates (with a completely different crystal structure).

The type(s) of clay minerals formed depend on the same factors that control soil formation: *parent material, climate, topography, vegetation and time.*

4.2.1 Processes

4.2.1.1 Diagenesis

Weathering of micas can be the start of a number of consecutive transformations whereby the basic layer lattice is preserved. Withdrawal of K^+ leads to hydromicas which form illites by re-taking up K^+. When all K^+ is removed vermiculite or smectite result (depending on the tetrahedral or octahedral isomorphous substitution, respectively). Other cations such as Mg^{2+}, Ca^{2+} or Na^+, as well as H_2O molecules can be adsorbed and the interlayer has to expand to accommodate these larger cations (see Fig. 25 A). Under certain acid conditions, these cations may be exchanged for Al^{3+}, which may polymerize and thus produce, by *alumination,* soil chlorite (Fig. 25 B,C).
Kaolinite may be formed by splitting up pedogenic chlorite (Fig. 25 C) or by stripping silicon sheets from montmorillonite (Fig. 25 D). This process is called *desilication.* Further desilication of kaolinite may lead to the formation of gibbsite since only the octahedral Al layer remains.

4.2.1.2 Neogenesis (or authigenesis)

In principle, all clay minerals can be formed by crystallization from a solution containing the necessary components. Thus, kaolinite can be formed from a soil solution containing predominantly Si and Al; illite from Si, Al and K; montmorillonite from Si, Al, Ca and Mg, providing the pH values are favourable (see next section). Crystallization can be either directly from solution or via an amorphous precipitate (allophane). It is noteworthy that trioctahedral minerals are generally much less stable than dioctahedral types (e.g. biotite weathers much faster than muscovite). Therefore, newly formed clay minerals will almost exclusively be of the dioctadral type. The most important transformations are summarized in Table 5.

4.2.1.3 Environmental conditions

As stated before, the environmental conditions determining the species of clay mineral that is formed, are controlled by the soil forming factors. Some of the most prominent differentiating factors will be discussed.

Generally, wherever in the universe, the type of mineral to be formed depends on the thermodynamic conditions, i.e. temperature, pressure, kind of components present and their abundance (concentration or activity). Since in soils temperature and pressure fluctuate between rather narrow limits the most significant variable is the *concentration of the components* in the soil solution.

The composition and concentration of the soil solution, including the pH, depends on such factors as mineral composition, particle size, effective rainfall, vegetation (humus type and content) vertical and lateral water movement (including stagnation). Temperature controls the rate of reaction and sometimes the type of mineral formed. Indirect influence on the type is exerted by temperature through the vegetation.

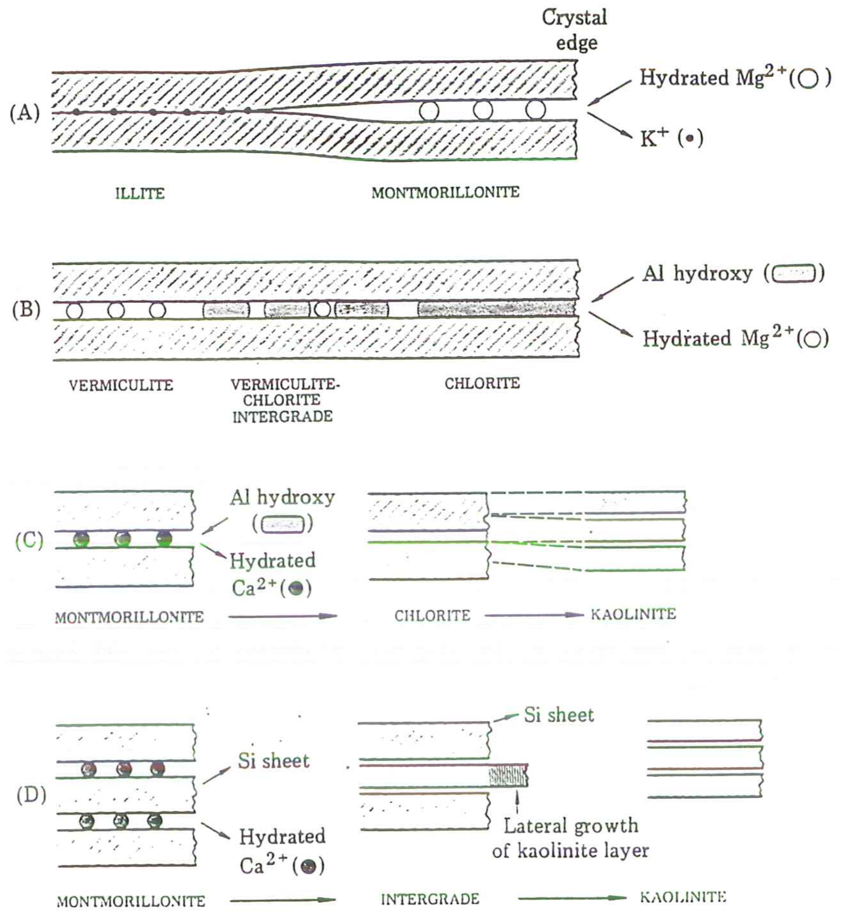


Fig. 25. Several clay mineral alteration schemes.

(A) Alteration of illite to vermiculite or smectite by K^+ removal.

(B) Alteration of vermiculite to chlorite by exchange of interlayer ions by Al^{3+} and subsequent polymerization of these.

(C) Alteration of smectite to chlorite to kaolinite. The silica tetrahedra in the middle kaolinite layer depicted must invert so that the interlinked bases of the tetrahedra all face upward.

(D) Alteration of smectite to kaolinite by stripping silicon sheets from smectite. Kaolinite growth can take place laterally from a newly formed kaolinite layer, and additional layers in turn can become oriented on these layers. Hydroxyls occupy the oxygen sites in the octahedral layers left vacant by silicon sheet removal.

Table 5. Some possible clay transformations occurring in the soil

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1) Mica $\xrightarrow{-K}$ illite $\underset{+K}{\overset{-K}{\rightleftarrows}}$ vermiculite, montmorillonite $\underset{-Al}{\overset{+Al}{\rightleftarrows}}$ sec. chlorite

2) Feldspar, Amphibole, Pyroxene, Olivine, etc. $\xrightarrow{-(Ca,\ Mg,\ K,\ Na)}$ decay products [SiO_2, Al_2O_3, Fe_2O_3, etc.]

$\longrightarrow$ allophane, kaolinite, halloysite, gibbsite
+K $\longrightarrow$ illite
+(Mg,Ca) $\longrightarrow$ e.g. montmorillonite
Al, Fe oxides and oxyhydroxides

3) Clay minerals (1:2 and 2:2 types) $\underset{+(Ca,\ Mg,\ K,\ Na,\ SiO_2)}{\overset{-(Ca,\ Mg,\ K,\ Na,\ SiO_2)}{\rightleftarrows}}$ Kaolinite, halloysite $\underset{+\ SiO_2}{\overset{-\ SiO_2}{\rightleftarrows}}$ gibbsite (and Al-oxide and oxyhydroxide)

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1) and 3): diagenesis 2): neogenesis

The generalized conditions for formation and persistence of the common clay minerals and (hydr)oxides can be summed up as follows:

Smectites (Montmorillonite and Relatives)
A relatively high ionic concentration of silicon and of magnesium is required for synthesis of montmorillonites. Generally these conditions are met in the vicinity of decomposing silicate minerals rich in magnesium (and iron). High Si concentration is maintained by slow movement or stagnation of soil water. Montmorillonite is unstable under acid conditions and rapid leaching. Therefore, it may survives in dense clay layers in which the rate of leaching is slow.

Vermiculite
This mineral is formed under mildly acid conditions such that potassium (and magnesium) are completely removed from interlayers. Mica must be present in the initial material. Concentration of Si must be high. But concentration of Al in solution must be low, otherwise this will be precipitated in interlayers to form 2:1-2:2 intergrades (vermiculite-chlorite).

Illite
This mineral forms where mica is present in the initial material, under conditions of moderate to low acid concentration necessary for partial stripping of K from interlayers. Moderate to relatively high concentrations of Si and Al are required for stability. Moderate to low pH causes instability and disappearance, as illite is converted to vermiculite.

Pedogenic or soil chlorite
This mineral is formed under conditions of moderate to low pH and medium to high Al and Si concentration. The interlayer spaces serve as a sink for the Al in solution: an *"anti-gibbsite"* effect. Mica, vermiculite or montmorillonite is required in the initial material as a precursor.

Kaolinite
This mineral is synthesized under conditions of approximately equal concentrations of Si and Al, with low pH and essentially an absence of Ca, Mg and other bases (i.e. chemically "poor" conditions). Formation is aided by the presence of layer silicates as "templates" or patterns for the 1:1 sheet structure.

Halloysite
This mineral forms where concentration of Al approximately equals that of Si, as a result of rapid weathering of feldspars. It can also be formed by organization of an amorphous aluminosilicate such as allophane into more crystalline halloysite over time. It requires low pH and zero to low concentration of bases.

Allophane
This "amorphous" material forms under conditions of medium to low pH (> 4.9) in highly moist or wet systems, by rapid weathering of non-crystalline volcanic ash, or to a limited extent by rapid weathering of feldspars.

Gibbsite
This mineral forms and persists under conditions of low Si concentration and low pH, with absence or low concentration of bases. It may form by aging of aluminous allophanes or gels.

Goethite - Hematite
Both these minerals can be formed when ferric ions are available. Their formation and the conditions favouring either of these species are rather complicated and still under study. Goethite seems to form directly by precipitation and crystallization whereas hematite seems to form from ferrihydrite by a dehydration-rearrangement process. The goethite-hematite pair is the most widespread association of two pedogenic Fe-oxides in soils of the warmer climates. The ratio of goethite to hematite is governed by a number of factors among which soil temperature, soil water activity, organic matter, pH and Al activity are the most prominent. The two pathways of formation are considered competitive. Hematite is generally favoured by a higher temperature and lower water activity. Increases in the proportion of hematite to goethite have been found to be related to redder colours in soils.

4.3 SOME PROPERTIES OF CLAY MINERALS

Three of the most important features of clay minerals are:

a) the cation exchange capacity (CEC), i.e. their capacity to *adsorb cations* (to a generally much lesser extent clays may also adsorb *anions:* AEC)
b) the capacity to adsorb water *(swelling properties)*
c) the ionic or electric double layer *(flocculation and peptization)*

Because of their variation in structure and particle size, the CEC of clay minerals differ widely. In addition, as stated before, the CEC is for most clay minerals to a varying degree dependent on the pH of the soil solution. Electrostatic charge on clay minerals originate from:

a) Isomorphous substitution in the crystal lattice (permanent charge)
b) OH-groups on the crystal planes (pH-dependent charge)
c) OH-groups on the frayed "ends" of crystals (broken bonds; pH-dependent charge)

These OH-groups may either be attached to Si- or Al- in the lattice, and protonation or deprotonation in dependence of the pH can be represented by the following equations:

$$X\text{-}O^- + H^+ \leftrightarrow X\text{-}OH^0 \qquad (21)$$

$$X\text{-}OH^0 + H^+ \leftrightarrow X\text{-}OH_2^+ \qquad (22)$$

in which X represents the clay lattice. It can be seen that the charge is negative (at higher pH values), zero or positive (at lower pH values). Usually, the net charge will be negative although especially for kaolinite, which has no isomorphous substitution, the CEC is very low at low pH values. When relatively high contents of (hydr)oxides are present (Ferralsols, Oxisols), the net charge may sometimes even be positive. It is clear that the CEC is closely related to the specific surface area of the materials (see Fig. 27). The CEC of some more materials in soils is given in Table 6.

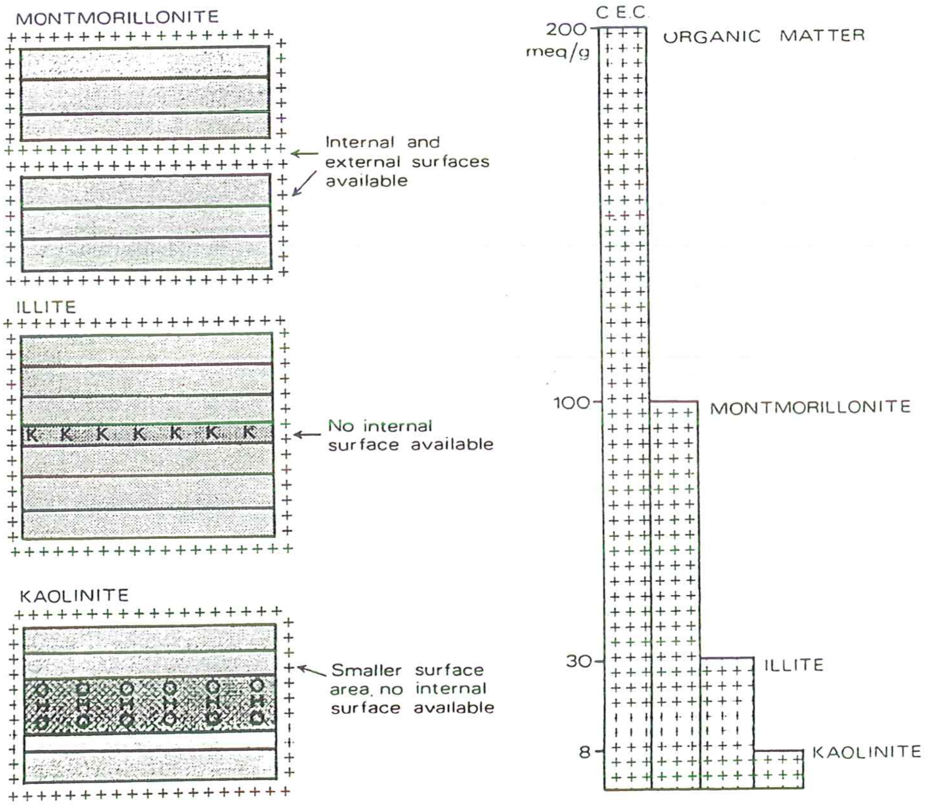


Fig. 27. Cation exchange capacity.

Table 6. Approximate cation exchange capacities (CEC) of some soil components.

Material	CEC (cmol_c/kg or me/100 g)
Organic matter	150 - 500
Kaolinite	3 - 15
Halloysite	5 - 10
Hydrated halloysite	40 - 50
Illite	10 - 40
Chlorite	10 - 40
Montmorillonite	80 - 120
Vermiculite	100 - 150 (or higher)
Allophane	25 - 125
Hydrous oxides	4 (or lower)
Feldspars	1 - 2
Quartz	1 - 2
Zeolites ("open" tectosilicates)	200 - 600

The absorption (and desorption) of water is expressed by the swelling and shrinking of clays. Although all clays have this property to some extent (most heavy clays will crack upon drying) the feature is most prominent in the smectites. *Gilgai* results from the repeated wetting and drying of montmorillonitic soils. These properties are further discussed in Sections 6 and 8.

5. SOIL FORMATION

Soils are formed in parent material that was derived from parent rock by weathering processes (and possible subsequent transport) although these processes usually overlap.
One could say that weathered rock is not soil but has to be turned into soil by the operation of certain factors and processes. It is generally recognized that these soil forming factors are:

- **Climate** (temperature and precipitation)
- **Biosphere** (especially the native vegetation)
- **Parent material** (texture, structure and composition)
- **Topography** (sloping, relief
- **Time** (duration of soil formation)

The first two factors are called the *active* soil forming factors, whereas the others are the *passive* factors.
Soil forming factors do not act in isolation but always together which sets limits to the operation as a whole. In certain situations it may be that one factor is dominant and then its effect can be clearly distinguished.
Soil formation is a complex of processes operating within boundaries determined by an interacting set of factors, the processes and boundaries being subject to variation during time. More precisely, it may be stated that ’*a soil is formed by action of climate and biosphere on parent material over a certain period of time and modified by topography*’.

5.1 CLIMATE

5.1.1 Zonality concept

By and large, climate is perhaps the most influential factor. This is borne out by the fact that there are many parallels in the distribution of climate and soil on the earth's surface. The concept of *zonality* of soils was based on this observation. The concept says that if the active factors (climate and biosphere) are similar the soils formed will be similar also when the parent materials are different. Alternatively, similar parent materials would give rise to different soils if the active soil forming factors are different.

The broad schematic soil map of the world (Fig. 24) shows how soils coincide with climatic zones. Soils that comply with the climatic zone rule are called *zonal* soils. Soils that do not comply (because other factors dominate) are called *intrazonal,* whereas soils that are too young to show horizon differentiation are called *azonal* soils.
Although the concept proved to be a great stimulus in soil science, it appeared to be untenable since too many deviations and exceptions were encountered.

The effect of climate on soil formation is *direct* and *indirect.* The direct effect is that of *precipitation* and *temperature,* the indirect effect is exerted through its influence on the biosphere (flora and fauna). It can be stated that climate is the determining factor for the vegetation type. Parent material can modify this but is far less important than climate.

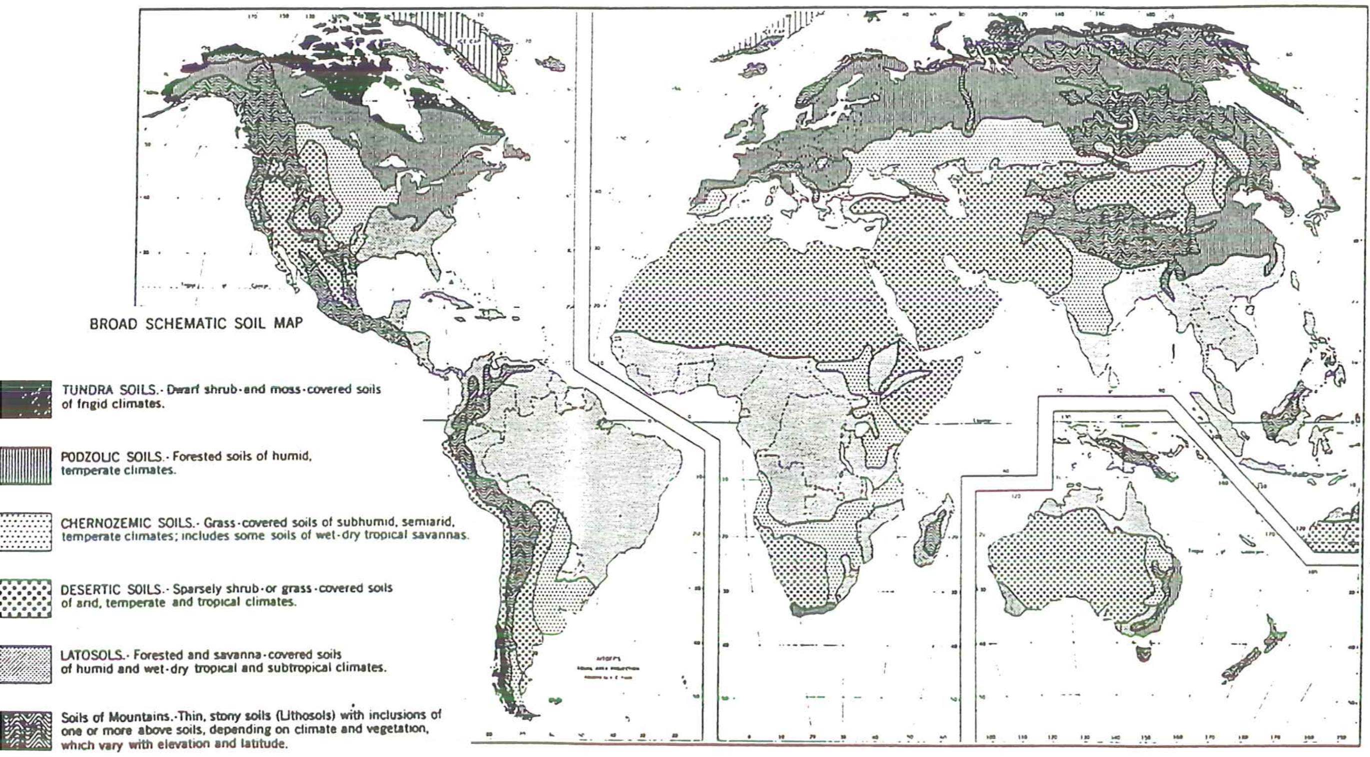


Fig. 28. World distribution of Great Soil Groups showing zonality.

5.1.2 Precipitation

If rain is falling on the earth's surface, part of it penetrates into the earth, part runs along the surface if topography is favourable for that, and part is evaporated. It is generally assumed that roughly 0-50% of the total rainfall percolates and reaches the groundwater. The part that penetrates and percolates, the *effective rainfall,* is very important with respect to soil formation. This water reacts with the mineral particles of the parent material, dissolves some constituents, removes this dissolved material from the surface and deposits it at a lower part of the solum. In this way the mass is differentiated into horizons and the soil body is formed. This process is the more effective, if a vegetation has established. In the beginning there is no vegetation and then erosion (surface run-off) is important.

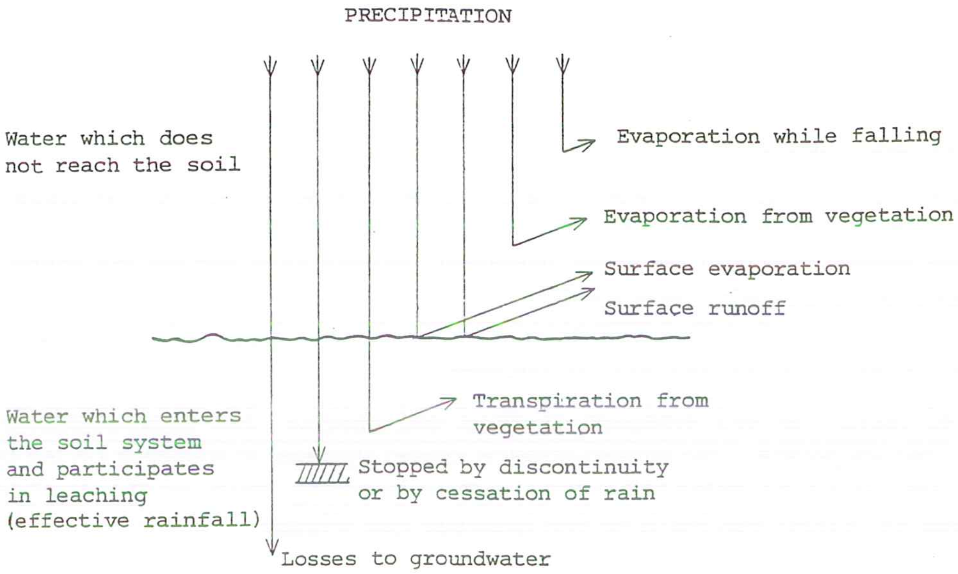


Fig. 29. Soil water balance.

But as soon as a vegetation is present, the roots protect the material from being eroded, the leaves protect the soil against the impact of raindrops, and decomposition products of the organic material become increasingly important as they contribute effectively in the reactions between the percolating waters and the materials with which they come in contact.

The water balance, yielding the effective rainfall, is depicted in Figure 29. It shows that not all effective rainfall will reach the groundwater but that part may be stopped at some depth by lack of further rain or by a discontinuity in the soil (pan; clayey layer etc.) and that another part is "caught" by the vegetation.
It is obvious that under different climates the contribution of the various components of the water balance will be different. For instance, at high humidity evaporation losses will be small. In a tropical rain forest the surface evaporation will be small because the relative humidity in such a forest is 100%. Transpiration, however, can be very high because the tops of the trees are in quite different conditions. Wind, by affecting evaporation and transpiration, can play an important role too. The fate of moisture under different climates is schematically represented in Figure 30.

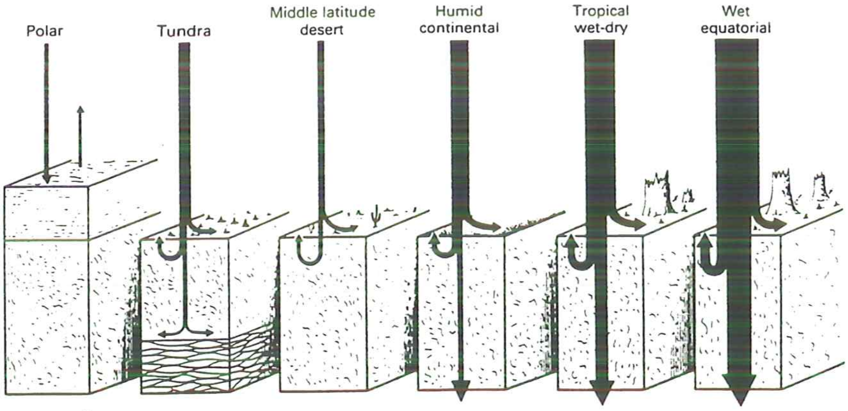


Fig. 30. The fate of moisture falling on the surface under a number of different climatic conditions. (After FitzPatrick).

Desert
In the rainless desert there is no profile differentiation, because there is no water for percolation. In many places in the desert soil profiles are found, however, but these are then evidence of earlier more humid climates. If precipitation increases, more percolation takes place and the formation of soil profiles is the result.

Equatorial belt
Precipitation culminates in the equatorial belt, where rainfall reaches the startling figures of 3000 to 7000 mm a year. Although a considerable part runs off and evapotranspires, such a large amount of water percolates that only few dissolved substances are retained in the soil profile, if drainage is not impeded. Consequently, a peculiar type of soil formation must occur here (ferralization, see Section 6.2).

Mediterranean climate
Not only the total annual amount of effective rainfall is important in soil formation, but also its *seasonal distribution.* The mediterranean climate, for instance, is characterized by rainfall in the winter, whereas the summer is rainless or nearly rainless; percolation effects and hence translocation of soil constituents are limited to the winter season. This must lead to the formation of soils, that differ from soils formed under a more uniform rainfall. Also the monsoon areas of the humid tropics have soils that differ strongly from those in the constantly humid areas (Vertisols - Oxisols).

Semi-arid climates
The form of the precipitation is also important. This effect can be seen in the semi-arid climates. Although here evapotranspiration is larger than precipitation, leaching of certain constituents can still take place. If rainfall occurs, it is in a form of a thunderstorm. Certain dissolved constituents can then be leached deep into the soil. Upon drying these leached substances cannot be transported upwards with the ascending moisture to re-establish the original conditions, because a considerable part of the ascending moisture moves in the vapour phase. Thus, soils with a calcic horizon (horizon of secondary lime accumulation) are very common in those areas.

Subhumid climates
In the subhumid regions, such as the Great Plains of North America, and the *chernozem* areas of Europe and Asia, the seasonal distribution of rainfall determines the formed soils. The long winter keeps the ground frozen for a considerable period in the year and in spring these soils are completely saturated; the molten snow cannot percolate, because the subsoil is still frozen.
The water does not percolate until the last trace of ice has disappeared. The summer, which is rather dry, shows an upward moisture movement. In other words, percolation is here restricted to springtime and is then sudden and strong.

Cold climates
In regions of the arctic, subarctic, and cold (tundra) climates, very little water is available for percolation. As the snow melts, most of the water disappears as surface runoff. Moreover, the subsoils are permanently frozen in the arctic areas, so that percolation is minimal.

5.1.3 Temperature

All the just mentioned sources of water losses are influenced by temperature. The higher the temperature, the greater the evaporation and transpiration and the lower the effective precipitation that percolates through the soil. The result is that, with the same precipitation in two different isothermal belts, different types of soil profiles will develop.

Cold climates
In the subartic, in the low temperature regions, and in the high mountain country long and severe winters occur. Hence, percolation is strongly hindered with the result that many lakes, swamps and peats are formed. The conditions for peat formation are favourable in spite of the scare vegetation, because the microbial activity, responsible for the rapid destruction of the organic material, is strongly restricted during the long winters. Prolonged freezing and, consequently, drying favours the

stabilization of humus, rendering it more resistant to decomposition by micro-organisms. The O-horizon is therefore very thick, 30 cm or more. Below the O-horizon there is generally no soil formation because the decomposition products formed in the O-horizon cannot percolate into the mineral mass and, consequently, cannot react with it.

In a somewhat less cold climate when percolation is not hindered by where microbial activity is still rather low, complexing organic compounds can be abundant and podzolization will take place.

Warm climates
In the humid tropics and subtropics a very dense vegetation occurs, but at the same time an intense activity of micro-organisms. The plant residues are therefore rapidly mineralized. The result is that very little organic matter accumulates in the O-layer, and ferralization will be the dominant soil forming process. Only in areas of poor drainage and at high elevations where temperatures are low, appreciable organic matter may accumulate. Therefore, also in the tropics peat and podzols can occur.

Temperature can also be affected by slope position. Figure 31 shows how differences in exposition to sun rays give rise to different climatic conditions and, hence, to different soils.

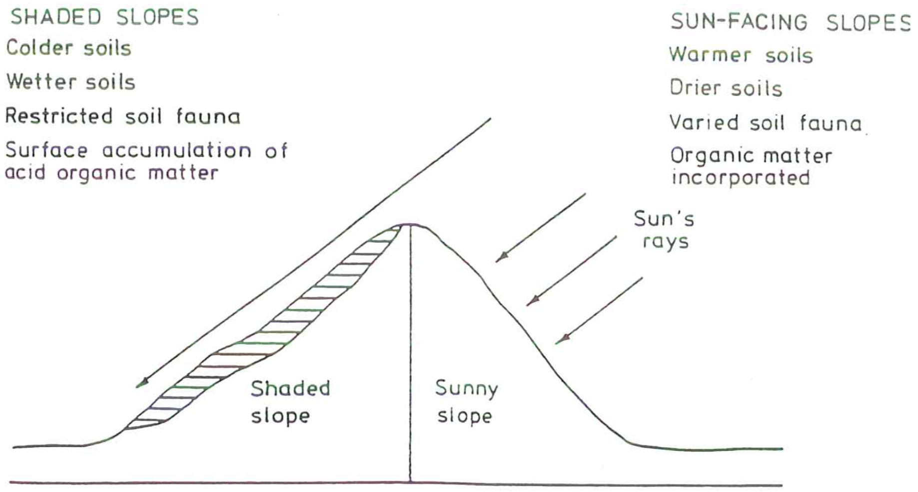


Fig. 31. Effect of slope aspect.

5.2 BIOSPHERE

The biosphere is composed of two elements, the *plant* kingdom, *flora* or *phytosphere,* and the *animal* kingdom, *fauna* or *zoosphere.* The two elements have opposing effects on soil formation. The phytosphere promotes soil horizon formation, the zoosphere retards the differentiation of the soil into horizons. This will be elucidated in the following.

5.2.1 Phytosphere

The role of organisms in soil formation is related to that of climate. Climate has strong control over both density and type of fauna which in one form or the other influence soil formation. It also controls the rate and type of organochemical reactions as was discussed earlier.
As in the case of climatic elements, the role of plants can be either *direct* or *indirect.*

Direct role
The direct role is mainly the supply of organic matter from leaf fall, dead roots, tree trunks, twigs, etc. The process by which organic matter is worked into the soil is called *humification,* a process

which involves also the soil fauna. Bacteria and fungi decompose the plant tissue whereas earthworms, termites and other animals then work the products into the soil. These decomposed plant materials supply carbon, nitrogen, sulphur and acids which are very important in soil formation. A very direct action is the excretion of CO_2 by living roots which to no small extent may contribute to acidity. The acids react with the minerals or their weathering products and the reaction products can be transported with the percolating waters and can either be accumulated at some depth in the profile or be removed with the draining water. In this way a differentiation of the soil into horizons is initiated.

Indirect role

The indirect role is pluriform. The roots penetrate rocks and soils and so facilitate entrance of solutions and colloids. Vegetation also affects climate. Generally, forests tend to make the climate milder, reduce wind speed and so decrease the evaporation especially in warm regions and seasons. They reduce the effective rainfall by interception and last but not least: they protect the soil against erosion.

A change in type of vegetation may be accompanied by a change in soil type. For instance, oak wood replacing a grass cover has led to marked leaching of the original chernozems. The changes that occur as a result of forest clearing are even more marked. Apart from promoting erosion, the supply of organic matter is interrupted, the plant nutrients are depleted by ensuing crops and possibly added fertilizers change the base status of the soil and, consequently, the microflora and fauna. This *human activity* is sometimes considered the *sixth soil forming factor*.

Humid and subhumid climates

In humid and subhumid temperate climates, because of the abundance of organic matter, soil formation is dominated by processes such as chelation, hydrolysis, oxidation and reduction. The results are a redistribution, including leaching, of iron, aluminium, silicon, calcium, magnesium and clay in the profile.

Arid and semi-arid climates

In arid and semi-arid climates, where the effect of organic matter is negligible, solubilization and precipitation of soluble salts are most prominent.

Humid hot climates

In humid hot climates with their rapid and almost complete decomposition of organic matter in the upper centimetres, hydrolysis and oxidation are the main processes leading to very typical soils (Oxisols).

A schematic *climosequence* across the U.S.A. is given in Figure 32, while Figure 33 shows the latitudinal zonation of soils from the north of Russia to the Caspian Sea in the south and from the centre of the Sahara Desert to Zaire.

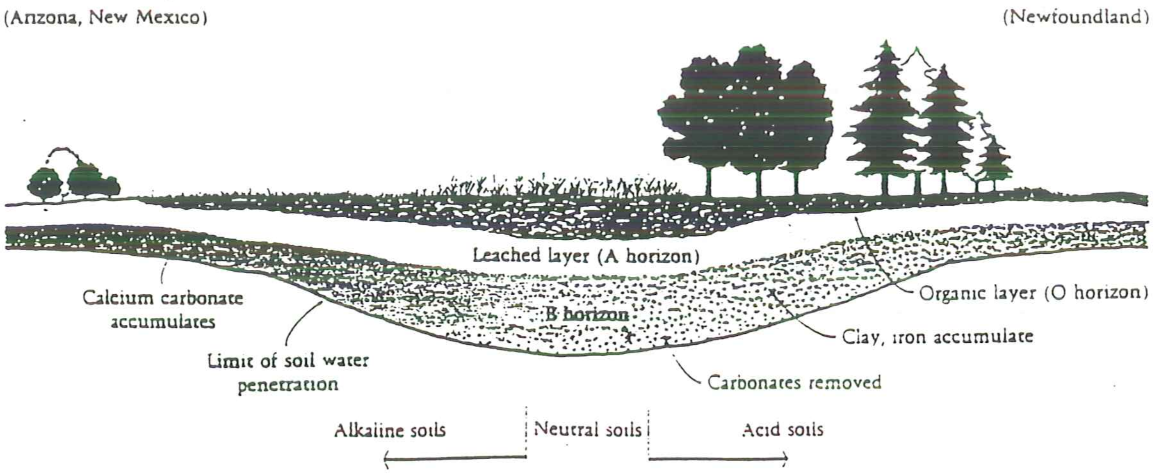


Fig. 32. Climosequence of soils across North America.

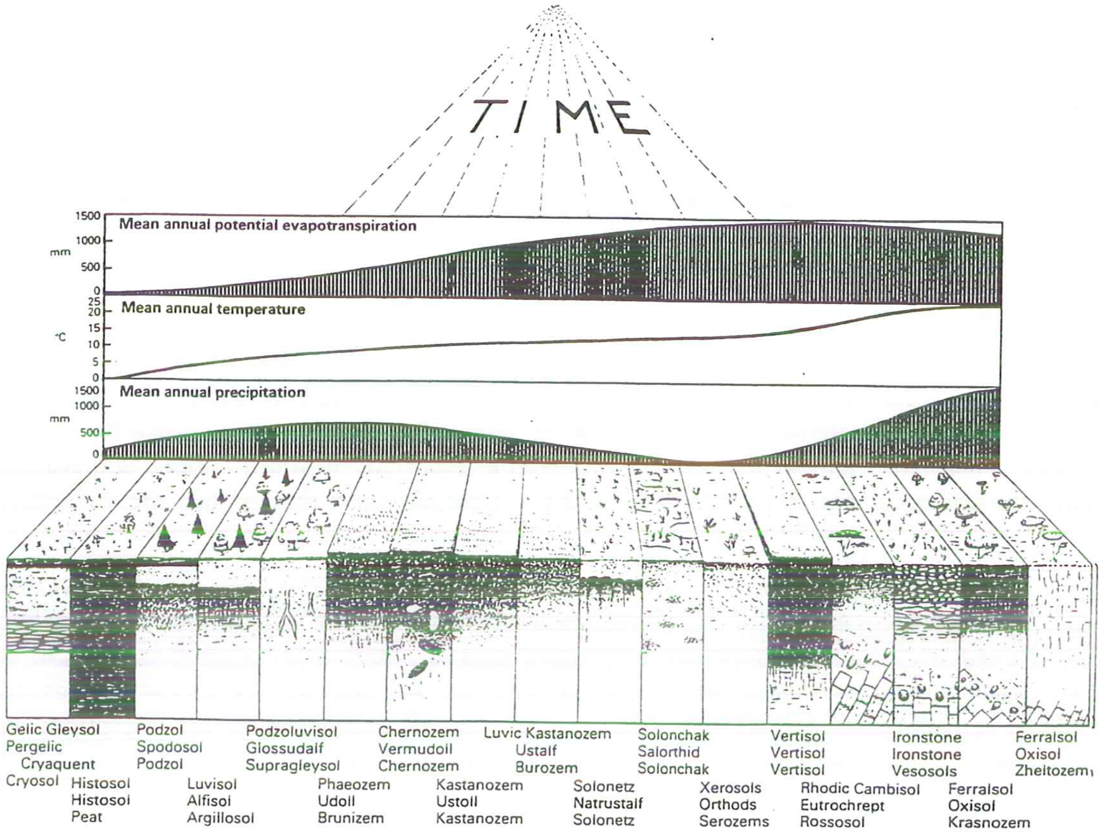


Fig. 33. Latitudinal zonation of soils. North: left - South: right. (After E.A. FitzPatrick).

5.2.2 Zoosphere

The contribution of the animal kingdom to the processes of soil formation is primarily mechanical in nature. Whereas the phytosphere causes a *differentiation* of the soil, the zoosphere has the tendency to hamper and even to nullify this. They *homogenize* the soil to a greater or lesser extent. At least they disturb the profile. *Rodents* (mice, rabbits, etc.) carry down much material from the A-horizon down into their nests (mostly located in the B-horizon) and bring up B-material to the A-horizon. This animal effect may be very important from the viewpoint of agriculture, because plant roots frequently follow the abandoned tunnel borings of rodents, worms, and insects.

Ants and termites carry material from lower to higher levels, build galleries and underground passages, throw out a great deal of material to the surface and frequently form hills and mounds. Ants are able to transfer soil material from the depth to the surface sufficient to form a layer of at least half a centimetre during a year.

5.3 PARENT MATERIAL

The nature of parent material has a decisive effect on the properties of young soils and may still exert an influence on even the oldest soils. Where parent material is derived from consolidated rock, the formation of parent material and the soil may occur simultaneously. The most important properties of parent material with respect to soil formation are:

1) texture (particle-size distribution)
2) mineralogical composition
3) degree of stratification

5.3.1 Texture

Parent materials with a fine texture, and consequently lower permeability, give generally rise to shallow soils. Coarse textured, permeable soils are favourable for the formation of deep soils.

Figure 34 gives an example of a *textural lithosequence.* The most prominent aspect is the deeper position of the B horizon in the coarser material due to "faster" movement of material.

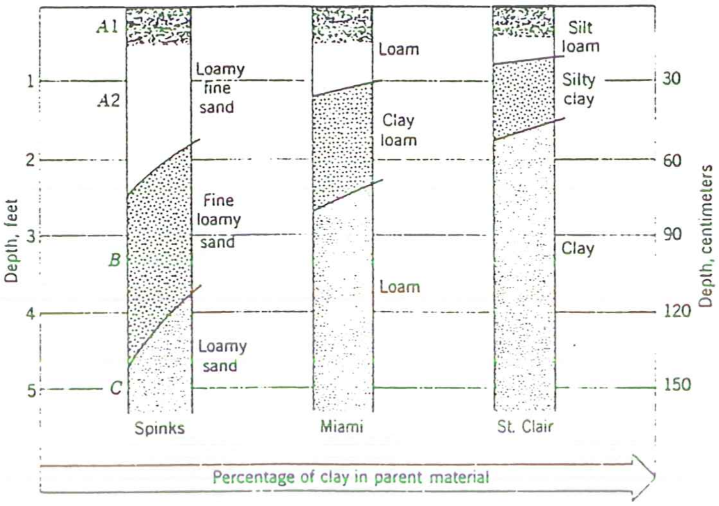


Fig. 34. Relationship between the texture of the parent material, and the thickness and texture of the horizons of three forest soils of the north central United States (Alfisols).

5.3.2 Mineralogical composition

If other factors are equal, soils formed on basic consolidated rocks will be deeper than those on acidic rocks because mafic minerals are less resistant to weathering than felsic minerals. During soil formation, the dominance of parent material properties may change. For instance, a basic rock may initially form a deeper soils than a neighbouring acidic rock but weathering of basic rocks yields more clay than acid rocks do so that after some time the texture will become an important factor in soil formation.

In case of soil formation in unconsolidated parent rock (many sediments are unconsolidated) the depth of leaching may be the same at different sites but the composition of the leaching water will then be the dominant factor. This is illustrated in Figure 4.9 which shows a *compositional lithosequence.* The high-lime till gives rise to a Mollisol without much evidence of degradation. at somewhat lower lime content, the lime may be leached out in the topsoil and clay may become mobile and transported downward *(lessivage)* forming an Alfisol.

In the non-calcareous material intrazonal Podzols are formed because of the low activity of the micro-life and the resulting slow decomposition of organic matter.

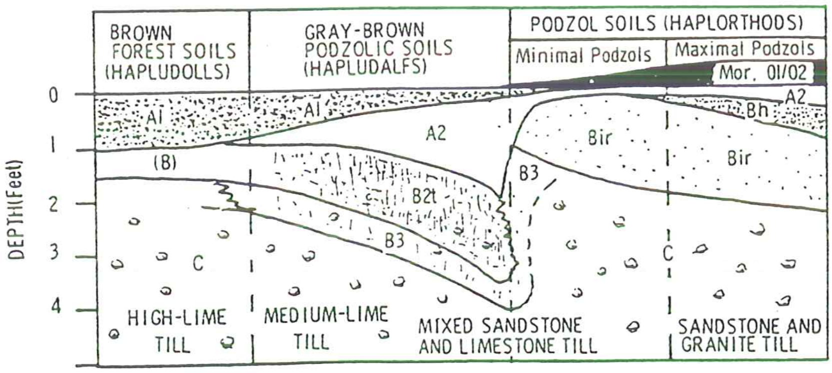


Fig. 35. Brown Forest-Podzol sequence in New York State (after Cline).

5.3.3 Stratification

Layers with deviating texture causing *stratification* often occur in parent materials. Such a *textural discontinuity* form a hindrance to the downward movement of leaching water and thus affect profile development. Layers of heavier texture (clay layers) allow only slow water movement and, moreover, water will only enter the underlying coarser layer at near saturation with water since the small pores (capillaries) will retain the water until that stage because they will only let the water go when the curve of the water meniscus equals or is less (flatter) than the curve of the meniscus in the coarser pores (see Fig. 36). In this way, any coarser textured layer will hinder the flow of water. As a result, materials carried by the leaching water may be deposited at a certain depth in the profile such as clay, humus, sesquioxides, solutes (calcite, gypsum). Because these materials tend to stop up pores, this process tends to be self-generating and self-intensifying. The halting of a water front in a homogeneous parent material may deposit enough material to initiate a discontinuity.

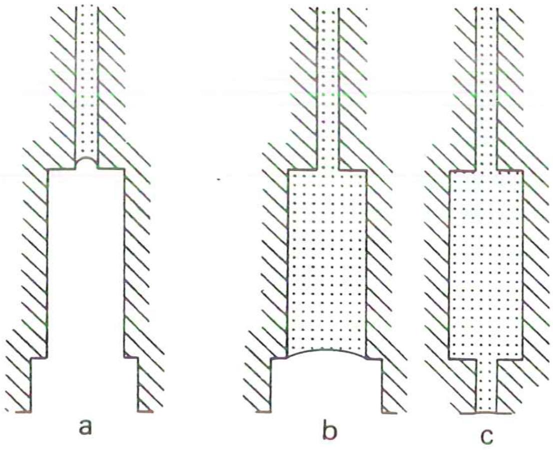


Fig. 36. The mechanism of water stagnation by a textural discontinuity.
(a) Water "hangs" in narrow pore of fine-textured layer. The water will not pass into the wider pore of the coarser-textured layer underneath as the moisture tension is too high (strongly curved water meniscus).
(b) Only when moisture tension is sufficiently lowered by water supply from above, water will flow into wider pore where water is held at lower moisture tension (less curved meniscus). The still wider pore further down will not be filled for reasons mentioned under *a*.
(c) Same moisture tension as *b* but the underlying layer is fine-textured with narrow pores. These will be immediately filled after the wider pores because their capillary force exceeds the low moisture tension prevailing in the wider pores. Thus, as soon as the wider pore section is filled, the vertical water flow is restored.

5.4 TOPOGRAPHY

Topography modifies soil profile development in three ways:

1. by influencing climate (exposure, temperature, rainfall)
2. by influencing hydrological situation
3. by influencing erosion (removal of soil from hill tops, deposition at hill foot)

5.4.1 Influence on climate

The climatic aspect of slope position was already shown in Fig. 31. Because of the difference in exposure to sun rays, there is a "cold" and a "warm" side of the hill or mountain. This often coincides with a "wet" and a "dry" side respectively. Thus, soil formation will be different on the opposite sides even if the parent material is similar.
Figure 37 shows a loess hill in the state of Washington, on the west coast of the U.S.A. The south-

facing slope features a non-calcareous Mollisol, considered normal in the area. The hill top has a thin Mollisol with also a thinner structure-B horizon but in addition a calcic horizon. These are due to a more arid micro-climate. The north-facing slope is characterized by a much thicker A horizon and even an eluvial (leached) horizon due to the higher effective rainfall and lower temperature.

In the *equatorial tropics* this effect of exposure is negligible because of the high elevation of the sun. However, the *monsoon* climates in this area result in different climatic conditions of the western and eastern slopes. The western slopes receive more rains during the wet west monsoon than the east side. In the east monsoon, being dry, both slopes get only little rain. as a consequence, the soils of the western slopes are deeper than those of the eastern exposures.

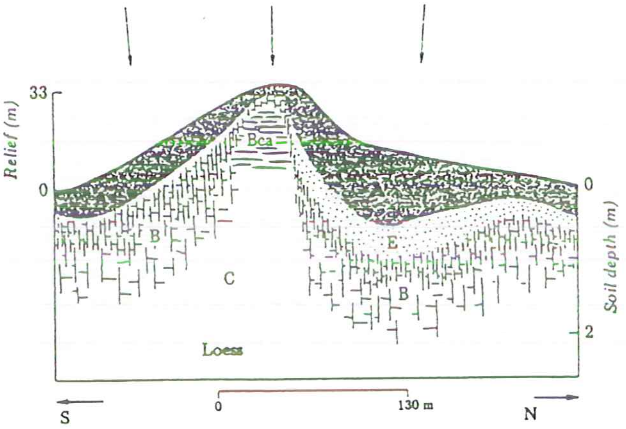


Fig. 37. Variation in soil properties with slope orientation in loess hills, eastern Washington.

5.4.2 Hydrological aspects

Relative differences in elevation may have great hydrological consequences. Due to lateral water movement and relative position of the ground water table, in higher areas soils may be dry whereas at lower places soils may be wet. Slight depressions and elevations (microtopography or microrelief) can give rise to locally deviating features such as *gley* or even more drastically deviating (intrazonal) soils. Examples are *saline* or *alkaline* soils in depressions in the area of the Chernozems, and Podzols in the area of Cambisols/Inceptisols.

