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LECTURE NOTES ON THE GEOGRAPHY, FORMATION, PROPERTIES AND USE OF THE MAJOR SOILS OF THE WORLD

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PREFACE

These Lecture Notes were co-produced by the Agricultural University Wageningen, the Netherlands, and the Faculty of Agricultural Sciences of the Catholic University of Leuven, Belgium, in support of introductory courses on Soil Geography and Classification.

The Revised Legend of the FAO-Unesco Soil Map of the World (FAO, 1988) is followed to structure the discussion of the distribution, formation, characteristics, management and use of the major soils of the world.

The following persons have contributed to this syllabus:

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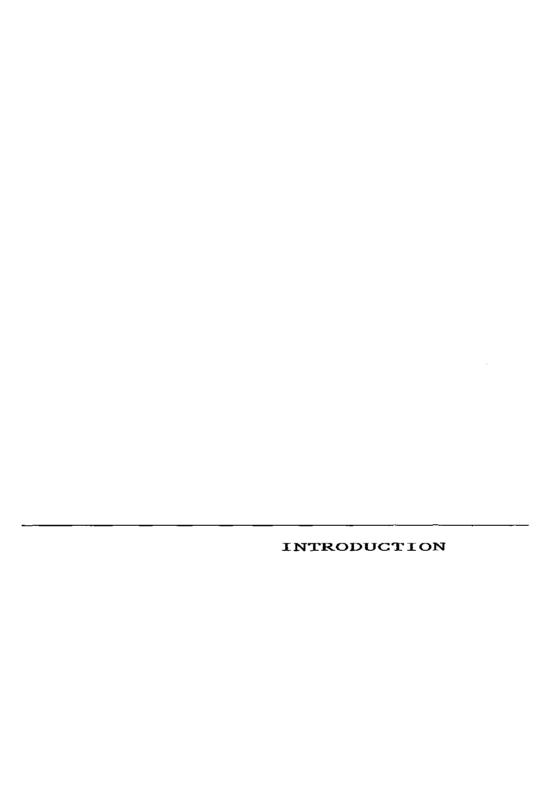
These Lecture Notes are for internal use only; they are still in a first stage of development and will be progressively improved. Suggestions for amendments or additions are most welcome.

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INTRODUCTION

Soils are formed through the impact of <u>climate</u>, <u>vegetation & fauna</u>, (including Man) and <u>topography</u> on the soil's <u>parent material</u>, over a variable <u>time</u> span. The relative importance of each of these five 'soil forming factors' in soil formation (or 'pedogenesis') varies among sites; this explains why there is such a great variety of soils. With few exceptions, soils are still in a process of change; they show in their 'soil profile' signs of differentiation or alteration of the soil material, indicative of a particular pedogenetic history.

Traditional soil names often indicate a domination of one soil forming factor over all other factors. Examples are 'desert soils' (the factor climate dominating), 'plaggen soils' (man), 'prairie soils' (vegetation), 'mountain soils' (topography), or 'volcanic ash soils' (parent material). Alternatively, soil names refer to a dominant single soil characteristic, e.g. the colour ('brown soils'), a chemical characteristic ('alkali soils'), or a particular physical attribute ('hydromorphic soils').

THE FAO-UNESCO CLASSIFICATION OF SOILS

Soils are identified through a system of soil classification. Many such systems have been developed; they differ widely because they are based on different appreciations of soil formation, and use different criteria and different hierarchical subdivisions. The soil classification system used in this syllabus is the one developed by FAO (1974, 1988) for the preparation of the FAO-Unesco Soil Map of the World at scale 1:5,000,000. This system has a geographic dimension through the World Soil Map.

In this syllabus, the 28 'Major Soil Groupings' that are distinguished at the highest taxonomic level in the FAO-Unesco classification, have been grouped in nine 'Sets' by applying a three-step key which addresses the evolutionary and geographic background of the soils:

First, a separation is made between organic soils and mineral soils; Secondly, mineral soils are clustered according to the dominant soil

forming factor or factors which conditioned soil formation;

Thirdly, clusters are subdivided into Major Soil Groupings by considering which combinations of soil forming factors occur within each Set.

Table 1 presents the nine Sets that evolve:

SET 1 includes the organic soils, i.e soils with more than a defined quantity of 'organic soil materials'. All organic soils are accommodated in only one Major Soil Grouping: the HISTOSOLS.

SET 2 contains the man-made soils.

These soils vary widely in properties and appearance and can occur in any environment influenced by human intervention. They too have been accommodated in only one Major Soil Grouping: the ANTHROSOLS.

SETS	DOMINANT IDENTIFYERS	MSGs
SET 1	Organic soils	HISTOSOLS
SET 2	Mineral soils in which soil formation is conditioned by Human influences (not confined to any particular region)	ANTHROSOLS
SET 3	Mineral soils in which soil formation is conditioned by Parent Material	
	 Soils developed in <u>volcanic material</u> Soils developed in <u>residual and shifting sands</u> Soils developed in <u>expanding clays</u> 	ANDOSOLS ARENOSOLS VERTISOLS
SET 4	Mineral soils in which soil formation is conditioned by Topography/Physiography	
	- Soils in <u>lowlands (wetlands) with level topograph</u>	FLUVISOLS GLEYSOLS
	- Soils in <u>elevated regions with non-level topograph</u>	
SET 5	Mineral soils in which soil formation is conditioned by their limited Age (not confined to any particular region)	CAMBISOLS
SET 6	Mineral soils in which soil formation is conditioned by the Climate (& Vegetation) of wet tropical and subtropical regions	PLINTHOSOLS FERRALSOLS NITISOLS ACRISOLS ALISOLS LIXISOLS
SET 7	Mineral soils in which soil formation is conditioned by the Climate (& Vegetation) of arid and semi-arid regions	SOLONCHAKS SOLONETZ GYPSISOLS CALCISOLS
SET 8	Mineral soils in which soil formation is conditioned by the Climate (& Vegetation) of steppes and steppic regions	KASTANOZEMS CHERNOZEMS PHAEOZEMS GREYZEMS
SET 9	Mineral soils in which soil formation is conditioned by the Climate (& Vegetation) of subhumid forest and grassland regions	LUVISOLS PODZOLUVISOLS PLANOSOLS PODZOLS

SET 3 includes the mineral soils whose formation is conditioned by the particular properties of their parent material.

The Set includes three Major Soil Groupings:

- (1) the dark and fluffy ANDOSOLS of volcanic regions,
- (2) the <u>sandy</u> ARENOSOLS of desert areas, beach ridges, inland dunes, areas with highly weathered sandstones, etc., and
- (3) the swelling and shrinking <u>heavy clayey</u> VERTISOLS of backswamps, river basins, lake bottoms, and other periodically wet areas with a high content of expanding 2:1 lattice clays.

SET 4 is composed of mineral soils whose formation was markedly influenced by their topographic/physiographic setting.

Such soils can be found in a low terrain position, associated with recurrent floods and/or prolonged wetness, but also in elevated and/or accidented terrain where soil formation is hindered by low temperatures and/or erosion. The Set holds four Major Soil Groupings:

- (1) In low areas, they are young alluvial FLUVISOLS which show stratification or other evidence of recent sedimentation, and
- (2) non-stratified GLEYSOLS in waterlogged areas that do not receive regular additions of sediment.
- (3) In elevated and/or eroding areas, they are the shallow LEPTOSOLS over hard rock or highly calcareous material, and
- (4) the deeper REGOSOLS which occur in unconsolidated materials and which have only surficial profile development, e.g. because of low soil temperatures, prolonged dryness or erosion.

SET 5 accommodates (other) soils that are only moderately developed on account of their limited pedogenetic age or of rejuvenation. Moderately developed soils occur in all environments, from the sea level to the highlands, from the equator to the boreal regions, and under all kinds of vegetation. They have not more in common than 'signs of beginning soil formation' so that there is considerable diversity among the soils in this Set. Nonetheless, they all belong to only one Major Soil Grouping: the CAMBISOLS.

The remaining 18 Major Soil Groupings of the FAO-Unesco system all represent mineral soils whose formation was largely conditioned by the past or present climate and by the climax vegetation which developed under this climate.

SET 6 accommodates the red and yellow soils of wet tropical and subtropical regions.

High soil temperatures and (in times) ample water promote rock weathering and mineralization of soil organic matter. The Major Soil Groupings in this Set have in common that a long history of rapid dissolution and transport of weathering products has resulted in deep and genetically mature soils:

- (1) the PLINTHOSOLS of old weathering surfaces, with a mixture of clay and quartz that hardens irreversibly upon exposure to the open air,
- (2) the deeply weathered FERRALSOLS that have a very low cation exchange capacity and are virtually devoid of weatherable minerals,

- (3) the deep, NITISOLS in richer parent material that have shiny, nutty structure elements and a stretched clay accumulation horizon.
- (4) the strongly leached, red and yellow ACRISOLS on acid parent rock, with a clay accumulation horizon, low base saturation and low activity clays,
- (5) the less common ALISOLS with low base saturation and high activity clays, and
- (6) the LIXISOLS with a high base saturation and a low cation retention capacity.

SET 7 accommodates Major Soil Groupings that are typical of <u>arid and</u> semi-arid regions.

An important mechanism of horizon differentiation in soils in the dry zone is by redistribution of calcium carbonate and gypsum. Soluble salts may accumulate at some depth or, alternatively, near the soil surface in areas with shallow groundwater. The various Major Soil Groupings in Set 7 are:

- the SOLONCHAKS of desert and semi-desert regions; these soils have a high content of accumulated soluble salts,
- (2) the SOLONETZ with a high percentage of adsorbed sodium ions,
- (3) the GYPSISOLS with a horizon of secondary gypsum enrichment, and
- (4) the CALCISOLS with secondary carbonate enrichment.

SET 8 holds soils that occur in the <u>steppic zone</u> between the dry climates and the humid temperate regions at higher latitudes.

This transition zone has a climax vegetation of ephemeric grasses and dry forest; its location corresponds roughly with the transition from a dominance of accumulation processes in soil formation to a dominance of leaching processes. Set 8 includes four Major Soil Groupings:

- the KASTANOZEMS with carbonate and/or gypsum accumulation at some depth; these soils occur in the driest parts of the steppe zone,
- (2) the deep, very dark CHERNOZEMS with carbonate enrichment in the subsoil,
- (3) the dusky PHAEOZEMS of prairie regions with a high base saturation but no visible signs of secondary carbonate accumulation, and
- (4) the dark GREYZEMS with high organic matter contents in their surface soils and signs of bleaching along ped faces. These soils occur in the wettest parts of the zone.

The soils in this Set show evidence of clay and/or organic matter redistribution. The cool climate and the short genetic history of most soils in this zone explain why some soils are still relatively rich in bases despite a dominance of eluviation over enrichment processes. Eluviation and illuviation of metal-humus complexes produce the (greyish) bleaching and (brown to black) coatings that are common in this zone. Set 9 contains four Major Soil Groupings:

- (1) the base-rich LUVISOLS with a distinct clay accumulation horizon,
- (2) the base-poor PODZOLUVISOLS with a bleached eluviation horizon tonguing into a clay enriched horizon,
- (3) the PLANOSOLS with a bleached topsoil over a dense, slowly permeable subsoil, and
- (4) the acid PODZOLS with a bleached eluviation horizon over an accumulation horizon of organic matter with aluminium or iron or with both.

NOTE THAT Table 1 indicates merely the most common relationships between Major Soil Groupings and soil forming factors. The Major Soil Groupings in Sets 6 through 9 represent soils, which occur predominantly in specific climate zones. Such soils are traditionally referred to as 'zonal soils'. However, not all soils in Sets 6 through 9 are zonal soils, nor are soils in other Sets always non-zonal. Podzols, for instance, are most common in (sub)humid temperate climates but they are also found in the humid tropics; Planosols may equally occur in subtropical and steppe climates; Regosols are not necessarily confined to eroding uplands; Ferralsols may occur as remnants outside the humid tropics...

Soils whose characteristics result from a strong <u>local</u> dominance of a soil forming factor other than 'climate' are not zonal soils. They are 'intrazonal soils'. In other words, there are zonal and intrazonal Podzols, zonal and intrazonal Gleysols, zonal and intrazonal Histosols, and many more.

There are also 'azonal soils', i.e. soils that are too young to reflect the influence of site-specific conditions in their profile characteristics. Young alluvial soils (Fluvisols) and soils in recent hillwash (e.g. Cambisols) are examples of azonal soils.

The zonality concept helps to explain the worldwide distribution of soils but cannot be used as a basis for soil classification. The Sets presented in the foregoing may therefore not be taken as high level classification units but merely as a reflection of common relationships between soils and the factors which govern their formation.

It was already mentioned that the FAO-Unesco classification system has a hierarchical structure. The 28 Major Soil Groupings at the highest level of generalization are subdivided into a total of 153 Soil Units at the second level. In the 'Revised Legend' of 1988, a third hierarchical level, that of the Soil Subunits was introduced. Soil Subunits are either intergrades between Soil Units, or mark characteristics in addition to those used in the definition of Soil Units. This further subdivision proved necessary when the system became used in connection with soil mapping at increasingly larger scales.

The system accommodates also surface or subsurface features of the land that stem from or influence agricultural use; such 'phases' are designations that have practical significance but are non-taxonomic and can cut across the boundaries of Soil Units and/or Soil Subunits.

DIAGNOSTIC HORIZONS AND DIAGNOSTIC PROPERTIES

The taxonomic units of the FAO-Unesco classification were selected on the basis of (present knowledge of) the distribution, formation and characteristics of the soils of the world, and in line with their importance as natural resources and their environmental significance. To facilitate soil identification and correlation in areas far apart, the various taxonomic units in the FAO-Unesco system have been defined in terms of measurable and observable key properties. The differentiating criteria are essentially properties of the soil itself and have been selected on the basis of generally accepted principles of soil formation. Selected key properties are clustered to 'diagnostic horizons', the basic identifiers in soil classification. The system uses individual 'diagnostic properties' as further differentiating characteristics.

The definitions and nomenclature of diagnostic horizons and diagnostic properties are very similar to the concepts developed by the USDA Soil Taxonomy classification system (Soil Survey Staff, 1975). However, some of the definitions have been simplified or modified in accordance with the requirements of the Soil Map of the World legend.

The various diagnostic horizons and diagnostic soil properties are summarized in Tables 2 and 3.

TABLE 2. Summary of diagnost	,	full definitions).
DIAGNOSTIC HORIZON		

-	
DIAGNOSTIC HORIZON	MOST PROMINENT FEATURES
histic H-horizon	peaty surface soil of 20 to 40 cm depth; in some cases till 60 cm.
mollic A-horizon	surface horizon with dark colour due to organic matter; base saturation exceeds 50 percent.
umbric A-horizon	similar to a mollic A-horizon but base saturation lower than 50 percent.
Fimic A-horizon	man-made surface layer, 50 cm or more thick, produced by long-continued manuring.
ochric A-horizon	surface horizon without stratification and lacking the characteristics of a histic H-horizon, or a mollic, umbric or fimic A-horizon.
albic E-horizon	bleached eluviation horizon with the colour of uncoated primary soil material, usually overlying an illuviation horizon.
argic B-horizon	clay accumulation horizon lacking properties of a natric B-horizon and/or a ferralic B-horizon.
natric B-horizon	clay accumulation horizon with more than 15 percent exchangeable sodium, usually with a columnar or prismatic structure.
spodic B-horizon	horizon with illuviation of organic matter with iron or aluminium or with both.
ferralic B-horizon	highly weathered horizon, at least 30 cm thick, with a cation exchange capacity of 16 cmol(+)/kg clay or less.
cambic B-horizon	genetically young B-horizon (therefore not meeting the criteria for argic, natric, spodic or ferralic horizons) showing evidence of alteration: modified

	the criteria for argic, natric, spodic or ferralic
	horizons) showing evidence of alteration: modified
	colour, removal of carbonates or presence of soil
	structure.
calcic horizon	horizon with distinct calcium carbonate enrichment.
gypsic horizon	horizon with distinct calcium sulphate enrichment.
petrocalcic horizon	a continuous cemented or indurated calcic horizon.

petrogypsic horizon a gypsic horizon hardened to the extent that dry fragments do not slake in water and roots cannot enter.

sulfuric horizon horizon of at least 15 cm thick, having jarosite mottles and $pH(H_2O, 1:1) < 3.5$.

NOTE THAT a distinction must be made between the soil horizon designations that are given in a soil profile description, and diagnostic horizons as used in soil classification. The former belong to a nomenclature in which master horizon codes (H, O, A, E, B, C and R; see Annex 4) are assigned to the various soil layers described and interpreted in the field. The choice of horizon code is by personal judgement of the soil surveyor and based on presumed processes of soil formation. In contrast, diagnostic horizons are quantitatively defined and their presence or absence can be ascertained on the basis of unambiguous field and/or laboratory measurements.

Some of the diagnostic horizons in the FAO-Unesco system are special forms of A- or B-horizons, e.g. a 'mollic' A-horizon, or a 'ferralic' B-horizon. Other diagnostic horizons are not necessarily A- or B-horizons, e.g. a 'calcic' or a 'gypsic' horizon. In these cases the horizon code is not added to the name of the diagnostic horizon.

TABLE 3. Summary of diagnostic properties (see Annex 2 for full definitions) DIAGNOSTIC MOST PROMINENT FEATURES **PROPERTIES**

refer to largely pyroclastic (weathering) material; andic properties normally high in extractable aluminium.

ferralic properties mark a cation exchange capacity (by 1M NH4OAc at pH 7.0) of less than 24 cmol(+)/kg clay in Cambisols and Arenosols.

mark the presence of Fe-enriched mottles or nodules ferric properties

in Alisols, Lixisols and Acrisols.

fluvic properties mark ongoing sedimentation or stratification or an irregular organic carbon profile in recent alluvial

sediments.

mark soil materials having 1.5 cmol(+)/kg soil or geric properties less of extractable bases plus aluminium and a pH(1M KCl) of 5.0 or more, or having a delta-pH (pH KCl

minus pH H₂O) of -0.1 or more.

gleyic and stagnic present visible evidence of prolonged waterlogging properties either by shallow groundwater (gleyic properties) or by a perched water table (stagnic properties).

mark a moderate to strong angular blocky soil nitic properties

structure, easily falling apart into smaller blocky elements, showing shiny ped faces.

mark an electrical conductivity of the saturation salic properties extract of more than 15 dS/m, or of more than 4 dS/m

if the $pH(H_2O,1:1)$ exceeds 8.5.

mark high saturation of the exchange complex by sodic properties (15 percent or more) or by sodium plus

magnesium (50 percent or more).

mark cracks, slickensides, wedge-shaped or paralvertic properties lelipiped structural aggregates that are not in a combination, or are not sufficiently expressed for

the soil to qualify as a Vertisol.

abrupt textural marks a doubling of the clay content, or an increase change of the clay content by 20 percent, over a vertical

distance of less than 5 cm.

Table 3 cont'd

calcareous refers to soil material which shows strong effervescence in contact with HCl and/or having more than 2

percent of CaCOz-equivalent.

continuous hard material which is sufficient coherent and hard when

rock moist to make digging with a spade impracticable.

gypsiferous refers to soil material which contains 5 percent or

more gypsum.

interfingering narrow penetrations of an albic E-horizon into an

underlying argic or natric B-horizon along mainly

vertical ped faces.

organic soil saturated soil materials (unless drained), having materials >=18 percent organic carbon if having >=60 percent

clay, or having >=12 percent carbon if without clay, or having a proportional carbon content if the clay content is between zero and 60 percent; or soil materials that are never saturated for more than a

few days having >=20 percent organic carbon.

permafrost the condition of soil temperatures being perennially

at or below 0 °C.

plinthite an iron-rich, humus-poor mixture of clay and quartz

that hardens irreversibly on drying.

slickensides polished and grooved surfaces that are produced by

one soil mass sliding past another.

smeary consistence a consistence which changes under pressure and

returns to the original state after the pressure is released ('thixotropic' materials in Andosols).

soft powdery lime calcium carbonate precipitated in situ and soft

enough to be cut with a finger nail.

strongly humic refers to soil materials having more than 1.4 percent

organic carbon as a weighted average over a depth of

100 cm from the surface.

sulfidic materials waterlogged mineral or organic soil material contain-

ing 0.75 percent or more sulfur and less than three

times as much carbonates as sulfur.

tonguing relatively wide penetrations of an albic E-horizon

into an underlying argic or natric B-horizon mainly

along vertical ped faces.

weatherable minerals that release plant nutrients and iron or

minerals aluminium by weathering.

NOTE THAT the generalized descriptions of diagnostic horizons and diagnostic soil properties given in Tables 2 and 3 are solely meant as a first introduction to FAO-Unesco terminology. Actual soil classification requires that exact concepts and full definitions be used. These are presented in Annexes 1 and 2.

Final taxon identification is done by identifying the Major Soil Grouping with the 'KEY TO MAJOR SOIL GROUPINGS', presented hereafter. The identification of the Soil Unit (within the selected Major Soil Grouping) is done using the 'KEY TO SOIL UNITS'. A full Key to Soil Units is contained in each chapter which describes (the Regional Distribution, Genetic History, Characteristics, Management and Use of) a particular Major Soil Grouping.

KEY TO MAJOR SOIL GROUPINGS

This 'Key to Major Soil Groupings' refers to the chapter in this text in which a particular Major Soil Grouping is described. Chapters on related Major Soil Groupings are preceded by a description of the major landforms with which the groupings are commonly associated.

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 $0.1~\text{Mg/m}^3$) 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 if it rests on rocks or on fragmented material of which the interstices are filled with organic matter:

HISTOSOLS (HS)

(page 19)

Other soils in which human activities resulted in profound modification of the original soil characteristics:

ANTHROSOLS (AT)

(page 35)

Other soils which are limited in depth by continuous hard rock or highly calcareous materials or a continuous cemented layer within 30 cm of the surface, or having less than 20 percent fine earth over a depth of 75 cm from the surface; having no diagnostic horizons other than a mollic, umbric, or ochric A-horizon with or without a cambic B-horizon:

LEPTOSOLS (LP)

(page 115)

Other soils having, after the upper 20 cm have been mixed, 30 percent or more clay in all horizons to a depth of at least 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 parallelipiped structural aggregates at some depth between 25 and 100 cm from the surface:

VERTISOLS (VR)

(page 67)

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

FLUVISOLS (FL)

(page 93)

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

SOLONCHAKS (SC)

(page 181)

Other soils 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 calcic or a gypsic horizon; lacking plinthite within 125 cm of the surface:

GLEYSOLS (GL)

(page 105)

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)

(page 47)

Other soils which are coarser than sandy loam to a depth of at least 100 cm from the surface, having no diagnostic horizons other than an ochric A-horizon or an albic E-horizon:

ARENOSOLS (AR)

(page 59)

Other soils having no diagnostic horizons other than an ochric or umbric A-horizon:

REGOSOLS (RG)

(page 119)

Other soils having a spodic B-horizon:

PODZOLS (PZ)

(page 259)

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 if 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)

(page 139)

Other soils having a ferralic B-horizon:

FERRALSOLS (FR)

(page 147)

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, exclusive of a natric or a spodic B-horizon:

PLANOSOLS (PL)

(page 253)

Other soils having a natric B-horizon:

SOLONETZ (SN)

(page 191)

Other soils having a mollic A-horizon with a moist chroma of 2 or less to a depth of at least 15 cm, showing uncoated silt and quartz grains on structural ped surfaces; having an argic B-horizon:

GREYZEMS (GR)

(page 231)

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 horizon, or concentrations of soft powdery lime within 125 cm of the surface, or both:

CHERNOZEMS (CH)

(page 219)

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 gypsic horizon, or concentrations of soft powdery lime within 125 cm of the surface:

KASTANOZEMS (KS)

(page 215)

Other soils having a mollic A-horizon; having a base saturation (by 1M NH₄OAc at pH 7.0) of 50 percent or more throughout the upper 125 cm of the soil:

PHAEOZEMS (PH)

(page 227)

Other soils having an argic B-horizon showing an irregular or broken upper boundary resulting from deep tonguing of the E-horizon 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 with iron and having redder hues and stronger chromas than the interiors:

PODZOLUVISOLS (PD)

(page 247)

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 invaded by gypsum or calcium carbonate, a calcic or a petrocalcic horizon:

GYPSISOLS (GY)

(page 197)

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 invaded by calcium carbonate:

CALCISOLS (CL)

(page 203)

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 A- and B-horizons; having nitic properties in some subhorizon within 125 cm of the surface;

NITISOLS (NT)

(page 157)

Other soils having an argic B-horizon which has a cation exchange capacity equal to or more than 24 cmol(+)/kg clay and a base saturation (by 1M NH₄OAc at pH 7.0) of less than 50 percent at least in some part of the B-horizon within 125 cm of the surface;

ALISOLS (AL)
(page 167)

Other soils having an argic B-horizon which has a cation exchange capacity of less than 24 cmol(+)/kg clay and a base saturation (by 1M NH₄OAc at pH 7.0) of less than 50 percent in at least some part of the B-horizon within 125 cm of the surface:

ACRISOLS (AC)

(page 161)

Other soils having an argic B-horizon which has a cation exchange capacity equal to or more than 24 cmol(+)/kg clay and a base saturation (by 1M $\rm NH_4OAc$ at pH 7.0) of 50 percent or more throughout the B-horizon to a depth of 125 cm:

LUVISOLS (LV)

(page 241)

Other soils having an argic B-horizon which has a cation exchange capacity of less than 24 cmol(+)/kg clay and a base saturation (by 1M NH₄OAc at pH 7.0) of 50 percent or more throughout the B-horizon to a depth of 125 cm:

LIXISOLS (LX)

(page 171)

Other soils having a cambic B-horizon:

CAMBISOLS (CM)

(page 125)

CORRELATION

Since the beginnings of soil science, at the end of the 19th century, many soil classification systems have been developed. The first systems published reflected clearly their place of origin namely regions dominated by quaternary materials of periglacial origin in the USSR, Central and Western Europe and the USA. As more experience was gained in other regions as well, including the tropics and subtropics, novel approaches to soil classification were developed which embrace the soil cover at a global scale.

Truely international classification systems were developed in France, the USA and the USSR. The FAO-Unesco classification of soils was set up as a common denominator to facilitate correlation and comparison between systems. An in-depth discussion of the subject of correlation is clearly beyond the scope of this syllabus; it may suffice to comment on the analogies and differences which exist between the FAO-Unesco concepts and definitions and those of the USDA Soil Taxonomy.

The FAO-Unesco classification is less than a full-fledged taxonomic system in the sense that it has only two categories (Major Soil Groupings and Soil Units) with a third one (Soil Subunits) to be developed. It maintained traditional nomenclature as much as possible.

The USDA Soil Taxonomy comprises six categories: Orders, Suborders, Great Groups, Subgroups, Soil Families and Soil Series. At the highest hierarchical level Soil Taxonomy defines 11 Orders. The broad criteria by which Orders are differentiated are as follows:

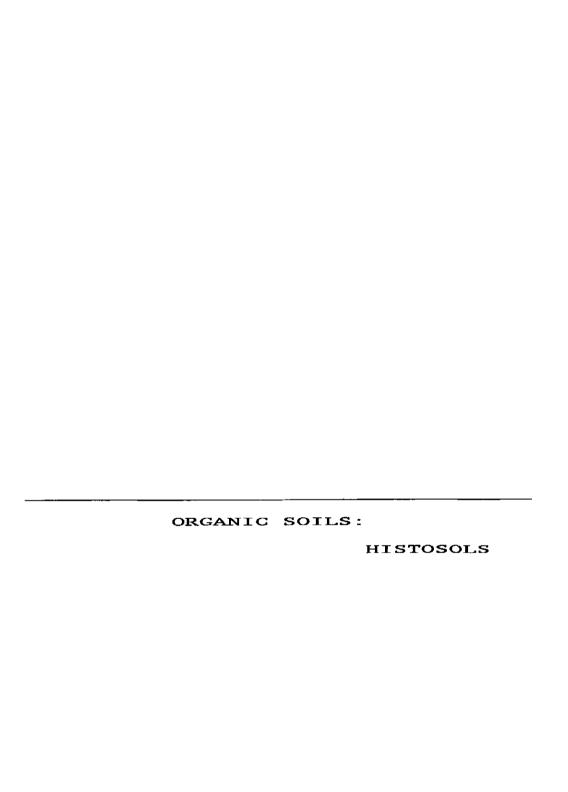
- (1) the production of organic materials over deterioration and destruction (Order of Histosols):
- (2) the production and illuvial accumulation of metal-humus complexes and associated compounds (Order of Spodosols);
- (3) the presence of low-charge, mainly pH-dependent constituents (Order of Oxisols):
- (4) a limited capability to form horizons due to a significant amount of shrinking and swelling clays. (Order of Vertisols);
- (5) a very low rate of transformation due to the lack of soil moisture (Order of Aridisols);
- (6) the presence of illuviated crystalline clays and a relatively high base reserve in the substratum (Order of Alfisols);
- (7) the production and maintenance of a base-rich, organic enriched surficial horizon and a high base reserve (Order of Mollisols);
- (8) the presence of illuviated crystalline clays and a relatively low base reserve in the substratum (Order of Ultisols);
- (9) the alteration of parent materials and initiation of horizon differentiation (Order of Inceptisols);
- (10) the presence of materials that may be altered to form horizons with the advent of more pedologic time (Order of Entisols);
- (11) the transformation of (mainly volcaniclastic) parent materials and the presence of extractable aluminium (Order of Andisols; proposed, at present in the Inceptisols).

Both systems use diagnostic horizons and diagnostic properties as taxonomic identifyers but there exist considerable differences between the two systems in the definitions of the criteria used and in the levels at which they are applied. One important difference is that, unlike Soil Taxonomy, the FAO-Unesco system does not use the soil temperature and soil moisture regime as distinctive criteria in the classification but projects these data on the soil map as an overlay.

There are also differences and similarities between both systems in the naming of taxonomic units. Both systems use formative elements. In part these were derived from traditional soil names that acquired general acceptance and are firmly established in the literature on soils. For a limited number of soils it was needed to coin new names in order to avoid confusion. In both systems names are meant to sum up, in an easily remembered term, a set of characteristics which have been found to be representative of a particular soil in different parts of the world. The FAO-Unesco classification uses one name to denote the Major Soil Grouping with adjectives for Soil Units and Soil Subunits. The formative elements in the names of the Soil Orders are used as name-endings at lower hierarchical levels, as in the following example:

SOIL ORDER	SOIL SUBORDER	GREAT SOIL GROUP	SOIL SUBGROUP
Sp <u>od</u> osol	Orth <u>od</u>	Fragiorth <u>od</u>	Typic Fragiorth <u>od</u>

The Major Soil Groupings of the FAO-Unesco classification compare with either Orders or Suborders of Soil Taxonomy; the Soil Units correspond roughly with Great Groups.



HISTOSOLS

Soils having 40 cm or more organic soil materials 0 (60 cm or more if the organic materials consist mainly of sphagnum or moss or have a bulk density of less than $0.1~{\rm Mg/m^{3}}$) either extending down from the surface or taken cumulatively within the upper 80 cm of the soil; the thickness of the organic surface horizon may be less if it rests on rock or on fragmental material in which the interstices are filled with organic matter.

Key to Histosol (HS) Soil Units

Histosols having permafrost^a within 200 cm of the surface.

Gelic Histosols (HSi)

Other Histosols having a sulfuric * horizon or sulfidic materials a at less than 125 cm from the surface.

Thionic Histosols (HSt)

Other Histosols that are well drained and are never saturated with water for more than a few days.

Folic Histosols (HS1)

Other Histosols having raw or weakly decomposed organic materials^a, the fiber content of which is dominant to a depth of 35 cm or more from the surface; having very poor drainage or being undrained.

Fibric Histosols (HSf)

Other Histosols having highly decomposed organic materials with strongly reduced amounts of visible plant fibers and a very dark grey to black colour to a depth of 35 cm or more from the surface, having an imperfect to poor drainage.

Terric Histosols (HSs)

SUMMARY DESCRIPTION OF HISTOSOLS

Connotation: peat and muck soils; from Gr. histos, tissue.

<u>Parent material</u>: incompletely decomposed plant remains, with or without admixtures of sand, silt or clay.

<u>Environment</u>: the majority of all Histosols have formed in boreal regions. Elsewhere, Histosols are confined to poorly drained basins and depressions, swamp and marsh lands with shallow groundwater, and highland areas with a high precipitation/evapotranspiration ratio.

^{*} Diagnostic horizon; see Annex 1 for full definition.

a Diagnostic property; see Annex 2 for full definition.

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<u>Profile development</u>: mostly H or HCr profiles. Transformation of plant remains through biochemical desintegration and formation of humic substances create a surface layer of mould. Translocated organic material may accumulate in deeper tiers but is more often leached from the soil.

<u>Use</u>: peat lands are used for various forms of extensive forestry and/or grazing or lie idle. If carefully reclaimed and managed, Histosols can be very productive under capital-intensive forms of arable cropping/horticulture. Deep Histosols are best left untouched.

REGIONAL DISTRIBUTION OF HISTOSOLS

Peats cover an estimated 240 million hectares worldwide (Figure 1). Some 200 million hectares of peat land are situated in the boreal and temperate regions of North America, Europe and Asia. The rest is in the tropics and subtropics with the most prominent formations occurring in the coastal lowlands of southeast Asia, where some 20 million hectares of acid forest peat border the Sunda Flat.

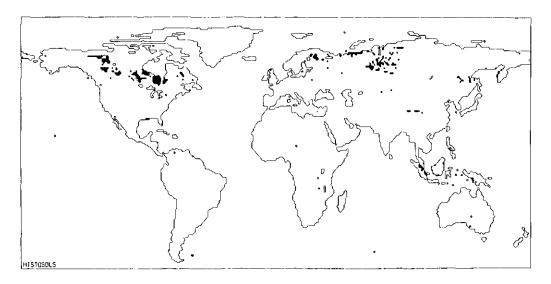


Fig. 1. Histosols worldwide.

GENESIS OF HISTOSOLS

Peat accumulation is always associated with conditions where organic material is produced by an adapted (climax) vegetation, and bio-chemical decomposition of plant debris is retarded due to:

- (1) low temperatures, and/or
- (2) persistent waterlogging, and/or

- (3) extreme acidity or oligotrophy, and/or
- (4) the presence of high levels of electrolytes or organic toxins. Figure 2 indicates that a net surplus of organic material can build up under swamp conditions even in the tropics.

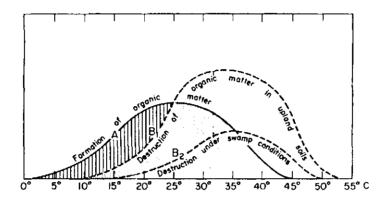


Fig. 2. Comparative rates of production (A) and decomposition (B) of organic matter as a function of mean annual temperature and aeration of the soil. Source: Van Dam, 1971.

In view of the limited agricultural significance of the (extensive) boreal moss peats, the following discussion of Histosol development will focus on peats in temperate and tropical climates. The examples and illustrations used will mainly refer to peat areas in the (humid) tropics where soil forming processes are accelerated and soil formation manifests itself more clearly than in cooler climates. However, the mechanisms discussed apply to peat bogs and Histosols in all climates.

The majority of all peat bogs occur in lowlands areas, e.g. in stagnant coastal plains and in lagoons that are separated from the sea by beach ridges parallel to the shore (Andriesse, 1974). High groundwater, frequent flooding and sealing with mineral sediment and, in coastal peats, high salt contents retard the decomposition of plant debris. As a consequence, near-stagnant pools in depression areas are gradually filled in with the remains of aquatic plants in their deeper parts and with reeds, sedges, grasses and ferns in their fringe areas. The shallow fringe areas of a depression are the first to become filled in and 'dry'. This prompts the vegetation, differentiated in floral belts in response to different degrees of wetness, to shift toward the centre of the depression. Eventually, the entire depression is filled with 'topogenous peat' (groundwater peat). There may be an abrupt change from the mineral substrata to the overlying peat body or the transition may be gradual; a thin transitional layer of black, smeary and totally decomposed organic sediment ('gyttja') can occur as well (see Figure 3).

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Topogenous peats are shallow by nature. Only where their accumulation coincides with gradual tectonic lowering of the land surface can they reach a great depth. The topogenous peat deposits in the Drama Plain, Greece, for instance, are in places deeper than 300 meters.

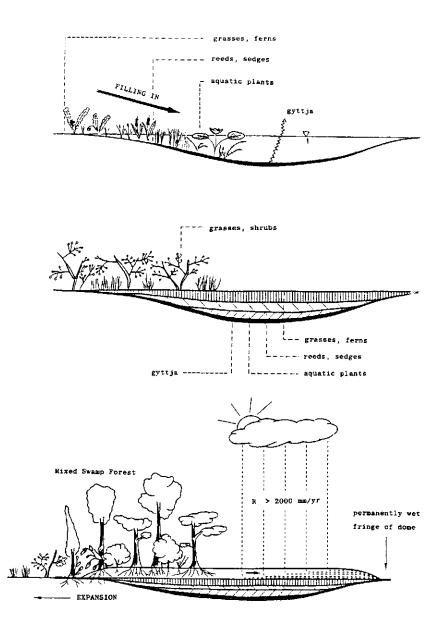


Fig. 3. A depression is gradually filled in with topogenous peat which is then overgrown by a laterally expanding ombrogenous peat mass. Note the changing composition of the vegetation.

After a depression is completely filled with topogenous peat, further accumulation of peat <u>may</u> take place if rainfall is high and evenly spread over the year, and microbial activity is suppressed by severe acidity, oligotrophy and/or organic toxins. The then formed rain-dependent 'ombrogenous peat' mass rises over the mean water level and is no longer enriched with mineral material (clay, nutrients) from outside. The continual precipitation surplus (a precondition for the formation of ombrogenous peat) drains away to the fringe of the bog where it creates a wet zone. Topogenous peat can grow here and become covered with ombrogenous peat later on. Peat bodies from various nuclei (depressed areas) in a plain will eventually merge into one coherent peat formation.

<u>In temperate regions</u>, topogenous 'low moor' peats include woody forest peats as well as peats derived from a grassy marsh vegetation. The 'high moor' peats of ombrogenous raised bogs are mostly moss (Sphagnum) peats. <u>In the tropics</u>, almost all lowland peats are ombrogenous; the peat consists of woody rain forest debris. At high elevations and in the subtropics, peats are predominantly topogenous and less woody (Papyrus swamps, sawgrass peats, etc.)

As long as a peat body is still shallow, the vegetation can draw nutrients from the underlying mineral base. Once the peat has grown to a depth that puts the subsoil out of reach of the living roots, losses of nutrients through leaching, fixation or otherwise, force the vegetation to survive on a gradually decreasing quantity of cycling nutrients. The (climax) vegetation adapts by becoming poorer in quality and species composition. An initially heavy mixed swamp forest degrades slowly into a light monotonous forest, and ultimately into a stunted forest which produces insufficient organic material for further vertical growth of the bog. This limit to vertical growth, in combination with continuing lateral expansion, explains the characteristic dome shape of ombrogenous 'raised bogs'.

The rate of vertical peat accumulation depends on the vegetation, on the genetic age of the bog, and on a number of outside factors. The growth rate seems to decrease with time according to a roughly exponential pattern. Carbon datings of bogs in Sarawak and Indonesia suggest that the initial accumulation rate of 2.5 to 4.5 mm/yr decreases in the course of 3 to 5 millennia to 0.5 mm/yr and less in deep (8 to 12 m) dome areas (Anderson, 1964). Figure 4 presents a view from the flat top of a (drained) ombrogenous raised bog.

<u>Topogenous peats</u> are often well decomposed throughout. Botanical stratification is partly still recognizable but most of the identifiable plant fibers desintegrate upon gentle rubbing. Such **sapric** material makes up the body of many groundwater peats but there are numerous exceptions. Mangrove peat of raw and brittle fibric material is an example. Intermediate hemic material is also quite common in topogenous peats.



Fig. 4. The steep edge of a peat dome near Pontianak, Indonesia (site at km 11 in Figure 5). Note the steep sides and flat top of the dome which extends 7 meters over the mineral coastal plain in the background.

The <u>ombrogenous peats</u> of the temperate zone are mostly moss peats; they are less decomposed near the surface than at some depth whereas the ombrogenous (forest) peats of the wet tropics show the highest degree of decomposition in the top 10-30 cm layer.

The situation in Sphagnum domes is clear: the mosses grow at their top ends and the dead base parts become more and more humified.

The greater decomposition near the surface of tropical raised bogs than at some depth is a consequence of the better aeration and higher nutrient levels of the surface tiers which promote microbial activity. Once the relatively well decomposed surface material becomes covered with younger peat in the course of further vertical growth of the bog, the water table rises (bogs are sponges!) and brings the former surface horizon within the permanently saturated zone. This retards further decomposition of the material (see Figure 1) while soluble and fine-grained insoluble decomposition products continue to be removed with effluent water. The remaining lignin/wood-rich skeleton forms the loose and coarse fibric subsurface peat typical of well developed dome formations in the tropics.

In horizontal direction there is a similar differentiation. Permanently saturated central dome areas consist of less decomposed peat than comparable layers nearer to the fringe of the dome where the peat is younger, (relatively) rich in nutrients and occasionally aerated and subject to more intensive microbial attack. Figure 5 shows the different peat types and the different degrees of peat decomposition in a tropical raised bog.

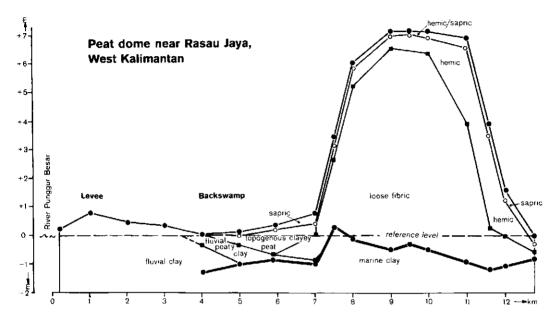


Fig. 5. Cross section of a small coastal peat formation near Pontianak, Indonesia. Note the fibric core of the ombrogenous dome, the better decomposed (hemic and sapric) shell, and the occurrence of sapric and clayey topogenous peat at the foot of the dome. Source: Driessen, 1978.

CHARACTERISTICS OF HISTOSOLS

It is evident that the characteristics of a Histosol are determined by the type of the peat bog, its depth, the floristic composition of the peat, the stratification/homogenization, packing density, wood content, mineral content and degree of decomposition of the peat material, and a score of other factors. Hence, only a general description of 'peat soil characteristics' is possible.

Most Histosols have an H-horizon of more than 40 cm over a strongly reduced mineral subsoil which is often at great depth. H, H/Cr and HCr profiles are the commonest configurations. Soils with an organic surface horizon shallower than 40 cm qualify as Histosols only if the organic materials rest directly on bedrock; more commonly they are Fluvisols, Gleysols or (Gleyic) Podzols, Major Soil Groupings which commonly occur in association with Histosols (Schmidt-Lorenz, 1986).

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Hydrological characteristics

Most virgin peats are permanently wet, with the water table close to the surface of the soil. The central areas of virgin topogenous peat bodies and of extensive ombrogenous formations are always saturated with (near) stagnant water. The fringe areas of extensive raised bogs have a less monotonous water regime, with drier areas near natural depressions due to gravity drainage of the immediate surroundings. The opposite seems true for smaller domes where the water regime is less buffered; radial drainage of such domes results in occasional floods near the margins while the centre may, at times, be quite dry at the surface.

Physical characteristics

Ombrogenous Fibric Histosols are loosely packed in the natural state, with a bulk density (BD) that is typically between 0.05 and 0.1 Mg/m^3 ; the surface tiers contain (still) more solid material than the subsurface layers and have bulk densities of 0.1 to 0.2 Mg/m^3 . See Table 1. Reclaimed (drained and cropped) peats acquire a higher bulk density of (up to 0.4 Mg/m^3) after a few years of consolidation and decomposition of the

to 0.4 Mg/m^3) after a few years of consolidation and decomposition of the peat (Terric Histosols).

The specific gravity (SG) of peat material with low contents of mineral constituents (less than 3 percent by weight) is always close to 1.4 Mg/m³. It follows that the total pore space fraction (= 1-BD/SG) of fibric ombrogenous peat exceeds 0.9 m³/m³; skeletal subsurface tiers consist for only 5 to 7 volume percent of solid matter! Values measured on the peat dome of Figure 5 are given in Table 1. Note the vertical and horizontal differentiation in packing density and its correlation with the vegetation type.

TABLE 1.

Total Pore Fractions (TPF = 1 - BD/SD) calculated for a peat dome near Pontianak, Indonesia.

		ed Swamp Forest (dome fringe)	•		Monotonous Forest (dome center)		
		13	11.4	10.9 9.4		8.0	
10-20 cm	BD:	0.20	·0.15	0.13	0.14	0.13	
	SG:	1.39	1.28	1.42	1.52	1.44	
	TPF:	0.86	0.88	0.91	0.91	0.91	
70-80 cm	BD:	0.23*	0.11	0.10	0.10	0.09	
	SG:	1.67*	1.24	1.35	1.48	1.29	
	TPF:	0.86	0.91	0.93	0.93	0.93	

clayey peat from shallow fringe area

Virgin peats can retain considerable quantities of water. However, if dried to the extent that adsorptive water is lost, irreversible changes occur in the colloidal components of the peat resulting in a marked and

^{**} See Figure 5.

permanent reduction of the water retention capacity. Overheated peat becomes hydrophobic and shrinks to a granular powder with unattractive physical and agricultural properties and a high sensitivity to erosion.

In peat areas under monotonous forest, where the peat consists for a large part of recognizable wood in various stages of decay, the peat material shrinks less than in the better decomposed fringe areas under mixed swamp forest. Fibric peat in particular has many wide pores. Its saturated hydraulic conductivity exceeds 1.6 m/d and may even be greater than 30 m/d. Well decomposed sapric peat has finer pores and lower hydraulic conductivity values.

NOTE THAT the above statements are generalizations to which there are numerous exceptions. Woody peats, for instance, are nearly always very permeable to water. Compacted (reclaimed/drained) peats have much lower Total Pore Fractions than virgin peats and stratified peats may be virtually impermeable, irrespective of their fiber content.