The position with respect to groundwater is illustrated by the formation of groundwater laterite in small depressions in an area occupied by latosols (plinthic Ferralsol or Gleysol *vs.* orthic Ferralsol; Aquox *vs.* Orthox). Also the occurrence of low humic podzols in depressions of an area of Podzols (Umbraquepts *vs.* Typorthods).

Percolating water becomes more concentrated and higher in pH in lower-slope soils. The oxides may differ in degree of hydration and are usually yellowish in lower-slope soils and brown to red in upper-slope soils.

The hydrological position also affects the vegetation and, consequently, soil formation. The transition sequence of soil types along a slope is called *toposequence* or *catena.*

5.4.3 Erosion

With regard to position along the slope, it is important to note that soil material is washed down from the upper to the lower slope. Thus, generally, the soil depth would increase downslope. However, the picture is usually more complex. Figure 38 shows that the soil depth is related to the slope angle rather than to the soil position. The steeper the slope, the shallower the soil.

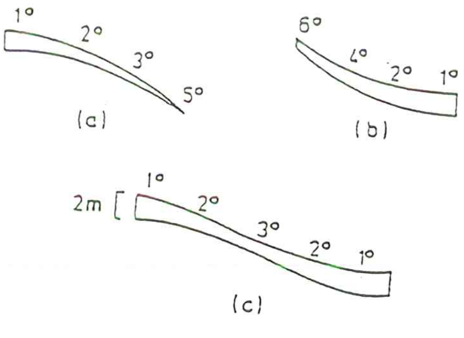


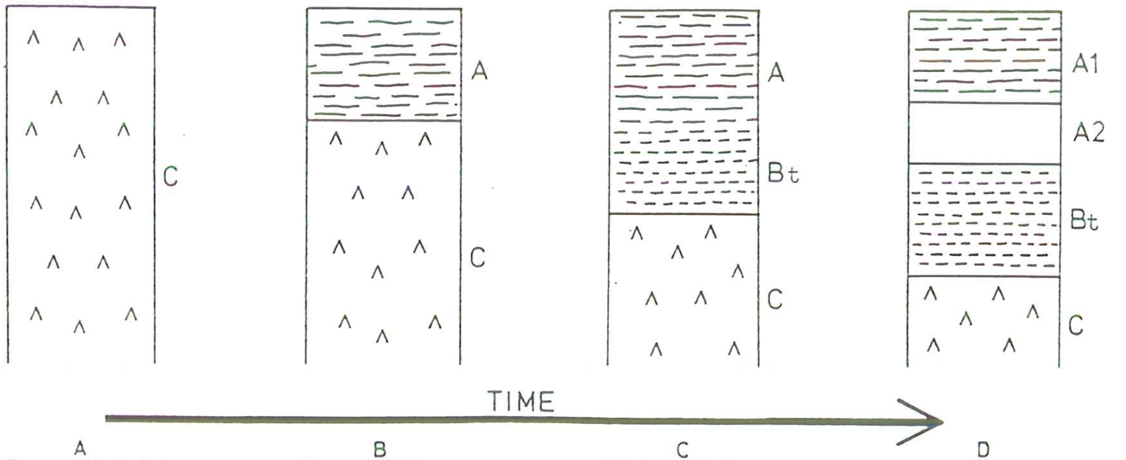
Fig. 38. Changes in solum thickness with slope angle.

5.5 TIME

In soil formation, time is a very relative factor. Development of a soil from hard rock may take thousands of years, whereas development in permeable, unconsolidated material can be fairly rapid: from some tens to some hundreds of years. A given period of time may produce much change in one soil and little in another. Also, soil formation may keep pace with the change of climate, a very gradual process. On the other hand, changes in environmental condition may be marked or catastrophic, e.g. by floods or mass-flows leading to truncation or burial of an original soil profile or to both. Therefore, counting years is of little use and time can better be considered in terms of stage of development or of maturity.

The parent material may be transformed into an *immature* or young soil in a relatively short period. There may be some organic matter accumulation in the surface layer accompanied by some weathering and leaching. Only A and C horizons are present. The *mature* stage may be attained at a later stage: the profile is then in a "steady state" or in "dynamic equilibrium" since there is just as much formation as destruction. However, conditions may allow that after sufficient time an *old-age* stage may be reached, characterized by a strong differentiation in A and B horizons. Such a *chronosequence* occurring in the central U.S. is represented in Figure 39.

The effect of time is often obscured because in past times climates can have occurred that are different from the present-day climate. In such cases it is difficult to establish whether the soils are different as a consequence of time or of climate.



A Parent Material	B Young Soil (Regosol/Entisol)	C Mature Soil (Luvisol/Alfisol)	D Old Soil (Planosol/Ultisol)
Original material before soil development begins	Thin, solum, organic matter accumulation in A horizon, from which carbonates have been leached. Minimum weathering and eluviation.	Organic matter content is at a maximum. Has moderate clay accumulation in the B horizon and the solum is acid.	Very acid in reaction, severely weathered and has less organic matter than mature stage. Clay accumulation in B horizon has formed a clay pan. An A2 horizon exists.

Fig. 39. Time factor in soil formation.

6. SOIL FORMING PROCESSES

The action and particularly the interaction of the soil forming factors is expressed in processes that actually form the soils. A number of important soil forming processes will be discussed in this chapter: podzolization, ferralization, laterization (plinthization), gleization, movement of clay (lessivage), formation of calcic horizon, salinization, sodication and alkalinization.

6.1 PODZOLIZATION

Under certain conditions microbial decomposition of organic material is hampered and formation of stable, highly polymerized humus does not take place. Instead, there is a production of large amounts of fulvic and humic acids as well as low-molecular acids such as oxalic acid, citric acid, etc.. These water-soluble organic compounds move downward through the profile taking along iron and aluminium with which they form stable complexes (see Complexation, Section 3.2.6). This is the only way in which iron can move to any significant extent in well-aerated soils. The solubility of Fe-organic complexes may be several orders of magnitude higher than that of ferric ions.

At some depth a large proportion of these complexes is precipitated and partly decomposed. A certain part of the complexes will be lost to the groundwater. The major horizons that are formed in this way are (see Fig. 40):

O : Accumulated organic matter
A1 or Ah : Organic matter rich mineral topsoil
A2 (or E) : Leached (and often bleached "albic" horizon
B1 or Bh : Organic matter enriched horizon ┐ "spodic"
B2 or Bs : Sesquioxide-enriched horizon ┘ horizon
B3 : Transition horizon
C : Unaffected parent material (possibly with gley)

The most important conditions causing the accumulation of organic matter are: low pH, low base status, poor microflora (mainly fungi), low temperature, periodical saturation with water. Such conditions occur notably in the cooler humid climates (zonal podzols) but are also frequently found in the wet tropics on very poor parent materials or at higher altitudes where temperatures are sufficiently low.

The process
Because of the production of complexing organic compounds, chemical weathering in the A1 and A2 horizons is very intense. Iron and aluminium as well as divalent cations (Ca, Mg) are removed from the minerals and translocated downward as soluble organic complexes. When such complexes become saturated with metal ions they are insoluble and will precipitate. Supply of fresh, unsaturated complexes will remove some of the metal ions from the precipitated complexes and thus rendering them soluble again. In the beginning, the metal ion supply may be good enough to retard podzolization. After some time of leaching (*cheluviation*) a bleached A2, E, or *albic* horizon may become visible.

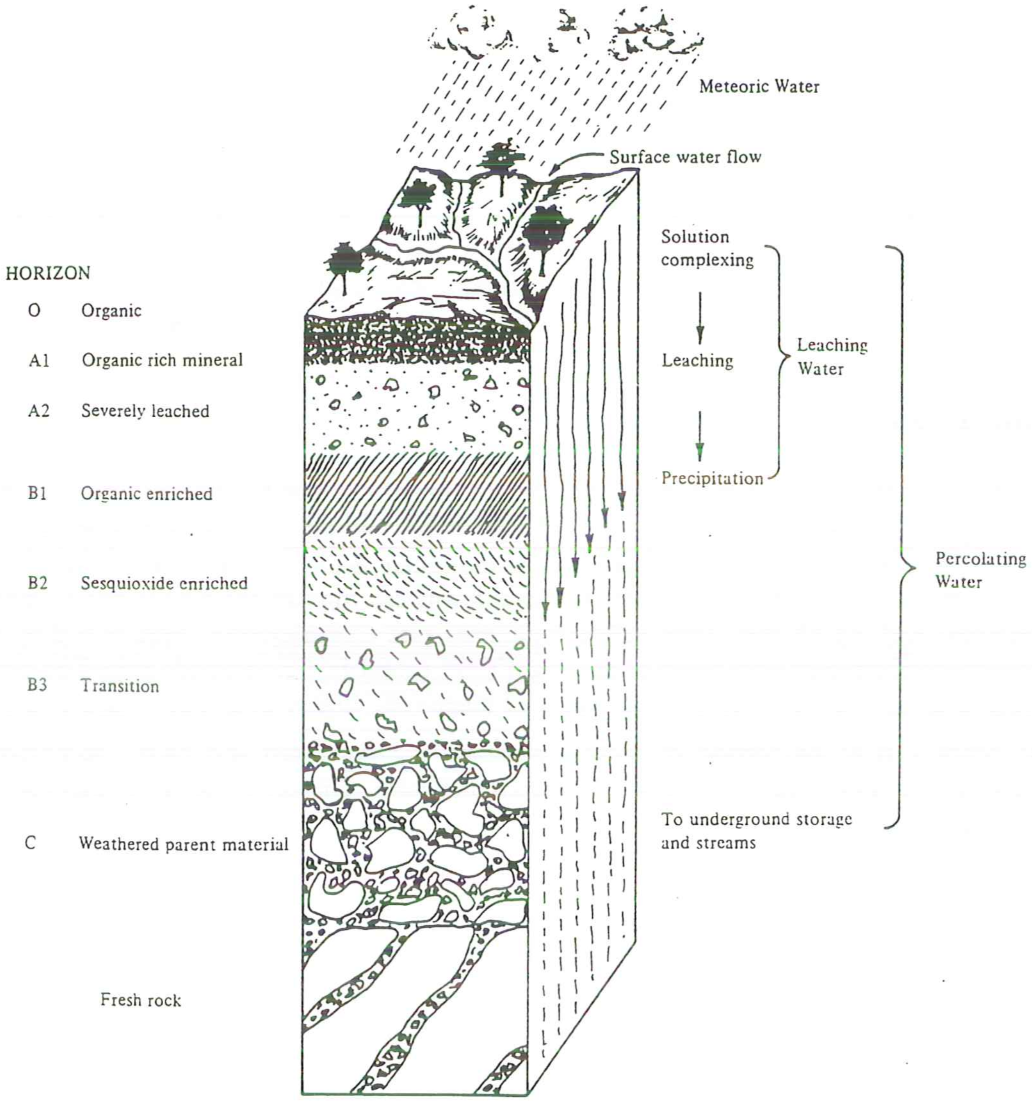


Fig. 40. A generalized podzol profile. The notes on the left side indicate a dominant characteristic for each horizon. The right side indicates water movement through the profile and the horizons subject to leaching and precipitation. Some leaching affects the whole profile. (From Finkl).

At some depth in the profile the organo-metallic complexes precipitate forming a B horizon or spodic horizon. This precipitation is due to one or more of the following processes:

a) formation of insoluble complexes
b) hydrolysis of soluble complexes (into organic acid and metal hydroxide)
c) decomposition (oxidation) of the organic component
d) stagnation at textural discontinuity.

Because in the A horizon sesquioxides are preferentially depleted, the remaining material becomes relatively richer in silica. Under such residually enriched conditions chlorite and mica will disappear but montmorillonite may persists for a long time although kaolinite is the only stable clay mineral in the albic horizon.

After deposition of the organo-metallic complexes in the B horizon, a redistribution of materials occurs. Because unsaturated organic materials reach the B horizon, the upper part of this will be depleted of sesquioxides - (hydr)oxides of Fe and Al) which are precipitated somewhat further down. This partial segregation leads to a Bh and a Bs horizon. Decomposition of the organo-metallic complexes liberate the sesquioxides. Aluminium may be used in clay mineral formation (chloritization) or co-precipitate with silicon in solution to form allophane and imogolite. Iron will precipitate as hydrous ferric oxides (e.g. goethite) and is often redistributed under reducing conditions in the wetter podzols which in extreme cases may result in a *placic* horizon *(iron pan).* In many podzols the Bs horizon is not or not conspicuously present because the sesquioxide content of the parent material is too low (Humods).

The often occurring fibres (micro-B horizons) under the main B horizon occur by precipitation of organometallic complexes that escaped from the overlying B horizon.

It is clear that podzolization is a very dynamic process since precipitation can ever be followed by redissolution, which may lead to very deep profile development *(giant podzols).*

6.2. FERRALIZATION (or FERRALITIZATION)

This process of soil formation occurs in a warm and humid climate under conditions of perfect drainage (no groundwater table, dry seasons may occur). Weathering consists mainly of dissolution, hydrolysis, hydration, carbonation and oxidation. There is an intensive biological activity causing a considerable CO_2 partial pressure in the soil. Furthermore, accumulation of organic matter does not take place and *chelation plays no role.* The result is a strong leaching of alkali and alkaline earth metals and silicic acid set free by the weathering. The residue consists of oxides and hydroxides of aluminium and iron which have a lower solubility than silica under the present conditions (pH 5-6, see Fig.41). The only clay mineral stable in this environment is kaolinite.

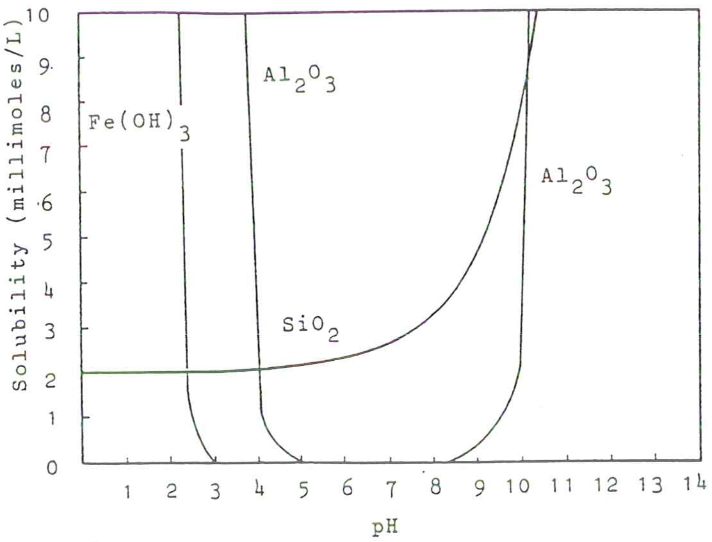


Fig. 41. The solubility of silica and sesquioxides.

The *Ferralsols/Oxisols* (formerly: *Latosols*) formed in this way can be very deep (tens of metres) and have indistinct horizon boundaries. The *structure* is granular or subangular blocky and *very stable,* clay movement is generally absent. The soils are very *porous* and have a very *good permeability.* The colours are red, reddish-brown, yellowish-brown and yellow, depending on the parent rock. The oxic horizon does not harden upon exposure to air (in contrast to plinthite). If the solum is very deep (e.g. thicker than 4 m), it is sometimes underlain by a mottled clay horizon. Otherwise the solum lies directly on the rotten rock (saprolite). Concretions are absent or few in amount and are then generally soft, only manganese concretions are frequently present. Due to the high porosity, stable structure, and depth, most Ferralsols are far less susceptible to erosion than other soils under similar topographical conditions.

6.3 LATERIZATION AND PLINTHIZATION

When a *groundwater table* is present, soil formation in a humid tropical climate with alternating dry and wet seasons leads to an entirely different situation than described in the preceding section for a freely drained system. Under such conditions gley-like features will develop which may lead to the formation of *plinthite* and, eventually, to *ironstone.* Formerly, in pedology, such a development was referred to as *laterization.* Of late, the definition of laterite has been rephrased several times by workers in different disciplines so that at present a somewhat confusing situation exists. We will here discuss laterization in the traditional sense as in detail described by Mohr, Van Baren and Van Schuylenborgh (1972).

The process can be divided into three stages (see Fig. 42).

Stage 1
The first stage is characterized by free or nearly free drainage and soil formation is similar to that described above for a Ferralsol.

Stage 2
Upon continuous weathering, the contents of clay minerals and sesquioxide colloids increase and the base level of erosion is approached, leading at a certain moment to the stage that the permeation of the soil with rainwater is slow enough to prevent free drainage in the wet season. Consequently a temporary groundwater table is established. In the dry season, however, the water can still drain off sufficiently and disappears. In this second stage (appearance and disappearance of groundwater), reduction processes will start, especially if the roots are penetrating deep enough. The second stage is therefore characterized by the occurrence of weathering and soil formation reactions under alternating reductive and oxidative conditions. The transformations during this stage are more or less comparable with those involved in *pseudo-gley* formation and they will be different from that of Stage 1. The material formed will show *mottling* (Zone III), whereas the upper part of the profile will have a uniform colour. Clearly, in this Stage 2 the *plinthization* process begins.

Stage 3
Upon further weathering and approach of the base level of erosion, drainage and evapotranspiration are no longer capable to remove all water even in the dry season and a permanent groundwater table will be established. *Normal gley* formation will be superimposed upon that of *pseudogley.* A zone of permanent reduction comes into existence (Zone IV, *"pallid"* zone) and rock weathering will then proceed under reducing conditions (Zone V). The groundwater table is still fluctuating, although the highest water level is lowered gradually upon weathering and the mottled horizon grows thicker. Part of the originally formed mottled clay is now only moistened by capillary rise above the water table and the upper part (Zone II) will not be moistened any longer. This part can dry to some extent and this is assumed to lead to the special characteristics of the plinthite material.
The drying causes a certain dehydration and crystallization of silica and "amorphous" iron oxihydrates into cryptocrystalline and crystalline hydroxides and oxides, thus cementing the material. Identification of plinthite in the field is more of an art than of scientific knowledge.

After some time, due to erosion of the surface soil, the plinthite may be at the surface as a *laterite* or *ironstone* crust.

It should be mentioned that the described process can be influenced by lateral influx of material and/or by influx with groundwater. This can lead to strong enrichment with both silica and iron oxides. Under certain conditions, iron moving laterally in percolating drainage water may eventually be precipitated at some point in the landscape where better aeration supervenes, and ironstone deposits form (see Fig. 43).

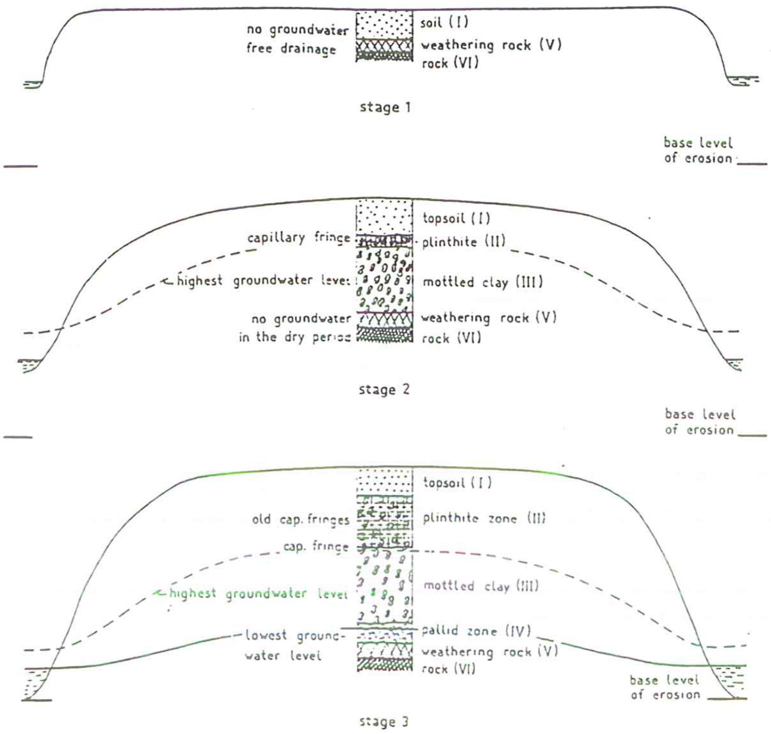


Fig. 42. Schematic representation of the three stages in the genesis of an Oxisol with plinthite. (After Mohr *et al.*, 1972).

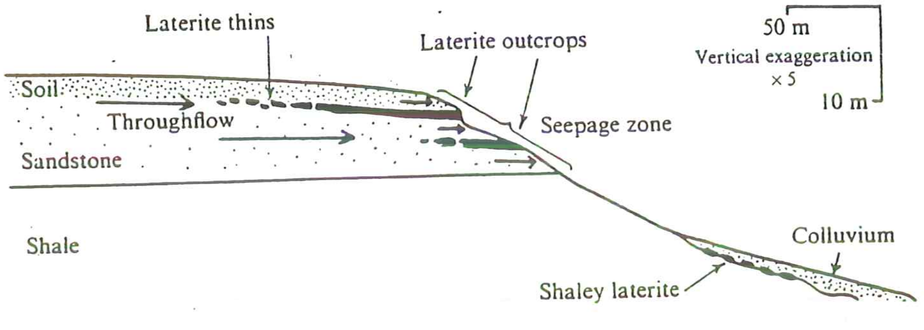


Fig.43. Groundwater laterite outcrops.

A model description of the zones can be as follows:

I. *Surface zone*
Red colour, some iron concretions may be present. Thickness up to 60 cm. Clear lower boundary.

II. *Plinthite zone*
Red material, mottled with yellow, white and even violet; generally clayey texture, but, if formed from material rich in quartz, often sandy. Granular to subangular blocky structure. Firm when moist, however hardens strongly upon exposure to the air and can then be used as building material. Pores frequently filled with whitish or greyish material. Upper part often hardened to *laterite* or *ironstone,* in particular if Zone I has been partly or completely eroded. Sometimes rich in gibbsite and silicates (especially kaolinite). May have a considerable thickness: 2 m is common. Diffuse lower boundary.

III. *Mottled clay zone*
Very similar in appearance to II, but soft and does not harden upon exposure to the atmosphere. May be absent in the profile or may have considerable thickness (10 m). Clear lower boundary.

IV.*Pallid zone*
A bleached zone poor in iron. Sometimes rock structure still present. May be absent and, if present, may be the actual or paleo zone of permanent reduction. May also be very thick (several metres). Diffuse lower boundary.

V. *Weathering rock, rotten rock, or saprolite zone*
Consists of weathered rock and has light grey colour if Zone IV is present, or is strongly mottled with red, yellowish brown and grey when Zone IV is absent. The zone is thin (a few millimetres to a few centimetres) in the case of basic rocks and several metres thick in the case of acidic rocks. Generally rich in kaolinite.

VI. *Parent rock*

The presence of all mentioned zones is exceptional, because the soils are either very old (truncated profiles) or not old enough, so that time was insufficient to form all zones.

6.4 GLEYING OR GLEIZATION

6.4.1 Introduction

When (part of) a soil remains saturated with water for some time (not all the time) it is subject to gleying. Saturation with water (that is, all pores are filled with water) does not mean that all oxygen is immediately excluded as some can be dissolved in water. However, oxygen will be quickly used up by the micro-organisms that use it for their metabolism. Under such conditions the *redox potential (Eh)* is lowered. This may reach the stage (below about 250 mV) that certain soil components are reduced, notably oxides of iron and manganese, but also nitrates, sulphates and elemental sulphur. This can happen because after depletion of gaseous oxygen, anaerobic and facultative anaerobic micro-organisms take over from the aerobic species and they are able to use the oxygen of oxidized components for their metabolism. Figure 44 shows some threshold redox potentials of some components.

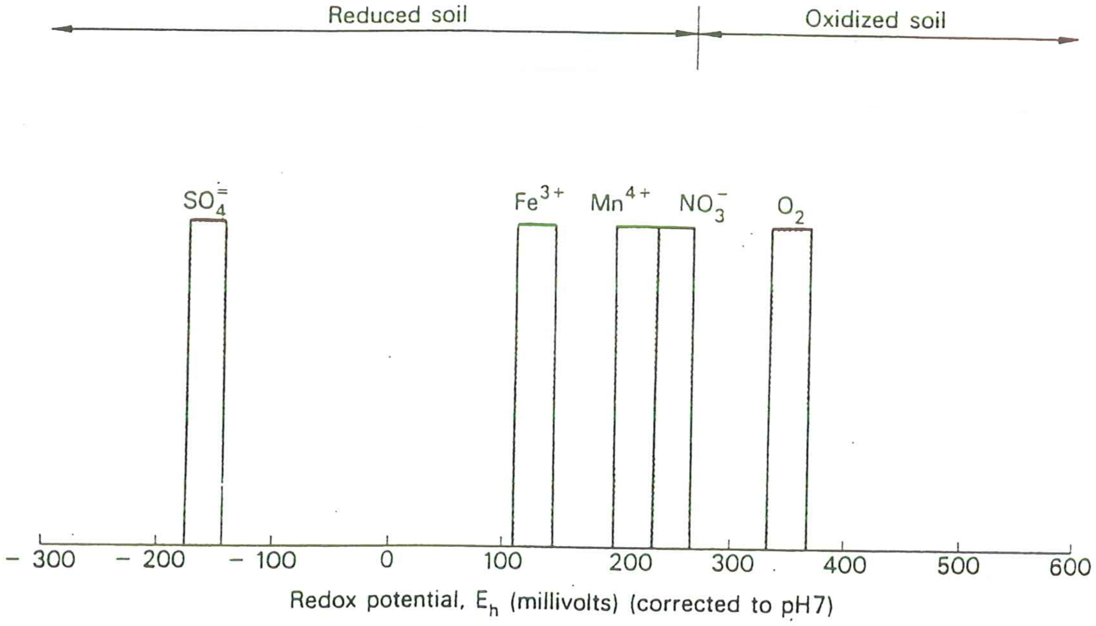


Fig. 44. Threshold redox potentials at which oxidized species become unstable. (After Patrick and Mahapatra).

Reduction gives soils characteristic greenish or bluish shades. The most distinctive feature of gley, however, is a scattering of brown/red and black mottles. They result from a long-term seasonally alternating dry and wet conditions causing alternating oxidation and reduction. Precipitation of newly oxidized compounds takes place preferentially on already existing oxide surfaces. Ensuing reduction is not capable to dissolve these coarse grained concretions within one season. The zones where gleying is most active are along the wider pores in the soil since those are the first to be filled with water when the rain starts (conducting pores) and also the first to be emptied when conditions become dryer (see Fig. 36).

Two fundamentally different types of gley can be distinguished:

1) gley caused by groundwater: *groundwater gley* or *normal gley*
2) gley caused by a perched (hanging) water table: *surfacewater gley* or *pseudogley.*

A very schematical picture of the two types of gley is drawn in Figure 45. It shows the situation along a crack between two structural elements (peds) but the situation can equally be found along other pores such as old root and worm channels.

6.4.2 Groundwater gley or normal gley

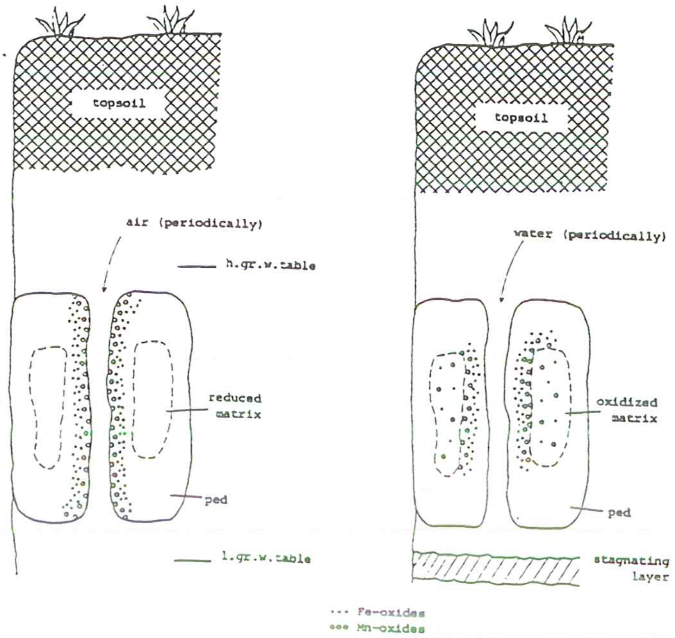
This type of gley is due to wetness caused by a groundwater table and thus results from *limited external drainage.* At the end of the wet season the groundwater is at its highest level and the solum is in the reduced state. After the dry (or less humid) season has started, the groundwater table is lowered. The soil dries and forms cracks. These cracks and other larger pores will be filled with air and thus, oxygen enters the soil. The reduced Fe^{2+} and Mn^{2+} ions, present in the soil solution in the smaller pores along the cracks will oxidize and precipitate along these cracks forming brown and black coatings. As a result, the concentration of Fe^{2+} and Mn^{2+} in this zone is lowered whereas in the wet matrix it is still high. Consequently, Fe^{2+} and Mn^{2+} ions diffuse from the interior of the structural units towards the air-filled spaces and, arrived at these places, oxidize. In this way diffusion contributes to the already started precipitation and accumulation of iron and manganese oxides and hydroxides. The interior of the structural units is therefore gradually deprived of iron and manganese and its colour turns into the colour of the soil minerals, which are usually white, grey or green. After air entrance into the wider pores, some oxygen will diffuse into the still wet matrix causing precipitation at some distance from the ped surface. Here, *segregation* of the Fe and Mn oxides will occur because Fe^{2+} oxidizes "easier" (at a lower Eh or oxygen partial pressure) than Mn^{2+} (see Fig. 44) so that generally the Fe oxides occur further away from the cracks than the Mn oxides. Of course, there is a gross overlap between the two since the zone of precipitation will gradually move into the soil matrix as a result of drying. Furthermore, each year the rate and pattern of drying and wetting will be different. In the course of many years the iron and manganese accumulations turn into concretion-like bodies and even into real and hard concretions. Below the lowest groundwater table the soil remains always saturated with water: oxygen never enters into this system and oxidation never occurs. Gradually the subsoil will be deprived of iron and thus obtains its homogeneous grey, green or blue colour.

6.4.3 Surfacewater gley or pseudogley

This type of gley is caused by wetness due to *limited internal drainage* of the profile causing a perched or hanging water table. Such impeded drainage is usually caused by a stagnating layer somewhere in the profile, e.g. a textural horizon or parent rock, or the soil material itself when it is sufficiently heavy textured.

At the end of the dry season the surface soil is dry and the subsoil still somewhat moist; the whole solum is well-aerated. Narrow cracks (the width depending on the clay content) are observed between structural (frequently prismatic) units. It appears that the roots have followed, in search for water, the paths of least resistance, viz. the cracks and root channels.

Now, the first rains start; frequently, these are heavy showers. The water follows preferably the paths of least resistance (the cracks and worm tracks) and moisten the soil along these pathways.



Groundwater gley
or Normal gley

Surface water gley
or Pseudogley

Fig. 45. Diagram indicating the sites of precipitation of iron and manganese oxides in normal and pseudogley.

Arrived at the water-stagnating layer, it stops and the cracks are completely filled with water. The new situation is that the structural units are surrounded by water-saturated layers. The inside of the units remains relatively dry. Because of the swelling of the saturated layers, the cracks narrow and close; the roots are deprived of O_2 and die. Upon microbial attack of the dead roots and other organic matter present, the oxygen is rapidly consumed and strongly anaerobic conditions are created along the former cracks. Fe(III) and Mn(IV) oxides are reduced to Fe(II) and Mn(II) ions which diffuse to the interior of the structural units where conditions are still aerobic. The Fe^{2+} and Mn^{2+} ions are then re-oxidized at those places where enough O_2 is present to do this (or better: where the Eh is high enough). In this way the immediate surroundings of the cracks are deprived of iron and manganese and a rim of Fe(III)-oxides is formed at some distance from the cracks. The Mn^{2+} ions diffuse further away than the Fe^{2+} ions as Mn^{2+} ions need a higher Eh to be oxidized than Fe^{2+} ions (see Fig. 44). As a result, the MnO_2-separations are observed in the innermost parts of the structural units: the place of the highest O_2-pressures. When this process is frequently repeated, broad white or grey tongues are formed along the cracks with a clear rim of iron oxide separations, whereas the manganese oxide accumulations are found in the very centre of the structural units.

Some reactions involved in the oxidation/reduction of iron and the function of organic matter and micro-organisms.

$4Fe^{2+} = 4Fe^{3+} + 4e^-$ (oxidation of ferrous iron) (1)

$O_2 + 4e^- = 2O^{2-}$ (reduction of oxygen) (2)

$2O^{2-} + 2HOH = 4OH^-$ (reaction of oxygen radicals with water) (3)

$4Fe^{3+} + 12HOH = 4Fe(OH)_3 + 12H^+$ (hydrolysis of ferric iron) (4)

$\overline{\phantom{4Fe^{3+} + 12HOH = 4Fe(OH)_3 + 12H^+}}$ +

$4Fe^{2+} + O_2 + 12H_2O = 4Fe(OH)_3 + 8H^+$ (hence 2 H^+ are produced per oxidized Fe) (5)

Oxidation of Organic Matter (by bacterial activity) uses O_2:

Overall reaction: $(CH_2O) + O_2 = CO_2 + H_2O + \text{energy}$ (6)

Half-reaction: $(CH_2O) = C^{4-} + H_2O + 4e^-$ (oxidation of carbon) (7)

Half-reaction: $O_2 + 4e^- = 2O^{2-}$ (reduction of oxygen) (8)

Finally: $C^{4+} + 2O^{2-} = CO_2$ (reaction of radicals) (9)

If oxygen is depleted, e.g. because of water-logging, then the bacteria will turn to other acceptors of the electrons that are liberated during the oxidation (digestion) of the organic matter (their food!). Such acceptors are various oxides in the soil, notably those of Fe and Mn but also of N and S (nitrates and sulphates).

For example: $4FeOOH + 4e^- + H_2O = 4Fe^{2+} + 12OH^-$ (10)

Then, instead of (9): $C^{4+} + 2H_2O = CO_2 + 4H^+$ (11)

Thus, reduction of iron oxide raises the pH of the soil (excess of OH^-). Usually, however, carbonic acid formed by the CO_2 reduces this effect.

The formed Fe^{2+} is quite soluble and may be translocated in the profile (or even be removed). If not removed it will at a later stage, when oxygen re-enters the soil, be oxidized as indicated above.

6.5 CLAY MIGRATION

The occurrence of an argillic, argic, kandic, natric or, in general, a Bt horizon of some depth in the soil profile usually implies that physical transport of clay has taken place. Although some of the clay can have been formed by *in situ* weathering or by neoformation from components dissolved in percolating water, it is generally believed that the major process is migration of clay particles suspended in percolating water. It is, therefore, interesting to investigate:

1. the conditions under which clay becomes mobile;
2. the conditions of transport of clay;
3. the conditions of deposition of the clay.

6.5.1 Mobilization

Clay migration requires that the clay be *dispersed* or *peptized* so that it can remain in suspension to be transported. *Flocculated* clay, by contrast, is physically stable and immobile. In the peptized state, clay particles repel each other so that they remain apart; in the flocculated state the clay particles attract each other to form some kind of aggregate (see Fig. 46).

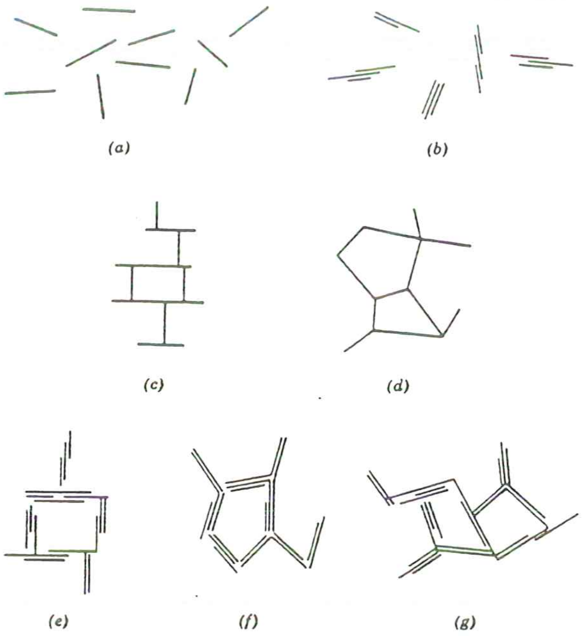


Fig. 46. Modes of particle association in clay suspensions.
(a) peptized, *(b)* aggregated but peptized, *(c)* to *(g)* flocculated.

Whether there is peptization or flocculation depends on the thickness of the layer of ions that satisfies the negative charge of the particle ("electrical double layer", see Fig. 47). Ions attracted to particle surfaces are distributed so that their concentration is highest close to the surface, and concentration decreases away from the surface. When two particles, each with positive ions attracted (adsorbed) to their surface, move toward one another, the initial reaction is one of repulsion. This results in peptization. If, however, under certain conditions the clay particles can move closer together, Van der Waals forces take over and attraction and flocculation occur. In practice, two major conditions can effect such a closer approach of the clay particles:

1. at high particle concentration in suspension
2. when the electrical double layer is thin, or rather, compressed

The first condition is simply a function of the water content of the suspension (which makes lighter textured soils more liable to clay migration), whereas the thickness of the double layer is a function of *a)* the type of attracted cation and *b)* the salt concentration of the soil solution. The

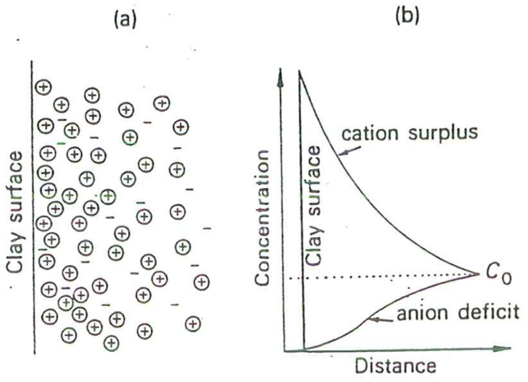


Fig. 47. The ionic distribution at a negatively charged clay surface.

higher the valency of the cation, and the higher the concentration of the soil solution, the more compressed the double layer. Thus, Ca-saturated clays are better flocculated than Na-clays and, therefore, clay migration seldom takes place in calcareous soils but is common in sodic soils (solonetz). In saline soils however, because of a high salt concentration, the clay is flocculated and clay migration does not take place (not even when the salt is mainly Na-salt). Table 7 shows the relative effect of valency and ion concentration.

Table 7. Approximate "thickness" of the electric double layer as a function of electrolyte concentration at a constant surface potential.

Concentration of ions of opposite charge to that of the particle, mmol/dm^3	"Thickness" of the double layer, Å	
	Monovalent ions	Divalent ions
0.01	1000	500
1.0	100	50
100.	10	5

Low pH values (<4.5) in the soil are generally favourable for flocculation as Al^{3+} ions are then usually present and will be adsorbed. The type of clay mineral also has an influence. Generally, the higher the charge density, the easier to peptize the clay. Thus, montmorillonitic clay will easier migrate than kaolinite. However, this aspect is obscured by the particle-size effect: the smaller the clay particles, the easier they migrate in the soil. Montmorillonitic clay is usually finer grained than kaolinitic clay.

6.5.2 Transport and deposition *(argeluviation and argilluviation)*

In Soil Taxonomy a satisfactory explanation of the mechanism of the formation of the argillic horizon is given:

'Wetting of a dry soil seems to lead to disruption of the fabric and to dispersion of clay. Once dispersed, the clay moves with the percolating water and stops where the percolating water stops. Water percolating in non-capillary voids commonly is stopped by capillary withdrawal into the soil fabric. During this withdrawal the clay is filtered out and deposited on the walls of the non-capillary voids. This explains why illuvial clays are so commonly plastered on the ped faces and on the walls of pores. Such a mechanism for clay movement and deposition is favoured in several ways by a seasonal moisture deficit:

a) as mentioned above, wetting a dry soil favours the dispersion of clay
b) on drying, cracks form in which percolation of gravitational water or water held with

low tension can take place

c) the halting of percolating water by capillary withdrawal is favoured by the strong tendency for dry soil to take up moisture.'

'Water ordinarily does not remain in the non-capillary voids; rather, it will move into and be transported through material having pores of capillary size. Movement of water in non-capillary voids occurs only if the rate or water addition exceeds the capillary conductivity of the soil. As conditions for water percolation in large voids do not usually prevail, clay movement and deposition are likely to be infrequent and sporadic. Thus, a long period of time usually is necessary before the cumulative effect of clay movement becomes readily discernible.'

'Capillarity can have still more to do with the downward movement of water (as well as into the peds). If a lower soil horizon is considerably coarser in texture, capillaries are broken and the water tends to hang in the fine capillaries above (see Fig. 36). As the water evaporates or is withdrawn by roots, suspended or dissolved materials are left. This action even accentuates the original differences in fineness of capillaries.'

Some additional observations are made. Peptization in the surface soil is favoured by the relatively lower electrolyte concentration due to leaching; at some depth the concentration will be higher and flocculating can take place.

An essential prerequisite for clay movement is the presence of pores and cracks that are sufficiently wide for clay transport. The minimum diameter seems to be 20 μm. Especially in the beginning of the wet season transport is facilitated by cracks developed during the preceding dry season.

6.6 THE CALCIC HORIZON

Soils in semi-arid and arid regions commonly have a calcic horizon at some depth below the surface. If the climate is dry enough or the surface erosion intensive enough, these may extend to the surface.

The origin of the calcic horizon involves the following equilibria:

$$CO_2 + H_2O \leftrightarrow H_2CO_3^0 \qquad (23)$$

$$H_2CO_3^0 \leftrightarrow HCO_3^- + H^+ \qquad pK = 6.4 \qquad (24)$$

$$HCO_3^- \leftrightarrow CO_3^{2-} + H^+ \qquad pK = 10.3 \qquad (25)$$

$$CaCO_3 \leftrightarrow Ca^{2+} + CO_3^{2-} \qquad pK = 8.3 \qquad (26)$$

These equilibria show the relation between dissolved gaseous CO_2, the pH and the solubility of calcite. The relation of the carbonate species with the pH, using the equilibrium constants, is shown in Figure 48. (The pK of an equilibrium reaction is the pH value at which the reactants and products are present in a 1 : 1 molar ratio).

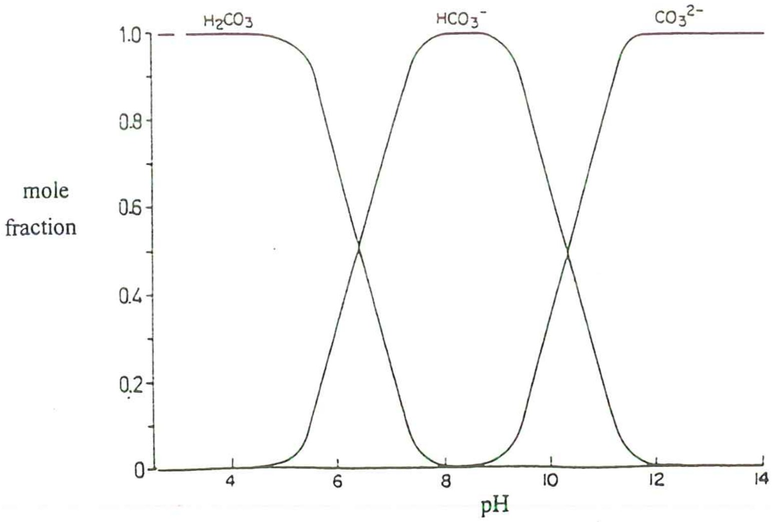


Fig. 48. Distribution of aqueous CO_2 species with pH.

In practice, for the calcic horizon the relevant equilibria can be condensed as follows:

$$\begin{array}{c} CO_2 + H_2O \\ \uparrow\downarrow \\ CaCO_3 + H_2CO_3 \leftrightarrow Ca^{2+} + 2\ HCO_3^- \end{array}$$

An increase in CO_2 content in the soil air or a (consequent) decrease in pH will drive the reaction to the right; calcite will dissolve and Ca^{2+} and HCO_3^- will move with the soil water. Dissolution is also enhanced by increasing the amount of (fresh) water moving through the soil as this decreases the concentration of Ca^{2+}. Reversely, precipitation of carbonate occurs under conditions that drive the reaction to the left, i.e. a lowering of CO_2 pressure, a (consequent) rise in pH, or an increase in ion concentration to the point where saturation is reached (where the solubility product of Ca^{2+} and CO_3^{2-} is exceeded) and precipitation takes place.

The relation between the CO_2 pressure, the pH and the Ca^{2+} concentration of the soil solution is given in Figure 49.

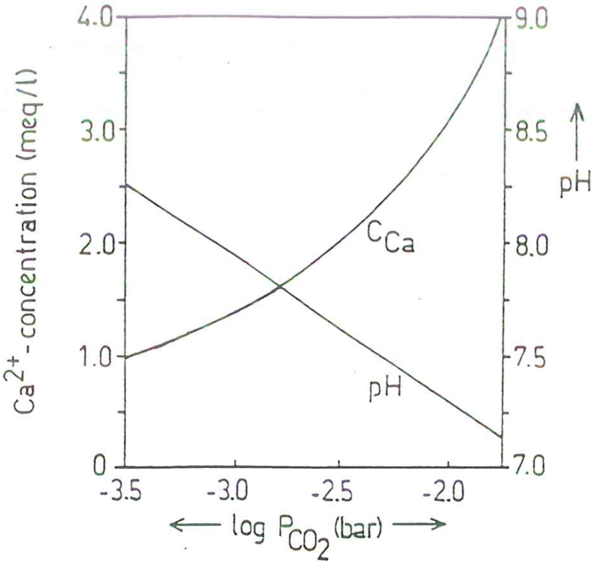


Fig. 49. Solubility of calcite and resulting pH-value in water at different CO_2-pressures.

Because the CO_2 pressure and the pH are directly related it can be stated that the precipitation/dissolution equilibrium of carbonates in the soil is governed by two factors:

1) the CO_2 pressure
2) the moisture content of the soil

It is now easy to explain the formation of the calcic horizon. The biological activity is highest in the A horizon in most soils. The partial pressure of CO_2 is high as a result of CO_2 produced by root and micro-organism respiration (organic matter decomposition). This pressure in soil air may be 10× to 100× that in the atmosphere! (see also Table 3). Consequently, calcite will dissolve and move downward. On its way down, the water may take up more dissolved calcite. This increase in concentration together with a decrease in CO_2 pressure (fewer roots and micro-organisms lower in the profile) will result in a precipitation of the calcite at a certain depth. Precipitation of calcite from ascending groundwater occurs when the solubility product is exceeded by evaporation of the groundwater at a certain level in the profile (and water vapour cannot carry solids).

Deposition of $CaCO_3$ along and in old root channels and worm tracks can be explained in a similar way. When the soil water reaches such a channel, water will evaporate when the channel is empty and the carbonate precipitates. This is enhanced by the much lower CO_2 pressure in the channel as compared with the soil since the channel is usually in better contact with the atmosphere.

6.7 SALT-AFFECTED (HALOMORPHIC) SOILS

Whenever there is no net downward movement of water through the profile there exists a hazard of *salinization* (= accumulation of salt). Thus, all arid and semi-arid region soils should be considered to be subject to potential salinization. Irrigation may strongly add to this hazard, especially if all water is consumed by the vegetation or if no excess water passes through the profile.

Halomorphic soils are not very extensive but their agricultural significance is considerable. Salinity causes problems for agriculture as the high salt concentration is accompanied by a high osmotic pressure reducing or prohibiting plant growth. Furthermore, the composition of the exchange complex is changed: usually the proportion of Na is increased *(sodication;* formerly: *alkalization)* yielding physically inferior soils due to peptization of the clay.

6.7.1 Processes

6.7.1.1 Salinization

This is the accumulation of soluble salts in the profile, whether by natural processes or as a result of irrigation. The groundwater table or the capillary zone above it must reach into the soil profile. Accumulation is effected by evaporation of water above the capillary zone. Depending on the groundwater level, capillaries may reach the surface causing salt to be deposited on top of the soil: such a saline soil is called an *"external solonchak"*. If the salt is precipitated below the surface it can be called an *"internal solonchak"*. The salt accumulation is illustrated in Figure 50. Salinity is measured indirectly as the *electrical conductivity (EC)* of a soil extract, usually a saturation extract. As indicated earlier, saline soils usually have a good structure (apart from clogging by salt accumulation) because of the flocculated state of the clay.

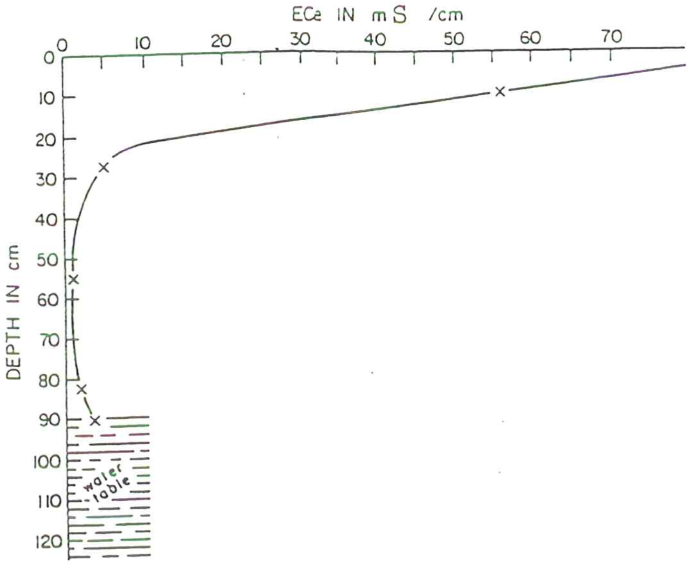


Fig. 50. Typical salinity profile in soil exposed to a high water table.

6.7.1.2 Sodication or alkalization (or solonization)

This is the process whereby the exchange complex acquires an appreciable saturation with *Na* ions. This is effected by prolonged or repeated contact with water in which sodium predominates over other cations. Since sodium has the lowest replacement ability of all the usual exchangeable cations it is necessary that its concentration in the soil solution substantially exceeds that of other cations.

Therefore, it is possible that salinization occurs without sodication (e.g. in the presence of Ca and Mg chlorides and sulphates). Sodication may be caused by groundwater, irrigation water or flooding with seawater and can take place within a few years.
Sodic soils or *solonetz* usually have a bad structure since Na saturation easily causes peptization of the clay and compact structures result. Under sufficient leaching clay may migrate downward causing formation of an argillic or argic B horizon often with a characteristic columnar structure. Irrigation has the hazard of sodication if not sufficient calcium ions are present or added. (This cannot cause a Bt within a few years of course).
An early stage of the solonetz may be recognized as a *solonetzic solonchak.*

6.7.1.3 Alkalinization

When sodication occurs by sodium carbonates rather than sulphates or chlorides (irrigation!) this will be accompanied by a *rise in pH* (due to hydrolysis: carbonic acid is a weak acid and NaOH a strong base). Such high pH values are precluded by the presence of Ca and Mg ions since their carbonates are only slightly soluble and the carbonates will precipitate.
Alkaline sodic soils will occur when in solution ($CO_3^{2-} + HCO_3^-$) > ($Ca^{2+} + Mg^{2+}$) , i.e. when there is so-called *"residual alkalinity".*
High pH values are very restrictive for plant growth since it renders many (micro) nutrients insoluble and thus unavailable while others can become toxic.
The chemical properties of halomorphic soils are summarized in Table 8.

Table 8. Chemical criteria for classification of salt-affected soils.

Soil	EC_e (mS/cm)	ESP*	pH
Saline	> 4	< 15	< 8.5
Saline-sodic	> 4	> 15	< 8.5
Sodic	< 4	> 15	> 8.5

* ESP = Exchangeable Sodium Percentage

6.7.1.4 Solodization

This is a degradation process caused by leaching with fresh (or salt-poor) water of a previously sodicated (solonized) soil. By hydrolysis, the exchangeable sodium is replaced by hydrogen ions and the soil eventually becomes acid. The clay remains peptized and there is intense clay translocation. The topsoil becomes sandy and bleached. Wind erosion may follow. These soils are called *solodized solonetz* and in their final stage *solod.*

The maturity sequence from solonchak to solod caused by a gradually falling water table is shown in Figure 51.

6.7.2 Reclamation

In principle, salinization is easily reversed by leaching with fresh water. The problem, however, is that leaching implies dilution of the soil solution which may lead to peptization (if a substantial proportion of the adsorbed cations consists of Na^+) causing a structure deterioration and consequently permeability and drainage problems. Therefore, sodic soils are difficult to reclaim. Liming can only be effective if the pH is not too high. Otherwise gypsum, having a higher solubility, is to be preferred. Even gypsum may have too low a solubility in certain cases and then the expensive $CaCl_2$ can be the answer. In some cases, elemental sulphur may be used, which is oxidized in the soil to sulphuric acid. This lowers the pH, dissolves the lime while the structure remains intact. Clearly, in all cases good drainage is essential.

Sodium Adsorption Ratio (SAR)

Sodium saturation of the exchange complex takes place when the soil comes into contact with a solution having a high value of the *"reduced ratio"* of the concentration of Na ions to divalent (Ca and Mg) ions, also referred to as the *Sodium Adsorption Ratio (SAR):*

$$SAR = Na / \sqrt{½(Ca + Mg)}$$

(concentrations expressed as *me/L* or $mmol_c/L$)

Such a condition is likely to occur upon concentrating the soil solution (as is typical for the salinization process). If a solution containing Na, Ca, and Mg ions is concentrated with a factor x, then the *SAR* increases with a factor $\sqrt{x}$! This means that the effect of Na increases when the solution evaporates so that *salinization* may well lead to *sodication* (alkalization). The reverse is also true, i.e. dilution of the soil solution with fresh water (e.g. rain water) will reduce the *SAR* and decrease the proportion of Na on the complex (de-sodication).

Sodication may be enhanced if the anions in the system are dominantly sulphate as the solubility of $CaSO_4$ is rather limited, whereas Na_2SO_4 has a high solubility.

A similar problem occurs when irrigation water is rich in carbonate. Such water can only be high in carbonate when the cations are alkali (usually Na) which is expressed by a high pH (> 8.5). The solubility of Ca and Mg carbonates at such high pH values is negligible *(residual alkalinity,* see main text) so that water high in carbonate will have a high *SAR* even if the total salt content is not very high. In this way *alkalinization* (= high pH) will also lead to sodication.

Thus, sodication of soils can result from two processes: salinization and/or alkalinization. To assess the (potential) hazard of the occurrence of these processes in a particular case, determination of the *SAR* of the soil solution or the irrigation water is very useful.

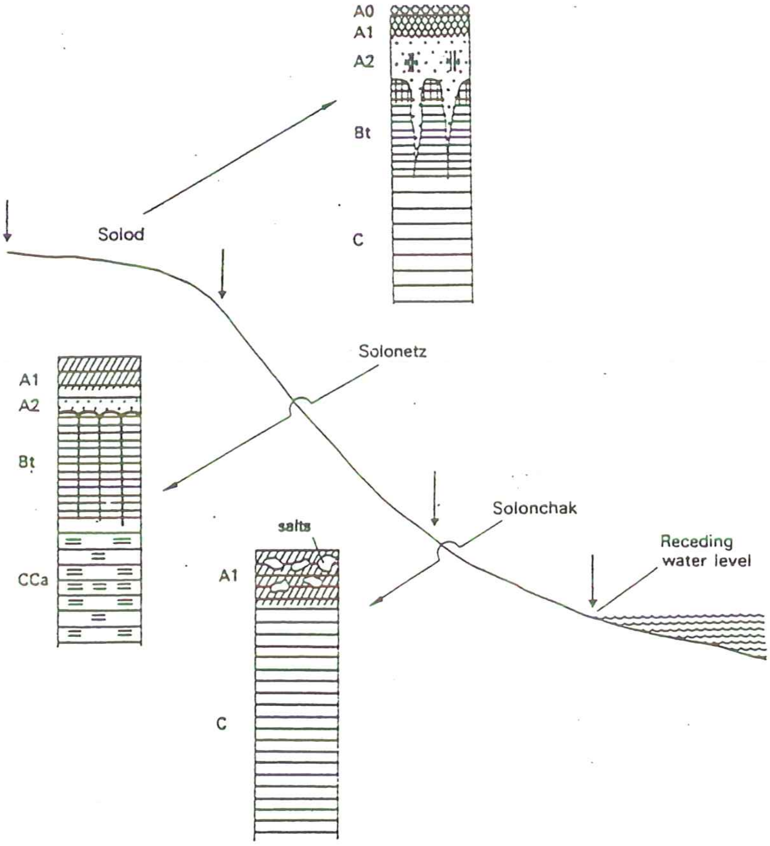


Fig. 51. Solonchak-solonetz-solod maturity sequence resulting from a falling water table.

7. SOIL FERTILITY

Soil is the substrate on and in which plants grow. The essential factors are light, nutrients, air, and water. The latter two will be discussed in section 1.8 on soil physics. The present section deals briefly with the chemical aspects of soil fertility.

The chemical environment relevant to plants is determined by the pH of the soil (acidity) and the availability of macronutrients and micronutrients. The relative uptake of nutrients is schematically represented in Figure 52.

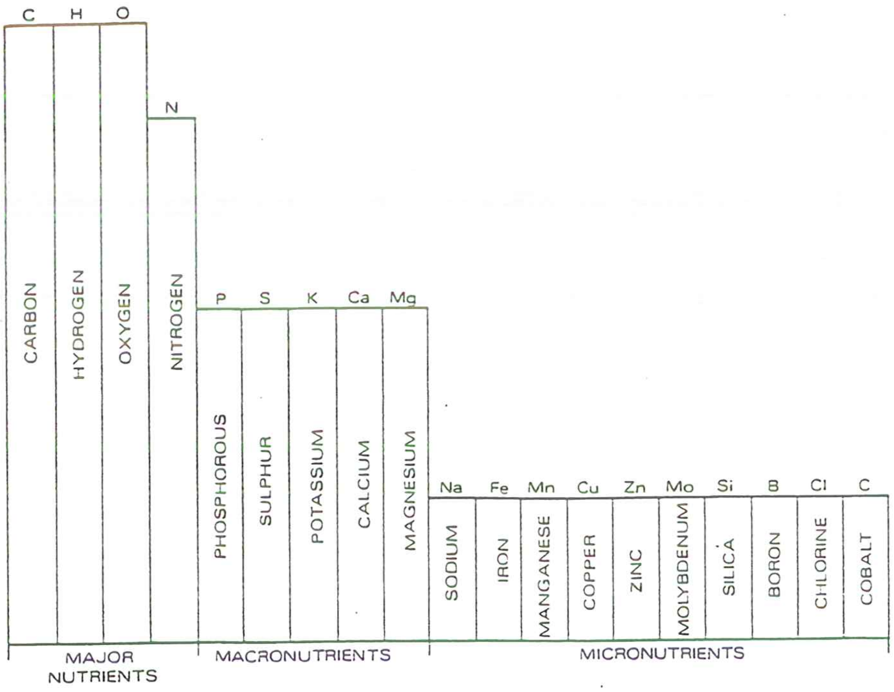


Fig. 52. Schematic relative uptake of nutrients by plants.

7.1 SOIL REACTION, ACIDITY (pH)

Soil acidity is expressed as the concentration H^+ ions (protons) in the soil solution. The pH of soils can vary widely from somewhat below 3 to over 11 (see Fig. 53). The significance of the pH is that it determines the availability of nutrients and the living conditions of micro-organisms.

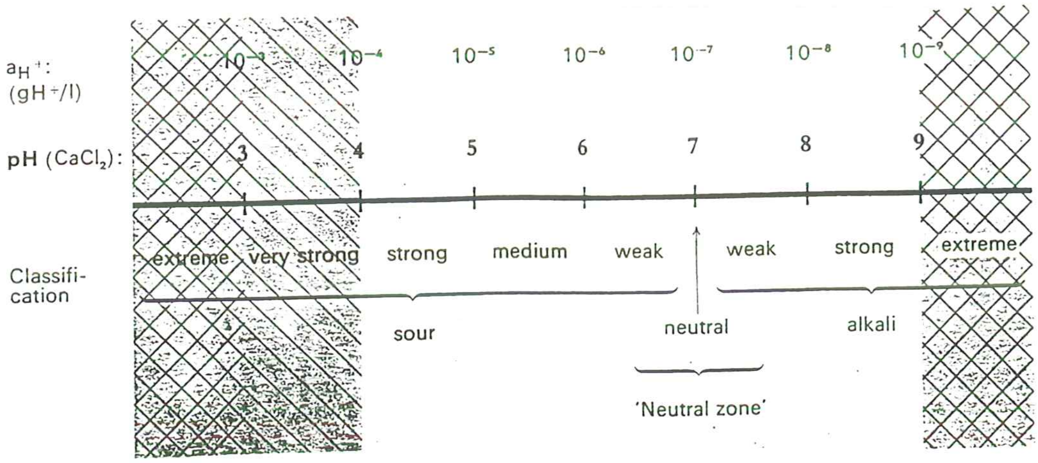


Fig. 53. The pH range in soils.

7.1.1 Measurement of acidity

The pH is measured in a suspension of soil particles with water. Accurate measurements can be done with a pH meter, a rough and quick estimate can be obtained by dipping indicator paper in the suspension. The pH is an *intensity* factor, it gives no information as to the amount (*capacity*) of acidity. This can be obtained by titrating a weighed amount of soil.
Often the pH is measured in two ways: one in a suspension with water, and the other in a suspension of a salt solution (usually KCl or $CaCl_2$). The cations will bring all adsorbed H^+ ions into the suspension by ion exchange thus giving some information about the amount of acidity. Therefore, the pH-KCl or pH-$CaCl_2$ is usually (not always) lower than the pH-H_2O

7.1.2 Sources of acidity

There are various sources of H^+ in the soil:

- From CO_2 respiration of plants and micro-life: $CO_2 + H_2O \leftrightarrow H_2CO_3^0$ (27, 23)
- Direct production by plant roots (see Fig. 55)
- Humification of organic matter producing multitude of organic acids
- Oxidation reactions of weathering (e.g. Reactions 16 and 17) and biological oxidation (e.g. NH_4 to NO_3).
- Pollution (e.g. SO_2, NO_x)
- Manuring with acid fertilizers e.g., ammonium sulphate.

7.1.3 Nutrient availability

The availability of nutrients is related to the pH:

- through the solubility of the compounds in which the nutrients occur.
- competition of H^+ with nutrient cations such as K^+, NH_4^+ Ca^{++}, Mg^{++}.

At very high and very low pH values some elements become unavailable whereas others may be available in too high concentrations and become toxic. The competition aspect entails that at low pH the concentration of other cations than H^+ will be low. Generally, pH values close to neutral (6-7) are most favourable. This is illustrated in Figure 54.

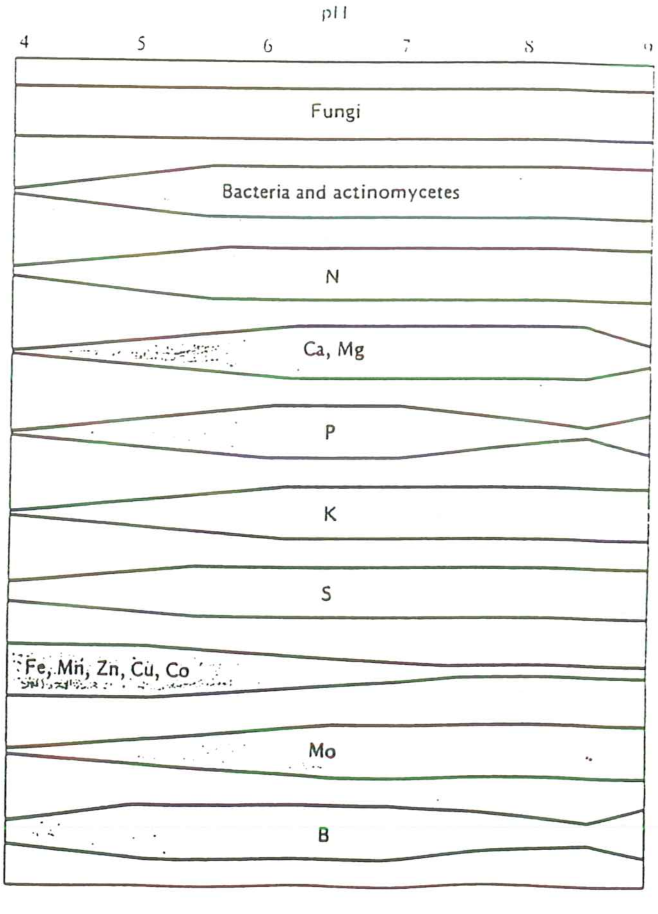


Fig. 54.

Relationships existing in mineral soils between pH on the one hand and the activity of microorganisms and the availability of plant nutrients on the other. The wide portions of the bands indicate the zones of greatest microbial activity and the most ready availability of nutrients. When the correlations as a whole are taken into consideration, a pH range of about 6–7 seems to promote best the availability of plant nutrients. In short, if soil pH is suitably adjusted for phosphorus, the other plant nutrients, if present in adequate amounts, will be satisfactorily available in most cases.

7.2. Ion Exchange

As was mentioned before, cations can be held by adsorption on negatively charged surfaces of clay minerals and soil organic matter. Figure 55 schematically shows this way of storage of ionic nutrients. After release form the minerals through weathering (or form other sources such as fertilizers) the ions pass to the charged surface usually via the soil solution. By exchange for (usually) H^+ the nutrient ions are released and given back to the soil solution after which they can be taken up by the plants (see Fig. 56).

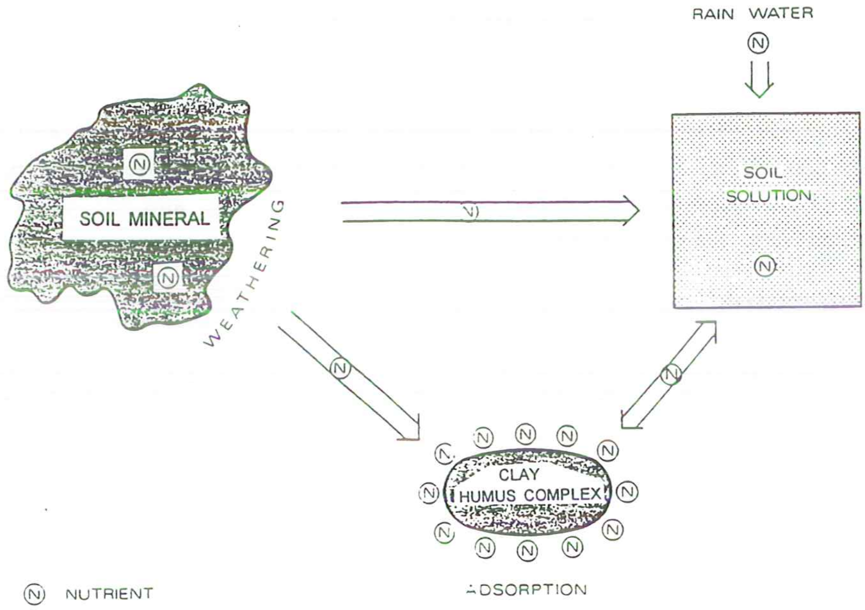


Fig. 55. Adsorption of nutrients.

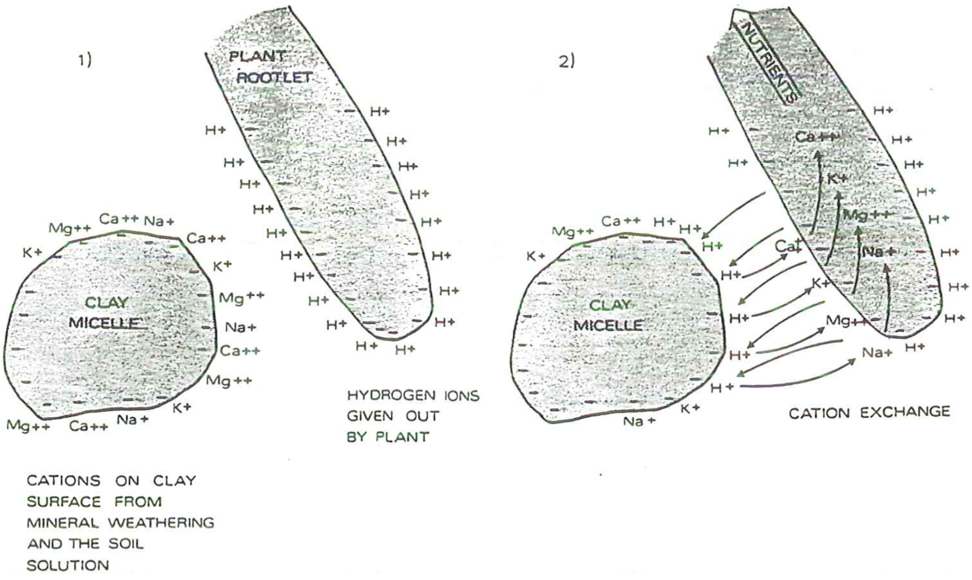


Fig. 56. Cation exchange.

The base saturation is the proportion of basic cations (K, Na, Ca, Mg) of the total cation exchange capacity CEC (in %), the acid cations being H^+ and Al^{3+} (the latter reacts acid through hydrolysis). Both the CEC and exchangeable cations are routinely determined in the soil laboratory.

Anionic nutrients such as NO_3^- and PO_4^{3-}, having a negative charge, cannot be adsorbed on negatively charged surfaces, These will be adsorbed on positively charged sites which may be present as "broken bonds" of the octahedral layer of clay minerals (See Equation 22) and on the surface of (hydr)oxides of Fe and Al (sesquioxides). In most soils (Ferralsols/Oxisols excepted) the negative charges will outnumber the positive charges. Therefore, particularly nitrates will be lost to the groundwater, phosphate (and sulphate) may often precipitate as Ca, Fe, of Al-phosphates.

8. SOIL PHYSICS

In this section a number of properties will be discussed that have a bearing on soil as a physical support for plant growth as well as on soil as an engineering substrate. These properties are: structure, porosity, bulk density, consistence, swelling and shrinking, and soil-water relationships (infiltration, permeability, water retention).

8.1 SOIL STRUCTURE

The primary soil particles may be aggregated to larger units, structural elements (peds), in various manners (see Fig. 57). The factors determining the type and stability of the structure are the soil forming factors in general (see Section 1.5) and more particularly the type and content of clay (texture), organic matter, and the oxides.

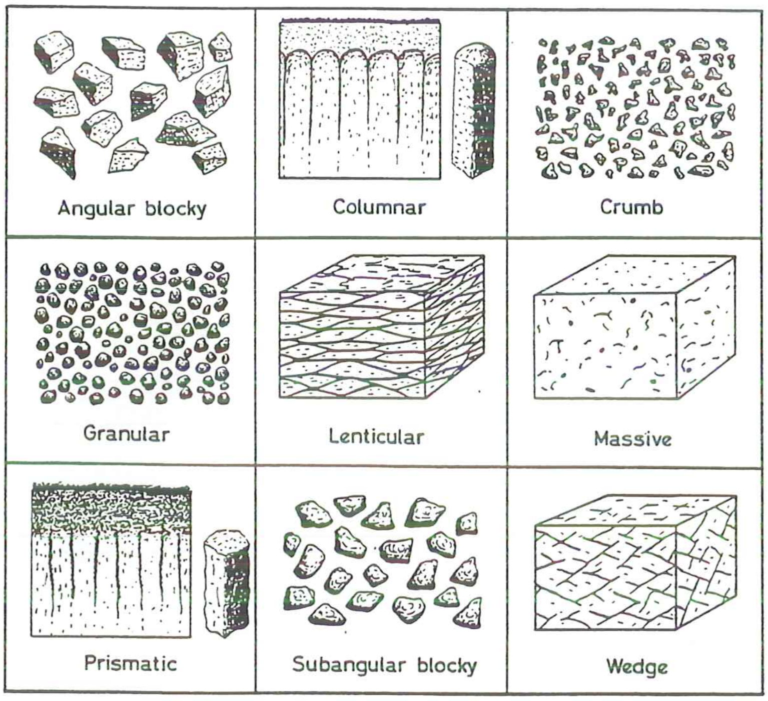


Fig. 57. Diagrammatic representation of some major types of soil structure. (After FitzPatrick)

The soil pore system includes discrete pores within the structural elements and continuous (conducting) pores between the peds. The size and continuity of the pores are important for water movement and retention. The large pores (> 10 μm) conduct water, the medium-sized pores (10 - 0.2 μm) hold water that is available to plants while the finer pores (< 0.2 μm) retain water so strongly that it is unavailable.

Clearly, not only the structural elements are important but also the total build-up of the soil which may consist of more than one type of structure, often differentiated in horizontal layers or horizons. Such stratification may be inherited from the parent material (sediments!) or may result from soil formation, e.g. the textural B horizon. Layering hinders the vertical movement of water (see Fig. 36) and consequently the root development of plants. The difference between a gradual and an abrupt transition is shown in Figure 58.

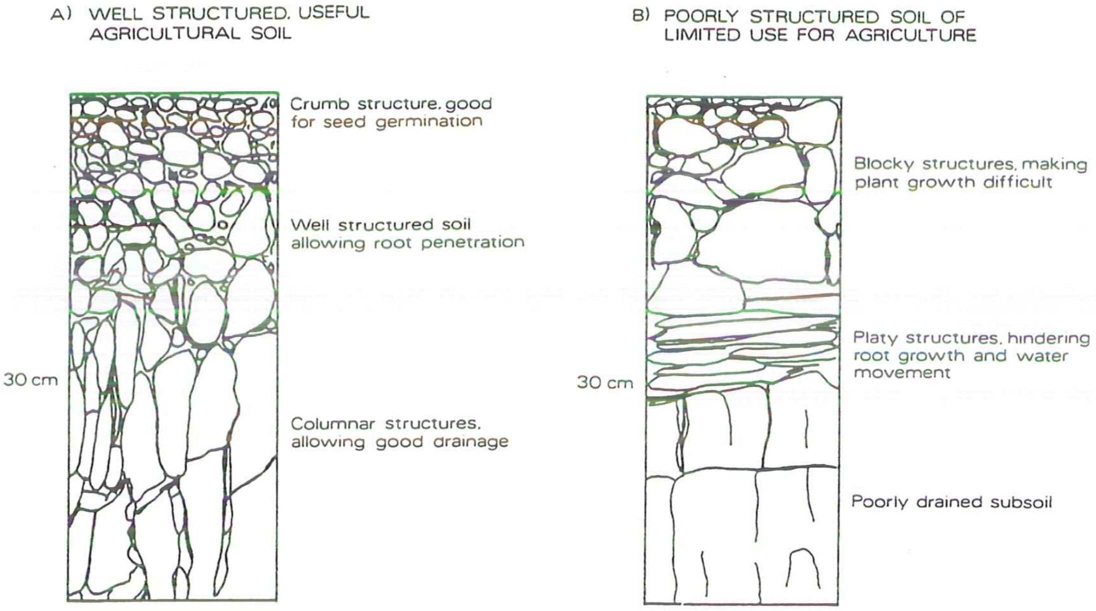


Fig. 58. Schematic drawing of gradual and abrupt transitions in soil structure.

8.2 POROSITY AND BULK DENSITY

Porosity or pore volume is an important soil property determining many other properties such as water and air holding and transport capacity, rootability, temperature behaviour (heat conductance), tillage behaviour. Naturally, the distribution of the pore size is of equal importance: generally, the more evenly distributed the better the agronomic properties.
The pore volume is defined as the *pore space ratio (PSR)* and is the volume of pores divided by the total soil volume. The easiest way to measure this is to saturate an undisturbed core sample (of known volume) with water and determine the water content. (Beware that soil volume may change with the water content due to possible swell and shrink). An other measure for the pore volume (not for the distribution) is the *bulk density (BD)*. This is also determined on a core sample of known volume. The bulk density is defined as the (oven-dry) weight in kg of a litre of soil (kg/L).

8.3 SOIL CONSISTENCE

This is a term to describe the resistance of a soil at various moisture contents to mechanical stress and manipulations. It is determined by forces of mutual attraction among soil particles (influenced by water). It is commonly measured by feeling and manipulating the soil by hand or by pulling a tillage instrument through it. Table 9 summarizes the terms used to describe soil consistence.

8.4 AGGREGATE STABILITY

Important for the structure and pore space is the stability of the structure. One way to assess this is to assess the stability of the aggregates. When immersed in water, weakly structured aggregates

Table 9. Terms to describe soil consistence.

Conditions of least coherence are represented by terms at the top of each column. Greater coherence characterizes terms as you move down the column.

	Wet soils			
	Stickiness	*Plasticity*	*Moist soils*	*Dry soils*
Increasing coherence ↓	Nonsticky	Nonplastic	Loose	Loose
	Slightly sticky	Slightly plastic	Very friable	Soft
	Sticky	Plastic	Friable	Slightly hard
	Very sticky	Very plastic	Firm	Hard
			Very firm	Very hard
			Extremely firm	Extremely hard

may slake and disperse. Measurement may be done by wet-sieving: a sample is transferred to the top sieve of a nest of sieves with decreasing mesh size. The whole set is immersed in a water tank and mechanically moved up and down. After the test the amount of soil material on the sieves is dried and weighed.

8.5 SWELLING AND SHRINKING

Soils will generally swell upon wetting and shrink upon drying. Both for agricultural and engineering purposes this is an adverse property. The extent of swelling and shrinking depend on the type of clay mineral (smectites!), type of adsorbed cations and the organic matter content. The function of the electrical double layer in this respect was discussed in Section 1.6.5 (clay migration). A measure of swelling is the COLE value (coefficient of linear extensibility). Determination is done by measuring the volume increase of a clod upon water uptake.

8.6 WATER RETENTION

The amount of water retained and the rate of release by soil is a very important quality aspect. The influence of the pore size on the force with which water is held is shown in Figure 59. It shows that narrower pores retain water with greater force than wider pores (earlier shown in Fig. 36). This relation between the moisture tension and the size of the pores that can still hold water is given in Figure 60.

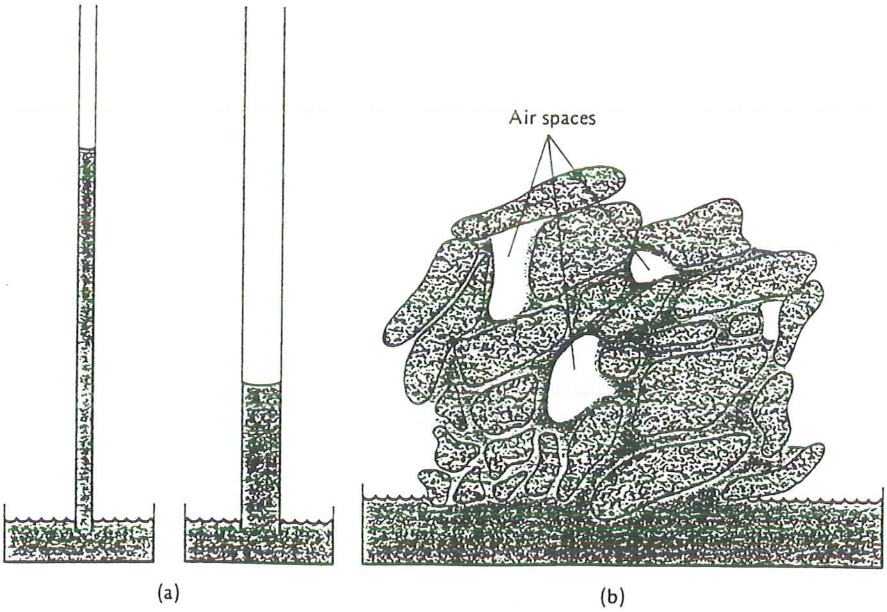


Fig. 59.
Upward movement by capillarity (a) in glass tubes of different sizes and (b) in soils. Although the mechanism is the same in the tubes and in the soil, adjustments are extremely irregular in soil because of the tortuous shape and variability in size of the soil pores and because of entrapped air.

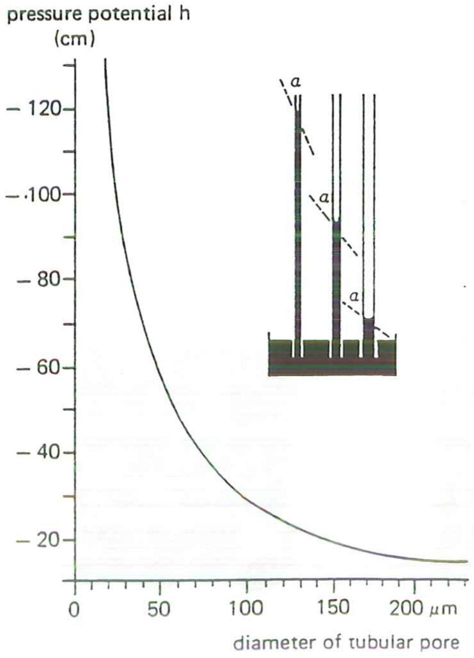


Fig. 60. The relationship between pore size and the corresponding water potential (tension).

An important quality aspect of soil is the *soil water retention characteristic* (old names: pF curve, water release curve). This gives the relation between the amount of water held by soil and the soil moisture tension or the *water potential* (the water potential is the sum of several components: matric potential, osmotic potential, gravitational potential, etc.). Figure 61 gives an example of such curves.

Note: pF is an obsolete but still much used convenient term. It is defined as the logarithm of the moisture suction or as the log of the negative moisture tension, both in cm head of water (compare pH!). For instance pF 2 corresponds with a *suction* of 100 cm (hanging) water column or a *tension* of -100 cm head of water (100 cm water ≈ 0.1 bar or 10 kPa). In the literature all these terms and units can be encountered which may lead to confusion.

Parameters related to the moisture retention characteristic are (see Fig. 61 and 62):

1. *Water content at saturation:* all pores are filled with water (pF ≈ 0)
2. *Field capacity:* the water content after saturation and free drainage (pF ≈ 2.2)
3. *Wilting point:* the water content where plants wilt (pF ≈ 4.2)
4. *Available water:* amount of water held between field capacity and wilting point

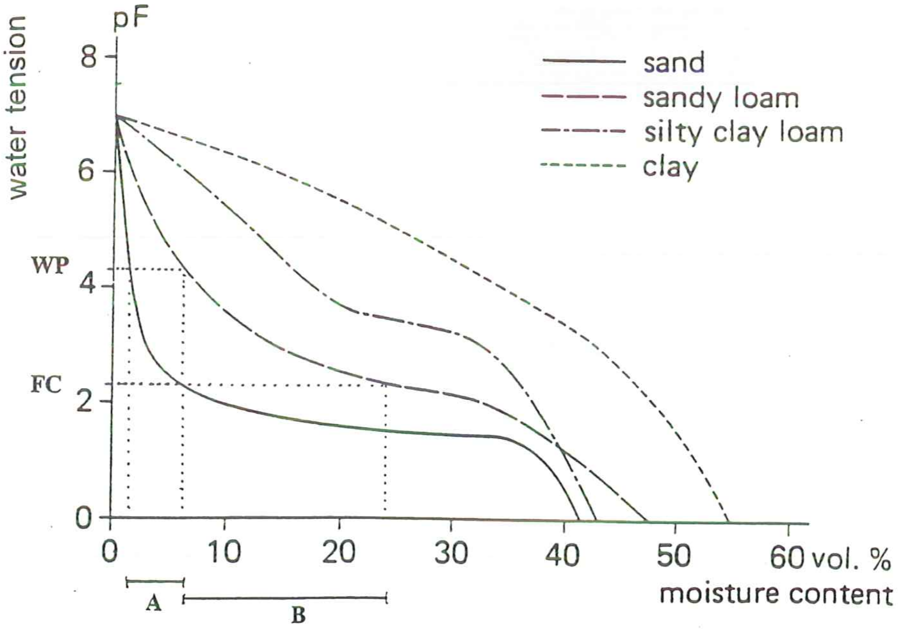


Fig. 61. Soil water retention characteristics (pF-curves) of four soil materials with different textures. Note the large difference in *available moisture* between sand (A) and the sandy loam (B). Already at field capacity the sand has lost most of its water. (FC = Field Capacity; WP = Wilting Point).

The principle of the determination of water retention is to measure the moisture content after equilibrating a soil sample at various moisture tension values. The tension can either be applied by *suction* for the lower tension range (up to 50 kPa or pF 2.7 on a silt/kaolin bath) or by *pressure* for the high tension range (in a pressure plate extractor, over 100 kPa or pF 3).

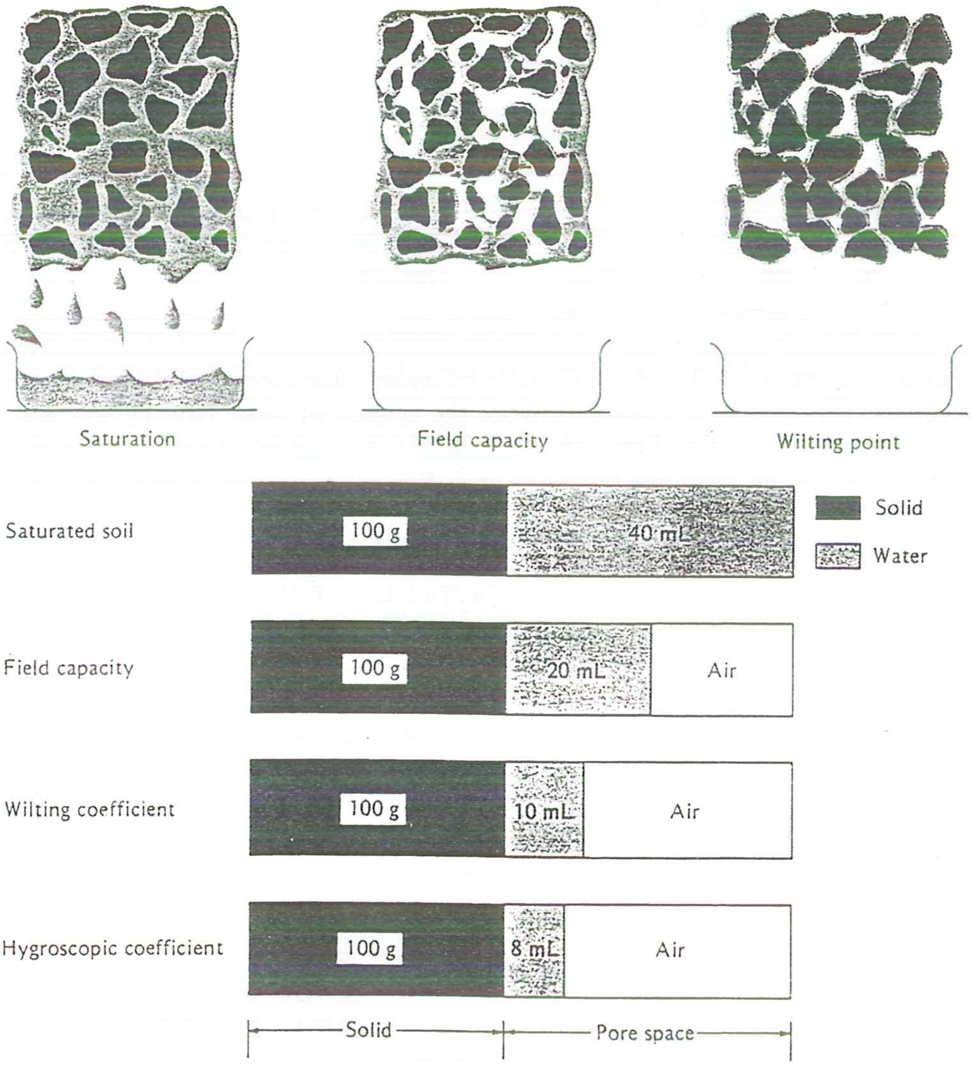


Fig. 62.

Volumes of water and air associated with 100 g of a well-granulated silt loam at different moisture levels. The top bar shows the situation when a representative soil is completely saturated with moisture. This situation will usually occur for short periods of time during a rain or when the soil is being irrigated. Water will soon drain out of the larger pores (*macropores*). The soil is then said to be at the *field capacity*. Plants will remove moisture from the soil quite rapidly until they begin to wilt. When permanent wilting of the plants occurs, the soil moisture is said to be at the *wilting coefficient*. There is still considerable moisture in the soil, but it is held too tightly to permit its absorption by plant roots. A further reduction in moisture content to the *hygroscopic coefficient* is illustrated in the bottom bar. At this point the water is held very tightly, mostly by the soil colloids. [Top drawings modified from *Irrigation on Western Farms* published by the U.S. Departments of Agriculture and Interior]

8.7 HYDRAULIC CONDUCTIVITY

This concerns the *dynamic* aspect of water in soils (water retention being *static*). The hydraulic conductivity (K_h, expressed in cm/day) determines such properties as permeability and infiltration rate. The transport of water through the soil depends on the size of the pores, the continuity of them (in a closed pore no transport will take place) and on the water potential. The latter determines the proportion of the pores that are filled with water and thus can take part in the transport. (Note that at higher water potentials the larger pores are empty). Thus, a distinction must be made between saturated and unsaturated hydraulic conductivity. The relationship between moisture content and hydraulic conductivity is shown in Figure 63.

For the determination of K_h various methods have been developed. The main problems with the laboratory methods is the handling of the samples: they need to remain undisturbed and preferable as large as possible to be representative. Therefore, when possible, field methods are preferred.

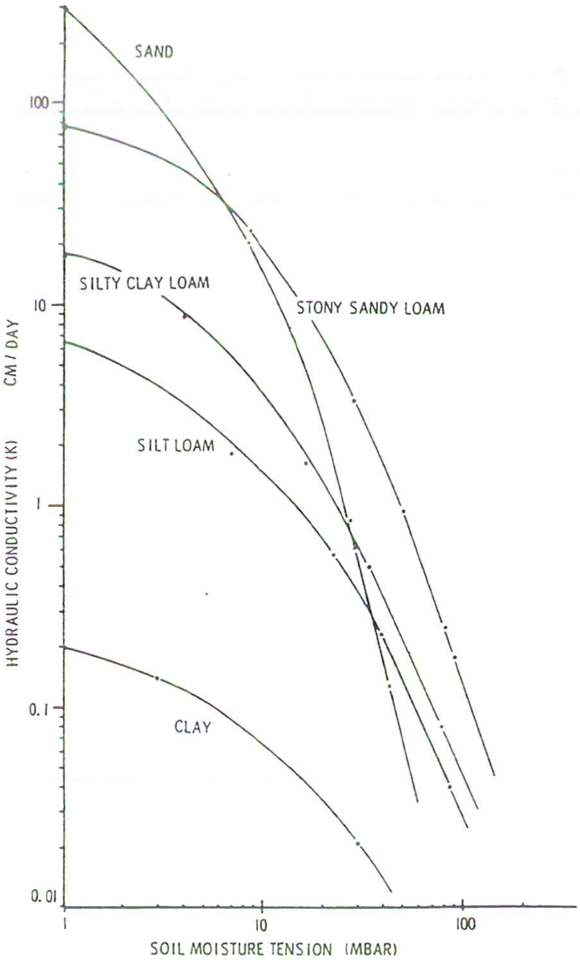


Fig. 63. Relationship between hydraulic conductivity and soil moisture tension for soils with various different textures.

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Part II

Soil Classification and problem soils in the Tropics, Subtropics and Arid Regions

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CONTENTS

II.1 SOIL CLASSIFICATION

Soil classification is probably as old as farming. The fact that around 8,000 BP the first farming communities in Europe settled on the better loess soils indicates that during these times farmers were already capable of distinguishing between the more and less productive soils. The oldest historical record of soil classification is most likely the Chinese book "Yugong" in which the soils of China were classified into three categories and nine classes, based on soil colour, texture and hydrological features (Gong, 1994). Even today such criteria are still in use by farmers to differentiate soils. Studies on indigenous soil knowledge in northern Ghana, for example, have shown that farmers use texture, colour, stoniness and soil depth to stratify the soils (Asiamah et al., 1997).

This chapter will describe subsequently the Legend of the Soil Map of the World (FAO-UNESCO, 1974), its revised version (FAO, 1988), and the World Reference Base for Soil Resources (ISSS-ISRIC-FAO, 1998) as the international system of soil classification adopted by the International Union of Soil Sciences (IUSS).

II.1.1 The FAO-UNESCO Legend of the Soil Map of the World 1 : 5 000 000

II.1.1.1 Introduction

The Soil Map of the World is a response to recommendations made by the international soils community in the 1950s to give special attention to developing the classification and correlation of the soils of great regions of the world (FAO-UNESCO, 1974). It was the first attempt to prepare, on the basis of international co-operation, a soil map covering all continents of the world using a uniform legend, thus enabling the correlation of soil units and the comparison of soils on a global scale. It should be borne in mind that the Soil Map of the World project produced a legend accompanying the map and not a global soil classification system.

II.1.1.2 History

The Food and Agriculture Organization (FAO) and United Nations Educational, Scientific, and Cultural Organization (UNESCO), in association with the International Society of Soil Science (ISSS), jointly took up the recommendations made during the society's sixth and seventh congresses in 1956 and 1960, to prepare a soil map of the world at scale 1 : 5 000 000. The project started in 1961 and was based on the compilation of available soil survey material and field correlation. A scientific advisory panel was convened to study the scientific and methodological problems relative to the preparation of such a soil map of the world.

During a number of meetings the advisory panel worked out the organization of the field correlation, selected the scale of the map and its topographic base, and prepared the first draft definitions of soil units. These were presented to the eighth ISSS congress in 1964. In 1966, a general agreement was reached on the principles for constructing the international legend, on the preparation of the definitions of soil units, and on the adoption of a unified nomenclature. The first draft was presented in 1968 to the ninth ISSS congress, which approved the outline of the legend, the definitions and the nomenclature. Moreover, it recommended that the Soil Map of the World be published as soon as possible.

II.1.1.3 Objectives

The objectives of the Soil Map of the World project (FAO-UNESCO, 1974) were to:

(1) make a first appraisal of the world's soil resources,
(2) supply a scientific basis for the transfer of experience between areas with similar environments,
(3) promote the establishment of a generally accepted soil classification and nomenclature,
(4) establish a common framework for more detailed investigations in developing areas,
(5) serve as a basic document for educational, research, and development activities, and
(6) strengthen international contacts in the field of soil science.

II.1.1.4 The Soil Units

The soil units, which form the basis of the FAO-UNESCO Legend, have been defined in terms of measurable and observable properties of the soil itself. They form a monocategorical and not a taxonomic system, with different levels of generalization. However, for the sake of logical presentation, they can be grouped on generally accepted principles of soil formation (FAO-UNESCO, 1974).

Based on soil development status, material, and major geographical zone, 24 major soils and 106 soil units are distinguished (Table II.1).

The presence or absence of diagnostic horizons and properties characterises the soil units. Key properties have been selected on the basis of generally accepted principles of soil formation so as to correlate with as many other characteristics as possible. Clusters of properties have been combined into so called "diagnostic horizons" which have been adopted to formulate the definitions of the soil units. The definitions and nomenclature of the diagnostic horizons and properties used are drawn from those adopted in Soil Taxonomy (Soil Survey Staff, 1975), but the definitions have been summarized and sometimes simplified to serve the purpose of the legend. For full background and details, the user is referred to Soil Taxonomy (Soil Survey Staff, 1975). A brief overview of the diagnostic horizons and properties used in the FAO-UNESCO Legend and their meaning is given in Table II.2.