The loose structure and flexible peat fibers are also accountable for the low bearing capacity and poor trafficability of most peats. The low penetration resistance makes it difficult to use normal farm machinery and even light equipment may get stuck because of high rolling resistance and slip. The bearing capacity is a function of the soil moisture potential, internal friction and 'effective normal stress' (stress transmitted through the peat skeleton). Effective normal stress is influenced by the bulk density of the peat; the bearing capacity therefore increases upon reclamation (consolidation) of the peat.

Chemical characteristics

The wide variation in the physical characteristics of Histosols is matched by an equally wide variation in chemical soil conditions. Most peats are acid; mature ombrogenous peats are very acid (pH 3 to 4.5) and contain less than 5 percent inorganic constituents. The organic fraction consists of lignin, cellulose, hemicellulose and small quantities of proteins, waxes, tannins, resin, suberin, etc. Ombrogenous moss peats in temperate climates consist largely of cellulose whereas deep lowland forest peats from Indonesia appeared to consist for two-thirds of lignin with cellulose and hemicellulose accounting for only 1-10 percent of the dry sample weight. The cellulose contents of tropical topogenous peats may well be higher than the lignin contents (as in peats in temperate regions) but there is little quantitative information on tropical topogenous peats.

One important organic fraction in organic soil materials is not contained in fresh plant debris but is synthesized in the course of microbial transformation of the organic materials: 'humic substances', a mixture of humins and humic and fulvic acids. Humic substances form stable complexes with metal ions. These are easily leached out of the peat mass with effluent water. (Geographic names such as Rio Negro, Blakkawatra, Cola creek, Air hitam, Zwarte water, and many more testify of the constant leaching of organic compounds from peat bogs.) Table 2 presents ranges in total microelement contents as observed in a number of deep and extremely poor Indonesian peat soils. The higher element contents of surface tiers reflect

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the cycling of elements by the vegetation. Note that a considerable part of all micro-elements in the system are stored in the vegetation; the values presented in Table 2 are by no means indicative of the total quantities present immediately after felling and burning of the (forest) vegetation.

TABLE 2. Micronutrient contents (kg/ha) measured in surface and subsurface tiers of deep ombrogenous forest peats from Indonesia.

	0 - 25 cm80 - 100 cm		
Cobalt	0.1-0.2	0.05-0.1	
Copper	0.8-8.0	0.2-0.8	
Iron	143-175	67-220	
Manganese	4.1-25	1.1-7.1	
Molybdenum	0.6-1.0	0.3-0.6	
Zinc	2.8-4.4	1.8-4.8	

With the respective differences considered, the same applies to the levels of macro- and secondary nutrients in (virgin) peat soils: element contents are highest in the top 25 cm of the soil. Felling of the natural forest (part of the 'reclamation' process of peat lands) upsets this pattern because of interrupted nutrient cycling, release of nutrients from decaying organic materials and biomass, increased leaching of nutrients, volatilization upon heating (burning) of the peat, etc.

TABLE 3. Nitrogen and total ash contents and C/N-ratios in a moderately deep (A) and a deep (B) Fibric Histosol from the Netherlands. Source: Dinc et al, 1976.

	·			 -		
DEPTH	BD ~	pH(H ₂ O)	TOTAL ASH	CARBON	NITROGEN	C/N
(cm)	(Mg/m ³)		(%)	(%)	(%)	
A						
0-7	0.20	3.1	5.8	54.3	1.01	53.8
7-18	0.14	3.0	3.1	58.8	1.03	57.1
18-92	0.12	3.3	2.7	62.9	1.13	55.7
92-117	0.10	4.3	3.9	61.7	1.55	39.8
117-127	0.14	4.9	84.2	6.8	0.22	30.9
В						
0-7	0.24	3.1	8.8	56.1	1.28	45.8
7-24	0.15	3.1	1.2	57.1	0.69	82.7
24-35	0.14	3.1	0.8	62.7	0.82	76.5
35-55	0.13	3.0	1.2	57.9	0.80	72.4
55-86	0.10	3.2	5.8	53.7	0.64	83.9
86-120	0.13	3.0	1.4	61.3	0.96	63.9
120-142	0.10	3.1	1.0	64.1	0.93	68.9
>142	n.d.	3.4	3.1	56.9	0.65	87.5

The contents of 'total ash', K_2O , P_2O_5 and SiO_2 of the surface soil decrease after clearing of the forest vegetation whereas CaO and MgO contents tend to increase. The quantities of Na, Cl and SO_4 depend strongly on local conditions such as the distance from the sea and the presence of pyrite in the peat (see also the chapter on Fluvisols).

The quantities of nitrogen contained in the peat are of the order of 2,000 to 4,000 kg N/ha per 0-20 cm, of which only a very small portion is available to plants. Table 3 illustrates the foregoing with data from two Fibric Histosols from The Netherlands (note the high C/N-ratios of the subsurface peat).

Figure 6 presents the distribution of mineral constituents in a young and a mature Fibric Histosol from Kalimantan. As in the Dutch Histosols, nutrient elements are concentrated (and available) in the upper 10 cm of the soil where a dense root mat occurs, whereas living roots are absent (and ash contents low) in deeper layers.

Virgin shallow or moderately deep peats are less depleted of plant nutrients than deep peats. Deficiencies are most severe in genetically old peats, with the exception of topogenous peats which receive nutrients from outside and cannot be included in any generalized account of the chemical properties of peat soils.

MANAGEMENT AND USE OF HISTOSOLS

Natural peats need to be drained to permit the cultivation of dryland crops. Centrally guided reclamation projects are (almost) exclusive to the temperate zone where millions of hectares have been drained. Tropical peat lands are more commonly opened by individual farmers although some medium-sized government sponsored projects have been carried out as well. Despite small capital inputs, farmers are sometimes more successful than government-inspired reclamation projects because they are more flexible in the selection of their locations and use less advanced but time-tested techniques.

The reclamation of peat lands starts nearly always with the construction of shallow drainage ditches. As a rule, the vegetation is left standing because it helps drying of the peat. One-meter-deep drains at 20-40 m intervals are satisfactory in most cases but well decomposed or clayey peats may require narrower spacings whereas undecomposed woody dome peats could in some cases be drained with ditches 100 m apart. A complex drainage system is not advisable because land subsidence is likely to disturb the connections between sucker drains and collecting drains. In practice, draining peat lands is a matter of experience and the standard formulas applicable to mineral soils are of little value.

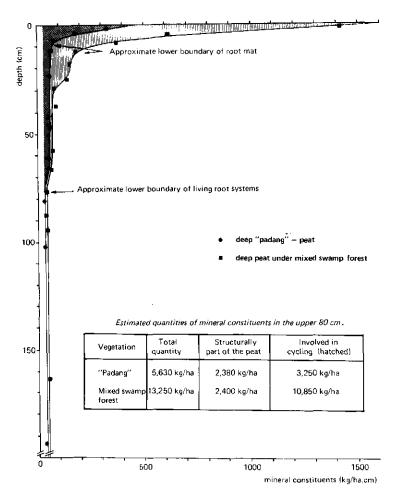


Fig. 6. Distribution of mineral constituents in two deep virgin ombrogenous forest peats in Kalimantan.

After some time of operation, the drainage system will have to be adjusted because peat properties change. The soil's hydraulic conductivity might decrease in the course of drainage because of the collapse of large pores, the formation of an illuvial horizon, or the effects of tillage. On the other hand, the soil may actually become more permeable after some time because of decaying wood providing passage for the escape of water or because of increasing biological activity (roots, animals) or the formation of cracks. Farmers in the tropics prefer shallow hand-dug drainage ditches that can be deepened as needed.



Fig. 7. Controlled burning of peat to raise the pH of the surface soil and 'liberate' nutrient elements before planting.

It is difficult to say exactly <u>when</u> peat reclamation is completed. Reclamation and cultivation overlap and in most instances cropping is even part of reclamation. In the first few years after the removal of the natural vegetation, suitable annual crops may produce fair yields, thanks to the nutrients that are still contained in the surface soil (and in the ashes of burnt plant material). The uneven distribution of these nutrients over a field explains the irregular growth typical of young reclamations. After a few years, when subsidence of the land surface has slowed, trees and shrubs can be planted. They may grow satisfactorily for some time but yields will eventually decrease if the land is not fertilized.

Where fertilizers are not sufficiently available, farmers often turn to some sort of controlled burning of the past to 'liberate' putriouts and

Where fertilizers are not sufficiently available, farmers often turn to some sort of controlled burning of the peat to 'liberate' nutrients and to raise the pH of the surface soil. See Figure 7.

Burning has undoubtedly a stimulating effect on plant performance, but the desirability of burning and its precise effects are still open to discussion. Those in favour of controlled burning claim that it is not more destructive than oxidation in the long run but concentrates certain nutrients (N, P, K, Ca, Mg, S) and renders them more available to the

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plant. Others are of the opinion that burning should be discouraged altogether because most of the liberated nitrogen and sulphur are lost to the atmosphere and other nutrients are largely leached out of the surface soil. The overall deterioration of the soil structure (the burnt layer is usually by far the best part of the profile) and the resulting uneven soil surface are additional arguments against burning.

Liming and full fertilization are needed for good yields on ombrogenous peat. In addition to massive applications of ground (magnesium) limestone, N, P and K fertilizers must be applied together with small doses of sulfur, copper, zinc, manganese, molybdenum, and iron.

The reclamation of peats is further complicated by the fact that the loose and perishable nature of most peats leads to considerable subsidence of the land surface once the land is drained and the suspending force of the groundwater removed. The total loss in surface elevation is determined by the combined effects of consolidation of the peat mass (through settlement, shrinkage and compaction of the peat), and of mineralization of the peat through biochemical oxidation and burning. Removal of peat, e.g. by wind erosion, is a problem in many Histosol areas in the temperate zone.

Initial subsidence starts directly after the water table is lowered and can reach values of one meter and more in the first year in deeply drained ombrogenous peats. This high rate levels off to a much lower continuous subsidence after a number of years. The initial subsidence is primarily caused by consolidation of the peat mass after the loss of groundwater buoyancy and - in the case of woody forest peats - the decay of the roots and timber skeleton of the peat. Continuous subsidence is mainly a matter of mineralization of the peat and represents a true loss of soil material rather than a loss of soil volume. In no case should drainage of the land be deeper than strictly necessary.

MINERAL SOILS CONDITIONED BY HUMAN INFLUENCES: **ANTHROSOLS**

ANTHROSOLS

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

Key to Anthrosol (AT) Soil Units

Anthrosols showing remnants of diagnostic horizons due to deep cultivation.

Aric Anthrosols (ATa)

Other Anthrosols having a fimic* A-horizon.

Fimic Anthrosols (ATf)

Other Anthrosols having an accumulation layer of fine sediments, thicker than 50 cm and resulting from long-continued irrigation or artificial raising of the soil surface.

Cumulic Anthrosols (ATc)

Other Anthrosols having, to a depth of more than 50 cm, admixtures of wastes from mines, town refuses, fills from urban developments, etc.

Urbic Anthrosols (ATu)

SUMMARY DESCRIPTION OF ANTHROSOLS

<u>Connotation</u>: soils whose most prominent characteristics result from the activities of man; from Gr. <u>anthropos</u>, man.

<u>Parent material</u>: various parent materials, modified by man through deep cultivation or through addition of material from elsewhere.

<u>Environment</u>: Anthrosols occur in all environments; Fimic Anthrosols are most common in north-west Europe where they occur in areas of quartzitic sands with a long history of mixed farming.

<u>Profile development</u>: the influence of man is mostly restricted to the surface horizon(s); a buried soil can still be intact at some depth and testify of soil conditions as existed before Man modified the land. In recent reclamations, heavy equipment often destroyed all previous soil development to a great depth or it buried existing soil profiles under thick layers of allochtonous material.

<u>Use</u>: European Anthrosols were traditionally grown to winter rye, oats, and barley; Fimic Anthrosols were valuable tobacco soils. At present, they are

^{*} Diagnostic horizon; see Annex 1 for full definition.

36 Anthrosols

used for forage maize, potatoes, horticultural crops (strawberries) and tree nurseries; in places also for pasture, or for vegetable and flower production in greenhouses. Cumulic Anthrosols in irrigation areas are grown to cash crops or food crops.

REGIONAL DISTRIBUTION OF ANTHROSOLS

Small areas of Anthrosols are found wherever people have lived for a long time. Fimic Anthrosols are the best known man-made soils; they occupy a total of some 0.5 million hectares in western Europe (Holland, Belgium, Federal Republic of Germany; see Figure 1.); smaller occurrences are found along the english coast and in Scotland, Wales and Ireland.

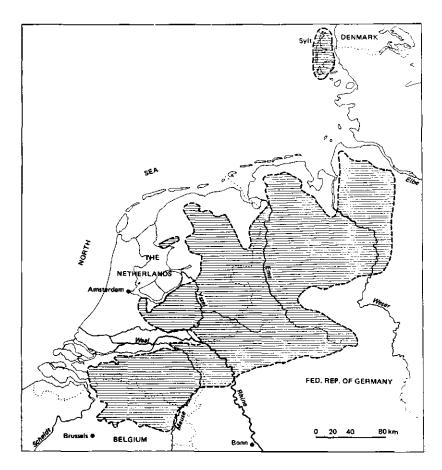


Fig. 1. Fimic Anthrosols in northwestern Europe. Source: Pape, 1970.

GENESIS OF ANTHROSOLS

Aric, Cumulic and Urbic Anthrosols were mostly formed by short term human interference such as deep cultivation, irrigation or 'landscaping' in the context of land reclamation, and regional development and/or urbanization schemes.

<u>Aric Anthrosols</u> formed as a result of deep cultivation. In Europe, for instance, sandy Podzols with compacted or cemented B-horizons at shallow depth were widely mixed and are now Aric Anthrosols. Deep cultivation of land with a shallow histic topsoil over sand produced Aric Anthrosols with alternating oblique layers of peat and sand, that are excellently suited for arable cropping (the peat layers act as a sink in times of wetness whereas the sand layers keep capillary rise going in times of drought).

Most Cumulic Anthrosols formed through prolonged sedimentation of silt from irrigation water. In depressed areas, trees are commonly planted on man-made ridges that alternate with drainage furrows. Eventually, the original soil profile of the ridge areas is buried under a thick cover of soil and mud. The ridge-and-furrow system is known from such different environments as the wet forests of western Europe and the coastal swamps of southeast Asia where the ridges are planted to dryland crops and rice is grown in the shallow ditch areas (the 'sorjan' system; see also the chapter on Fluvisols).

The Urbic Anthrosols are a very heterogeneous taxonomic unit. Well known examples are soils in mine spoils, or in filled-in open pit mining areas and soils in wetlands that were covered with layers of sand as part of urbanization schemes.

<u>Fimic Anthrosols</u> have a characteristic diagnostic horizon: the fimic A-horizon, produced by long continued addition of a mixture of organic manure and earth. The man-made character of the fimic A-horizon is evident from bits of brick and pottery and/or from high levels of extractable phosphorus (normally more than 250 mg P₂O₅ per kg by 1 percent citric acid).

The formation of most fimic A-horizons started in the Middle Ages when inorganic fertilizers were virtually nonexistent. Where natural soil fertility was insufficient for sustained production of arable crops, the farmers used a system of mixed farming in which arable fields were combined with sheep grazing on communal pasture land. During the nights and in wintertime, sheep and cattle were kept in stables with bedding material of thin sods of heath and/or forest litter to absorb the valuable dung. Fresh bedding material was regularly provided until the bedding became too thick and had to be removed. It was then spread out on the fields as an 'organic earth manure'.

Evidently, this manure could decompose only in part; the mineral fraction remained behind, with some of the organic matter, and raised the land surface by some 1 millimeter per annum. In places, this system was in use for more than one thousand years and produced overthickened fimic A-horizons of more than 100 cm.

Depending on the composition of the bedding material, the fimic A-horizon is black (bedding material from heathlands with Podzols) or brown (from forest litter) in colour. In places, sods from low-lying grasslands were incorporated in the earth manure. This gave the A-horizon a somewhat higher clay content than the deeper solum. Some 10 hectares of heathland were needed to maintain the nutrient level of one hectare of arable land. Removal of the sods made the heathland susceptible to wind erosion and not seldom large tracts of heathland turned to barren shifting sands that went completely out of control (see also under Arenosols).

Most arable fields were situated on favourable sites near villages and were well drained even before acquiring a fimic A-horizon. They were selected, among other reasons, on account of a lower frost hazard to the winter rye crop; only in areas with a high population density were less well drained soils also used. This arrangement of arable fields on well drained (higher) terrain positions and pastures in nearby depressions can still be seen in Anthrosol areas in western Europe (see Figure 2).



Fig. 2. Mixed farming near Winterswijk, The Netherlands. Arable cropping on well drained Fimic Anthrosols (in foreground) is combined with dairy farming in depressed areas. Source: Van de Westeringh, 1979.

The mixed farming system described produced the world's largest continuous area of Fimic Anthrosols (see Figure 1). Different types of fimic

A-horizons occur elsewhere in western Europe, formed for instance by gradually covering peat soils with layers of sand, bagger from drainage ditches, and organic manure.

In parts of Ireland and England, calcareous beach sands were carted to areas with acid Arenosols, Podzols, Podzoluvisols and Histosols which thus became Fimic and/or Cumulic Anthrosols. Other examples of Fimic Anthrosols are found along river terraces in southern Maryland, U.S.A., where deep, black A-horizons formed in layers of kitchen refuse (mainly oyster shells) from early Indian habitation. Similar soils occur along the Amazon river in Brasil ('Terra Preta'). Many countries possess small areas of soils that were modified by early inhabitants.

CHARACTERISTICS OF ANTHROSOLS

Anthrosols cannot be characterised by any particular arrangement of soil horizons; the only horizon that is diagnostic is the fimic A-horizon of Fimic Anthrosols. This surface horizon has a thickness of 50 cm or more, a black or brown colour, and is in almost all cases free of mottles. The surface horizon of an underlying buried soil may be incorporated in the (lower part of the) A-horizon; spade marks may still be detectable there.

Hydrological characteristics

Fimic Anthrosols are well drained because of their overthickened A-horizon and because the system of earth manuring was only practiced on selected well drained sites. The internal drainage of the fimic A-horizon is excellent because of its sandy texture and high porosity (high biological activity); aeration is never a problem. Soil units other than Fimic Anthrosols range from well drained to poorly drained depending on their texture, their porosity and their position in the landscape.

Physical characteristics

The physical characteristics of the fimic A-horizon are excellent: the total pore fraction is nearly always close to $0.5~{\rm cm}^3/{\rm cm}^3$, pores are interconnected and of different sizes, the penetration resistance is low and allows unhindered rooting and, last but not least, the storage capacity of 'available' soil moisture is high if compared to that of the underlying soil material. The type of organic matter ('moder') furthers stable soil structures and limits the susceptibility to slaking. The upper part of a fimic A-horizon may become somewhat dense if tillage is done with heavy (vibrating) machinery.

Chemical characteristics

Most fimic A-horizons have pH(KCl) values between 4 and 4.5, except where foreign calcareous material was brought into the soil (e.g. sea sand). The organic carbon contents of fimic A-horizons are typically between 1 percent and 5 percent. The organic carbon content of black fimic A-horizons is higher than of brown ones; the C/N ratio is generally between 10 and 20, with the higher values in black soils. The CEC depends largely on the

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quantity of organic matter in the soil and on its composition; reported CEC values are between 5 and 15 $\rm cmol(+)/kg$ soil. Characteristic of all fimic A-horizons is the high total phosphorus content which, in places, built up to 2000 mg/kg and more. Other elements are often in short supply (N, K, Mn, B, Cu).

MANAGEMENT AND USE OF ANTHROSOLS

Aric, Cumulic and Urbic Anthrosols vary widely in properties and requirements and cannot be included in any generalized account of the management and use of Anthrosols. Fimic Anthrosols are characterised by favourable physical properties (porosity, rootability, moisture availability) and rather unsatisfactory chemical properties (acidity, nutrients). Rye yields obtained without the use of fertilizers were 700 to 1100 kg per hectare, or 4 to 5 times the quantity of seed used. The good drainage and the dark colour of the surface soil make it possible to till and sow or plant early in the season. Rye, oats, barley, potato and also the more demanding sugar beet and summer wheat are common crops on Fimic Anthrosols. In places, the soils are used for tree nurseries and horticulture, notably for vegetable, small fruit and flower production. Until the 1950's, soils with deep fimic A-horizons in The Netherlands were in demand for the cultivation of tobacco. With modern fertilization, yields of 5000 kg rye per hectare, 4500 kg barley and some 5500 kg summer wheat can be obtained. The production figures for sugar beet and potato are of the order of 40 to 50 tons per hectare. Fimic Anthrosols are increasingly used for production of silage maize and grass; yields of 12 to 13 tons of dry maize silage per hectare and of 10 to 13 tons of dry grass are considered normal.

MINERAL SOILS CONDITIONED BY THE PARENT MATERIAL:

ANDOSOLS ARENOSOLS VERTISOLS

MAJOR LANDFORMS IN VOLCANIC REGIONS

Volcanic regions differ from all other regions in two ways:

- (1) by their structure and origin, and
- (2) by the particular composition of the volcanic materials.

The world distribution of active volcanoes is shown in Figure 1. The chemical and mineralogical composition of volcanic materials is marked by an abundance of easily weatherable components, accountable for the remarkable properties that soils in volcanic regions have in common. In spite of the overriding importance of the parent material for soil formation in volcanic regions, different soils (may) occur in different landforms.

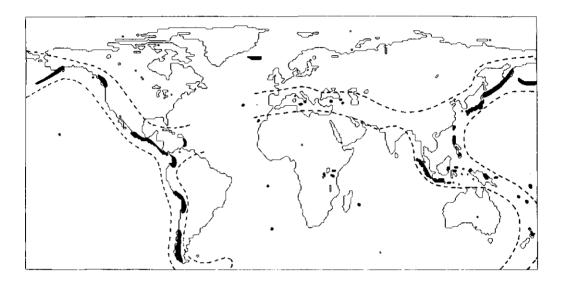


Fig. 1. Volcanic regions of the world. Note that most volcanoes are found near plate margins, in zones of young orogeny. Based on: Robinson, 1975

As magma composition has a strong bearing on the nature of the erupted products but also on the explosivity, character, and morphology of volcanic phenomena, we shall discuss the major landforms of volcanic regions by compositional group, viz. for regions with basaltic, andesitic and rhyolitic volcanism.

LANDFORMS IN REGIONS WITH BASALTIC VOLCANISM

Basaltic volcanism is most commonly found in situations of divergent plate margins, in continental rift valleys, and in hot-spot areas. The classic example of basaltic hot-spot volcanism is Hawaii, the largest shield volcano in the world, with a diameter of 250 km at the base (on the ocean floor) and a total height of 9 km. Basaltic magma is comparatively fluid and gases escape easily. Therefore, eruptions are relatively quiet

and produce lava flows, lava lakes and lava fountains, but little ash.

Basaltic volcanism in continental settings (e.g. the East-African rift valley, the Baikal graben, or the Rhine - Rhone and related rift valleys) did produce 'strombolian' cones and maare but, here too, ash production is limited and localized. The ash deposits extend seldom outside the volcanic areas themselves, and where extensive ash blankets occur, as in some rift valleys, they are usually more acid in composition.

Basaltic lava flows, because of their fluidity, may follow river valleys over considerable distances from the eruption centres. Later differential erosion may result in relief inversion with the former basaltic valley fills extending as plateaus above the surroundings.

LANDFORMS IN REGIONS WITH ANDESITIC VOLCANISM

Andesitic volcanism is a characteristic element in 'Cordillera'-type mountain belts such as the Andes and the Barisan range in Indonesia, and in island arcs. The classic volcano type is the stratovolcano. Literally, the term means 'stratified' volcano, which is misleading in the sense that all volcanoes are built up of layers, be it of basalt flows, as in the Hawaiian shield volcanoes, or of pyroclastics, as the scoria cones of the Eifel. What the term indicates, actually, is that this type of volcano is composed of alternating layers of lavas and pyroclastic rock, mostly of andesitic composition.

Andesitic magmas hold an intermediate position between basaltic and rhyolitic magmas with regard to their SiO₂ content, viscosity and gas content. Whereas basaltic, low viscosity magmas hardly produce pyroclastics ('tephra'), and high-viscosity rhyolites hardly produce lavas, andesitic magmas will normally produce both. Because of the greater viscosity of the magma, a higher pressure must build up before an eruption can occur; eruptions are less frequent and more violent than in basaltic volcanism.

Lava flows emitted by stratovolcanoes are more viscous than those of basaltic shield volcanoes, and do not extend as far from their point of emission, usually not more than a few kilometres. This explains why stratovolcanoes have the 'classical' cone shape and steeper slopes than shield volcanoes.

Lahars are volcanic mudflows. They can form in several ways:

- (1) when a crater lake is thrown out during an eruption, or
- (2) when numerous condensation nuclei in the air (volcanic ash) generate heavy rain showers, or
- (3) when the volcano was covered by snow or glaciers before the eruption. The term lahar stems from Indonesia, where volcanic mudflows of the first type were particularly disastrous.

Pyroclastic flows are frothy masses of ash and pumice, deposited by glowing avalanches. The resulting rocks are known as **ignimbrites**. They can have a variety of structures depending on the flow conditions during emplacement and on the degree of postdepositional welding.

Volcanic air-fall ashes are widespread in volcanic regions. Whereas lavas and pyroclastic flows are confined to the immediate vicinity of volcanoes, ashes are blown over vast distances and travel hundreds of kilometres. The thickness of ash deposits decreases with increasing distance from the point of origin. It may be difficult to recognize the presence of volcanic ashes

in soils as it is incorporated in the solum by burrowing animals and weathers rapidly. Nonetheless, 'rejuvenation' of a soil material with fresh volcanic ash is often of great importance as it restores or improves soil fertility and promotes physical soil stability.

LANDFORMS IN REGIONS WITH RHYOLITIC VOLCANISM

Partial melting of the continental crust in Cordilleran mountain ranges may produce rhyolitic magma. Rhyolites represent the extrusive equivalents of granites; they are acid volcanic rocks with a high content of SiO2. Rhyolitic magmas are therefore very viscous and withstand very high gas pressures. As a result, rhyolitic eruptions are rare, but also extremely violent. If a rhyolitic magma chamber is present below a stratovolcano, it builds up such tremendous pressures that, once a vent for eruption is opened, the magma chamber empties itself completely, leaving a cavity in the earth's crust in which the entire stratovolcano collapses. In this way, craters of several kilometres in diameter are formed: the calderas. The main emission products are ashes, in astonishing quantities and spread over vast areas, and ignimbrites. The latter stem from pyroclastic flows of a fluid, glass-rich mass; the flows (can) extend over several tens of kilometres and fill depressions and valleys to depths of tens or even hundreds of metres. In contrast with the irregular surfaces of lava flows and lahars, ignimbrite surfaces are completely flat and featureless. White, porous and fibrous pumice inclusions are common.

Both ashes and ignimbrites consist for the greater part of volcanic glass and weather rapidly. Phenocrysts (mainly biotite and hornblende) make up less than 20 percent of the ash. The only historic ignimbrite-forming eruption was that of the Katmai in Alaska in 1912. The eruption of the Krakatoa, in 1883, did produce a caldera, but as the main eruption took place under water, ignimbrite formation was not conspicuous on the land surface. The largest eruption in comparatively recent times took place some 40,000 years ago and led to the formation of Lake Toba in Sumatra.

Rapid weathering of porous volcanic material in environments with sufficient rainfall, translocation of (part of) the weathering products and accumulation of stable organo-mineral complexes and of short-range-order minerals such as allophane, imagolite and ferrihydrite are essential processes in the formation of the characteristic soils of volcanic regions: the ANDOSOLS.

NOTES

ANDOSOLS

Soils showing andic^a 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; lacking gleyic^a properties within 50 cm of the surface; lacking the characteristics which are diagnostic for Vertisols; lacking salic^a properties.

Key to Andosol (AN) Soil Units

Andosols having permafrost within 200 cm of the surface.

Gelic Andosols (ANI)

Other Andosols showing gleyic^a properties within 100 cm of the surface.

Gleyic Andosols (ANg)

Other Andosols lacking a smeary consistence^a, or a texture which is silt loam or finer on the weighted average for all horizons within 100 cm of the surface, or both.

Vitric Andosols (ANz)

Other Andosols having a mollic* A-horizon.

Mollic Andosols (ANm)

Other Andosols having an umbric * A-horizon.

Umbric Andosols (ANu)

Other Andosols.

Haplic Andosols (ANh)

SUMMARY DESCRIPTION OF ANDOSOLS

Connotation: soils in volcanic materials; from Jap. an, dark, and do, soil.

<u>Parent material</u>: pyroclastic material, notably volcanic ash, but also tuff, pumice, cinders and other volcanic ejecta of various composition. Exceptionally, weathering of non-volcanic material may result in andic soil properties and hence in Andosols.

Environment: undulating to mountainous, humid, arctic to tropical climates
with a wide range of vegetation types.

<u>Profile development</u>: AC- or ABC-profile. Rapid weathering of the porous material results in accumulation of stable organo-mineral complexes and of short-range-order minerals such as allophane, imagolite and ferrihydrite.

Diagnostic horizon; see Annex 1 for full definition.

Diagnostic property; see Annex 2 for full definition.

<u>Use</u>: many Andosols are intensively cultivated with a variety of crops, their major limitation being the high phosphate fixation capacity. In places, steep topography is the chief limitation.

REGIONAL DISTRIBUTION OF ANDOSOLS

Andosols occur in volcanic regions all over the earth. Important concentrations are found near the Pacific Basin: on the west coast of South America, Central America, the Rocky Mountains, Alaska, Japan, the Philippine Archipelago, Indonesia, Papua New Guinea and New Zealand. They are also prominent on many islands in the Pacific: Fiji, Vanuata, New Hebrides, New Caledonia, Samoa, Hawaii.

In Africa, Andosols occur in Ethiopia, Kenya, Rwanda, Cameroon and Madagascar. Other regions with Andosols are the West Indies, Canary Islands, Italy, France, Germany, Iceland. The total Andosol area is estimated at some 124 million hectares or 0.84 percent of the global land surface. Some 80 percent of the Andosol area is potential crop land which corresponds with about 2 percent of the total potential crop land area. More than half of this is situated in the tropics. Figure 1 shows the worldwide distribution of Andosols.

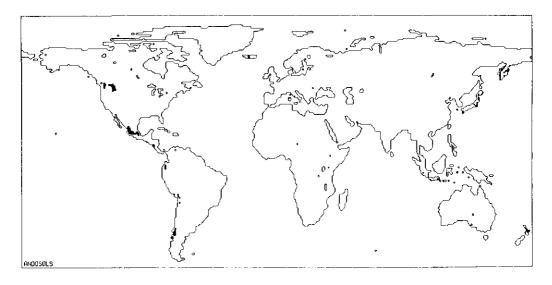


Fig. 1. Andosols worldwide.

GENESIS OF ANDOSOLS

Andosol formation depends essentially on the rapid chemical weathering of porous, permeable, fine-grained parent material containing 'volcanic glass' in the presence of organic matter. The hydrolysis of the primary

minerals microcline and augite may serve to illustrate this type of weathering ('glass' being an amorphous mixture with similar chemical composition and reacting in the same way):

Under relatively acid conditions (pH <5), this hydrolysis is fostered by protons from organic acids; the protons stem from carbonic acid if the pH is higher. Depending on the intensity of leaching, the liberated basic cations are to a great extent washed out whereas the Al- and Fe-ions, and most prominently the former, are tied up in stable complexes with humus. The ferrous iron is first oxidized to the ferric state and is then for the greater part precipitated as ferrihydrite¹:

$$Fe^{2+} = Fe^{3+} + e$$

$$Fe^{3+} + 3 H_2O = Fe(OH)_3 + 3 H^+$$
ferrihydrite
$$(or: 2 Fe^{2+} + \frac{1}{2} O_2 + 5 H_2O = 2 Fe(OH)_3 + 4 H^+)$$

The Al in the complexes protects the organic part against bio-degradation (toxic to micro-organisms). The mobility of these complexes is rather limited because rapid weathering yields sufficient Al and Fe to produce complexes with a high metal/organic ratio that are only sparingly soluble. This combination of low mobility and high resistance against biological attack promotes the accumulation of organic matter in the topsoil. By contrast, a similar combination in Podzols leads generally to metal-unsaturated complexes which are much more mobile.

The fate of the liberated silica depends largely on the extent to which Al is complexed by the humus. If most or all Al is complexed, the silica concentration of the soil solution will increase and while part of the silica is washed out, another part precipitates as opaline silica. If not all Al is complexed, the remainder may co-precipitate with Si to form

¹Ferrihydrite is the approved new name for hydrous iron oxides of short-range-order previous-ly termed 'amorphous ferric oxide' or 'iron oxide gel'. Neither its structure nor its composition has yet been established beyond doubt; a good approximation is probably: Fe₂O₃.2FeOOH. 2.6H₂O.

Ferrihydrite is the dominant iron oxide mineral of most volcanic soils and some of the properties ascribed to allophane may in part be due to ferrihydrite. Recent evidence suggests that much, if not all, of the organically bound Fe (as extracted by pyrophosphate) is ferrihydrite-Fe.

allophane² of varying composition and often imagolite³, as well as, under certain conditions, halloysite. The formation of Al-humus complexes and the formation of allophane associations are mutually competitive. Both groups appear to occur in inverse proportions; this constitutes the binary composition of Andosols. It seems that allophane and imagolite are stable under mildly acid to neutral conditions (pH >5) whereas Al-humus complexes are dominant under more acid conditions (pH <5). If there is any Al still available under the latter conditions, this may combine with excess Si to form 2:1 and 2:1:1 type phyllosilicate clay minerals (e.g. chlorite) as are often found in association with Al-humus complexes. Because of the acid conditions, these soils may have exchangeable Al which is not found on allophane. The stability conferred on the organic matter by Al is no less in the presence of allophane. Apparently, the activity of Al in allophane is high enough to interact with organic molecules and prevent bio-degradation and leaching.

The described competition between humus and silica for Al is influenced by a number of factors:

- (1) The Al-humus complex + opaline silica + phyllosilicate clay association is most pronounced in acidic types of volcanic ash that are subject to strong leaching. Primary quartz, a typical indicator of acid parent materials, is often present in the Al-humus complex association. In practice, there is a continuous range in the binary composition of Andosols from a pure Al-humus complexes association ('non-allophanic') to a pure allophane/imogolite association ('allophanic') in which the extremes are rare. This variation occurs both within one profile and between profiles.
- (2) In the very early stage of Andosol formation under humid temperate conditions, (near-)complete Al-complexation by organic matter may constrain the formation of allophane. Only when humus accumulation levels off is aluminium becoming available for mineral formation. This explains why B-horizons in Andosols are usually much richer in allophane and imagolite than A-horizons: the weathering of primary minerals proceeds but the supply of organic matter is limited so that Al is hardly used for complexation.

During the intensive weathering processes, the pore space of the already quite porous material is greatly increased, typically from some 50 volume percent to more than 75 percent, because of losses by leaching and strong stabilization of the remaining material by organic matter and weathering

²Allophanes are non-crystalline hydrous aluminosilicates with Al/Si molar ratios typically between 1 and 2 (Al/Si ratio of kaolinite is 1). They consist of hollow spherules with a diameter of 3.5 - 5 nm and have a very large (reactive) specific surface area.

³Imogolite is a paracrystalline aluminosilicate consisting of smooth and curved threads with diameters varying from 10 to 30 nm and several thousands of nm in length. The threads consists of finer tube units of 2 nm outer diameter; their outer wall consists of a gibbsite (Al) sheet and their inner wall of a silica sheet. The Al/Si molar ratio is 2.

products (silica, allophane, imogolite, ferrihydrite), some of which have quite open structures themselves. Biologic upgrading of the pore space is of less importance in Andosols.

The genesis of Andosols is often complicated by repeated deposition of fresh ash. This may involve surficial rejuvenation of the soil material with thin ash layers but also complete burial of the profile. A new profile will then develop in the fresh deposit while soil formation in the buried A-horizon is affected by the sudden decrease in organic matter supply and a different composition of the soil moisture.

When Andosols get older, the clay mineralogy changes, particularly that of the subsoil, as allophane and imogolite are transformed to halloysite, kaolinite or gibbsite, depending on the silica concentration of the soil solution. In addition, Al from the humus complexes will gradually become available and ferrihydrite turns into goethite. Evidently, this process is strongly influenced by such factors as the rate of rejuvenation, the depth and composition of the overburden, the composition of the remaining material and the moisture regime. Eventually, an Andosol may grade into a 'normal' soil, e.g. a podzolic soil, or a soil with ferric properties, or with clay illuviation.

CHARACTERISTICS OF ANDOSOLS

The typical Andosol has an AC or ABC profile with a dark Ah-horizon from 20 to 50 cm thick (thinner or thicker occurs) over a brown B- or C-horizon. There is a clear difference in colour between the topsoil and the subsoil but colours are generally darker in cool regions than in tropical climates where net accumulation of organic matter is less. The organic matter content averages about 8 percent but may reach 30 percent in the darkest profiles. The A-horizon is very porous, very friable, fluffy, non-plastic and non-sticky, and has a crumb or granular structure. In the field, the soil material may be smeary and feel greasy or unctuous; it becomes almost liquid when rubbed ('thixotropy'). Ash layers are often clearly visible.

<u>Hydrological characteristics</u>

Because of their high porosity and their occurrence in high terrain positions, most Andosols have excellent internal drainage. Gleyic soil properties occur where the groundwater is present at shallow depth.

Mineralogical characteristics

The quantities of volcanic glass, ferromagnesian minerals (olivine, pyroxenes, amphiboles), feldspars and quartz in the silt and sand fractions of Andosol material varies with the locality. Some of the mineral grains may have a 'coating' of volcanic glass.

The mineral composition of the clay of Andosols varies with the genetic age of the soil, the composition of the parent material, the pH, the base status, the moisture regime, the thickness of overburden ash deposits, and the organic matter. The 'X-ray amorphous materials' in the clay fraction of Andosols include allophane and, less commonly, imagolite, and/or humus

complexes of Al and Fe together with opaline silica. Allophane/imogolite and Al-humus complexes may occur together although the two groups have conflicting conditions of formation. Besides primary minerals, other components may occur such as ferrihydrite, (disordered) halloysite and kaolinite, gibbsite and various 2:1 and 2:1:1 layer silicates and intergrades.

Physical characteristics

The high aggregate stability of Andosols and their high permeability to water make that these soils are not very susceptible to water erosion. Exceptions to this rule are highly hydrated types of Andosols when they dry out strongly (e.g. upon deforestation). The difficulty to disperse Andosol material gives problems in texture analysis; caution should be taken when interpreting such data.

The bulk density of Andosol material is low, not just in the surface horizon; it is typically less than 0.9 $\rm Mg/m^3$ but values as low as 0.3 $\rm Mg/m^3$ have been recorded in highly hydrated Andosols. The bulk density does not change much over a suction range of 1500 kPa (limited shrink and swell). Therefore, the bulk density in the field-moist condition can in practice be substituted for the bulk density at 33 kPa suction, which is a diagnostic criterion.

Unless the soil is very young, the moisture content at 1500 kPa suction is high. Yet, the quantity of 'available water' in Andosols (difference in water content retained at 33 and 1500 kPa suction) is generally not lower and often higher than in other mineral soils. Air-drying of Andosol material may irreversibly lower the water holding capacity, the ion exchange capacity, volume, liquid limit and plastic limit. This effect is stronger with the more hydrated types. Air-drying may also cause loss in cohesion of the particles; the resulting dust is very susceptible to wind erosion.

Chemical characteristics

Andosols are notorious for their highly variable exchange properties: the charge is dependent on pH and electrolyte concentration and, therefore, very similar to that described for the Ferralsols (see chapter on Ferralsols). A difference is that the negative charge of Andosols can reach much higher values than that of Ferralsols because of the high contents of soil organic matter and allophane. The CEC can reach values of 100 cmol(+)/kg soil or higher. The variation in charge as a function of the pH is given for some Andosol components in Figure 2; montmorillonite, having a dominantly permanent charge, is included for comparison.

Because of the variable charge properties, the base saturation values are also variable. With the exception of very young Andosols and Andosols in non-humid areas, the base saturation values are generally low because of strong leaching.

If the clay assemblage is dominated by allophane (and imogolite), the pH of the soil is normally higher (pH >5) than if the clay consists largely of Al- and Fe-humus complexes together with phyllosilicates (pH <5). In the latter case exchangeable Al is often present in quantities that are

toxic to plants. The strong affinity for phosphate ions ('chemisorption') is a diagnostic 'andic' property.

The strong chemical reactivity of Andosols has long been attributed to X-ray amorphous compounds of Al, Si and humus. It is more appropriate, however, to ascribe this Andosol characteristic to the presence of 'active aluminium' which may occur in various forms:

- in short-range-order or paracrystalline aluminosilicates such as allophane and imogolite.
- 2. as interlayer Al-ions in 2:1 and 2:1:1 layer silicates.
- 3. in Al-humus complexes, and
- 4. as exchangeable Al-ions on layer silicates.

The role of active Fe may often not be ignored but is generally considered of less importance than that of active Al.

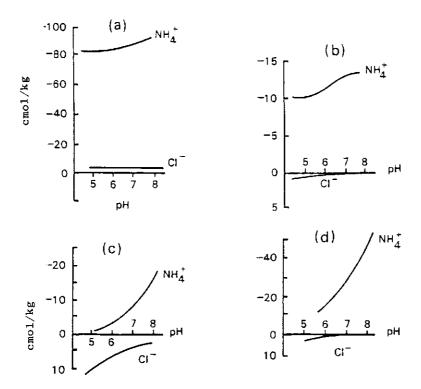


Fig. 2. NH₄⁺ and Cl⁻ retention curves using 0.01 M NH₄Cl (0.1 M NH₄Cl for montmorillonite). (a) Montmorillonite; (b) Halloysite; (c) Allophane 905 (Al:Si=2:1, containing some imagolite); (d) Allophane PA (Al:Si=1:1). After Wada & Okamura, 1977.

MANAGEMENT AND USE OF ANDOSOLS

Andosols have a high potential for agricultural production but many of them are not used to their capacity. Their natural fertility is high, particularly of soils developed in intermediate or basic volcanic ash and not exposed to excessive leaching. A major problem is the strong phosphate fixation of Andosols. Amelioration involves measures to reduce this effect (caused by active Al) through application of lime, silica, organic material and 'phosphate' fertilizer. Because of the strong buffering, raising the pH is difficult and expensive.

Andosols are easy to cultivate and have a good rootability and water availability. The most hydrated types, however, have only limited trafficability and soil may stick to the ploughshare during tillage.

Andosols are used for a wide variety of crops such as sugarcame, tobacco, sweet potato (little sensitive to phosphate fixation), tea, vegetables, wheat, orchard crops and forest (on steep slopes). Paddy rice cultivation requires that drainage is impeded by high groundwater or a dense subsoil layer.

MAJOR LANDFORMS IN REGIONS WITH SANDS

Sandy regions can be divided in two broad categories:

- (1) <u>residual sands</u>, formed upon weathering of old, usually quartz-rich soil material or rock, commonly under tropical conditions.
- (2) <u>shifting or only recently deposited sands</u>, e.g. in deserts and beach lands. (Regions with on-going deposition of fluvial sands will be discussed in the chapter on lowland formation.)

LANDFORMS IN RESIDUAL SANDS

Large expanses of horizontal sandstone plateaus in tropical shield areas have a deep weathering mantle of white sands. Well known examples are the Precambrian Roraima sandstones on the Guiana Shield and the Voltaian sandstone in Western Africa. They have in common with formations of unconsolidated fluvial sands, such as the Tertiary White Sands in Guyana and Suriname and their equivalents in eastern Perú, northeastern Brazil, or Liberia (western Africa), that they are all completely white, very rich in quartz, poor in clay, and excessively drained.

Most rivers on old sandstone plateaus and in white sand areas have black water of pH 4.0 or lower. The dark colour indicates that Fe-humus complexes are leached from the sands; these lose the iron coatings around the grains and become increasingly white. In fact, an extremely thick albic E-horizon is formed, with a spodic B-horizon either at several metres depth, or lacking altogether. However, because of the geologic dimensions of these processes, the albic horizons are parent material for the formation of 'normal' soils: ARENOSOLS.

LANDFORMS IN SHIFTING OR RECENTLY DEPOSITED SANDS

Sandy parent materials are also abundant in areas where sand accumulates by the selective action of wind or water.

Winds carry silt-sized particles over very large distances (Sahara 'dust' settles regularly in central Europe); sand-sized particles travel less far (by saltation) and fine gravel is only transported by creep. This explains why the purest sands, with a uniform particle size, are found in eolian deposits. Many such deposits show a characteristic large-scale cross bedding, indicative of sand deposition on the slip faces of dunes.

Fixed dunes are formed when transported sand settles in the lee of an obstacle such as a brush or a piece of rock. The obstacle thus grows in size and more sand settles: the dune grows. When the wind drives the sand grains to the top of the dune, its carrying capacity decreases so that an increasing part of the transported sand settles before reaching the crest of the dune. This steepens the angle of the slope, particularly near the crest. When the slope angle exceeds the angle of rest of the deposited sand (typically 34° for dry particles), shearing takes place along a slightly

56 Landforms in sands

less steep plane. Thus, a slip face is formed on the lee-ward side of the dune by a process of oversteepening and shearing.

Along coasts, the dunes are often 'parabolic dunes', shaped in part by landward migration of the dune strips.

Free dunes have no fixed position but migrate downwind by erosion on the gently inclined windward side and deposition on the leeward side (slip face) in the same way as described above. The smallest free dunes are ordinary wind ripples and measure only a few centimetres in height. Larger structures are indigenous to desert areas and large coastal sandflats.

If the winds are predominantly uni-directional, free dunes can assume a crescent shape ('barchan dunes'). As the rate of advancement of the sand is roughly inversely proportional to the height of the crest, the flanks of a shifting dune advance more rapidly than the central part until they become sheltered by the main mass of the dune. The speed at which barchan dunes move varies; values measured in the Middle East ranged between 5 and 20 metres per year. Figure 1 summarizes the process of barchan dune formation.