Volume I of the Legend of the Soil Map of the World provides a key to the soil units which can be used to identify the soil (FAO-UNESCO, 1974). An abbreviated version listing the major soil units is reproduced in Table II.3.

II.1.2 The Revised Legend of the FAO-UNESCO Soil Map of the World

II.1.2.1 Introduction

A revised version of the Legend of the Soil Map of the World was issued (FAO, 1988) which assessed to what extent the objectives of the original Legend of the Soil Map of the World (FAO-UNESCO, 1974) were met, and analyzed its present day function. It was realized that, in order to keep the maps and accompanying legend up-to-date, revisions were necessary. New knowledge on soils, particularly from the developing world, had emerged and more recent soil surveys had yielded better insight into the distribution of soils in the world. The Revised Legend does not replace the 1974 Legend, which continues to serve as reference for the Soil Map of the World. The Revised Legend is to be used for updating the Soil Map of the World, contained in the UNEP-ISSS-ISRIC-FAO sponsored programme of SOTER, and serves as framework for the establishment by the ISSS of the World Reference Base for Soil Resources.

II.1.2.2 Amendments to the 1974 Legend of the Soil Map of the World

The monocategorial character of the 1974 Legend, using only soil units, was transformed into a multicategorial system with Major Soil Groupings (MSG)(e.g. Fluvisols), soil units (e.g. Dystric Fluvisols), and soil subunits (e.g. Gleyi-dystric Fluvisols) (FAO, 1988). The soil subunit level is described in more detail by Nachtergaele et al. (1994), providing guidelines for distinguishing soil subunits.

The introduction of a multicategorical system was necessitated by the increasing use of the 1974 Legend in more detailed surveys, particularly in Africa, where during the 1980s 1 : 1 000 000 soil maps were produced of, for example, Kenya, Botswana and Zambia.

The Revised Legend of the Soil Map of the World distinguishes 28 MSGs, four more than the 1974 Legend, and 153 soil units, 47 more than in 1974. Major changes are:

(1) amalgamation of the Lithosols, Rendzinas and Rankers into one major soil grouping (Leptosols), since the three had been difficult to show on the map,
(2) deletion of the Xerosols and Yermosols, which were characterized by an aridic moisture regime (as a general principle, climatic criteria were not to be used in separating the soil units.),
(3) division of the Acrisols and Luvisols, in 1974 distinguished by the base saturation, into four MSGs, introducing the clay activity as an additional separating criterion,
(4) introduction of new MSGs, (Alisols, Calcisols, Gypsisols, Lixisols, Plinthosols and Anthrosols), and
(5) renaming of Nitosols (Nitisols).

A listing of the MSGs and the soil units is given in Table II.4.

The diagnostic horizons and properties have also been largely adapted. In 1974, they were fairly similar to those in Soil Taxonomy (Soil Survey Staff, 1975) but in 1988, many were redefined and renamed, and new additions made. The argillic and oxic B horizons were renamed argic and ferralic B horizons, respectively. The argic B horizon now includes both the argillic and the kandic horizon as defined in Soil Taxonomy (Soil Survey Staff, 1999), while the ferralic B horizon includes additional criteria such as silt/clay ratio and water dispersible clay content. An addition is the thick, manmade, fimic A horizon which includes both the anthropic and plaggen epipedons of Soil Taxonomy.

Newly defined diagnostic properties include continuous hard rock, fluvic, geric, nitic, and sodic properties. The 1974 hydromorphic properties were split into gleyic (wetness conditioned by groundwater) and stagnic (wetness conditioned by surface water) properties. Definitions of albic material, aridic moisture regime, high organic matter content and thin iron pan were deleted, as they are no longer used in defining soil units. Andic properties have replaced the 1974 exchange complex dominated by amorphous material.

A brief overview of the diagnostic horizons and properties used in the Revised Legend and their meaning is given in Table II.5. A key to the Major Soil Groupings in the Revised Legend is given in Table II.6.

II.1.3 The World Reference Base for Soil Resources

II.1.3.1 Introduction

The World Reference Base for Soil Resources (WRB) was initiated by the ISSS, FAO, and ISRIC, to "..... provide scientific depth and background to the 1988 FAO-UNESCO-ISRIC Revised Legend of the Soil Map of the World, so that it incorporates the latest knowledge relating to global soil resources and interrelationships." (ISSS-ISRIC-FAO, 1998).

II.1.3.2 History

The history of the initiative is closely related to the Soil Map of the World 1 : 5 000 000 project of FAO and UNESCO. After its completion in the early 1980s it was realized that 20 years had elapsed since the project had started. During this period, numerous new soil surveys, often using local or national soil classification systems, had been carried out both in developing and developed countries, generating new knowledge and insight on the distribution and potential of our soil resources. If the Soil Map of the World were to retain its value as a global soil resource inventory, it needed regular updating with the most recent information.

Preliminary discussions were started on the establishment of an International Reference Base for Soil Classification (IRB), an initiative undertaken by FAO and UNESCO, supported by the United Nations Environmental Program (UNEP) and the ISSS. The intention of the IRB project was to work toward the establishment of a framework through which existing soil classification systems could be correlated and ongoing soil classification work could be harmonized (Dudal, 1990). Meetings were organized in Sofia, Bulgaria, to commence such an international program. The outcomes were draft definitions of 16 major soil groups occurring globally.

During the ISSS congresses in 1982 and 1986 and expert consultations in 1987 and 1988, the IRB took form and as a result 20 major soil groupings were identified and agreed upon as being representative of the principal components of the world's soil cover.

Subsequently, it became clear that some of the proposed 20 major soil groupings were too broad to be defined consistently and, consequently, had to be subdivided. By doing so the list of major soil groupings became very close to those of the Revised Legend of the Soil Map of the World (FAO, 1988). As a result it was decided in 1992 to adopt the Revised Legend as the frame for further IRB work. This was also prompted by the fact that both the Revised Legend and the International Reference Base for Soil Classification were supported by the ISSS and that it would be inappropriate to pursue two programs which essentially had the same goal, namely to arrive at a rational inventory of global soil resources (ISSS-ISRIC-FAO, 1994). The two programs were therefore merged under the name "World Reference Base for Soil Resources (WRB)", an ISSS/FAO/ISRIC undertaking.

II.1.3.3 Objectives

The specific objectives of the WRB are to:

1. develop an internationally acceptable framework for delineating soil resources to which national classifications can be attached and related, using the FAO Revised Legend as a guideline,
2. provide this framework with a sound scientific base so that it can also serve different applications in related fields such as agriculture, geology, hydrology and ecology,
3. acknowledge in the framework important lateral aspects of soils and soil horizon distribution as characterized by topo- and chronosequences, and
4. emphasize the morphological characterization of soils rather than to follow a purely analytical approach.

II.1.3.4 Concepts and Principles

For describing and defining the reference soil groups of the WRB, use is made of characteristics, properties and horizons, which combined define soils and their relationships. Soil characteristics are single parameters, which are observable or measurable in the field or laboratory, or can be analyzed using microscope techniques. They include such characteristics as colour, texture and structure of the soil, features of biological activity, arrangement of voids and pedogenic concentrations (mottles, cutans, nodules, etc.) as well as analytical determinations (soil reaction, particle size distribution, CEC, exchangeable cations, amount and nature of soluble salts, etc.).

Soil properties are combinations (assemblages) of soil characteristics which are known to occur in soils and indicative of present or past soil forming processes (e.g. vertic properties, which are a combination of heavy texture, smectitic mineralogy, slickensides, hard consistence when dry, sticky when wet, shrinking when dry and swelling when wet).

Soil horizons are three-dimensional pedological bodies, which are more or less parallel to the earth's surface. Each horizon exhibits one or more property, occurring over a certain depth, which characterizes it and permits its recognition. The thickness varies from a few cm to several meters. The upper and lower limits (boundaries) are more or less clear, and progressive, or abrupt. Laterally, the extension of a soil horizon varies greatly, from a meter to several km. However, a soil horizon is never infinite. Laterally, it disappears or grades into another horizon.

Soils are defined by the vertical combination of horizons, occurring within a defined depth, and by the lateral organization (sequence) or lack thereof of the soil horizons at a scale reflecting the relief or a land unit. Soil horizons and properties are intended to reflect the expression of genetic processes, which are widely recognized as occurring in soils. They are considered to be diagnostic when they reach a minimum degree of expression, which is determined by visibility, prominence, measurability, importance and relevance for soil formation and soil use. To be diagnostic, soil horizons also require a minimum thickness, which must be appraised in relation to bioclimatic factors (e.g. a spodic horizon in boreal regions can be expected to be thinner than in the tropics).

A brief description of the diagnostic horizons, properties and materials recognized in the WRB is presented in Table II.7.

The general principles governing the construction of the WRB can be summarized as follows:

(1) classification of soils is based on soil properties defined in terms of diagnostic horizons, properties and materials, which to the greatest extent possible should be measurable and observable in the field;
(2) selection of diagnostic horizons, properties and materials takes into account their relationship with soil forming processes; however, at a high level of generalization, it also attempts to select, to the extent possible, diagnostic features which are of significance for management purposes;
(3) climatic parameters are not applied;
(4) the "Reference Base" is limited to the highest level only and comprises 30 reference soil groups; listing of adjectives to the reference soil group names, each with a unique meaning, is provided to enable precise characterization and classification of individual soil bodies;
(5) the reference soil groups identified are representative of major soil regions so as to provide a comprehensive overview of the world's soil cover;
(6) the "Reference Base" is not meant to substitute for national classification systems but rather to serve as a common denominator for communication at international level and as a link between existing soil classification systems; in addition, WRB may also serve as a consistent communication tool for compiling global soil databases and for inventory and monitoring of the world's soil resources;
(7) the FAO-UNESCO-ISRIC Revised Legend of the Soil Map of the World is used as base to develop the WRB to take advantage of international soil correlation work which has already been conducted in the course of the project;
(8) where possible, definitions and descriptions of soils reflect both vertical and lateral variations in soil characteristics to account for spatial linkages in the landscape; and
(9) the nomenclature used to distinguish reference soil groups and soil units will have terms which are traditionally used or which can easily be introduced in current language.

II.1.3.5 The WRB Reference Soil Groups

Thirty reference soil groups are now identified (ISSS-ISRIC-FAO, 1998). Compared to FAO Revised Legend, one major soil grouping has been omitted (Greyzems), and three new ones are introduced (Cryosols, Durisols and Umbrisols). Greyzems were deleted, as they constitute the smallest major soil grouping and were amalgamated with the Phaeozems. Cryosols were newly introduced to identify a group of soils, which occur under the unique environmental conditions of thawing and freezing. Durisols have been added to group soils together, which have accumulation of secondary silica, analogous to the Calcisols and Gypsisols. Umbrisols constitute the group of soils, which have a thick accumulation of desaturated organic matter at the surface, and are the natural counterpart of Chernozems, Kastanozems and Phaeozems. The key to the reference soil groups of the WRB is presented in Table II.8, with a simplified key in Table II.9.

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Table II.1 Listing of the 106 soil units of the FAO-UNESCO Legend of the Soil Map of the World 1 : 5 000 000

J	**FLUVISOLS**
Je	Eutric Fluvisols
Jc	Calcaric Fluvisols
Jd	Dystric Fluvisols
Jt	Thionic Fluvisols

G	**GLEYSOLS**
Ge	Eutric Gleysols
Gc	Calcaric Gleysols
Gd	Dystric Gleysols
Gm	Mollic Gleysols
Gh	Humic Gleysols
Gp	Plinthic Gleysols
Gx	Gelic Gleysols

R	**REGOSOLS**
Re	Eutric Regosols
Rc	Calcaric Regosols
Rd	Dystric Regosols
Rx	Gelic Regosols

I	**LITHOSOLS**

Q	**ARENOSOLS**
Qc	Cambic Arenosols
Ql	Luvic Arenosols
Qf	Ferralic Arenosols
Qa	Albic Arenosols

E	**RENDZINAS**

U	**RANKERS**

T	**ANDOSOLS**
To	Ochric Andosols
Tm	Mollic Andosols
Th	Humic Andosols
Tv	Vitric Andosols

V	**VERTISOLS**
Vp	Pellic Vertisols
Vc	Chromic Vertisols

Z	**SOLONCHAKS**
Zo	Orthic Solonchaks
Zm	Mollic Solonchaks
Zt	Takyric Solonchaks
Zg	Gleyic Solonchaks

S	**SOLONETZ**
So	Orthic Solonetz
Sm	Mollic Solonetz
Sg	Gleyic Solonetz

Y	**YERMOSOLS**
Yh	Haplic Yermosols
Yk	Calcic Yermosols
Yy	Gypsic Yermosols
Yl	Luvic Yermosols
Yt	Takyric Yermosols

X	**XEROSOLS**
Xh	Haplic Xerosols
Xk	Calcic Xerosols
Xy	Gypsic Xerosols
Xl	Luvic Xerosols

K	**KASTANOZEMS**
Kh	Haplic Kastanozems
Kk	Calcic Kastanozems
Kl	Luvic Kastanozems

C	**CHERNOZEMS**
Ch	Haplic Chernozems
Ck	Calcic Chernozems
Cl	Luvic Chernozems
Cg	Glossic Chernozems

H	**PHAEOZEMS**
Hh	Haplic Phaeozems
Hc	Calcaric Phaeozems
Hl	Luvic Phaeozems
Hg	Gleyic Phaeozems

M	**GREYZEMS**
Mo	Orthic Greyzems
Mg	Gleyic Greyzems

B	**CAMBISOLS**
Be	Eutric Cambisols
Bd	Dystric Cambisols
Bh	Humic Cambisols
Bg	Gleyic Cambisols
Bx	Gelic Cambisols
Bk	Calcic Cambisols
Bc	Chromic Cambisols
Bv	Vertic Cambisols
Bf	Ferralic Cambisols

L	**LUVISOLS**
Lo	Orthic Luvisols
Lc	Chromic Luvisols
Lk	Calcic Luvisols
Lv	Vertic Luvisols
Lf	Ferric Luvisols
La	Albic Luvisols
Lp	Plinthic Luvisols
Lg	Gleyic Luvisols

D	**PODZOLUVISOLS**
De	Eutric Podzoluvisols
Dd	Dystric Podzoluvisols
Dg	Gleyic Podzoluvisols

P	**PODZOLS**
Po	Orthic Podzols
Pl	Leptic Podzols
Pf	Ferric Podzols
Ph	Humic Podzols
Pp	Placic Podzols
Pg	Gleyic Podzols

W	**PLANOSOLS**
We	Eutric Planosols
Wd	Dystric Planosols
Wm	Mollic Planosols
Wh	Humic Planosols
Ws	Solodic Planosols
Wx	Gelic Planosols

A	**ACRISOLS**
Ao	Orthic Acrisols
Af	Ferric Acrisols
Ah	Humic Acrisols
Ap	Plinthic Acrisols
Ag	Gleyic Acrisols

N	**NITOSOLS**
Ne	Eutric Nitosols
Nd	Dystric Nitosols
Nh	Humic Nitosols

F	**FERRALSOLS**
Fo	Orthic Ferralsols
Fx	Xanthic Ferralsols
Fr	Rhodic Ferralsols
Fh	Humic Ferralsols
Fa	Acric Ferralsols
Fp	Plinthic Ferralsols

O	**HISTOSOLS**
Oe	Eutric Histosols
Od	Dystric Histosols
Ox	Gelic Histosols

Table II.2 Diagnostic horizons, properties and materials in the 1974 FAO-UNESCO Legend of the Soil Map of the World 1 : 5 000 000

Horizons

Albic E	light coloured eluvial horizon generally associated with argic and spodic horizons.
Argillic B	subsurface horizon with distinct clay accumulation.
Calcic	horizon with accumulation of calcium carbonate.
Cambic B	subsurface horizon showing evidence of alteration relative to the underlying horizon(s).
Gypsic	horizon with accumulation of gypsum.
Histic H	poorly aerated, waterlogged, highly organic surface horizon.
Mollic A	thick, dark coloured surface horizon with high base saturation and moderate to high organic matter content.
Natric B	subsurface horizon with distinct clay accumulation and a high exchangeable sodium percentage.
Ochric A	weakly developed surface horizon, either light-coloured, or thin, or having a low organic matter content.
Oxic B	strongly weathered subsurface horizon with low cation exchange capacity.
Spodic B	dark coloured subsurface horizon with illuvial alumino-organic complexes, with or without iron.
Sulfuric	extremely acid subsurface horizon with sulphuric acid resulting from oxidation of sulphides.
Umbric A	thick, dark coloured surface horizon with low base saturation and moderate to high organic matter content.

Properties and materials

Abrupt textural change	sharp increase in clay content within a limited depth range.
Albic material	light coloured mineral soil material.
Aridic moisture regime	no available water in any part of the moisture control section for as long as 90 consecutive days, or more than half the time when soil temperature (at 50 cm depth) is above 5°C.
Exchange complex dominated by amorphous material	CEC (pH 8.2) more than 150 $\mathrm{cmol_c\ kg^{-1}}$ clay; if pF 15-bar is 20% or more, pH NaF is more than 9.4; ratio 15-bar water content to clay more than 1.0; more than 0.6% organic carbon; DTA shows low temperature endotherm; bulk density is 0.85 $\mathrm{g\ cm^{-3}}$ at 1/3 bar tension.
Ferralic	low cation exchange capacity (<24 $\mathrm{cmol_c\ kg^{-1}}$ clay by NH_4Cl).
Ferric	iron concentrated in large mottles or concretions, or low cation exchange capacity (<24 $\mathrm{cmol_c\ kg^{-1}}$ clay by NH_4Cl).
Gilgai microrelief	succession of enclosed microbasins and microknolls in level areas, or microvalleys and microridges on slopes.
High organic matter content	organic matter content of 1.35% or more averaged to a depth of 100 cm, or 1.5% organic matter in the upper part of the B horizon (Acrisols only).
High salinity	EC of saturation extract more than 15 $\mathrm{dS\ m^{-1}}$ at 25°C at specified depths, or more than 4 $\mathrm{dS\ m^{-1}}$ within 25 cm of the surface if pH (H_2O, 1:1) exceeds 8.5.
Hydromorphic properties	saturation with groundwater; occurrence of a histic H horizon; dominant neutral (N) hues or hues bluer than 10Y; saturation with water at some period of the year (unless artificially drained) with evidence of reduction or of reduction and segregation
Interfingering	penetrations of an albic E horizon into an underlying argillic or natric B horizon, not wide enough to constitute tonguing.
Permafrost	perennial temperature at or below 0°C.
Plinthite	iron-rich, humus-poor soil material, which hardens irreversibly upon repeated wetting and drying.
Slickensides	polished and grooved ped surfaces that are produced by one mass sliding past another.
Smeary consistence	presence of thixotropic soil material.
Soft powdery lime	accumulation of translocated calcium carbonate in soft powdery form.
Sulfidic materials	waterlogged deposit containing 0.75% or more sulphur, and less than three time as much carbonates ($CaCO_3$ equivalent) as sulphur.
Takyric features	combination of heavy texture, polygonal cracks and platy or massive surface crust.
Thin iron pan	black to dark reddish layer cemented by iron, by iron and manganese, or by iron-organic matter complexes.
Tonguing	penetrations of an albic E horizon into an argillic B horizon with specified dimensions.
Vertic properties	cracks 1 cm or more wide within 50 cm of the upper boundary of the B horizon, extending to the surface or at least to the upper part of the B horizon.
Weatherable minerals	presence of minerals considered to be unstable relative to other minerals such as quartz and 1:1 lattice clays, and which produce plant nutrients upon weathering.

Table II.3 Key to the Major Soil Units of the 1974 FAO-UNESCO Legend of the Soil Map of the World 1 : 5 000 000

Soils having an H horizon of 40 cm or more (60 cm or more if the organic material consists mainly of sphagnum or moss or has a bulk density of less than 0.1 g cm^{-3}) either extending down from the surface or taken cumulatively within the upper 80 cm of the soil; the thickness of the H horizon may be less when it rests on rock or on fragmental material of which the interstices are filled with organic matter.

HISTOSOLS (O)

Other soils which are limited in depth by continuous coherent and hard rock within 10 cm of the surface.

LITHOSOLS (I)

Other soils which, after the upper 20 cm have been mixed, have 30 percent or more clay in all horizons to at least 50 cm from the surface; at some period in most years have cracks at least 1 cm wide at a depth of 50 cm, unless irrigated, and have one or more of the following characteristics: gilgai microrelief, intersecting slickensides or wedgeshaped or parallelepiped structural aggregates at some depth between 25 and 100 cm from the surface.

VERTISOLS (V)

Other soils developed from recent alluvial deposits, having no diagnostic horizons other than (unless buried by 50 cm or more new material) an ochric or an umbric A horizon, a histic H horizon, or a sulfuric horizon.

FLUVISOLS (J)

Other soils having high salinity and having no diagnostic horizons (unless buried by 50 cm or more new material) an A horizon, an H horizon, a cambic B horizon, a calcic or a gypsic horizon.

SOLONCHAKS (Z)

Other soils showing hydromorphic properties within 50 cm of the surface; having no diagnostic horizons other than (unless buried by 50 cm or more new material) an A horizon, an H horizon, a cambic B horizon, a calcic or gypsic horizon.

GLEYSOLS (G)

Other soils having either a mollic or an umbric A horizon possibly overlying a cambic B horizon, or an ochric A horizon and a cambic B horizon; having no other diagnostic horizons (unless buried by 50 cm or more new material); having to a depth of 35 cm or more one or both of: (a) a bulk density (at 1/3 bar water retention) of the fine earth (less than 2 mm) fraction of the soil of less than 0.85 g/cm^3 and an exchange complex dominated by amorphous material; (b) 60 percent or more vitric volcanic ash, cinders, or other vitric pyroclastic material in the silt, sand and gravel fractions.

ANDOSOLS (T)

Other soils of coarse texture consisting of albic material occurring over a depth of at least 50 cm from the surface, or showing characteristics of argillic, cambic or oxic B horizons which, however, do not qualify as diagnostic horizons because of textural requirements; having no diagnostic horizons other than (unless buried by 50 cm or more new material) an ochric A horizon.

ARENOSOLS (Q)

Other soils having no diagnostic horizons or non other than (unless buried by 50 cm or more new material) an ochric A horizon.

REGOSOLS (R)

Other soils having an umbric A horizon, which is not more than 25 cm thick; having no other diagnostic horizons (unless buried by 50 cm or more new material).

RANKERS (U)

Other soils having a mollic A horizon which contains or immediately overlies calcareous material with a calcium carbonate equivalent of more than 40 percent (when the A horizon contains a high amount of finely divided calcium carbonate the colour requirements of the mollic A horizon may be waived).

RENDZINAS (E)

Other soils having a spodic B horizon.

PODZOLS (P)

Table II.3 Key to the Major Soil Units of the 1974 FAO-UNESCO Legend of the Soil Map of the World 1 : 5 000 000 *(cont'd)*

Other soils having an oxic B horizon.

FERRALSOLS (F)

Other soils having an albic E horizon overlying a slowly permeable horizon (for example, an argillic or natric B horizon showing an abrupt textural change, a heavy clay, a fragipan) within 125 cm of the surface; showing hydromorphic properties at least in a part of the E horizon.

PLANOSOLS (W)

Other soils having a natric B horizon.

SOLONETZ (S)

Other soils having a mollic A horizon with a moist chroma of 2 or less to a depth of at least 15 cm, showing bleached coatings on structural ped surfaces.

GREYZEMS (M)

Other soils having a mollic A horizon with a moist chroma of 2 or less to a depth of at least 15 cm; having one or more of the following: a calcic or a gypsic horizon, or concentrations of soft powdery lime within 125 cm of the surface when the weighted average textural class is coarse, within 90 cm for medium textures, within 75 cm for fine textures.

CHERNOZEMS (C)

Other soils having a mollic A horizon with a moist chroma of more than 2 to a depth of at least 15 cm; having one or more of the following: a calcic or a gypsic horizon, or concentrations of soft powdery lime within 125 cm of the surface when the weighted average textural class is coarse, within 90 cm for medium textures, within 75 cm for fine textures.

KASTANOZEMS (K)

Other soils having a mollic A horizon.

PHAEOZEMS (H)

Other soils having an argillic B horizon showing an irregular or broken upper boundary resulting from deep tonguing of the E into the B horizon or from the formation of discrete nodules (ranging from 2 to 5 cm up to 30 cm in diameter) the exteriors of which are enriched and weakly cemented or indurated with iron and having redder hues and stronger chroma than the interiors.

PODZOLUVISOLS (D)

Other soils having a weak ochric A horizon and an aridic moisture regime; lacking permafrost within 200 cm of the surface.

XEROSOLS (X)

Other soils having a very weak ochric A horizon and an aridi moisture regime; lacking permafrost within 200 cm of the surface.

YERMOSOLS (Y)

Other soils having an argillic B horizon with a clay distribution where the percentage of clay does not decrease from its maximum amount by as much as 20 percent within 150cm of the surface; lacking plinthite within 125 cm of the surface; lacking vertic and ferric properties.

NITOSOLS (N)

Other soils having an argillic B horizon; having a base saturation which is less than 50 percent (by NH_4OAc) in at least some part of the B horizon within 125 cm of the surface.

ACRISOLS (A)

Other soils having an argillic B horizon.

LUVISOLS (L)

Other soils having a cambic B horizon or an umbric A horizon which is more than 25 cm thick.

CAMBISOLS (B)

Table II.4 Major Soil Groupings and Soil Units of FAO's 1988 Revised Legend of the Soil Map of the World

FL	**FLUVISOLS**
FLe	Eutric Fluvisols
FLc	Calcaric Fluvisols
FLd	Dystric Fluvisols
FLm	Mollic Fluvisols
FLu	Umbric Fluvisols
FLt	Thionic Fluvisols
FLs	Salic Fluvisols
GL	**GLEYSOLS**
GLe	Eutric Gleysols
GLk	Calcic Gleysols
GLd	Dystric Gleysols
GLa	Andic Gleysols
GLm	Mollic Gleysols
GLu	Umbric Gleysols
GLt	Thionic Gleysols
GLi	Gelic Gleysols
RG	**REGOSOLS**
RGe	Eutric Regosols
RGc	Calcaric Regosols
RGy	Gypsiric Regosols
RGd	Dystric Regosols
RGu	Umbric Regosols
RGi	Gelic Regosols
LP	**LEPTOSOLS**
LPe	Eutric Leptosols
LPd	Dystric Leptosols
LPk	Rendzic Leptosols
LPm	Mollic Leptosols
LPu	Umbric Leptosols
LPq	Lithic Leptosols
LPi	Gelic Leptosols
AR	**ARENOSOLS**
ARh	Haplic Arenosols
ARb	Cambic Arenosols
ARl	Luvic Arenosols
ARo	Ferralic Arenosols
ARa	Albic Arenosols
ARc	Calcaric Arenosols
ARg	Gleyic Arenosols
AN	**ANDOSOLS**
ANh	Haplic Andosols
ANm	Mollic Andosols
ANu	Umbric Andosols
ANz	Vitric Andosols
ANg	Gleyic Andosols
ANi	Gelic Andosols
VR	**VERTISOLS**
VRe	Eutric Vertisols
VRd	Dystric Vertisols
VRk	Calcic Vertisols
VRy	Gypsic Vertisols
CM	**CAMBISOLS**
CMe	Eutric Cambisols
CMd	Dystric Cambisols
CMu	Humic Cambisols
CMc	Calcaric Cambisols
CMv	Vertic Cambisols
CMo	Ferralic Cambisols
CMg	Gleyic Cambisols
CMi	Gelic Cambisols
HS	**HISTOSOLS**
HSl	Folic Histosols
HSs	Terric Histosols
HSf	Fibric Histosols
HSt	Thionic Histosols
HSi	Gelic Histosols
AT	**ANTHROSOLS**
ATa	Aric Anthrosols
ATc	Cumulic Anthrosols
ATf	Fimic Anthrosols
ATu	Urbic Anthrosols
CL	**CALCISOLS**
CLh	Haplic Calcisols
CLc	Luvic Calcisols
CLp	Petric Calcisols
GY	**GYPSISOLS**
GYh	Haplic Gypsisols
GYk	Calcic Gypsisols
GYl	Luvic Gypsisols
GYp	Petric Gyopsisols
SN	**SOLONETZ**
SNh	Haplic Solonetz
SNm	Mollic Solonetz
SNk	Calcic Solonetz
SNy	Gypsic Solonetz
SNj	Stagnic Solonetz
SNg	Gleyic Solonetz
SC	**SOLONCHAKS**
SCh	Haplic Solonchaks
SCm	Mollic Solonchak
SCk	Calcic Solonchaks
SCy	Gypsic Solonchaks
SCn	Sodic Solonchaks
SCg	Gleyic Solonchaks
SCi	Gelic Solonchaks
KS	**KASTANOZEMS**
KSh	Haplic Kastanozems
KSl	Luvic Kastanozems
KSk	Calcic Kastanozems
KSy	Gypsic Kastanozems
CH	**CHERNOZEMS**
CHh	Haplic Chernozems
CHk	Calcic Chernozems
CHl	Luvic Chernozems
CHw	Glossic Chernozems
CHg	Gleyic Chernozems
PH	**PHAEOZEMS**
PHh	Haplic Phaeozems
PHc	Calcaric Phaeozems
PHl	Luvic Phaeozems
PHj	Stagnic Phaeozems
PHg	Gleyic Phaeozems
GR	**GREYZEMS**
GRh	Haplic Greyzems
GRg	Gleyic Greyzems
LV	**LUVISOLS**
LVh	Haplic Luvisols
LVf	Ferric Luvisols
LVx	Chromic Luvisols
LVk	Calcic Luvisols
LVv	Vertic Luvisols
LVa	Albic Luvisols
LVj	Stagnic Luvisols
LVg	Gleyic Luvisols
PL	**PLANOSOLS**
PLe	Eutric Planosols
PLd	Dystric Planosols
PLm	Mollic Planosols
PLu	Umbric Planosols
PLi	Gelic Planosols
PD	**PODZOLUVISOLS**
PDe	Eutric Podzoluvisols
PDd	Dystric Podzoluvisols
PDj	Stagnic Podzoluvisols
PDg	Gleyic Podzoluvisols
PDi	Gelic Podzoluvisols
PZ	**PODZOLS**
PZh	Haplic Podzols
PZb	Cambic Podzols
PZf	Ferric Podzols
PZc	Carbic Podzols
PZg	Gleyic Podzols
PZi	Gelic Podzols
LX	**LIXISOLS**
LXh	Haplic Lixisols
LXf	Ferric Lixisols
LXp	Plinthic Lixisols
LXa	Albic Lixisols
LXj	Stagnic Lixisols
LXg	Gleyic Lixisols
AC	**ACRISOLS**
ACh	Haplic Acrisols
ACf	Ferric Acrisols
ACu	Humic Acrisols
ACp	Plinthic Acrisols
ACg	Gleyic Acrisols
AL	**ALISOLS**
ALh	Haplic Alisols
ALf	Ferric Alisols
ALu	Humic Alisols
ALp	Plinthic Alisols
ALj	Stagnic Alisols
ALg	Gleyic Alisols
NT	**NITISOLS**
NTh	Haplic Nitisols
NTr	Rhodic Nitisols
NTu	Humic Nitisols
FR	**FERRALSOLS**
FRh	Haplic Ferralsols
FRx	Xanthic Ferralsols
FRr	Rhodic Ferralsols
FRu	Humic Ferralsols
FRg	Geric Ferralsols
FRp	Plinthic Ferralsols

Table II.5 Diagnostic horizons and properties of FAO's 1988 Revised Legend of the Soil Map of the World

Horizons	
Albic E	light coloured eluvial horizon generally associated with argic and spodic horizons.
Argic B	subsurface horizon with distinct clay accumulation.
Calcic	horizon with accumulation of calcium carbonate.
Cambic B	subsurface horizon showing evidence of alteration relative to the underlying horizon(s).
Ferralic B	strongly weathered subsurface horizon with low cation exchange capacity.
Fimic A	surface and subsurface horizons resulting from long continued cultivation.
Gypsic	horizon with accumulation of gypsum.
Histic H	poorly aerated, waterlogged, highly organic surface horizon.
Mollic A	thick, dark coloured surface horizon with high base saturation and moderate to high OM content.
Natric B	subsurface horizon with distinct clay accumulation and a high exchangeable sodium percentage.
Ochric A	weakly developed surface horizon, either light coloured, or thin, or having a low OM content.
Petrocalcic	continuous cemented or indurated calcic horizon.
Petrogypsic	continuous cemented or indurated gypsic horizon.
Spodic B	dark coloured subsurface horizon with illuvial alumino-organic complexes, with or without iron.
Sulfuric	extremely acid subsurface horizon with sulphuric acid resulting from oxidation of sulphides.
Umbric A	thick, dark coloured surface horizon with low base saturation and moderate to high OM content.

Properties	
Abrupt textural change	sharp increase in clay content within a limited depth range.
Andic	high acid oxalate extractable aluminium and iron content, low bulk density, high phosphate retention, high amount of coarse volcanoclastic material.
Calcareous	strong effervescence with 10 percent HCl (more than 2 perent calcium carbonate).
Calcaric	presence of calcareous soil material between 20 and 50 cm depth.
Continuous hard rock	presence of coherent rock, practically impermeable for roots.
Ferralic	low cation exchange capacity (less than 24 cmol kg^{-1} clay or 4 cmol$_c$ kg^{-1} fine earth).
Ferric	presence of many coarse mottles or large iron concretions.
Fluvic	presence of fresh fluviatile, lacustrine or marine sediments at the surface.
Geric	extremely low to negative effective cation exchange capacity.
Gleyic	wetness producing reduced conditions caused by groundwater.
Gypsiferous	presence of 5 percent or more gypsum.
Interfingering	narrow penetrations of an albic E horizon into an argic or natric B horizon.
Nitic	presence of strongly developed, nutshaped structure and shiny pedfaces.
Organic material	material containing a very high amount of organic debris.
Permafrost	perennial temperature at or below 0°C.
Plinthite	presence of iron-rich, humus-poor material that hardens irreversibly upon repeated wetting and drying.
Salic	high soluble salt content (EC$_e$ of more than 15 dS m^{-1}, or more than 4 if pH exceeds 8.5).
Slickensides	presence of polished and grooved surfaces produced by one mass sliding past another.
Smeary consistence	presence of thixotropic soil material.
Sodic	exchangeable sodium percentage of 15 percent or more, or exchangeable Na + Mg of 50 percent or more.
Soft powdery lime	accumulation of translocated calcium carbonate in soft powdery form.
Stagnic	wetness producing reduced conditions caused by stagnating surface water.
Strongly humic	high organic matter content.
Sulfidic material	waterlogged deposit containing sulphides, and only moderate amounts of calcium carbonate.
Tonguing	wide penetrations of an albic E horizon into an argic B horizon, or penetrations of a mollic A horizon into an underlying cambic B horizon or into a C horizon (Chernozems only).
Vertic	presence of cracks, slickensides, wedge-shaped or parallelepiped structural aggregates.
Weatherable minerals	presence of minerals unstable in a humid climate relative to other minerals.

Table II.6 Key to the Major Soil Groupings of FAO's 1988 Revised Legend of the Soil Map of the World

Soils having an H horizon, or an O horizon, of 40 cm or more (60 cm or more if the organic material consists mainly of sphagnum or moss or has a bulk density of less than 0.1 Mg m^{-3}) either extending down from the surface or taken cumulatively within the upper 80 cm of the soil; the thickness of the H or O horizon may be less when it rests on rocks or on fragmental material of which the interstices are filled with organic matter.

HISTOSOLS (HS)

Other soils in which human activities have resulted in a profound modification or burial of the original soil horizons, through removal or disturbance of surface horizons, cuts and fills, secular additions of organic materials, long-continued irrigation, etc.

ANTHROSOLS (AT)

Other soils, which are limited in depth by continuous hard rock or highly calcareous materials (calcium carbonate equivalent of more than 40 percent) or a continuous cemented layer within 30 cm of the surface or having less than 20 percent of fine earth over a depth of 75 cm from the surface. Diagnostic horizons may be present.

LEPTOSOLS (LP)

Other soils having, after the upper 18 cm have been mixed, 30 percent or more clay in all horizons to a depth of 50 cm; developing cracks from the soil surface downward which at some period in most years (unless the soil is irrigated) are at least 1 cm wide to a depth of 50 cm; having one or more of the following: intersecting slickensides or wedge-shaped or parallelepiped structural aggregates at some depth between 25 and 100 cm from the surface.

VERTISOLS (VR)

Other soils showing fluvic properties and having no diagnostic horizons other than an ochric, mollic, an umbric A horizon, or a histic H horizon, or a sulfuric horizon, or sulfidic material within 125 cm of the surface.

FLUVISOLS (FL)

Other soils showing salic properties and having no diagnostic horizons other than an ochric, umbric or mollic A horizon, a histic H horizon, a cambic B horizon, a calcic or a gypsic horizon.

SOLONCHAKS (SC)

Other soils, exclusive of coarse textured materials (except when a histic H horizon is present), showing gleyic properties within 50 cm of the surface; having no diagnostic horizons other than an A horizon, a histic H horizon, a cambic B horizon, a sulfuric horizon, a calcic or a gypsic horizon; lacking plinthite within 125 cm of the surface.

GLEYSOLS (GL)

Other soils showing andic properties to a depth of 35 cm or more from the surface and having a mollic or an umbric A horizon possibly overlying a cambic B horizon, or an ochric A horizon and a cambic B horizon; having no other diagnostic horizons.

ANDOSOLS (AN)

Other soils which are coarser than sandy loam to a depth of at least 100 cm from the surface, having less than 35 percent of rock fragments or other coarse fragments in all subhorizons within 100 cm of the surface, having no diagnostic horizons other than an ochric A horizon or an albic E horizon.

ARENOSOLS (AR)

Other soils having no diagnostic horizons other than an ochric or umbric A horizon; lacking soft powdery lime.

REGOSOLS (RG)

Other soils having a spodic B horizon.

PODZOLS (PZ)

Other soils having 25 percent or more plinthite by volume in a horizon which is at least 15 cm thick within 50 cm of the surface or within a depth of 125 cm when underlying an albic E horizon or a horizon which shows stagnic properties within 50 cm of the surface or gleyic properties within 100 cm of the surface.

PLINTHOSOLS (PT)

Other soils having a ferralic B horizon.

FERRALSOLS (FR)

Table II.6 Key to the Major Soil Groupings of FAO's 1988 Revised Legend of the Soil Map of the World *(cont'd)*

Other soils having an E horizon showing stagnic properties at least in part of the horizon and abruptly overlying a slowly permeable horizon within 125 cm of the surface, and lacking a natric or a spodic B horizon.

PLANOSOLS (PL)

Other soils having a natric B horizon.

SOLONETZ (SN)

Other soils having a mollic A horizon with a moist chroma of 2 or less to a depth of at least 15 cm, shoeing uncoated silt and sand grains on structural pedfaces; having an argic B horizon.

GREYZEMS (GR)

Other soils having a mollic A horizon with a moist chroma of 2 or less to a depth of at least 15 cm; having a calcic or petrocalcic horizon, or concentrations of soft powdery lime within 125 cm of the surface, or both.

CHERNOZEMS (CH)

Other soils having a mollic A horizon with a moist chroma of more than 2 to a depth of at least 15 cm; having one or more of the following: a calcic, petrocalcic or gypsic horizon, or concentrations of soft powdery lime within 125 cm of the surface.

KASTANOZEMS (KS)

Other soils having a mollic A horizon; having a base saturation (by NH_4OAc) of 50 percent or more throughout the upper 125cm of the soil.

PHAEOZEMS (PH)

Other soils having an argic B horizon showing an irregular or broken upper boundary resulting from deep tonguing of the A into the B horizon or from the formation of discrete nodules larger than 2 cm, the exteriors of which are enriched and weakly cemented or indurated and have redder hues and stronger chromas than the interiors.

PODZOLUVISOLS (PD)

Other soils having a gypsic or a petrogypsic horizon within 125 cm of the surface; having no diagnostic horizons other than an ochric A horizon, a cambic B horizon or an argic B horizon permeated with gypsum or calcium carbonate, a calcic or petrocalcic horizon.

GYPSISOLS (GY)

Other soils having a calcic or a petrocalcic horizon, or a concentration of soft powdery lime, within 125 cm of the surface; having no diagnostic horizons other than an ochric A horizon, a cambic B horizon, or an argic B horizon which is calcareous.

CALCISOLS (CL)

Other soils having an argic B horizon with a clay distribution which does not show a relative decrease from its maximum of more than 20 percent within 150 cm of the surface; showing gradual to diffuse horizon boundaries between the A and B horizons; having nitic properties in some subhorizon within 125 cm of the surface.

NITISOLS (NT)

Other soils having an argic B horizon which has a cation exchange capacity equal to or more than 24 $cmol_c$ kg^{-1}clay and a base saturation (by NH_4OAc) of less than 50 percent in at least some part of the B horizon within 125 cm of the surface.

ALISOLS (AL)

Other soils having an argic B horizon which has a cation exchange capacity of less than 24 $cmol_c$ kg^{-1}clay and a base saturation (by NH_4OAc) of less than 50percent in at least some part of the B horizon within 125 cm of the surface.

ACRISOLS (AC)

Other soils having an argic B horizon which has a cation exchange capacity equal to or more than 24mol_c kg^{-1} clay and a base saturation (by NH_4OAc) of 50 percent or more throughout the B horizon to a depth of 125 cm.

LUVISOLS (LV)

Other soils having an argic B horizon which has a cation exchange capacity of less than 24 $cmol_c$ kg^{-1}clay and a base saturation (by NH_4OAc) of 50 percent or more throughout the B horizon to a depth of 125 cm.

LIXISOLS (LX)

Other soils having a cambic B horizon.

CAMBISOLS (CM)

Table II.7 Diagnostic horizons, properties and materials in World Reference Base for Soil Resources

Horizons	
Albic	light coloured eluvial horizon generally associated with argic and spodic horizons.
Andic	moderately weathered horizon in pyroclastic material dominated by short-range order minerals.
Anthropedogenic	surface and subsurface horizons resulting from long continued cultivation.
Argic	subsurface horizon with distinct clay accumulation.
Calcic	horizon with accumulation of calcium carbonate.
Cambic	subsurface horizon showing evidence of alteration relative to the underlying horizon(s).
Chernic	thick, well structured, black, base-saturated surface horizon, rich in OM and biological activity.
Cryic	perennially frozen mineral or organic soil horizon.
Duric	subsurface horizon with weakly cemented to indurated nodules cemented by silica ("durinodules").
Ferralic	strongly weathered subsurface horizon with low cation exchange capacity.
Ferric	horizon in which iron is concentrated in large mottles or concretions.
Folic	well aerated, highly organic surface horizon.
Fragic	natural, non-cemented subsurface horizon with access for roots and water only to interped faces.
Fulvic	thick, black, OM-rich horizon, associated with pyroclastic deposits and vegetation other than grassland.
Gypsic	horizon with accumulation of gypsum.
Histic	poorly aerated, waterlogged, highly organic surface horizon.
Melanic	thick, black, OM-rich horizon, associated with pyroclastic deposits and grassland vegetation.
Mollic	thick, dark coloured surface horizon with high base saturation and moderate to high OM content.
Natric	subsurface horizon with distinct clay accumulation and a high exchangeable sodium percentage.
Nitic	clayey subsurface horizon with strongly developed, nut-shaped structure and shiny pedfaces.
Ochric	weakly developed surface horizon, either light-coloured, or thin, or having a low OM content.
Petrocalcic	continuous cemented or indurated calcic horizon.
Petroduric	continuous cemented or indurated duric horizon.
Petrogypsic	continuous cemented or indurated gypsic horizon.
Petroplinthic	continuous cemented or indurated plinthic horizon.
Plinthic	iron-rich, humus-poor subsurface horizon irreversibly hardening upon repeated wetting and drying.
Salic	surface or shallow subsurface horizon with a high soluble salt content.
Spodic	dark coloured subsurface horizon with illuvial aluminoorganic complexes, with or without iron.
Sulfuric	extremely acid subsurface horizon with sulphuric acid resulting from oxidation of sulphides.
Takyric	heavy textured crusted surface horizon occurring under arid conditions.
Umbric	thick, dark coloured surface horizon with low base saturation and moderate to high OM content.
Vertic	clayey subsurface horizon dominated by shrink-swell clays.
Vitric	horizon dominated by volcanic glass and other primary minerals derived from volcanic ejecta.
Yermic	surface horizon with desert pavement or a loamy vesicular crust covered by windblown deposits.
Properties	
Abrupt textural change	sharp increase in clay content within a limited depth range.
Albeluvic tonguing	penetrations of clay and iron-depleted material into an argic horizon.
Alic	very acid mineral soil material with a high amount of exchangeable aluminium.
Aridic	presence of properties (low organic matter, aeolian activity, light colours, high base saturation) in surface horizons, characteristic of arid environments.
Continuous hard rock	presence of coherent rock, practically impermeable for roots.
Ferralic	low cation exchange capacity.
Geric	extremely low to negative effective cation exchange capacity.
Gleyic	wetness producing reduced conditions caused by groundwater.
Permafrost	perennial temperature at or below 0°C.
Soft powdery lime	accumulation of translocated calcium carbonate in soft powdery form.
Stagnic	wetness producing reduced conditions caused by stagnating surface water.
Strongly humic	high OM content.
Soil material	
Anthropogenic	unconsolidated mineral or organic material produced by human activity.
Calcaric	material containing calcium carbonate.
Fluvic	fresh fluviatile, lacustrine or marine sediments.
Gypsiric	material containing gypsum.
Organic	material containing a very high amount of organic debris.
Sulfidic	waterlogged deposit containing sulphides, and only moderate amounts of calcium carbonate.
Tephric	unconsolidated, non or only slightly weathered pyroclastic products.

Table II.8 Key to the Reference Soil Groups of World Reference Base for Soil Resources

Soils having a *histic* or *folic* horizon,
1. *either* a. 10 cm or more thick from the soil surface to a lithic or paralithic contact;
   or b. 40 cm or more thick and starting within 30 cm from the soil surface; *and*
2. lacking an *andic* horizon starting within 30 cm from the soil surface.

HISTOSOLS

Other soils having one or more *cryic* horizons within 100 cm from the soil surface.

CRYOSOLS

Other soils having *either*
1. a *hortic, irragric, plaggic* or *terric* horizon 50 cm or more thick; *or*
2. an *anthraquic* horizon and an underlying *hydragric* horizon with a combined thickness of 50 cm or more.

ANTHROSOLS

Other soils, which are *either*
1. limited in depth by *continuous hard rock* within 25 cm from the soil surface; *or*
2. overlying material with a calcium carbonate equivalent of more than 40 percent, both within 25 cm from the soil surface; *or*
3. containing less than 10 percent (by weight) fine earth to a depth of 75 cm or more from the soil surface; *and*
4. having no diagnostic horizons other than a *mollic, ochric, umbric, yermic* or *vertic* horizon.

LEPTOSOLS

Other soils having
1. a *vertic* horizon within 100 cm from the soil surface; *and*
2. after the upper 20 cm have been mixed, 30 percent or more clay in all horizons to a depth of 100 cm or more, or to a contrasting layer (lithic or paralithic contact, *petrocalcic, petroduric* or *petrogypsic* horizons, sedimentary discontinuity, etc.) between 50 and 100 cm; *and*
3. cracks which open and close periodically.

VERTISOLS

Other soils having
1. *fluvic* soil material within 25 cm from the soil surface; *and*
2. no diagnostic horizons other than a *histic, mollic, ochric, takyric, umbric, yermic, salic* or *sulfuric* horizon.

FLUVISOLS

Other soils having
1. a *salic* horizon starting within 50 cm from the soil surface; *and*
2. no diagnostic horizons other than a *histic, mollic, ochric, takyric, yermic, calcic, cambic, duric, gypsic* or *vertic* horizon.