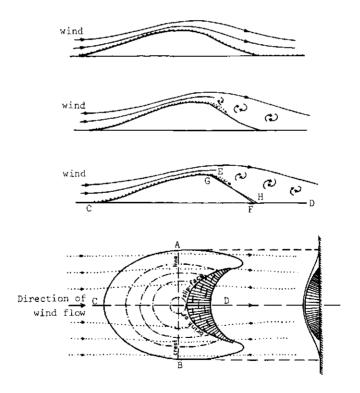


Fig. 1. The formation of barchan dunes. Source: Bagnold, 1965.

Coalescing barchans produce transverse dunes, whereas longitudinal dunes parallel to the wind direction ('seifs') are formed by spiralling 'screw-driver' winds. More or less stable star dunes ('ghourds') form with varying wind directions.

The area of dune formation is delimited by the 150 mm/yr isohyet. This boundary appears to have moved up and down in the recent past. Between 20,000 and 13,000 BP, the southern limit of active dune formation in the Sahara was 800 km south of its present position and most of the present Sahelian zone was an active dune area at that time. These dunes, mostly of the longitudinal type, are now fixed by vegetation, but their eolian parentage is still obvious. A similar story can be told for the Kalahari sands. Overgrazing in recent times has reactivated the eolian transport of these sands.

The cover sands and parabolic dunes of the temperate climate zone formed under periglacial conditions during the same arid interval between 20,000 and 13,000 BP. Forest, notably pine forest, re-established itself on most of these sands but overgrazing by sheep in medieval times led to renewed (wind) erosion, so that young anthropogenic dunes with ARENOSOLS are found in many parts of northwestern Europe. See Figure 2.

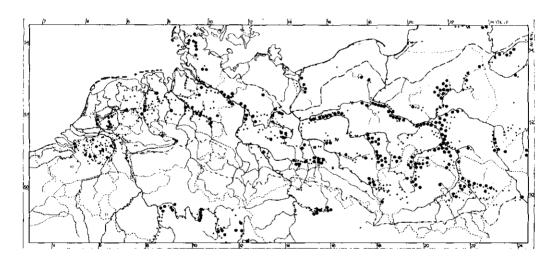


Fig. 2. Cover sands in Europe, Source: Koster, 1978.

NOTES

ARENOSOLS

Soils which are coarser than sandy loam to a depth of at least 100 cm of the surface, exclusive of materials which show fluvic^a or andic^a properties; having no diagnostic horizons other than an ochric^{*} A-horizon or an albic^{*} E-horizon.

Key to Arenosol (AR) Soil Units

Arenosols showing gleyic[®] properties within 100 cm of the surface.

Gleyic Arenosols (ARg)

Other Arenosols having an albic * E-horizon with a minimum thickness of 50 cm within 125 cm from the surface.

Albic Arenosols (ARa)

Other Arenosols which are calcareous $^{\hat{\omega}}$ at least between 20 and 50 cm from the surface.

Calcaric Arenosols (ARc)

Other Arenosols showing some clay increase or lamellae of clay accumulation within 125 cm of the surface.

Luvic Arenosols (AR1)

Other Arenosols showing ferralic^a properties within 125 cm of the surface.

Ferralic Arenosols (ARo)

Other Arenosols showing colouring or alteration characteristic of a cambic * B-horizon.

Cambic Arenosols (ARb)

Other Arenosols.

Haplic Arenosols (ARh)

SUMMARY DESCRIPTION OF ARENOSOLS

Connotation: weakly developed soils of coarse texture; from L. arena, sand.

<u>Parent material</u>: mostly unconsolidated, in places calcareous, translocated material of sand texture; in the wet tropics, relatively small areas of Arenosols occur on residual, intensively leached sandstone weathering.

<u>Environment</u>: from arid to (per)humid and from extremely cold to extremely hot; landforms vary from recent dunes, beach ridges and sandy plains under a scattered (mostly grassy) vegetation, to very old plateaus under a light forest.

Diagnostic horizon; see Annex 1 for full definition.

a Diagnostic property; see Annex 2 for full definition.

<u>Profile development</u>: A(E)C profiles. In the dry zone, an ochric A-horizon is the only diagnostic horizon. Arenosols in the perhumid tropics often have a thick albic E-horizon whereas most Arenosols of the humid temperate zone show signs of alteration or transport of humus, iron or clay, but too weak to be diagnostic at the Major Soil Grouping level.

<u>Use</u>: most Arenosols in the dry zone are used for little more than extensive grazing but they could be used for arable cropping if irrigated. They are permeable soils with good workability but their low organic matter content, low CEC, low water holding capacity and low coherence require adapted management. Arenosols in the temperate zone are moderately suitable for mixed arable cropping and grazing; supplemental (sprinkler) irrigation is needed during dry spells. Arenosols in the perhumid tropics are chemically exhausted and highly sensitive to erosion. They are best left under their natural vegetation.

REGIONAL DISTRIBUTION OF ARENOSOLS

Arenosols cover some 400 million hectares, mainly in the dry zone, viz. in the southern Sahara, southwest Africa and western Australia. Several million hectares of highly leached Arenosols are found in the perhumid tropics, notably in South America (Brazil, the Guianas) and in parts of southeast Asia (e.g. Sumatra, Kalimantan, Sarawak, The Pilippines). Small areas of (young) Arenosols occur in all parts of the world; they are not represented in Figure 1.

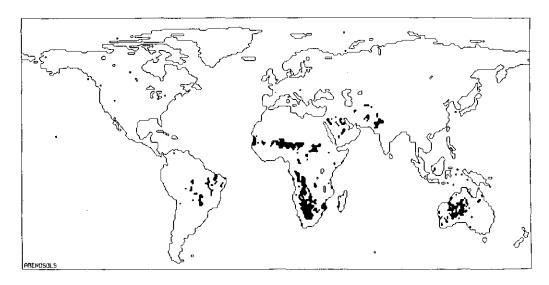


Fig. 1. Arenosols worldwide.

GENESIS OF ARENOSOLS

The history of the Arenosols of the dry zone is distinctly different from that of Arenosols in the wet tropics. The former show minimal profile development because soil forming processes are at a standstill during long periods of drought and/or because the parent material is of young age. The latter formed in young sandy deposits or, the other extreme, constitute the thick albic E-horizon of a Giant Podzol and represent the ultimate in soil formation.

Arenosols of the dry zone

In the dry zone, Arenosols are frequently associated with areas with areas of (shifting) sand dunes. Evidently, soil formation in such dune sand is minimal until the dune is colonized by a vegetation and held in place. Then, some humus can accumulate in the surface soil and a shallow, ochric A-horizon can develop. The individual sand grains of arid Arenosols have not seldom a coating of (brownish) clay and/or carbonates or gypsum while, in places, the desert sand is deep red by coatings of goethite (ferrugination, a relic feature according to some). Where the parent material is gravelly, sand is blown out of the surface layer and the coarser constituents remain behind at the soil surface as a 'desert pavement' of polished pebbles and stones.

Arenosols of the temperate zone

Arenosols in the temperate zone show signs of more advanced soil formation than those in arid regions. They occur predominantly in fluvioglacial, alluvial, lacustrine, marine or eolian quartzitic sands of very young to Tertiary age. Calcareous materials are often deeply decalcified and beginning podzolization (see under Podzols) with accumulation of Feand Al-humus complexes in thin lamellae is not uncommon. A true spodic B-horizon is not (yet) formed, not even in the poorest Arenosols. In loamy sands that are relatively rich, biological homogenization counteracts cheluviation (see under Podzols) and deep, homogeneous, brown or reddish profiles develop. Their ochric A-horizon contains humus of the 'moder' type that consists for the greater part of excrements. Many have an orange-red colour below the ochric A-horizon, attributable to thin (<10⁻⁵ m) iron coatings on the sand grains.

Arenosols in the humid tropics

Arenosols in the humid tropics either are young soils in coarsely textured alluvial, lacustrine or eolian deposits, or they are very old soils in residual acid rock weathering which lost all primary minerals other than (coarse grained) quartz in the course of an impressive pedogenetic history.

The <u>young Arenosols</u> of beach ridges and coastal plains, are <u>azonal</u> soils; they merely have a thin brown ochric A-horizon over a deep subsoil that may show gleyic properties below a depth of 50 cm and/or signs of beginning horizon differentiation that are taxonomically insignificant.

62 Arenosols

The <u>old (Albic) Arenosols</u> constitute the deep, bleached surface soils of Giant Podzols which extend downward to a depth of more than 125 cm. These <u>zonal</u> soils are the result of intense and prolonged dissociation of weatherable minerals and translocation of the weathering products through cheluviation. (The processes of podzolization, cheluviation and chilluviation are discussed in the chapter on Podzols.) Figure 2 shows an area with Albic Arenosols developed in sandstone weathering in Sarawak.



Fig. 2. Albic Arenosols in the deep eluvial horizon of a Giant Podzol in Sarawak. Photo by courtesy of ISRIC, Wageningen.

CHARACTERISTICS OF ARENOSOLS

The Arenosols of the arid zone have a beginning A-horizon with weak single grain or crumb structure over a massive C-horizon. Those of the temperate zone have better developed but still ochric A-horizons over a substratum which may have thin iron coatings throughout, or contain lamellae of illuviated humus, clay or iron compounds that are too thin, too few or contain too little humus to qualify as a diagnostic B-horizon. Young tropical Arenosols are morphologically not very different from those of the temperate zone. Old tropical Arenosols under forest on residual quartzitic rock weathering have a dark brown O-horizon over a shallow greyish brown A-horizon which tops a deep, grey to white coarse sandy E-horizon. A shallow mini-Podzol may form in the A-horizon; it remains intact because of the virtual absence of biological activity.

Mineralogical characteristics

With particle <u>size</u> being the classification criterion and not particle composition, there is considerable mineralogical diversity among Arenosols. Those in temperate and humid tropical climates contain a high proportion of weathering resistant minerals, mainly quartz but also zircon, tourmaline, rutile, staurolite, etc. The Arenosols of the dry zone can have any mineralogical composition; many are calcareous and/or gypsiferous. The parent material may contain some material finer than sand which, upon weathering, releases iron compounds that colour the soil matrix orange or red.

Hydrological characteristics

Arenosols are very permeable soils; their saturated hydraulic conductivity varies with the packing density of the sand and have any value between 300 and 30,000 cm/day. Arenosols store only between 5 and 12 percent 'available' water; in arid regions they need irrigation for good production but with infiltration rates ranging from 4 to 400 cm/hour, many of them are unfit for surface irrigation due to high percolation losses.

Physical characteristics

Due to their low coherence, Arenosols are sensitive to compaction; total pore fractions vary between some 0.35 and 0.55 cm 3 /cm 3 . A cemented or indurated layer may occur at some depth but tillage and/or rooting are normally not severely hindered by it.

Chemical characteristics

Arenosols in humid temperate or tropical regions are leached and often deeply decalcified soils with a low capacity to store bases. Their A-horizons are shallow and/or contain only a few percent organic matter or less. On poor quartzitic sands, the natural (forest) vegetation survives on cycling nutrients and roots almost exclusively in the O-horizon and in a shallow A-horizon. In the richer, loamy Arenosols of the temperate zone, rooting is deeper and nutrient cycling less vital to the vegetation. The Arenosols of the dry zone are mostly rich in bases. Because of the high permeability, the low water storage capacity and the low biological activity, there can be shallow decalcification despite an extremely low annual precipitation sum.

MANAGEMENT AND USE OF ARENOSOLS

Arenosols occur in widely different environments and vary accordingly in genesis, composition and properties. The only characteristic that all Arenosols have in common is their coarse texture, accountable for their generally high permeability and low water storage capacity.

64 Arenosols

Arenosols in arid lands, where the annual rainfall sum is lower than 300 mm, are predominantly used for extensive (nomadic) grazing. From 300 mm/year onwards, dry farming is possible. High yields of small grains, melons, pulses and fodder crops have been realized on irrigated Arenosols but high percolation losses often make surface irrigation impracticable. The low coherence, low nutrient storage capacity and high sensitivity to erosion are further limitations of Arenosols in the dry zone.

Arenosols in the (sub)humid temperate zone have similar limitations as those of the dry zone albeit that drought is a less serious constraint. In some instances, e.g. in capital-intensive horticulture, the low water storage of Arenosols is considered advantageous because the soils warm up early in the season. In (much more common) mixed farming systems with cereals, fodder crops and grassland, supplemental sprinkler irrigation is applied to prevent drought stress during dry spells. A large part of the Arenosols of the temperate zone are under forest, either production forest or 'natural' stands in carefully managed nature reserves.

Arenosols in the humid tropics are best left under their natural vegetation, particularly so the deeply weathered Albic Arenosols. As nutrient elements are all concentrated in the biomass and in the top 20 cm of the soil, clearing of the land will inevitably produce infertile badlands without ecological or economic value. Under forest, the land can still produce some timber (e.g. Agathis) and wood for the paper industry.

MAJOR LANDFORMS IN SMECTITE REGIONS

Soil materials whose properties are dominated by an abundance of expanding 2:1 lattice clays can occur in a many landscape elements. Extensive areas are found in:

- (1) (Former) sedimentary lowlands, and
- (2) Denudational plains

LANDFORMS IN (FORMER) SEDIMENTARY LOWLANDS

Sedimentary lowlands of 'heavy' smectitic clays cover vast tracts along the southern border of the Sahara, in the Sahelian zone. These lands were lakes and floodplains between 12,000 and 8,000 BP, when the climate was more humid than at present. The Saharian lakes, notably Lake Chad, the inland delta of the Niger and the alluvial plains of the Nile (in the present Sudan) expanded in that period. The level of Lake Chad was in times 40 m higher than today and the lake had the size of the present Caspian Sea. Most rivers in the Sahelian zone, even those which are intermittent in our times, flowed continuously in meandering channels and deposited finely textured sediments.

Much of what is known about the former expansion of the Saharian lakes was revealed by palynological studies of diatoms and pollen in the lake sediments. Apparently, a savannah vegetation colonized large parts of the Sahara. Rock drawings in the area suggest that ostriches, rhinoceroses, crocodiles and giraffes lived there at the time. The more humid climate is attributed to southward penetration of polar air masses when subtropical high-pressure cells were weaker than at present. Later in the Holocene, notably after 5000 BP, the climate became drier again; lake levels subsided and rivers became intermittent. Under this regime of alternating dry and wet spells, VERTISOLS could form in the alluvial deposits.

LANDFORMS IN DENUDATIONAL PLAINS

Denudational plains of smectitic clays occur in the same, semi-desertic climate zone, but are restricted to areas where the parent rock was rich enough in Ca, Mg, and Na for smectites to form. These are essentially basic volcanic rocks, such as the Deccan Traps basalts in India, and basic basement rocks, such as amphibolites and greenschists.

VERTISOL formation is especially plausible where shallow groundwater carried these ions in solution and neoformation of smectites could occur.

NOTES

VERTISOLS

Soils having, after the upper 20 cm have been mixed, 30 percent or more clay in all horizons to a depth of at least 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 intersecting slickensides or wedge-shaped or parallellepiped structural aggregates at some depth between 25 and 100 cm from the surface, with or without gilgai.

Key to Vertisol (VR) Soil Units

Vertisols having a gypsic* horizon within 125 cm of the surface.

Gypsic Vertisols (VRj)

Other Vertisols having a calcic * horizon or concentrations of soft powdery lime a within 125 cm of the surface.

Calcic Vertisols (VRc)

Other Vertisols having a base saturation (by $1M \times 14^{\circ}$ NH₄OAc at pH 7.0) of less than 50 percent at least between 20 and 50 cm from the surface.

Dystric Vertisols (VRd)

Other Vertisols.

Eutric Vertisols (VRe)

* Diagnostic horizon; see Annex 1 for full definition.

Diagnostic property; see Annex 2 for full definition,

* Phase: see Annex 3 for full definition.

SUMMARY DESCRIPTION OF VERTISOLS

Connotation: churning heavy clay soils; from L. vertere, to turn.

<u>Parent material</u>: sediments that are finely textured and contain a high proportion of smectite clay, or products of rock weathering that have these characteristics.

<u>Environment</u>: 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.

<u>Profile development</u>: A(B)C-profiles. Alternate swelling and shrinking of the expanding clay produces deep cracks during the dry season, and slickensides and wedge-shaped structural elements in the subsurface soil. In many areas a 'gilgai' microrelief occurs.

68 Vertisols

<u>Use</u>: Fine texture and poor internal drainage account for the often poor workability of Vertisols, both in wet and dry conditions. Vertisols are productive soils if properly managed.

REGIONAL DISTRIBUTION OF VERTISOLS

Vertisols cover a total of 311 million hectares or 2.4 percent of the global land area. An estimated 150 million hectares is potential crop land. In tropical areas, Vertisols cover some 200 million hectares or 4 percent of the 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 (Figure 1). Smaller areas of Vertisols exist in wetter, drier and cooler climates. The annual rainfall may be as low as 150 mm (e.g. Sudan) or as high as 3000 mm (e.g. Tripidad).

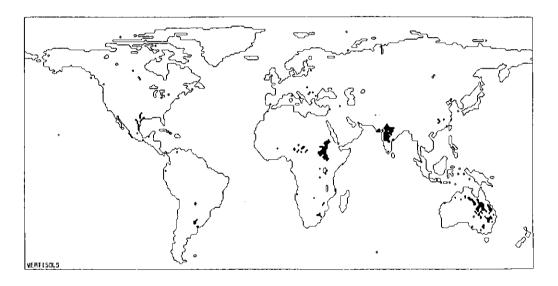


Fig. 1. Vertisols worldwide.

GENESIS OF VERTISOLS

The formation of a vertic structure

The genesis of Vertisols centers around the formation of characteristic structural aggregates ('vertic structure') that occur throughout most of the solum; the aggregates change only gradually with depth in grade of development and size of peds. The formation of a vertic structure will be explained assuming a level plain with smectitic clayey sediments and a semi-arid tropical climate with a distinct rainy season:

At the end of the rainy season, the clay plain is flooded but most of the excess water will eventually evaporate. When the saturated surface soil starts to dry out, there is initially one-dimensional shrinkage, and the soil surface subsides without cracking. Upon further drying, the soil loses its plasticity; tension in the soil material increases until its tensile strength is locally exceeded and the soil cracks. Cracks are formed in a pattern that becomes finer as desiccation proceeds. In most Vertisols, complete drying out of the surface soil turns it into a 'surface mulch' with a granular or crumb structure. Vertisols are 'self-mulching' if they develop a surface mulch. Granules or crumbs of the mulch horizon fall into the cracks partly filling them up (Figure 2).

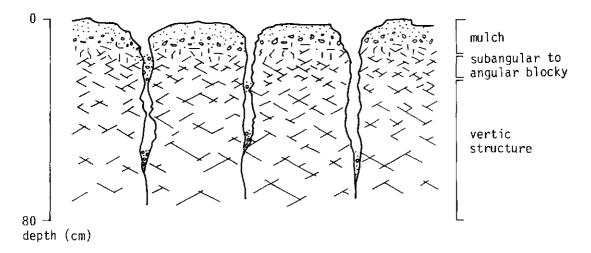


Fig. 2. Cracks, surface mulch and soil structure in a Vertisol during the dry season.

Upon rewetting, part of the space that the soil requires for its increased volume is occupied by mulch material. Continued water uptake causes pressures that result in shearing: the sliding of soil masses along each other. Shearing occurs as soon as the 'shear stress' that acts upon a given volume of soil exceeds its 'shear strength'. The swelling pressure, acting in all directions, is resolved by mass movement along oblique planes at an angle of 20 to 30 degrees with the horizontal plane (Figure 3). The shear planes are known as slickensides, polished surfaces that are grooved in the direction of shear. Intersecting shear planes define wedge-shaped angular blocky peds. The ped surfaces are slickensides or parts thereof, sometimes known as pressure faces. Although the structure conforms to the definition of angular blocky, the specific shape of the peds has prompted authors to coin special names such as 'lentils', 'wedge-shaped peds', 'tilted wedges', 'parallelepipeds' and 'bicuneate peds'; the type of structure is sometimes

called 'lenticular' or 'bicuneate' but shall be referred to as 'a vertic structure' in the following.

The size of the peds increases with depth. In uniform soil material this is attributable to:

- (1) the moisture gradient during drying and wetting. This gradient is steepest near the surface where small aggregates are formed in a loose packing ('mulch'). The moisture gradient decreases with depth except around cracks where wetting and drying are much more rapid than in the interior of crack-bounded soil prisms.
- (2) the increasing overburden, i.e. the increasing load of the overlying soil. At greater depths, higher swelling pressures are needed to exceed the soil's shear strength; only a large volume of swelling soil can generate such pressures and, consequently, structural aggregates will be larger.

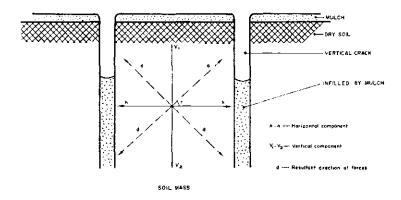


Fig. 3. Idealized stress diagram. Soil at three-dimensional expansion stage (source: De Vos & Virgo, 1969).

The typical vertic structure (see Figure 4) extends from some 15 or 20 cm below the surface mulch down to the transition of solum to substratum, just below the depth of cracking. There are no seasonal moisture changes in the substratum; if it has a vertic structure, then that structure is fossil. Vertisols with a very deep, fossil vertic structure are common in situations where sedimentation has alternated with periods of geogenetic stand-still.

The sliding of crumb surface soil into the cracks, and the resultant shearing have some important consequences:

(1) Subsurface soil is pushed upwards while surface soil falls into cracks. In this way surface soil and subsurface soil are mixed, a process known as churning or (mechanical) pedoturbation. Churning has long been considered an essential item in Vertisol formation; this is also suggested by the name (the Latin 'vertere' means 'to turn'). However, morphological studies and radio-carbon dating in recent

- years showed that many Vertisols do not exhibit strong homogenization. In such Vertisols shearing is not necessarily absent but it may be limited to up-and-down sliding of soil bodies along shear planes.
- (2) In churning Vertisols, coarse fragments like quartz gravel and hard rounded carbonatic nodules are concentrated at the surface, leaving the solum gravel-free. The coarse fragments are pushed upwards with the swelling soil, but most of the desiccation fissures that develop in the dry season are too narrow to let them fall back.
- (3) Aggregates of soft powdery lime indicate absence of churning, unless such aggregates are very small and form rapidly. Soft powdery lime is a substratum feature in Vertisols.

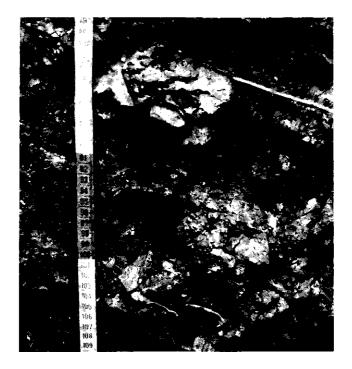


Fig. 4. The typical vertic structure of Vertisols.

Not all Vertisols develop a surface mulch. Some develop a hard surface crust. Cracks in such soils are sharp-edged, remain open throughout the dry season, and little surface soil falls into them. Because of differential wetting between adjoining parts of soil, swelling pressures can still develop, and such soils do have a vertic structure but the grade of structure is weaker than in self-mulching Vertisols.

Crusty Vertisols are but one example of the variation in structure formation among Vertisols. Factors that have a bearing on structure formation usually act through their effects on the tensile and shear strengths of the

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soil material: fine peds or, alternately, cracks at close intervals, are generally formed in soil materials that have low tensile and shear strengths under moist conditions, whereas large peds and cracks at wider intervals are formed in soil materials with high tensile and shear strengths. Vertisols that have a high ESP have greater tensile and shear strengths than soils with lower sodium saturation and commonly have a surface crust rather than a mulch. If the ESP is low and there is much finely divided lime, surface mulching is at a maximum and peds are fine. Vertic structure forming processes become stronger with increasing clay content and with a higher proportion of swelling clay minerals. Sandy Vertisols have limited swell and shrink; they develop narrow cracks and a surface crust.

The formation of a gilgai surface topography

A typical self-mulching Vertisol has an uneven surface topography: the edges of crack-bounded soil prisms crumble down, whereas the centres are pushed upward. The scale of this surface irregularity is that of the cracking pattern, usually a few decimeters. Microrelief on a larger scale, superimposed on this unevenness, is known as 'gilgai' and consists on level terrain of small mounds surrounded by a continuous network of small depressions, or depressions surrounded by a continuous network of narrow ridges.

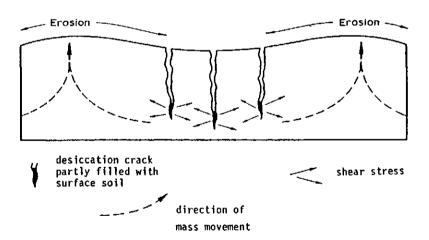


Fig. 5. Sketch showing the kinematics of mass movement in Vertisols that result in gilgai microrelief (after Beinroth, 1965).

Several hypotheses have been put forward to explain the gilgai microrelief. These have in common that they relate gilgai to mass movement in swell/shrink soils. The soil must have sufficient cohesion in order to transfer pressures all the way to the soil surface. Gilgai is sometimes explained by sloughing of surface mulch into cracks and an upward thrust of soil between cracks as a result of subsurface soil swelling. However, gilgai is a feature that is superimposed over the cracking pattern; it originates in the subsurface soil and substratum.

There are two observations to support a subsurface origin of the gilgai microrelief:

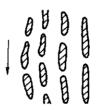
- (1) A trench profile through a complete 'wave' of mound and depression shows the slickensides in the lower solum and upper substratum to be continuous from below the centre of the depression toward a higher position below the centre of the mound. The oblique shear planes show a preferential direction. Substratum material is pushed upwards alongside such sets of parallel slickensides. See Figure 5.
- (2) A gilgaied land surface that is levelled will have gilgai reappearing in a few years.

The commonest form of gilgai is the normal or round gilgai. On slightly sloping terrain (0.5 to 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.

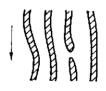
In most gilgais, the wave length (from centre of mound to centre of depression) is between 2 and 8 m; the vertical interval or 'amplitudo' is often between 15 and 50 cm. Figure 6 presents some common forms of gilgai.



normal or round gilgai



lattice gilgal



linear or wavy gilgai

//// mound or micro-ridge
direction of slape

Fig. 6. The commonest forms of gilgai.

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Most gilgaied areas have Vertisols, but not all Vertisols develop a gilgai microrelief. Some areas with gilgaied Vertisols have been mapped as a 'gilgai phase' on the 1:5,000,000 Soil Map of the World. In the Sudan, where Vertisols occur in a more or less continuous clay plain over a distance of some 700 km from north to south, and where the annual rainfall increases in that direction from 150 to 1000 mm, the gilgai microrelief is restricted to the 500-1000 mm rainfall zone. The gilgaied Vertisols have a thinner and less clearly expressed surface mulch and are less calcareous than the non-gilgaied Vertisols in the north.

The profile morphology of gilgaied Vertisols is different between mound and depression. On the mounds, the A-horizon is thin whereas the depression profile has a deeper (thickened) and usually darker A-horizon. Coarse components of substratum material which reach the soil surface at the mound site such as quartz gravel and carbonatic concretions, remain at the surface whereas finer soil material is washed down to the depressions. In Australia, 'high gilgais' occur with wavelengths up to 120 m and amplitudes of up to 240 cm. These high gilgais may well have formed in an entirely different way.

The formation of the smectite clay parent material

The environmental conditions that are instrumental in the formation of a vertic soil structure also promote the formation of suitable parent materials. Rainfall must be sufficient to enable weathering but not so high that leaching of bases occurs. Dry periods are required for the crystallization of clay minerals that form upon rock or sediment weathering. Impeded drainage hinders leaching and curbs the loss of weathering products. High temperatures, finally, promote weathering processes. Under such conditions smectites can be formed in the presence of silica and basic cations - especially Ca and Mg - and at a pH above neutral.

The largest Vertisol areas are found on sediments that have a high content of smectite clays or produce such clays upon post-depositional weathering (e.g. Sudan, described by Buursink, 1971), and on extensive basalt plateaus (e.g. India).

The formation of Vertisol parent materials and Vertisol profiles can be illustrated by examining 'red-black' soil catenas, as abundant in Africa, the Indian subcontinent and Australia. Usually there are red soils (Luvisols) on crest and upper slope positions, shallow or moderately deep red soils (Leptosols and Cambisols) on steeper sections of the slope, and black soils (Vertisols) on lower-lying sites. Depending on parent rock and environmental conditions, Vertisols cover only valley bottom positions or, in addition to the valley bottoms, also contiguous lower footslopes, or even, as residual soils, (gently) sloping hillsides.

In the semi-arid to subhumid tropics, smectite is the first secondary mineral that forms upon rock weathering (Kantor & Schwertmann, 1974). It retains most of the ions that have been liberated from the primary silicates, like Ca and Mg. Iron, contained as ${\rm Fe}^{2^+}$ in primary minerals, is preserved in the smectite crystal lattice as ${\rm Fe}^{3^+}$. As weathering proceeds, smectites become unstable. The clay minerals decompose, and the released basic cations and part of the silica are removed by leaching. ${\rm Fe}^{3^+}$ -compounds

remain in the soil, lending it a reddish colour; aluminium is retained in kaolinite and Al-oxides. The leached soil components accumulate at the lower terrain positions where they precipitate and form smectites. Smectites in these poorly drained positions are stable as long as the pH is above neutral.

There are some additional reasons why there is a relative dominance of smectite in the lower members of the catena;

- lateral (surficial and sub-surficial) physical transport of clay, particularly of fine clay in which the proportion of smectites is greater than in the coarse clay, and
- decreasing drainage and leaching of soluble compounds from high to low terrain positions. Internal drainage is impeded by the formation of smectites. (It is increased when kaolinite forms: ferric iron, released from the smectite lattice, cements soil particles to stable structural peds and maintains a permanent system of pores in the soil.)

The processes of rock weathering, breakdown of primary and formation of secondary minerals, and transport of soil components result in the catenary differentiation mentioned above: reddish well-drained soils on higher positions, and black, poorly drained soils in depressions (see Table 1).

Minor colour differences between Vertisols are often an indication of differences in drainage status. The more reddish hue or stronger chroma of relatively better-drained Vertisols is due to a higher content of free iron-oxides. Poorly drained Vertisols are low in kaolinite; the hue is less reddish and the chroma weaker. Formerly, colour chroma was used as a differentiating characteristic between Vertisols (Soil Taxonomy still does so) but this practice was left in the FAO/Unesco classification system when it became clear that pale-coloured, apparently poorly drained Vertisols which were thought to be typical of level and/or depressed areas, occurred also on gentle and even moderate slopes. The simple landscape-soil relation could then no longer be held upright.

TABLE 1
Analytical data of the highest (Luvisol) and the lowest member (Vertisol) in a 'red-black' soil catena in the Sudan.

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PROFILE	ABC	DEPTH (cm)	CLAY (%)	pН	CEC CE (cmol(+		BS (%)	SiO_2/Al_2O_3 (clay fr.)	orgC (%)	lime (%)
LUVISOL	A	0-10 10-30	4 15	6.6	nd 9.0	nd	nd	3.8	0.5	0
	AB Btl	30-60	23	6.1 4.7	15.5	60 68	71 49	3,3 3,5	0.6	0
	Bt2	60-85	33	4.5	21.4	66	45	3.3	0.3	Ö
	Bt3	85-105	43	4.5	22.6	50	51	3.1	0.2	0
	BC	105-135	39	4.4	27.4	69	47	3.0	0.1	0
	С	135-160	39	4.7	27.4	71	57	3.2	tr	0
VERTISOL	A	0-30	78	6.6	66.6	86	100	4.3	0.9	0.7
	Bw1	30-90	78	7.2	78.4	100	100	4.6	0.9	1.2
	Bw2	90-150	81	7.3	78.2	96	100	4.6	0.7	1.2
	BCwk	150-180	79	7.3	80.2	103	100	4.5	0.4	1.5

CHARACTERISTICS OF VERTISOLS

Vertisols are commonly seen as AC-profiles, the A-horizon consisting of both the surface mulch (or crust) and the underlying structure profile that changes only gradually with depth. Surely, the subsurface soil with its distinct vertic structure conforms to the definition of a cambic B-horizon, but where does the A-horizon end and the B-horizon begin? Soil colour, texture, element composition, cation exchange capacity, etc are all uniform throughout the solum. There is hardly any movement of soluble or colloidal soil components. (If such transport occurs, it is counteracted by the pedoturbation process.)

There is, nevertheless, a growing tendency to describe Vertisols as ABC-profiles, and to define as a (cambic) B-horizon that profile section that has a vertic structure. The A-horizon is then considered limited to the surface mulch (or crust) and an underlying subangular blocky zone of a few decimeters thickness. A calcic horizon or a concentration of soft powdery lime is often present in or below the horizon(s) with a vertic structure, whereas gypsum may have accumulated in places, either uniformly distributed over the matrix or in nests of gypsum crystals.

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 occurs easily because infiltration of water in a moist Vertisol is extremely slow. Vertisols with 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 waterholding capacity of Vertisols differ widely which may be due to the complex pore space dynamics. Water is retained both at the 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 (Farbrother, 1972) 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 invariate (at about 20 percent, corresponding to a matric suction of some 1500 kPa) throughout the year.

Physical characteristics

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 very 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.

Chemical 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 6.0 to 8.0. Higher pH's (8.0-9.5) occur in Vertisols with a high ESP. The cation exchange capacity is very high, in the range of 30 to 80 cmol(+)/kg of dry soil. The CEC of the clay is of the order of 50 to 100 cmol(+)/kg clay. 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.

Saline, sodic and saline/sodic phases occur in the more arid parts of the Vertisol coverage. Sodicity occurs incidentally in higher-rainfall areas, e.g. in depressions without outlet.

The effect of sodicity on the physical properties of Vertisols is still a subject of debate. As stated earlier, Na-clays have greater tensile and shear strengths than Ca-clays, and a high ESP produces a soil structure of a relatively 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. Clay dispersion accompanied by clay movement, the normal consequence of high sodium saturation in clay soils, cannot take place in soils with such a low hydraulic conductivity 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; an ESP value of 16 is considered optimal for irrigated cotton cultivation on saline/sodic Vertisols in the Sudan Gezira (possibly because a relatively 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/Manaquil irrigation scheme in the Sudan, is indispensible for maintaining a low salinity level in the surface soil.

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

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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 a 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 land slides 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 must therefore be discouraged.

Examples from the Sudan and India may illustrate the potential and the problems encountered in farming pursuits on Vertisols:

The 740,000 hectares Gezira/Manaqil 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, groundnuts and sorghum.

In addition to irrigated farming, extensive forms of rainfed cropping are practiced in large Vertisol plains; sorghum is the 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; increased infiltration of water, reduced erosion hazard, and optimum use of the available water. Some of the measures applied at the Soil and Water Conservation Research Centre at Bellary, Karnataka State, are a good illustration (Rama Mohan Rao & Seshachalam, 1976):

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 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 management operations. The system is described by Kanwar et al. (1982).

NOTES

MINERAL SOILS CONDITIONED BY THE TOPOGRAPHY/PHYSIOGRAPHY:

FLUVISOLS GLEYSOLS LEPTOSOLS REGOSOLS



MAJOR LANDFORMS IN ALLUVIAL LOWLANDS

In the present context, the term 'lowlands' connotes flat and level wetlands, normally situated at or near sea level and consisting of Pleistocene and/or Holocene sediments that are too young to be strongly weathered. Sedimentary lowlands are prominent in fluvial, lacustrine, marine and glacial landscapes, and particularly in areas of crustal subsidence. Climate played an important role in the shaping of most lowlands, next to tectonic processes. Climate determines the sediment load and bedform patterns in fluvial environments; the climate determines whether a sediment-trapping mangrove vegetation develops along a coast, and the climate influences the weathering and ripening of the deposited sediments.

Climate changes in the Quaternary have greatly influenced today's low-lands. Most coastal lowlands were shaped to their present form and extent after the melting of Pleistocene ice caps stopped some 6,000 years ago, and some 15 percent of the present land surface was influenced by Pleistocene glaciers that have now disappeared. It is impossible to understand the complexity of the landforms in lowlands without referring to both their present climate and their climatic history.

The major landforms in alluvial lowlands will be discussed for two broad categories of lowlands:

- (1) (inland) fluvial lowlands, and
- (2) (coastal) marine lowlands.

LANDFORMS IN (INLAND) FLUVIAL LOWLANDS

River systems are complex systems and a watershed area will normally harbour a variety of landforms. These correspond with the nature of the river and with the position in the river system. In most systems, three zones can be distinguished:

- the upper parts, where erosion outweighs sediment deposition. These stretches are normally mountainous or hilly but become gradually levelled by erosion. These parts will not be discussed here, as they are usually not lowlands.
- (2) the middle stretch, in which erosion and deposition roughly compensate each other. This zone is largely a zone of transport; rivers flow in their own alluvium and have meandering or braided channels, in places with terraces or alluvial fans.
- (3) the lower parts with net sediment accumulation. This zone usually grades into a delta, an estuary or some other coastal form, or (in arid regions) it fades away in a dry drainless basin.

In the following, attention will be given to some common landforms in the systems of three types of rivers:

- (1) braided rivers,
- (2) meandering rivers, and
- (3) anastomosing rivers.

Braided rivers

Braided rivers have numerous shallow channels of low sinuosity. At maximum discharge, the sediment islands between the channels are usually submerged but they reappear when the water level falls, not seldom at a completely different place. Gravel or coarse sand is deposited at the downstream ends of the islands whereas the sides are eroded as the channels shift. This river type with its continually changing pattern of shallow channels and gravel bars, occurs in watershed areas with a highly <u>irregular discharge</u> and an <u>abundance of coarse weathering products</u>. Such conditions prevail in areas with a sparse vegetation cover and torrential rain showers such as arid and periglacial regions and along glacier fronts, where sudden peaks occur in the discharge of water and sediment due to melting of the ice in spring.

The actual river bed is normally unsuitable for arable farming because of the risk of flooding but agriculture may be practised on river terraces which extend above the flood plain. Terraces are not real wetlands; they are usually remnants of ice-age floodplains with a higher base level which became dissected later, in (more humid) interglacial times.

Braided river deposits contain alternating areas (lenses) of coarse gravel and sand and only minor inclusions of finer sediment. The coarse deposits may have become covered with fine-grained sediment later, e.g. when the braided stream turned into a meandering river or during exceptional floods ('Hochflutlehm'). Not seldom, one can detect former gravel bars and gullies in the microrelief (less than a metre high) of areas underlain by braided river deposits. Figure 1 shows the basic elements of a typical braided river system.

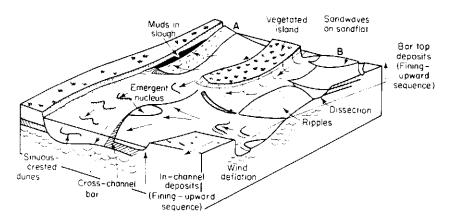


Fig. 1. Basic elements of a typical braided river. Bar A is being driven laterally toward the far bank and is forming a protected part of the channel (slough) in which mud is deposited. B is a complex sand flat, originally formed from growth of an emergent nucleus such as that shown in the left central part of the diagram. Source: Blatt, Middleton & Murray, 1980; based on the work of D.J. Cant in the South Saskatchewan River.

Alluvial fans are formations which have the coarse sediments and shifting channels of braided rivers. They develop where the gradient of a stream decreases sharply, e.g. where a tributary stream leaves the mountains and enters the floodplain. There, the sediment load of the river can no longer be carried and most of it is dropped right at the entrance to the plain. This rapidly blocks the channel which then sweeps left and right to obviate the obstacle. The result is a fan-shaped, low-angle sediment cone. There is often a grain size gradient, from coarse to fine, from the 'proximal' part of the fan (close to the apex) to the 'distal' part (far into the plain). The fans are excellent aquifers, and ideal sites for towns and villages.

Meandering rivers

Meandering rivers consist normally of one single channel of high sinuosity. The channel is bordered by 'natural levees' behind which the 'basins' or 'backswamps' occur. Laboratory experiments have confirmed that meandering channels are typical of streams with a regular, rather steady discharge pattern and a bedload of relatively fine-grained materials (sand, silt, clay). See Figure 2.

Meandering rivers occur in vegetated areas in a humid climate. The dense vegetation cover is associated with a preponderance of chemical weathering, with formation of clay in soil and saprolite, and with stable slopes with little surface runoff and much underground feeding of rivers (throughflow).

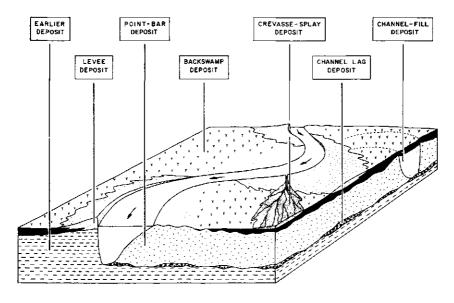


Fig. 2. The classical point bar model for a meandering stream. Source: Allen, 1964.

Meandering rivers may show considerable fluctuations in discharge, due to seasonality of rainfall and snow melting, but they never run dry and sudden floods are rare. If the channel is filled to the top of the natural levees, the river is in 'bank-full stage'. Occasionally, more water is supplied than can be held between the natural levees. Then, the river overflows its levees and the basins are flooded.

Sedimentation takes place in the channel, on the levees and in the basins. The coarsest sediments are found at the bottom of the channel ('lag deposits') and consist normally of coarse sand or gravel. Finer sand settles along the inner bends of the rivers, on the 'point bars'. During floodings, fine sand or silt is deposited on top of the levees, and clay in the basins. (Peat may accumulate there as well.)

The whole system of lag gravels, pointbar sands, levee silts and back-swamp clays migrates laterally as the meander loops erode the outer banks and deposit sediments on the inner point bars. In this way a 'fining upwards' sedimentary sequence may develop. 'Crevasse splays' are deposited in the basins where the levees break; they are 'coarsening upwards' sediment sequences. Channel cut-offs, ('oxbow lakes') become eventually filled in with clay and/or peat.

Although meandering rivers are most abundant in humid regions, they occur also in arid environments but their sources are invariably outside the arid region. The Euphrates and Tigris, for instance, originate in the snow-clad Taurus range in Turkey before they reach the Mesopotamian desert.

Anastomosing rivers

Anastomosing rivers resemble the braided rivers in having several branching channels, but have their fine-grained sediments and their well-developed levees and backswamps (often covered with peat) in common with the meandering rivers. They differ from the braided and the meandering rivers in that the position of the channels changes little unless a sudden breach of the levee ('avulsion') occurs. Anastomosing rivers have very low gradients. There is little erosion of the outer banks and little lateral displacement of the meander belt. Anastomosing rivers occur mainly in rapidly subsiding interior basins; examples are the Lower Magdalena Basin in Colombia, or areas close to the coast such as the Alblasserwaard in the Netherlands.

LANDFORMS IN (COASTAL) MARINE LOWLANDS

The coastal lowlands contain a great variety of landforms. For all practical purposes, the lowlands can be divided in three broad groupings:

- deltas.
- (2) wave-influenced coastal lowlands, and
- (3) tide-dominated coastal lowlands.

<u>Deltas</u>

Deltas are formed where silt-loaded rivers debouch into the sea or a lake. As the oceans reached their present level only some 6000 years ago, most surface features of deltas are quite recent. However, drillings in large delta bodies have revealed long and complex histories. Such large delta bodies can be more than 10 kilometres thick and their sediments contain evidence of numerous sea level fluctuations and changes of the drainage base. The different sedimentary facies of delta bodies reflect changes in one or more of the following external factors:

- the density of the river water relative to the density of the sea or lake water.
- (2) the nature and volume of the sediment carried by the river,
- (3) the influence of waves and tides,
- (4) the climate, and
- (5) the rate of (tectonic) subsidence of the basin.

The oldest description of deltas dates back to 1885 and is given in a publication on the outflow of glacially fed rivers into the former (Pleistocene) Lake Bonneville. In this case, in which no differences in the apparent densities of river water and lake water existed and the influence of waves and tides was negligible, a fan-shaped classical delta or 'Gilbert-type delta' is formed. In this type of delta, the coarsest (sand-sized) material is deposited in the delta plain close to the mouth; finer (silty) sediments accumulate in the submerged delta slope and the finest clay particles travel farthest (to the 'prodelta'). Thus, a grain size gradation evolves across the delta. When the delta progrades under continuing sediment supply, progressively coarser sediments cover (or 'onlap' as sedimentologists say) finer sediments: a coarsening upwards sedimentary sequence results. Such sequences may be tens or even hundreds of metres thick. The fine prodelta sediments are the 'bottom-set beds', the sloping delta-front sediments the 'fore-set beds', and the topmost delta plain sediments the 'top-set beds'.

If (low density) fresh water flows out in the sea, it acts as a buoyant jet stream driving its suspended sediment far into the sea. With the river's natural levees extending into the sea, the delta system expands rapidly, particularly if little wave or tidal action disturbs sedimentation. In this way, a 'birdfoot delta' is formed, in which the tributary channels of the delta plain form the 'toes' of the foot. Swamp lands form between the toes; they are the brackish equivalents of the backswamps in purely fluvial environments. Occasionally, a channel breaks through its levees and deposits a crevasse splay; the channel may then be abandoned and a new one formed on top of the crevasse sediments. So-formed fining upwards sequences can be several metres thick.

Where tidal influence is strong, the channels widen to funnels similar to the ones found in estuaries (see the paragraph on tide-dominated lowlands). The areas in between the channels develop into 'tidal flats'. In tropical areas, these are normally colonized by mangroves which add much organic debris to the fresh sediment.

This combination of a shallow, brackish, tidal swamp with reduced sediments and much organic matter promotes the microbial reduction of sulfates

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(from the sea water) to 'pyrite' (iron sulfide). Pyritic sediments are the parent material of 'Acid Sulfate soils'.

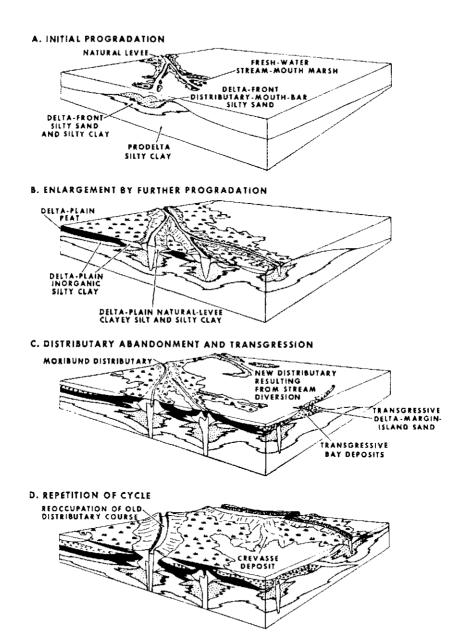


Fig. 3. Development of delta sequences in the Mississippi delta. Source: Frazier, 1967.

Wave-influenced delta plains have commonly cuspate or straight outlines formed by sand ridges, in places topped by eolian dunes. Lagoons, commonly filled in with peat, lie behind the (parallel) ridges. Any sediment which enters the sea is quickly dispersed by wave action.

The Mississippi delta is an example of a birdfoot delta without strong tidal influence (see Figure 3); the deltas of the Mekong and Brahmaputra rivers are tide-dominated delta plains. Wave-dominated deltas are, for instance, the deltas of the Rhone, Ebro and Nile in the Mediterranean Sea, and the Danube delta in the Black Sea.

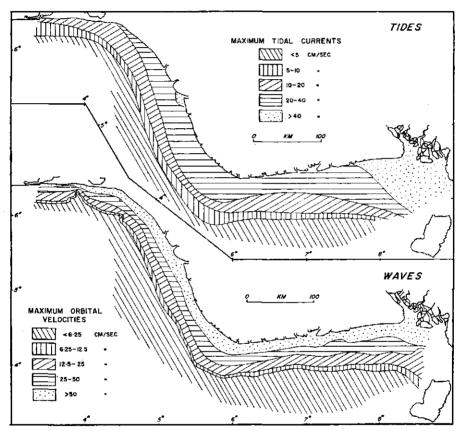


Fig. 4. Wave and tidal currents acting offshore in modern Niger delta. Source: Allen, 1965.

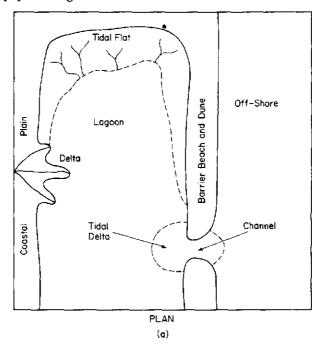
Wave-influenced coastal lands

The importance of the tides is usually measured by the tidal amplitude: coasts with a tidal amplitude of 2 metres or less are called 'microtidal', those with 2 to 4 metres amplitude are 'mesotidal' and coasts with a tidal amplitude of more than 4 metres are 'macrotidal' coasts.

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Microtidal coasts are shaped by wave action rather than tidal action. Waves rolling up to the coast produce a surf (wash and backwash) resulting in net transport of sediment towards the coast. Compensating currents occur parallel to the shore: 'longshore currents'. If the waves approach the coast under an angle, longshore currents and 'beach drift' displace the sediment parallel to the coast. See Figure 4.

Most wave-dominated coasts have sandy beaches which are attached to the hinterland; if they are separated from the hinterland by a narrow strip of sea, they are called 'barriers'. See Figure 5. Along meso- and macrotidal coasts, inlets are present between individual barriers and tidal flats are formed behind the barriers. These are inundated at high tides and are bare or carry a halophytic vegetation.



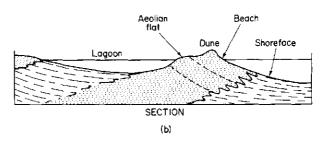


Fig. 5. Main morphological components of the barrier model. Source: Blatt, Middleton & Murray, 1972.

Some wave-dominated coasts consist mainly of mud. Small sand ridges may occur within these clay plains; they are called 'cheniers', and the plains themselves are 'chenier plains'. Examples of coasts with chenier plains are the coasts of the Guianas and the coast of Louisiana. The Surinam chenier plain has grown seaward over tens of kilometres during the last 6000 years.

Delta growth is directed away from the hinterland; in the case of sandy beaches and barrier coasts the direction of landform development is not so clear. Whether accretion will take place or transgression depends on such factors as the along-shore sediment supply, the slope of the foreshore and the subsidence rate of the basin, all factors which influence the sedimentary sequences found in wave-built sandy coastal lowlands. Seaward progradation produces coarsening-upwards sequences where the sea bottom is formed by finer material than the sediments added, and where the sands near the shoreline are finer-grained than those higher up. On the other hand, landward shifting of barriers may cause the deposition of sand on top of older lagoon material, so that coarsening upwards sequences may develop. The interpretation of observed sedimentary patterns is complicated by the fact that regression and progradation have in places alternated during the Hologene.

An overview of the history of the Dutch central coast, between Hoek van Holland and Den Helder, illustrates how the landforming processes discussed in the foregoing have shaped this wave-dominated coastal lowland (the present tidal amplitude is only 1.5 metres).

When the post-glacial sea level rise was still rapid, between >7,000 and 5,000 years BP, the then existing coastal barriers moved inland. They were separated by inlets behind which clayey tidal flats were formed, the 'Calais deposits'. On the landward side, the Calais deposits interfinger with fluvial clays and peat. When the sea level rise slowed down (5000-2000 BP), the barrier system closed and grew seawards, forming the 'Old Barriers' ('Oude Strandwallen') that are particularly well developed in the stretch between The Hague and Haarlem. The slow rise of the drainage base during this period created ideal conditions for the accumulation of thick layers of peat (the 'Holland peat') on top of the Calais deposits. Much of the peat was later mined for fuel, until the Calais deposits were reached. The resulting lakes have largely been drained and form some of the lowest polders of the Netherlands.

Tide-dominated coastal lowlands

Especially along macrotidal coasts, the river mouths are strongly widened by incoming tides. 'Estuaries' are formed which have funnel shaped channels with extensive sand flats in between. Sand waves may develop there with wave lengths of several metres ('megaripples'); the interior parts are commonly brackish or saline.

Where the tides are high enough to flood parts of deltaic or estuarine systems, they build tidal flats. The tides enter and leave through deep gullies (geulen) between barrier islands. The gullies are permanently submerged and conduct huge masses of water and (coarse) sediment at incoming and outgoing tides. The sediment reaches the flats through small creeks

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(prielen). These have a meandering pattern and widen towards their mouths. 'Intertidal flats' (platen), between mean high water level and mean low water level, are flooded twice a day. Fine sandy and clayey sediment settles on top of these flats. The parts which are flooded only at extreme tides are called 'supratidal flats' (salt marshes, kwelders, schorren).

In temperate regions, intratidal flats are normally barren whereas supratidal flats carry a halophytic vegetation of herbs and shrubs. This vegetation traps sediment when flooded; the resulting deposits are stratified with a typical alternation of fine and coarse layers (kweldergelaagdheid). Tidal flats in the humid tropics carry a mangrove vegetation already in the intertidal zone and silt up much more rapidly than most supratidal flats in temperate regions. Shifting of creeks and gullies leads to fining upwards sequences, with fine supratidal and intertidal sediments on top of coarser creek and gully sediments.

The southwestern part of the Dutch coastal plain is an example of a tide-dominated coastal lowland. Its 'Duinkerken deposits' were formed by continuously shifting estuary channels, mainly in connection with extreme floods. The oldest parts of the province of Zeeland (Oudland) are estuarine tidal flats dating back to pre-Roman times. In places, sandy creek fills in (former) tidal flats stand out as a result of relief inversion by differential compaction of clays and peat around the creeks. The youngest parts in this area (Nieuwland) are estuarine channel fills that were deposited only in the 16th century.

The soils of alluvial lowlands are marked by prolonged wetness (unless empoldered) and young age. Where sedimentation is still going on, stratified FLUVISOLS occur. GLEYSOLS are found in depressions which do not receive regular additions of sediment; their profiles testify of a shallow water table during all or most of the year. Both Fluvisols and Gleysols include Thionic Soil Units which formed in pyritic sediments.

FLUVISOLS

(with special reference to Thionic Soils)

Soils showing fluvic^a properties and having no diagnostic horizons other than an ochric^{*}, a mollic^{*} or an umbric^{*} A-horizon, or a histic^{*} H-horizon, or a sulfuric^{*} horizon, or sulfidic materials^a within 125 cm of the surface.

Key to Fluvisol (FL) Soil Units

Fluvisols having a sulfuric* horizon or sulfidic^a material, or both, at less than 125 cm from the surface.

Thionic Fluvisols (FLt)

Other Fluvisols having a mollic* A-horizon or a eutrica histic* H-horizon.

Mollic Fluvisols (FLm)

Other Fluvisols which are calcareous 0 at least between 20 and 50 cm from the surface.

Calcaric Fluvisols (FLc)

Other Fluvisols having an umbric *A-horizon or a dystric histic H-horizon.

Umbric Fluvisols (FLu)

Other Fluvisols having a base saturation (by 1M NH_4OAc at pH 7.0) of less than 50 percent, at least between 20 and 50 cm from the surface.

Dystric Fluvisols (FLd)

Other Fluvisols showing salic properties.

Salic Fluvisols (FLs)

Other Fluvisols.

Eutric Fluvisols (FLe)

SUMMARY DESCRIPTION OF FLUVISOLS

Connotation: soils developed in alluvial deposits; from L. fluvius, river.

<u>Parent material</u>: mostly recent, medium- and fine-textured fluviatile, lacustrine or marine deposits.

<u>Environment</u>: periodically flooded areas (unless empoldered) in alluvial plains, valleys and (tidal) marshes, in climates ranging from arctic to equatorial, and from semi-arid to perhumid.

Diagnostic horizon; see Annex 1 for full definition.

a Diagnostic property; see Annex 2 for full definition.

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<u>Profile development</u>: AC-profiles; a distinct Ah-horizon may be present. The lower horizons show evidence of stratification and have no or only a weak soil structure. Gleying is common in the lower part of the profile.

<u>Use</u>: Fluvisols are used for a wide range of crops, or for grazing. Flood control, drainage and/or irrigation are normally required. Thionic Fluvisols suffer from severe soil acidity and high levels of noxious Al-ions.

REGIONAL DISTRIBUTION OF FLUVISOLS

Fluvisols occur on all continents and in all kinds of climates. Out of the 316 million hectares of Fluvisols worldwide, some 200 million hectares are in the tropics. Major areas can be found:

- a) along rivers and lakes, e.g. the Amazon basin, the Ganges plain of India, the plains near Lake Chad in Central Africa, or the marsh lands of Bolivia and northern Argentina;
- b) in deltaic areas, e.g. the deltas of the Ganges/Brahmaputra, Indus, Mekong, Nile, Zambesi, Orinoco, Rio Plata;
- c) in areas with recent marine deposits, e.g. the coastal lowlands of Sumatra, Kalimantan, and Irian in Indonesia, and the coastal zones of the above mentioned deltas.

Thionic Fluvisols ('Acid Sulfate Soils') cover an estimated 13 million hectares; the largest concentration (6.7 million hectares) is in the coastal lowlands of southeast Asia (Indonesia, Vietnam, Thailand) but they occur also in West Africa (Senegal, the Gambia, Guinea Bissao, Sierra Leone, Liberia) and along the north-eastern coast of South America (Venezuela, the Guyanas). Figure 1 shows the worldwide distribution of Fluvisols.

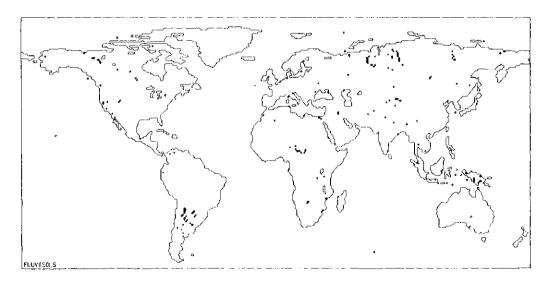


Fig. 1. Fluvisols worldwide.

GENESIS OF FLUVISOLS

Fluvisols are are young alluvial soils which have 'fluvic soil properties'. For all practical purposes this means that they receive fresh sediment during regular floods (unless empoldered) and have still stratification and/or an irregular organic matter profile.

In the upstream part of river systems, soils with fluvic soil properties are normally confined to narrow strips alongside the actual river bed. In the middle and lower stretches, the flood plain is normally wider and has the classical arrangement of levees and backswamps, with coarsely textured Fluvisols on the levees and more finely textured soils in the backswamps further away from the river.

In areas with marine sediments, relatively coarse-textured Fluvisols are present on barriers, cheniers, sand flats and crevasse splays; finely textured Fluvisols are present on clayey tidal flats and in chenier plains in between the ridges. Where rivers carry only fine grained material to the sea, the coastal plains (and their Fluvisols) are entirely clayey.

Genesis of Acid Sulfate Soils

The only difference between the parent material of Thionic Fluvisols and that of other Fluvisols is the presence of pyrite (FeS₂) in Thionic Fluvisols.

<u>Formation of pyrite</u> can take place in marine environments if the following conditions are met during sedimentation:

- Iron must be present. Easily reducible iron oxides or hydroxides are normally present in coastal sediments.
- (2) <u>Sulfur</u> must be present. This originates from sulfates in the sea water.
- (3) <u>Anaerobic conditions</u> must prevail to allow reduction of sulfate and iron oxides. This condition is met in fresh coastal sediments.
- (4) <u>Iron- and sulfate-reducing microbes</u> must be present; they occur in all coastal sediments.
- (5) Organic matter is needed as a source of energy for the microbes; it is supplied by a lush pallustric vegetation (mangrove forests, reeds, sedges).
- (6) <u>Tidal flushing</u> must remove the alkalinity which forms in the process of pyrite formation.
- (7) A slow sedimentation rate is necessary. Otherwise, the time will be too short to form sufficient pyrite for a potentially acid sediment.

The mechanism of pyrite accumulation boils down to the following: under oxygen-poor conditions (under water), microbes reduce ferric (Fe $^{3+}$) iron to ferrous (Fe $^{2+}$) ions, and sulfate (SO $_4^{2-}$) to sulfide (S $^{2-}$). During this process, organic matter is decomposed, ultimately to bicarbonates. Thus, a potentially acid part (the pyrite) and an alkaline part (bicarbonates) are formed in an initially neutral system. Tidal flushing removes the alkalinity (HCO $_3^{-1}$), and potentially acid pyrite remains behind.

Pyrite is formed in a number of steps, but the 'overall' reaction equation is:

$$Fe_2O_3 + 4 SO_4^{2} + 8 CH_2O + 1/2 O_2 = 2 FeS_2 + 8 HCO_3^2 + 4 H_2O$$

Important intermediate products in this process are soluble ferrous sulfate (FeSO₄) and (meta-stable) jarosite (KFe(SO₄)₂(OH)₆). Jarosite is formed at a pH below 4 if sufficient K*-ions are present. Jarosite has a typical straw-yellow colour which is easily recognized in the field and is indicative of Actual Acid Sulfate Soils (as distinguished from Potential Acid Sulfate Soils; the difference will be discussed later).

The composition of the various iron minerals which form in the process of pyrite oxidation is largely determined by the redox potential of the soil material (a measure of the quantity of oxygen present) and the pH. The stability of the iron components, and their dependence on redox potential and pH, is shown in Figure 2.

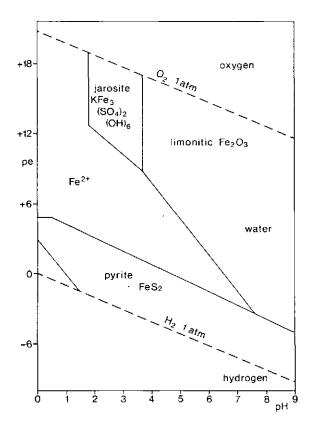


Fig. 2: pe:pH diagram of pyrite, limonitic Fe_2O_3 , jarosite and dissolved K^* , $SO_4^{2^*}$, Fe^{2^*} and Fe^{3^*} at $log[SO_4^{2^*}] = -2.3$, $log[K^*] = -3.3$, $log[Fe^{2^*}+Fe^{3^*}] = -5$ and 1 atm. total pressure. Source: Miedema, Jongmans & Slager, 1973.

When oxygen penetrates a reduced pyritic sediment, the pyrite is oxidized:

$$FeS_2 + 7/2 O_2 + H_2O = Fe^{2+} + 2 SO_4^{2-} + 2 H^+$$
 (pyrite)

The ferrous iron ions formed are oxidized to ferric hydroxide:

$$6 \text{ Fe}^{2+} + 15 \text{ H}_2\text{O} + 2 \text{ O}_2 = 6 \text{ Fe}(\text{OH})_3 + 12 \text{ H}^+$$

<u>In the presence of carbonates</u>, no lowering of the pH takes place even though much acidity is released. Gypsum is formed:

$$CaCO_3 + 2 H^+ = Ca^{2+} + H^2O + 6 CO_2$$
 and: (carbonates)

$$Ca^{2+} + SO_4^{2-} + 2 H_2O = CaSO_4.2H_2O$$
 (gypsum)

<u>In the absence of carbonates</u>, the hydrogen produced is not neutralized and the pH of the sediment falls sharply. The ferric hydroxyde is transformed to jarosite:

3
$$Fe(OH)_3 + K^+ + 2 SO_4^{2-} + 3H^+ = KFe_3(SO_4)_2(OH)_6 + 3 H_2O$$
(jarosite)

Some of the soluble ferrous sulphate that is formed upon oxidation of pyrite, might diffuse to lower layers where and react directly to jarosite:

$$Fe^{2+} + 2/3 SO_L^{2-} + 1/3 K^+ + 1/4 O_2 + 3/2 H_2O = 1/3 Jarosite + H^+$$

With time, the hydrogen ions will be removed from the system by percolation or lateral flushing, and jarosite will be hydrolyzed to ferric hydroxide:

$$1/3 \text{ Jar} + \text{H}_2\text{O} = 1/3 \text{ K}^+ + \text{Fe(OH)}_3 + 2/3 \text{ SO}_4^{2-} + \text{H}^+$$

The poorly crystallized ferric hydroxide will eventually be transformed to goethite:

$$Fe(OH)_3 = FeOOH + 3H_2O$$
 (goethite)

The foregoing shows that aeration of fresh, non-calcareous pyritic sediments, e.g. by forced drainage, results in large quantities of H^{\star} -ions being released to the soil solution. These H^{\star} -ions lower the pH of the soil and exchange with bases at the cation exchange complex. Once the pH has fallen to a level between pH 3 and pH 4, the clay minerals themselves are

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attacked. Mg and Fe, but mostly Al are released from the clay lattices, and noxious ${\rm Al}^{3+}$ -ions become dominant in the soil solution and at the exchange complex.



Fig. 3. Geology and physiography of the Mekong Delta, Vietnam. Legend: (1) Granite hills, (2) Pleistocene alluvial terraces, (3) Holocene complex of river levees and backswamps, (4) Holocene brackish water sediments in wide depressions, (5) Holocene marine sand ridges and clay plains, and (6) Holocene peat dome.

Figure 3 shows the geology and physiography of the Mekong delta in Vietnam, to illustrate the genesis and physiographic position of Fluvisols (map units 3, 4 and 5) in general and of Thionic Fluvisols in particular. Pyrite occurs in the sediments of broad depressions (map unit 4) where extensive areas of Acid Sulfate Soils are found. There is no pyrite in the soils of map units 3 (fresh water deposits; no sulfate present during sedimentation), and 5 (sedimentation rate too high). The lower tiers of topogenous deltaic peat (map unit 6) may well contain pyrite. Eutric Fluvisols

are confined to narrow strips along the rivers in map unit 3. A large part of the river sediments and most of the marine sediments have lost their fluvic properties, i.e. they are no longer stratified or have developed a cambic B-horizon. As gley phenomena are normally present within 50 cm, they classify as Gleysols. Large areas with Thionic Fluvisols occur in the broad depressions on both sides of the rivers.

CHARACTERISTICS OF FLUVISOLS

Fluvisols are very young soils with weak horizon differentiation; they are mostly AC-profiles and predominantly brown (aerated soils) and/or grey (waterlogged soils). Their texture can vary from coarse sand in levee soils to heavy clays in backswamps. Most Fluvisols show mottling due to alternating reducing and oxidizing conditions. Even if such 'gleyic soil properties' occur in the upper 50 cm of the profile, the soils are not classified as Gleysols because their fluvic properties have priority in the key to Major Soil Groupings.

Fluvisols may show some structure development, but only in part of the profile. Heavy clay soils have commonly a coarse blocky or prismatic structure, whereas lighter textured soils can have a variety of structure types (granular, crumb, subangular blocky). Strongly stratified sediments tend to develop platy structures upon drying. It is evident that the characteristics of Fluvisols are dominated by their recent sedimentation and wetness: stratification, beginning ripening, chemical properties influenced by alternate reducing and oxidizing conditions, and in some environments also soil salinity. Rather special are the characteristics of Thionic Fluvisols; they will be addressed in some detail.

Hydrological characteristics

Most Fluvisols are wet in all or part of the profile due to stagnating groundwater and/or flood water from rivers or tides. Terraces are a part of many river systems but are much better drained than the active flood plain; terrace soils are normally well homogenized and lack fluvic properties.

Physical characteristics

The 'ripening stage' of sedimentary material is expressed by its 'n-value' which indicates the approximate quantity of water that is absorbed by one gram of clay. In the field, the ripening stage is judged by sqeezing a lump of soil material through one's fingers and interpreting the resistance felt. Wet clays and silts which have lost little water since deposition, are soft and unripe. Such soils pose problems for agricultural use; they have a low bearing capacity and machines cannot be used on them. Many clayey Fluvisols have few pores and a low hydraulic conductivity. However, many of coastal landforms were once colonized by a pallustric vegetation (e.g. mangroves or reeds) which left large tubular pores in the sediment that are often still intact. Fluvisols on river levees and coastal sand ridges are porous as well.

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Chemical characteristics

Most Fluvisols are fertile soils; they have neutral or near-neutral pH values which do not impair the availability of nutrients. Coastal sediments contain usually some calcium carbonate (sea shells !), and the exchange complex is saturated with bases from the sea water. On the other hand, an unfavourably high sodium saturation is not uncommon and high salt levels in the soil moisture can be a problem.

The hydrological and physical properties of Thionic Fluvisols are largely similar to those of other Fluvisols but their chemical characteristics are decidedly different. Within the Thionic Fluvisols, a distinction must be made between Potential Acid Sulphate Soils which are not yet oxidized but contain pyrite in the soil material, and Actual Acid Sulfate Soils which are oxidized and acidified.

Unfavourable properties of Potential Acid Sulfate Soils are:

- (1) <u>Salinity</u>: Potential Acid Sulfate Soils are mostly situated in coastal areas with tidal influence.
- (2) Strong acidification upon drainage.
- (3) <u>Low accessibility/trafficability</u>: Potential Acid Sulfate Soils occur in unripe, soft muds.
- (4) <u>High permeability</u>: the root channels of the (former) pallustric vegetation made the sediments excessively permeable to water.
- (5) Flooding of the land: flooding at spring tide may cause damage to crops.
- (6) <u>Engineering problems</u> arise when dikes, etc are constructed on the soft muds. Acidity from the oxidized dikes attacks steel and concrete structures.

Unfavourable properties of Actual Acid Sulfate Soils are:

- (1) Low pH: most plants can tolerate pH values as low as pH 4, but only if the supply of nutrients is well balanced.
- (2) Aluminium toxocity can occur 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 crop.
- (3) <u>Salinity</u> in Actual Acid Sulfate Soils is not always caused by salts from sea water; sulfate can build up in the soil solution to the extent that the soil is to be regarded saline.
- (4) <u>Phosphorus deficiency</u>: high aluminum levels in the soil solution cause precipitation of insoluble Al-phosphates.
- (5) Ferrous iron (Fe²⁺) toxicity is a common problem when rice is cultivated on Actual Acid Sulfate Soils. Reduction of insoluble ferric iron to soluble ferrous iron components takes place in flooded rice fields.
- (6) <u>Acidification of the surface water</u>: when Actual Acid Sulfate Soils are flooded for rice cultivation, soluble ferrous iron can diffuse to the surface water. This surface water contains oxygen, and the ferrous iron can be oxidized to ferric iron again:

$$2 \text{ Fe}^{2+} + 1/2 \text{ O}_2 + 5 \text{ H}_2\text{O} = 2 \text{ Fe}(\text{OH})_3 + 4 \text{ H}^+$$

This process acidifies the surface water and can cause irreparable damage to engineering structures and fish in a very short time.

- (7) N-deficiency: mineralization of organic matter by microbial activity is slow in the wet Actual Acid Sulfate Soils.
- (8) Engineering problems: acidity from surface water attacks steel and concrete structures.
- (9) H₂S toxicity becomes a problem when Actual Acid Sulfate Soils are flooded for long periods (a year or longer). Sulfate can then be reduced to H₂S, which is toxic at very low concentrations.

MANAGEMENT AND USE OF FLUVISOLS

The high natural fertility of most Fluvisols allows cultivation of a wide range of dryland crops on river levees and on higher parts in marine landscapes.

In tropical lowlands with a year-round supply of fresh water, three crops per year are possible. Such places are among the most densily populated parts of the world, and have been under intensive use since pre-historic times. Land tenure is often characterized by small plots, and most production is for home consumption or for local markets.

Paddy rice cultivation is widespread on tropical Fluvisols with satisfactory irrigation and drainage. Paddy land should be dry for at least a few weeks in every year to prevent the soil's redox potential from becoming so low that nutritional problems (iron, H₂S) develop. Besides, a dry period stimulates microbial activity and promotes mineralization of organic matter. Other suitable crops besides rice may 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.

Acid Sulfate Soils are widely left idle. Reclaimed and/or carefully managed Acid Sulfate Soils in the tropics are predominantly used for rice growing; those in the temperate zone are in use as pasture land.

There are two (opposite) strategies thinkable for reclaiming and using <u>Potential Acid Sulfate Soils</u>:

The first one is to drain and completely oxidize 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 expell undesirable aluminium ions from the exchange complex. This approach solves the problem once and for all but has the disadvantages that it is expensive, 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 was desastrous in Senegal, where insufficient water was available for leaching, and in the Netherlands where the first generation of settlers barely survived the reclamation of the Haarlemmermeer polder.

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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 was 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 <u>Actual Acid Sulfate Soils</u>, 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 water 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.

In many coastal lowlands, the 'intensive shallow drainage system' is practiced: shallow ditches are dug at narrow spacings. This system relies on sufficient 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 subsurface soil) to the top. Rice is grown in the shallow depressions between the raised beds.

It is virtually impossible to eliminate the acidity problem completely by applying lime to the soil. Table 1 shows the lime requirement for complete neutralization of soils with various contents of oxidizable sulfur. The neutralizing capacity of a 10 cm layer (without lime; only neutralization by exchange ions) is also given. The table demonstrates the practical impossibility of the liming option: very few farmers can afford to apply to an average Acid Sulfate Soil (say 1.5 percent sulfur and an apparent density of 1.0 Mg/m³) a total of (47 minus 19), or 28, tons of lime. And that covers only the needs of the top 10 cm, assuming that no new acid forms during a subsequent dry spell.

TABLE 1. Lime requirement for complete neutralization of a 10 cm soil layer. After Dent & Raiswell, 1982.

Lime requirement of a 10 cm layer, in tons of lime/ha							Neutralizing capacity of the 10 cm layer, (no lime present) in a clayey soil			
Apparent density										
of the soil (Mg/m ³)	0.5	1	1.5	2	3	4				
0.6	9	19	28	37	56	74	11			
0.8	12	25	37	50	74	112	14			
1.0	16	31	47	62			19			
1.2	19	37	56	74			22			

Many lands with Actual Acid Sulfate Soils are not used for agriculture at all. Such hostile 'wet desert' lands have a pallustric vegetation, a limited fauna, and acid surface water, especially early in the wet season when acid substances dissolve. The situation seems to be a little less bleak in the equatorial climatic zone than in monsoonal climates. There, Acid Sulfate Soils will not dry out as easily as in the monsoon area, and more water is available for management measures throughout the year.

NOTES

GLEYSOLS

Soils in unconsolidated materials, exclusive of coarse-textured materials and alluvial deposits which show fluvic® properties; 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 calcic* or a gypsic* horizon; lacking the characteristics which are diagnostic for Vertisols or Arenosols; lacking salic® properties; lacking plinthite® within 125 cm of the surface.

Key to Gleysol (GL) Soil Units

Gleysols having permafrost^a within 200 cm from the surface.

Gelic Gleysols (GLi)

Other Gleysols having a sulfuric * horizon at less than 125 cm from the surface.

Thionic Gleysols (GLt)

Other Gleysols having andica properties.

Andic Gleysols (GLa)

Other Gleysols having a mollic* A-horizon or a eutric[®] histic* H-horizon.

Mollic Gleysols (GLm)

Other Gleysols having an umbric *A-horizon or a dystric histic H-horizon.

Umbric Gleysols (GLu)

Other Gleysols having a calcic * horizon or a gypsic * horizon, or both, within 125 cm of the surface.

Calcic Gleysols (GLk)

Other Gleysols having a base saturation (by 1M NH_4OAc at pH 7.0) of less than 50 percent, at least between 20 and 50 cm from the surface.

Dystric Gleysols (GLd)

Other Gleysols.

Eutric Gleysols (GLe)

SUMMARY DESCRIPTION OF GLEYSOLS

<u>Connotation</u>: soils with clear signs of excess wetness; from R. <u>gley</u>, mucky soil mass.

<u>Parent material</u>: a wide range of unconsolidated materials, mainly sediments of Pleistocene or Holocene age, with basic to acidic mineralogy.

^{*} Diagnostic horizon; see Annex 1 for full definition.

a Diagnostic property; see Annex 2 for full definition.

106 Gleysols

Environment: depressed areas with shallow groundwater.

<u>Profile development</u>: mostly A(Bg)Cr or H(Bg)Cr profiles. Evidence of reduction processes with or without segregation of iron compounds within 50 cm of the surface.

<u>Use</u>: waterlogging is the main limitation. Gleysols are mostly used for grazing or are covered with swamp forest. Gleysols in the tropics and subtropics are widely planted to rice.

REGIONAL DISTRIBUTION OF GLEYSOLS

The greatest extent of Gleysols is in the boreal and cool humid parts of the world. Gleysols occupy an estimated 623 million hectares worldwide of which some 200 million hectares are in the tropics and subtropics. See Figure 1.

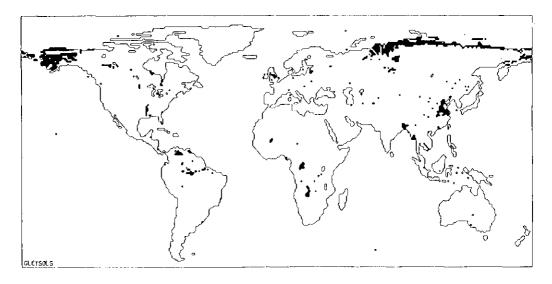


Fig. 1. Gleysols worldwide.

GENESIS OF GLEYSOLS

The formation of Gleysols is conditioned by excessive wetness at shallow depth (less than 50 cm from the soil surface) in some period of the year or throughout the year. Low-redox conditions, brought about by prolonged saturation of the soil material in the presence of organic matter, result in the reduction of ferric iron compounds to (mobile) ferrous compounds. This explains why the permanently saturated subsoil layers of Gleysols have

grey, olive or blue matrix colours: with the iron compounds mobilized and removed, the soil material shows its own colour. Note that the characteristic gley colours will only form if the soil is periodically saturated with water which contains the dissolved products of organic matter decomposition

Subsequent oxidation of transported ferrous compounds (back) to ferric iron oxides can take place near fissures or cracks in the soil and along former root channels where there is a supply of oxygen. Hysteresis between (comparatively rapid) oxidation and (slow) reduction processes leads to a net accumulation of ferric compounds near such aerated spots in the otherwise reduced soil matrix: the subsoil develops a pattern of mottles (around air pockets) and 'root prints' (former root holes lined with iron oxide).

NOTE THAT such 'gleyic soil properties' are strictly associated with movement of the groundwater table; mottled, oxidized horizons occur on top of a 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': a reduced horizon occurs on top of an oxidized subsurface horizon. If peds occur, the reduced horizon tongues into the oxidized subsoil along the faces of the peds (Figure 2).

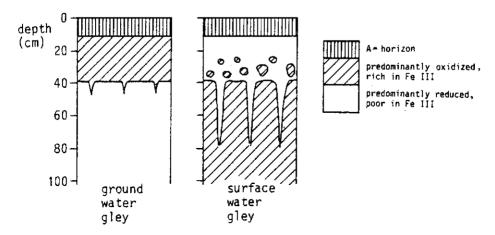


Fig. 2. The different morphologies of soils with gleyic properties ('groundwater gley') and soils with stagnic properties ('surface water gley').

Stagnic properties are marked by the following morphometric characteristics:

(1) Evidence of reduction as in materials with gleyic properties, and

108 Gleysols

- (2) If mottling is present, a dominant moist chroma of 2 or less on the surface of peds and mottles of higher chroma occurring within the peds, or a dominant moist chroma of 2 or less in the soil matrix and mottles of higher chroma or iron-manganese concretions or both occurring within the soil material. If mottling is absent, a dominant moist chroma of 1 or less on the surfaces of peds or in the soil matrix, and
- (3) The dominant moist chroma of the soil matrix and of the surfaces of peds increases in value with to depth.

Soils with stagnic soil properties <u>are not Gleysols</u>; they key out either at the Major Soil Grouping level as Planosols or Plinthosols, or at the second level as 'stagnic' Soil Units in other Major Soil Groupings. Soils which are subject to flooding or show reduction as a result of irrigation are marked by 'inundic' and 'anthraquic' phases.

A special category of hydromorphic soils are the soils of the rice paddies. They may be true Gleysols, but are more often an anthraquic phase of some other Major Soil Grouping which developed stagnic soil properties due to long continued irrigation or inundation. This means that there are two different types of 'Paddy Soils':

- the Gleysols and Fluvisols of wetlands (with a completely reduced subsoil), and
- (2) anthraquic Paddy Soils formed in originally well drained land.

A thin, compacted and slowly permeable plough sole is formed at the depth of cultivation in all Paddy Soils. Anthraquic Paddy Soils often possess a reddish brown to black accumulation horizon of iron and/or manganese oxides at some depth below the plough sole. This accumulation layer formed as a result of intensive reduction processes in the reduced surface layer and translocation and precipitation of iron and manganese compounds in the (oxidized) subsoil. The accumulation layer may indurate to an impenetrable hardpan.

CHARACTERISTICS OF GLEYSOLS

Gleysols have normally a spongy or matted litter layer resting on a dark grey Ah-horizon. This horizon changes sharply into a mottled grey or olive cambic Bg-horizon. With depth, the Bg-horizon grades into a grey, olive or blue, completely anaerobic Cr-horizon.

In Gleysols of the savannahs, the topsoil consists normally of a very dark grey to black heavy clay. In some profiles this continues to 2 m or more, whereas in others it gives place to a pale grey horizon with prominent mottling (Fitzpatrick, 1986). Close to the surface, the soil structure is medium blocky or even crumb, but it becomes coarsely prismatic at some depth. The soil material is hard when dry and sticky when wet.

Where Gleysols remain waterlogged throughout the year, except perhaps for short periods, the topsoil is typically a mixed organic and mineral (muck) H-horizon; it overlies a mottled clay or sandy clay below which is material subject to permanent anaerobic weathering.

The dominant attribute of Gleysols is their prolonged saturation with water, associated with lack of aeration, an unfavourable root environment and poor conditions for the soil fauna. Repeated wetting and drying may also cause soil densification due to weakening of interparticle bonds during saturation and contraction of soil particles upon desaturation. Gleysols in depressions or at the foot of slopes compare favourably 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 levels of phosphorus and potassium. This is partly the result of a generally finer texture and slower organic matter decomposition, but is also caused by influx of ions from adjacent (higher) lands.



Fig. 3. Wetland rice fields in a river valley near Rokupr, Sierra Leone. The surrounding uplands are planted to oil palm.

MANAGEMENT AND USE OF GLEYSOLS

The main obstacle to the utilisation of Gleysols is the necessity to install a drainage system, either designed to lower the groundwater table, or to intercept seepage or surface runoff water. Adequately drained

110 Gleysols

Gleysols are widely used for arable cropping, dairy farming or horticulture.

Where the surface soil is high in organic matter and pH values are low, liming creates a better habitat for micro- and meso-organisms and enhances the decomposition of soil organic matter. If (too) wet soils are cultivated, the structure of the soil will be destroyed for a long time. Gleysols in low-lying areas with unsatisfactory drainage possibilities are therefore best kept under a permanent grass cover or under (swamp) forest.

The use of Gleysols for wetland rice cultivation has already been mentioned. The difficulties discussed for Thionic Fluvisols apply also to Thionic Gleysols (see under Fluvisols). Where Gleysols occur in narrow valley floors, these are modified into broad and level steps, each with a water-retaining bund. Figure 3 presents a picture of wetland rice fields on Gleysols in a small valley in Sierra Leone.

Gleysols can be put under tree crops only after the water table has been lowered with deep drainage ditches. Alternatively, the trees are planted on ridges that alternate with shallow depressions in which rice is grown. This 'sorjan' system is widely applied in tidal swamp areas with pyritic sediments in south-east Asia.

MAJOR LANDFORMS IN ERODING UPLANDS

Eroding uplands are marked by the occurrence of unstable rocky slopes and outcrops of bedrock. They are particularly common in the following environments:

- (1) <u>High mountain areas</u> which were exposed to glaciation in the recent past; any old soil cover was scraped off during the latest glacial advances). These regions include almost all mountain chains in the temperate zone (Alps, Canadian Rockies) and much of the tropical high mountain belts above the limit of Pleistocene glaciation which lies above 3000 m at the equator.
- (2) <u>Middle mountains</u> (Caledonian and Hercynian orogenic areas), mainly in the temperate zone. These regions were not or only slightly glaciated in the recent geological past but suffered so much periglacial slope action that they possess extremely stony soils.
- (3) Arctic shield areas such as Scandinavia and northern Canada, which experienced glaciation and consist predominantly of solid rock (scraped clean) or of stony moraine deposits.

LANDFORMS IN HIGH MOUNTAIN AREAS

The highest mountains in the world were all formed comparatively recently, viz. in the Tertiary and Quaternary. They belong to mountain <u>belts</u> that are situated at plate boundaries and their properties depend to some extent on the nature of these plate boundaries. There are three basic types:

- (1) Island arcs (a curved array of volcanic islands) form where one oceanic plate is being subducted below another. In an early stage of the subduction process, the islands are almost entirely volcanic; examples of such islands are the Lesser Antilles, or the Aleutian Islands off Alaska. Later, the islands acquire a more complex geological structure such as Japan and the Philippines. Volcanism is initially basaltic to andesitic and becomes increasingly andesitic later on.
- (2) Cordilleran mountains form where an oceanic crust is being subducted below a continental plate. Examples of this type are the Andes and the American Cordilleras. Andesitic and rhyolitic volcanism are prominent in cordilleran mountains. The uplift to the present elevation is rather recent (Plio-Pleistocene) in all cordilleran mountains but orogeny (mountain formation) began normally already in the Paleozoic or even in the Precambrian. Folded sedimentary rocks, igneous intrusive bodies (granite, granodiorite), and older metamorphic rocks from all geologic periods can be found in these mountain belts.
- (3) Continent-continent collision mountains differ from the previous types in that volcanism plays a much more restricted role. Moreover, compression is much stronger than in cordilleran belts, so that 'nappes', low-angle overthrust sheets which may have been displaced hundreds of kilometres from their original position, are very common.

112 Landforms in eroding uplands

Otherwise, a similar array of igneous, sedimentary and metamorphic rocks of greatly differing ages can be found as in cordilleran belts. Examples of continent-continent collision mountains are the Alps and their continuation into Eastern Europe, Iran and the Himalayas; these mountains are being formed as a result of the collision of the African Shield, the Arabian Peninsula and the Indian Shield with the Eurasian continent. That orogeny continues in these areas is demonstrated by the fact that the Kaukasus is still being uplifted at a rate of 2 cm per year. The underthrusting of the Indian Plate below the Eurasian continent has made the Himalayas the highest mountains on earth.

Rock outcrops and shallow, stony weathering mantles are prominent in all three mountain types, especially in the zone between the actual snow limit and the lowermost extension of the glaciers in the last glaciation. (The present glaciers and snow fields are not considered here.) Typical erosional forms in this zone are the 'cirques', amphitheatre-like former glacier basins, and trough valleys, former fluvial valleys modified into a U-shape by glacial abrasion. Both have steep upper slopes, usually consisting of rough rock outcrops, and extensive lower slopes of rock debris. The size of the rock fragments varies with their lithology; highly fissile rocks such as shales, schists and dolomite desintegrate to fine debris whereas granites give huge angular boulders.

Besides rock outcrops and in-situ rock weathering, this zone contains soil parent materials of glacial, fluvioglacial and lacustrine origin. The glacial deposits consist of unsorted, clay-rich 'glacial tills' on the bottoms of trough valleys and cirques, and 'moraine ridges' of coarse, unsorted, and usually clay-poor material.

The fluvioglacial deposits are meltwater deposits of well-sorted sands and gravels similar to the deposits of (other) braided streams.

<u>The lacustrine deposits</u>, formed in temporary lakes at retreating glacier margins, are stratified as a result of seasonal variations in sediment influx ('varves').

Soil formation in these materials is slow (high altitude, low temperature) and went on for only 10,000 years or less. As a result, (poorly developed) LEPTOSOLS and REGOSOLS are strongly represented in high mountain areas.

The zone below the Pleistocene glacial limit contains older weathering profiles, with more developed soils than occur in the recently deglaciated regions.

LANDFORMS IN MIDDLE MOUNTAINS

The 'middle mountains' of Central Europe and eastern North America are gigantic continental collisional structures. They formed on both sides of the Atlantic and are known as the Caledonian and Hercynian Massifs. They extend from northern Norway to Mauretania in Eur-Africa and from eastern Greenland to the southern USA. Their folded sedimentary and metamorphic rocks indicate a complex orogenic history.

Both massifs have an internal zone of crystalline rocks, and an external zone of low-grade metamorphic and sedimentary rocks. Typical internal-zone crystalline provinces are the Piedmont province of the USA, Brittany, and the Central Massif, Vosges, Black Forest and Bohemian Massif in central Europe. The Valley and Ridge province in the USA, and the Ardennes, the Rheinische Schiefergebirge and Devonshire in Europe are mainly sedimentary. As orogenesis ended long ago, essentially before the start of the Mesozoic, most Caledonian and Hercynian Massifs have degraded to low peneplains such as Brittany (France). Where differences in elevation still exist in such areas, they are caused by:

- (1) <u>different rock types</u> with different resistance: hard limestones, dolomites, sandstones or quartzites form 'cuestas' or 'hogbacks'; soft shales and marls form gently sloping valleys. Granites form 'tors' (huge heaps of boulders) embedded in sandy saprolite, such as those at Dartmoor. Much of this structural relief was smoothed by slope processes under (peri)glacial conditions during the ice ages.
- (2) <u>later uplift</u> generated by younger orogeny outside the massifs themselves; the Alpine orogeny caused strong uplift and incision of entrenched meanders in the Ardennes and Rheinische Schiefergebirge. The Norwegian Caledonides and the Appalachians were uplifted as a corollary of the separation of North America and Europe in the Mesozoic and Tertiary.

Peneplain formation has continued until the Tertiary in many of these areas; relic (sub)tropical soils in some shielded areas on peneplain remnants testify to the occurrence of warmer climates in the past.

The Caledonian ranges of western Norway, Scotland and Ireland were strongly modified by glacial action. The fjords of Norway are exceptional in their enormous depth, down to 1300 m, more than 1 km below the continental shelf.

In contrast, Hercynian massifs in Central Europe were hardly affected by glacial action because they remained below the snowline. But they have experienced harsh climates: there is ample evidence of frost shattering, frost heaving, periglacial slope processes and typical structures related to thawing and freezing on top of permafrost. All these processes were instrumental in the mixing of existing soil material with fresh rock fragments from deeper down, so that many soils, also on the uplifted plateaus, are very stony.

Rock outcrops are rare in plateau areas of sedimentary and low-grade metamorphic rocks but very common on the steep slopes of rejuvenated valleys. LEPTOSOLS and REGOSOLS are also common there.

LANDFORMS ON ARCTIC SHIELDS

Precambrian shields such as the Scandinavian and Canadian shields are highly complex structures; their geologic history spans more than three billion years. Arctic shield areas have in common with high mountain areas that they were glaciated in the Pleistocene and consist of fresh rock outcrops alternating with glacial and fluvioglacial deposits. The latter

114 Landforms in eroding uplands

stem almost exclusively from the last deglaciation when the vast continental ice sheets of the Northern Hemisphere melted down.

The arctic shield areas are generally subdued lowlands with little relief; frost weathering, steep rock outcrops and slopes are much less common than in the high mountains. Otherwise, much of what was said in the preceding paragraphs about landforms in glacial deposits applies also to arctic shields. LEPTOSOLS and REGOSOLS are common soils there.

LEPTOSOLS

Soils which are limited in depth by continuous hard rock® or highly calcareous® material 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 of the surface; having no diagnostic horizons other than a mollic*, umbric*, or ochric* A-horizon, with or without a cambic* B-horizon.

Key to Leptosol (LP) Soil Units

Leptosols which are limited in depth by continuous hard rock^a within 10 cm of the surface.

Lithic Leptosols (LPs)

Other Leptosols having permafrost $^{\mathrm{a}}$ within 200 cm of the surface.

Gelic Leptosols (LPi)

Other Leptosols having a mollic * A-horizon which contains or immediately overlies calcareous $^{\hat{a}}$ material with a calcium carbonate equivalent of more than 40 percent.