SOLONCHAKS

Other soils having
1. *gleyic* properties within 50 cm from the soil surface; *and*
2. no diagnostic horizons other than a *histic, mollic, ochric, takyric, umbric, andic, calcic, cambic, gypsic, plinthic, salic* or *sulfuric* horizon within 100 cm from the soil surface.

GLEYSOLS

Other soils having
1. *either* a *vitric or andic* horizon, both starting within 25 cm from the soil surface; *and*
2. having no diagnostic horizons (unless buried deeper than 50 cm) other than a *histic, fulvic, melanic, mollic, umbric, ochric, duric* or *cambic* horizon.

ANDOSOLS

Other soils having a *spodic* horizon starting within 200 cm from the soil surface, underlying an *albic, umbric* or *ochric* horizon, or an *anthropedogenic* horizon less than 50 cm thick.

PODZOLS

Table II.8 Key to the reference soil groups of World Reference Base for Soil Resources *(cont'd)*

Other soils having ***either***
1. a *petroplinthic* horizon starting within 50 cm from the soil surface; ***or***
2. a *plinthic* horizon starting within 50 cm from the soil surface; ***or***
3. a *plinthic* horizon starting within 100 cm from the soil surface when underlying either an *albic* horizon or a horizon with *stagnic* properties.

PLINTHOSOLS

Other soils
1. having a *ferralic* horizon at some depth between 25 and 200 cm from the soil surface; ***and***
2. lacking a *nitic* horizon; ***and***
3. lacking a layer which fulfils the requirements of an *argic* horizon and which has in the upper 30 cm 10 percent or more water-dispersible clay (unless the soil material has *geric* properties or more than 1.4 percent organic carbon).

FERRALSOLS

Other soils having
1. an eluvial horizon, the lower boundary of which is marked, within 100 cm from the soil surface, by an *abrupt textural change* associated with *stagnic* properties above that boundary; ***and***
2. no *albeluvic tonguing*.

PLANOSOLS

Other soils having a *natric* horizon within 100 cm from the soil surface.

SOLONETZ

Other soils having
1. a *chernic* horizon or a *mollic* horizon with a moist chroma of 2 or less to a depth of at least 20 cm or directly below any plough layer; ***and***
2. concentrations of *soft powdery lime* starting within 50 cm of the lower limit of the Ah horizon but within 200 cm from the soil surface; ***and***
3. no *petrocalcic* horizon between 25 and 100 cm from the soil surface; ***and***
4. no secondary gypsum; ***and***
5. no uncoated silt and sand grains on structural ped surfaces.

CHERNOZEMS

Other soils having
1. a *mollic* horizon with a moist chroma of more than 2 to a depth of at least 20 cm or directly below any plough layer; ***and***
2. concentrations of *soft powdery lime* within 100 cm from the soil surface; ***and***
3. no diagnostic horizons other than an *argic, calcic, cambic, gypsic* or *vertic* horizon.

KASTANOZEMS

Other soils having
1. a *mollic* horizon; ***and***
2. a base saturation (by 1*M* NH_4OAc, pH7) of 50 percent or more at least to a depth of 100 cm from the soil surface, or to a contrasting layer (lithic or paralithic contact, *petrocalcic* horizon) between 25 and 100 cm; ***and***
3. no concentrations of *soft powdery lime* within 200 cm from the soil surface; ***and***
4. no diagnostic horizons other than an *albic, argic, cambic* or *vertic* horizon or a *petrocalcic* horizon in the substratum.

PHAEOZEMS

Other soils having
1. ***either*** a *gypsic* or *petrogypsic* horizon within 100 cm from the soil surface, ***or*** 15 percent (by volume) or more gypsum, which is accumulated under hydromorphic conditions, averaged over a depth of 100 cm within 1.5 m from the soil surface; ***and***
2. no diagnostic horizons other than an *ochric* or *cambic* horizon, an *argic* horizon permeated with gypsum or calcium carbonate, or a *calcic* or *petrocalcic* horizon underlying the gypsic horizon.

GYPSISOLS

Other soils having a *duric* or *petroduric* horizon within 100 cm from the soil surface.

DURISOLS

Table II.8 Key to the reference soil groups of World Reference Base for Soil Resources *(cont'd)*

Other soils having
1. a *calcic* or *petrocalcic* horizon within 100 cm of the surface; ***and***
2. no diagnostic horizons other than an *ochric* or *cambic* horizon, an *argic* horizon which is calcareous, or a *gypsic* horizon underlying a petrocalcic horizon.

CALCISOLS

Other soils having an *argic* horizon within 100 cm from the soil surface with an irregular upper boundary resulting from *albeluvic tonguing* into the argic horizon.

ALBELUVISOLS

Other soils having
1. *alic* properties in the major part between 25 and 100 cm from the soil surface; ***and***
2. an *argic* horizon, which has a cation exchange capacity (by M NH$_4$OAc, pH7) of 24 cmol$_c$ kg^{-1} clay or more in some part, either starting within 100 cm from the soil surface, or within 200 cm from the soil surface if the argic horizon is overlain by loamy sand or coarser textures throughout; ***and***
3. no diagnostic horizons other than an *ochric, umbric, albic, andic, ferric, nitic, plinthic* or *vertic* horizon.

ALISOLS

Other soils having
1. a *nitic* horizon starting within 100 cm from the soil surface; ***and***
2. gradual to diffuse horizon boundaries between the surface and the underlying horizons; ***and***
3. no *ferric, plinthic* or *vertic* horizon within 100 cm from the soil surface.

NITISOLS

Other soils having
1. a base saturation (by 1*M* NH$_4$OAc, pH7) of less than 50 percent in the major part between 25 and 100 cm; ***and***
2. an *argic* horizon, which has a cation exchange capacity (by M NH$_4$OAc, pH7) of less than 24 cmol$_c$ kg^{-1} clay in some part, either starting within 100 cm from the soil surface, or within 200 cm from the soil surface if the argic horizon is overlain by loamy sand or coarser textures throughout.

ACRISOLS

Other soils having an *argic* horizon with a cation exchange capacity (by M NH$_4$OAc, pH7) equal to or more than 24 cmol$_c$ kg^{-1} clay throughout, either starting within 100 cm from the soil surface, or within 200 cm from the soil surface if the argic horizon is overlain by loamy sand or coarser textures throughout.

LUVISOLS

Other soils having an *argic* horizon, either starting within 100cm from the soil surface, or within 200 cm from the soil surface if the argic horizon is overlain by loamy sand or coarser textures throughout.

LIXISOLS

Other soils having
1. an *umbric* horizon; ***and***
2. no diagnostic horizons other than an *anthropedogenic* horizon less than 50 cm thick, or an *albic* or *cambic* horizon.

UMBRISOLS

Other soils having ***either***
1. a *cambic* horizon; ***or***
2. a *mollic* horizon overlying a subsoil which has a base saturation (by M NH$_4$OAc, pH7) of less than 50 percent in some part within 100 cm from the soil surface; ***or***
3. one of the following diagnostic horizons within the specified depth from the soil surface:
   a. an *andic* or *vitric* horizon between 25 and 100 cm;
   b. a *plinthic, petroplinthic* or *salic* horizon between 50 and 100 cm, in absence of loamy sand or coarser textures to a depth of at least 100 cm.

CAMBISOLS

Table II.8 Key to the reference soil groups of World Reference Base for Soil Resources *(cont'd)*

Other soils having

1. a texture which is loamy sand or coarser to a depth of at least 100 cm from the surface; ***and***
2. less than 35 percent (by volume) of rock fragments or other coarse fragments within 100 cm from the soil surface; ***and***
3. no diagnostic horizons other than an *ochric, yermic* or *albic* horizon, or a *plinthic, petroplinthic* or *salic* horizon below 50 cm from the soil surface.

ARENOSOLS

Other soils

REGOSOLS

II.2 PROBLEM SOILS IN THE TROPICS, SUBTROPICS AND ARID REGIONS

II.2.1 The tropical environment

The ***tropical regions*** are located roughly between 23.5° north and south of the equator. They cover approximately 38% of the earth surface and are home to about ***half*** of the world population. Tropical regions have a special environment and therefore special soils and soil problems.

II.2.1.1 Temperature

The temperature in tropical lowlands is high (more than 18° C, but usually more than 22-25° C) and more or less constant (variation is normally less than 5° C between the averages of the three warmest and three coldest months). In tropical highlands temperature is lower but also more or less constant.

Elevation (m)	Mean annual temperature °C	Classification
0 – 600	22° +	Isohyperthermic
600 – 1800	15 - 22°	Isothermic
1800 – 3000	8 - 15°	Isomesic

The ***soil temperature*** variations in the subsoil are negligible due to the low heat capacity of soil material, which is about one-fifth of that of water (see example below).

Table II.2.1 Mean monthly and daily soil temperature variations in °C at Jakarta,Indonesia

Soil depth (cm)	Highest month	Lowest month	Daily variation
Air	26.6	22.5	6.9
3	29.9	28.3	5.2
5	29.9	28.7	5.0
10	29.9	28.9	3.1
15	30.0	28.7	1.5
30	30.0	28.5	0.3
60	30.8	28.5	0.05
90	29.8	28.7	0.04
110	29.7	28.8	0.04

II.2.1.2 Solar radiation for photo-synthesis

Tropical regions receive more solar radiation than other climatic zones because of:

- the tilt of the earth axis; and
- the passage of the sun rays through a thinner atmosphere

The photoperiod, important for the photo-synthesis, is ***shorter*** in tropical regions compared to other regions of the world because of the daylength during the growing season, but it is more or less constant throughout the year, in contrast to other areas where longer periods during the summer season alternate with shorter periods during the winter season. Therefore tropical crops can be described as typical "short-day" crops.

II.2.1.3 Rainfall

Rainfall is the most important factor for tropical agriculture, and the rainfall distribution is therefore one of the basic elements of the agro-climatic classification and the agro-ecological zonation (AEZ).

Table II.2.2 Distribution of *Potentially Arable Land with Soil Moisture Limitations* in tropical regions having no temperature limitation, in millions of hectares

Months of moisture limited crop growth	Tropical America	Tropical Africa	Tropical Asia and Pacific	Total	Percentage
0	315	109	81	505	28
4	82	0	56	138	8
6	260	223	130	613	34
8	32	206	89	327	18
10	16	114	72	202	11
12	1	16	12	29	1
Total	706	668	440	1814	100

Note: More than half (52%) of the tropical regions have a long dry season (6-10 months), especially in Africa (64%).

Critically is the distribution of the dry seasons. The occurrence of two dry seasons means a very short growing period, as demonstrated in the diagram below.

Equally important is the *rainfall variability*. The average values given in climatic summaries indicate the "broad picture" only. However, rainfall variability is large and has a large impact on practical farming (when to prepare the land, when to plant seeds, how much to plant, etc.) and land-use planning. Therefore, ***probability estimates*** per week or decade (10 days) are very useful.

II.2.2 Soil-water retention in tropical soils

The differences in structure between different broad types of tropical soils produce dramatic differences in their water retention properties. An Arenosol (sandy soil) empties its pores of gravitational water at tensions close to 0.1 bar, while a Vertisol (clayey soil) does this at 0.5 bar. The moisture retention pattern of a clayey, well aggregated Ferralsol (strongly weathered tropical soil) is hybrid of those two. It holds as much water as the Vertisol up to approximately 0.1 bar, but drains its macropores at about that tension, thus reaching "field capacity" like the Arenosol. At higher tensions the well-aggregated clayey Ferralsol holds a larger amount of water than Arenosols, but less than Vertisols of similar clay content. These Ferralsols act like sands in terms of water movement at low tensions but hold water like clays at higher tensions. Therefore they have a narrower available water range than other clayey soils. Many of these well-aggregated Ferralsols have ***drought problems*** completely out of proportion to their measured clay and water contents. The aggregates may be close to saturation, while the macropores between the aggregates are totally depleted of moisture available to crops. These unique properties are due to the aggregate size distribution. Andosols (volcanic soils) also possess unique water retention characteristics. They hold a considerable more water at lower tensions than other soils because of their high porosity and size of water-stable aggregates. If water content is calculated on weight basis, the figures may range from 100 to 300 percent water. If calculated on a volume basis, their moisture contents at 0.3 bar may run as high as 84 percent and at 15 bar to 45 percent because of their low bulk density. Nevertheless, Andosols hold more water at a given tension than most other soils. Although the quick achievement of "field capacity" is due to the drainage of large pores between large and very large stable aggregates, the high macroporosity within the aggregates makes them hold more water.

II.2.3 Selected tropical soils and their problems

Important soils in the humid tropical regions are Ferralsols, Acrisols, Alisols and Plinthosols. Lixisols, Vertisols, Planosols and Solonetz are important soils in the seasonally dry tropics, whereas Gleysols and acid sulphate soils (Thionic Fluvisols and Thionic Gleysols) occur in both environments.

II.2.3.1 Ferralsols

Environment
Ferralsols occur on level to undulating stable land surfaces of Pleistocene age or older, or younger land if on easy weatherable rocks. Perhumid or humid climates prevail, although sometimes they may be encountered in the seasonally dry tropics as well, as a remnant from past eras with wetter climates than at present. Natural vegetation comprises tropical rain forest, semi-deciduous forest and savannah.

Profile development
Ferralsols typically have an A-Bo-C profile. They are deep and intensely weathered, which has resulted in residual concentration of resistant primary minerals such as quartz, and the formation of stable secondary minerals with a low cation exchange capacity, viz. kaolinitic clays and iron and aluminium (hydr-)oxides ("ferralization"). This mineralogy and the low pH account for a strongly developed microstructure ("pseudo-sand") and yellowish (goethite) or reddish (hematite) colours.

Morphological features
Ferralsols have

- a deep solum, usually several meters, over weathering rock;
- diffuse or gradual transitions between the horizons, unless a stone-line is present;
- a high iron content in the ferralic horizon, together with the good internal drainage responsible for distinct red (hematite) or yellow (goethite) matrix colours, usually without mottles;
- well developed microstructure: kaolinite with a negative surface charge combined with sesquioxides with a positive charge forms strong micro-aggregates of silt ("pseudo-silt") or sand ("pseudo-sand ") size. Because of this process, soils with a very high clay content (60 percent or more) feel loamy in the field and actually have the same mechanical properties as medium or even light textured soils;
- weakly developed macro-structure: absence of well developed blocky or prismatic structures, very fine granules that are more or less coherent in a porous, friable soil mass.

Chemical and physical properties
Ferralsols have as serious limitation a low natural fertility status, virtual absence of weatherable minerals (which can recharge plant nutrients), and a very low water retention capacity. Due to the strong micro-aggregation and the occurrence of large interstitial pores, Ferralsols are normally well drained and fairly resistant to water erosion. Under traditional management, shifting cultivation ("slash-and-burn") is the most common land use. Liming and application of phosphorus and other plant nutrients are essential for permanent cultivation.

As bases are leached from the soil, the adsorption sites that are associated with the permanent charge become increasingly occupied by aluminium. In typical Ferralsols, aluminium saturation is not a great problem because the permanent charge is very small. In

Ferralsols, which intergrade to less weathered soils, aluminium saturation may become important in the sense that Al-toxicity, can hamper the development of non-tolerant crops.

Biological characteristics
The typical homogeneity of Ferralsols (gradual or diffuse transitions between the horizons) may at least be partly attributed to termites. These animals also increase the depth of the solum by destroying remnants of stratification or rock structure. Their nests, tunnels and ventilation shafts increase the porosity and permeability of the soil. As termites preferentially move fine and medium sized particles, they leave the coarse sand, gravel and stones behind. This often results in the formation of a ***stoneline*** in or below the ferralic horizon. The depth of the stoneline reflects the depth of the termite activity.

Management and use
Ferralsols have rather good physical but poor chemical properties. Their great depth, high permeability and stable microstructure make them less susceptible to erosion than many other soils on land with the same slope. Exceptions are the shallow and sandy types.
Ferralsols are easy to work because their consistence is friable under moist conditions. They are well drained but may be droughty because of their low (available) water holding capacity.

The natural fertility status of Ferralsols is poor; weatherable minerals are (almost) absent and the cation retention by the mineral soil fraction is low. In Ferralsols under natural vegetation, the bulk (80-90%) of all "available" plant nutrients (and all living plant roots) is concentrated in the upper 10-50 cm of the soil, because elements that are taken up by the roots are eventually returned to the surface soil with falling leaves and other plant debris. If this process of nutrient cycling is interrupted, e.g. by introduction of low input sedentary subsistence farming, the root zone will rapidly become depleted of plant nutrients.
Maintenance of the organic matter by manuring, mulching or adequate fallow periods and prevention of surface soil erosion are important management requirements.

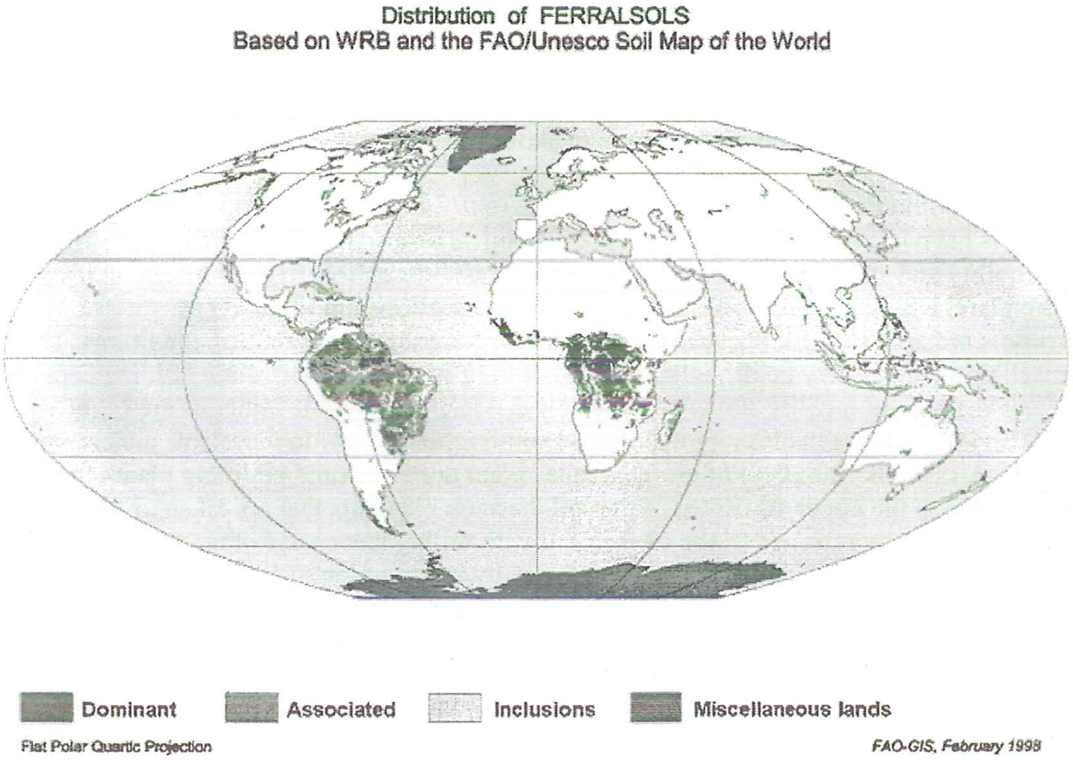
A special problem with Ferralsols (and other soils with a high content of sesquioxides) is the strong sorption of phosphates. Other macro-nutrients, nitrogen and potassium, often are deficient as well, just as the secondary nutrients calcium, magnesium and sulphur and a score of micro-nutrients. Due to the high leaching intensity, even silica deficiency is possible if silica demanding crops (e.g. grasses) are grown. Elements such as manganese and zinc, which are very soluble at low pH, might reach toxic levels, but they may also become deficient on account of their high susceptibility to leaching.

The easily upset ion balance, high losses of cations through leaching (low CEC), and high phosphate fixation complicate liming and fertiliser application. Liming is a means to raise the pH of the rooted surface soil; this combats aluminium toxicity and raises the CEC. On the other hand, liming lowers the anion exchange capacity (AEC), which may lead to collapse of structural elements and slaking of the soil surface. Frequent application ("split application") of small doses of lime or basic slag is therefore preferable over one massive dose. The effects of phosphate fixation can be mitigated by adapting the fertiliser selection, and the mode and/or timing of the P-fertiliser application. Broadcast slow release (rock) phosphate, for instance, is a popular phosphorus source in Ferralsol areas; it is applied at a rate of several tons per hectare and eliminates phosphorus deficiency for a number of years. Alternatively, it could be profitable to use the much more soluble (Triple) Super Phosphate, applied in much smaller quantities and placed in the direct vicinity of the roots.

Ferralsols are grown to a variety of annual and perennial crops. Shifting cultivation and grazing are common as well, and considerable areas are not used for agriculture at all. The good physical conditions and the often level topography would encourage more intensive forms of land use if the problems with respect to their poor natural soil fertility could be overcome.

Regional distribution of Ferralsols

There are close to 750M ha of Ferralsols world-wide. Nearly all of this area is situated in the tropics where Ferralsols occupy some 20 percent of the land surface. They are especially common on the old and stable continental shields of South America and central Africa.



Distribution of FERRALSOLS
Based on WRB and the FAO/Unesco Soil Map of the World

Flat Polar Quartic Projection

FAO-GIS, February 1998

II.2.3.2 Acrisols and Alisols

Environment

Acrisols are most common on old land surfaces with an undulating to hilly topography, in wet and seasonally dry ("monsoon") climates. They occur over any weathering material, but especially over acid rocks (granites, gneiss, etc.). Light tropical (rain) forest to savannah woodlands is the natural vegetation type.

Alisols normally develop in young deposits, often of sedimentary origin, under humid climatic conditions. The vegetation type is normally tropical rainforest, although localised they may occur under open or savannah woodland.

Profile development

Acrisols and Alisols typically have A-(E)-Bt-C profiles. Variations among them are mainly connected with variations in environmental conditions (drainage, seepage, and slope). They have a distinct subsurface horizon, which is richer in clay than the overlying horizons ("argic horizon"). ***Acrisols*** represent strongly weathered soils with a low CEC and a low base saturation (less than 50%), and typically have a kaolinitic mineralogy. ***Alisols***, on the other hand, represent intermediately weathered soils, dominated by 2:1 lattice clays, which are unstable and subject to alteration. Weathering of these clay minerals releases large amounts of magnesium and aluminium, which is characteristic for Alisols.

Chemical and physical properties

The prolonged weathering and advanced soil formation in ***Acrisols*** have caused a dominance of low activity clays and a general paucity of plant nutrients, also reflected in the low pH

Acrisols have. Their nutritional limitations include widespread aluminium toxicity and strong P-sorption as in Ferralsols. They have a weak micro-structure and a massive, though porous macro-structure. The structural weakness of the (surface) soil is caused by the low organic matter content and/or even distribution of sesquioxides over the soil mass with low amounts residing in the surface horizon(s). The bonding between sesquioxides (AEC) and the negatively charged, low activity clay minerals (CEC) is less stable than in Ferralsols.

Alisols, on the other hand, have medium to high activity clays, which result in a high CEC. Due to the rapid weathering of these clay minerals, large amounts of magnesium and aluminium are released, resulting in a very low pH (pH KCl < 4). The aluminium saturation is high, more than 60% and absolutely more than 12 cmol kg^{-1} clay. They normally have strongly developed structures and are clayey.

Management and use

Aluminium toxicity, strong phosphorus sorption, slaking/crusting and a high susceptibility to erosion strongly limit the suitability of ***Acrisols*** for permanent cropping; large areas are used under some form of shifting cultivation. If carefully managed, Acrisols can produce fair yields of various food and fibre crops. In practice, however, they are not very productive unless grown to undemanding, acid-tolerant crops such as cashew nut and pineapple. In view of their high input requirements (lime, fertiliser) and the serious hazard to erosion, Acrisols are best left under their natural vegetation.

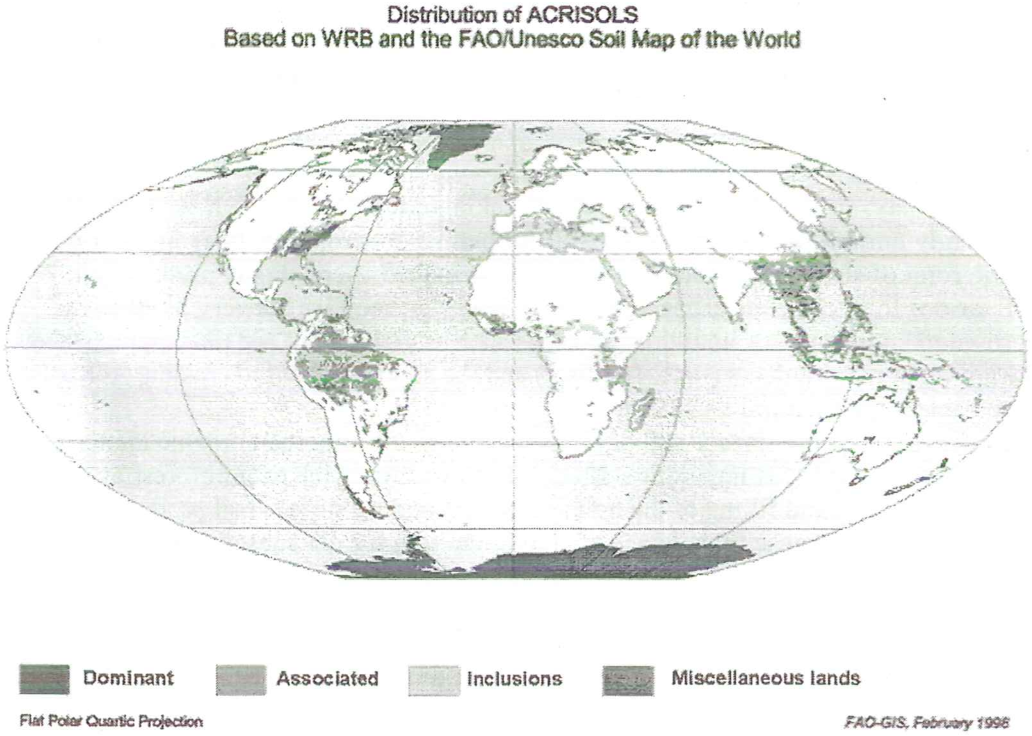
Preservation of the surface soil of Acrisols with its all-important organic matter and (micro-) biological activity is imperative. Mechanical clearing of the natural forest by extraction of root balls and filling of the holes with surrounding surface soil produces land that is largely sterile, because toxic levels of aluminium (the former subsoil !) kill off any seedling planted outside the filled in spots. Adapted cropping systems with careful management, including liming and full fertilisation, are required if resident farming is to be taken up on Acrisols. The commonly used “slash-and-burn” agriculture may seem primitive at first sight, but is really a sophisticated type of land use, developed over centuries of trial and error. If occupation periods are short (one to two years) and followed by a sufficiently long regeneration period (up to several decades), this system probably makes the best use of the marginal Acrisol qualities. On the whole, low input farming is not feasible and sustainable on Acrisols.

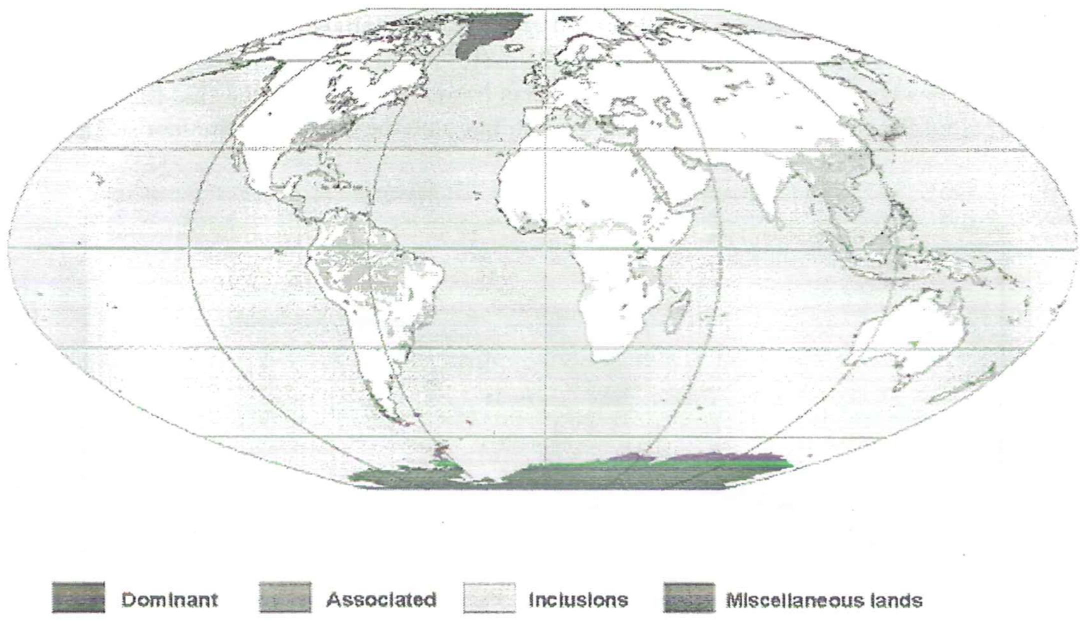
Alisols are short of macro- and micro-nutrients while free aluminium is present in toxic quantities. Liming (to correct the pH of the soil and to suppress the level of free aluminium) and full fertilisation are needed for sustained crop production, but these measures are not always economically feasible. Alisols are traditionally used in shifting cultivation or for low volume production of aluminium-tolerant crops. Alisols on steep lands are best left under their natural vegetation.

Liming, intended to increase the soil pH and to decrease the level of free aluminium, implicitly leads to an increase in CEC and a decrease in AEC. On one hand, this may lower soil structural stability; on the other hand liming boosts (the activity of) the soil fauna which might actually improve the stability of the structure. On average, Alisols are unstable soils, susceptible to erosion. Perennial crops, which minimise soil disturbance by tillage and tree crops are to be preferred over annual crops, particularly on sloping lands. As Alisols still have reasonable nutrient reserves, they are well capable of regenerating after chemical exhaustion. Alisols are only marginally suited for permanent agriculture; low transport rates and aluminium toxicity at shallow depth cause rooting depths to be shallow as well, with the possible consequence of water stress in the dry season. Deep placement of soil amendment (lime, P-fertiliser) proved beneficial in a number of experiments in regions with a pronounced dry season.

Regional distribution of Acrisols and Alisols

Most extensive on acid rocks in Southeast Asia, Southeast USA, the southern fringes of the Amazon basin and in both east and west Africa, Acrisols world-wide cover about 1000M ha, being one of the largest major soil groups. The total area of Alisols has not been ascertained as these soils have been mapped in association with Acrisols . However, it is thought to be in the region of 100M ha. Alisols are typically found in the south-eastern USA, Latin America, Indonesia and China.





Distribution of ALISOLS
Based on WRB and the FAO/Unesco Soil Map of the World

Flat Polar Quartic Projection

FAO-GIS, February 1998

II.2.3.3 Plinthosols

Environment
Plinthosols, soils with mottled clayey materials that become hard as brick when repeatedly exposed to the open air, are found mainly in level to gently undulating sloping areas with a fluctuating water table. They include soils of which the plinthite has hardened into petroplinthite. Plinthosols are more common in weathering material from basic rocks than from acid rock. In any case, sufficient iron needs to be present, originating either from the parent material itself or brought in by seepage water from an adjacent upland area. Plinthosols are associated with rain forest areas, Petric Plinthosols and soils build up from gravely petroplinthic material are more common in the savannah zone.

Plinthosols occur in three physiographically distinct landscape positions (see below):

- as soils with indurated ironstone (massive iron pan or gravel) capping on old erosion surfaces;
- as soils in a low level plain (e.g. a river terrace) with periods of high groundwater;
- on banks of rivers where plinthite becomes exposed and hardens to petroplinthite.

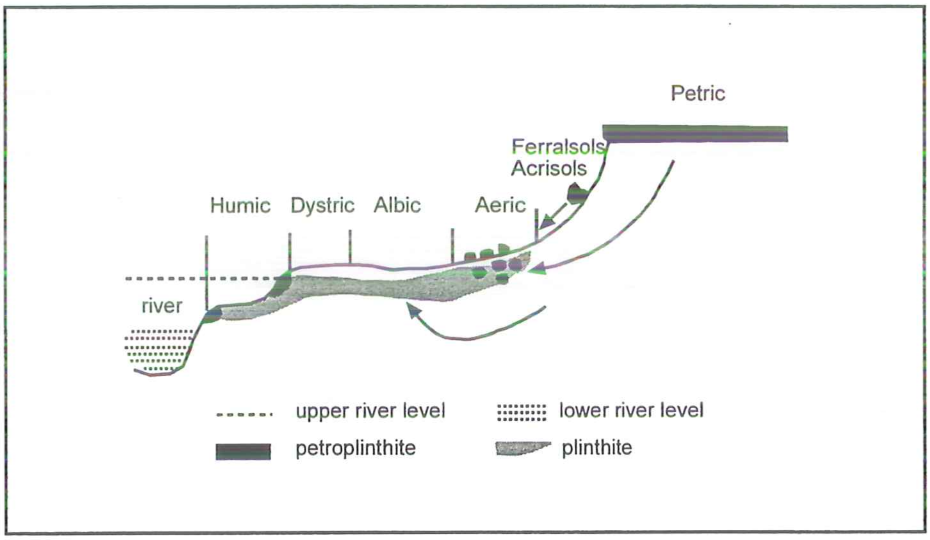
Profile development
Plinthosols have an A-B-C or an A-E-B-C profile. Accumulation of sesquioxides occurs by fertilisation and enrichment from outside sources. Segregation of iron mottles takes place in the zone with a fluctuating water table.

Genesis of plinthite
Plinthite is an iron-rich, humus-poor mixture of clay and quartz. Its formation involves:

1) accumulation of sesquioxides:
   a) relative accumulation because of removal of silica and bases by ferralitization
   b) absolute accumulation through enrichment with sesquioxides from outside
   In most cases both processes act together.

2) segregation of iron mottles, caused by alternating reduction and oxidation as a result of a fluctuating water table. In times of water saturation, much of the iron is in ferrous form (Fe^{2+}), has a high mobility and is easily redistributed. When the water table falls, this iron precipitates as ferric (Fe^{3+}) oxides that are relatively insoluble and will not, or only partially, redissolve in the next wet season.

Plinthite hardens to petroplinthite ("ironstone") when the groundwater table is lowered, e.g. due to a change in the base level or a change in climate. Plinthite forms in low and wet areas; after the land has become drier and the plinthite has hardened to a hardpan, this form shields against erosion. Parts that initially were the lowest, become the highest in the landscape.



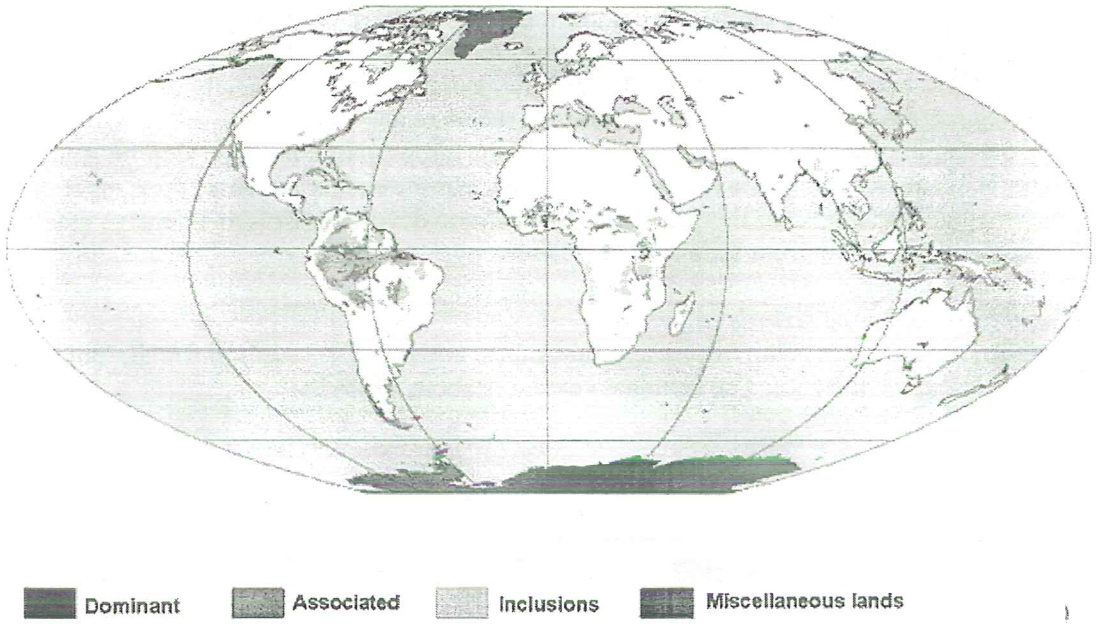
Management and use

Plinthosols are real problem soils. Their nutrient status is low and they pose similar (management) problems as encountered in Ferralsols. Moreover, if the plinthite layer becomes exposed to the surface, ironstone becomes a serious problem, in cases rendering the land useless. Ironstone limits the possibilities for root growth and lowers the water storage capacity and the chemical fertility of the soil.

Civil engineers have a different appreciation of ironstone and plinthite than agronomists have. Plinthite is a valuable material to make bricks for construction purposes. Massive ironstone may be cut into large building blocks. Ironstone gravel can be used for foundations and surfacing of roads and airfields. In some instances it is a valuable ore or iron, aluminium, manganese or titanium.

Regional distribution of Plinthosols

Plinthosols are found in western India (where Buchanan first used the term laterite in 1807), West Africa and parts of South America. All these areas have a hot and humid climate with a very high annual rainfall (in places more than 2000 mm/year) and a short dry season.



Distribution of PLINTHOSOLS
Based on WRB and the FAO/Unesco Soil Map of the World

Flat Polar Quartic Projection

FAO-GIS, February 1998

II.2.3.4 Lixisols

Environment
Lixisols occur on flat to sloping land predominantly in monsoonal and semi-arid regions. Conceivably, Lixisols are polygenetic soils with characteristics formed under a more humid climate than at present. They occur over thoroughly weathered and strongly leached unconsolidated materials, mainly colluvial and alluvial deposits.

Profile development
Lixisols have an A-(E)-Bt-C profile. In places the argic horizon is at the surface or at shallow depth because of erosion.

Chemical and physical characteristics
Lixisols have a higher pH and lower AEC than most other weathered tropical soils. Consequently, the structural stability is lower than, for instance, in Acrisols and Ferralsols (for explanation see section under Ferralsols). Stable pseudo-sand or pseudo-silt is (almost) absent from Lixisols what explains why the moisture content at low pF values is higher than in Ferralsols or Acrisols with the same amount of organic matter and total clay. In the field, the low structural stability may lead to severe slaking and crusting of the surface soil. In extreme cases, water infiltration is reduced to the extent that the soil never attains field capacity.

Being highly weathered, Lixisols have only low levels of available nutrients and low nutrient reserves. The high base saturation of 50 percent or more, one of the characteristics of Lixisols, is as such not a sign of adequate nutrient supply; soils with a low CEC need only small quantities of bases to saturate 50 percent or more of the exchange sites. The chemical properties of Lixisols may yet be better than those of Ferralsols or Acrisols because their higher soil-pH and the absence of serious aluminium toxicity.

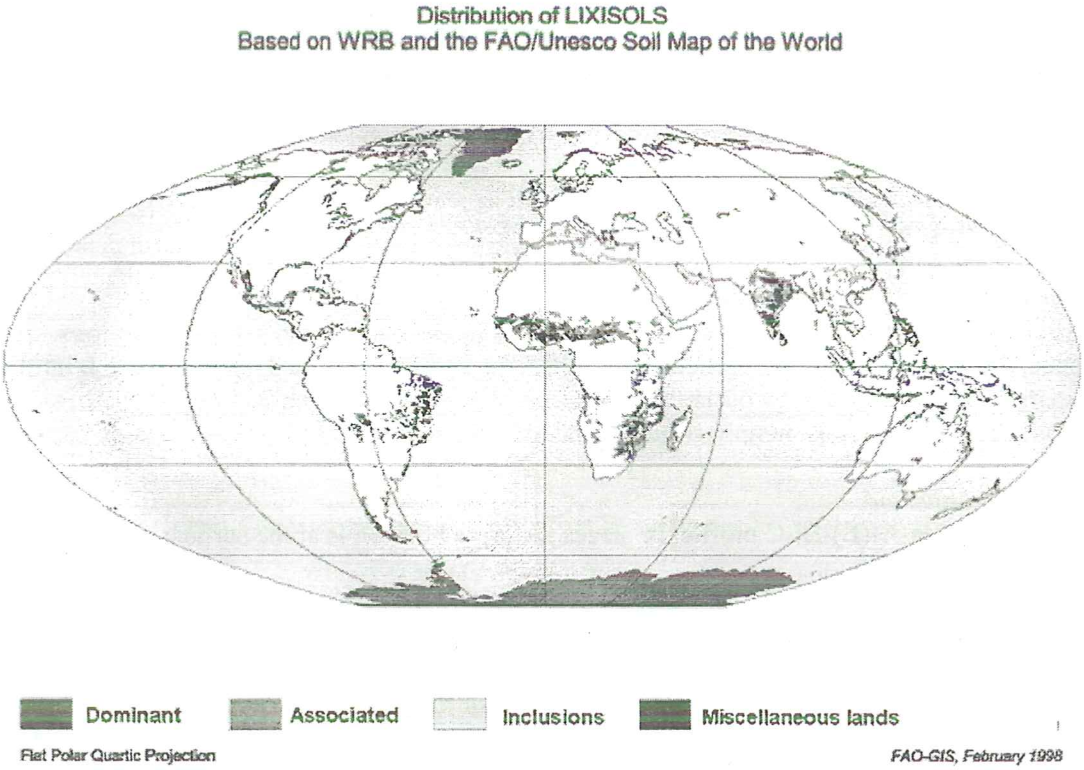
Management and use

Lixisols have low nutrient reserves. They are only marginally suitable for arable farming. The low aggregate stability of the surface horizon is conducive to slaking and/or erosion if the topsoil is exposed to the direct impact of raindrops. Tillage of wet soil or the use of heavy machinery can cause serious deterioration of the surface soil and interfere with the rooting of crops. Minimum tillage and/or erosion control measures such as terracing, contour ploughing, mulching and the use of cover crops help to conserve the soil. Split applications of fertilisers (low CEC !) are needed for good yields. Chemically exhausted and physically deteriorated Lixisols regenerate very slowly unless effectively reclaimed.

Perennial crops are to be preferred over annual crops, particularly on sloping land. Cultivation of tuber crops (cassava, sweet potato, etc.) or groundnut increases the danger of soil deterioration and erosion. In most cases it is better to devote Lixisols to extensive grazing or forestry and to abstain from sedentary agriculture.

Regional distribution of Lixisols

Lixisols cover vast areas in the tropics and subtropics, notably in east-central Brazil, India and in west and Southeast Africa. The estimated extent is about 435M ha.

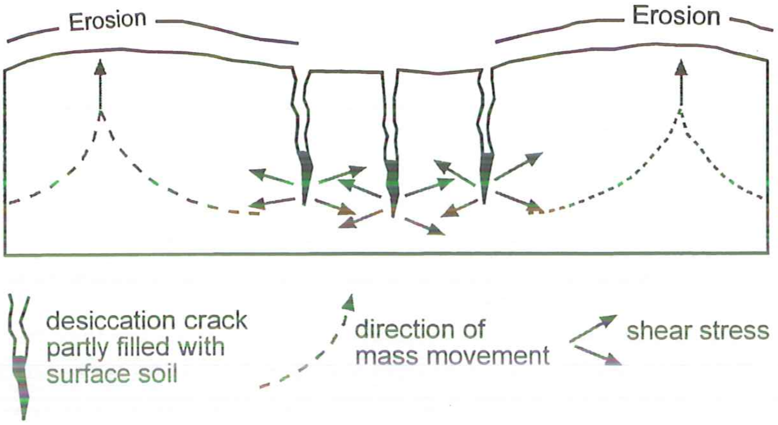


II.2.3.5 Vertisols

Environment

Vertisols occur in depressions and level to undulating areas, mainly in tropical, semi-arid to (sub)humid and Mediterranean climates with an alternation of distinct wet and dry seasons. The climax vegetation is savannah, natural grassland and/or woodland. They are formed in sediments that are finely textured and contain a high proportion of smectitic clay, or products from rock weathering that have these characteristics.

Profile development
Vertisols have an A-B-C profile. Alternating swelling and shrinking of the expanding clays produces deep cracks during the dry season, and slickensides ("shiny surfaces on peds") and wedge-shaped structural elements in the subsurface horizon ("vertic horizon"). The topsoil structure is often fine granular (called soil "mulch"). In many areas a "gilgai" microrelief occurs.



The most common form of gilgai is the normal or round gilgai. On slightly sloping terrain (0.5-2 percent slope) wavy or linear gilgai occurs; lattice gilgai is a transitional form on very slight slopes. Wavy gilgai consists of parallel microridges and microvalleys that run with the slope, i.e. at right angles to the contours. In most gilgai the wavelength (from centre of the mound to centre of the depression) is between 2 and 8 meters, and the vertical interval or "amplitude" is 15 to 50 centimetre.

Chemical and physical characteristics
Most Vertisols have a high cation exchange capacity and a high base saturation. The soil reaction varies from weakly acid to weakly alkaline, pH-values are in the range of 6.0 to 8.0. Higher pH's (8.0-9.5) occur in Vertisols with a high exchangeable sodium percentage (ESP). The CEC is very high, in the range of 30 to 80 $\text{cmol}_c\ \text{kg}^{-1}$ of dry soil. The CEC of the clay is of the order of 50 to 100 $\text{cmol}_c\ \text{kg}^{-1}$. The base saturation percentage is high, usually above 50 and often close to 100 percent with Ca and Mg occupying more than 90 percent of the exchange sites; the Ca/Mg ratio lies between 3 and 1.

Vertisols that are subject to strong pedoturbation are uniform in particle size throughout the solum. An abrupt change may occur when the substratum is reached. When dry, Vertisols have a very hard consistence; they are plastic and sticky when wet and friable only over a rather narrow moisture range. Their physical conditions are greatly influenced by high levels of soluble salts and/or adsorbed sodium.

Saline, sodic and saline/sodic Vertisols occur in the more arid parts of the world. Sodicity occurs incidentally in higher-rainfall areas, e.g. in depressions without outlet. The effect of sodicity on the physical properties of Vertisols is particularly noteworthy. As stated earlier, Na-clays have greater tensile and shear strengths than Ca-clays, and a high ESP produces a soil structure of relative coarse class. In Vertisols that are both saline and sodic, the effect that a high ESP has on the diffuse double layer is commonly offset by the high ionic strength of the soil solution. There is, indisputably, an effect of Na^+ on the structure of Vertisols but the issue is rather controversial. Clay dispersion accompanied by clay movement, the classical consequence of high sodium saturation of clay soils, cannot take place in soils with such a low hydraulic conductivity (see below) and such a low volume of soil that ever becomes saturated with water. In India, ESP's above 7 were found to have an effect on the hydraulic conductivity of Vertisols; in the Sudan Gezira an ESP value of 16 is considered optimal for irrigated cotton cultivation on saline/sodic Vertisols, possibly because a relative high ESP increases the water holding capacity of Vertisols.

Salinity in Vertisols may be inherited from the parent material or may be caused by irrigation. Leaching of excess salt is hardly possible; it is, however, possible to flush salts that have accumulated on the walls of cracks. There are strong indications that the fallow year observed in the rotations of the Gezira/Managil irrigation scheme in the Sudan, is indispensable for maintaining a low salinity level in the surface soil.