Rendzic Leptosols (LPk)

Other Leptosols having a mollic* A-horizon.

Mollic Leptosols (LPm)

Other Leptosols having an umbric * A-horizon.

Umbric Leptosols (LPu)

Other Leptosols having an ochric * A-horizon and a base saturation (by 1M NH_{ℓ}OAc at pH 7.0) of less than 50 percent.

Dystric Leptosols (LPd)

Other Leptosols.

Eutric Leptosols (LPe)

SUMMARY DESCRIPTION OF LEPTOSOLS

Connotation: weakly developed, shallow soils; from Gr. leptos, thin.

<u>Parent material</u>: various kinds of rock, or unconsolidated materials with less than 20 percent fine earth.

<u>Environment</u>: mostly lands at high or medium altitude, and with strongly dissected topography. Leptosols are found in all climate zones, in particular in areas with a high rate of erosion.

Diagnostic horizon; see Annex 1 for full definition.

Diagnostic property; see Annex 2 for full definition.

116 Leptosols

<u>Profile development</u>: mostly A(B)R or A(B)C profiles with a thin A-horizon. Leptosols in calcareous weathering material have normally a mollic A-horizon with a high degree of biological activity.

<u>Use</u>: unattractive soils for arable cropping; limited potential for tree crop production or extensive grazing. Leptosols are best kept under forest.

REGIONAL DISTRIBUTION OF LEPTOSOLS

Leptosols constitute the most extensive Major Soil Grouping on earth, with an estimated total area of 2260 million hectares (900 million hectares in the tropics and subtropics). Lithic Leptosols in montane regions form the majority of all Leptosols. See Figure 1.

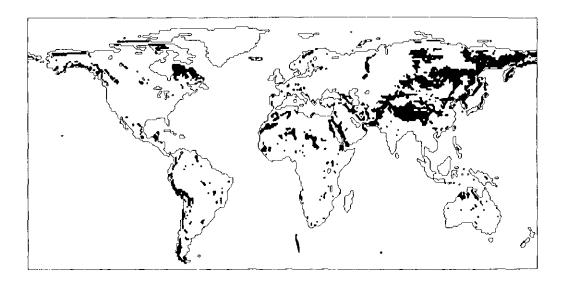


Fig. 1. Leptosols worldwide.

GENESIS OF LEPTOSOLS

Leptosols are genetically young soils and evidence of soil formation is normally limited to a thin A-horizon over a beginning (cambic) B-horizon or directly over the unaltered parent material.

The principal soil forming process in Rendzic and Mollic Leptosols is the dissolution and subsequent removal of carbonates. A relatively small residue remains behind and is thoroughly mixed with stable humifying organic matter and, in Rendzic Leptosols, fragments of undissolved limestone. Swelling and shrinking smectite clays in the mineral residue are accountable for the dominance of blocky structures. Umbric Leptosols occur mostly

on siliceous parent rock in (montane) regions with a cool climate and a high precipitation sum,

Leptosols are highly variable in soil depth, texture and composition, particularly in areas with active erosion where they often occur in association with Cambisols.

CHARACTERISTICS OF LEPTOSOLS

Most Leptosols have an A(B)R configuration of only weakly expressed horizons. Rendzic and Mollic Leptosols have pronounced morphological features. Their dark brown or black calcareous organo-mineral surface soil, in Rendzic Leptosols speckled with white fragments of limestone rock, has a well developed crumb or granular structure, or a vermicular structure with abundant earth worm casts. Below, there is an abrupt change to the underlying rock or there may be a narrow transition horizon.

The physical, chemical and biological properties of non-calcareous Leptosols vary considerably because they are largely determined by the characteristics of the parent material and the climate. Calcareous Leptosols have generally better physical and chemical properties than non-calcareous ones and are also less diverse.

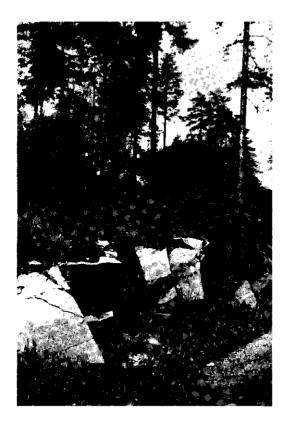


Fig. 2. A Dystric Leptosol under pine forest in Norway.

118 Leptosols

All Leptosols are freely drained; they lack gleyic or stagnic properties at shallow depth and they are also free from high levels of soluble salts. Their shallowness and/or stoniness, associated with a low water holding capacity, are serious limitations. The natural vegetation varies with the climate but is generally richer on calcareous Leptosols than on acid ones. Earth worms, enchytraeid worms, arthropods and bacteria are the chief soil organisms. The soil fauna may be temporarily inactive due to drought.

MANAGEMENT AND USE OF LEPTOSOLS

Most non-calcareous Leptosols are not cultivated. They have a resource potential for wet-season grazing and as forest land. Rendzic Leptosols in southeast Asia are planted to teak and mahogony; those in the temperate zone are under deciduous forest whereas acid Lithic, Umbric and Dystric Leptosols are more commonly under pine forest (Figure 2).

The chemical soil fertility of Leptosols is often higher on hill slopes than on more level land. One or a few good crops can perhaps be grown on such slopes but at the price of severe soil erosion.

Erosion is a serious problem, particularly in montane areas where a high population pressure promotes overexploitation of the land. Summer tourism, the construction of skiing grounds and the recent deterioration of forests by acid atmospheric deposition threaten large areas of vulnerable Leptosols in temperate regions.

Steep slopes with shallow and stony soils can be transformed into cultivable land through terracing, the removal of stones by hand and their use as terrace fronts. Agro-forestry (a combination or rotation of arable crops and forest under strict control) holds promise but is largely still in an experimental stage. The often excessive internal drainage of Leptosols can cause drought even in a humid environment.

REGOSOLS

Soils from unconsolidated materials, exclusive of materials that are coarse-textured or show fluvic® properties, having no diagnostic horizons other than an ochric* or umbric* A-horizon; lacking gleyic® properties within 50 cm of the surface; lacking the characteristics which are diagnostic for Vertisols or Andosols; lacking salic® properties.

Key to Regosol (RG) Soil Units

Regosols having permafrost^a within 200 cm of the surface.

Gelic Regosols (RGi)

Other Regosols having an umbric * A-horizon.

Umbric Regosols (RGu)

Other Regosols which are gypsiferous $^{\mathbf{D}}$ at least between 20 and 50 cm from the surface.

Gypsic Regosols (RGj)

Other Regosols which are calcareous $^{\rm a}$ at least from 20 to 50 cm from the surface.

Calcaric Regosols (RGc)

Other Regosols having a base saturation (by 1M NH_4OAc at pH 7.0) of less than 50 percent at least from 20 to 50 cm from the surface.

Dystric Regosols (RGd)

Other Regosols.

Eutric Regosols (RGe)

SUMMARY DESCRIPTION OF REGOSOLS

Connotation: soils in the weathered shell of the earth; from Gr. rhegos, blanket.

Parent material: finely grained unconsolidated weathering material.

<u>Environment</u>: all climate zones, both at low and high altitudes; mostly in land of level to rolling topography.

<u>Profile development</u>: AC-profiles, with no other diagnostic horizon than an othric or an umbric A-horizon. Profile development is minimal as a consequence of young age and/or slow soil formation e.g. because of low soil temperatures or prolonged dryness.

^{*} Diagnostic horizon; see Annex 1 for full definition.

a Diagnostic property; see Annex 2 for full definition.

120 Regosols

<u>Use</u>: land use and management vary widely among the different soils in this grouping, from capital-intensive irrigated farming to low volume grazing. Many Regosols are not used at all.

REGIONAL DISTRIBUTION OF REGOSOLS

Regosols cover an estimated 900 million hectares worldwide, mainly in the arctic region. Some 100 million hectares of Calcaric and Gypsic Regosols lie in the semi-arid (sub)tropics. Regosols occur in montane areas above the vegetation zone but also at sea level. See Figure 1.

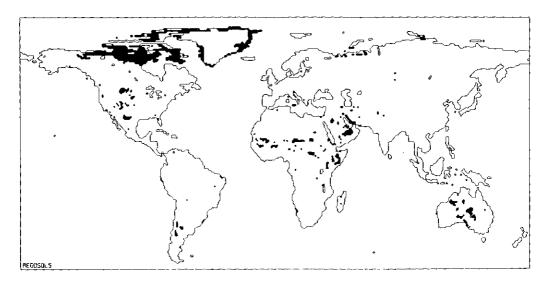


Fig. 1. Regosols worldwide.

GENESIS OF REGOSOLS

Soil forming processes have had a very limited effect in Regosols, either because pedogenesis went on for a relatively short period of time or because of conditions which retard soil formation such as a dry and hot desert climate or permafrost.

Profile development is typically limited to a thin ochric A-horizon over the unaltered parent material. The paucity of pedogenetic transformation products accounts for the low coherence of the matrix material and makes that soil colours are normally (still) determined by the composition of the mineral soil fraction (Schmidt-Lorenz, 1986).

Regosols in acid parent material may develop an umbric A-horizon if the soil organic matter decomposes slowly, e.g. in regions with a short summer and a long and cold winter season.

In regions with a considerable evaporation surplus over precipitation, some lime and/or gypsum may have accumulated at shallow depth in the profile but not to the extent that a calcic or gypsic horizon is present.

CHARACTERISTICS OF REGOSOLS

The morphology of Regosols is primarily determined by the character of the parent material and the climate. In cold climates, the A-horizon contains poorly decomposed organic matter whereas (ochric) A-horizons tend to be thin, low in organic matter and generally weakly developed in hot and dry climates. The content of weatherable minerals varies from low to extremely high (little transformation). The low coherence of the matrix material makes most Regosols sensitive to erosion; the low water holding capacity and high permeability to water make them sensitive to drought.

MANAGEMENT AND USE OF REGOSOLS

Land use and management procedures applied on Regosols vary widely. The extensive arctic Regosols have minimal agricultural significance. Regosols in the steppe region with 500-1000 mm of rainfall per annum need irrigation for good production. The low moisture holding capacity of these soils makes frequent applications necessary; sprinkler or trickle irrigation solves the problem, but is rarely economic.

In areas where the total rainfall sum exceeds 750 mm per annum, the entire profile is raised to its (low) field capacity quite early in the wet season and, there, improvement in methods of dry farming may be a better form of investment (Fitzpatrick, 1980). Regosols in tropical montane areas are mainly used for extensive grazing; many are not used at all.

NOTES

MINERAL SOILS CONDITIONED BY THEIR LIMITED AGE:

CAMBISOLS

CAMBISOLS

Soils having a cambic B-horizon and no diagnostic horizons other than an ochric or an umbric A-horizon or a mollic A-horizon overlying a cambic B-horizon with a base saturation (by 1M NH₄OAc at pH 7.0) of less than 50 percent; lacking salic properties; lacking the characteristics diagnostic for Vertisols or Andosols; lacking gleyic properties within 50 cm of the surface.

Key to Cambisol (CM) Soil Units

Cambisols having permafrost within 200 cm of the surface.

Gelic Cambisols (CMi)

Other Cambisols showing gleyic^a properties within 100 cm of the surface.

Gleyic Cambisols (CMg)

Other Cambisols showing vertic^a properties.

Vertic Cambisols (CMv)

Other Cambisols having an umbric* A-horizon or a mollic* A-horizon overlying a cambic* B-horizon with a base saturation (by 1M NH₄OAc at pH 7.0) of less than 50 percent.

Humic Cambisols (CMu)

Other Cambisols which are calcareous a at least between 20 and 50 cm from the surface.

Calcaric Cambisols (CMc)

Other Cambisols having a cambic *B-horizon with ferralic a properties.

Ferralic Cambisols (CMo)

Other Cambisols having a base saturation of less than 50 percent (by 1M NH_2OAc at pH 7.0) at least in some part of the B-horizon.

Dystric Cambisols (CMd)

Other Cambisols which have a strong brown to red B-horizon (rubbed soil has a hue of 7.5YR and a chroma of more than 4, or a hue redder than 7.5YR).

Chromic Cambisols (CMx)

Other Cambisols.

Eutric Cambisols (CMe)

^{*} Diagnostic horizon; see Annex 1 for full definition.

Diagnostic property; see Annex 2 for full definition.

126 Cambisols

SUMMARY DESCRIPTION OF CAMBISOLS

<u>Connotation</u>: soils with beginning horizon differentiation through changes in colour, structure, and/or texture; from L. <u>cambiare</u>, to change.

<u>Parent material</u>: medium and fine-textured materials derived from a wide range of rocks, mostly in colluvial, alluvial or eolian deposits.

<u>Profile development</u>: ABwC profiles. Cambisols are moderately developed soils characterized by slight or moderate weathering of the parent material and by absence of appreciable quantities of illuviated clay, organic matter, aluminium and/or iron compounds.

<u>Environment</u>: level to mountainous terrain; arctic to tropical climates; wide range of vegetation types.

<u>Use</u>: a wide variety of agricultural uses; climate, topography, shallowness, stoniness, or low base status may pose restrictions on the use. In steep lands mainly grazing and/or forestry.

REGIONAL DISTRIBUTION OF CAMBISOLS

Cambisols cover 925 million hectares, or 7 percent of the earth's land surface; 500 million hectares are actual or potential crop land. Cambisols are particularly common in temperate and boreal regions that were under the influence of ice during recent glacial periods, partly because the soil's parent material is still young, but also because soil formation is slow under the low temperatures (or even permafrost) of northern regions. See Figure 1.

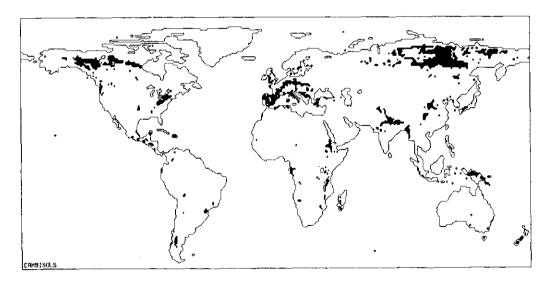


Fig. 1. Cambisols worldwide.

In the tropics, the largest continuous surface of Cambisols is in the (young) alluvial plains and terraces of the Ganges-Brahmaputra system. Cambisols are widespread in areas with active geologic erosion, where they may occur in association with highly developed soils such as Acrisols and Ferralsols; the mountainous regions of Papua New Guinea are an example. As Cambisols occur often in association with other soils, many Cambisol areas are of limited extent and not represented in Figure 1.

GENESIS OF CAMBISOLS

The cambic B-horizon is the only diagnostic feature that all Cambisols have in common. A cambic B-horizon can also occur in other Major Soil Groupings but it is not a differentiating characteristic there because other properties are given a higher priority (salic properties in Solonchaks, gleyic properties in Gleysols, andic properties in Andosols, etc.) The fact that Cambisols key out last in the taxonomic hierarchy of Major Soil Groupings implies that Cambisols include many soils that just missed out on one or more requirements for other soil groupings. In other words, many Cambisols are in a transitional stage of development from a young soil to a mature soil with an argic, natric, spodic, or ferralic B-horizon. Nonetheless, a cambic B-horizon can be quite stable, viz. where the environment counteracts pedogenetic change, e.g. by low temperatures or even permafrost, or by low precipitation, or impeded drainage, or highly calcareous or weathering-resistant parent materials, or by a continuous supply of ions to replenish ions lost by leaching, or by a slow but continuous rate of erosion that is in equilibrium with weathering processes.

A cambic B-horizon must be seen as a 'minimum B-horizon' with <u>beginning</u> soil formation. It is marked by:

- recognizeable soil structure or, alternatively, absence of rock structure, and/or
- (2) greyish reduction colours, and/or
- (3) a stronger chroma, redder hue or higher clay content than the underlying horizon.

In practice, a cambic B-horizon is any section of a soil profile situated between an A-horizon and a relatively unaltered C-horizon, that has soil structure rather than rock structure and that differs from the C-horizon in colour and/or clay content. The fact that Cambisols are in an early stage of soil formation is frequently evidenced by the presence of appreciable quantities of weatherable minerals and the absence of any signs of advanced pedogenesis.

Most cambic B-horizons formed upon weathering/transformation of primary minerals in a situation of <u>free internal and external drainage</u>. In a weakly acid environment, hydrolysis of iron-containing minerals (biotite, olivine, pyroxenes, amphiboles, etc) produces ferrous iron that is oxidized to ferric oxides and hydroxides (e.g. goethite, haematite). This 'free iron' coats sand and silt particles, and aggregates clay, silt and sand to peds.

The soil becomes structured and yellowish brown to reddish in colour. Besides ferric oxides, aluminium oxides and hydroxides, and silicate clays are formed. There is some leaching of bases but no clear migration of Fe, Al, organic matter or clay. This oxidative weathering process is not limited to the cambic B-horizon; it occurs just as well in the A-horizon and may even be stronger there than in the B-horizon, but its effects are outweighed by the dark colours of accumulating soil organic matter.

Weathering under conditions of <u>impeded drainage</u> leads to new formation of clays, and segregation or removal of free iron oxides at some depth. Soil forming processes are impeded by the lack of drainage, so that there is only limited leaching of soluble compounds, and little clay illuviation.

The processes leading to the formation of a cambic B-horizon are fundamentally the same in all climate zones but the intensities of chemical and biological transformations are considerably higher in the (humid) tropics than elsewhere. In the humid tropics, Cambisols can form in a few years time. In dry tropical regions, soil formation is interrupted for shorter or longer periods.

The wide variation among Cambisols is perhaps best illustrated by mentioning a number of typical situations where these soils occur:

<u>In the humid tropics</u>, Cambisols are widespread in highland regions (Umbric Cambisols) and in hilly to mountainous terrain, mainly at medium altitudes. The steepest slopes have no soil at all, or only Lithic Leptosols; Dystric or Ferralic Cambisols occur on the moderately steep hillsides and (residual) Acrisols or Ferralsols in more stable sites.

<u>In the drier subtropics</u>, Calcaric and Chromic Cambisols may form upon erosion of Luvisols or Kastanozems. Vertic Cambisols occur in association with Vertisols on the Deccan Plateau in India, where long-continued cultivation and soil erosion have produced shallow soils that do not qualify as Vertisols.

<u>In the temperate zone</u>, Cambisols are particularly common in alluvial, colluvial and eolian deposits; they are predominantly Eutric or Dystric Cambisols, depending on the nature of the parent material. The majority of all Cambisols in the northern hemisphere are Gelic Cambisols.

<u>In wetlands</u>, Gleyic Cambisols can be found in association with Gleysols and Fluvisols, and, in somewhat better drained positions such as terraces, together with Luvisols, Acrisols, Plinthosols.

CHARACTERISTICS OF CAMBISOLS

The typical Cambisol profile has an ABwC horizon sequence with an ochric, mollic or umbric A-horizon over a cambic B-horizon that has normally a yellowish-brown colour but that may also be an intense red. In poorly drained terrain positions, oxidation/reduction mottles occur in the cambic B-horizon. Textures are loamy to clayey. Some clay coatings may be detectable in the cambic B-horizon but the clay content is commonly (still) highest in the A-horizon.

It is not well possible to sum up all mineralogical, physical and chemical characteristics of Cambisols in one generalized account because Cambisols occur in such widely differing environments. However:

- (1) most Cambisols contain at least some weatherable minerals in the silt and sand fractions.
- (2) most Cambisols occur in regions with a precipitation surplus but in terrain positions that permit surficial discharge of excess water.
- (3) most Cambisols are medium-textured and have a good structural stability, a high porosity, a good water holding capacity and good internal drainage.
- (4) most Cambisols have a neutral to weakly acid soil reaction, a satisfactory chemical fertility and an active soil fauna.

NOTE THAT there are numerous exceptions to the above generalizations !

MANAGEMENT AND USE OF CAMBISOLS

On the whole, Cambisols make good agricultural land and are intensively used. The Eutric Cambisols of the temperate zone are among the most productive soils on earth. Distric Cambisols, though less fertile, are used for (mixed) arable farming and as grazing and forest land. Stoniness (rudic phase; see Annex 3) and shallowness are the commonest limitations in the temperate zone. Cambisols on steep slopes are best kept under forest; this is particularly true for the Umbric Cambisols of highlands.

The Vertic and Calcaric Cambisols in (irrigated) alluvial plains in the dry zone are intensively used for the production of food and oil crops. Eutric, Calcaric and Chromic Cambisols in undulating or hilly (mainly colluvial) terrain are planted to a variety of annual and perennial crops and tree crops or are used for grazing.

The Dystric and Ferralic Cambisols of the humid tropics are poor in nutrients but still richer than neighbouring Acrisols or Ferralsols and they have a higher cation exchange capacity. The Gleyic Cambisols of alluvial plains under paddy rice are highly productive soils.

NOTES

MINERAL SOILS CONDITIONED BY A WET (SUB)TROPICAL CLIMATE:

PLINTHOSOLS FERRALSOLS NITISOLS ACRISOLS ALISOLS LIXISOLS

MAJOR LANDFORMS IN TROPICAL REGIONS

Large parts of the tropics belong to one of the following three morphostructural units:

- Precambrian shield areas, which constitute major parts of eastern South America, equatorial Africa, and central and southern India;
- (2) Young alpine fold belts, such as the equatorial Andes and Central America, and major parts of Southeast Asia;
- (3) <u>Tropical alluvial plains</u>, including fluvial sedimentary basins such as the Amazon basin, the Zaïre basin and the Indus-Ganges basin, as well as coastal plains, e.g. the coastal plains of the Guianas, the Niger delta and the Mekong delta.

Each of these units have their characteristic landforms, developed in characteristic parent materials through characteristic geomorphic and tectonic processes. Landforms in high mountain areas (above 3000 m) have been discussed in a previous chapter, just as the (tropical) lowlands. The present chapter is concerned with the major landforms in Precambrian shield areas and in the lower ranges of young alpine fold belts (below 3000 m) in the humid tropics.

LANDFORMS IN PRECAMBRIAN SHIELD AREAS

Precambrian shield areas constitute the oldest cores of continents. They are remnants of mountains which formed more than 600 million years ago and have since eroded down to undulating plains that extend only a few hundred metres above the present sea level. Shield areas form part of the lithospheric plates which move along the earth's surface at a rate of several centimetres per year. Horizontal movement brings about vertical adjustment, and therefore zones of active subsidence are found next to zones of active uplifting (though much less vigorous than in young mountain belts).

The Precambrian spans 80 percent of the geological history of the earth and includes many periods of mountain building, erosion and sedimentation. Igneous, sedimentary and metamorphic rocks of Precambrian age exist in great variety but crystalline (plutonic and metamorphic) rocks predominate. The Precambrian formations are made up of the following broad units:

- (1) High-grade metamorphic belts are usually narrow (tens of kilometres across) and consist largely of strongly metamorphosed rocks, many of which originate from sedimentary rocks. The lithology of the metamorphic belts is accordingly varied: metamorphosed limestones (marbles), metamorphosed sandstones (quartzites), metamorphosed basalt flows or dykes (amphibolites), and a considerable proportion of rocks of more or less granitic composition. The variation in chemical and mineralogical composition of these rocks explains the wide variation in landforms and soils.
- (2) Greenstone belts are narrow (a few tens or hundreds of kilometres across) but can stretch over thousands of kilometres. They consist mainly of volcanic rocks (basalts, andesites, rhyolites) with varying

amounts of intercalated sedimentary rocks that have normally been converted to schists and phyllites by low-grade metamorphism. A characteristic feature of greenstone belts is the occurrence of tonalite intrusions, normally with oval outlines on the geological map. Tonalite is a granite-like rock of plutonic origin and usually has plagioclase as the sole feldspar. As plagioclase weathers easily, tonalite areas are more deeply weathered than nearby granite areas. Examples are the Koidu Basin in Sierra Leone and the Brokopondo lake in Surinam.

- (3) Granite areas are often associated with either migmatites (banded rocks formed through partial melting of sediments deep in the crust), or rhyolitic volcanic rocks (the extrusive equivalent of granites).
- (4) Platform areas consist of horizontal sedimentary rocks, commonly sandstones, on top of the Precambrian shield; the surrounding, uncovered shield areas are referred to as 'basement' areas.

Tropical shield areas were not affected by glacial, periglacial or eolian processes in the recent past; they are mainly modified by the action of rain and by fluvial processes. The amount and intensity of precipitation on the one hand, and the protection provided to the land by the vegetation cover on the other, determined how water could shape the surface. We shall restrict the discussion to areas under either of two vegetation types:

- (1) a tropical rain forest, or
- (2) a savannah vegetation.

In areas under rain forest, most precipitation is intercepted by the canopy from where it trickles down to the forest floor and infiltrates into the soil. There, it reaches the groundwater and is eventually discharged by rivers. On its path, it contributes to strong chemical weathering of rock because dissolution processes are favoured by the low ionic strength and comparatively high temperature of the water. The saprolite (rotten rock) under a rain forest vegetation extends often to tens of metres depth, or even to hundreds of metres. The saprolite is less thick on granite (say, 10-20 metres) than on metamorphic rock (40-70 metres; data from Surinam). The strong chemical weathering and comparatively low surface runoff cause gradual deepening of the weathering front, a process known as 'etching'. The saprolite is normally clayey because all feldspars and ferromagnesian minerals are weathered to clay minerals and (sesqui-)oxides; the sand content of the saprolite reflects the content of coarse quartz in the original parent rock. Thoroughly weathered saprolites are chemically very poor, despite the luxuriant rain forest vegetation.

<u>In very arid areas</u>, the vegetation is widely spaced because each individual plant needs a large volume of soil for its water supply. In such open land, a single downpour can cause torrential 'sheet floods'. Higher precipitation sums allow a denser vegetation cover but go with increased erosivity of the rain. In the intermediate situation, i.e. in the semi-arid savannahs and prairies, surface runoff and denudation are most severe.

Under a sparse vegetation, the rate of erosion may well exceed the rate of weathering and soil formation which leads to 'stripping' of the land. Under more protective vegetation types, e.g. under rain forest, etching

predominates. The present humid tropics have experienced profound climate changes in the recent past. Many of today's rain forest areas were savannahs during glacial periods and the present savannahs were even drier than they are today. Figure 1 presents a schematic model of the development of summit levels by differential etching and stripping. The balance between etching and stripping was almost certainly different from the present situation for long periods.

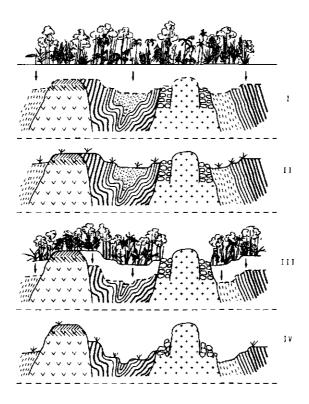


Fig. 1. The development of summit levels by differential etching and stripping. Source: Kroonenberg & Melitz, 1983.

Many shield areas consist of vast dissected plains with solitary elevated areas that either are bare, dome-shaped granite hills ('inselbergs') such as the sugar loaf of Rio de Janeiro, or heaps of huge granite boulders known as 'tors' (see Figure 2). The plains consist of a deep, flat or undulating, residual weathering crust, dissected by a network of small V-shaped valleys that are only a few metres deep. Where the drainage pattern is widely spaced, remnants of the original flat surface may still be detectable but in areas with densely spaced drains there remain only rounded, convex hills. (The drainage pattern is usually determined by the structure of the underlying bedrock; low ridges and depressions result from differential etching and stripping of resistant and less resistant rocks.)



Fig. 2. Granite tors in Botswana extends over the surroundings after a long history of weathering. Photograph by courtesy of Paulien Brombacher.

The pattern of isolated inselbergs and tors amidst vast expanses of undulating lowland is more clear in savannah regions than in areas with rain forest. A difference is that the valleys (called 'dambo' or 'vlei' in Africa) are broad and shallow in savannah regions as a result of colluviation and slope wash. Surface runoff is higher there than in rain forest areas where infiltration rates are higher and (linear) erosion along channels predominates. Flat surfaces in savannah areas are essentially wash pediments that were shaped by sheet floods and surficial wash whereas any (fossil) wash pediments in rain forest areas are merely dissected by channeled vertical erosion.

The origin of inselbergs and tors is connected with climate fluctuations in the past. In wet eras, some rock masses were less deeply etched than others because of differences in chemical composition and/or joint spacing. In a subsequent drier period, these more resistant parts carried a less dense vegetation cover and were subject to less intense chemical weathering. Eventually, they came to stand out above their surroundings.

Climate fluctuations have also played a role in the formation of the characteristic 'laterite plateaus'. Highly weathered saprolite with quartzrich clays ('plinthite') formed during a humid period and became truncated and hardened to 'ironstone' during a subsequent drier era. The plateaus either constitute weathering residues under ironstone, or formed through relief inversion of iron-cemented valley fills.

LANDFORMS IN ALPINE FOLD BELTS (LOWER THAN 3000 METRES)

High mountain areas in the Tropical Zone were partly glaciated in the Pleistocene but tracts below 3,000 metres were never reached by descending glaciers.

Rain forest is the preponderant vegetation type below this altitude in humid tropical mountain. The relation between rainfall and land surface transformation is similar to that in the shield areas. Infiltration reaches deep levels in the rain forest zone and weathering is advanced, despite steep slopes. Fresh rock is difficult to find, even in deeply dissected terrain, and tors and inselbergs are absent.

The dominant geomorphic controls in humid tropical mountain areas are:

- (1) the presence of deep weathering mantles;
- (2) strong tectonic uplifting;
- (3) rapid downcutting by rivers, and
- (4) undermining of slopes and subsequent mass movements.

Slope forms are mainly determined by landslides which continue further downwards as mudflows. The landslides are often triggered by earthquakes and are especially common in lands with primary forest because the trees add to the weight of the unstable soil masses. In the forested mountain areas of New Guinea, Hawaii, the Andes, and elsewhere, shallow landslides are very common and a provisional chronology can often be established by considering the degree of forest regeneration. Counter to common belief, the forest vegetation cannot prevent landslides from happening because the sliding land mass detaches itself at the 'weathering front', i.e. the contact plane between saprolite and fresh rock which is beyond the reach of the roots.

Regions with crystalline rocks have often symmetrical hills with sharp crests and rectilinear slopes, separated by steep V-shaped valleys. The topography resembles that of other eroded badlands in homogeneous materials, only much enlarged (hence the name 'megabadlands'). Interfluve and slope morphologies are dominated by landslide scars.

<u>Siliceous sedimentary rocks</u> have less deep weathering mantles than crystalline rocks. The alternation of resistant and less resistant strata in fold patterns is the main controlling factor. Often, landslides occur along beds of impermeable shales, especially in times of heavy rainfall. The alternation of strata accounts also for the occurrence of mesas, cuestas, and especially hogback morphology in an Appalachian style.

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<u>Calcareous rocks</u> are associated with typical tropical karst phenomena, starting with the formation of sink holes and caves, but ultimately leading to far more advanced dissolution phenomena than found outside the tropics. 'Tower karst' is an example, just as 'cockpit' or 'mogote' hills, or razor-sharp ridges of limestone between deep holes and ravines, as in the forest-clad 'broken bottle country' of New Guinea.

<u>Volcanic rocks</u> and the characteristic landforms of regions with volcanic materials have been discussed in a previous chapter of this text.

In summary, the wet tropics and subtropics are marked by large, deeply weathered areas on old, geologically stable shields and by (smaller) occurrences of younger materials, e.g. residual material which became exposed by truncation of the soil, transported and mixed colluvial and alluvial deposits and materials which became rejuvenated by (volcanic) foreign admixtures. The characteristic soils of the wet tropics are red or yellow in colour, old and strongly leached. They are deep, finely textured, contain no more than traces of weatherable minerals, have low-activity clays, less than 5 percent recognizable rock structure and gradual soil boundaries. Typical Major Soil Groupings are the plinthite-containing PLINTHOSOLS, the deeply weathered and chemically poor FERRALSOLS, the richer NITISOLS, strongly leached ACRISOLS with a clay illuviation horizon, ALISOLS with low base saturation and high activity clays, and LIXISOLS with high base saturation and low CEC. The differences among the soils in the wet tropics can largely be attributed to differences in lithology and (past) moisture regime.

PLINTHOSOLS

Soils having 25 percent or more plinthite^a 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 gleyic^a or stagnic^a properties within 100 cm of the surface.

Key to Plinthosol (PT) Soil Units

Plinthosols having an albic* E-horizon.

Albic Plinthosols (PTa)

Other Plinthosols having an umbric* A-horizon or a dystric* histic* H-horizon and which are strongly humic*.

Humic Plinthosols (PTu)

Other Plinthosols having an ochric* A-horizon and a base saturation (by 1M $NH_{\lambda}OAc$ at pH 7.0) of less than 50 percent throughout the upper 50 cm of the plinthite* horizon.

Dystric Plinthosols (PTd)

Other Plinthosols.

Eutric Plinthosols (PTe)

SUMMARY DESCRIPTION OF PLINTHOSOLS

<u>Connotation</u>: soils with mottled clayey materials which become as hard as a brick when exposed to the open air; from Gr. <u>plinthos</u>, brick.

<u>Parent material</u>: plinthite is more common in weathering material from basic rock than from acid rock. In any case it is essential that sufficient iron is present, originating either from the parent material itself or brought in by seepage water from an adjacent upland area.

Environment: plinthite is formed in level to gently sloping areas with a fluctuating water table. A petroferric phase with continuous 'ironstone' develops where plinthite becomes exposed to the surface, e.g. on old erosion surfaces that are above the present drainage base. Skeletic soils with a concretionary layer of hardened plinthite that is not continuously cemented, occur mostly in colluvial and alluvial materials. Plinthite is associated with rain forest areas; petroferric and skeletic soils are more common in the savannah zone.

<u>Profile development</u>: mostly ABC or AEBC profiles. Accumulation of sesquioxides by ferralitization and enrichment from outside sources; segregation of iron mottles in a zone with a fluctuating water table.

^{*} Diagnostic horizon; see Annex 1 for full definition.

Diagnostic property; see Annex 2 for full definition.

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<u>Use</u>: low agricultural value because of frequent waterlogging, dense plinthite layer and low chemical fertility. The iron-rich material is used for building purposes and road construction.

REGIONAL DISTRIBUTION OF PLINTHOSOLS

Areas where the formation of 'plinthite' (an iron-rich, humus-poor mixture of clay and quartz) is reportedly very active include western India (where Buchanan first used the term 'laterite' for this material in 1807), West Africa and parts of South America. All these areas have a hot and humid climate with a high annual rainfall sum (in places more than 2000 mm) and a short dry season.

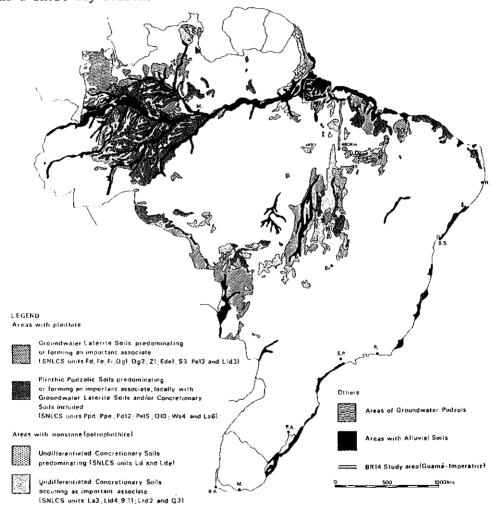


Fig. 1. Occurrence of plinthite and ironstone soils in Brasil. Source: Sombroek & Camargo, 1983.

Areas with 'ironstone' (which develops upon repeated wetting and drying of plinthite) are far more extensive than those with plinthite or Plinthosols. The reason is that, once formed, ironstone is highly resistant to weathering. Ironstone materials are often relics from periods with a different climate than at present. Figure 1 shows the occurrence of plinthite and ironstone in Brasil; both materials are particularly widespread in the Amazon region.

GENESIS OF PLINTHOSOLS

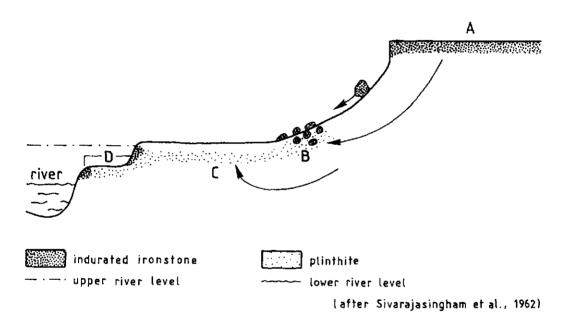


Fig. 2. Four physiographically distinct landscape positions where plinthite and ironstone occur.

- A: indurated ironstone (massive iron pan or gravel) capping an old erosion surface.
- B: plinthite and ironstone (gravel and boulders) in a colluvial footslope (subject to iron-rich water seepage).
- C: plinthite in soils of a low level plain (river terrace) with periods of high groundwater.
- D: along the banks of rivers where plinthite becomes exposed and hardens to ironstone.

The formation of plinthite

Plinthite is an iron-rich, humus-poor mixture of clay and quartz. Its formation involves the following processes:

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- (1) accumulation of sesquioxides, through:
 - (a) Relative accumulation as a consequence of the removal of silica and bases by ferralitization (see under Ferralsols), and/or
 - (b) Absolute accumulation through enrichment with sesquioxides from outside (vertical or lateral, see Figure 2).
- (2) <u>segregation of iron mottles</u>, caused by alternating reduction and oxidation. In times of water saturation, much of the iron is in the ferrous form, has a high mobility and is easily redistributed. When the water table falls, this iron precipitates as ferric oxide that will not, or only partially, redissolve in the next wet season.

The hardening of plinthite to ironstone

Plinthite forms in low and wet areas; if the land becomes drier, e.g. because of a change in base level or a change in climate, the plinthite hardens to ironstone. This hardening involves the following processes:

- crystallization of amorphous iron compounds into continuous aggregates or networks of iron oxide minerals, especially goethite.
- (2) <u>dehydration</u> of goethite (FeOOH) to hematite (Fe₂O₃) and of gibbsite $(Al_2O_3.3H_2O)$, if present, to boehmite $(Al_2O_3.H_2O)$.

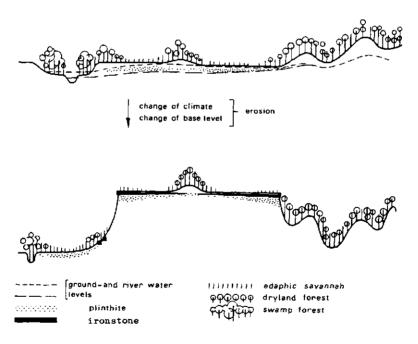


Fig. 3. Inversion of relief in an eroding landscape, Hardening of exposed plinthite produces a protective (petroferric) shield against further erosion. Source: Buringh & Sombroek, 1971.

Hardening of plinthite is often initiated by removal of the vegetation, especially forest, as this induces erosion and exposure of plinthite to the air. Ironstone materials occur in many tropical soils, either in a 'skeletic' (concretionary) or a 'petroferric' (continuous pan) form. They occur in the rain forest zone and in the savannahs but are especially abundant in the transition zone: the wet climate of the rain forest promotes the formation of plinthite, whereas the drier savannah environment is inducive to its hardening. In places, ironstone caps form a shield against erosion. The result is an inversion of the original relief; parts that initially were the lowest of the landscape become the highest (see Figure 3).

The main diagnostic feature of Plinthosols is the presence of at least 25 percent (by volume) plinthite. Not every red mottled clay horizon contains plinthite.

It is not always easy to distinguish between red rust mottles, plinthite and ironstone gravel because they grade into each other. Important field criteria for identification of red mottles that qualify as plinthite are:

- the red mottles are firm to very firm when moist and hard to very hard when dry;
- (2) they can be cut with a knife but only with considerable difficulty;
- (3) they have sharp boundaries;
- (4) they hardly stain the fingers when rubbed;
- (5) they do not slake in water.

The main criterion is, of course, the irreversible hardening to ironstone upon repeated wetting and drying, but this cannot always be established in the field.

Ironstone (also referred to as 'laterite', 'murram', 'ferricrete', or 'petroplinthite') can be divided, on basis of its morphology, in the following types:

- (1) Massive iron pans (the petroferric phase; see Annex 3), either:
 - (a) residual: formed by hardening of plinthite where the red mottles are interconnected; the lighter coloured parts that do not harden form irregular cavities ('vesicular' laterite); in some instances the original rock structure may still be visible ('ferruginized rock'), or
 - (b) recemented: colluvial ironstone gravel, stones and boulders that are cemented together.
- (2) <u>Discontinuous ironstone</u> (the skeletic phase; see Annex 3), either:
 - (a) residual gravels formed by hardening of plinthite where the red mottles make up 40 percent of the soil volume or more, and are not connected. The gravels are round concretions ('pisolithes' or 'pea iron') or irregular nodules, or
 - (b) colluvial gravel, stones and boulders formed by disintegration of a massive iron pan.

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The petroferric layer contains little or no organic matter and occurs at shallow depth. Figure 4 shows an ironstone 'duricrust' (Fr. 'cuirasse') developed on a decapitated (former) Plinthosol. Skeletic layers may consist for a considerable part of ironstone (up to 80 percent).

Mineralogical characteristics

Plinthite and ironstone have a high content of sesquioxides. Free iron is present as the oxide minerals lepidocrocite (FeOOH), goethite (FeOOH) or hematite (Fe $_2$ O $_3$); free aluminium occurs as gibbsite (Al $_2$ O $_3$.3H $_2$ O) and/or boehmite (Al $_2$ O $_3$.H $_2$ O). Old ironstone crusts contain more hematite and boehmite and less hydrated Fe- and Al-oxides than plinthite. Free silica is present as quartz, inherited from the parent material. Easily weatherable primary minerals have disappeared. The dominant clay mineral is kaolinite.



Fig. 4. Petroferric ironstone 'duricrust' in Ivory Coast. Photograph by courtesy of ISRIC, Wageningen.

Hydrological characteristics

Plinthosols are indigenous to regions with a distinct annual precipitation surplus over evaporation. Percolating rain may cause eluviation symptoms such as an albic E-horizon, particularly where the surface horizon is strongly humic. Plinthosols with gleyic or stagnic soil properties are common phenomena in bottom lands.

Physical characteristics

The plinthite layer is dense and obstructs the flow of water as well as deep penetration of roots. The capacity to harden is a potential danger of all Plinthosols. The specific density of ironstone ranges from 2.5 to 3.6 Mg/m^3 and increases with increasing iron content.

Chemical characteristics

All Plinthosols have a high content of iron and/or aluminium, with proportions varying from more than 80 percent iron oxides with little aluminium, to about 40 percent of each. Most Plinthosols have a low CEC and a low base saturation but there are exceptions.

MANAGEMENT AND USE OF PLINTHOSOLS

Plinthosols come with serious management problems. Waterlogging and low natural fertility are their main limitations. If the plinthite layer hardens, e.g. because of deep drainage or erosion, ironstone will limit the possibilities for root growth and lower the water storage capacity of the soil. Most petroferric soils are unsuitable for arable farming; they are used for extensive grazing and firewood production. Skeletic soils with high but variable contents of pea iron are planted to food crops and tree crops (e.g. cocoa in West Africa) but the crops suffer from drought in the dry season.

Civil engineers have a different appreciation of ironstone and plinthite than agronomists have. To them, plinthite is a valuable material to make bricks (massive ironstone can be cut into building blocks too). Ironstone gravel can be used for foundations and as a surfacing material for roads and air fields. In some instances it is a valuable ore of iron, aluminium, manganese or titanium.

NOTES

FERRALSOLS

Soils having a ferralic* B-horizon.

Key to Ferralsol (FR) Soil Units

Ferralsols having plinthite within 125 cm of the surface.

Plinthic Ferralsols (FRp)

Other Ferralsols having geric⁰ properties in at least some part of the ferralic¹ B-horizon within 125 cm of the surface.

Geric Ferralsols (FRg)

Other Ferralsols which are strongly humic^a, having an umbric^{*} A-horizon, or a mollic^{*} A-horizon and a base saturation (by 1M $\rm NH_4OAc$ at pH 7.0) of less than 50 percent in at least a part of the B-horizon within 100 cm of the surface.

Humic Ferralsols (FRu)

Other Ferralsols having a red to dusky red B-horizon (rubbed soil has hues redder than 5YR with a moist value of less than 4 and a dry value not more than one unit higher than the moist value).

Rhodic Ferralsols (FRr)

Other Ferralsols having a yellow to pale yellow B-horizon (rubbed soil has hues of 7.5YR or yellower with a moist value of 4 or more and a moist chroma of 5 or more).

Xanthic Ferralsols (FRx)

Other Ferralsols.

Haplic Ferralsols (FRh)

SUMMARY DESCRIPTION OF FERRALSOLS

<u>Connotation</u>: red and yellow tropical soils with a high content of sesquioxides; from L. <u>ferrum</u>, iron and <u>alumen</u>, alum.

<u>Parent material</u>: pre-weathered, mostly transported materials of old age, derived from a wide variety of rocks; more often in weathering material from basic rock than in siliceous material.

<u>Environment</u>: level to undulating stable land surfaces of Pleistocene age or older; less commonly in younger land on easily weatherable rocks. Perhumid or humid tropical climates; small occurrences elsewhere are relics from past eras with a wetter climate than today. Tropical rain forest, semi-deciduous forest and savannah.

Diagnostic horizon; see Annex 1 for full definition.

a Diagnostic property; see Annex 2 for full definition.

<u>Profile development</u>: ABC profiles. Deep and intensive weathering has resulted in a residual concentration of resistant primary minerals, and the formation of kaolinitic clays and iron and aluminium oxides and hydroxides. This mineralogy and the low pH account for the stable microstructure (pseudo-sand) and yellowish (goethite) or reddish (hematite) colours of Ferralsols.

<u>Use</u>: Ferralsols have good physical properties but are chemically poor. Their low natural fertility, virtual absence of weatherable minerals, and very low cation retention capacity are serious limitations. Many Ferralsols are (still) used for shifting cultivation. Liming and full fertilization are required for sustainable sedentary agriculture.

REGIONAL DISTRIBUTION OF FERRALSOLS

There are close to one billion hectares of Ferralsols worldwide. Nearly all Ferralsols are situated in the tropics where they occupy some 20 percent of the land surface with major occurrences on the continental shields of South America and central Africa. These regions have not been affected by intensive folding or glacial action during recent geological periods. Outside the continental shields, Ferralsols are restricted to regions with easily weatherable basic rock and a hot and humid climate, e.g. in southeast Asia and on some Pacific islands. The main Ferralsol areas are shown in Figure 1.

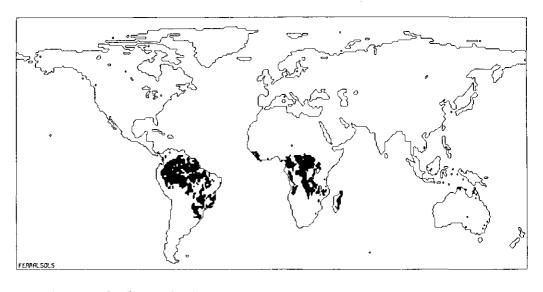


Fig. 1. Ferralsols worldwide.

GENESIS OF FERRALSOLS

Water affects primary minerals though the processes of 'hydration' and 'hydrolysis'. Hydration is water absorption by solid particles. In hydrolysis, H*-ions penetrate the primary silicate structures of minerals such as feldspars which then release cations (K, Na, Ca, Mg). As the hydrogen ion is much smaller than any of the cations it replaces, the structure of the minerals is weakened. This facilitates dissolution of Si and Al from the clay lattices.

'Ferralitization' is hydrolysis in an advanced stage. A combination of slow release and leaching of Si, K, Na, Mg and Ca keeps the concentrations of these ions in the soil solution low. If the soil temperature is high and percolation intense, all weatherable primary minerals will ultimately be removed from the soil mass. Less soluble compounds such as iron and aluminium oxides and hydroxides and coarse quartz grains, remain behind. The process of ferralitization (or 'desilication' as it is also called) is accelerated by the following conditions:

- (1) A relatively low pH and low concentrations of dissolved weathering products in the soil solution promote desilication and enrichment of the soil with (residual) Fe and Al. High production of CO₂ (from respiration by roots and soil organisms feeding on organic matter) and high percolation rates depress the pH and lower the concentrations of weathering products.
- (2) Geomorphic stability over prolonged periods of time is a precondition. Ferralitization is a very slow process, even in the tropics where high temperatures increase reaction rates and solubility limits. In temperate regions, old erosion surfaces are rare because of the influence of recent glacial processes on landscape formation.
- (3) Ferralitization proceeds most strongly in basic parent materials which contain relatively much iron and aluminium in easily weatherable minerals, and little silica. On weathering, iron remains behind in crystalline compounds, mainly goethite (Fe00H) and hematite (Fe203), and aluminium in gibbsite (AL(OH)3). In soils from acid rock, ferralitization proceeds much slower. There are less easily weatherable minerals and more quartz. Although most silica disappears through leaching (hence 'desilication'), the silica content of the soil solution remains higher than in soils from basic rock. This silica combines with aluminium to the 1:1 clay mineral kaolinite (kaolinitization). Gibbsite is normally absent. Impeded internal drainage may enhance the process because silica in solution is not removed quickly enough (see Table 1).

TABLE 1. The occurrence of gibbsite and kaolinite in strongly weathered soils under various drainage conditions.

PARENT MATERIAL	INTERNAL DRAINAGE very good good moderate poor				
Mafic rock	gibbsite	gibbsite	kaolinite	2:1 clays	
Felsic rock	gibbsite	kaolinite	kaolinite	kaolinite	

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Upon weathering, ferrihydrite $(Fe(OH)_3)$; see also the chapter on Andosols) forms if the iron concentration is high. Hematite, the mineral which gives many tropical soils their bright red colour, forms out of ferrihydrite if:

- (1) the iron concentration is high, and
- (2) the organic matter content is low (Fe-humus complexes inactivate Fe), and
- (3) the temperature is high (accelerates dehydration of ferrihydrite and decomposition of organic matter), and
- (4) the soil-pH is higher than 4.0 (else $Fe(OH)_2^+$ -monomers are formed). Goethite, more orange in colour than the bright red hematite, is formed when one or more of the above conditions are not (fully) met.

CHARACTERISTICS OF FERRALSOLS

The essential diagnostic feature of Ferralsols is the presence of a ferralic B-horizon. A typical horizon sequence in a Ferralsol would be:

0- 30 cm A ochric A-horizon

30- 60 cm AB transitional horizon, A-characteristics dominant

60- 90 cm BA transitional horizon, B-characteristics dominant

90-210 cm B ferralic B-horizon

210-240 cm BC transitional horizon to C-horizon

Characteristic morphological features are:

- (1) a deep solum, usually several meters thick, over weathering rock;
- (2) diffuse or gradual horizon boundaries, unless a stoneline occurs.
- (3) a high iron content in the ferralic B-horizon, together with the good internal drainage responsible for distinct red (hematite) or yellow (goethite) matrix colours, usually without mottles;
- (4) a well developed microstructure; kaolinite with a negative surface charge combines with sesquioxides with a positive surface charge to form strong micro-aggregates of silt (pseudo-silt) or sand (pseudosand) size. Soils with a clay content of 60 percent or more 'feel loamy' in the field and have the same mechanical properties as medium or even light textured soils;
- (5) a weak 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.

Mineralogical characteristics

Ferralsols are strongly weathered and therefore characterized by primary and secondary minerals of great stability. Easily weatherable primary minerals such as glasses and ferro-magnesian minerals and even the more resistant feldspars and micas have disappeared completely. Quartz is the main primary mineral (if originally present in the parent rock). The clay fraction consists mainly of kaolinite, goethite, hematite and gibbsite in varying amounts and reflects the kind of parent rock and the drainage conditions (see Table 1).

Hydrological characteristics

A consequence of the advanced weathering of Ferralsols is that most of these soils have a high clay content and strong water retention at permanent wilting point. The formation of micro-aggregates reduces their moisture storage at field capacity which explains the rather limited 'available' water holding capacity: some 10 mm of available water per 10 cm soil depth is a rule of thumb. Most Ferralsols are sensitive to drought, particularly those in elevated positions.

Physical characteristics

Stable micro-aggregates and many biopores account for the excellent porosity, good permeability and high infiltration rates of Ferralsols. Clayey soils with a high content of (positively charged) iron oxides and (negatively charged) kaolinite, have a particularly stable soil structure due to bonding of opposite elements. Ferralsols with a low iron content, low pH, high Al-saturation and low organic matter content, as occur in Surinam and Brasil (Xanthic Ferralsols), have less stable structures, especially the sandy ones. As long as such soils remain under forest, there is no problem, but surface sealing and compaction become serious problems once such soils are taken into cultivation.

The strong cohesion of the (micro-)aggregates and the rapid reflocculation of suspended particles complicate the determination of the particle size distribution of Ferralsol material. The percentage clay which is found after removal of the iron and addition of a peptising agent is called the 'total clay' content. The amount of clay that is determined by shaking an aliquot of soil with distilled water (without removing the iron and adding dispersion agents) is called the 'natural clay'. In ferralic B-horizons, the natural clay content is very low, which tallies with the high stability of the aggregates and the absence of clay movement. The ratio of natural clay over total clay may be used as an index of stability; it is one of the criteria by which Ferralsols are distinguished from the (less stable) Acrisols.

Chemical characteristics

Ferralsols are chemically poor soils with a low ion exchange capacity. The exchange of cations and anions between the solid and liquid phases in soils is conditioned by the type and quantity of the clay, the oxides and the organic matter. The exchange capacity is composed of a permanent and a variable component:

The 'permanent charge' is caused by isomorphic substitution of e.g. Si^{4+} by Al^{3+} , or Al^{3+} by Mg^{2+} , in the crystal lattice of the clay minerals. The resulting negative charge is independent of soil pH or ion concentrations of the soil solution. Kaolinite is the main clay mineral in Ferralsols and has only a very small permanent charge.

The 'variable charge' is due to:

 dissociation of H⁺-ions from molecules located at the perifery of the exchange complex (creating negative sites), or

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(2) <u>protonation</u> which results in a positive charge. Protons (H⁺) may, for example, be released by acid groups at the edges of clay particles, or by carboxylic or phenolic groups in the organic matter, or by aluminium and iron hydroxides.

Dissociation of H^+ -ions is strongest at high concentrations of OH^- in the soil solution and is therefore called the 'pH dependent' part of the cation exchange capacity (CEC).

The cation exchange capacity of a ferralic B-horizon may, by definition, not be in excess of 16 cmol(+)/kg of clay; the CEC is determined in a 1M NH₄OAc solution, buffered to pH 7. A Ferralsol with 50 percent clay in its B-horizon contains less than 8 cmol(+) cations per kg at pH 7.0. However, the field-pH of Ferralsols is normally around 5. The net negative charge of the exchange complex is neutralized by exchangeable bases (Na † , K † , Ca 2† , Mg 2†) plus exchangeable acidity (Al 3† + H †). Therefore, the 'effective CEC, or ECEC, is the sum of the bases plus the exchangeable acidity. The ECEC represents the cation exchange capacity at field conditions and thus of great practical importance.

NOTE THAT the ECEC of Ferralsols is much lower than the CEC and the actual cation adsorption is often not higher than a mere 3 or 4 cmol(+) per kg soil.

Protonation of hydroxylic groups at low pH-values may boost the anion exchange capacity (AEC) of the soil to the extent that the AEC equals or exceeds the CEC. This can be detected by comparing the pH of two samples of the same soil, one in suspension in $\rm H_2O$ and the other in 1M KCl. Normally (with a net negative charge), the pH(KGl) is lower than pH($\rm H_2O$), but with a net positive charge the reverse is true.

In publications on the exchange properties of strongly weathered tropical soils, the following terminology is used:

- The pH value at which the AEC fully compensates the CEC (permanent plus variable charges) is called the 'point of zero net charge' (PZNC).
- The difference between pH(KCl) and pH(H2O) is called 'delta pH'.

Figure 2 presents in a schematic way the exchange characteristics of strongly weathered tropical soils at different pH levels.

Biological characteristics

The gradual horizon boundaries in Ferralsol profiles must, according to some, at least partly be attributed to termites. These animals 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 explains the frequent occurrence of a stoneline in or below the ferralic B-horizon; the depth of the stoneline reflects the depth of termite activity.

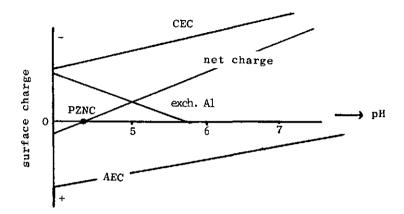


Fig. 2. Schematic relation between exchangeable aluminium level, AEC, CEC, net surface charge and the $pH(H_2O)$ of the soil.

MANAGEMENT AND USE OF FERRALSOLS

Ferralsols have good physical properties and poor chemical properties. Their great depth, high permeability and stable microstructure make them less susceptible to erosion than many other soils, with exception of the shallow and the sandy types. All Ferralsols have a friable consistence under most conditions and are easy to work. They are well drained but may in times be droughty because of their low 'available' water storage capacity.

The chemical fertility of Ferralsols is poor; weatherable minerals are absent and cation retention by the mineral soil fraction is low. In Ferralsols under a natural vegetation, the bulk of all 'available' plant nutrients (and of all living plant roots) is concentrated in the upper 10 to 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 subsistance farming, the root zone will rapidly become depleted of plant nutrients. Maintenance of the organic matter content 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 inactivation of phosporus. Nitrogen and potassium are often deficient too, just as the secondary nutrients calcium, magnesium and sulphur and a score of micronutrients. 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

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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 phosphorus fixation complicate liming and fertilizer application.

<u>Liming</u> is a means to raise the pH of the rooted surface soil. This combats aluminium toxicity and raises the GEC. On the other hand, liming lowers the AEC which might lead to collapse of structure elements and slaking at the soil surface. Frequent application of small doses of lime or basic slag is therefore preferable over one massive dose.

Fertilizer selection and the mode/timing of fertilizer application determine to a great extent the success of fertilizer use on Ferralsols. 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.



Fig. 3. Xanthic Ferralsols under equatorial forest in the Amazon basin. Photograph by courtesy of the Conselho Nacional de Geografia, Brasil.

Ferralsols are grown to a variety of annual and perennial crops, by sedentary subsistence farmers and shifting cultivators. Grazing is also widely practiced 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 the poor chemical soil properties could be overcome. Between Ferralsols, the Rhodic Ferralsols (basic parent material, high clay content) and the Humic Ferralsols make the best arable land. The Plinthic (plinthite), Geric (very low CEC) and Xanthic Ferralsols (especially the sandy ones; see Figure 3) have severe limitations. Haplic Ferralsols take up an intermediate position.

NOTES

NITISOLS

Soils having an argic* B-horizon showing 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 A- and B-horizons; having nitic* properties in some subhorizon within 125 cm of the surface; lacking the tonguing* which is diagnostic for Podzoluvisols; lacking ferric* and/or vertic* properties; lacking plinthite* within 125 cm of the surface.

Key to Nitisol (NT) Soil Units

Nitisols which are strongly humic^a, having an umbric^{*} A-horizon, or a mollic^{*} A-horizon and a base saturation (by 1M $\rm NH_4OAc$ at pH 7.0) of less than 50 percent in at least a part of the B-horizon within 125 cm of the surface.

Humic Nitisols (NTu)

Other Nitisols having a red to dusky red argic* B-horizon (rubbed soil having hues redder than 5YR with a moist value of less than 4 and a dry value not more than one unit higher than the moist value).

Rhodic Nitisols (NTr)

Other Nitisols.

Haplic Nitisols (NTh)

* Diagnostic horizon; see Annex 1 for full definition.

Diagnostic property; see Annex 2 for full definition.

SUMMARY DESCRIPTION OF NITISOLS

<u>Connotation</u>: deep, red tropical soils with clay illuviation and shiny peds; from L. <u>nitidus</u>, shiny.

<u>Parent material</u>: finely textured weathering products from intermediate to basic parent rock; in places rejuvenated by admixtures of volcanic ash.

<u>Environment</u>: level to hilly land under tropical rain forest or a savannah vegetation.

<u>Profile development</u>: ABtC profiles. Reddish brown clayey soils with a deeply developed clay illuviation horizon of high structural stability.

Use: moderately to highly productive under a wide range of crops.

REGIONAL DISTRIBUTION OF NITISOLS

There are approximately 250 million hectares of Nitisols worldwide. More than half of all Nitisols occur in Africa, especially in Kenya, Zaire and Cameroon but they are also represented in tropical Asia (22 percent), South America (12 percent), Central America (5 percent) and Australia (5 percent). See Figure 1.

Nitisols are most extensive in regions with clearly defined wet and dry seasons but they occur also under a (per)udic moisture regime. Their occurrence in dry regions e.g. in parts of Australia, is considered a paleo-feature. Nitisols are most common on slightly undulating to dissected surfaces of early to middle Pleistocene age. In East Africa, they occur mainly at altitudes of 1000 m or more but they are well represented at lower altitudes in South America and southeast Asia.

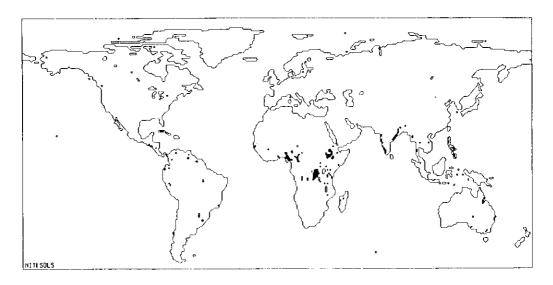


Fig. 1. Nitisols worldwide.

GENESIS OF NITISOLS

Nitisols are formed in intermediate to basic parent materials under a vegetation type ranging from wooded grassland to (montane) rain forest. The formation of Nitisols involves the following processes:

(1) ferralitization: intensive hydrolysis of weatherable minerals combined with leaching of silica and bases, and relative accumulation of kaolinite and sesquioxides. The process is the same as in the formation of Ferralsols but it is still in an early stage.

- (2) translocation of clay from the surface to the subsoil. Clay cutans are present on ped faces and in voids in the Bt-horizon. Because of the well drained and permeable nature of the profile, clay can be transported down to considerable depth. This process is responsible for the stretched bulge in clay content that is typical of Nitisols.
- (3) nitidization: formation of strong angular blocks that fall apart into smaller angular blocks with shiny ped faces ('nitic properties'). The exact process is still unknown. It is probably a combination of clay translocation and micro-swelling and shrinking which caused the formation of pressure faces.
- (4) homogenization: (biological pedoturbation) by termites, ants, worms and other soil fauna. This process is particularly prominent in the top 100 cm soil layer where it causes destruction of clay cutans and formation of crumb and subangular blocky structures and of gradual or diffuse soil horizon boundaries.

CHARACTERISTICS OF NITISOLS

Nitisols are well drained soils with nitic properties and a deeply developed argic B-horizon. A typical horizon sequence is shown below:

0-30 cm Ap ochric, mollic or umbric A-horizon

30-90 cm AB transitional horizon between A- and B-horizon

90-150 cm Bt argic B-horizon with nitic properties

150-210 cm BC transitional horizon to C-horizon

The solum extends normally down to a depth of $150\,\mathrm{cm}$ or more and includes an argic B-horizon of more than $100\,\mathrm{cm}$ thick. The boundaries between the A- and the B-horizon, and between the B- and the C-horizon are gradual to diffuse, and there are normally transitional horizons.

Most Nitisols are red soils, with Munsell notations in the 2.5YR and 5YR hues; yellower (10YR, 7.5YR) or redder (10R) soils are less common.

The soil structure is moderate to strong. Crumb and subangular blocky structures are present in the surface soil; there are mainly subangular blocky elements between 50 and 100 cm, and very fine to medium angular blocky elements deeper down. The angular blocks break down upon pressure into ever smaller units that have shiny surfaces, especially under moist conditions. The aggregate stability is very high. During dry periods the soil material shrinks and cracks develop in the surface soil.

Most Nitisols contain more than 35 percent clay throughout, with a gradual increase of the clay content from the topsoil down to the B-horizon and a very gradual decrease, if any, of the clay percentage in the subsoil.

Mineralogical characteristics

Kaolinite is the commonest clay mineral in Nitisols, followed by (meta)-halloysite. Minor quantities of illite, chloritized vermiculite and randomly interstratified clay minerals may also be present, just as hematite, goethite and gibbsite. An amorphous iron content (Fe $_2$ 0 $_3$ by acid oxalate extraction) that is at least 5 percent of the free iron content (Fe $_2$ 0 $_3$ by dithionite-citrate extraction) is typical.

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The mineralogical composition of the sand fraction depends strongly on the nature of the parent material. Although weathering-resistant minerals (notably quartz) predominate, minor quantities of more easily weatherable minerals (feldspars, volcanic glass, apatite, amphiboles) may be present and show that Nitisols are less strongly weathered than Ferralsols.

Physical characteristics

Nitisols are very porous and have a high moisture storage capacity. Their permeability is rapid to moderate, even in the deeper subsoil. A Nitisol can normally be tilled within 24 hours after it was moistened, without a detrimental effect on the soil structure.

Chemical characteristics

The cation exchange capacity of Nitisols is high if compared to that of other tropical soils such as Ferralsols, Lixisols and Acrisols. The reasons are:

- (1) Although the CEC of the clay (by 1M NH_4OAc at pH 7.0) is often less than 16 cmol(+)/kg, the clay content is high (more than 35 percent and not seldom more than 60 percent); and
- (2) the contribution of the organic matter to the CEC is considerable, especially if a mollic or umbric A-horizon is present.
 The base saturation varies from less than 10 to more than 90 percent. The

The base saturation varies from less than 10 to more than 90 percent. The pH usually lies between 5.0 and 6.5; P-fixation may be a problem.

Biological characteristics

Faunal activity is to some extent accountable for the typical gradual horizon boundaries. Termites homogenize the soil; volcanic glass deposited on the (present) surface was found at a depth of 7 m in Nitisols in Kenya (Wielemaker, 1984).

MANAGEMENT AND USE OF NITISOLS

Nitisols are among the most productive soils of the humid tropics. Their high porosity and deep solum allow deep rooting. This, and their stable soil structure, makes Nitisols less susceptible to erosion than many other soils. Their internal drainage, water holding capacity and workability are good. Their chemical fertility compares favourably to other tropical soils because of their moderate CEC, the relatively high organic matter content and the presence of (some) weatherable minerals.

Nitisols are intensively used for plantation crops such as cocoa, coffee and pineapple, and for food crop production. They respond well to fertilizer (N,P) applications. Erosion control measures help to conserve the organic matter content of the A-horizon which contributes much to the productive capacity of Nitisols.

ACRISOLS

Soils having an argic* B-horizon which has a cation exchange capacity of less than 24 cmol(+)/kg clay and a base saturation (by 1M NH₄OAc at pH 7.0) of less than 50 percent in at least some part of the B-horizon within 125 cm of the surface; lacking the E-horizon abruptly overlying a slowly permeable horizon, the distribution pattern of the clay and the tonguing® which are diagnostic for Planosols, Nitisols and Podzoluvisols respectively.

Key to Acrisol (AC) Soil Units

Acrisols having plinthite within 125 cm of the surface.

Plinthic Acrisols (ACp)

Other Acrisols showing gleyic^a properties within 100 cm of the surface.

Gleyic Acrisols (ACg)

Other Acrisols which are strongly humic^a, having an umbric^{*} or a mollic^{*} A-horizon.

Humic Acrisols (ACu)

Other Acrisols showing ferrica properties.

Ferric Acrisols (ACf)

Other Acrisols.

Haplic Acrisols (ACh)

SUMMARY DESCRIPTION OF ACRISOLS

Connotation: strongly weathered acid soils with a low base saturation; from L. acer, very acid.

Parent material: mainly silica-rich weathering materials.

<u>Environment</u>: most common in old land surfaces with a hilly or undulating topography, in wet tropical and monsoonal climates. Light tropical (rain) forest is the natural vegetation type.

<u>Profile development</u>: mostly A(E)BtC profiles. Variations among Acrisols are mainly connected with variations in terrain conditions (drainage, seepage). The shallow A-horizon with dark, raw and acid organic matter grades into a yellowish E-horizon. The Bt-horizon has stronger or more reddish colours than the overlying E-horizon.

Diagnostic horizon; see Annex 1 for full definition.

Diagnostic property; see Annex 2 for full definition.

162 Acrisols

<u>Use</u>: prolonged weathering and advanced soil formation have led to a dominance of low activity clays and a general paucity of plant nutrients. Aluminium toxicity, strong phosphorus sorption, slaking/crusting and high sensitivity to erosion pose severe limitations to permanent cropping. Large areas of Acrisols are used for subsistence farming, partly under shifting cultivation. These soils are not very productive but can be grown to undemanding, acidity-tolerant crops such as cashew nut and pineapple.

REGIONAL DISTRIBUTION OF ACRISOLS

Acrisols and Alisols together cover some 800 million hectares, situated almost exclusively in the equatorial tropics. See Figure 1. As Alisols were only recently introduced as a Major Soil Grouping, the exact division of the Acrisol-Alisol area is not yet clear; it is assumed that more than half of the total area consists of Acrisols, with major occurrences on old, deeply weathered land surfaces in southeast Asia, west Africa and the central part of South America.

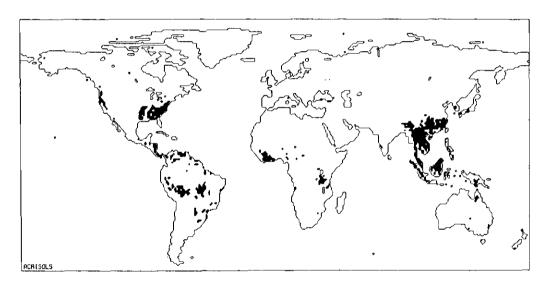


Fig. 1. Areas with Acrisols and Alisols.

GENESIS OF ACRISOLS

Acrisols are characterized by the presence of an argic B-horizon, a dominance of stable low activity clays and a general paucity of bases.

For a description of the processes of clay <u>dispersion</u>, clay <u>transport</u> and clay <u>accumulation</u> in a Bt-horizon, reference is made to the chapter on Luvisols. It must be mentioned here, however, that some authors dismiss all clay illuviation horizons in highly weathered soils in the wet tropics as relics from a distant past.

The process of ferralitization by which sesquioxides accumulate in the profile, is described in some detail in the chapter on Ferralsols. Subsequent redistribution of iron compounds by cheluviation/chilluviation (see under Podzols) has not seldom resulted in a colour differentiation directly below an A(h)-horizon where an eluviation layer with yellowish colours lies on top of a more reddish coloured Bst-horizon (the 'Red-Yellow Podzolics' of southeast Asia). In situations with periodic waterlogging, the effluent percolation water can be black as in Podzols. Acrisols are more strongly weathered than Alisols, have less primary mine-

Acrisols are more strongly weathered than Alisols, have less primary minerals and a strong dominance of well crystallized 1:1 clays.

CHARACTERISTICS OF ACRISOLS

Most Acrisols have a thin, brown, ochric A-horizon; darker colours occur where mineralization of soil organic matter is hindered by (periodic) waterlogging, augmented perhaps by oligotrophy. (A mollic A-horizon, though separately mentioned in the definition of Humic Soil Units, is atypical.) Most Acrisols have bright red and yellow subsurface colours. However, gleyic soil properties and/or plinthite are common in Acrisols in low terrain positions. The structure of the surface soil is weak; the individual elements collapse under the impact of heavy tropical rain showers, particularly where the organic matter content of the A-horizon is low and/or the E-horizon has become exposed by erosion (see Figure 2). The structure of the Bt-horizon is more stable.

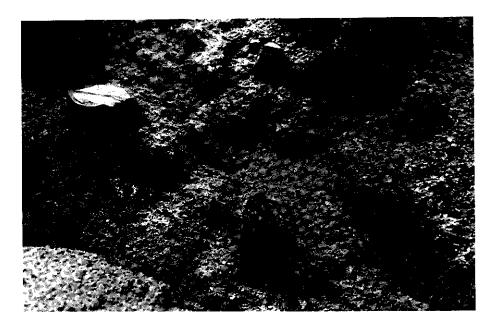


Fig. 2. Severe erosion of (denudated) Acrisols under the impact of rain.

Mineralogical characteristics

Acrisols have little weatherable material left. The contents of Fe-, Aland Ti-oxides are comparable to those of Ferralsols or slightly lower; the $\mathrm{SiO_2/Al_2O_3}$ ratio is 2 or less. The clay fraction consists mainly of well crystallized kaolinite and some gibbsite.

Hydrological characteristics

Under protective natural forest, Acrisols have a porous surface soil which permits adequate infiltration of (rain) water. If the forest is cleared, the surface soil will lose its organic matter and slake, with crusting, surface runoff during rain showers and (in sloping positions) devastating erosion as a result. Many Acrisols show signs of periodic water saturation, particularly those in depressed areas. There, the A-horizons are very dark brown to black whereas matrix colours may be close to white directly below the Bt-horizon.

Physical characteristics

Most Acrisols have a weak microstructure and a massive macrostructure, particularly where the organic matter content of the surface soil is low, because of the uneven distribution of sesquioxides over the soil mass with low amounts residing in the surface horizon(s). The bonding between sesquioxides (AEC) and negatively charged low activity clays (CEC) is less stable than in Ferralsols. The ratio of water dispersible 'natural clay' over 'total clay' (see under Ferralsols) is higher than in Ferralsols.

Chemical characteristics

Acrisols have poor chemical characteristics; their nutritional limitations include widespread aluminium toxicity and strong P-sorption as in Ferralsols. The $pH(H_2O)$ is close to 5.5. As Acrisols are very poor soils with little biological activity and no expanding clay, natural regeneration of the surface soil, e.g. degraded by mechanized agriculture, is very slow.

MANAGEMENT AND USE OF ACRISOLS

As with other highly weathered tropical soils, preservation of the surface soil with its all-important organic matter 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 seedlings planted outside the filled-in spots.

Adapted cropping systems with careful management, including liming and full fertilization, are required if sedentary farming is to be taken up on Acrisols. The commonly used 'slash-and-burn' agriculture ('shifting cultivation') may seem primitive at first sight but is really a sophisti-

cated type of land use, developed over centuries of try-and-error. If occupation periods are short (one or a few years only) and followed by a sufficiently long regeneration period (up to several decades), this system probably makes the best use of the limited possibilities of Acrisols. On the whole, low input farming is not possible on Acrisols. Undemanding acidity-tolerant crops such as cashew nut and/or pineapple can be grown with success.

NOTES

ALISOLS

Soils having an argic* B-horizon which has a cation exchange capacity equal to or more than 24 cmol(+)/kg clay and a base saturation (by 1M NH₄OAc at pH 7.0) of less than 50 percent in at least some part of the B-horizon within 125 cm of the surface; lacking the E-horizon abruptly overlying a slowly permeable horizon, the distribution pattern of the clay and the tonguing which are diagnostic for Planosols, Nitisols and Podzoluvisols respectively.

Key to Alisol (AL) Soil Units

Alisols having plinthite within 125 cm of the surface.

Plinthic Alisols (ALp)

Other Alisols showing gleyic^a properties within 100 cm of the surface.

Gleyic Alisols (ALg)

Other Alisols showing stagnic^a properties within 50 cm of the surface.

Stagnic Alisols (ALs)

Other Alisols which are strongly humic^a, having an umbric^{*} or a mollic^{*} A-horizon.

Humic Alisols (ALu)

Other Alisols showing ferric properties.

Ferric Alisols (ALf)

Other Alisols.

Haplic Alisols (ALh)

SUMMARY DESCRIPTION OF ALISOLS

 $\underline{\text{Connotation}}$: acid soils containing high levels of 'free' aluminium; from L. $\underline{\text{alumen}}$, alum.

<u>Parent material</u>: given the right conditions, Alisols can form in nearly any weathering material.

<u>Environment</u>: most common in old land surfaces with a hilly or undulating topography, in wet tropical and monsoonal climates. Light forest is the natural vegetation type.

<u>Profile development</u>: mostly ABtC profiles. Variations among Alisols are mainly connected with variations in environmental conditions (irrigation, drainage, seepage).

^{*} Diagnostic horizon; see Annex 1 for full definition.

a Diagnostic property; see Annex 2 for full definition.

168 Alisols

<u>Use</u>: these weathered soils have a general paucity of macro- and micronutrients; free aluminium is present in toxic quantities. Liming (to correct the pH of the soil and to depress the level of free Al) and full fertilization are needed for permanent cropping but are not always economic. Alisols are traditionally used in shifting cultivation or for low volume production of aluminium tolerant crops. They have some potential for the production of estate crops (oil palm). Alisols in steep land are best left under their natural vegetation cover.

REGIONAL DISTRIBUTION OF ALISOLS

Alisols are exclusive to the tropics where an estimated 100 million hectares are being used for agriculture. The biggest concentrations of Alisols are found in southeast Asia, west Africa, the central part of South America and in the southeastern USA. As Alisols were only recently introduced as a separate Major Soil Grouping, their exact regional distribution is not yet clear. An indication of their main areas of occurrence is given in the chapter on Acrisols.

GENESIS OF ALISOLS

Alisols are characterized by the presence of an argic B-horizon, a mixed clay assemblage that is in a state of transition, high aluminium levels in the subsoil, and a general paucity of bases.

For a description of the processes of clay <u>dispersion</u>, clay <u>transport</u> and clay <u>accumulation</u> in an illuviation horizon, reference is made to the chapter on Luvisols. Note that some authors regard all clay illuviation horizons in highly weathered soils in the wet tropics as relics from a distant past. The process of **ferralitization** by which sesquioxides accumulate in the profile, is described in some detail in the chapter on Ferralsols.

Alisols are comparable with Acrisols but are less strongly weathered, and contain (still) some weatherable primary minerals. The formation of secondary minerals is less advanced than in Acrisols.

CHARACTERISTICS OF ALISOLS

Most Alisols have a brown ochric A-horizon; darker colours occur under a virgin forest vegetation and/or where mineralization of soil organic matter is hindered by (periodic) waterlogging. Stagnic soil properties develop where downward percolation is hindered by a dense argic B-horizon and/or the soil is used for wetland rice cultivation, as in Alisol areas in southeast Asia. Gleyic soil properties and/or plinthite are common in low positions in the landscape. The surface soil structure is rather weak, particularly where the organic matter content of the A-horizon is low. The structure of the Bt-horizon is clearly more stable than that of the surface horizon(s).

Mineralogical characteristics

Dissociation and transformation of 2:1 clay minerals is in process, resulting in the release of considerable quantities of free aluminium and in a gradually decreasing cation exchange capacity of the clay which becomes more and more kaolinitic. Gibbsite is present in negligible quantities. As weathering is still going on, the silt content is relatively high and the content of sesquioxides has not yet reached its maximum. Therefore, Alisols have a lower AEC than Acrisols.

Hydrological characteristics

Alisols with a dense argic B-horizon that are used for wetland rice cultivation, develop pronounced stagnic soil properties ('Anthraquic phase'; see Annex 3). Gleyic properties and/or plinthite are common in Alisols in depressed areas and in lower footslope positions.

Physical properties

The presence of swelling and shrinking clays explains the relatively dense, prismatic Bt-horizons of many Alisols. Their high silt-to-clay ratio and relatively low sesquioxides content, on the other hand, limit (micro)-structure stability.

Another reason for the unstable surface horizons may lie in the low biological activity which is a consequence of the poor nutrient status of the surface soil and the high aluminium content of the subsoil. The level of sesquioxides in the soil is (still) comparatively low and/or the sesquioxides are unevenly distributed over the profile, as witnessed by pale colours in combination with mottles or nodules. Therefore, the positive effects of sesquioxides on the stability of the soil structure are less than in Acrisols and much less than in Ferralsols. Slaking of the surface soil and reduced permeability restrict the infiltration of rain water, reduce the internal drainage of the soil and increase the erosion hazard in sloping land (see Figure 1).

Chemical characteristics

Where easily weatherable minerals are present in an aggressive environment (low in bases, low in pH), aluminium is released from decaying minerals. The low permeability and poor internal drainage prevent adequate discharge of this aluminium. As a result, a considerable part of the cation exchange capacity of the subsoil is occupied by aluminium ions and many crops on Alisols suffer from aluminium toxicity.

Alisols have low levels of 'available' nutrients but have still some nutrient reserves. This makes these soils suitable for crops which grow and produce over a longer period of time (tree crops, forestry).

170 Alisols

MANAGEMENT AND USE OF ALISOLS

Liming, needed to increase the soil-pH and to decrease the level of free aluminium, raises the CEC and depresses the AEC. On the one hand, this may lower soil structure stability but, 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 and susceptible to erosion. Tree crops and perennial crops which minimize soil disturbance by tillage, are to be preferred over annual crops, particularly on sloping land. Oil palm is successfully grown in plantations on Alisols in Malaysia and on Sumatra.



Fig. 1. Slaking and aluminium toxicity harmed upland rice on this eroding hillside with Alisols in Central Sumatra. Note that preserving the surface soil is essential for (food) crop production on these soils.

Alisols have a limited capacity to recover from chemical exhaustion (e.g. in shifting cultivation) or physical degradation. They are only marginally suited for sedentary agriculture. Aluminium toxicity at shallow depth can restrict rooting and cause water stress in the dry season. Deep placement of soil amendments (lime, phosphorus fertilizer) proved beneficial in (experiments in) regions with a pronounced dry season.

LIXISOLS

Soils having an argic* B-horizon which has a cation exchange capacity of less than 24 cmol(+)/kg clay at least in some part of the B-horizon and a base saturation (by 1M NH₄OAc at pH 7.0) of 50 percent or more throughout the B-horizon; lacking a mollic* A-horizon; lacking the E-horizon abruptly overlying a slowly permeable horizon, the distribution pattern of the clay and the tonguing which are diagnostic for Planosols, Nitisols and Podzoluvisols respectively.

Key to Lixisol (LX) Soil Units

Lixisols having plinthite within 125 cm of the surface.

Plinthic Lixisols (LXp)

Other Lixisols showing gleyic properties within 100 cm of the surface.

Gleyic Lixisols (LXg)

Other Lixisols showing stagnic^a properties within 50 cm of the surface.

Stagnic Lixisols (LXs)

Other Lixisols having an albic* E-horizon.

Albic Lixisols (LXa)

Other Lixisols showing ferric properties.

Ferric Lixisols (LXf)

Other Lixisols.

Haplic Lixisols (LXh)

SUMMARY DESCRIPTION OF LIXISOLS

<u>Connotation</u>: strongly weathered soils in which clay is washed down from the surface soil to an accumulation horizon at some depth; from L. <u>lixivia</u>, washed out substances.

<u>Parent material</u>: thoroughly weathered and strongly leached unconsolidated materials; mainly alluvial and colluvial deposits.

<u>Environment</u>: flat to sloping land, predominantly in monsoonal and semi-arid regions. Lixisols are widely seen as polygenetic soils with characteristics formed under a more humid climate in the past.

<u>Profile development</u>: mature ABtC profiles; in places, the argic B-horizon is at the surface or at shallow depth because of erosion.

^{*} Diagnostic horizon; see Annex 1 for full definition.

Diagnostic property; see Annex 2 for full definition.

172 Lixisols

<u>Use</u>: Lixisols have a low nutrient reserve. The unstable (surface) soil structure of Lixisols makes them prone to slaking and to serious erosion in sloping land; Conserving management is important.

REGIONAL DISTRIBUTION OF LIXISOLS

Lixisols cover vast areas in the tropics and subtropics, notably in east-central Brasil, the Indian subcontinent and in west and southeast Africa (Figure 1). Due to the recent introduction of Lixisols as a Major Soil Grouping in the FAO-Unesco Legend, it is not yet possible to provide a reliable figure for their worldwide extent; an estimate of 200 million hectares is probably on the conservative side.

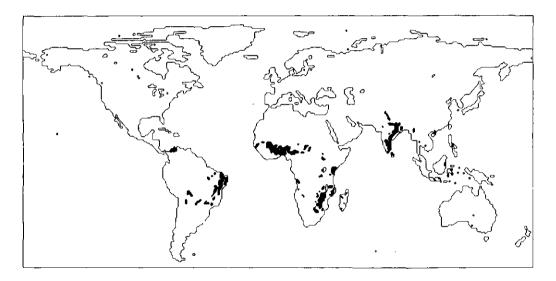


Fig. 1. Lixisol areas worldwide

GENESIS OF LIXISOLS

It is widely accepted that (many) Lixisols started their development under a wetter climate than prevails today. Strong weathering during the early stage of soil formation could be followed by chemical enrichment in more recent times when the climate had changed towards an annual evaporation surplus. Lixisols could also be enriched by base-rich eolian deposits, by biological activity (import of bases from the subsoil), or by lateral seepage. The occurrence of fossil plinthite, and/or coarse reddish iron mottles or indurated iron nodules in the subsurface layers of some Lixisols are also indications of wetness in the past.

Many Lixisols have a reddish or yellowish argic B-horizon as a result of

'rubefaction', a process of dehydration of iron compounds in long dry seasons. The minerals which cause these colours are normally goethite (in the subtropics) and hematite but other iron compounds may be involved as well.

CHARACTERISTICS OF LIXISOLS

Lixisols have normally a brown ochric A-horizon (not seldom shallow as a result of erosion) over a brown or reddish brown argic B-horizon. The subsurface soil may show signs of iron redistribution. The horizon of eluviation of clay and free iron oxides may be sufficiently developed to qualify as an albic E-horizon.

Mineralogical characteristics

Advanced weathering accounts for a low silt-to-clay ratio, a dominance of 1:1 clays (leaching of SiO_2), and for Fe-, Al-and Ti-oxide contents that are higher than those of the less weathered Luvisols. The $\mathrm{SiO}_2/\mathrm{Al}_2\mathrm{O}_3$ ratio of Lixisols is lower than 2.0; gibbsite contents are only slightly lower than in most Ferralsols.

Hydrological characteristics

Lixisols with evidence of periodic water saturation in the upper metre of the profile occur in depressed areas with shallow groundwater but more often because a perched water table forms above the argic B-horizon in periods of wetness.

Physical characteristics

Lixisols have a higher pH and lower AEC than most other weathered tropical soils. Consequently, the structure stability is lower than, for instance, in Acrisols and Ferralsols (for an explanation see the chapter on Ferralsols) and slaking and caking of the surface soil is a serious problem. Stable pseudosand and pseudosilt are virtually absent from Lixisols; the moisture content at low pF values is higher than in Ferralsols or Acrisols with the same contents of organic matter and clay.

Chemical characteristics

As all highly weathered soils, Lixisols have only low levels of available nutrients and low nutrient reserves. Yet, the chemical properties of Lixisols are generally better than those of Ferralsols or Acrisols because of their higher soil-pH and the absence of serious Al-toxicity.

MANAGEMENT AND USE OF LIXISOLS

The low aggregate stability in the surface horizon(s) of Lixisols is conducive to slaking and/or erosion if the topsoil is exposed to the direct impact of rain drops. Tillage of wet soil or the use of heavy machinery can

174 Lixisols

cause serious structure deterioration and compaction of the surface soil and interfere with the rooting of crops. Minimum tillage and erosion control measures such as terracing, contour ploughing, mulching and the use of cover crops help to conserve the soil. Split applications of fertilizers (low CEC!) are needed for good yields.



Fig. 2. Where Lixisols are limited in depth by concretionary layers, farmers in Zambia increase the rooting volume by scraping surface soil on to little mounds. Photograph by courtesy of FAO.

Chemically exhausted and physically deteriorated Lixisols regenerate very slowly unless actively 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. It is often better to devote Lixisols to extensive grazing or forestry.

MINERAL SOILS CONDITIONED BY A (SEMI-)ARID CLIMATE:

SOLONCHAKS SOLONETZ **GYPSISOLS** CALCISOLS

MAJOR LANDFORMS IN ARID REGIONS

Arid and semi-arid regions are, as such, not a morphotectonic category. Dry lands occur on Precambrian shields and in platform areas in the subtropics, e.g. in the Sahara, the Kalahari and in Western Australia, but also in montane rain shadow areas such as in central Asia, north of the Alpine-Himalayan chain from Turkey to China, and in some arctic areas (see Figure 1).

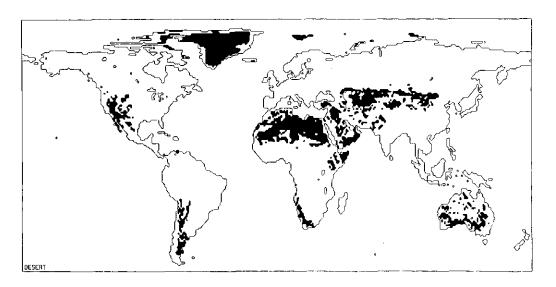


Fig. 1. Desert areas of the world.

Nevertheless, it is evident that important geomorphic processes, and therewith landforms, differ from those in humid regions:

- (1) streams are intermittent or ephemeral;
- (2) braided rivers and unconfined sheet floods are more prominent;
- (3) many rivers do not debouch into the sea (their base level is set by inland depressions without outlet);
- (4) salt lakes are a common landscape feature, and
- (5) eolian processes play an important role, particularly in areas below the 150 mm/yr isohyet.

Many regions that are arid today have known a more humid climate in the past. Conversely, many of the present humid regions were much drier in glacial periods, especially between 20,000 and 13,000 BP when eolian processes influenced land formation more than at present.

This chapter will only concern itself with <u>fluvial</u> and <u>lacustrine</u> landforms in arid environments; sandy eolian deposits were treated in an earlier chapter and loess deposits will be dealt with later, when the major landforms of steppes and steppic regions are discussed.

FLUVIAL LANDFORMS IN ARID AND SEMI-ARID REGIONS

Contrary to popular belief, desert areas are by no means endless seas of sand; only some 15 percent of today's desert areas is covered by sands and a far greater portion consists of rocks and/or gravel. Air photos of desert lands show deep V-shaped river valleys that are now completely dry and filled with sediment. Such valleys ('wadis' or 'oueds') formed during a more humid climatic episode, e.g. between 13,000 and 8,000 BP, at the transition from the Last Glacial to the Early Holocene.

In real deserts, wadis carry water only after torrential rain storms that normally occur once in a few years. At the onset of the rains, the water can still infiltrate into the soil but if the downpours continue, the supply of water soon exceeds the infiltration capacity of the soil and the excess water runs off over the surface. Slaking and caking of the soil surface are common in such barren lands and enhance run-off towards the wadis which become torrential braided streams with a high sediment load. When the stream reaches an (inland) basin, the sediment settles quickly and forms an alluvial fan, not seldom deep in the basin.

At mountain fronts, gentle pediment slopes are formed by the vehement sheet floods. French and Spanish scientists refer to gently sloping fans as 'glacis d'accumulation' and to erosive pediment forms as 'glacis d'érosion'. Pediments are commonly composed of angular, poorly sorted gravel embedded in mud.

LACUSTRINE LANDFORMS IN ARID AND SEMI-ARID REGIONS

When the flood water evaporates, its solutes precipitate in the lowest parts of the basin. First, $CaCO_3$ precipitates as calcite or aragonite. As the brine is further concentrated, gypsum $(CaSO_4.2H_2O)$ segregates, and still later, when the lake is almost dry, halite (NaCl) and other soluble salts. Such salt lakes are called 'playas' or 'shott'. When playa lakes dry out, the mud on the lake floor shrinks and cracks; salt crusts form on the playa floor and in the cracks. Much of the accumulated salt stems from evaporitic marine sediments outside the basin; many Mesozoic (Triassic, Jurassic) and Tertiary sediments are very rich in evaporites.

It depends on local hydrographic conditions whether a playa is wet around the year or dries out. Some playas are fed by groundwater and stay wet (almost) permanently, even in closed basins. Basins, such as the Dead Sea, are fed by perennial rivers and will never dry out but their water is so salty that salts precipitate. Laminated evaporites of considerable thickness can form in this way; the lamination reflects the periodicity of the seasons. The largest evaporite basin in the recent geological history

is the Mediterranean basin: when the Street of Gibraltar became blocked by mountain building some 6 million years ago, a closed or almost closed basin remained. Before the Street opened again, half a million years later, a layer of 1 kilometre of evaporites accumulated on the basin floor.