Hydrological characteristics

Dry (cracked) Vertisols with a surface mulch or a fine tilth have a high initial infiltration rate. However, once the surface soil is thoroughly wetted and the soil has swollen and the cracks closed, further infiltration is almost nil. The very process of swell/shrink implies a discontinuous and non-permanent pore system. If, at this stage, the rains continue (or irrigation is prolonged), flooding will easily result because infiltration of water in a moist Vertisol is extremely slow. Vertisols have a considerable shrink/swell capacity, but maintaining a relatively fine class of structure, have the highest infiltration rates. Not only the cracks transmit water from the (first) rains but also the pore space that developed between slickensided ped surfaces as the peds shrunk.

Data on the water holding capacity of Vertisols differ widely which may be due to the complex pore space dynamics. Water is retained both at clay surfaces and between the crystal lattice layers. A large proportion of all water, and notably the water held between the basic crystal units, is generally unavailable to plants. By and large, however, Vertisols are considered soils with a relatively good water holding capacity. Investigations in the Sudan Gezira showed that when the clay plain was flooded for several days or even several weeks, the soil moisture content midway between large cracks had hardly changed; it decreased gradually from more than 50 percent in the upper 20 cm to 30 percent at 50 cm depth. Deeper than 100 cm, the soil moisture content was almost invariant (at about 20 percent, corresponding to a matrix suction of some 1500 kPa) throughout the year.

Management and use of Vertisols

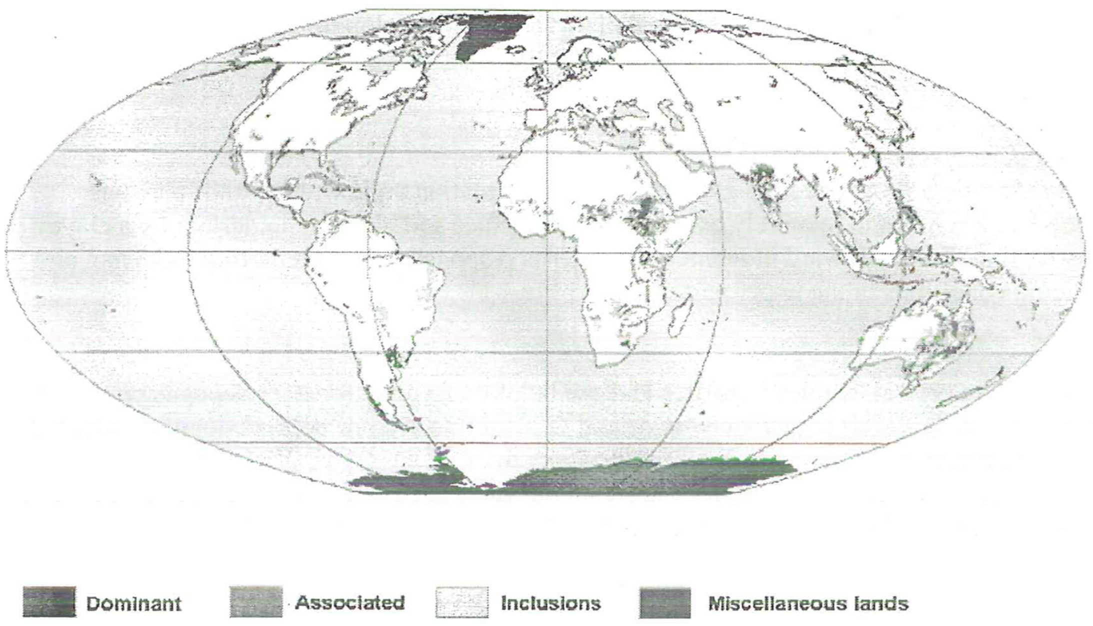
Large areas of Vertisols in the semi-arid tropics are still unused or used only for rough grazing, wood chopping, charcoal burning and the like. Areas that are in use for rainfed and irrigated agriculture are evidence that these soils form a great agricultural potential but special management practices are required to secure sustained production. Problems are in the areas of physical soil characteristics and water management. Assets are the rather high chemical fertility, and the location of Vertisols in extensive level plains, where reclamation and mechanical cultivation are relatively easy. Vertisol plains lend themselves to large-scale mechanised forms of agriculture and are less suited to low-technology farming on account of their poor workability.

Soil erosion takes place even in slightly sloping plains and landslides occur where a wet and plastic surface soil slides over a coherent subsurface soil. Cultivation of annual crops on land with a slope of more than 5 degrees should therefore be discouraged.

Regional distribution of Vertisols

Vertisols cover a total of 335M ha or 2.5 percent of the global land area. An estimated 150M ha is potential crop land. In tropical areas, Vertisols cover some 200M ha or 4 percent of the tropical land surface. A quarter of this is considered to be useful land. Most Vertisols occur in the semi-arid tropics, with an average annual rainfall between 500 and 1000 mm. Large Vertisol areas occur in the Sudan, in India and in Australia. Smaller areas of Vertisols exist in wetter, drier or cooler climates. The annual rainfall may be as low as 150 mm (Sudan) or as high as 3000 mm (Trinidad).

Distribution of VERTISOLS
Based on WRB and the FAO/Unesco Soil Map of the World



Flat Polar Quartic Projection

FAO-GIS, February 1998

Examples of the Sudan and India illustrating the potential and the problems of Vertisols

The 740 000 hectares Gezira/Managuil scheme in the Sudan, gravity-irrigated with Blue Nile water of excellent quality, is the best-known example of large-scale irrigated farming on Vertisols. Cotton is the principal cash crop; other crops grown are wheat, groundnut and sorghum. In addition to irrigated farming, extensive forms of rainfed cropping are practised with sorghum is main crop. This low-input agriculture is increasingly confronted with soil deterioration. With rainfed cropping spreading out onto slightly sloping clay plains, soil erosion has become a major problem.

In India, very slightly sloping Vertisol plains cover the basaltic Deccan plateau. The Vertisols are partly in residual materials and partly in colluvium, and they are also found in alluvial plains. In places, the rainfall sum is low and the pattern erratic. Management practices have therefore been developed that serve three purposes: (1) increased infiltration of water, (2) reduced erosion hazard, and (3) optimum use of the available water.

Some of the measures applied at the Soil and Water Conservation Research Centre at Bellary, Karnataka State, are good examples:

- ☐ *Contour cultivation and contour bunding improve infiltration and make it possible to make better use of the available rain water. These measures also reduce the slope length but there is a danger of water stagnating against bunds and breaking through. This danger is less when bunds are graded. Graded bunds follow "graded contours" which have a very slight slope. Channels are dug on the upslope side of the bund and these drain towards grassed waterways. The waterways end in farm ponds where the excess water is stored for supplemental irrigation of rainfed crops. Broad-base terraces, consisting of a shallow channel and a broad-base low ridge, are also laid out on graded contours. The terraces drain towards grassed waterways.*
- ☐ *Vertical mulching is a method whereby sorghum stubble is placed vertically in trenches, 30 cm deep and 15 cm wide, with the stubble protruding 10 cm above the soil surface. Trenches are 4 to 5 m apart; they are laid out on the contours. Sorghum yields reportedly increased up to 50 percent by vertical mulching.*
- ☐ *The broad bed and furrow system is another example how soil and water conservation can be achieved through improved infiltration and re-use of excess of rain water. It is similar to the broad-base terrace system outlined above but the difference is that two crops are grown, one in the rainy season and one immediately after the rainy season (or a combination of a short-season and a long-season crop). Traditionally only one crop is grown, viz. directly after the rainy season. Double-cropping became possible after a detailed analysis of the rainfall regime over many years permitted optimum timing of the various management options.*

II.2.3.6 Planosols

Environment

Planosols occur in level (depressed) areas in flat to gently undulating terrain in tropical, sub-tropical, semi-arid, sub-humid and humid climates. They develop mostly in alluvial and colluvial deposits of clayey weathering material from various rocks. Natural vegetation type is savannah, woodland savannah or natural grassland.

Profile development

Planosols have a pronounced A-E-Bt-C profile. Destruction and/or removal of clay has created a bleached and relatively coarse textured surface soil abruptly underlain by a clayey subsoil. Impeded downward movement of water is responsible for alternating oxidising and reducing conditions in the upper part of the profile.

Genesis

Planosols are typical "duplex" profiles and can be of geogenetic (coarser deposits on top of finer textured deposits) or pedogenetic origin, or a combination of both. Pedogenetically the following theory of "ferrolysis" has been developed.

In the absence of oxygen (i.e. in water-saturated soils with reducing organic matter), Fe^{3+}-iron from oxides and hydroxides is reduced to Fe^{2+}-ions which go into solution:

$$CH_2O + 4Fe(OH)_3 + 7H^+ \Leftrightarrow 4Fe^{2+} + HCO_3^- + 10H_2O$$

The Fe^{2+}-ions replace absorbed basic cations and aluminium at the exchange complex; the replaced ions are partly leached out. When air re-enters the soil in the next dry period, the exchangeable Fe^{2+} is oxidised again to Fe^{3+}-(hydr)oxide. This produces two H^+-ions of each Fe^{2+}-ion oxidised:

$$4Fe^{2+} + O_2 + 10H_2O \Leftrightarrow 4Fe(OH)_3 + 8H^+$$

Part of the hydrogen ions replace aluminium or basic cations in the clay structure (acid attack); the cations become exchangeable. During the next wet season Fe^{2+}-ions are again formed from iron(hydr)oxides and a new cycle starts.

Chemical and physical characteristics

Planosols with signs of progressed ferrolysis are chemically degraded. The surface soil has lost (much of) its clay; the exchange capacity has become very low and the material is largely depleted of bases because of leaching.

The surface horizons have a low structural stability; if silty, they become hard as concrete in the dry season whereas they turn to a heavy mud with extremely low bearing capacity when waterlogged or inundated in the wet season. The sandy topsoil is hard when dry but not cemented. The poor structural stability of the topsoil, the compactness of the subsoil and the abrupt transition between the topsoil and the subsoil are all detrimental to the rooting possibilities of crops.

Biological characteristics

Waterlogging hampers the growth of most shrubs and trees, particularly if wetness alternates with periods of severe drought. Consequently, the natural vegetation of areas with Planosols is mostly herbaceous. Where trees grow, it concerns species with extensive, shallow rooting systems that are capable of withstanding severe drought and strong winds. Their productivity is generally only a fraction of that in areas with well drained soils of comparable mineralogy and nutrient levels. In line with the adverse physical and chemical soil conditions in Planosols, the soil fauna is not very diverse and the population densities are only low.

Management and use of Planosols

Planosols are used for a variety of land utilisation types, albeit at a lower intensity than most other soils under the same climate. Large areas are used for extensive grazing. Depending on the amount and distribution of the annual rainfall, Planosols are colonised by a sparse grass and sedge vegetation or by grass with scattered bushes, trees or forest. The wood production of trees on Planosols is often only half to less than one third of that on other soils from the same parent material.

Planosols in Southeast Asia are normally planted to a single crop of paddy rice, produced in bunded field that are inundated during the rainy season. Efforts to produce dryland crops on the same land with irrigation during the dry season have met with little success; the soils seem better suited to a second crop of rice with dry season irrigation and fertilisation.

When paddy rice is grown, (improved) irrigation might be double-cropping possible. Besides irrigation, fertilisers are needed to increase rice yields. The soils should be allowed to dry out at least once a year to prevent or minimise micro-nutrient deficiencies or toxicity associated with recurrent soil reduction. Some Planosols will require application of more than just macro-nutrients and their low fertility level may prove difficult to correct.

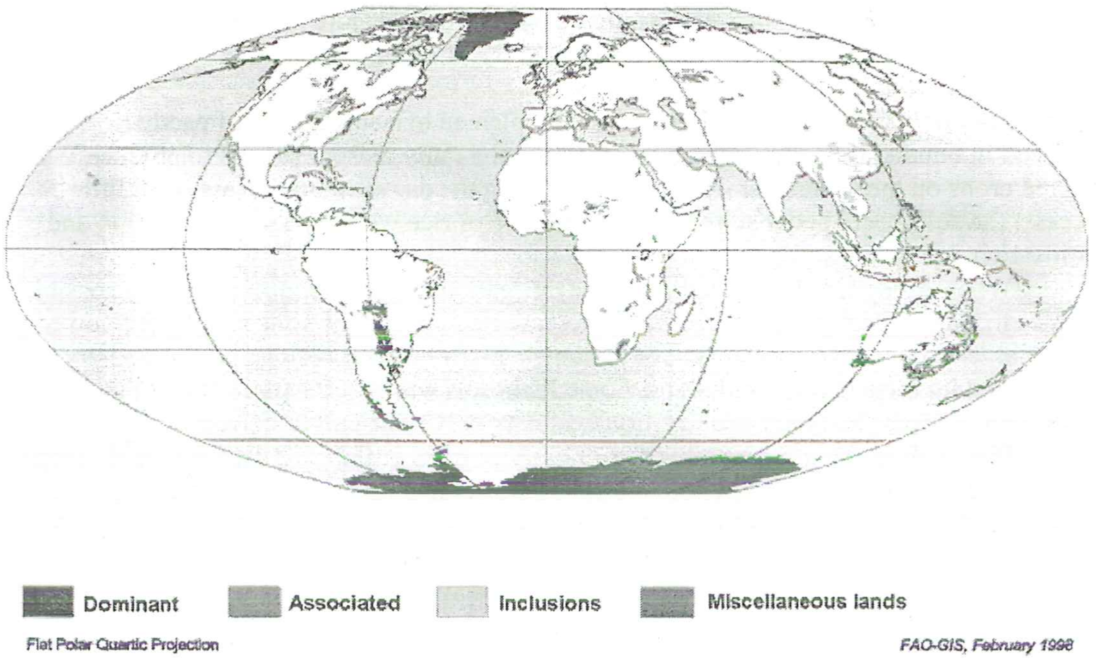
Planosols in the temperate zone are mainly in grass or they are used for arable crops such as wheat or sugarbeet. There too, the yields are comparatively low because of adverse physical and chemical soil conditions. Narrow drain spacing are required because of the low hydraulic conductivity at shallow depth. Besides oxygen deficiency during wet periods, the excessive soil density and toxic levels of aluminium may hinder root growth. Experiments in Europe, spanning a number of years after drainage and deep loosening of the subsoil, with and without placement of lime, failed to show conclusive results.

In climates with long dry periods and short, infrequent wet spells, irrigation of grassland in the dry season may hold promise. Where temperate regimes are favourable for rice, paddy rice cultivation is probably superior to any other kind of land use. Strongly developed Planosols with a silty or sandy surface are perhaps best left in their natural state, without efforts at improvement.

Regional distribution of Planosols

The total area of Planosols is estimated to be 130M ha worldwide, with the most important occurrences in South America (60M ha) and Australia (45M ha). The largest areas of Planosols are in climates with a clear alternation between wet and dry seasons, e.g. in the sub-humid and (semi-)arid tropics of Brazil and Uruguay, the Sahelian countries, and eastern Asia, and in the subtropical areas of Australia, South Africa, South America, southern USA and China. Smaller areas occur outside the tropics and subtropics, e.g. in the mid-western USA and in Russia. Planosols occur predominantly in flat lands with incised drainage ways but they can also be found in other positions in the landscape, e.g. at the foot of slopes, in a strip just above the valley bottom, or in a strip somewhere along the contours of a long and relatively gentle slope.

Distribution of PLANOSOLS
Based on WRB and the FAO/Unesco Soil Map of the World



II.2.3.7 Solonetzes

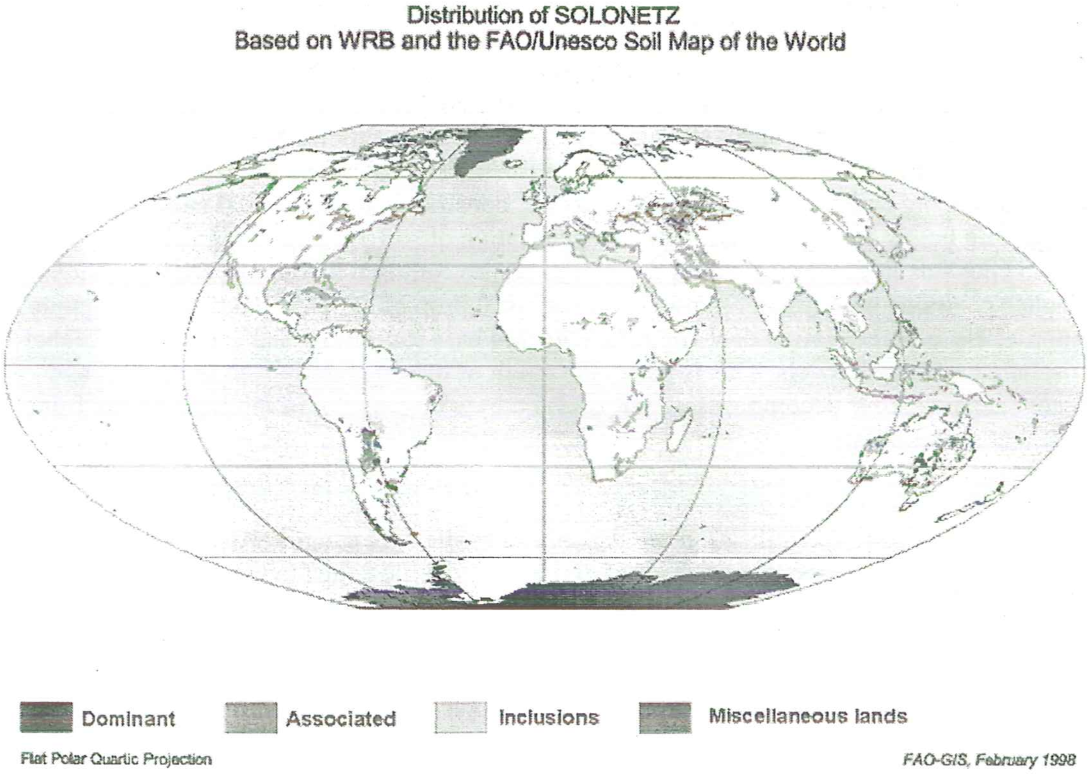
Environment
Solonetzes occur mainly in seasonally dry tropics and sub-tropics, semi-arid and sub-humid climates associated with sodium-bearing parent material or other external sources of sodium. They are found in the lower part of the landscape. The natural vegetation is mainly composed of *Halophytes* and grasses.

Profile development
Solonetzes typically have an A-E-Btn-C profile. The transition between the A or E horizon to the Btn ("natric horizon") is usually abrupt. Characteristically, the top part of the Btn horizon has a prismatic or columnar structure.

Formation
Solonetzes are the result of "alkalinization" (sodification), which is the formation of soils with a high percentage of exchangeable sodium (ESP); for Solonetzes an ESP of 15 or more is required. Under such circumstances, clay particles and humus become dispersed, resulting in sodium-rich clay accumulation in the subsurface horizon. In extreme cases, sodium carbonate can be formed, giving rise to soil pHs of 8.5 or more.

Management and use of Solonetzes
Essential in the reclamation of Solonetzes is the replacement of sodium by calcium, at the same time flushing out the excess sodium. This can be done by liming or, where gypsum is found in the subsoil, by deep ploughing, thus bringing the gypsum to the surface to replace the sodium by the calcium from the gypsum. However, unless there are overriding reasons, these soils can best be left under their natural vegetation, as the dispersed state of clay and humus makes them very difficult to cultivate.

Regional distribution of Solonetzes
Solonetzes are found in scattered areas throughout the world where there is a dominance of sodium over calcium salts. There are about 135M ha of Solonetzes world-wide, mainly in Australia and central Asia.



Distribution of SOLONETZ
Based on WRB and the FAO/Unesco Soil Map of the World

Flat Polar Quartic Projection

FAO-GIS, February 1998

II.2.3.8 Gleysols

Environment
Gleysols are found in depressions with shallow groundwater. They occur over a wide range of unconsolidated materials, mainly sediments of Pleistocene or Holocene age, with a basic to acid mineralogy.

Profile development
Gleysols mostly have an A-(Bg)-Cr of H-(Bg)-Cr profiles. There is evidence of reduction processes with or without segregation of iron ("gleyic properties") within 50 cm of the surface.

Genesis of Gleysols
The genesis of Gleysols is dominated by excessive wetness at some period of the year or throughout the year, which manifests itself in reduction of ferric iron to the mobile form of ferrous iron in the upper 50 cm of the soil. Subsequent oxidation of ferrous iron to (ferric) iron oxides may take place where there is a supply of oxygen. This is the case near larger fissures or cracks and along former root channels ("rusty root channels"). The grey, olive or blue subsoil is permanently saturated and shows no mottling. The presence of organic matter promotes reduction of iron compounds. Only when horizons at or near to the surface are periodically saturated with water in which the products of organic matter decomposition are dissolved, will the characteristic gleyic colour pattern form

Gleyic soil properties are strictly associated with movement of a groundwater table; mottled, oxidised horizons occur on the top of the fully reduced subsoil.

A different type of mottling is found where perched water occurs on top of a slowly permeable subsurface horizon, while the real groundwater occurs at greater depth. Such soils have "stagnic" properties and form units in other major soil groups; a reduced horizon on top of an oxidised subsurface horizon. If peds occur, the reduced horizon tongues into the oxidised subsoil along the faces of the peds.

Chemical, physical and biological characteristics

The dominant characteristic of Gleysols is their prolonged saturation with water, which creates a specific set of physical, chemical and biological conditions. Saturation implies lack of aeration, creating an unfavourable root environment and poor conditions for the soil fauna. Repeated wetting and drying also may lead to soil densification due to weakening on interparticle bonds during saturation and concentration of soil particles upon desaturation.

Gleysols in depressions or at the foot of slopes compare favourable with similar soils at higher positions with respect to nearly all chemical properties. They have higher organic matter levels, a higher cation exchange capacity and base saturation, and usually also higher phosphate and potash levels. This is partly the result of their generally heavier texture and slower organic matter decomposition, but is also due to the transfer in solution of ions from the higher surroundings.

Management and use of Gleysols

The major obstacle to the utilisation of Gleysols and soils with stagnic properties is the necessity to install and adequate system of drainage, either designed to lower the groundwater table, or to intercept seepage or surface runoff water. Where the surface soil is high in organic matter and pH values are low, the soil reaction and the aeration of the surface horizon can be improved by incorporation of lime. This creates a better habitat of micro- and meso-organisms and enhances the decomposition of soil organic matter. Fertilisers must normally be used where permanent agriculture is practised. When adequate drainage is accomplished, Gleysols can be used for arable cropping, dairying or horticulture. If (too) wet soils are cultivated, the structure of the soil may be destroyed for a considerable period of time. Gleysols in low-lying areas with unsatisfactory drainage possibilities are best kept under a permanent grass cover.

The possibility of using Gleysols for wetland rice cultivation was already mentioned. The difficulties discussed for Thionic Fluvisols are also applicable to Thionic Gleysols (see under Fluvisols). In Asia, Gleysols in humid climates are almost invariably planted to wetland rice. Where they occur in narrow valley floors, these are modified into broad and level steps, each with a water-retaining bund. In places, Gleysols are used for tree crops after the water table has been lowered with deep drainage ditches, or by planting trees on parallel ridges that alternate with wide (10-20 cm) but shallow depressions that are planted to rice in the wet season.

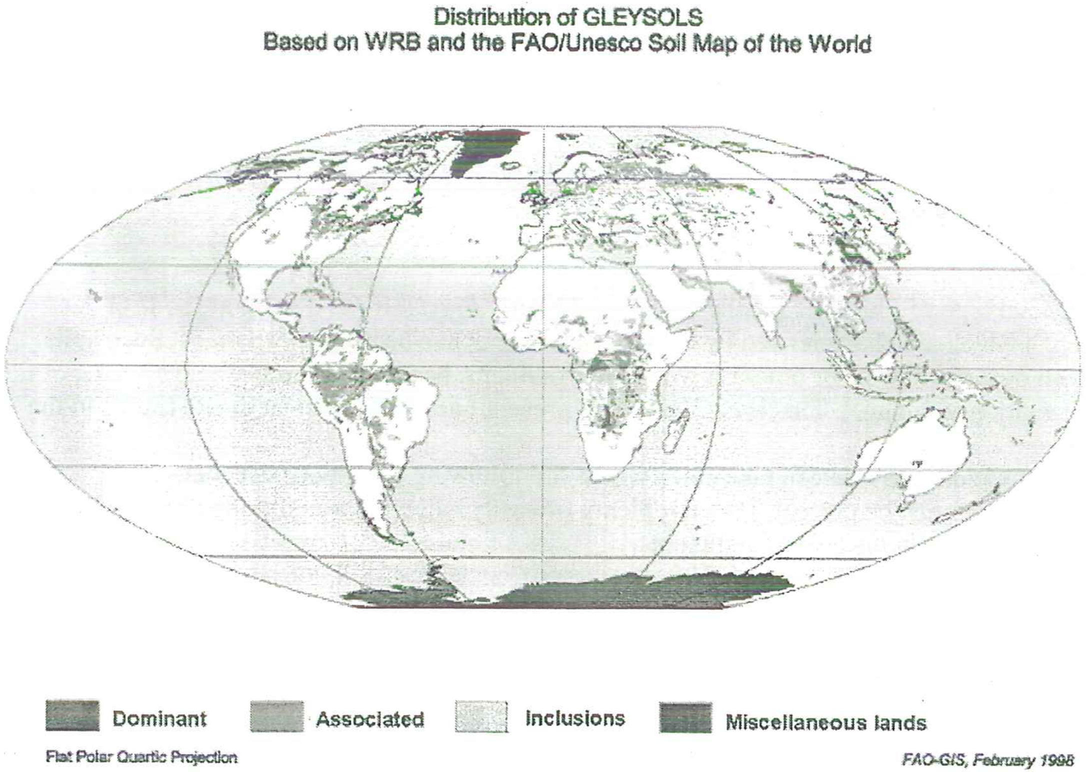
Paddy soils (rice-soils)

A special category of hydromorphic soils is the so-called paddy soils. They may be true Gleysols (or wet Fluvisols), but they also form the separate reference soil unit of ***Hydragric Anthrosols***, soils of which the characteristics have been profoundly modified by human action, in this case by long-continued cultivation under wet conditions (e.g. puddling). This means that there are basically three different types of paddy soils:

1) the (groundwater-influenced) Gleysols and Fluvisols of wet alluvial plains (with a completely reduced subsoil);
2) other soil groups which have stagnic properties due to the frequent inundation for wet rice cultivation, but without sufficient signs of long-continued wet cultivation; and

3) the Hydragric Anthrosols, which show all signs of long-continued wet cultivation (puddled layer, plough pan with vesicular pores, illuvial iron and manganese coatings).

Regional distribution of Gleysols
The greatest extent of Gleysols is in the cool humid part of the world. Gleysols occupy some 720M ha world-wide, and some 200M ha in the tropics and subtropics.



II.2.3.9 Fluvisols

Environment
Fluvisols occur in all climates on almost all level areas in low-lying situations, which are subject to periodic flooding. A wide range of profile features have developed as a result of the different particle size and drainage characteristics of the sediments on floodplains, levees and back-swamps. Most Fluvisols have a grassland vegetation, especially when the water table is high, while the relatively high-lying Fluvisols, e.g. the levees, may sustain even a tropical rain forest.

Profile development
Fluvisols have an A-C or A-Cr profiles. Many of the have a high groundwater table and exhibit gleyic properties. Depending on the environment in arid regions Fluvisols may contain accumulated salts, while in marine environments sulphur deposits in the form of pyrite may occur.

Chemical and physical characteristics
Most Fluvisols are very fertile; they have neutral or near-neutral pH values, which do not impair the availability to nutrients. Coastal sediments usually contain $CaCO_3$ (shells!), and the

exchange complex is saturated with bases from the seawater. Unfavourably high sodium saturation is not uncommon and a high electrical conductivity of the soil moisture is often a problem. This phenomenon may be seasonal in a monsoon climate where rain leaches excess salts out of the root zone. Soil salinity is particularly widespread in Fluvisols of the dry zone; in equatorial climates, the salinity problem is limited to coastal areas that are under the direct influence of the tides.

The ripening stage is an indication of the maturity of alluvial sediments. The n-value, which represents the quantity of water absorbed per gram clay is a measure for the ripening stage but in the field ripening is judged by squeezing a lump of soil material through one's fingers. Depending on the resistance felt the soil material is classified ***ripe*** or ***unripe***. Such soils pose problems for agricultural use; they have a low bearing capacity and machines cannot be used on them.

One would expect clayey Fluvisols to have few pores and low hydraulic conductivity. However, many of the coastal plains are or were once colonised by adapted vegetation (e.g. mangroves or reeds), that makes large tubular pores. Fluvisols on river levees and coastal sand ridges are porous as well.

Thionic Fluvisols

A special type of Fluvisols are the so-called ***Thionic Fluvisols*** (acid sulphate soils or "cat clays") which are characterised by the presence of sulphur-bearing compounds. Such soils, which occur in marine or brackish water environments, become extremely acid after reclamation (drainage). This is caused by the presence and behaviour of pyrite (FeS_2) in the soil.

Formation of pyrite can take place where the following conditions are met:

a) iron must be present. Usually, plenty of easily reducible iron oxides or hydroxides are present in the coastal sediments.
b) sulphur must be present. This originates from sulphates in the seawater.
c) anaerobic conditions must prevail to allow reduction of sulphate and iron oxides. This condition is met in fresh coastal waters.
d) iron- and sulphate-reducing microbes must be present; they occur in all coastal sediments.
e) organic matter is needed as a source of energy for the microbes; it is supplied by lush pallustric vegetation (mangrove forests, reeds, and sedges).
f) tidal flushing must remove the alkalinity formed in the process of pyrite formation.
g) a slow sedimentation rate is necessary. Otherwise, the time will be too short to form sufficient pyrite for a potentially acid sediment.

Pyrite is formed as follows: in reduced conditions (under water) microbes reduce ferric (Fe^{3+}) iron to ferrous (Fe^{2+}) ions, and sulphate (SO_4^{2-}) to sulphide (S^{2-}). During the process, organic matter is decomposed, ultimately to bicarbonates. In essence, a potentially acid part (the pyrite) and an alkaline part (bicarbonates) is formed in an initially neutral system. Tidal flushing removes the alkalinity (HCO_3^-), and potentially acid pyrite remains behind in the sediment. Pyrite is formed in a number of steps, but the "overall" reaction equation is:

$$Fe_2O_3 + 4SO_4^{2-} + 8CH_2O + \frac{1}{2} O_2 \Leftrightarrow 2FeS_2 + 8HCO_3^- + 4H_2O$$

When oxygen enters a pyritic sediment, pyrite will be oxidised, by microbial intervention, to sulphuric acid (H_2SO_4) and iron hydroxide ($Fe(OH)_3$). Important intermediate products in this process are soluble ferrous sulphate ($FeSO_4$) and the meta-stable mineral jarosite ($KFe(SO_4)_2(OH)_6$). Jarosite has a typical straw-yellow colour which is easily recognised in the field and is typical of ***actual acid sulphate soils*** (as distinguished from ***potential acid sulphate soils;*** the difference will be discussed later).

The oxidation of pyrite takes place in steps, depending on the quantity of oxygen present. The first step (at a low redox potential) is oxidation of pyrite to soluble ferrous sulphate:

$$FeS_2 + 3\tfrac{1}{2}O_2 + H_2O \Leftrightarrow Fe^{2+} + 2SO_4^{2-} + H^+$$

Next, at a higher redox potential, soluble ferrous sulphate is transformed to jarosite:

$$Fe^{2+} + {}^2/_3SO_4^{2-} + {}^1/_3K^+ + {}^1/_4O_2 + {}^3/_2H_2O \Leftrightarrow {}^1/_3\text{Jarosite} + H^+$$

Finally, slow hydrolysis of jarosite to ferric hydroxide takes place:

$${}^1/_3\text{Jarosite} + H_2O \Leftrightarrow {}^1/_3K^+ + Fe(OH)_3 + {}^2/_3SO_4^{2-} + H^+$$

Adding up these reactions, the total reaction from pyrite to ferric hydroxide reads:

$$FeS_2 + {}^{15}/_4O_2 + {}^7/_2H_2O \Leftrightarrow Fe(OH)_3 + 2SO_4^{2-} + 4H^+$$

The above reaction equations indicate that one mole of pyrite ultimately leads to the release of four moles of acid; two moles during the oxidation to ferrous sulphate, one more during oxidation of jarosite, and the final one during hydrolysis of jarosite to ferric hydroxide.

If sediment contains a surplus of lime or other carbonates over pyrite, the (sulphuric) acid formed will be neutralised and gypsum ($CaSO_4.2H_2O$) will be formed. With little or no lime present, the pH of the soil will drop dramatically and large quantities of H^+-ions will be released to the soil solution. First, the H^+-ions will exchange with bases at the cation exchange complex but once the pH has dropped to a level between 3 and 4, clay minerals are attacked, preferentially smectites with their high specific surface. Mg and Fe, but mostly Al are released from the clay lattices, and Al^{3+}-ions become dominant in the soil solution and at the exchange complex.

The ***hydrological and physical properties*** of Thionic Fluvisols are similar to those of other Fluvisols but their chemical characteristics are decidedly different. Within the Thionic Fluvisols, a distinction can be made between Protothionic Fluvisols (potential acid sulphate soils), which are not yet oxidised but contain pyrite in the soil material, and Orthithionic Fluvisols (actual acid sulphate soils), which are oxidised and acidified.

The unfavourable properties of ***Protothionic Fluvisols*** are:

- salinity: Protothionic Fluvisols are mostly situated in coastal areas with tidal influence.
- strong acidification upon drainage.
- low accessibility/trafficability: Protothionic Fluvisols occur in unripe, soft mud.
- high permeability: the roots of the (former) pallustric vegetation made the sediments excessively permeable to water.
- flooding of the land: flooding at spring tide may cause damage to crops.
- engineering problems arise when dikes, etc. are constructed on the soft mud. Acidity from the oxidised dikes attacks steel and concrete structure.

The unfavourable properties of ***Orthithionic Fluvisols*** are:

- low pH: most plants can tolerate pH values as low as pH 4, but only if the supply of nutrients is well balanced.
- aluminium toxicity: occurs in soils with a pH below 4.5; generally valid toxicity limits cannot be given as the toxicity depends on the availability of nutrients, the growth stage of the plant, and the kind of crop.
- salinity: in Orthithionic Fluvisols not always caused by salts from sea water; sulphate can build up in the soil solution to the extent that the soil is to be regarded saline.
- phosphorus deficiency: high aluminium levels in the soil solution cause precipitation of insoluble Al-phosphates, thus immobilising P.
- ferrous-iron (Fe^{2+}) toxicity is a common problem when rice is cultivated on Orthithionic Fluvisols. Reduction of insoluble ferric iron (Fe^{3+}) to soluble ferrous iron components takes place in flooded rice fields.

- acidification of the surface water: when Orthithionic Fluvisols are flooded for rice cultivation, soluble ferrous iron can diffuse to the surface water. This water contains oxygen, and the ferrous iron can be oxidised to ferric iron again. This process acidifies the surface water and can cause irreparable damage to engineering structures and fish in a very short time.
- N-deficiency: mineralisation of organic matter by microbial activity is slow in wet Orthithionic Fluvisols.
- engineering problems: acidity from soil surface water attacks steel and concrete structures.
- H_2S-toxicity: becomes a problem when Orthithionic Fluvisols are flooded for long periods (a year or longer). Sulphate can then be reduced to H_2S, which is toxic at very low concentrations.

Management and use of Fluvisols

The high natural fertility of most Fluvisols suggests that there are possibilities for a wide range of agricultural uses, but that is only partly the case. Levee soils and the higher parts of marine landscapes, which do not suffer from wetness, can be planted to a wide range of dryland crops. In tropical lowlands with a year-round supply of fresh water, three crops per year are possible. Such places are among the most densely populated parts of the world, and often have been under intensive use since historical times. Land tenure is often characterised by small plots, and most production is for home consumption or for local markets.

The most obvious and widespread use of tropical Fluvisols is paddy rice cultivation. The requirements in terms of irrigation and drainage facilities are high. For good yields, paddy land should be dry for at least a few weeks every year. Else, the soil's redox potential will become so low that nutritional toxicity's (iron, H_2S) develop. Besides, a dry period stimulates microbial activity and promotes mineralization of organic matter. For all these reasons, Fluvisol areas under rice normally have complex canal systems. Other suitable crops besides rice can be jute, kenaf and various tuber crops (*Colocasia, Xanthosoma*). Coconut survives periodic flooding and some degree of salinity. Tidal lands that are strongly saline are normally under mangroves or some other salt-tolerant vegetation. Such areas are used for fishing, hunting, salt pans, or wood cutting for charcoal or firewood.

Thionic Fluvisols are widely left idle. Reclaimed and/or carefully managed, they can be used as (wet) rice lands in the tropics and as pasture land in the temperate zone.

There are two (opposite) strategies thinkable for reclaiming and using ***Protothionic Fluvisols***:

1. The first one is to drain and completely oxidise the soil, and leach the acidity formed. Leaching can be done with saline or brackish water; this will not only remove soluble acidity, but also expel undesirable cations (like Al) from the exchange complex. The advantage of this approach is that the problem is solved once and for all. The disadvantages are that it is an expensive method, poses a threat to the environment (acid drain water!) and depletes the soil of useful elements together with the undesirable ones. The method has been applied with some success in coastal rice growing areas in Sierra Leone and in areas with fish ponds in the Philippines, but turned out disastrous in Senegal, where insufficient water was available for leaching, and in the Netherlands where one generation of farmers barely survived the reclamation of the Haarlemmermeer polder.
2. The second strategy is to try to limit pyrite oxidation by maintaining a high groundwater table. A precondition is the availability of sufficient water. This method also requires rather heavy investments in water management, and the potential danger of acidification remains present. The method is widely applied, both in temperate regions and in the tropics, often with ingenious adaptations to suit local conditions and practices.

When discussing management and use of **Orthithionic Fluvisols**, a distinction should be made between areas with shallow inundation (less than 60 cm) and areas with deep inundation. In areas with shallow inundation and at least some tidal influence in creeks or canals, flap gates can be used for water control. In the rainy season, water can be discharged at low tide, or irrigation can be applied as needed. In some areas, the spring tide can be used for irrigation in the dry season, provided that the tide is freshwater tide. If the tide is not high enough to flood the land for rice, dryland crops can be cultivated by using flap gates to maintain a shallow groundwater table.

For rainfed rice, the "intensive shallow drainage system" is sometimes practised. Shallow ditches are dug at short spacing and the spill is spread over the land. This system ensures efficient leaching of the surface soil at the start of the rainy season.

Dryland crops can only be grown on raised beds whereby care must be taken not to turn the profile upside down and bring the most acid part, the subsoil, to the top.

Table II.2.3 *Lime requirement for complete neutralisation of a 10 cm thick soil layer of a Thionic Fluvisol, in tonnes of lime ha^{-1}.*

Apparent density of the soil g cm^{-3}	**Percent oxidizable sulphur**					
	0.5	1	1.5	2	3	4
0.6	9	19	28	37	56	74
0.8	12	25	37	50	74	112
1.0	16	31	47	62		
1.2	19	37	56	74		

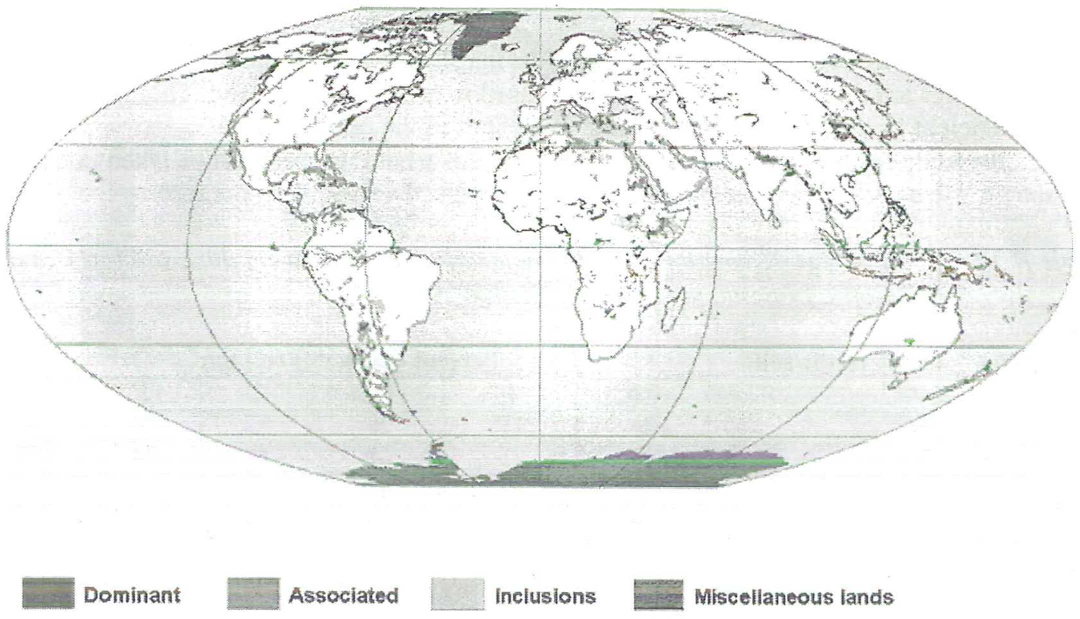
It is virtually impossible to eliminate the acidity problem completely by applying lime to the soil. The table above shows the lime requirement for complete neutralisation of soils with various contents of oxidizable sulphur. The neutralising capacity of a 10 cm thick layer (without lime; only neutralisation by exchange ions) is also given. The table shows the practical impossibility of the liming option: no farmer can afford to apply on an average Thionic Fluvisol (say 1.5 percent sulphur and an apparent density of 1.0 g cm-3) a total of 47 tonnes of lime. And that covers only the need of the top 10 cm, assuming that no new acid forms during the next dry season.

Yields on Thionic Fluvisols are commonly low, or very low. Most land with Orthithionic Fluvisols having a "sulfuric horizon" within 50 cm of the soil surface, is not used for agriculture at all. This particularly hostile "wet desert" land has a pallustric vegetation, a limited fauna, and acid surface water, especially early in the wet season when acid substances such as aluminium-sulphate dissolve. The situation seems to be a little less bleak in Thionic Fluvisol areas in the equatorial climatic zone than in monsoonal climates. The soils will not dry out as easily as in the monsoon are, and more water is available for water management measures throughout the year.

Regional distribution of Fluvisols

Fluvisols occur world-wide on river floodplains, deltaic areas and coastal marine lowlands. The estimated total area is 350M ha.

Distribution of FLUVISOLS
Based on WRB and the FAO/Unesco Soil Map of the World



Flat Polar Quartic Projection

FAO-GIS, February 1998

II.2.4 The arid and semi-arid environment

Arid and semi-arid environments are climatologically determined by having an annual evapotranspiration exceeding the annual precipitation. There are several ways to define and classify the arid and semi-arid zones, two of them mentioned in these lecture notes, viz. the ***bio-climatic classification of UNESCO*** and the ***agro-ecological zonation (AEZ)*** of FAO. The bio-climatic classification is described below, for the agro-ecological zonation the reader is referred to the chapter on Land Evaluation in Part III.

II.2.4.1 The bio-climatic classification of UNESCO

The bio-climatic classification developed by UNESCO makes use of the ration between the mean annual precipitation (P) and the mean annual evapotranspiration (ETP). The following four zones are defined:

1. The ***hyper-arid*** zone with a P/ETP ratio of less than 0.03. It is less determined by climatic criteria rather than by the absence of vegetation except for ephemera's and shrubs in riverbeds. This zone corresponds to the extreme desert and is virtually unsettled. Mean average annual rainfall ranges from 0 – 150 mm.
2. The ***arid*** zone with a P/ETP ratio between 0.03 and 0.20. It comprises dryland areas with sparse perennial and annual vegetation. Nomadic pastoralism can be practised here, but rainfed agriculture is not possible. Mean average annual rainfall ranges from 150 – 300 mm.
3. The ***semi-arid*** zone with a P/ETP ratio between 0.20 and 0.50. It includes steppe and tropical shrubland with a discontinuous herbaceous cover and increased frequency of perennials. Livestock breeding and rainfed agriculture (dry farming) are both possible in this zone. Mean average annual rainfall ranges from 300 – 500 mm.
4. The ***sub-humid*** zone with a P/ETP ration between 0.50 and 0.75. It is characterised by a more dense vegetation than in the previous zone. It includes

tropical savannah, Mediterranean maquis and “chaparral“, and steppes. Rainfed agriculture is common for crops adapted to seasonal drought. Mean average annual rainfall ranges from 500 – 800 mm.

II.2.4.2 Arid landforms, weathering and soil formation

Weathering in the arid climates is mainly physical, producing debris, immediately followed by water and wind erosion. It results in typical arid landforms such as planation surfaces (“pediplains”), colluvial deposits (fans, debris cones), deep wide valleys (“wadis”), sediment slopes, and dunes and depressions (“playas”). Consequently, arid soils can be considered as young soils, in which the parent material (debris), climate and topography are the dominating soil forming factors. Less important are the vegetation (because of the low production level of organic matter), time (processes are slow to very slow) and man (unless irrigation is involved). The ***major soil forming processes*** in the arid to sub-humid regions are:

- Salinisation and alkalinization (“sodification”)
- Redistribution of carbonates and gypsum
- Silification (“duripan” development)
- Translocation of clay (mainly in the sub-humid zone)

These processes give rise to a number of soils, viz. ***Calcisols*** (accumulation of calcium carbonate), ***Durisols*** (accumulation of silica), ***Gypsisols*** (accumulation of gypsum) and ***Solonchaks*** (accumulation of salts more soluble than gypsum). Out of these four only the salt-affected soils (Solonchaks) will be discussed here.

II.2.4.3 Land use in arid regions

In the hyper-arid and arid zones (0 – 300 mm of rainfall) only extensive grazing is possible. The semi-arid zone (300 – 500 mm of rainfall) is suitable for grazing and dry farming, while in the sub-humid zone (500 – 800 mm of rainfall, with one or two dry periods) grazing and rainfed farming is possible.

Dry-farming is a crop-fallow rotation, by cultivating the land one year and leaving it fallow for one or two years to conserve moisture. This method is based on two principles:

- Increase in moisture perception
- Reduction of evaporation or evapotranspiration

Both aspects can be positively affected by frequent soil tillage at regular intervals during the fallow year. This helps increasing the infiltration rate during rain and by reducing capillary moisture uptake to the topsoil and hence evaporation. It also helps to control weeds, but this can also be done with weedkillers.

The effect of the fallow year can be expressed as fallow-efficiency, i.e. the percentage of the precipitation in the fallow year that is stored in the soil for the crop year. In zones with an annual precipitation of about 400 mm, fallow-efficiency may range from 20 – 30 %. Fallow-efficiency is strongly related with soil properties like texture, structure, structural stability and infiltration rate, as well as with soil depth, topography, rooting depth and rainfall intensity.

It should be noted that farming may deteriorate topsoil structure. This means that soils with a weak structural stability can better be used for range land rather than for dry-farming, provided that the land will not be overgrazed, and drinking water is available. Thus, dry-farming is a technique which conserves moisture in the soil to enable one crop every two or three years, instead of no crop at all. However, there remains the risk of irregular rainfall (amount and distribution) during the crop year.

Rainfed farming relies directly on the rainfall and is possible in the sub-humid zone with a mean average annual precipitation of 500 – 800 mm. It strongly depends on the distribution and amount (1 or 2 dry seasons may be possible) of the rainfall, and the reliability of the first rains make all the difference as regard success or failure. Under this system, sandy soils are more risky than loamy or clayey soils. A good intake rate (permeable topsoil, no crusting) is of paramount importance.

Irrigated farming removes radically the major constraint for crop growth in the arid regions: lack of moisture. At present about $^1/_6$th of the global arable area is under permanent or seasonal irrigation (also outside the arid zones). The potential irrigable area is estimated to be twice the present area.