Many lakes in present-day arid regions were freshwater lakes in the wet period between 12,000 and 8,000 BP. Terraces and/or shorelines from that period extend well above the present lake or lacustrine plain. The same lakes were completely dry in the arid Late Pleniglacial (20,000-13,000 BP). It is evident that a comparatively minor change in climate could cause a fundamental change in the sedimentation regimes of arid lands.

Arid and semi-arid regions harbour a wide variety of soils that occur also in other environments (e.g. Leptosols, Regosols, Arenosols, Fluvisols). Typical dry zone soils are soils whose formation was conditioned by aridity; such soils are marked by accumulation and/or redistribution of anorganic compounds.

High levels of soluble salts built up in SOLONCHAKS; these soils are particularly common in drainless depression areas such as playas and inland basins. SOLONETZ are not marked by a high electrolyte content but by a high proportion of sodium ions in the soil solution; they occur predominantly in temperate and subtropical semi-arid environments. GYPSISOLS have a gypsic or petrogypsic horizon within 125 cm of the surface and CALCISOLS are marked by redistribution of calcium carbonate. The latter two Major Soil Groupings occur in a wide range of landforms, including pediments, lake bottoms, terraces and alluvial fans.

NOTES

SOLONCHAKS

Soils which do not show fluvic^a properties, having salic^a properties and having no diagnostic horizons other than an A-horizon, a histic^{*} H-horizon, a cambic^{*} B-horizon, a calcic^{*} or a gypsic^{*} horizon.

Key to Solonchak (SC) Soil Units

Solonchaks having permafrost[®] within 200 cm of the surface.

Gelic Solonchaks (SCi)

Other Solonchaks showing gleyic^a properties within 100 cm of the surface.

Gleyic Solonchaks (SCg)

Other Solonchaks having a mollic* A-horizon.

Mollic Solonchaks (SCm)

Other Solonchaks having a gypsic* horizon within 125 cm of the surface.

Gypsic Solonchaks (SCj)

Other Solonchaks having a calcic* horizon within 125 cm of the surface.

Calcic Solonchaks (SCk)

Other Solonchaks showing sodic a properties at least between 20 and 50 cm of the surface.

Sodic Solonchaks (SCn)

Other Solonchaks,

Haplic Solonchaks (SCh)

SUMMARY DESCRIPTION OF SOLONCHAKS

Connotation: saline soils; from R. sol, salt, and R. chak, salty area.

<u>Parent material</u>: virtually any unconsolidated material, except for young alluvial sediments that meet the criteria for fluvic properties.

Environment: most common in seasonally or permanently waterlogged areas in (semi-)arid regions with a vegetation of grasses and/or halophytic herbs. Also in (previously inundated) coastal areas in all climates.

<u>Profile development</u>: mostly AC or ABC profiles, often with gleyic properties at some depth. In low-lying areas with a shallow water table, salt accumulation is strongest in the upper few centimeters of the soil. Solonchaks with a deep groundwater table have their highest salt concentration at some depth below the surface.

^{*} Diagnostic horizon; see Annex 1 for full definition. ^a Diagnostic property; see Annex 2 for full definition.

182 Solonchaks

<u>Use</u>: Solonchaks have a limited potential for cultivation of salt-tolerant crops. Many are used for extensive grazing or lie idle.

REGIONAL DISTRIBUTION OF SOLONCHAKS

Solonchaks cover an estimated 260 million hectares worldwide, most of which are in the arid subtropics. They are most extensive in the northern hemisphere, notably in North America, northern Africa, the Middle East and central Asia, but cover also vast areas in South America and in Australia. Figure 1 shows the major occurrences of Solonchaks.

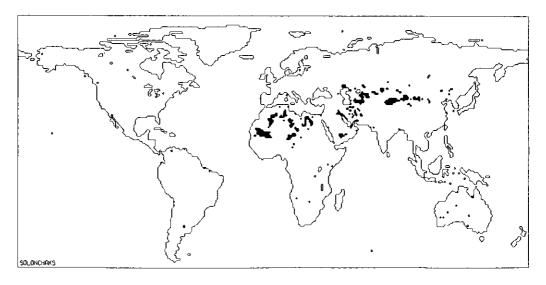


Fig. 1. Solonchaks worldwide.

Solonchaks occur predominantly in inland river basins, (former) lake bottoms, and depressed areas which collect seepage water from surrounding uplands. Coastal lowlands with Solonchaks are much less extensive. The most pronounced Solonchaks are found in regions that were once the bread baskets of prosperous civilisations (e.g. Mesopotamia, the Nile delta, Asia Minor, the Indus floodplain) but succumbed to overpopulation, mismanagement and man's disregard of the limits to ecosystem resilience. The same processes are going on today and for the same reasons, both in developed countries and in developing countries. The world's Solonchak area is growing at an alarming pace.

GENESIS OF SOLONCHAKS

Solonchaks form where there is a considerable evapo(transpi)ration surplus over precipitation (plus irrigation), at least during part of the year. Salts dissolved in the soil moisture remain behind after evaporation of the water and accumulate at the surface of the soil ('external Solonchaks') or at some depth ('internal Solonchaks').

External Solonchaks form in areas with shallow groundwater; internal Solonchaks develop where groundwater is deeper and capillary rise cannot fully replenish evaporation losses in the dry season. Internal Solonchaks may also form through leaching of salts from the surface to deeper layers during wet spells.

The accumulation of salts in the soil is enhanced if a cool wet season alternates with a hot dry season: the solubility of most salts increases in the warm dry season when there is a net upward water flux from the groundwater table to the surface soil, and decreases in the cooler wet season when salts are leached from the surface soil by surplus rainfall. See Figure 2.

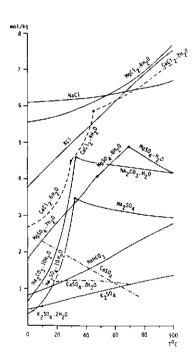


Fig. 2. The solubility of common salts in Solonchaks, expressed in mole anhydrous salt per kg H₂O, as a function of the soil temperature (Braitsch, 1962).

The 'critical depth' of the groundwater, i.e. the depth below which there is little danger that salic properties will develop in the rooted surface soil, depends on soil physical characteristics but also on the evaporative demand of the atmosphere. The USDA Soil Survey Staff considers a depth of 6 feet critical, "especially if the surface is barren and capillary rise is moderate to high".

184 Solonchaks

Normally, the bulk of the salts that accumulate in an area are imported from elsewhere, carried on by rivers from far-away catchment areas or with seepage water or surface run off from adjacent uplands. Accumulated salts can often be traced to deeper geological strata of marine origin (chlorides) or of volcanic origin (sulphates, nitrates).

Figure 3 presents a diagram of a common situation with Solonchaks in bottom land that receives water (and salts) from adjacent uplands.

Spatial and temporal variations in groundwater depth and soil temperature make certain salts accumulate faster than others. This is especially evident in the case of external Solonchaks whereby the <u>mineralogy</u> of salt efflorescences is of particular importance.

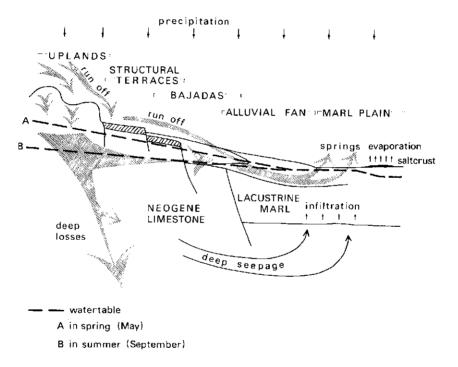


Fig. 3. Import and redistribution of salts in the Great Konya Basin, Turkey. External Solonchaks develop in the bottom lands. Source: Driessen & v.d. Linden, 1970.

Figure 4 presents the stability diagram of minerals in an NaCl-saturated $NaCl-Na_2SO_4-MgCl_2-H_2O$ system. The diagram demonstrates that annual and diurnal temperature fluctuations induce mineralogical transformations.

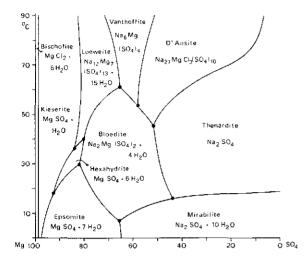


Fig. 4. Stability diagram of minerals in an NaCl-saturated NaCl-Na₂SO₄-MgCl₂-H₂O system. Source: Braitsch, 1962.

An example of a specific type of Solonchak which forms under the influence of <u>diurnal</u> fluctuations in the morphology of salts is the 'puffed Solonchak', an externally saline soil in which the greater part of all salt consists of sodium sulphate. At night, when the temperature at the soil surface is relatively low and air humidity high, crystalline sodium sulphate is present in the surface soil as needle-shaped Mirabilite (Na₂SO₄.10H₂O). The particular morphology of this mineral makes that it pushes the soil aggregates apart when it is formed and this creates a fluffy salt-soil mixture, directly at the surface of the soil. When the temperature rises again during the day, the Mirabilite is converted to water-free Thenardite (Na₂SO₄) crystals that have the appearance of fine flour. Figure 5 shows the soft and fluffy nature of the surface soil of a puffed Solonchak.

Another example of diurnally changing external Solonchaks concerns soils with a dominance of hygroscopic salts such as CaCl₂ or MgCl₂. They form so-called 'Sabakh' soils (sabakh is arabic for morning) that are dark coloured in the morning as a result of moisture absorption during the night. Sabakh soils lose their dark colour again in the course of the day when the temperature rises and air humidity drops to a low value.

An example of an <u>annual</u> cycle in which the morphology of salt minerals plays a role is the formation of 'slick spots', isolated patches of muddy and very saline soil in a field. Slick spots develop early in the dry season in shallow depressions (often hardly recognizable with the naked eye) that are covered with a salt crust. These crusts, e.g. a glass-like Halite (NaCl) crust, are so effective in sealing the underlying saline mud from the air that the soil remains wet throughout the dry season and forms no pores or cracks that can provide passage to rain or leaching water. The crust may dissolve in a subsequent wet season but the unripe mud remains saline and restores its protective crust as soon as the wet season is over.

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The spots thus remain soft, untrafficable and very saline and cannot be reclaimed with conventional (leaching) techniques.



Fig. 5. A spot of puffed Solonchaks in a former lake bottom in Central Anatolia, Turkey. Note the fluffy nature of the surface soil caused by an abundance of needle-shaped mirabilite crystals.

CHARACTERISTICS OF SOLONCHAKS

The horizon differentiation of Solonchaks is normally determined by other factors than their high salt content. Many saline soils in water-logged backswamps are Gleyic Solonchaks; they would have been Gleysols if it were not for their salic properties. Likewise, many Mollic Solonchaks have the appearance of a Chernozem, Kastanozem or Phaeozem, and Calcic and Gypsic Solonchaks are often strongly saline Calcisols and Gypsisols. Saline Histosols, Vertisols and Fluvisols occur as well; they are not classified as Solonchaks because Histosols, Vertisols and Fluvisols key out before Solonchaks. Their salinity is acknowledged in a salic phase. (See Annex 3 for a definition of the salic phase.)

The <u>structure</u> of the surface soil of Solonchaks is often influenced by the high salt content of the soil. The surface layer of Sabakh soils is a muddy mixture of salt and soil particles during early morning hours but becomes a hard crust later in the day. On Solonchaks with a high component of nitrates (and possibly other salts) the soft crust is pushed upwards by gases escaping from the underlying mud; prints of gas bubbles remain visible when the crust, detached from the underlying wet soil, dries out. The fluffy top layer of puffed Solonchaks is a morphological feature that is exclusive to Solonchaks with a high content of sodium sulphate. The most common type of salt crust, however, is a loose cover of salt crystals. Especially in heavy clays, very saline surface layers may exist without any clear efflorescence of salts. Examination with a lens reveals tiny crystals on the faces of crumb or granular structure elements. Very saline pseudo-sand may even be formed that accumulates to clay dunes when exposed to strong winds.

The morphology of internal Solonchaks differs little from that of comparable non-saline soils. Solonchaks have, perhaps, a somewhat stronger subsoil soil structure with, in very saline soils, tiny salt crystals on the faces of the structure elements.

With salic soil properties as the only common characteristic, there is considerable diversity among Solonchaks and a detailed account of their hydrological, physical, chemical and biological properties is not well possible. A few general trends deserve attention in the present context.

Hydrological characteristics

Internal Solonchaks occur in land that lies well above the drainage base. When leached, they may actually furnish (part of) the salts that accumulate in contiguous bottom land with external Solonchaks. Extreme salinization with thick surface crusts occurs in depressions that collect surface run off water from surrounding (higher) land in the winter but dry out in the warm season ('flooded' Solonchaks).

Figure 6 presents a cross-section through an inland basin with severe soil salinity:

Where the water table is at shallow depth, strongly externally saline Gleyic Solonchaks occur (site 8). Slightly above the base level (site 9), the soils are still strongly saline but the highest concentration of salts is at some depth in the soil. This results from a combination of upward salt transport from the groundwater through capillary rise and downward leaching of salt from the surface soil to the zone with the highest salt concentration. In still higher areas with a very deep water table (site 10), non-saline Calcisols and internally saline Calcic Solonchaks occur. The sites 6 and 7 are located on an alluvial fan which drains freely to the low centre of the basin. Soils on the upper part of the fan are invariably non-saline; there is beginning salinization in the lower tract where the groundwater table is at shallow depth. (Note the different scales of the x-axis.)

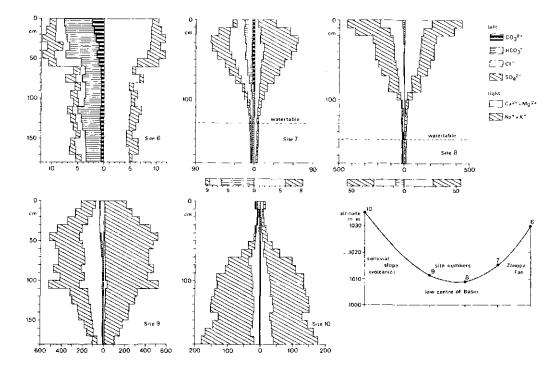


Fig. 6. A cross-section of the Great Konya Basin, Turkey, demonstrates how salinity patterns are influenced by topography and hydrology. All concentrations are expressed in mmol/liter saturation extract or groundwater.

Physical characteristics

Solonchaks which dry out during part of the year have normally strong soil structures. When the salt content is lowered by winter rains or irrigation water, these structures may collapse, particularly if the salts have a large component of sodium and/or magnesium compounds. Strong peptisation of clays may make the soil virtually impermeable to water. In the extreme case, Sodic Solonchaks or even Solonetz may be formed with a dense subsoil and adverse physical properties.

Chemical characteristics

Solonchaks have an 'electric conductivity' value (ECe) in excess of 15 dS/m at 25 °C within 30 cm of the surface at some time of the year, or more than 4 dS/m if the $pH(H_2O, 1:1)$ exceeds 8.5. An ECe of 15 dS/m corresponds with some 0.65 percent salt, or with 150 cmol(+) per liter soil moisture in a completely water-saturated soil paste (see also Figure 6). The ECe is the reciprocal value of the electric resistance (in ohm/cm) and is expressed in mho/cm (in older literature) or dS/m (S stands for 'Siemens').

For a quick orientation, the electric conductivity is often determined on 1:1 or 1:5 soil extracts (EC $_1$ or EC $_5$) but extracts of saturated soil pastes are used in base laboratory work. Values obtained with different methods cannot always be compared, among other reasons because the 'suspension effect' (different at different dilution ratios) influences the outcome of the conductivity measurement.

Table 1 suggests an indicative grading of salt-affected soils on the basis of their salt contents.

TABLE 1.
Indicative soil salinity classes and the effects of soil salinity on crop performance.

ECe at 25 °C (dS/m)	Salt Comextract	ncentration soil) (percent)	FAO-Unesco qualification	Effect on Crops
<2.0 2.0-4.0	<2 2-4	<0.15	- -	mostly negligible some damage to sensitive crops
4.0-8.0	4-8	0.15-0.35	salic phase (Solonchak)	serious damage to
8.0-15	8-15	0.35-0.65	salic phase (Solonchak)	only tolerant crops succeed
>15	>15	>0.65	Solonchak	few crops survive

As mentioned, the <u>quantity</u> of salts varies among Solonchaks but also the salt <u>composition</u>. Russian soil scientists characterize salt affected soils (Solonchaks and salic phases in other soil groupings) on the basis of anion ratios in the saturation extract. See Table 2. The Solonchaks in Figure 6 qualify as sulphate-chloride soils according to this classification.

TABLE 2. Classification of saline soils based on anion ratios (Plyusnin, 196?).

Sulphate soils Chloride-sulphate soils Sulphate-chloride soils Chloride soils	C1/SO ₄ C1/SO ₄ C1/SO ₄ C1/SO ₄	Pljusnin <0.5 0.5-1.0 1.0-5.0 >5.0	<0.2 0.2-1.0	Sadovnikov <0.2 0.2-1.0 1.0-5.0 >5.0
Soda soils Sulphate-soda soils Soda-sulphate soils	CO ₃ /SO ₄ CO ₃ /SO ₄ CO ₃ /SO ₄		46 	<0.05 0.05-0.16 >0.16

190 Solonchaks

Biological characteristics

Faunal activity is depressed in most Solonchaks and ceases entirely in soils with 3 percent salt or more. In severely salt-affected lands, the vegetation is sparse and restricted to halophytic shrubs, herbs and grasses that tolerate severe physiological drought (and can cope with periods of excessive wetness in areas with seasonally flooded Solonchaks).

MANAGEMENT AND USE OF SOLONCHAKS

Salt accumulation affects plant growth in two ways:

- (1) <u>indirectly</u> by skewing the composition of the soil solution which upsets the availability of plant nutrients, and
- (2) <u>directly</u> by inducing physiological drought as a consequence of the high osmotic pressure of the soil moisture.

High contents of a specific ion in the soil solution can interfere with the uptake of other ions. Such antagonistic effects are known to exist, for example, between sodium and potassium, between sodium and calcium and between magnesium and potassium. Very harmful are excess levels of sodium salts, chlorides (disturb normal N-metabolism) and MgSO4. The main damage is done, however, because the plant cannot compensate for the combined hydrostatic, adsorptive and osmotic forces that hinder water uptake. In practice, salinity augments the effects of aridity.

Solonchaks cannot be used for normal cropping. Only after the salts are leached out of the soil (which then ceases to be a Solonchak) can good yields be expected. Normal irrigation is inadequate for desalinization. Often, the reverse happens: salts contained in the irrigation water remain behind in the soil and the salt level builds up. The use of saline irrigation water must be avoided whenever possible, and excess water must be applied above the irrigation requirement to maintain a downward water flow in the soil. Just as important: drainage facilities must be designed to keep the groundwater table at sufficient depth. Amendments such as gypsum are sometimes used to correct imbalances at the exchange complex and to maintain a good permeability of the soil.

SOLONETZ

Soils having a natric * B-horizon.

Key to Solonetz (SN) Soil Units

Solonetz showing gleyic^a properties within 100 cm of the surface.

Gleyic Solonetz (SNg)

Other Solonetz showing stagnic^a properties within 50 cm of the surface.

Stagnic Solonetz (SNz)

Other Solonetz having a mollic* A-horizon.

Mollic Solonetz (SNm)

Other Solonetz having a gypsic* horizon within 125 cm of the surface.

Gypsic Solonetz (SNj)

Other Solonetz having a calcic* horizon within 125 cm of the surface.

Calcic Solonetz (SNk)

Other Solonetz.

Haplic Solonetz (SNh)

SUMMARY DESCRIPTION OF SOLONETZ

Connotation: 'sodic' or 'alkali' soils with high sodium saturation; from R. sol, salt, and R. etz, strongly expressed.

Parent material: unconsolidated materials, mostly fine-textured.

<u>Environment</u>: major concentrations of Solonetz are in flat or gently sloping natural grasslands in semi-arid, temperate and subtropical regions.

<u>Profile development</u>: ABtnC and AEBtnC profiles with a black or brown A-horizon over a natric B-horizon. Well developed Solonetz can have a (beginning) albic E-horizon directly over a natric B-horizon having strong prismatic or columnar structure elements with rounded tops. A calcic and/or gypsic horizon may be present below the natric B-horizon. In many Solonetz, the pH of the soil is around 8.5 due to dominance of sodium carbonate in the soil solution.

<u>Use</u>: high levels of exchangeable sodium affect plant performance, either directly (toxicity) or indirectly (structure deterioration). Solonetz in temperate regions are used for arable farming or grazing; subtropical Solonetz are mostly in use as range land or lie idle.

^{*} Diagnostic horizon; see Annex 1 for full definition.

Diagnostic property; see Annex 2 for full definition.

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REGIONAL DISTRIBUTION OF SOLONETZ

Solonetz occur predominantly in the dry interior parts of North America, Eurasia and Australia, with smaller occurrences in the semi-arid/subtropical Africa and South America, mostly in flat or gently sloping plains and not seldom in association with Solonchaks. Figure 1 presents the main areas of occurrence.

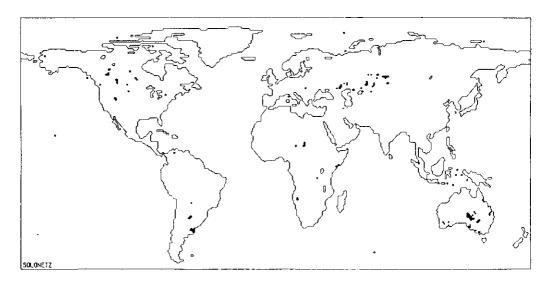


Fig. 1. Solonetz worldwide.

GENESIS OF SOLONETZ

The essential characteristic of Solonetz is the natric B-horizon which has an Exchangeable Sodium Percentage (ESP) of 15 or higher. The ESP is a function of the exchange properties of the soil material and the chemical composition of the soil solution in equilibrium with this soil material.

Not only the sodium content of the soil moisture is important but also the concentrations of other ions, particularly the divalent ions that are preferentially adsorbed at the exchange complex. A measure of the composition of the soil solution as relevant for Na[†] adsorption at the cation exchange complex is the Sodium Adsorption Ratio (SAR), defined as:

$$SAR = Na^{+}/[(Ca^{2+} + Mg^{2+})/2]^{0.5}$$

where Na^+ , Ca^{2+} and Mg^{2+} are expressed in cmol(+) per liter of solution.

($\underline{\text{NOTE THAT}}$ units are important here; other units, e.g. mmol/l, give different SAR values.)

The theoretical relation between SAR and ESP reads:

$$ESP/(100-ESP) = 0.031*K*SAR$$

where K has values of 0.75 and 1.5 $(\text{cmol}(+)/1)^{-0.5}$ for kandite and smectite clays respectively. The above relations have been found satisfactory in most circumstances, but poor correlation has occurred with high-carbonate waters. Various adapted relations have been proposed to cope with such conditions.

The SAR gives an indication of the chance that an unfavourably high ESP develops when a soil is brought in contact with a particular groundwater or irrigation water. The 'sodium hazard' is low if SAR<10, medium if between 10 and 18, high if between 18 and 26, and very high if SAR>26.

In areas with a marine history the sodium may originate from NaCl but many Solonetz in the (sub)tropics have soda (Na_2CO_3) as the dominant sodium compound. Soda can form in two ways:

- (1) by evaporation of water containing an excess of bicarbonate ions over divalent cations (Ca^{2+} and Mg^{2+}), and
- (2) biologically, by reduction of sodium sulphate.

If water with Ca- and Mg-bicarbonates evaporates, calcium and magnesium carbonates precipitate and the SAR value increases. Soda soils, with a high pH, develop if the bicarbonate content of the evaporating water is higher than its content of $(\text{Ca}^{2+}+\text{Mg}^{2+})$. Excess bicarbonate is in practice always sodium bicarbonate which is eventually transformed to Na_2CO_3 .

The biological formation of soda from sodium sulphate follows the sequence Na_2SO_4 --> Na_2S --> Na_2CO_3 + H_2S , whereby the hydrogen sulfide gas leaves the system. In addition to Na_2SO_4 , this reaction requires the presence of organic matter and (periods of) anaerobic conditions.

Soda formation raises the pH of the soil to a value near pH 8.5; silica and alumina can then dissolve from silicate clays. Clay destruction in the upper horizon and clay formation in the middle horizon, together with illuviation of clay, are the processes responsible for the formation of a natric B-horizon.

A mature Solonetz has a natric B-horizon with columnar structure elements that have rounded tops as a result of peptisation and destruction of clay. The process is limited to the top of the B-horizon because percolating water stagnates on this dense illuviation layer (see Figure 2). Stagnic properties may even develop directly above the natric horizon (Stagnic Solonetz).

As mentioned in the chapter on Solonchaks, Solonetz can also form through progressive leaching of salt-affected soils. Even soils that were initially rich in calcium may eventually develop a natric B-horizon. Prolonged leaching and exchange of adsorbed Na* by H* will ultimately produce an eluvial E-horizon and a low pH. Such strongly degraded soils are known as 'Solods'.

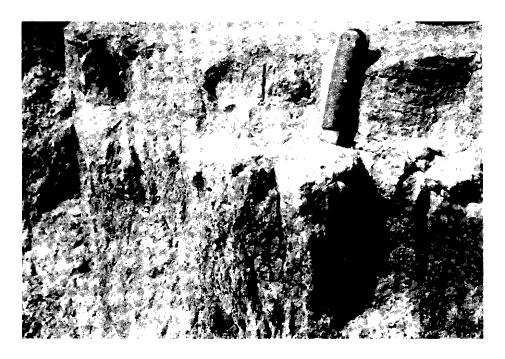


Fig. 2. A mature Solonetz. Note the columnar structure elements with rounded tops in the natric B-horizon. Photo by ISRIC, Wageningen.

CHARACTERISTICS OF SOLONETZ

Most (Haplic) Solonetz have a thin loose litter layer resting on black humified material about 2-3 cm thick. This overlies a brown granular ochric A-horizon which abruptly changes into a natric B-horizon with coarse (often round-topped) prismatic or columnar structure elements and grading with depth into a massive subsoil.

Mollic Solonetz have a thick, dark mollic A-horizon instead of an ochric A-horizon, and Calcic and Gypsic Solonetz have a calcic or gypsic horizon below the natric B-horizon. Gleyic Solonetz have a mottled subsoil and sometimes dispersed organic matter is translocated from the A-horizon to the top of the B-horizon; Stagnic Solonetz have stagnic properties with reduced surface soil material tonguing into the underlying, oxidized natric B-horizon.

Physical characteristics

Clayey Solonetz are waterlogged in the wet season and even ponding of water may occur. The dense subsurface soil hinders root penetration. When ploughed, many Solonetz have a lumpy surface with the top of the natric B-horizon showing up in a polygonal pattern in the bottom of the furrows. On the whole, Solonetz have poor physical properties.

Chemical characteristics

The high ESP of Solonetz is directly or indirectly detrimental to the suitability of these soils for cropping: <u>directly</u> because a high proportion of sodium ions in the soil is toxic to some plants and disturbs the uptake of essential plant nutrients, and <u>indirectly</u> because sodicity is associated with dense subsoils that interfere with downward percolation of water and hinder the growth of roots. The impression exists that sensitive crops show true sodium toxicity symptoms already at low ESP values whereas tolerant crops are stunted at high ESP values largely because of sodium-induced adverse physical soil conditions.

MANAGEMENT AND USE OF SOLONETZ

How detrimental high sodium saturation is in a certain situation is partly determined by other soil parameters such as the depth of the natric B-horizon and the presence or absence of gypsum, and by the sodicity tolerance of the crop.

Soils with a predominantly smectitic clay assemblage show already serious structure deterioration when the SAR exceeds 9; illitic and vermiculitic soils degrade at SAR>16 and the most stable soils (kaolinitic soils and soils rich in sesquioxides) deteriorate only if the SAR exceeds 26 in the absence of salinity. The indicative ranges for the sodium hazard of waters that were mentioned before, must be interpreted with the relative stability of the soil structures in mind. Also, it may not be forgotten that the presence of e.g. gypsum in a soil can mitigate the effects of high-SAR (irrigation) water. There are strong indications that a high percentage of exchangeable magnesium affects the soil structure in a similar manner as a high ESP.

The traditional way of reclaiming Solonetz is by flushing with calciumrich water. In the case of Gypsic Solonchaks, deep plowing may eliminate the need for (expensive) application of gypsum. Where a low hydraulic conductivity of the soil precludes effective leaching, one or more years under a sodium-tolerant, deeply rooting grass crop (e.g. Rhodes grass) may make the soil reclaimable. Solonetz in Canada are widely grown to wheat; those in sparsely populated areas in the semi-arid subtropics are commonly left idle or they are used for extensive grazing.

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GYPSISOLS

Soils having a gypsic* or a petrogypsic* horizon, or both, within 125 cm of the surface; having no diagnostic horizons other than an ochric* A-horizon, a cambic* B-horizon, an argic* B-horizon permeated with gypsum or calcium carbonate, a calcic* or a petrocalcic* horizon; lacking the characteristics which are diagnostic for Vertisols or Planosols; lacking salic* properties; lacking gleyic* properties within 100 cm of the surface.

Key to Gypsisol (GY) Soil Units

Gypsisols having a petrogypsic * horizon, the upper part of which occurs within 100 cm of the surface.

Petric Gypsisols (GYp)

Other Gypsisols having a calcic* horizon.

Calcic Gypsisols (GYk)

Other Gypsisols having an argic B-horizon.

Luvic Gypsisols (GY1)

Other Gypsisols.

Haplic Gypsisols (GYh)

SUMMARY DESCRIPTION OF GYPSISOLS

Connotation: soils with substantial accumulation of calcium sulfate; from L. gypsum, gypsum.

<u>Parent material</u>: unconsolidated alluvial, colluvial or eolian deposits rich in bases.

Environment: level to hilly land and depression areas (e.g. former inland lakes) in regions with an aridic moisture regime.

<u>Profile development</u>: AB(t)C profiles with a yellowish brown ochric A-horizon over a cambic or (relic?) argic B-horizon. Accumulation of calcium sulfate, with or without carbonates, is concentrated in and below the B-horizon.

<u>Use</u>: soils having less than 25 percent gypsum in the upper 30 cm of soil can be grown to cereal crops, alfalfa, cotton, etc. Yields may still be depressed due to nutrient imbalances, stoniness, and/or mechanical hindrances. Uneven subsidence of the land surface due to dissolution of gypsum in percolating (irrigation) water is a further limitation. Irrigation canals must be lined to prevent the walls from caving in.

^{*} Diagnostic horizon; see Annex 1 for full definition.

a Diagnostic property; see Annex 2 for full definition.

REGIONAL DISTRIBUTION OF GYPSISOLS

Gypsisols occur in the same regions as Calcisols but are much less widespread. The worldwide extent of Gypsisols is probably of the order of 100 to 150 million hectares; major areas are found in the Middle East and the southern USSR (see Figure 1), in southeast and central Australia and the southwestern USA.

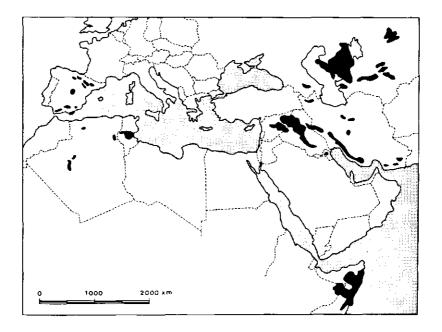


Fig. 1. Major occurrences of Gypsisols in Europe and the Near East. Source: Van Alphen & Romero, 1971.

GENESIS OF GYPSISOLS

Most of the gypsum in soils originates from Triassic, Jurassic and Cretaceous evaporites with gypsum ($CaSO_4$, $2H_2O$) and/or anhydrite ($CaSO_4$), and from (predominantly) Miocene gypsum deposits that are normally interbedded in marls or clays. Rarely is gypsum formed directly in the soil. This has, for example, been observed in areas with pyritic sediments in southwest Siberia where sulfate ions, formed when sulfides oxidized upon forced drainage of the land, precipitated as gypsum at depths of 20 to 150

cm below the surface. In Georgia, USSR, gypsum formation was observed where saline, ${\rm Na_2SO_4}\text{-containing}$ seepage water came in contact with Dolomite weathering.

Normally, Gypsisols form through dissolution of gypsum from gypsiferous weathering materials followed by downward transport with percolating soil moisture and precipitation of gypsum deeper down. The gypsum may precipitate as fine, white, powdery crystals in pockets or in former root channels ('gypsum pseudomycelium'), as pendants below pebbles and stones (see Figure 2), as coarse crystalline gypsum sand or even as rosettes ('desert roses'). In arid regions with hot, dry summers, gypsum (CaSO₄.2H₂O) dehydrates to loose, powdery hemihydrate (CaSO₄.0.5H₂O) which reverts to gypsum during the moist winters. If sufficiently abundant, so-formed (highly irregular) gypsum crystals may cluster together to compact layers or surface crusts that can become tens of centimetres thick.



Fig. 2. Gypsum pendants in a gravel terrace near Deir es Zor, Syria. Photograph by courtesy of ISRIC, Wageningen.

CHARACTERISTICS OF GYPSISOLS

The typical Gypsisol has a 20 to 40 cm thick, yellowish brown, loamy or clayey A-horizon over a pale brown B-horizon with distinct white gypsum pockets and/or pseudomycelium. The surface layer consists of strongly de-gypsified weathering residues and has a low organic matter content and a weak, subangular blocky structure. The gypsic horizon is most clearly developed in the lower B-horizon or slightly deeper and can be anything from a soft, powdery and highly porous mixture of gypsum, lime and clay, to a hard and massive layer of almost pure, coarse gypsum crystals.

Hydrological characteristics

Gypsisols have highly variable hydraulic properties. Saturated hydraulic conductivity values vary from 5 to >500 cm/d; infiltration of surface water is close to zero in some encrusted soils whereas very high percolation losses occur in soils in which dissolution of gypsum has widened fissures, holes and cracks to interconnected subterranean cavities. Infilling of the cavities with surface soil material makes it necessary to level the land surface each year which makes the valuable topsoil ever shallower. See Figure 3.

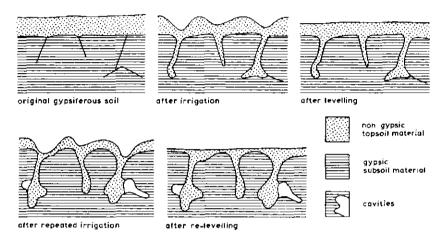


Fig. 3. Cavity formation, uneven subsidence, and stripping of the surface soil as a consequence of prolonged irrigation of shallow Gypsisols. Source: Van Alphen & Romero, 1971.

Physical characteristics

Most de-gypsified surface horizons contain 40 percent clay or more, and have an 'available' water holding capacity of 25 to 40 volume percent. Surface soils with more than 15 percent gypsum contain seldom more than 15 percent clay and their retention of 'available' soil moisture does not exceed 25 volume percent. Loamy surface soils slake easily to a finely platy surface crust which hinders infiltration of rain water and promotes sheet wash and gully erosion to the extent that deep (petro)gypsic horizons become exposed in spite of a low annual rainfall sum of only 200 to 450 mm.

Chemical characteristics

Small quantities of gypsum do not harm plants in any way but gypsum contents of more than 25 percent, as occur in many gypsiferous subsoils, upset the nutrient uptake by plants and lower the availability of phosphorus, potassium and magnesium.

The total element contents of Gypsisol surface horizons amount to less than 2500 mg N/kg, less than 1000 mg P_2O_5/kg (of which less than 60 mg/kg is

considered 'available'), and less than 2000 mg K_2O/kg : heavy fertilization is needed for good yields. The cation exchange capacity decreases with increasing gypsum content of the soil material and is typically around 20 cmol(+)/kg in the surface soil and around 10 cmol(+)/kg deeper down. The base saturation is nearly 100 percent.

MANAGEMENT AND USE OF GYPSISOLS

Gypsisols with not more than a few percent gypsum in the upper 30 cm layer (60 cm if irrigated) can be used for the production of small grains, cotton, alfalfa, etc. High irrigation rates in combination with forced drainage made it possible to produce excellent yields of alfalfa hay (10 tons per hectare), wheat, apricots, dates, maize and grapes on Gypsisols with more than 25 percent powdery gypsum.

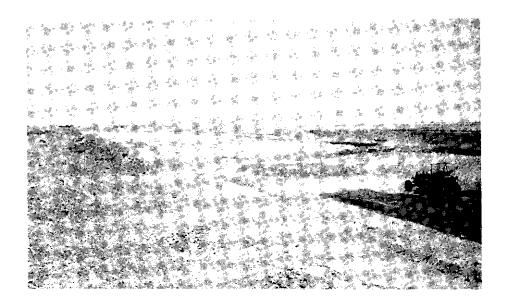


Fig. 4. Lining prevents caving in of irrigation canals in Gypsisols. Photograph by courtesy of ISRIC, Wageningen.

Irrigated agriculture on Gypsisols is plagued by quick dissolution of soil gypsum (see Figure 4) resulting in irregular subsidence of the land, caving in canal walls, and corrosion of concrete structures, and by the occurrence of shallow petrogypsic horizons. The latter obstruct root growth, and interfere with water supply to the crop and with soil drainage.

Dry farming on Gypsisols makes use of fallow years and other water harvesting techniques but is not very profitable; dry farming Gypsisols with more than 25 percent gypsum cannot be recommended.

NOTES

CALCISOLS

Soils having one or more of the following: a calcic* horizon, a petrocalcic* horizon or concentrations 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 permeated with calcium carbonate; lacking the characteristics which are diagnostic for Vertisols or Planosols; lacking salic® properties; lacking gleyic® properties within 100 cm of the surface.

Key to Calcisol (CL) Soil Units

Calcisols having a petrocalcic * horizon, the upper part of which occurs within $100~\mathrm{cm}$ of the surface.

Petric Calcisols (CLq)

Other Calcisols having an argic B-horizon.

Luvic Calcisols (CLe)

Other Calcisols.

Haplic Calcisols (CLh)

SUMMARY DESCRIPTION OF CALCISOLS

<u>Connotation</u>: soils with substantial accumulation of calcium carbonate; from L. <u>calcis</u>, lime.

<u>Parent material</u>: mostly alluvial (lacustrine) sediments and colluvial and eolian deposits of base-rich weathering material.

Environment: level to hilly land with an aridic moisture regime and a sparse vegetation of xerophytic shrubs and ephemeric grasses.

<u>Profile development</u>: AB(t)C-profiles with a pale brown ochric A-horizon over a cambic or argic B-horizon. Accumulation of carbonates at some depth below the soil surface.

<u>Use</u>: dryness, and in places also stoniness and/or the presence of a petrocalcic horizon, limit the suitability of these soils for agriculture. Drought resistant arable crops such as wheat and sunflower are grown under rainfed conditions. If irrigated, drained (to prevent salinization) and fertilized, Calcisols can produce good yields of fodder and vegetable crops. Many areas with Calcisols are used for low volume grazing of cattle and sheep.

^{*} Diagnostic horizon; see Annex 1 for full definition.
a Diagnostic property; see Annex 2 for full definition.

REGIONAL DISTRIBUTION OF CALCISOLS

It is difficult to estimate the worldwide extent of Calcisols with any measure of accuracy, partly because the grouping was only recently introduced, but also because many Calcisols occur together with Solonchaks that are actually salinized Calcisols, and/or with other soils with carbonate enrichment that do not key out as Calcisols. The total Calcisol area amounts probably to well over one billion hectares, nearly all of it in the arid and semi-arid subtropics of both hemispheres. Figure 1 gives an indication of the regions where Calcisols occur.

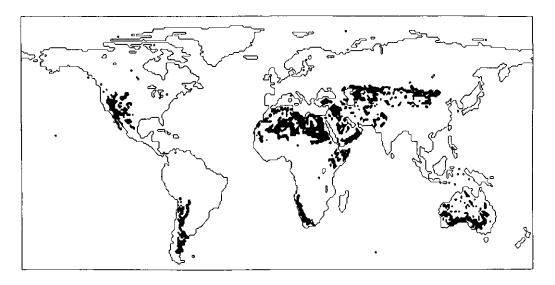


Fig. 1. Calcisols worldwide.

GENESIS OF CALCISOLS

Many Calcisols are old soils when counted in years, but their development was slowed down by recurrent periods of drought in which such important soil forming processes as chemical weathering, accumulation of organic matter, and translocation of clay, came to a virtual standstill. As a result, only an ochric A-horizon could develop and the modification of subsoil layers did normally not advance beyond the formation of a cambic B-horizon. Many Calcisols are 'polygenetic': their formation took different courses during different geologic eras with different climates. For instance, the argic B-horizon of Luvic Calcisols is widely considered a relic from eras with a more humid climate than at present.

The most prominent soil forming process in Calcisols - the process from which the soils derived their name - is the translocation of calcium carbonate from the surface horizon to an accumulation layer at some depth. Under certain conditions, e.g. in eroding land or in land that is intensively homogenized by burrowing animals, lime concretions occur right at the surface of the soil.

Dissolution of calcite $(CaCO_3)$ and subsequent accumulation in a calcic horizon is governed by two factors:

- (1) the CO2-pressure of the soil air, and
- (2) the concentrations of dissolved ions in the soil moisture.

The following equilibria are involved (the pH-ranges over which the equilibria are in operation are shown in Figure 2):

$$CO_2 + H_2O - H_2CO_3^{\circ}$$
 $H_2CO_3^{\circ} = HCO_3^{-} + H^{+}$
 $HCO_3^{-} = CO_3^{2-} + H^{+}$

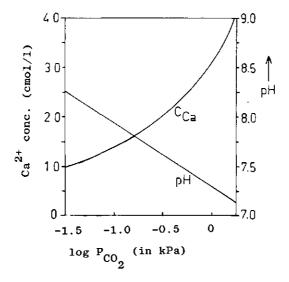
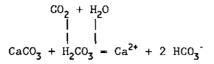


Fig. 2. Solubility of calcite at different ${\rm CO_2}$ -pressures and corresponding pH-value. Source: Bolt & Bruggenwert, 1979.

For all practical purposes, the equilibria which are relevant to the dissolution or precipitation of calcite in soils (pH <9) can be viewed as follows:



An increase in the CO_2 -content of the soil-air, associated with a decrease in soil-pH, drives the reaction to the right: calcite dissolves and the concentrations of Ca^{2+} - and HCO_3^{-1} ons in the soil solution rise. Alternatively, calcite dissolves if (rain) water with a low Ca^{2+} -concentration flushes the soil.

Precipitation of calcite occurs if the reaction is driven to the left, e.g. by a lowering of the $\rm CO_2$ -pressure (with a consequent rise in pH), or by an increase in ion concentrations to the point where the solubility product of $\rm Ca^{2+}$ and $\rm CO_3^{-2-}$ is exceeded.

The formation of a calcic horizon is now easily understood: The partial CO_2 -pressure of the soil-air is normally highest in the A-horizon where root activity and respiration by micro-organisms cause CO_2 contents to be 10 to 100 times higher than in the atmospheric air. As a consequence, calcite dissolves and Ca^{2+} - and HCO_3^{-1} ons move downward with percolating soil moisture, particularly during and directly after a rain shower. The water may take up more dissolved calcite on its way down.

Evaporation of water and a decrease in partial CO₂-pressure deeper in the profile (fewer roots and less soil organic matter and micro-organisms) cause saturation of the soil solution and precipitation of calcite. The precipitated calcite is not or only partly transported back with ascending water because much of this water moves in the vapour phase. (The water table in Calcisols is normally deep; where there is capillary rise to the solum, calcite accumulates at the depth where the capillary water evaporates.)

Calcite doesnot (always) precipitate evenly distributed over the soil matrix. Root channels and worm holes that are connected with the outside air act as ventilation shafts in which the partial ${\rm CO_2}$ -pressure is much lower than in the soil around it. When ${\rm Ca(HCO_3)_2}$ -containing soil water reaches such a channel, it loses ${\rm CO_2}$ and calcite precipitates along the walls. Continued accumulation produces the 'pockets of soft powdery lime' that are mentioned in the definition of Calcisols. Where narrow root channels become filled with calcite, so-called 'pseudomycelium' forms. Other characteristic forms of calcium carbonate accumulation in Calcisols are hard concretions ('calcretes') and calcite pendants below pebbles.

The high soil temperature and high pH of Calcisols enhance dissolution of silica from feldspars, ferromagnesian minerals, etc. Where there is (or was) sufficient moisture in some period of the year for translocation of dissolved silica, this may have augmented the induration of the layer with calcite accumulation. However, cementation of a petrocalcic horizon is in first instance by calcium and magnesium carbonates.

CHARACTERISTICS OF CALCISOLS

Most Calcisols have a thin (=<10 cm), brown or pale brown A-horizon over a slightly darker Bck-horizon and/or a yellowish brown Cck- or Cmk-horizon that is speckled with white calcite mottles. The organic matter content of the surface soil is low on account of a sparse vegetation and rapid decomposition of vegetal debris. The surface soil is crumb or granular, but platy structures can occur as well, possibly enhanced by a high percentage of adsorbed magnesium. The subsurface soil has weak blocky structures or is structureless; the structures are coarser, stronger and often more reddish in colour in an illuvial Bt-horizon than in soils without clay translocation.

The highest calcite concentration is normally found in the deeper B-horizon and directly below the B-horizon. Burrowing mammals homogenize the soil and bring hardened carbonate nodules to the surface; their backfilled holes often extend deep into the C-horizon ('krotovinas').

Hydrological characteristics

Calcisols are well drained and are wet only in part of the (short) rainy season when there is just enough downward percolation to flush soluble salts to the deep subsoil. One reason why Calcisols as a taxonomic unit have good drainage properties is that (carbonatic) soils in wet positions (depressions, seepage areas) quickly develop salic properties in their aridic environment and key out as Solonchaks.

Physical characteristics

Most Calcisols have a medium or fine texture and a good water holding capacity. Where surface soils are silty, slaking and crust formation may hinder the infiltration of rain and irrigation water. Surface run-off over the bare soil causes sheet wash and gully erosion and, in places, exposure of a petrocalcic horizon.