Arid zones have much potential if irrigation water is available, because:

- They are relatively flat and level;
- There is an optimal CO_2 assimilation;
- Soils have favourable chemical and physical properties for crops;
- Pests and diseases are few (dryness);
- Land is available.

These are the reasons why irrigated areas in arid zones have expanded enormously. However, many projects have failed because of flooding and/or salinisation and also for health (malaria), social and economic reasons.

Irrigation will change the soil in several respects. The rapid alternation of wetting and drying may result in:

- Increased weathering of the soil minerals;
- Change of soil micro-flora and increase of microbial activity
- Increased supply of organic matter and increased demineralisation;
- decrease of mean annual soil temperature;
- Leaching of salts and minerals, especially Ca and Mg, resulting in a drop of pH
- Accumulation of soluble salts;
- Degradation of soil structure by compaction of the surface (irrigation crust) or deeper (irrigation layer);
- Reduction of the surface soil;
- Erosion.

Salinity and alkalinity (sodicity)

Saline soils occur in arid zones and in marine lowlands and deltas. The salts in these soils originate from rocks or geological salt deposits, or are indirectly derived from the sea or saline groundwater. Salts accumulate in the soil and at the soil surface because of strong evaporation (in arid zones between 1500 and 3000 mm per year) of groundwater and also of evaporating water in depressions.

Salts are harmful to plants and crops because:

- Osmotic pressure of the soil solution becomes too high, so the water uptake is hampered or even prevented;
- Sodium salts cause deterioration of the soil structure, hampering the root penetration and aeration;
- Certain sodium salts like soda and also borates are toxic to plants.

In general, the more soluble salts are, the more harmful they are for the plants.

Common salts in salt-affected soils

The most common salts in soils from arid and semi-arid regions are calcium carbonate, magnesium carbonate, sodium carbonate (soda), calcium sulphate (gypsum), magnesium sulphate, sodium sulphate, magnesium chloride, sodium chloride, potassium chloride, nitric acid salts and boric acid salts.

Calcium carbonate ($CaCO_3$) is a salt with extreme low solubility ($0.012\ g\ l^{-1}$). In the presence of carbonic acid, as a result of the formation of calcium bicarbonates ($CaCO_3 + H_2CO_3 \Leftrightarrow Ca(HCO_3)_2$), the solubility of calcium carbonates increases to $0.06 - 0.14\ g\ l^{-1}$. In view of its low solubility, the presence of calcium carbonate in soils is not harmful for the majority of agricultural crops, although its occurrence may cause an imbalance in the ions of the absorption complex.

Magnesium carbonate ($MgCO_3$) has a much greater solubility than calcium carbonate. In the presence of carbonic acid, its solubility increases sharply, and its solution is highly alkaline (pH about 10). This alkalinity may cause stress in plants. However, the accumulation of magnesium carbonate in soils in free form is rare, due to the absorption of Mg by clays and due to the formation in the arid zone of dolomite ($CaMg(CO_3)_2$, an insoluble compound.

Sodium carbonate is commonly found in nature and it sometimes accumulates in soils in considerable quantities. In soils it crystallises with varying quantities of water ($Na_2CO_3.10H_2O$). Sodium carbonate is highly soluble (178 g l^{-1} at 20°C). As a result of hydrolysis, it provokes acute alkalinization of the soil, up to pH 12. It is therefore extremely toxic for the majority of plants. Moreover, the presence of Na_2CO_3 in soil solutions causes peptisation of soil colloids, collapse of soil aggregates and hence low water permeability, even at quantities of 0.05 to 0.1%.

Bicarbonate of sodium ($NaHCO_3$) is both less alkaline and less toxic than normal soda. This is due to the fact, that the sodium bicarbonate is partially neutralised by carbonic acid. The formation of $NaHCO_3$ is increased in conditions of decomposition of organic matter and low temperature: $Na_2CO_3 + H_2O + CO_2 \Leftrightarrow 2NaHCO_3$.

It is comparatively rare that considerable quantities of soda accumulate in soils, due to the fact that in most arid soils an accumulation of gypsum ($CaSO_4.2H_2O$) is present, in the presence of which sodium turns into calcium carbonate ($Na_2CO_3 + CaSO_4 \Leftrightarrow CaCO_3 + Na_2SO_4$). Hence, it is only in gypsum-free sediments that soda and sodium bicarbonate can accumulate in significant amounts.

Calcium sulphate ($CaSO_4.2H_2O$) has a low solubility (1.9 g l^{-1}) and is physiologically harmless to plants. But as gypsum may accumulate in arid soils in large quantities (up to 50 – 90%), causing cementation of the entire horizon, plant growth can be hampered because of physical reasons.

Magnesium sulphate ($MgSO_4$) is a typical component of saline soils, accumulating in quantities of several percent. Owing to its high solubility (262 g l^{-1}), it is one of the most harmful salts to plants. $MgSO_4.7H_2O$ never accumulates in the soil in pure form, but always in combination with other easily soluble salts.

Sodium sulphate (Na_2SO_4) is also a typical component of saline soils, saline water and saline mud. Its toxicity is less than that of magnesium sulphate. Its solubility depends very much on temperature, increasing considerably when temperature rises. In the warm seasons of the year, sodium sulphate rise to the soil surface together with other soluble salts (magnesium sulphate, magnesium and sodium chlorides). In the cold season, on the other hand, sodium sulphate is hardly leached. Tenardite (Na_2SO_4) often accumulates in the warm season in the topsoil of saline soils together with gypsum, forming a powdery, “fluffy“ salty topsoil, containing up to 30% sodium sulphate. Reclamation of such soils should be carried out by leaching in the warm season, preferably in combination with irrigated rice.

Magnesium chloride ($MgCl_2$) is common in saline soils and groundwater. Owing to its high solubility (353 g l^{-1}) it is exceptionally harmful to plants. It is an extremely hygroscopic salt (like the not common calcium chloride $CaCl_2$), which absorbs moisture from the atmosphere at night. Saline soils, which contain MgC12 and CaC12, remain humid and sticky a long time after dew or rain. Reclamation by leaching is very difficult.

Sodium chloride (NaCl) is (together with Na_2SO_4 and $MgSO_4$) the most common and widespread component of saline soils. Its solubility is high (264 g l^{-1}) and therefore also its harmfulness. Plants start suffering with a content of 0.1% NaCl and many saline soils containing 2.5% NaCl are totally barren. Reclamation by leaching is very easy if the soil also contains gypsum. If no gypsum is present, leaching is difficult, because sodium assumes an exchangeable form, the alkalinity is high and the surface soil becomes peptised.

Potassium chloride (KCl) has chemical properties similar to NaCl. However, as a rule its content in saline soils is not high, since potassium is consumed by organisms and is subject to adsorption by clays.

Nitric acid salts (nitrates) do not accumulate in large quantities in soils. In hyper-arid deserts sodium and potassium nitrates cause acute salinisation in the same way as sodium and potassium chlorides. Nitrates in soils of over 0.07 – 0.1% NO_3 are much more harmful to crops than chlorides. However, nitrates can be easily washed out of the soil.

Boric acid salts (borates) are very rare and can be found in the proximity of volcanoes. Borates are extremely toxic even in small quantities.

Estimation of salinity and alkalinity

Salinity is estimated by measuring the electrical conductivity of a standardised saturated soil extract (EC_e, expressed as mS cm^{-1}). Alkalinity (sodicity) is estimated by measuring the concentration of Na^+, Ca^{2+} and Mg^{2+} and calculate the sodium absorption ratio (SAR):

$$SAR = \frac{Na^+}{\frac{1}{2}\,(Ca^{2+} + Mg^{2+})}$$

The sodium adsorption ratio and the electrical conductivity are related and used to characterise the soil:

Saline soil:	$EC_e > 4$	ESP < 15	pH > 8.5
Alkali soil:	$EC_e < 4$	ESP > 15	pH > 8.5
Saline alkali soil:	$EC_e > 4$	ESP > 15	pH > 8.5

These criteria are meaningful because plants start suffering at EC_e of 4 and at an ESP > 15.

Soil profile development

Salinisation and alkalinization are major soil forming processes in arid zones. Salinisation occurs as result of uptake of saline groundwater, after which a profile forms with salt accumulation at the surface (crust or fluffy layer), and often a saline horizon ("salic horizon") in the subsoil. Soil structure is loose and/or granular. These soils are called ***Solonchaks***. Alkalinization or sodification is the formation of soils with a high percentage of exchangeable sodium. Clay particles and humus become dispersed, resulting in sodium-rich clay accumulation in subsurface horizons with a prismatic or columnar structure ("natric horizon"). These soils are called ***Solonetzes*** (see section II.2.3.7). Nitrification is the oxidation of ammonia into nitrate, which may accumulate. Nitrate salinisation is rare but may occur in arid regions.

Reclamation (rehabilitation) and management

Distinction has to be made between non-saline arid soils and salt-affected arid soils. No major difficulties can be foreseen for rainfed farming on non-saline arid soils, except when strongly expressed calcic or petrocalcic horizons, or a duripan occur. Large areas of flat steppe and savannah soil have been (and still are) plowed for grain cultivation. However, yields are low and there is danger of water- and wind erosion, structure deterioration and depletion of nutrients ("soil mining").

Salt-affected arid soils can only be used after reclamation. Irrigation is a suitable tool to achieve this. Leaching the soluble salts into the subsoil is most effective, while increasing the permeability will help leaching. Several methods of reclamation are used:

- Vegetative method: rice followed by alfalfa followed by grain or other crops. This may take several years.
- Mechanical methods: deep ploughing, subsoiling, sanding and profile inversion are the most common ones.
- Chemical methods: applications to neutralise the soil reaction, to react on free soda, and to replace exchangeable sodium by calcium. Amendments to achieve this are soluble $CaCl_2$ and $CaSO_4$, slowly soluble limestone ($CaCO_3$), and application of acidifying compounds such as sulphuric acid, sulphur and iron sulphate in order to neutralise soda and to react with lime to form gypsum.

To prevent deterioration of irrigated areas, it is of major importance to leach and drain without over-irrigation, which may lead to rising of the groundwater table and subsequent re-salinisation. If the subsoil is permeable, no special drainage is required. If not, deep and effective drains are necessary. Drain spacing has to be adapted to the vertical and horizontal permeability of the soil, while careful levelling is also important.

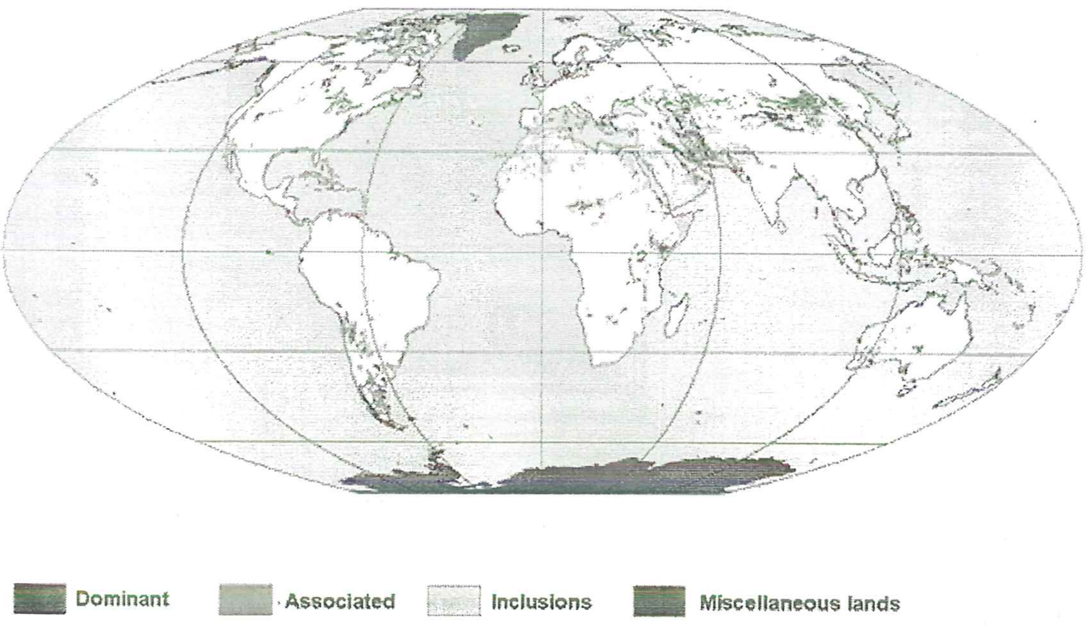
Summarising, saline/alkali soil quality management requires:

- Well-planned irrigation layout, based on survey;
- Careful levelling;
- Selected crops and crop rotation;
- Adequate irrigation and drainage system;
- Very skilful irrigation management;
- Effective advisory service to farmers;
- Maintenance of irrigation and drainage channels;
- Reliable water supply from the main channels.

Regional distribution of Solonchacks

Solonchaks are widespread though scattered in Saharan Africa, East Africa, Namibia, Central Asia, Australia and South America. Estimates of their area vary between 260 and 340M ha.

Distribution of SOLONCHAKS
Based on WRB and the FAO/Unesco Soil Map of the World



Flat Polar Quartic Projection

FAO-GIS, February 1998

PART THREE:

LAND EVALUATION
AND
LAND RESOURCES
INFORMATION SYSTEMS

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TABLE OF CONTENTS

1. INTRODUCTION

Man and the land...

From the earliest times man assesses the environment for his needs. Starting out as hunters, and later becoming sedentary farmers in small communities, most societies were able to live in balance with nature. With increasing population, the impact of man on the environment increased, however this was a gradual process in most places and social groups were able to develop often complex systems for exploiting natural resources on a sustainable basis (FAO, 1980). Today population pressure is high in many parts of the world and increases at an alarming rate. In an increasing number of developing countries the food security situation is deteriorating. The necessary intensification of existing land-use systems to meet future food demands requires inputs that are not normally available. Consequently, farmers are opening new land for agriculture. However, the prospect of expansion of agricultural land as a way to increase food production is bleak. Unused land areas tend to be only marginally suitable for arable cropping (Driessen, 1997), and farmers are using technologies –suited to the high potential regions- that are not adapted to ecological fragile areas (Ndegwa, 1997). At the same time changing socio-economic conditions have undermined traditional systems of land resources management (FAO, 1980), with degradation of the land resources as a severe consequence (See Box 1).

A troubled partnership...

Box 1: GLASOD – Global Assessment of Human Induced Soil Degradation

Human-induced soil degradation by water erosion is one of the most destructive and certainly most extensive phenomena world-wide, and is fast becoming recognized as a key issue in affecting global food security e.g. (Barrow, 1991). Of the total land surface of 13,069 x 10^6 ha about 1,094 x 10^6 ha or 8% is affected by some form of water erosion. When taken as a proportion of the used land (Oldeman, 1994) it increases towards 24%. The major cause for soil degradation is deforestation and removal of natural vegetation (43%), 29% is due to overgrazing, 24% is caused by improper management of the agricultural land, and 4% is a result of over-exploitation of the natural vegetation (Oldeman, 1991).

Methods for evaluation of land

In order to plan sustainable land management effectively we need information on what uses are possible in an area, and what practice is required that gives a high profitability, with a low risk, and improves –or at least does not degrade- environmental quality. This involves and analysis of *constraints* (limitations for given land uses) and an analysis of *potential* (management intervention required to optimize production), (DHV-Consultants, 1996). Land evaluation is often part of an overall land use planning. Land may be evaluated for different objectives: assessment of suitability for crops, suitability for urban development, suitability for nature conservation, land taxation, possibilities for land reclamation, etc. Farming systems analysis is focussed on the development of agricultural technology for small farmers. On a global scale many different local evaluation procedures, each with their own terminology and definitions, are/have been in use. It was only since the early seventies that a theoretical basis was provided for land evaluation. The Framework for Land Evaluation (FAO, 1976) has become an international standard for the appraisal of land. The Agro Ecological Zones Project (FAO, 1978) aimed to provide information on the adequacy of land resources to feed present and future populations. A procedure was developed by which small-scale land suitability and potential could be assessed. In spite of its limitations, the AEZ study is a milestone in the history of land evaluation. It introduced a promising new approach to land suitability assessment and sparked the development of quantified methods of land- use systems analysis (Driessen and Konijn, 1992). Increasing computational capacity facilitated the development and application of crop growth simulation models of increasing analytical complexity over the past decades. These models allow estimates of growth and prediction of the main agricultural crops under a wide range of environmental conditions (De Wit and Van Keulen, 1987). The dynamic nature of agricultural systems is accounted for and such quantified procedures find wide application as part of decisions support systems for policy support and land use planning.

2. THE FRAMEWORK FOR LAND EVALUATION

The Framework for land evaluation is a set of methodological guidelines, rather than a classification system, which is intended to be applicable to any land evaluation project in any environmental situation and at any scale. It is an ecological analysis whereby land mapping units are evaluated with reference to defined land utilization types which also incorporate social, economic, and technological descriptions (Davidson, 1992).

2.1 PRINCIPLES OF THE FAO FRAMEWORK

The approach to land evaluation in the Framework is based on a set of fundamental principles. These are as follows (FAO, 1976):

i. *Land suitability is assessed and classified with respect to specified kinds of use.* This principle recognizes that different kinds of land use have different requirements. Thus an area highly suitable for one type of land use may be unsuitable for another.

ii. *Evaluation requires a comparison of the benefits obtained and the inputs needed on different types of land.*
This involves a comparison of the production costs with the returns of the different land uses on an economic basis.

iii. *A multidisciplinary approach is required*
The evaluation process requires contributions from the fields of natural science, the technology of land use under consideration, economics and sociology. In quantitative evaluation the comparison of benefits and inputs in economic terms plays a major part in the determination of suitability.

iv. *Evaluation is made in terms relevant to the physical, economic and social context of the area*
The assumptions underlying evaluation will differ from one country to another and, to some extent, between different areas of the same country. Factors such as availability and cost of labour, distance to local or export markets must be taken into account.

v. *Suitability refers to use on a sustained basis*
The aspect of environmental degradation is taken into account when assessing suitability. And a proposed land use should not result in degradation such that it affects the functions and qualities of land.

vi. *Evaluation involves comparison of more than a single kind of use*
Evaluation is only reliable if benefits and inputs from any given kind of use can be compared with at least one, and usually several different, alternatives. If only one use is considered there is the danger that, whilst the land may indeed be suitable for that use, some other and more beneficial use may be ignored.

2.2 LAND EVALUATION CONCEPTS AND DEFINITIONS

Different kinds of land are unequally suited to various uses (Rossiter, 1990). Land evaluation is the assessment of the suitability of a tract of land for a specified kind of use; it provides objective sets of data on potentials and constraints, which can contribute to decisions on a sustainable land use (Van Lanen, 1991). In practice this involves the comparison ("matching") of the requirements of a specified land use with the properties of the land.

Land
The spatial entities that are evaluated are land units (LU), or Land Mapping Units (LMU). A LMU comprises an area on a map which is relatively homogenous in terms of soil, climate, topography, hydrology. A land (mapping) unit need not to be uniform in all aspects. Relevant is whether the variation that occurs affects the functioning of the land under the intended use; therefore the concept of 'land unit' is used for areas that can be considered uniform in view of the requirements of the defined (actual) or intended land use (Driessen and Konijn, 1992). Soil is but one aspect of land, alongside of terrain, climate, vegetation, hydrology, infrastructure, etc. and the socio-economic context within which a land unit is used.

Land Characteristics
A land unit is described by its major Land Characteristics (LCs). They can either be single or compound. Single land characteristics are properties of the land that can be measured or estimated, e.g. annual rainfall, dominant slope, soil drainage class and soil depth. Compound land characteristics are composed of associated single characteristics. 'Available water capacity' (AWC) is an example of a compound land characteristic, as it is a function of soil depth and matrix geometry.

Land Qualities
A Land Quality (LQ) is a set of interacting land characteristics which acts in a distinct manner in its influence on the suitability of land for a specified use. Examples of land qualities are the 'water availability to a crop' (a.o. influenced by AWC, rainfall, soil depth, hydraulic conductivity), 'availability of nutrients' and 'resistance to erosion'. See Table 1 for a list of land qualities that can be used for physical land evaluation for rainfed crops.

Land use
The Framework for Land Evaluation uses the concept of 'major kinds of land use' (e.g. 'deciduous forest', 'annual crops') and the more specific **Land Utilization Type (LUT).**

Land Utilization Type (LUT)
A Land Utilisation Type (LUT) is a kind of land use for which a set of technical specifications are described in a given physical, economical and social setting. The LUT is characterized by its **'key attributes'**. An example of a broadly defined LUT is: ‘low input, rainfed maize cultivation (single crop per year) with low technology’.

Key attributes
Key attributes are the biological, socio-economic and technical aspects of land use that are relevant to the productive capacity of a land unit (LU). Crop selection, labour intensity and management level are examples of key attributes.

Land use requirements
Land utilization types are characterized by a set of Land Use Requirements (LURs), which are 'the conditions of land necessary for the successful and sustained practice of a given LUT', (FAO, 1984). Where a land utilization type concerns the growth of a crop/variety, land use requirements are expressed mainly as crop requirements. Land use requirements express the demand of a land use, whereas land qualities express the supply, i.e. properties of a particular tract of land.

Matching
Matching is the comparison of land use requirements with land qualities of specified land units. The sufficiency of a land quality, which is the degree to which a requirement of land use is satisfied by a corresponding land quality, is expressed in a rating. In a broader sense, matching refers to the process of mutual adaption and adjustment of (descriptions of) land utilization types and land units, in order to find the best combination of (improved) land use and (improved) land qualities (FAO, 1983; Batjes, 1990). See Table 2 for an example a rating table for a single land quality.

Table 1. Evaluation of land qualities – example of evaluation sheet

vh = very high n = not present
h = high w = weak
m = moderate m = moderate
l = low s = serious
vl = very low vs = very serious

LMU JM 4
⇓

LAND QUALITY	Availability (1)	vh	h	m	l	vl
		n	w	m	s	vs
CLIMATE						
Radiation regime	1					
Temperature regime	1		▨			
Climatic hazards (hailstorm, wind, frost)	2					
Length growing season	1		▨			
Drought hazard during growing season	2					
SOIL						
Potential total soil moisture	1				▨	
Oxygen availability	1			▨		
Nutrient availability	1				▨	
Nutrient retention capacity	1			▨		
Rooting conditions	1				▨	
Conditions affecting germination	1		▨			
Excess of salts - salinity	2	▨				
- sodicity	2	▨				
Soil toxicities (e.g. high Al sat.)	2					▨
LAND MANAGEMENT						
Initial land preparation	1		▨			
Workability	1		▨			
Potential for mechanization	1		▨			
Accessibility - existing	1	▨				
- potential	1	▨				
Erosion hazard - wind		▨				
- water	2		▨			
Flooding hazard	2	▨				
Pest and diseases	2	▨				

COMMENTS Limitations of JM 04 for subsistence rainfed cultivation of maize, with low technology and capital intensity.

Table 2. Rating table for assessing land quality 'potential for mechanisation'.

Rating	Stoniness/rocks	Gravel content of profile	Slope	(effective) soil depth
	% surface cover	vol%	%	(cm)
1	< 0.1	< 2	< 2	> 60
2	0.1 - 3	2 - 15	2 - 6	> 60
3	3 - 15	15 - 50	6 - 13	30 - 60
4	15 - 90	50 - 90	13 - 25	15 - 30
5	> 90	> 90	> 25	< 15

Land use system
A land use systems is the combination of one land unit and one land utilization type. In general the land use system is analysed for land suitability. *Multiple land use systems* refer to the situation where there is more than one crop on the field at one moment in time. *Compound land use systems* refer to single or multiple land use systems in rotation.

Farming systems
Consist of one or more land use systems practised by one household or management unit.

Final suitability rating
In order to arrive at a final suitability rating, the individual ratings of the various land qualities are translated into a (comparative) suitability class of a land unit for the land use under study, using a conversion table (See Table 3). Often a simple limitation method is used in which the suitability class is determined by the highest severity level of one or more land qualities (Law of Liebig). The ratings of the land characteristics and the land qualities are crop specific. Distinction can be made between situations where a suitability class is determined by the severity level of one land quality (e.g. oxygen availability is limiting) or by several land qualities in the same severity level class (oxygen availability, availability of nutrients *and* conditions for germination are limiting).

In the Framework, two classes of land are recognized: *Suitable (S)* and *Not suitable (N)* for the projected use. The suitability class is usually subdivided into three comparative classes. S1 land is highly suitable, having no significant limitations to sustained application of a given use. S2 land is moderately suitable, having limitations that are moderately severe for sustained application of a given use (See Fig. 1). S3 land is marginally suitable, having severe limitations for sustained application of a given use. Class N1 land is currently not suitable with limitations of such severity that preclude successful sustained use of the land in the given manner. Limitations may be surmountable in time but cannot be corrected with existing knowledge at currently acceptable cost. Class N2 land is permanently not suitable, having limitations which are so severe that preclude any possibilities of sustained use of the land in the given manner.

Table 3. Example of a conversion table for physical land suitability rating.

Suitability	Land quality ratings[#]							
Class	Awc	Asal	Asod	Oxy	Ger	Nut	Roo	Wrk
S1 - Highly suitable	1	2	2	2	1	2	1	2
S2 - Mod. suitable	2	3	3	3	2	3	2	3
S3 - Marg. suitable	3	4	4	4	3	4	3	4
N2 - Perm. Not suitable	4	5	5	5	4	4	4	5

[#] = Example of rating specifications for a single land quality, see Table 1. Source: (Muchena, 1987).
Awc=available water capacity; Asal=absence of salinity; ASod=absence of sodicity; Oxy=availability of oxygen; Ger=conditions for germination; Nut=Nutrient availability; Roo=foothold for roots; Wrk=Workability.

Quantitative definition of these classes is normally unnecessary, since by definition both are uneconomic for the given use. The upper limit of Class N1 is already defined by the lower limit of the lowest suitable class in Order S. The boundary of Class N2, Permanently Not Suitable, is normally physical and permanent. In contrast, the boundary between the two orders, Suitable and Not Suitable is likely to be variable over time through changes in the economic and social context (FAO, 1976).

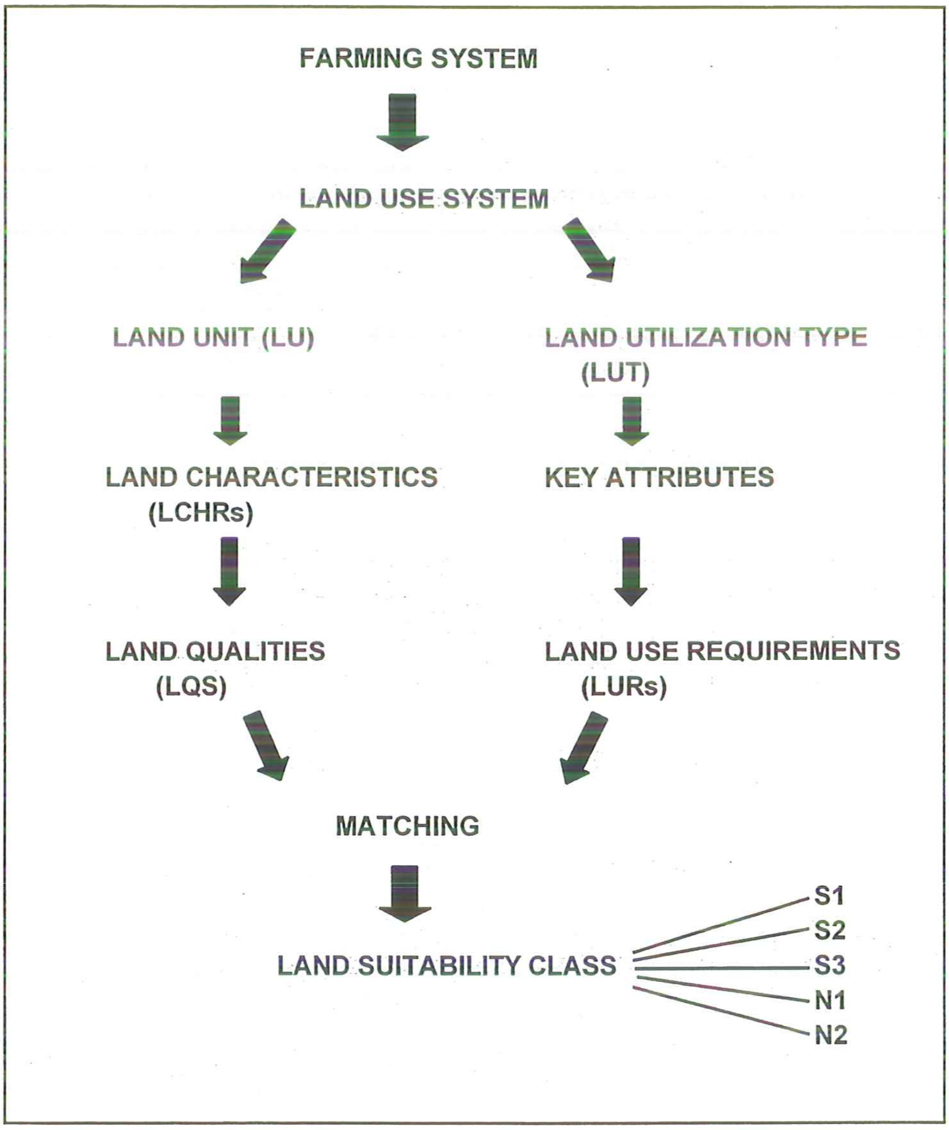


Fig. 1. Units and structure of a Framework evaluation.

A qualitative, physical land evaluation results in *relative* suitabilities of each land area for the set of defined land utilization types, based on the evaluator's judgement. This is useful within a single land utilization type. For determining the *best use* of a tract of land, i.e. comparison of land utilization types, more information should be included in the evaluation. A qualitative, physical land evaluation results in *relative* suitabilities of each land area for the set of defined land utilization types, based on the evaluator's judgement. This is useful within a *single land utilization type*. For determining the *best use* of a tract of land, i.e. comparison of land utilization types, more information should be included in the evaluation. In a quantitative classification, both inputs and benefits must be expressed in common measurable terms, normally economic (FAO, 1976). Differences in degrees of suitability are determined mainly by the relationship between benefits and inputs. It should be expected that boundaries between suitability classes need review and revision with time in the light of technical developments and economic and social changes.

The advantage of economic evaluation is that it allows for comparison between land utilization types. It is possible for instance, that a land unit is rated as 'highly suitable' for a specific land utilization type, and only 'marginally' suitable for another, but that the second LUT is preferred by farmers, because of economic considerations (more profitable). On the other hand farmers do not only make market oriented, economic considerations. Often considerations of risk avoidance (food security) have a higher priority to a farmer than increasing yields/economic return. Social factors should therefore be considered as well (the availability of labour, knowledge of farmers, etc.).

2.3 FARMING SYSTEMS ANALYSIS

The following paragraph on farming systems analysis is based on (Fresco *et al.*, 1990). Farming systems analysis (FSA) has emerged in response to the growing concern over the increasing gap between the yields obtained on experimental fields and actual farmer fields. In land evaluation the potentials and constraints of current and alternative land uses is calculated, whereas farming systems analysis is focussed on the development of agricultural technology for small farmers. FSA recognizes that farmers have multiple goals and it focuses on increasing long-term stability of yield and reduce risks, for example through diversification of crops.

The analytical procedure for farming systems analysis comprises the following steps:

-description of the **physical, biological and socio-economic environment** within which farmers operate.
-to understand the **skills and knowledge, aspirations** of farm households.
-to **evaluate existing systems**, i.e. their performance of processing inputs (labour, seeds, fertilizer, management, etc.) into outputs (crop and livestock products for cash, food, fiber, fuel, etc.).
-to **identify the most constraining factors** that research should focus on.

Farming systems analysis concentrates on a given area and analyses the problems faced by the farmers in that area. It identifies the target groups composed of farmers operating in about the same environment. This implies that farmers are part of similar systems at different levels of hierarchy: similar conditions at regional, village, farming and cropping system levels.

The scope of farming systems analysis is both narrower and wider than that of land evaluation. Farming systems analysis focuses on determining present uses of land, while general land evaluation emphasizes future and potential land uses. Farming systems analysis is more focussed on the management or household unit and is therefore mainly concerned with optimization of present land uses. Land evaluation evaluates land mapping units with socio-economic and technical specifications and scenario's and is more oriented on land use potential and management options for (alternative) and potential land uses. This, however, is more a reflection of past practices of both approaches rather than a methodological necessity.

2.4 LAND EVALUATION AS PART OF THE LAND USE PLANNING PROCESS

The land use planning process, of which land evaluation is a contributing component, is characterised by the following generalized sequence of activities and decisions (FAO, 1976):

i. recognition of a need for change;

ii. identification of aims;

iii. formulation of proposals, involving alternative forms of land use, and recognition of their main requirements;

iv. recognition and delineation of the different types of land present in the area;

v. comparison and evaluation of each type of land for the different uses;

vi. selection of a preferred use for each type of land;

vii. project design, or other detailed analysis of a selected set of alternatives for distinct parts of the area; This, in certain cases, may take the form of a feasibility study.

viii. decision to implement;

ix. implementation;

x. monitoring of the operation.

Land evaluation plays a major part in stages iii, iv and v of the above sequence, and contributes information to the subsequent activities. Thus land evaluation is preceded by the recognition of the need for some change in the use to which land is put; this may be the development of new productive uses, such as agricultural development schemes or forestry plantations, or the provision of services, such as the designation of a national park or recreational area.

2.5 TWO-STAGE AND PARALLEL APPROACHES TO LAND EVALUATION

The land use planning process requires assessment and evaluation of natural resources, economic analysis and social appraisal. In an integrated approach, the natural resources assessment and the socio-economic analyses are performed simultaneously. In a two-stage approach, the biophysical land evaluation precedes the socio-economic analysis (See Figure 2). Biophysical potentials and constraints are then calculated first for a range of possible land use systems. This information is then used to assess which, or under what conditions, land use alternatives are socially acceptable and economically viable. It is evident that socio-economic aspects are taken into account in farming systems analysis and land use planning. A two-stage analysis is often followed, as the temporal and spatial resolution of natural resources and socio-economic data often differs widely. E.g. economic information is often only available averaged at the district or county level, within which there is much variation in land use (possibilities). Interactive multiple goal 'linear programming' models have been developed for the purpose of analyses where trade-offs between different policy goals and the effects on land use are calculated in scenario's with varying socio-economic constraints or possibilities (e.g Njiti, 1994, Bouman *et al.*, 1999).

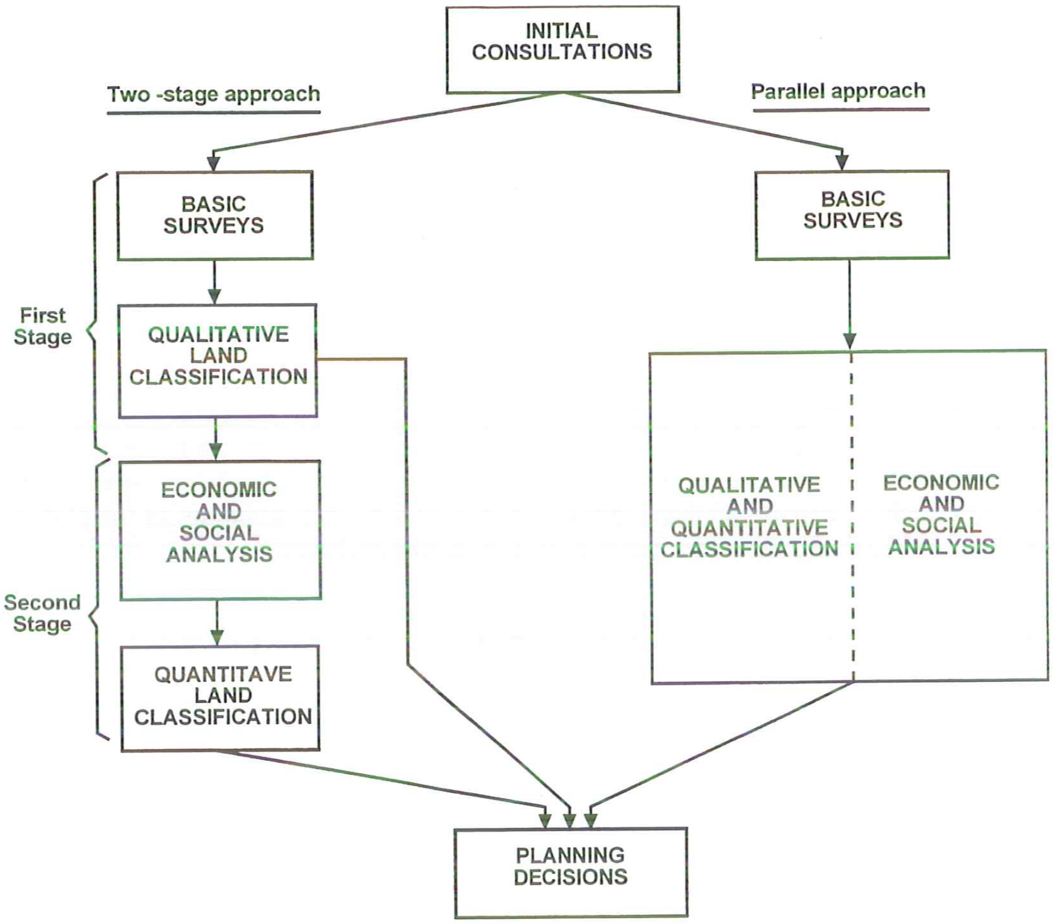


Fig. 2. Two-stage and parallel approach to land evaluation (Source: FAO, 1976).

2.6 SCALES AND LAND EVALUATION

The degree of detail of conclusions that can be derived from a land evaluation study is strongly determined by the level of spatial aggregation of the climate, terrain and soil maps/data, as well as the possible level of integration of the bio-physical and socio-economic information. At regional and national level scales, ‘micro-variations’ in the environmental features are de-emphasized so as to highlight regional trends (Batjes, 1990). The level of detail used for defining land qualities often depends on the amount, accuracy and availability of input data. Furthermore, the technical level of detail when defining land qualities is determined by the types of questions being asked (Bouma, 1989).

The mapping units in larger scale surveys, i.e. reconnaissance (100,000-250,000) and exploratory (1:250,000-1:1,000,000), are rarely homogeneous (Sombroek and Van de Weg, 1980). They consist of more than one component, which cannot be delineated on the map at the map scale. Their nature and properties may be known and quantified in the map legend or the database. Each component of a soil or land unit mapping component is characterized by a 'representative profile', which is described in detail. This 'representative profile' is selected from a number of reference profiles. When interpreting mapping unit (soil) component data for land evaluation the scale should be considered in the level of detail of the evaluation criteria. Some attributes may be representative or indicative, if however the variability of this attribute in the soil component is high, care should be taken to use this attribute (without probability statement) as diagnostic criterion in qualitative interpretations as well as in quantitative studies.

2.7 CONCLUSIONS

One of the major achievements of the FAO Framework is that it provided an internationally applicable standard. The matching of *functionally* relevant properties of land and land use requirements is an improvement over much local evaluation schemes where soils were given a value rating for general agriculture. The decision procedures and evaluation criteria used are made explicit within the Framework. This enables users to trace back what assumptions and rules were used.

The Framework requires specification of the kind of land use in a given physical, economical and social setting ('demand side'). This is described in the same terms as the land related qualities (supply-side), so that demand and supply can be matched to give suitability rating for a specified use. At the same time, the Frameworks' definitions became a source of confusion, liable to multiple interpretations and not fitting in the concept of other disciplines. With increased focus on quantitative and integrated analytical procedures the Framework as a document and, with it as background philosophy, is becoming outdated (Van Diepen *et al.*, 1991). In regional analyses, a qualitative evaluation in accordance with the Framework is often preceding a quantified evaluation in a mixed qualitative/quantitative analysis (Van Lanen *et al.*, 1992). The qualitative analysis is then used as a way to stratify the area for suitability. The area not suitable for the projected use can then be excluded for further quantified analysis. For the suitable areas, limitations relevant for land management can be indicated using the qualitative analysis (e.g. liming requirement). Crop production potentials and yield variation, risks of crop failure and requirements for adapted management (e.g. irrigation and drainage) can then be quantified with simulation models within the suitable areas.

3. QUANTIFIED ASSESSMENT OF LAND RESOURCES

3.1 AGRO ECOLOGICAL ZONING

In the late seventies FAO took the initiative for a global assessment of land resources that would 'assess the potential agricultural use of the world's resources' (FAO, 1978) and would make projections on world food security by the year 2000. For this project a procedure was developed for small-scale land suitability assessment, called 'Agro-Ecological Zoning'. The basic units for evaluation were created by overlay of climatic with soil information. The soil information was derived from the FAO Soil Map of the World (FAO, 1974). With regard to the alternative land uses considered for evaluation a selection was made of the major food and fiber crops. Crops not commonly grown throughout the developing countries, such as barley, oats, groundnuts, rye, peas, chickpeas, grapes and rapeseed, were then disregarded. Sugar cane was not taken into account because its production potential is highly dependent on factory management. This left 11 crops: five grain crops (wheat, rice, maize, pearl millet and sorghum), three root crops (white potato, sweet potato and cassava), two legumes (phaseolus bean and soybean), and one cash crop (cotton). The description of Land Utilization Types was restricted to two levels of management and technology: low and high input. In the AEZ study, low input yields were calculated as one-quarter of those possible with high inputs.

(Driessen and Konijn, 1992) summarized the procedures for agroecological zoning in seven preparatory steps for arriving at land suitability classes, see Figure 3 for relational context:

1) Soil related-requirements of the selected major crops are matched against the properties of all soil units distinguished on the Soil Map of the World. This results in **soil-unit ratings** for rainfed cropping with 'high' and 'low' inputs.
2) **Major climatic divisions** are defined from records of individual climate stations. The temperature specifications of the major climatic divisions are matched against the temperature requirements of the selected major crops to distinguish broad climatic regions that are 'not suitable' (N) for growing a particular major crop. Further analysis is restricted to regions with 'suitable' (S) climate.
3) Precipitation, potential evapotranspiration and temperature data of stations are analysed to determine beginning and end of a possible growing season and the **length of growing period** (LGP). Sites with comparable LGP are aggregated to **LGP zones**.
4) Radiation and temperature data of stations are matched against relevant climate-related crop requirements in a model that calculates **net biomass production** and **constraint-free yield**.
5) The step from potential ('constraint-free') yield to a more practical **anticipated yield** is made by yield deductions for likely **agro-climatic constraints** in a given LGP zone, under consideration of the available technology (i.e. 'input').
6) The anticipated yield is then matched against a reference yield (different for 'high input' and 'low input' cropping). The outcome of this matching is the **agro-climatic suitability**.
7) The soil-unit rating and the agro-climatic suitability determine the (preliminary) land suitability class for high input and low-input rainfed farming. The final land suitability class is obtained by correcting the preliminary land suitability class by a set of **phase, slope** and **texture rules**.

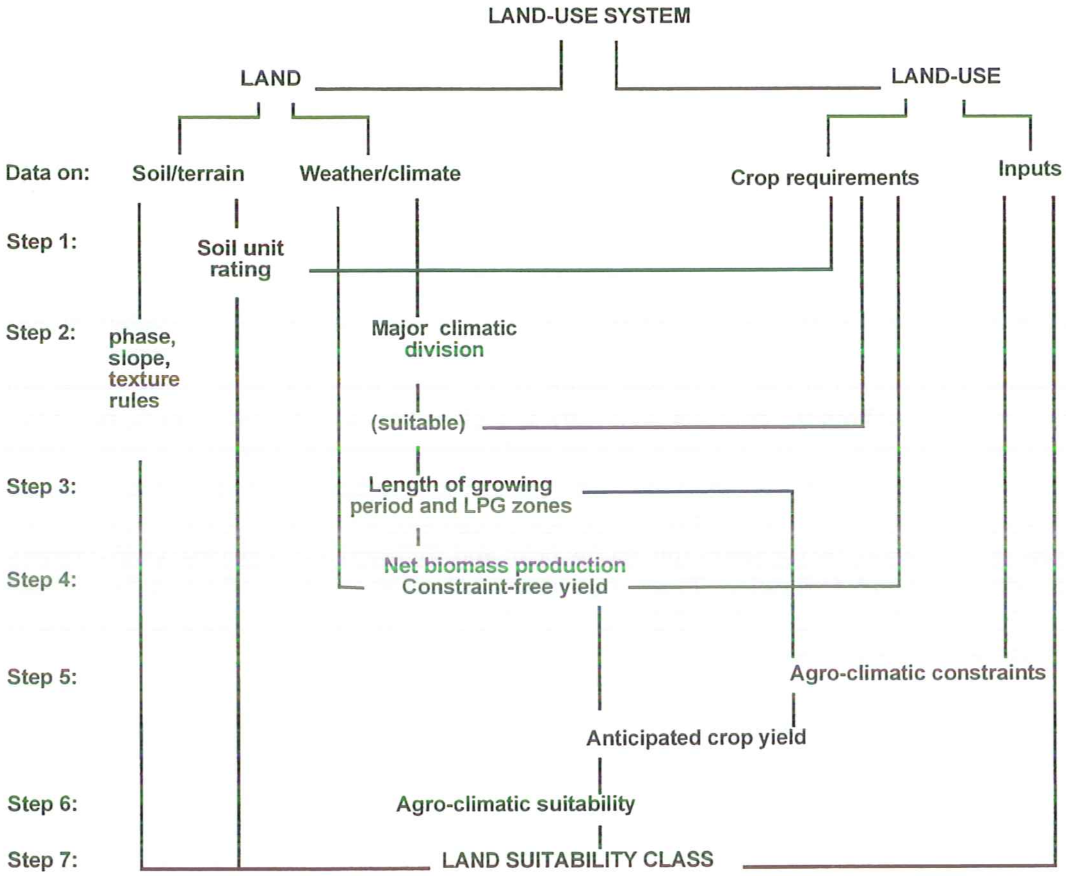


Fig. 2. AEZ matching procedures in their relational context. Source: (Driessen and Konijn, 1992).

The Length of Growing Period is one of the key-concepts in the AEZ-methodology. Potential crop production -or the 'biophysical production ceiling' (Driessen, 1997)- is determined by crop properties, solar radiation and temperature. Sufficiency of water and nutrient availability during the growth cycle, and a range of other factors -such as management, pest and diseases, etc-, determine the actual attainable yield. The AEZ study (FAO, 1978) defined the growing period to include two parts. The beginning of a growing period is defined as the moment when the amount of precipitation equals half the rate of evapotranspiration ($P > 0.5 * PET$) after a dry period (see Figures 4 and 5). When, after that, precipitation exceeds full evapotranspiration ($P > PET$) a humid period occurs, a condition which allows water to build up in the soil. The possible growing period ends when the amount of precipitation is lower than half the rate of evapotranspiration and extends into the dry season until all 'available' stored soil moisture (set at the theoretical amount of 10 cm) has been depleted. Finally, the lengths of growing period are adjusted to allow for periods when mean temperature falls below the minimum needed for crop growth - about 6.5°C. These periods are considered severely restricting for crop growth and are subtracted from the growing period as determined from water availability.

Assessment of the length of growing period is quick, few basic data are needed and a first indication of water availability is obtained. The drawbacks are that soil properties, apart from a static 'available water capacity', are not considered and plant characteristics are not included in the analysis (stored soil moisture is the same for all crops, while plants differ in drought resistance, soil moisture tension at which wilting occurs, etc) and *actual* precipitation is compared with *potential* evapotranspiration.

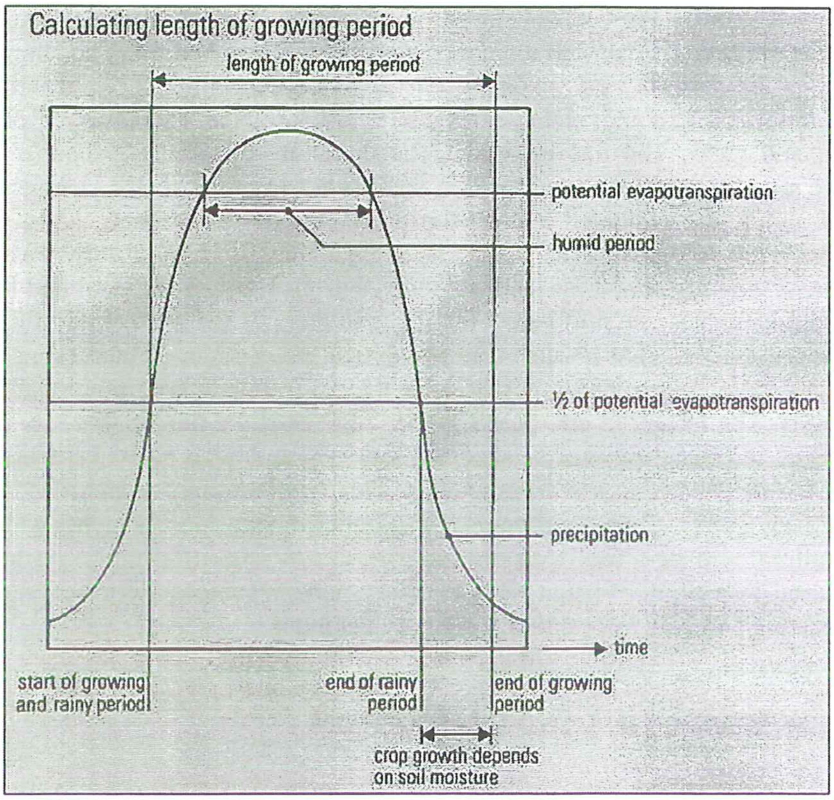


Fig. 4. The Length of Growing Period as determined by the AEZ project Source: FAO, 1991.

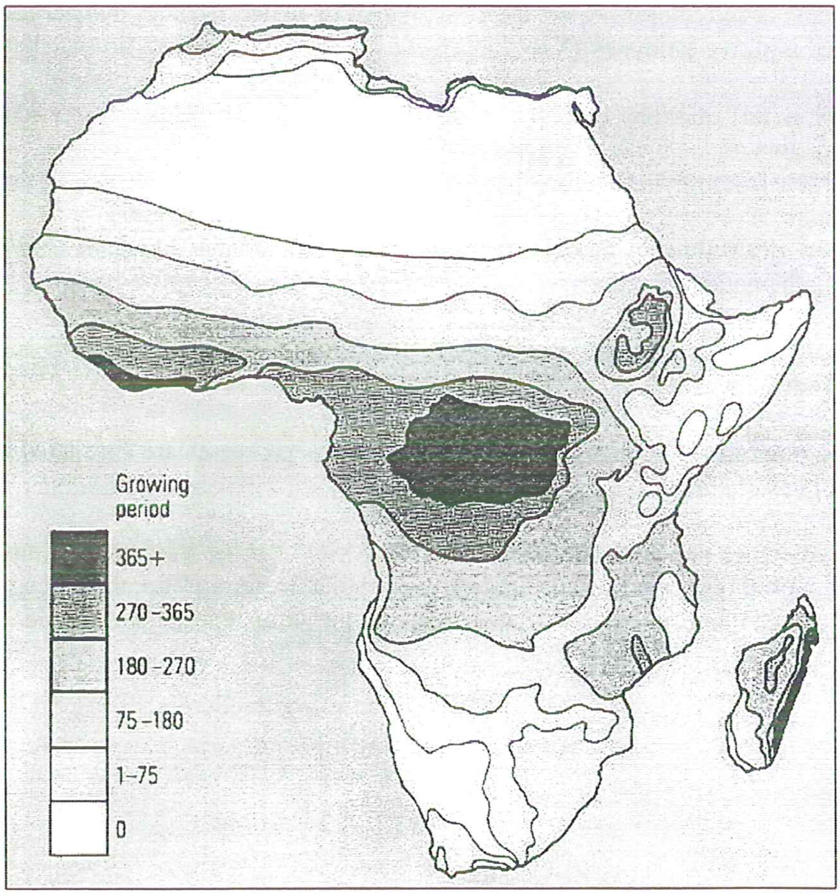


Fig. 5. Length of growing periods for Africa. Source: FAO, 1991.

Conclusions

The agro Ecological Zones Project developed a procedure for small-scale land suitability assessment and was the first serious step towards quantified land evaluation. The simplicity of the approach is both its strength and its weakness (Driessen and Konijn, 1992): the interpretation procedures are formalized and universally applicable, and produce estimates of potential and actual production and yield. The interpretation procedures are simple and require limited and standard data. The simplifications of the approach –that result in rather crude yield estimates- can be criticized, but should be evaluated against the objectives of the project; assessing the land potentials of the world at reconnaissance scale. The accuracy of the methodology was appropriate for the time-frame of the project and the scale of study. For planning at regional and lower levels, more accurate yield estimates are required. This requires making the procedures fully land-use-system-specific and dynamic. That implies using analytical procedures of increasing complexity, with a higher data demand and running costs. Since the AEZ project computational possibilities have evolved rapidly. The advent of computers greatly facilitated the development of crop growth simulation models of an increasing analytical complexity. These models allow estimates of growth and prediction of the main agricultural crops under a wide range of environmental conditions (De Wit and Van Keulen, 1987). Data requirements changed with the increased use of quantitative models for agricultural and land use decisions support. Quantified information on land and soil recourses is required in a digital format. Consequently, land resources information systems are compiled from existing and new information by survey institutes.

3.2 THE ROLE OF MODELS IN LAND EVALUATION

(Mantel and Van Engelen, 1997) review the role of quantified models in land evaluation studies. Models are analytical tools for schematizing a complex reality. They are attractive means for studying processes or systems, especially when the actual processes are expensive or difficult to measure. For estimation of crop performance there are many types of models, differing in level of complexity and data requirements. Apart from qualitative or expert models, which are exclusively based on 'reasoned intuitive estimates' (Van Diepen *et al.*, 1991), we can discriminate between:

1) statistical (empirical) models that relate growth or yield to selected environmental and management factors on the basis of regression analysis,
2) parametric models (multiplying factors that are considered relevant), which are in fact qualitative of nature. Results can be satisfactory when calibrated for specific conditions,
3) mechanistic or deterministic models that describe fundamental physical and/or chemical processes in mathematical functions.

Deterministic models calculate crop performance as a fundamental process over specific temporal intervals. The advantages of such models are (van Diepen *et al.*, 1991):

a) the dynamic nature of land qualities is taken into account as processes are simulated in time steps, depending on rates of change, and

b) as the basic processes are mathematically described, yield can be predicted at any location in principle, although models do need calibration as each parameter and function in a model has its own inaccuracy and the resulting errors accumulate in the simulated final crop yield.

Crop yield assessments for different scenarios

Introduction

The conceptual model for crop growth simulation as developed in Wageningen (De Wit and Van Keulen, 1987; Van Keulen and Wolf, 1986; Van Diepen *et al*., 1989; Driessen and Konijn, 1992) calculates crop yields under three principal growth constraints. This results in theoretically defined situations that are hierarchically ordered according to increasing analytical complexity. The effect of principal growth constraints are evaluated by making separate calculations of:

1) the constraint-free yield, or potential yield, reflecting the 'bio-physical production ceiling' determined by the crop's genetic potential under ambient radiation and temperature regime,

2) the water-limited yield, additionally reflecting the influence of limited or excessive water supply, and

3) the nutrient-limited yield.

Constraint-free yield

Dry-matter accumulation of the crop is quantified under the prevailing weather conditions, i.e. as a function of radiation, temperature and crop characteristics. Other factors influencing crop growth are considered optimal at this production level. **State variables** characterize the changing situation of the crop and the soil at any moment and are updated after each time-step. Radiation and temperature are examples of forcing variables that determine the rate of the simulated processes. Constants are for instance the carbon dioxide content of the air and soil depth. Yield potential varies with the absorption of light, the fraction of net assimilate production that is to be invested in new plant matter, the losses incurred in maintenance respiration, and the efficiency with which the remaining assimilates are converted to structural plant matter (De Wit and Van Keulen, 1987).

Water-limited yield

Assimilation, i.e. reduction of atmospheric CO_2 to carbohydrates $(CH_2O)_n$, is the fundamental process in plant growth. With the intake of carbon dioxide through the stomatal openings in plant leaves, water is lost through transpiration. When plants are exposed to drought they close their stomata, in order to avoid more water loss, thereby reducing carbon dioxide intake, and consequently assimilation, i.e. the reduction of CO_2 to carbohydrates $(CH_2O)_n$, (and thereby plant growth) is hampered. Water losses from the system are through transpiration and evaporation and supply is through rainfall. The buffering capacity of the soil, which determines how much of the rainfall will be available to the plant, causes the growth rate of plants to depend only indirectly of the forcing variable: rainfall. Soil constants needed to simulate the water fluxes in the soil, e.g. water retention and conductivity characteristics, are not measured on a routine basis in soil surveys and thus often absent in soil survey information. Hydraulic parameters are therefore frequently estimated on the basis standard measured soil survey data using pedo-transfer functions (Bouma and van Lanen, 1987).

Nutrient-limited yield

The availability of nutrients to crops is difficult to assess quantitatively, because key parameters are often very variable in both time and space. Furthermore, measurements of relevant indicators for the nutrient capacity of soils is problematic. Nutrient elements in a soil may occur in many forms, and are, depending on the conditions, more or less available to a crop. This availability does not only depend on soil factors, such as soil temperature, pH and moisture conditions, but also on weather conditions and farm management practices. Moreover, crops differ in their demand for nutrients and some crops are more efficient in extracting elements than others. Existing methodologies for calculating nutrient limited yield are generally empirical or statistical. Deterministic modelling of nutrient limited yield would lead to an elaborate model with massive data needs, which would probably generate results that were exceedingly expensive and not very accurate (Driessen, 1997). Quantification of relationship of local validity between soil test value and yield can be quantified with multiple regression techniques. However, as regression equations for soil test calibration are both site and season specific, one should repeat such work in both space and time in order to represent a range of uncontrollable variables (Sumner, 1987). When the potential production can be predicted with a reasonable accuracy, it is possible to calculate the approximate input of fertilizer required to meet a production target. This production target cannot be higher than the water-limited yield (Driessen and Konijn, 1992).

Example of the 'state-variable approach':

Quantification of the sufficiency of the land quality 'water availability', using a water balance model.

When uptake of water by roots cannot fully replenish transpiration losses, plants actively curb their water consumption (i.e. the rate of transpiration). There is a strong correlation between the rate of transpiration and the rate of assimilation of CO_2 from the atmosphere.

In a dynamic water balance model the values of some variables, e.g. precipitation (P), potential evapotranspiration (PET), actual soil moisture (SMAA) and rooted depth (RD) are considered constant for the duration of one interval. The values of P and PET are fixed for the duration of one time interval, for instance one day, after which new values of P and PET are called from the climate file for calculations for the next time interval. The value of SMAA is used to calculate actual evapotranspiration rate (ETA), after which the value of SMAA is updated by adding the water influx over the rooted depth (RD) and subtracting the (calculated) water losses in the interval. This updated value of SMAA is considered constant over the next interval, and so on. SMAA is a *'state variable'*; it is calculated anew for each interval in the crop cycle and expresses the state of the system during one interval. The sufficiency of water availability is often presented as the ratio of actual transpiration (TR) over transpiration maximum (TRM).

Static vs dynamic water balance model approach

The advantage of using dynamic water balance analysis as a tool for this purpose, over determining a fixed 'Length of Growing Period', is that it is land-use system (crop) specific. In addition to climatic factors, soil attributes, plant specific requirements and interactions are taken into account. With dynamic analysis, the temporal variability of land use requirements and land qualities are taken into account. However, a water balance model requires more data on soil, crop and climate compared to the LGP approach.

To resolve each problem, the appropriate tools should be selected. A quick and first indication of 'availability of moisture' is obtained using the LGP approach. More specific answers are obtained when a quantitative and dynamic simulation water balance model is used, which is more data demanding.

3.3 LAND RESOURCES INFORMATION SYSTEMS

Introduction

The outputs of traditional soil survey are maps and reports on land resources. A soil or land resources survey describes the spatial distribution of specific attributes of soil or land types. Sombroek and Van de Weg (1980) on quality and reliability of soil mapping:

The scale at which the survey determines the measure of detail at which one can represent the soil conditions of an area. There is a strict relation between accuracy, scale, and purpose of soil maps. The purpose determines the accuracy, the accuracy determines the scale, and the scale determines the possible uses of the map (i.e.the purpose). Each scale of mapping has its own recommended number of field observations –soil augerings and pits- per unit of surface area. In the area where soils are more homogeneous, usually the surveyor makes fewer observations.

In modern soil survey, data are stored in digital files, which are linked and connected to a Geographical Information System (GIS). The main product from a land inventory is now an information system. A soil map -or map of another land related theme- is no longer a survey end-product, but merely a theme that can be derived from the database.

Geographical information systems

Geographical information systems find wide application nowadays. Commercial software development and increased computer capacity have stimulated the increased use of GIS. Meijerink *et al.* (1994) summarize the four functions of a GIS: 1) data acquisition and pre-processing, 2) data-management, storage and retrieval, 3) manipulation and analysis, 4) product generation. In a GIS different information of the same area can be combined for analysis by an overlay procedure. Overlay techniques are a tool for modelling spatial data. An overlay of maps with different themes provides a map with new (combined) units showing spatially explicit combination of the single themes. GIS software maybe either 'raster' or 'vector' structured, or may handle both data types. 'Raster structured' means that spatial data are stored and processed as cells in a grid (X, Y, Z- co-ordinates). 'Vector structured' means that data are stored as polygons.

Land resources databases

Soil databases should be organised as a group of relational tables. Each table contains information relevant to an entity, and links to other tables. The coding of data of basic units or of same locations should be consistent. On the basis of the primary key, that is the data element that uniquely identifies the objects record (the site ID for instance), data can be extracted or files can be linked. Usually a code in terrain or soils databases for instance is given to data that reflects the basic mapping unit (an example of a code for land units is: 100/2/1, which means mapping unit 100, terrain component 2 (there may be more at a certain scale), soil component 1 within this terrain unit of the mapping unit). A typical soils database structure is (Meijerink *et al.*, 1994):

1) *Map units*, i.e. what is actually shown on the soil map: **key**: Map unit ID; **data**: component ID and their percentages within the map unit.
2) *Components* of map units: **key**: Component ID; **data**: landscape position within Map unit.
3) *Pedons* (representative soil profiles): **key**: Component ID + Pedon ID; **data**: location, site characteristics, whole-pedon attributes such as rootable depth.
4) *Layers* of pedons: **key**: Pedon ID + Layer ID; **data**: depth from surface, measured properties, morphology.
5) *Polygons* (or, delineations): **key**: Polygon ID (over the whole map); **data**: Map Unit ID, delineation attributes such as area. This is the link between the data and the polygons. It is stored in the GIS linkage file; the Polygon Attribute Table (PAT).

Soil horizon data are stored with data for each layer in a record, with an indication of the depth of each layer. Soil layers are within pedons (1-many relation). See Table 4 for a list of attributes stored in a land resources information system.

Mapping unit differentiating criteria of a land resources information system

The map legend (and database) is created from hierarchical classification of land. Landforms, parent rock, slope and soils are principal criteria to differentiate between units. Successive application of the differentiating criteria leads to definition and characterization of basic map units in terms of its terrain properties, terrain components and soil components (variation of terrain and soil conditions within the mapping unit). The different components of a soils and terrain unit are specified in the database with their proportion within the unit. The land variability is thus included in all calculations and interpretations.

An international methodology: Soil and Terrain Digital Databases (SOTERs)

SOTER is a methodology for design of map legend and database structure that was developed on the initiative of the International Union of Soil Sciences (IUSS). The methodology was designed for use at the national level (1:1M scale), but is in principle scale independent. It has been applied at several levels of detail, from the continental (1:5 M), to the national (1:1 M – 1:250,000) and local level (1:50,000-1:100,000). The SOTER methodology uses an integrated concept of land. Underlying SOTER approach is the identification of areas of land with a distinctive, often repetitive, pattern of landform, lithology, surface form, slope, parent material, and soil. Tracts of land distinguished in this manner are named SOTER units. Each SOTER unit thus represents a unique combination of terrain and soil characteristics (Van Engelen and Wen, 1995). The SOTER methodology is similar to the land systems approach developed in Australia. The major landforms are subdivided by lithology. The thus created map units are called *terrain units*. Within these units variation exists in surface form, slope gradient, soils, etc. (see Figure 6). There is a minimum size of units on the map at the given scale, and therefore not all variation that is characterized in the field, may be represented on the final map. To include the variation in the analyses, different components are described for terrain within a mapping or terrain unit. Its properties and proportions within the unit can be specified and the information in the database can be used for analysis. The proportions of terrain and soil components within the terrain unit count up to 100%. For each terrain component a soil type is identified. One terrain component may have more than one soil component. The proportions of the soil components are specified relative to the terrain unit. In this way the area can be calculated for each single database theme or model output, by soil component. The intermediate database files are the input files to the different models. In traditional survey such a cluster is known as a soil association or soil complex.

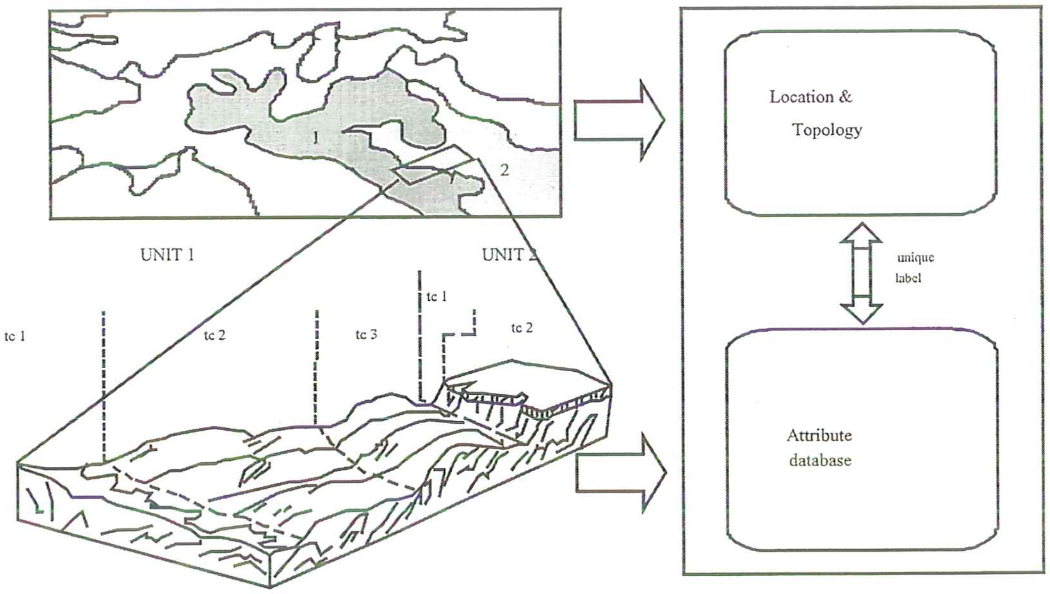


Fig. 6. SOTER units, their terrain components (tc), attributes, and location (Source: Van Engelen and Wen, 1995).

Table 4. Database items for description of a soil and terrain mapping unit.

TERRAIN

1 SOTER unit_ID
2 year of data collection
3 map_ID
4 minimum elevation
5 maximum elevation
6 slope gradient
7 relief intensity
8 major landform
9 regional slope
10 hypsometry
11 dissection
12 general lithology
13 permanent water surface

TERRAIN COMPONENT

14 SOTER unit_ID
15 terrain component number
16 proportion of SOTER unit
17 terrain component data_ID

TERRAIN COMPONENT DATA

18 terrain component data_ID
19 dominant slope
20 length of slope
21 form of slope
22 local surface form
23 average height
24 coverage
25 surface lithology
26 texture group non-consolidated parent material
27 depth to bedrock
28 surface drainage
29 depth to groundwater
30 frequency of flooding
31 duration of flooding
32 start of flooding

SOIL COMPONENT

33 SOTER unit_ID
34 terrain component number
35 soil component number
36 proportion of SOTER unit
37 profile_ID
38 number of reference profiles
39 position in terrain component
40 surface rockiness
41 surface stoniness
42 types of erosion/deposition
43 area affected
44 degree of erosion
45 sensitivity to capping
46 rootable depth
47 relation with other soil components

PROFILE

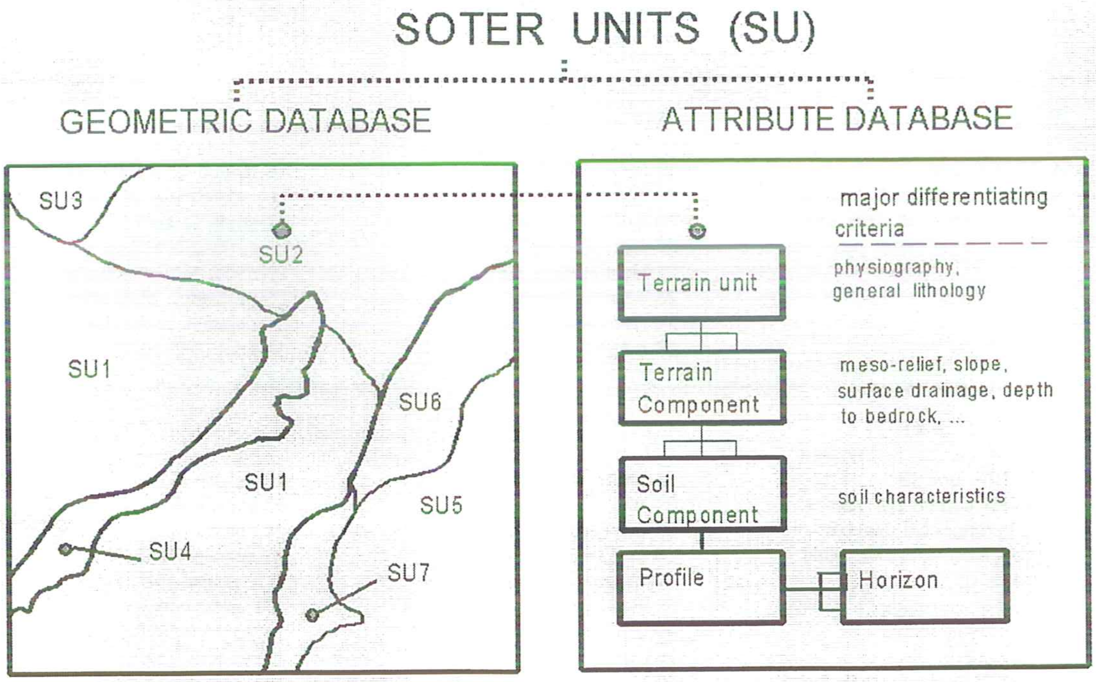
48 profile_ID
49 profile database_ID
50 latitude
51 longitude
52 elevation
53 sampling date
54 lab_ID
55 drainage
56 infiltration rate
57 surface organic matter
58 classification FAO
59 classification version
60 national classification
61 Soil Taxonomy
62 phase

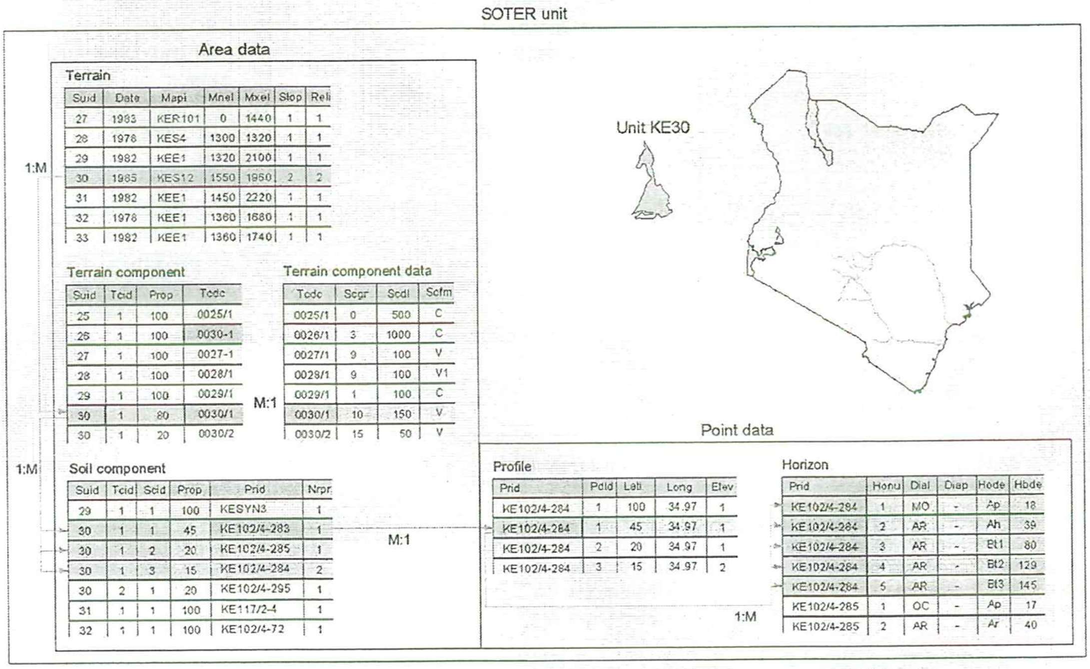
HORIZON (* = mandatory)

63 profile_ID*
64 horizon number*
65 diagnostic horizon*
66 diagnostic property*
67 horizon designation
68 lower depth*
69 distinctness of transition
70 moist colour*
71 dry colour
72 grade of structure
73 size of structure elements
74 type of structure*
75 abundance of coarse fragments*
76 size of coarse fragments
77 very coarse sand
78 coarse sand
79 medium sand
80 fine sand
81 very fine sand
82 total sand*
83 silt*
84 clay*
85 particle size class
86 bulk density*
87 moisture content at various tensions
88 hydraulic conductivity
89 infiltration rate
90 pH H_2O*
91 pH KCl
92 electrical conductivity
93 soluble Na^+
94 soluble Ca^{++}
95 soluble Mg^{++}
96 soluble K^+
97 soluble Cl^-
98 soluble SO_4^-
99 soluble HCO_3^-
100 soluble CO_3^-
101 exchangeable Ca^{++}
102 exchangeable Mg^{++}
103 exchangeable Na^+
104 exchangeable K^+
105 exchangeable Al^{+++}
106 exchangeable acidity
107 CEC soil*
108 total carbonate equivalent
109 gypsum
110 total carbon*
111 total nitrogen
112 P_2O_5
113 phosphate retention
114 Fe dithionite
115 Al dithionite
116 Fe pyrophosphate
117 Al pyrophosphate
118 clay mineralogy

Database design

The soils and terrain information in the database consists of a geometric component, which indicates the location and topology of SOTER units, and of an attribute part that describes the non-spatial SOTER unit characteristics. The geometry is stored in that part of the database that is handled by the Geographic Information System (GIS) software, while the attribute data is stored in a separate set of attribute files, manipulated by a Relational Database Management System (RDBMS). Figure 7 gives a schematic representation of the structure of the attribute database. The blocks represent tables in the database and the solid lines between the blocks indicate the links between the tables.





Terrain

Suid	Date	Mapi	Mnel	Mxel	Slop	Reli
27	1983	KER101	0	1440	1	1
28	1978	KES4	1300	1320	1	1
29	1982	KEE1	1320	2100	1	1
30	1985	KES12	1550	1950	2	2
31	1982	KEE1	1450	2220	1	1
32	1978	KEE1	1360	1680	1	1
33	1982	KEE1	1360	1740	1	1

Terrain component

Suid	Tcid	Prop	Tcdc
25	1	100	0025/1
26	1	100	0030-1
27	1	100	0027-1
28	1	100	0028/1
29	1	100	0029/1
30	1	80	0030/1
30	1	20	0030/2

Terrain component data

Tcdc	Scgr	Scdl	Scfm
0025/1	0	500	C
0026/1	3	1000	C
0027/1	9	100	V
0028/1	9	100	V1
0029/1	1	100	C
0030/1	10	150	V
0030/2	15	50	V

Soil component

Suid	Tcid	Scid	Prop	Prid	Nrpr
29	1	1	100	KESYN3	1
30	1	1	45	KE102/4-283	1
30	1	2	20	KE102/4-285	1
30	1	3	15	KE102/4-284	2
30	2	1	20	KE102/4-295	1
31	1	1	100	KE117/2-4	1
32	1	1	100	KE102/4-72	1

Profile

Prid	Pdid	Lati	Long	Elev
KE102/4-284	1	100	34.97	1
KE102/4-284	1	45	34.97	1
KE102/4-284	2	20	34.97	1
KE102/4-284	3	15	34.97	2

Horizon

Prid	Honu	Dial	Diap	Hode	Hbde
KE102/4-284	1	MO	-	Ap	18
KE102/4-284	2	AR	-	Ah	39
KE102/4-284	3	AR	-	Bt1	80
KE102/4-284	4	AR	-	Bt2	129
KE102/4-284	5	AR	-	Bt3	145
KE102/4-285	1	OC	-	Ap	17
KE102/4-285	2	AR	-	Ar	40

Fig. 7. SOTER attribute database structure with area and point data (1:M = one to many, M:1 = many to one relations).

Data quality

When handling large amounts of data using digital databases, checking of the database integrity and quality is essential. Updating of the database and addition of new data in time may introduce new errors or inconsistencies and especially with the use of automated procedures for extraction and interpretation the database quality and integrity must be relied upon. The database integrity relates for instance to the consistency in coding. Database quality relates to the accuracy and representativity of the attribute values, the degree to which data values were correctly entered, etc. There are many rules that the soil expert uses to assess the quality of soils data from field description and laboratory analysis. Both the 'logic' rules (to check coding f.i.) and the 'expert' rules can be formalized and translated into a computer programme for automated checking of the database. Part of the database checking is related to the quality of data entry and part is merely (cross-)checking of the laboratory data. Several reasons for analytical inaccuracies may be found. (Landon, 1991) summarizes:

1) *pretreatment effects*. Chemical changes in the soil sample as a consequence of the drying, grinding and sieving.
2) *intrinsic errors in measurement*. The accuracy of standard laboratory soil analysis is normally well within the spatial variability of soil properties.
3) *operator errors*. Operator faults, such as the incorrect preparation of standard solutions or incorrect calculations.
4) *contamination effects*. Dust contamination of the sample in the laboratory or during storage.
5) *variation between laboratories*. The results of duplicate sample analyses may often be consistent, but variability between laboratories for the same analytical method are notoriously high. The results of a world-wide laboratory exchange programme, co-ordinated by ISRIC, revealed that variation in laboratory results for the same soil samples after standardization of the methods, was not less than (van Reeuwijk, 1982; van Reeuwijk, 1984):

- ± 11% for clay determination
- ± 20% for CEC measurements
- ± 25% for CEC of the clay fraction
- ± 10% for the base saturation
- ± 0.2 units for the pH measurements

Examples of logic decision rules for verification of soil database quality:

1) The sum of the three texture classes (sand+silt+clay) should add up to 100%, with an allowed deviation to 105% and 95%.
2) pH-H_2O is in principle never smaller than 3 or greater than 8.
3) If pH-$H_2O > 5.4$ then the exchangeable acidity (sum of aluminium and hydrogen) = 0, and reversely, if pH-$H_2O < 5.4$ then the exchangeable acidity > 0.
4) The pH measured in water is in most soils greater than the pH measured in KCl (pH-$H_2O >$ pH-KCl + 0.1).
5) If pH-$H_2O > 6.5$ then the base saturation > 100%, and reversely, if pH-$H_2O < 6.5$ the base saturation < 100%.
6) The organic carbon content (C%) decreases with depth (only in Fluvisols or disturbed profiles this may not be the case), therefore C% should be lower for every consecutive deeper layer.
7) The following data-items can in principle not have the value zero: CEC, C%, pH-H_2O and pH-KCL.
8) The horizon number sequence should be such that it starts with the lowest value (1) and then increases in number, without skipping a value.

4. CASE STUDY: POTENTIAL FOR TRADITIONAL MAIZE GROWING IN KENYA

This case is based on a study by Mantel and Van Engelen (1999).

Introduction
The necessary intensification of existing land-use systems to meet future food demands in Kenya requires inputs that are not normally available. As a consequence, Kenyans are opening new land for agriculture. There is little prospect for expansion of agricultural land as a way to increase food production. Information on land potentials and constraints and required adapted management techniques may help to prioritize land use policies and research and support extension services to select appropriate land management strategies for land use systems.

Land Utilisation Type definition
Maize is the dominant food crop in Kenya. The LUT as broadly defined for this study was: 'low input, rainfed maize cultivation (single crop per year) with low technology'. Farm size is generally small (< 5 ha), cultivation practices include manual tillage ('hoe-farming') and stubble is left for grazing or is incorporated before sowing. No or limited use is made of chemical fertiliser and manure. Sowing dates vary with the Agro-Ecological Zone and range from March until April. Harvest is from July until September.

Data
Land unit data were taken from the Kenya National 1:1 M scale Soils and Terrain (SOTER) database. Natural resource surveys often do not put high priority on collection of hydraulic parameters and key soil fertility characteristics. Consequently, data gaps are a reality for any (small scale) database. These gaps were filled with 'pedo-transfer-functions' (Bouma and van Lanen, 1987) with which data can be estimated from correlated standard, measured soil survey data, such as soil texture.

Evaluation procedure
A mixed qualitative/quantitative approach (Van Lanen *et al.*, 1992) was followed, where a qualitative land evaluation is used to exclude the area not suitable for the projected use and to indicate the limitations for use within the suitable area. The quantified procedure is used to calculate the land use potential for the within the suitable units. The following steps were involved in the evaluation process:

Definition of evaluation units. An overlay of Agro-Climatic Zones (ACZ) with the land unit map provided the basic evaluation units. These Agro-Ecological Units (AEU) formed the basis for further evaluation. Only the spatially dominant soil component within each AEU was analysed. This condition limited computing time but still allowed for the analysis of 91% of the land surface of Kenya.

Suitability assessment for low input maize cultivation: The potentially suitable AEUs for the Land Utilization Type under consideration, were identified using an expert model for physical land evaluation defined in the Automated Land Evaluation System, ALES, (Rossiter, 1990). The land quality 'hazard for erosion' was determined separately for the units considered suitable for maize.

Calculation of crop yield potentials. The potential maize yield under the reigning agro-ecological conditions (soil and climate) was calculated using WOFOST4.1. This model calculates potential, water-limited and nutrient-limited crop production (Van Diepen *et al.*, 1989). The model incorporates files for a range of crops. The nutrient-limited production concerns limitations in the availability of soil nutrients and is calculated with a sub-module based on (Janssen *et al.*, 1990).

Results

Scope for expansion of agricultural land

A large part of Kenya is too dry for growing maize with a reasonable chance of success (see Figure 8). Soil fertility is a major constraint. Only 22% of the total area of Kenya were considered suitable for growing maize under the defined management level (see Table 5).

Table 5: Extent (km^2) of suitable areas for maize in Kenya.

Suitability class	Extent (km^2)
Highly suitable	26,644
Moderately suitable	28,224
Marginally suitable	56,090
Not suitable	399,671
No data	2,113

The physical suitability map was compared with actual land-use and it can be concluded that there is little scope for increasing maize production by expansion of agricultural land. In the areas currently not under mixed farming or annual field cropping, a limited area is moderately to marginally suitable for the defined LUT in physical terms. Significant increases in maize production can therefore only be expected from intensification of existing maize cropping systems.

Scope for intensification of maize cultivation systems

Comparing calculated Nutrient Limited Yield (NLY) and Water Limited Yield (WLY) with Constraint Free Yield (CFY) quantifies the sufficiency of water or nutrient availability. From such a sufficiency or yield gap analysis, we may identify those areas where adapted crop management techniques should have the highest effect. Limitations related to water availability are governed by natural conditions that are difficult to modify by management, such as rainfall amount and pattern, and soil physical and terrain conditions. Irrigation, for example, is often not possible. Assuming that sowing dates and tillage techniques are optimal, we may therefore consider the calculated WLY-potential as a reference-yield that is attainable if other conditions are optimal. The larger the gap between NLY and WLY, the more benefit may be expected from fertilizer input. The areas were soil conditions inhibit added nutrient uptake, such as alkaline soils were excluded in the suitability analysis.

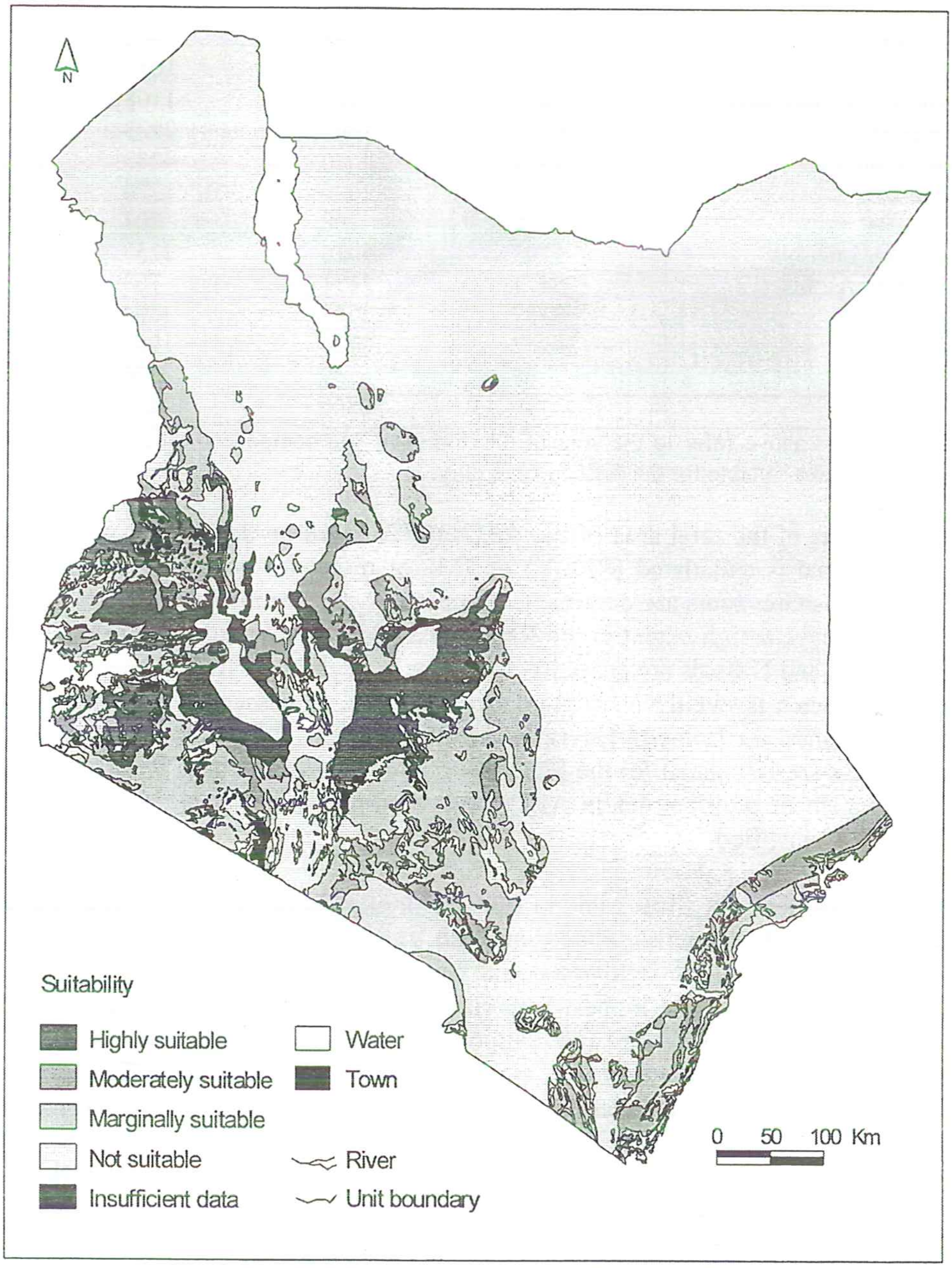


Figure 8 : Physical suitability of Kenyan land for low input maize cultivation.

Table 6 presents the extent (km^2) of the dominant soil components grouped according to the class of NLY-potential expressed as a percentage of the WLY-potential for actual conditions. Response of crops to soil fertility conditions varies with weather conditions. For reasons of legibility and to highlight general trends the NLY-gap analysis is presented by soil unit rather than by the basic evaluation unit, the AEU.

Table 6: Extent (km^2) of nutrient-limited yield classes relative to water-limited yield potential.

Spatially dominant soil	Classes of nutrient limited yield potential (NLY) as a percentage of water limited yield			
	<25%	25%-50%	50%-75%	> 75%
Andosols	1351	2197	441	1961
Vertisols	1015	3838	1053	2189
Phaeozems/ Greyzems	422	517	1108	9644
Planosols	6561	1007	2233	821
Cambisols	1432	4530	4283	4946
Regosols	418	1420	1724	908
Fluvisols	318	208	201	174
Lixisols / Luvisols	6293	3416	2157	885
Acrisols / Alisols	5471	2131	784	1737
Nitosols	3187	2789	3504	5708
Ferralsols	4202	8726	1905	628
Other	60	59	26	-
Total	30730	30838	19419	29601

Note: The numbers above refer to the area of the dominant soil component within a class. They were calculated for the area suitable for the defined LUT only.

In about a quarter of the total area of the AEUs (27% based on dominant soil) a NLY relative to the WLY-potential is calculated (RNLY) of 75% or more, for actual conditions. The relatively fertile Phaeozems/Greyzems are dominant, occupying a third of the total within smallest yield gap class. Vertisols have a high extent in the RNLY class 25%-50%, equivalent to a yield gap of 50%-75%. Cambisols and Nitisols are spatially dominant in the highest RNLY-class (lowest yield gap), but their occurrence is evenly distributed over the RNLY-classes higher than 25%. NLY-gaps (Table 6) are highest for Lixisols, Luvisols, Acrisols, Alisols, Ferralsols and Planosols. No or only small yield gaps are calculated for the Phaeozems and Greyzems. Focus of fertilizer and integrated nutrient management programmes in AEUs with dominantly Ferralsols, Planosols and soils with an argic-B seems justified.

Results from long-term fertilizer trials in Kenya (Smaling *et al.*, 1992; Smaling and Braun, 1996) indicated that on relatively fertile Phaeozems and Vertisols only responses to N-fertilizer are to be expected. A high response to fertilizer-P and farm yard manure additions was found on Nitisols. On Alisols and Arenosols the highest response was found for a combination of added P and N fertilizer. Use of farm yard manure and compost helps to maintain soil organic matter levels which is beneficial for soil buffering capacity, and therewith for fertilizer use efficiency. Smaling *et al.* (1992) concluded that recommendations on a district scale are constrained by spatial variation and that extension staff should give AEU-specific fertilizer recommendations as a general guideline. A stratification of farms based on yield level, and farm size and income within each AEU would be most useful, as profitability and applicability of fertilizer recommendations depend on farm capital and labour resources. They recommend that proper timing and dosage of fertilizer –complying with recommendations specific for AEU and crop type- should be accompanied by efficient use of manure and household waste, introduction of N-fixing species in the cropping system, field grazing of crops residue or incorporation in the soil or use as mulch, and appropriate tillage and soil and water conservation measures to combat degradation.

5. SOFTWARE TOOLS FOR LAND EVALUATION

PLANTGRO

PLANTGRO (Hackett, 1991) is a commercially available software for coarse, or broad scale, prediction of plant performance and was developed in a project for land evaluation for Papua New Guinea. PLANTGRO gives single ratings for land qualities and characteristics and a greatest limitation rating. The latest version includes 1700 plant files with environmental requirements. Pest and disease development over the growing season can be evaluated. The basic assumption underlying the methodology is that crop performance is determined by the level of the least favourable factor (Liebig's Law of the Minimum) or the most limiting or harmful factor.

http://www.ozemail.com.au/~chackett
http://phytoweb.org.au/intro

ALES

The Automated Land Evaluation System (ALES) is a computer program that allows land evaluators to build their own knowledge-based system with which they can compute the physical and economical suitability of map units in accordance with FAO's Framework for Land Evaluation. ALES is not by itself an expert system, and does not include by itself any knowledge about land and land use (Rossiter and Van Wambeke, 1993). ALES is merely a framework within which it is possible to build an evaluation model suited to the prevailing local conditions. In terminology of knowledge-based systems it is a *shell*, which provides a reasoning mechanism and constrains the evaluator to express inferences using this mechanism (Rossiter, 1990).

http://wwwscas.cit.cornell.edu/landeval/ales/ales.htm

FCC

Specialists often need to identify major limitations in Natural Resource Planning and Management, especially in reference to crop-related limitations of broad regions of land resources. Often there are regional limitations that have major policy implications while specific recommendation details can be handled by other forms of assistance to local scientists. One framework that identifies probable crop growth and productivity limitations is the Fertility Capability Classification System (FCC), (Buol *et al.*, 1975). The (free) software can be downloaded from:

http://www.fao.org/waicent/faoinfo/agricult/agl/AGLS/fcc3.htm

ECOCROP

ECOCROP is available as two separate programs that deliver: 1) environmental adaptability information and 2) environmental response information. Ecocrop 1 information permits the identification of 1710 plant species whose most important climate and soil requirements match the information on soil and climate entered by the user. It also permits the identification of plant species for defined uses. It can be used as a library of crop environmental requirements and it can provide plant species attribute files on crop environmental requirements to be compared with soil and climate maps in Agro-ecological zoning (AEZ) databases or Geographical Information System (GIS) map-based display. The database in Ecocrop 2 is designed as a library of studies on crop responses in relation to environmental and management factors. At present the database holds information on a number of varieties for 20 crops of world-wide importance. The information contained in the database can be illustrated on the screen in the form of response curves or it can be written out in statistical form.

http://www.fao.org/waicent/faoinfo/agricult/agl/AGLS/eco1.htm#goeco2

MICROLEIS

The MicroLEIS system is an interactive software package with comprehensive documentation for anyone planning, researching or teaching the sustainable use and management of rural resources, with special reference to the soils from Mediterranean regions. It is an expert-system for land evaluation and land management planning decision support. The (free) software and documentation can be downloaded from:

http://www.irnase.csic.es/microlei.htm

QUEFTS

The QUEFTS methodology (Quantitative Evaluation of the Fertility of Tropical Soils) provides a procedure to calculate nutrient-limited yield as a function of the availability of macro-nutrients, for which P-Olsen, exchangeable potassium and soil pH(H_2O) are diagnostic criteria (Janssen *et al.*, 1990; Smaling and Janssen, 1992). With the calculated nutrient-limited yield it is assumed that no other factors, e.g. water deficit, hampers growth. Optimum NPK fertilizer rates at a given soil and budget can be calculated. N:P:K fertilization aspects can be studied from economic as well as an environmental point of view A software package was developed for application of the methodology.

CAMASE

CAMASE is a project that has a useful site on the web where information can be found on models quantitative models for agricultural and environmental research. CAMASE is short for a Concerted Action for the development and testing of quantitative Methods for research on Agricultural Systems and the Environment. The objectives of CAMASE are to advance quantitative research on agricultural systems and their environment in the EU-countries, by improving systems research in participating institutes through exchange and standardization of concepts, approaches, knowledge, computer programs and data. Specific objectives are: 1) to produce a newsletter, 2) to produce a register of models, 3) to stimulate research on production systems.

http://www.bib.wau.nl/camase/cam-proj.html

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