Chemical characteristics

Calcisols are potentially fertile soils. They contain only 1 or 2 percent organic matter (C/N-ratio <10) but they are rich in nitrate and mineral nutrients. The pH(H₂0; 1:1) is 7 to 8 in the surface soil and slightly higher at a depth of 80 to 100 cm where the carbonate content may be 25 percent or more. The cation exchange capacity is highest in the A-horizon (10 to 25 cmol(+)/kg) and decreases slightly with depth. The exchange complex is completely saturated with bases; Ca^{2+} and Mg^{2+} make up more than 90 percent of all adsorbed cations.

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MANAGEMENT AND USE OF CALCISOLS

Vast areas of non-irrigated Calcisols are used for low-volume grazing. A crop of wheat or sunflower could be grown after one or a few fallow years, but Calcisols reach their full productive capacity only if they are carefully irrigated. Fodder crops such as 'el sabeem' (sorghum bicolor), Rhodes grass and alfalfa, are tolerant of high calcium levels. Cotton and a score of vegetable crops have successfully be grown on Calcisols fertilized with nitrogen, phosphorus and trace elements (Fe, Zn). Furrow irrigation is superior to basin irrigation on slaking Calcisols because it reduces seedling mortality due to surface crusting; pulse crops in particular are very vulnerable in the seedling stage.

In places, arable farming is hindered by stoniness of the surface soil and/or a petrocalcic horizon at shallow depth.

MINERAL SOILS CONDITIONED BY A STEPPIC CLIMATE:

KASTANOZEMS CHERNOZEMS PHAEOZEMS GREYZEMS

MAJOR LANDFORMS IN STEPPE REGIONS

Steppes and steppic regions have an annual rainfall sum of some 400 mm; more than twice the quantity that falls in true desert areas. There, the rainfall is insufficient to support a vegetation which could protect the land from erosion. In strong winds, sand saltates to a height of several metres above the ground, and its impact when falling back is such that even fine gravel is moved ('creep') and rock surfaces are polished. The dunes and sand plains which form in the process have been discussed earlier in this text, in a chapter on the landforms in residual and shifting sands. Finer particles than sand are transported in suspension over large distances until they settle as 'loess', predominantly in the steppe regions adjacent to the desert zone. The 'red rain' which falls in western Europe every now and then, is Saharan dust; it is actually a thin loess deposit.

LANDFORMS IN REGIONS WITH LOESS

Chinese records make mention of major (historical) periods of loess deposition between 400 and 600 AD, between 1000 and 1200 AD and between 1500 and 1900 AD (the 'Little Ice Age'), but the most extensive occurrences of loess on earth lie in the steppic regions of eastern Europe and the USA and are of Pleistocene age (see Figure 1). Therefore, some attention must be given to the relation between Ice Age aridity and loess deposition.

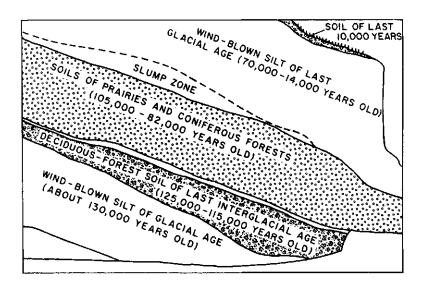


Fig. 1. Climatic history recorded in a Czechoslovakian brickyard. Events of the past 130,000 years are recorded as a sequence of soils and loesses in a quarry at Nove Mesto. (Based on the work of G.J. Kukla.)

During the Late Pleniglacial, between 20,000 and 13,000 BP, some 25 percent of the land surface became covered with continental ice sheets (versus some 10 percent today), the sea level sunk to about 100 metres below the present level, and large parts of the world were extremely arid. The Amazon rain forest dwindled to isolated refugia, European forests disappeared but for small sheltered areas, and large parts of the globe turned to tundra, steppe, savannah or desert.

Clearly, eolian processes were much more important at that time than at present. Large parts of the present temperate zone, from the cover sands of the Netherlands to the sand dunes in northeastern Siberia, are Ice Age (eolian) sands. South and east of this cover sand belt lies a belt of loess deposits, extending from France, across Belgium, the southern Netherlands, Germany and large parts of Eastern Europe into the vast steppes of the Soviet Union, and further east to Siberia and China. See Figure 2. A similar east-west loess belt exists in the USA and less extensive areas occur on the southern hemisphere, e.g. in the Argentinian pampas.

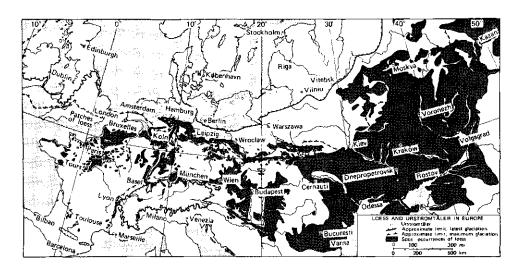


Fig. 2. Distribution of loess in Europe. Source: Flint, 1971.

Loess is a well-sorted, usually calcareous, unstratified, yellowish-grey, eclian clastic sediment. It consists predominantly of silt-sized quartz grains (2-50 $\mu \rm m$), and contains normally less than 20 percent clay particles and less than 15 percent sand. It covers the land surface as a blanket, which is less than 8 metres thick in the Netherlands (exceptionally 17 m), but reaches 40 metres in eastern Europe and 330 metres in China. Loess is a very porous material and vertical walls remain remarkably stable, but it slakes easily so that exposed surfaces are prone to erosion.

The loess material is probably produced by abrasion of rock surfaces by glaciers. It is difficult, however, to identify specific source areas for specific loess deposits, because the various loess deposits have a surprisingly homogeneous mineralogy. A possible explanation might be that glaciers abrade large surfaces of diverse mineralogy, so that the mineralogical variation between different source areas is averaged out. Loess is absent from regions which were covered by glaciers in the last glacial period, nor does it occur in the humid tropics.

The eolian origin of loess has long been a subject of dispute; until today, some workers regard loess as an 'alluvial-lacustrine deposit', although most loess deposits lack the stratification that typifies alluvial deposits. Those who advocate the eolian origin of loess have the following arguments:

- loess occurs as a blanket over a wide range of surfaces, more or less independent of the topography;
- (2) loess blankets are thickest on the leeward sides of topographic obstacles;
- (3) there is absolutely no (cor)relation between the mineralogy of the loess blanket and that of the subsurface strata (which rules out the possibility of weathering in-situ);
- (4) the grain size distribution of loess which is typical of eolian materials transported in suspension;
- (5) grain sizes show a downwind fining gradient, away from the source;
- (6) loess deposits become thicker towards their presumed source;
- (7) fossil terrestrial snails have been found in loess deposits, and
- (8) loess deposition is still going on around the present desert areas.

Loess settles when dust-laden winds slow down to 7 (on dry surfaces) to 14 metres per second (on moist surfaces). The particular pore distribution of loess makes that it is quickly retained by capillary forces if it lands on a moist surface. The presence of a vegetation cover may also enhance the rate of loess deposition, and many authors maintain that the northern limit of loess deposition coincides with the northernmost extent of grass steppes during arid periods in the Pleistocene.

It has already been said that small-scale stratification is usually absent from loess due to the extreme homogeneity of its grain size. Laminated 'loessoid' deposits were normally redistributed by postdepositional sheet wash. Larger-scale layering, (deci)metres thick, is often indicative of a certain periodicity in loess deposition. At least two separate loess sequences were identified in the Netherlands, a Weichselian one, and an older, Saalian, sequence. The boundary between the two sequences is marked by a paleosol (a relic soil profile) which developed in the Saalian loess during the intervening interglacial, the Eemian.

The vast loess plains are now colonized by a climax vegetation of grasses and/or forest, and are the home of some of the best soils of the world: the 'black earths'.

Deep, black CHERNOZEMS occupy the central part of the Eurasian and North

214 Landforms in steppe regions

American steppe zones. Dusky red PHAEOZEMS are the soils of slightly more humid areas such as the American prairies and pampas. Brown KASTANOZEMS occur in the drier parts of the steppe zone and border arid and semi-arid lands, whereas GREYZEMS are prominent in the forest-steppe transition zone to more humid temperate regions.

KASTANOZEMS

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 gypsic* horizon or concentrations of soft powdery lime® within 125 cm of the surface, lacking a natric* B-horizon; lacking the characteristics which are diagnostic for Vertisols, Planosols or Andosols; lacking salic® properties; lacking gleyic® properties within 50 cm of the surface when no argic* B-horizon is present.

Key to Kastanozem (KS) Soil Units

Kastanozems having a gypsic* horizon.

Gypsic Kastanozems (KSj)

Other Kastanozems having an argic* B-horizon.

Luvic Kastanozems (KS1)

Other Kastanozems having a calcic* horizon.

Calcic Kastanozems (KSk)

Other Kastanozems.

Haplic Kastanozems (KSh)

SUMMARY DESCRIPTION OF KASTANOZEMS

Connotation: (dark) brown soils rich in organic matter; from L. castanea, chestnut and from R. zemlja, earth, land.

<u>Parent material</u>: a wide range of unconsolidated deposits; a large part of all Kastanozems have developed in loess.

<u>Environment</u>: dry and warm; flat to undulating grasslands with ephemeric grasses.

<u>Profile development</u>: mostly AhBC profiles with a brown Ah-horizon of medium depth over a brown to cinnamon cambic or argic B-horizon and with lime and/or gypsic accumulation in or below the B-horizon.

<u>Use</u>: many Kastanozem areas are in use for extensive grazing; the principal arable land use is the production of small grains and (irrigated) food and vegetable crops. Drought and (wind and water) erosion are serious limitations.

^{*} Diagnostic horizon; see Annex 1 for full definition.

a Diagnostic property; see Annex 1 for full definition.

REGIONAL DISTRIBUTION OF KASTANOZEMS

Kastanozems occupy some 400 million hectares worldwide. Major areas of occurrence are in the southern USSR, in the USA, Mexico, southern Brasil, and in the drier parts of the pampa regions of Uruguay and Argentina. Figure 1 shows the regional distribution of the Kastanozems.

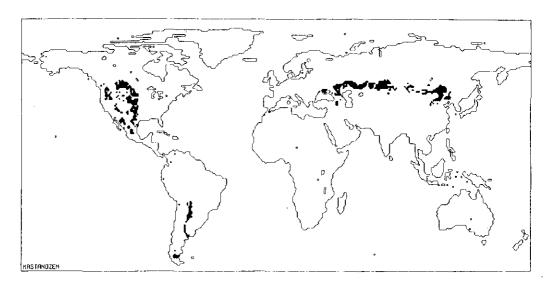


Fig. 1. Kastanozems worldwide.

GENESIS OF KASTANOZEMS

The genesis of Kastanozems is largely conditioned by the prevailing (climate-determined) type of vegetation. This is typically a short grass vegetation, scanty, poor in species and dominated by ephemers (early ripening grasses). The above-ground <u>dry</u> biomass amounts to only 0.8-1 tons/hectare, and the dry root mass to 3-4 tons/hectare.

More than 50 percent of all roots are concentrated in the upper 25 cm of the soil and there are few roots that extend deeper than 1 metre. The greater part of the vegetation dies each summer. In the equilibrium situation, the organic matter content of the Ah-horizon lies between 2 and 4 percent and seldom exceeds 5 percent.

Downward percolation in spring leaches solutes from the surface to the subsurface and subsoil layers. Lime accumulates at a depth of 90-100 cm, gypsum accumulation occurs commonly at a depth between 150 and 200 cm, and in the driest Kastanozems there may be a layer of salt accumulation deeper than 200 cm below the surface.

In places, a clay illuviation horizon is found as deep as 250-300 cm below the soil surface. The occurrence of argic B-horizons in Kastanozems

is still ill-understood. They may be fossil, as claimed by some Russian soil scientists, but there are also theories of a more recent formation, through 'normal' translocation of clay, or by destruction of clay or fine earth near the surface and reformation at greater depth.

Climatic gradients in the Kastanozem belt are reflected in pedogenic features. In the USSR, the darkest surface horizons occur in the north of the Kastanozem belt (bordering the Chernozems) whereas soils with shallower and lighter coloured horizons are more abundant in the south. There, effervescence with HCl starts already in the A-horizon and extends downward throughout the profile. The differentiation between horizons is clearer in the north than in the south of the Kastanozem belt which is a consequence of the decreasing length and intensity of soil formation with increasingly arid conditions.

CHARACTERISTICS OF KASTANOZEMS

The morphology of dark Kastanozems is not very different from that of the southern (drier) Chernozems whereas light Kastanozems (may) show a considerable resemblance with Calcisols. The northern Eurasian Kastanozem has an Ah-horizon of some 50 cm thick, dark brown and with a granular or fine blocky structure, grading into a cinnamon or pale yellow massive to coarse prismatic B-horizon. In the drier south, the Ah-horizon is only 25 cm thick and colours are lighter throughout the profile.

Reportedly, argic B-horizons have a "more intense coloration" in Luvic Kastanozems. Calcic and/or gypsic horizons, present in most Kastanozems, are particularly prominent in those of the southern dry steppes. Krotovinas occur in all Kastanozems but are less abundant than in Chernozems.

Hydrological characteristics

Kastanozems have an intermittent water regime. In the dry period, the soils dry out to great depth. In wet periods, the soils are often incompletely moistened, partly because of the low total precipitation sum and partly because the non-capillary porosity of Kastanozems is often rather low so that surface run-off losses during and after a heavy shower can be considerable. A 'dead dry horizon' occurs below the limit of wetting; this horizon receives neither percolation water from above nor capillary rise from below and is 'physiologically dead'.

Physical characteristics

The physical properties of Kastanozems are slightly less favourable than those of Chernozems but otherwise comparable. The lower humus content of the surface layer, particularly in the lighter Kastanozems, is associated with a lower degree of micro-aggregation, which translates into a lower total pore volume (40-55 percent), a lower moisture storage capacity, a denser packing of the soil and a lower permeability to water.

Chemical characteristics

Kastanozems are chemically rich soils with a cation exchange capacity of 25-30 cmol(+)/kg dry soil, and a base saturation percentage that is typically 95 percent or more. The majority of all adsorbed cations are Ca^{2+} and Mg^{2+} ; ESP-values of 4 to 20 have been reported.

The humous surface horizon contains some 0.1-0.2 percent nitrogen and 0.06-0.15 percent "phosphoric acid". The C/N-ratio of the organic soil fraction is around 10 as in Chernozems. The soil-pH is slightly above 7.0 but may increase to a value around 8.5 at some depth below the surface. Lime and gypsum accumulation are common; lime contents are some 10 to 20 percent higher in the accumulation horizon than in the deeper solum. More easily soluble salts may accumulate as well, deeper in the dark Kastanozems than in the lighter soils of the drier steppe. The salt content of the accumulation layer is commonly between 0.05 and 0.1 percent and doesnot seriously inhibit the growth of crops but salt levels may build up to 0.4 percent and higher.

MANAGEMENT AND USE OF KASTANOZEMS

Kastanozems are potentially rich soils; periodic lack of soil moisture is the main obstacle to high yields. Irrigation is nearly always necessary whereby care should be taken not to introduce secondary salinization of the surface soil. Small grains and (irrigated) food and vegetable crops are the principal commodities.

Extensive grazing is another important land use on Kastanozems but the sparsely vegetated grazing lands are inferior to the tall grass steppes with Chernozems.

CHERNOZEMS

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* horizon or concentrations of soft powdery lime® within 125 cm of the surface, or both; lacking a natric* B-horizon; lacking the characteristics which are diagnostic for Vertisols, Planosols or Andosols; lacking salic® properties; lacking gleyic® properties within 50 cm of the surface if no argic* B-horizon is present; lacking uncoated silt and quartz grains on structural ped surfaces.

Key to Chernozem (CH) Soil Units

Chernozems having an argic* B-horizon and showing gleyica properties within 100 cm of the surface.

Gleyic Chernozems (CHg)

Other Chernozems having an argic * B-horizon.

Luvic Chernozems (CH1)

Other Chernozems showing tonguing $^{\mathbf{a}}$ of the A-horizon into a cambic * B-horizon or into a C-horizon.

Glossic Chernozems (CHw)

Other Chernozems having a calcic* horizon.

Calcic Chernozems (CHk)

Other Chernozems.

Haplic Chernozems (CHh)

SUMMARY DESCRIPTION OF CHERNOZEMS

<u>Connotation</u>: black soils rich in organic matter; from R. <u>chern</u>, black, and <u>zemlja</u>, earth or land.

Parent material: mostly eolian sediments (loess).

<u>Environment</u>: flat to undulating plains with a vegetation of tall grasses (forest in the northern transitional zone) in regions with a continental climate with cold winters and hot summers.

<u>Profile development</u>: AhBC profiles with a dark brown to black mollic A-horizon over a cambic or argic B-horizon; commonly with redistribution of calcium carbonate to a calcic horizon or pockets of soft powdery lime in the subsurface soil.

^{*} Diagnostic horizon; see Annex 1 for full definition.

Diagnostic property; see Annex 2 for full definition.

<u>Use</u>: the high natural fertility of Chernozems and their favourable topography permit a wide range of agricultural uses including arable cropping (with supplemental irrigation in dry summers) and cattle ranging.

REGIONAL DISTRIBUTION OF CHERNOZEMS

Chernozems cover an estimated 300 million hectares worldwide, mainly in the middle latitude steppes of Eurasia and North America, north of a zone with Kastanozems. Figure 1 presents an overview of their main areas of occurrence.

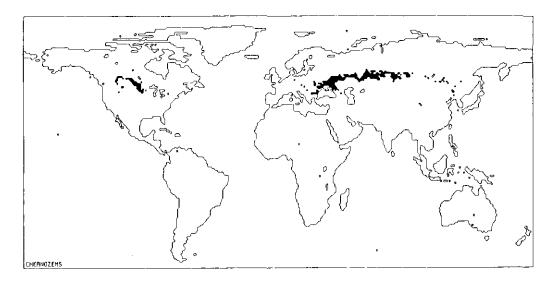


Fig. 1. Chernozems worldwide.

GENESIS OF CHERNOZEMS

Chernozems form in uniformly textured silty parent material (loess), under a tall grass vegetation with a vigorous development over a lengthy vegetative period. The above ground biomass amounts to some 1 to 1.5 tons of dry matter per hectare; the corresponding root mass, already incorporated in the soil, weighs 4 to 6 tons/hectare. The main concentration of roots is in the upper 60 cm of the soil, with 80 percent of all roots concentrated in the top 30 to 40 cm. See Figure 2.

Deep, humus-rich Chernozems occur in the <u>central</u> part of the steppe zone where the annual precipitation sum is approximately equal to the evaporation sum. Such Chernozems contain 10 to 16 percent organic matter, are neutral in reaction (pH 7.0, and about 7.5 in the subsoil), and highly

saturated with bases. Earthworms are very active there, in wet periods predominantly in the upper 50 cm layer but they move to deeper strata at the onset of the dry period. A considerable part of the surface soil consists of worm casts, a stable mixture of mineral and organic soil material. Further homogenization of the soil is brought about by burrowing small vertebrates, so that the Ah-horizon may extend to a depth of 2 meters.

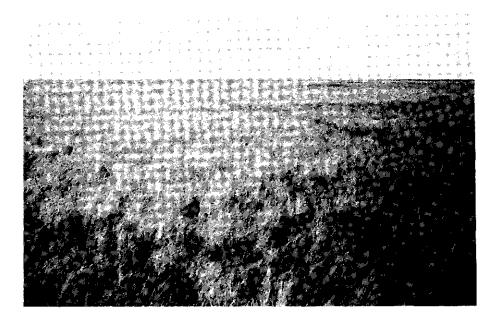


Fig. 2. Tall grass steppe in the Chernozem belt (Tselinograd, USSR). Photo by courtesy of ISRIC, Wageningen.

Deep percolation during wet spells has removed virtually all easily soluble salt from the profile; the salt content is seldom higher than 0.1 percent within the upper 4 meters of soil. At about the same depth, deeper in the north of the Chernozem belt and closer to the surface in the south, lies a 'dead dry horizon' that receives neither percolation water from above nor capillary rise from below. The depth at which it occurs and its thickness vary; the dead dry horizon is not continuous and may even be absent altogether.

At a depth of 2 to 3 meters (1.5 to 2.5 m in southern Chernozems), there may be some accumulation of gypsum and shallower still, say at a depth of about 1 metre from the surface, accumulation of lime. Migration of clay has resulted in a faint clay bulge at 50 to 200 cm depth in some Chernozems. These phenomena indicate that the central Chernozems are exposed to moderately strong leaching during wet periods.

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Towards the northern fringe of the Chernozem belt, the surface horizon with organic matter accumulation becomes shallower and more greyish in colour until signs of podsolization such as an eluvial E-horizon and/or horizontal lamellae begin to occur. The horizon with carbonate accumulation is often separated from the humic surface layer by a carbonate-free layer of appreciable thickness. The absence of a horizon with readily soluble salts, the slight acidity (pH 6-6.5) of the surface soil, and the presence of greyish colours are further indications that the northern Chernozems are subject to stronger leaching than those of the central steppe zone.

Towards the southern fringe of the steppe zone, the water regime becomes more and more intermittent, with increasingly longer dry periods. As a consequence, plants with a long vegetative period disappear and xerophytes and early ripening grasses move in. Moreover, the soil's humus undergoes increasingly intense mineralization and there is an increase in the content of readily soluble salts in the surface soil. In places, the soil's exchange complex may become saturated with sodium ions to the extent that the soils develop 'sodic properties' (See Annex 2) and soil structures collapse, which leads to a denser packing and a lower hydraulic conductivity than common in the soils of the central steppe zone.

The colour of the surface soil has diagnostic value: when the chroma of the (mollic) A-horizons has become higher than 2, this is seen as a sign that aridity is so severe that the soils are no longer true Chernozems; they are then classified as Kastanozems.

CHARACTERISTICS OF CHERNOZEMS

Most virgin Chernozems have a thin leafy litter layer on top of the dark grey to black A-horizon which is 'vermicular' (i.e. 'full of worm casts'). The surface horizon is normally between 50 and 80 cm thick but may extend to a depth of 2 m and more in well developed Chernozems. Worm casts and knotovinas testify of intense faunal activity.

In some instances, the Ah-horizon penetrates the B-horizon with tongues that are deeper than wide. Calcium carbonate accumulation may show up as pseudomycelium in the lower part of the surface soil and/or as lime nodules in a brownish grey to light cinnamon subsoil. The B-horizon has a blocky to prismatic structure.

Luvic Chernozems, having an argic B-horizon, are not uncommon in the north of the steppe zone where the grass vegetation grades into deciduous forest and the Chernozems grade into Greyzems or Luvisols. Chernozems in wet areas may develop gleyic properties; they are classified as Gleyic Chernozems only if they display clear signs of hydromorphy within one meter from the soil surface and have an argic B-horizon. Soils with a mollic A-horizon and with gleyic properties already in the upper 50 cm layer but not having an argic B-horizon key out as (Mollic) Gleysols.

Mineralogical characteristics

The mineral and mechanical properties of Chernozems are rather homogeneous, in line with the uniform composition of the parent material. Sesqui-

oxides and silica are uniformly distributed over the profile (no migration); the SiO_2/R_2O -ratio is high, about 2.0.

Hydrological characteristics

Although it is widely accepted that Chernozems formed under conditions of good drainage, there are also (Russian) soil scientists who maintain that certain Chernozems passed through a boggy phase of soil formation. At present, Chernozems are rarely water-saturated, apart from soils in (small) depressions with occasional shallow groundwater. There is approximate equity between the annual precipitation sum and evaporation, with a slight precipitation surplus in the north of the steppe zone and a slight deficit in the south. Table 1 presents an overview of the occurrence of Eurasian Chernozems as a function of the annual precipitation sum and the type of vegetation.

TABLE 1.
Typical Major Soil Groupings and Soil Units in the Eurasian steppe zone.

TEMPERATURE	PRECIPITATION	VEGETATION TYPE	SOIL GROUPING/UNIT
	> 550 mm	deciduous forest	Luvisols, Greyzems
	500 mm	steppe and forest	Luvic Chernozem
	500 mm	tall grass steppe	Haplic Chernozem
increase	450 mm	tall grass steppe	Calcic Chernozem
	200-400 mm	medium height grass	Kastanozems
l V	< 200 mm	steppe open vegetation	Calcisols

Physical characteristics

Chernozems possess favourable physical properties. Their total pore volume lies normally between 55 and 60 percent in the Ah-horizon and between 45 and 55 percent in the C-horizon. The soils have a high moisture holding capacity; reported soil moisture contents of some 33 percent at field capacity and 13 percent at permanent wilting point suggest an 'available moisture' content around 20 volume percent. The stable micro-aggregate structure of the humus-rich Ah-horizons (Table 2) ensures a favourable combination of capillary and non-capillary porosity and makes these soils suitable for irrigated farming.

TABLE 2.

Mechanical and micro-aggregate composition of the upper horizon of an ordinary (central) Chernozem. Source: N.A. Kachinsky (Plyusnin, 196?).

FRACTION (mm):	1-0.25	0.25-0.05	0.05-0.01	0.01-0.001	<0.001
after dispersion no dispersion:	: -	1.8 35.5	35.2 45.3	27.6 16.7	35.4 2.5

Chemical characteristics

Chernozem surface soils contain 5 to 15 percent 'mild' humus with a high proportion of humic acids and a C/N-ratio that is typically around 10. The surface horizon is neutral in reaction (pH 6.5-7.5) but the pH may reach a value of 7.5-8.5 in the subsoil, particularly where there is accumulation of lime. Chernozems have a good natural fertility status; the surface soil contains 0.2-0.5 percent nitrogen and 0.1 to >0.2 percent phosphorus. This phosphorus is only partly 'available'; crops on Chernozems tend to respond favourably to P-fertilizers. In southern Chernozems, the humus contents are lower (4-5 percent) and consequently also the cation exchange capacity: 20-35 cmol(+)/kg dry soil, versus 40-55 cmol(+) per kg in central Chernozems. Normally, the base saturation percentage lies close to 95 percent with Ca²⁺ and Mg²⁺ as the main adsorbed cations, but sodium adsorption may be high in southern Chernozems.

MANAGEMENT AND USE OF CHERNOZEMS

Russian soil scientists value the (deep, central) Chernozems among the best soils of the world. With less than half of all Chernozems in the USSR being used for arable cropping, these soils constitute a formidable resource for the future. See Figure 3.



Fig. 3. Arable farming on Chernozems in the USSR. Photo by courtesy of ISRIC, Wageningen.

Wheat, barley and maize are the principal crops grown, besides other food crops and vegetables; part of the Chernozem area is used for livestock rearing.

Summer drought makes moisture conservation imperative. Preservation of the favourable soil structure through timely cultivation and careful irrigation at low watering rates prevents ablation and erosion. Application of P-fertilizers is commonly required.

NOTES

PHAEOZEMS

Soils having a mollic* A-horizon; lacking a calcic* horizon, a gypsic* horizon, concentrations of soft powdery lime^a; having a base saturation (by lM NH₄OAc at pH 7.0) which is 50 percent or more throughout within 125 cm of the surface; lacking a ferralic* B-horizon; lacking a natric* B-horizon; lacking the characteristics which are diagnostic for Vertisols, Nitisols, Planosols or Andosols; lacking salic^a properties; lacking gleyic^a properties within 50 cm of the surface if no argic* B-horizon is present; lacking uncoated silt and quartz grains on structural ped surfaces if the mollic* A-horizon has a moist chroma of 2 or less to a depth of at least 15 cm.

Key to Phaeozem (PH) Soil Units

Phaeozems showing gleyic^a properties within 100 cm of the surface.

Gleyic Phaeozems (PHg)

Other Phaeozems showing stagnic properties within 50 cm of the surface.

Stagnic Phaeozems (PHs)

Other Phaeozems having an argic* B-horizon.

Luvic Phaeozems (PH1)

Other Phaeozems that are calcareous $^{\mathbf{a}}$ at least between 20 and 50 cm from the surface.

Calcaric Phaeozems (PHc)

Other Phaeozems.

Haplic Phaeozems (PHh)

SUMMARY DESCRIPTION OF PHAEOZEMS

Connotation: dark soils rich in organic matter; from Gr. phaios, dusky and R. zemlja, earth, land.

<u>Parent material</u>: eolian (loess) and other unconsolidated, predominantly basic materials.

<u>Environment</u>: flat to undulating land in warm to cool (e.g. tropical highland) regions, humid enough that there is, in most years, some percolation of the soil, but also with periods in which the soil dries out. The natural vegetation is tall (prairie) grass and/or forest.

^{*} Diagnostic horizon; see Annex 1 for full definition.

a Diagnostic property; see Annex 2 for full definition.

228 Phaeozems

<u>Profile development</u>: mostly AhBC profiles with a mollic A-horizon over a cambic or argic B-horizon.

<u>Use</u>: Phaeozems are fertile soils; they are grown to cereals and pulses or are used as grazing land. Periodic drought, wind and water erosion are the main limitations.

REGIONAL DISTRIBUTION OF PHAEOZEMS

Phaeozems cover an estimated 100 million hectares worldwide, mainly in the North American Prairie region, the pampas of Argentina and Uruguay and the subtropical steppe of eastern Asia. Smaller areas occur in mediterranean climates and in montane areas in the tropics, e.g. in eastern Africa. Figure 1 presents the major Phaeozem areas.

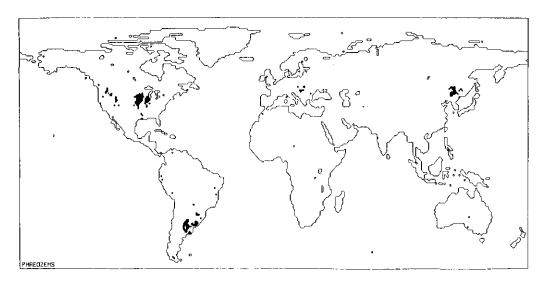


Fig. 1. Phaeozems worldwide.

GENESIS OF PHAEOZEMS

Phaeozems occur in more humid environments than Chernozems or Kastanozems and on fine-textured basic parent material. Consequently, the biomass production is high but also the rates of weathering and leaching. Calcium carbonate is leached out of the soil profile but leaching is not so intense that the soils have become depleted of bases. Besides, earthworms and burrowing mammals homogenize the soil and bring carbonates from below (back) to the leached surface soil. In places, the faunal activity is so intense that the mollic A-horizon is thickened and worm holes and krotovinas extend into the C-horizon.

Argic B-horizons do occur in Phaeozems, but they are widely regarded as relics from an earlier development towards Luvisols (in eras with a more humid climate).

CHARACTERISTICS OF PHAEOZEMS

Phaeozems have normally a brown to grey mollic A-horizon of 30-50 cm thickness over a brown cambic B-horizon or a yellowish brown C-horizon, or over a brown or reddish brown argic B-horizon. The A-horizons of Phaeozems are thinner than those of Chernozems and lighter in colour. Where groundwater is at shallow depth or temporarily a perched water table occurs, e.g. on top of an argic B-horizon, the surface soil may be mottled and/or dark. Luvic soil units, polygenetic or not, represent a more advanced stage of soil formation and have often more reddish colours than other Phaeozems; they include the former 'Reddish Prairie Soils'.

Hydrological characteristics

In spite of their generally good water storage capacity, many Phaeozems are short of water in the dry season.

Physical characteristics

Phaeozems are porous, well aerated soils with moderate to strong, very stable, crumb to blocky structures. Where clay illuviation occurs, the illuviation layer contains commonly 10-20 percent more clay than the overlying horizon.

Chemical characteristics

The organic matter content of the topsoil is typically around 5 percent; the C/N-ratio of the organic matter is 10-12. pH-values are between 5 and 7 and increase towards the C-horizon. The cation exchange capacity of Phaeozems is 25-30 cmol(+)/kg dry soil or somewhat lower; the base saturation percentage lies between 65 and 100 percent, with the highest values in the deeper subsoil.

MANAGEMENT AND USE OF PHAEOZEMS

Phaeozems are fertile soils and make excellent farm land. Luvic Phaeozems are said to be somewhat 'better' than Haplic Phaeozems because of their higher water holding capacity. In the USA and Argentina, Phaeozems are widely grown to wheat (and other small grains) and, in the USA, to soybean. Wind and water erosion are serious hazards. Large areas of Phaeozems are in use for cattle rearing/grazing.

NOTES

GREYZEMS

Soils having a mollic* A-horizon with a moist chroma of 2 or less to a depth of at least 15 cm and showing uncoated silt and quartz grains on structural ped surfaces; having an argic* B-horizon; lacking the characteristics which are diagnostic for Planosols.

Key to Greyzem (CR) Soil Units

Greyzems showing gleyic properties within 100 cm of the surface.

Gleyic Greyzems (GRg)

Other Greyzems.

Haplic Greyzems (GRh)

Diagnostic horizon; see Annex 1 for full definition.
Diagnostic property; see Annex 2 for full definition.

SUMMARY DESCRIPTION OF GREYZEMS

<u>Connotation</u>: dark soils with a grey tinge due to the presence of bleached (uncoated) quartz sand and silt in layers rich in organic matter, from E. grey and R. <u>zemlja</u>, land.

<u>Parent material</u>: decalcified unconsolidated materials including eolian, fluvial and lacustrine deposits, solifluction material and glacial till.

<u>Environment</u>: flat to gently undulating plains in mild temperate to cold, subhumid climates. Only in continental summer periods will the surface soil dry out. In montane areas, Greyzems occur at altitudes corresponding with the change of steppe to forest (vertical zonality).

<u>Profile development</u>: mostly AhBtC profiles with a mollic A-horizon over an argic B-horizon over unaltered parent material.

<u>Use</u>: Greyzems produce fairly good crops of cereals, sugar beets, peas, potatoes and fodder crops in the USSR and of small grains, oil seeds, forage and timber in Canada. In places, gleying and unfavourable climatic conditions are serious limitations.

REGIONAL DISTRIBUTION OF GREYZEMS

Greyzems cover an estimated 28 million hectares in the northern hemisphere where they occupy a (zonal) position between Luvic Chernozems and Luvisols at the transition from tall grass steppes to lands with deciduous mixed forests. Major areas of occurrence are indicated in Figure 1.

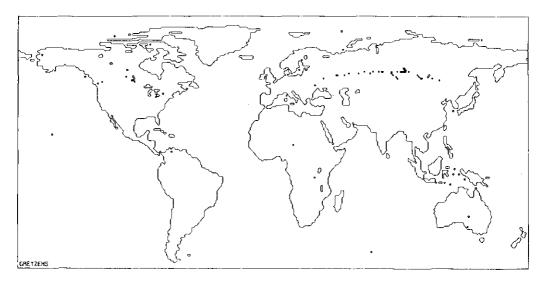


Fig. 1. Greyzems worldwide.

GENESIS OF GREYZEMS

Greyzems hold an intermediate position between Chernozems and Luvisols: they have the mollic A-horizon of the Chernozems and the argic B-horizon of the Luvisols.

It is unclear whether the argic B-horizon of Greyzems is older than the mollic A-horizon or, alternatively, degradation of the mollic A-horizon initiated the formation of the argic B-horizon.

If the first hypothesis holds, Greyzems are to be seen as Luvisols that have acquired a mollic A-horizon after the climate became wetter sometime in the Holocene. The steppe vegetation changed gradually into forest and the mollic A-horizon degraded under the increasing acidification of the forest floor.

In the second hypothesis, it also is acknowledged that the climatic conditions under which the mollic A-horizon developed do no longer exist and that the present forest invaded former grasslands. However, it is postulated that the formation of a litter layer and acidification of the surface soil under forest caused the degradation of clay-humus complexes in the mollic A-horizon, associated with bleaching of sand and/or silt and formation of an argic B-horizon upon translocation of destabilized fine material with percolating rain water.

Whatever the sequence of events may have been, there is clearly ongoing destruction of humus and degradation of the surface soils today. It is likely that the parent material of most Greyzems was calcareous once and that Greyzems result from a long continued process of decalcification and gradual loss of adsorbed bases.

CHARACTERISTICS OF GREYZEMS

The most striking characteristic of Greyzems is the presence of uncoated sand and silt particles in a mollic A-horizon. The bleached grains can be seen with a lens or even with the naked eye. They may appear in more or less horizontal bands associated with a weakly platy structure, but they occur also in spots ('pepper and salt' appearance). In Canada, Greyzems with a distinct, albeit thin, albic E-horizon are seen as intergrades to Albic Luvisols which lack the mollic A-horizon.

Hydrological characteristics

Most Greyzems are well drained, have a good moisture storage capacity and suffer from drought in the summer season only in positions near the drier end of their zone (close to the Chernozems). Greyzems in northern regions may suffer from wetness when the soils thaw in spring and the still frozen subsoil doesnot permit adequate discharge of water.

Physical characteristics

Properly managed Greyzems retain a reasonably stable cloddy to granular aggregation in the cultivated layer. Yet, the tendency to deteriorate in wet conditions and to form a surface crust when dry makes them susceptible to erosion in sloping terrain. The argic B-horizon may hinder root penetration if it is dense, but the texture change is normally rather gradual. On the whole, Greyzems are reasonably productive soils with a somewhat increased production risk.

Chemical characteristics

Greyzems have chemical characteristics similar to those of Chernozems: the organic carbon content of the surface soil lies between 3 and 5 percent, with a C/N-ratio around 10. They have favourably high CEC values (25 to 35 cmol(+)/kg dry soil) and a base saturation close to 100 percent, with ${\rm Ca}^{2+}$ as the dominant ion. The natural fertility status of Greyzems is good; their potassium level does not need correction, but they respond favourably to moderate applications of nitrogen and phosphorus fertilizers.

MANAGEMENT AND USE OF GREYZEMS

Farmers on Greyzems must minimize structure deterioration in the cultivated layer, and be aware of the tendency of these soils to puddle in (too) wet condition and to cake and crust in a subsequent dry spell.

234 Greyzems

Greyzems run a somewhat increased erosion risk in sloping terrain. Forestry and the growing of fodder crops minimize soil disturbance and structure deterioration. If arable crops (cereals, sugar beets, potatoes) are grown, tillage operations should be performed under favourable moisture conditions. The poor drainage of the (mostly eastern Siberian) Gleyic Greyzems limit the suitability of these soils.

MINERAL SOILS CONDITIONED BY A (SUB)HUMID TEMPERATE CLIMATE:

LUVISOLS PODZOLUVISOLS PLANOSOLS PODZOLS

MAJOR LANDFORMS IN TEMPERATE REGIONS

Most of the (present) temperate regions were covered with continental ice sheets when the Ice Ages had their maximum expanse, and massive glacial and fluvioglacial deposits were laid down in these regions when the ice melted. See Figure 1.

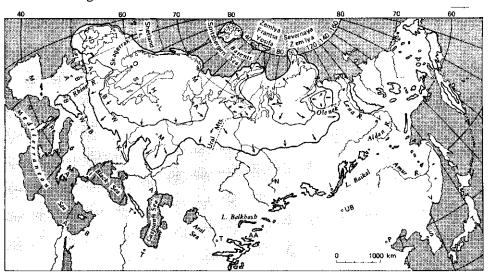


Fig. 1. Sketch map of Europe and northern Asia, showing supposed extent of maximum glaciated area in Pleistocene Ice Ages. Source: Flint, 1971.

Periglacial areas, adjacent to the ice-covered regions, are marked by sediments with characteristic deformation structures, incurred in repeated freezing and thawing. Strong winds blew sand and silt out of the frozen river plains; this material settled again as cover sands and dunes, and, at greater distance from the source, as loess blankets.

Virtually all landforms in the temperate zone show, in one way or another, evidence of the great variation in climate conditions during the past 100,000 years. But they have also a number of common characteristics which were brought about by their present, cool and (sub)humid climate:

- (most) rivers have a regular regime;
- meanders are the normal channel form in alluvial flood plains;
- rivers tend to incise rather than to aggradate;
- soil formation and weathering predominate over surface wash in sloping terrain as long as the natural vegetation cover remains intact.

Within the temperate zone, three broad morphotectonic categories can be distinguished:

- Pleistocene sedimentary lowlands with glacial, fluvioglacial, fluvioperiglacial and eolian deposits;
- (2) <u>uplifted and dissected sedimentary basins</u>, in places with (Mesozoic) limestones, or with sandstones, mudstones, and/or a loess blanket;
- (3) uplifted and dissected Caledonian and Hercynian Massifs, partly consisting of folded sedimentary and low-grade metamorphic rocks and partly of crystalline rocks.

For a discussion of the landforms in the third category, the reader is referred to the paragraph on the morphology of Middle Mountains.

LANDFORMS IN (PERI)GLACIAL AND EOLIAN SEDIMENTARY LOWLANDS

'Normal' fluvial and marine lowlands have been discussed in a preceding chapter; the following paragraphs will focus on areas that are underlain by glacial, fluvioglacial, periglacial or eolian deposits. Such areas are particularly extensive in the temperate regions of the northern hemisphere where the continental ice sheets had the greatest expanse during the Pleistocene glacial periods. The southern hemisphere simply lacked sufficient land at high latitude for the development of extensive ice sheets.

Reconstructions of the Pleistocene climate changes indicate that there were at least four advances of continental ice in Europe and on the North American continent. Each of these took place in a different glacial period. Three major advances could be identified in north-west Europe; they are known as the Elster, Saale and Weichsel glacial maxima; still older ones have been recognized elsewhere, e.g. in the Soviet Union.

Glacial advance and retreat conditioned some typical landforms. The commonest are:

- (1) till plains, e.g. the Drente Plateau in the Netherlands;
- (2) moraine complexes, e.g. the Salpausselkä in Finland and the Ra moraine in Norway;
- (3) ice-pushed ridges, e.g. the 'stuwwallen' in the Netherlands;
- (4) tongue basins such as the 'Gelderse Vallei' near Wageningen, and
- (5) outwash plains around terminal moraines or ice-pushed ridges.

Regions adjoining the actual ice-covered areas were perennially frozen ('permafrost'), so that hardly any vegetation could develop; the presence of ice caps was conducive to the occurrence of stronger winds than at present. Large areas became covered with eolian deposits; desert pavements, sand plains and dunes occur at short distances from the former ice front, and loess deposits farther away. The mechanism of dune formation was discussed in the chapter on residual and shifting sands; loess deposits were treated in the chapter on steppes and steppic regions.

LANDFORMS IN UPLIFTED SEDIMENTARY BASINS

Uplifted sedimentary basins formed in western Europe after the Hercynian orogeny. Their sediments, mostly shallow-water limestones, marls and calcareous sandstones, date back to Mesozoic transgressions. They were never folded, but differential subsidence and later uplift have in places resulted in tilting. Cuesta landscapes are a common feature; remnants of Tertiary soils on top of cuesta dipslopes, e.g. the 'clay-with-flint' on Cretaceous chalk ('limons à silex' in France; 'kleefaarde' or 'vuursteen-eluvium' in the Netherlands), indicate that these basins formed part of extensive peneplains in the Tertiary. Uplift as a result of orogeny in the nearby Alps caused the formation of river terraces and incised meanders; the 'Ile de Paris' is an example.

Many of these basins became partially covered with glacial or periglacial deposits. Moraines and fluvioglacial deposits are widespread on top of the Easteuropean Platform; parts of the Paris basin are covered with loess. Only where such covers are absent, e.g. on cuesta slopes that were too high or too steep to collect a thick loess blanket, did soils form in the original parent material.

PODZOLS are strongly represented in fluvioglacial and eolian sands; parabolic dunes of only 2-3 metres height contain fine podzolic catenas which reflect differences in groundwater depth.

LUVISOLS are among the commonest soils in the loess blankets in temperate regions; they grade into Chernozems towards the drier end of the zone. Luvisols occur also in (less rigidly sorted) fluvial deposits which illustrates that the occurrence of Podzols and Luvisols is not just a matter of climate, but also of parent material and consequently of Quaternary geological history. PODZOLUVISOLS developed in clayey glacial till and fine-textured materials of fluvioglacial or glaciolacustrine origin, but also in loess, in regions with cold winters and short, cool summers.

PLANOSOLS occur predominantly in subhumid and semi-arid regions on the southern hemisphere. In some instances they formed through degradation of Podzoluvisols.

NOTES

LUVISOLS

Soils having an argic B-horizon which has a cation exchange capacity equal to or more than 24 cmol(+)/kg clay and a base saturation (by 1M NH₄OAc at pH 7.0) of 50 percent or more throughout the B-horizon; lacking a mollic A-horizon; lacking the E-horizon abruptly overlying a slowly permeable horizon, the distribution pattern of the clay and the tonguing which are diagnostic for Planosols, Nitisols and Podzoluvisols respectively.

Key to Luvisol (LV) Soil Units

Luvisols showing gleyic^a properties within 100 cm of the surface.

Gleyic Luvisols (LVg)

Other Luvisols showing stagnic^a properties within 50 cm of the surface.

Stagnic Luvisols (LVs)

Other Luvisols having an albic E-horizon.

Albic Luvisols (LVa)

Other Luvisols showing vertic^a properties.

Vertic Luvisols (LVv)

Other Luvisols having a calcic * horizon or concentrations of soft powdery lime a within 125 cm of the surface.

Calcic Luvisols (LVk)

Other Luvisols showing ferric^a properties.

Ferric Luvisols (LVf)

Other Luvisols having a strong brown to red B-horizon (rubbed soil has a hue of 7.5YR and a chroma of more than 4, or a hue redder than 7.5YR).

Chromic Luvisols (LVx)

Other Luvisols,

Haplic Luvisols (LVh)

SUMMARY DESCRIPTION OF LUVISOLS

<u>Connotation</u>: soils in which clay is washed down from the surface soil to an accumulation horizon at some depth; from L. <u>luere</u>, to wash.

<u>Parent material</u>: a wide variety of unconsolidated materials including glacial till, and eolian, alluvial and colluvial deposits.

^{*} Diagnostic horizon; see Annex 1 for full definition.

a Diagnostic property; see Annex 2 for full definition.

242 Luvisols

<u>Environment</u>: most common in flat or gently sloping land in cool temperate regions and in warm (e.g. mediterranean) regions with distinct dry and wet seasons. Deciduous forest, coniferous forest and grasslands.

<u>Profile development</u>: ABtC profiles; intergrades to Podzoluvisols having an albic E-horizon between the A-horizon and the argic B-horizon are not rare. The considerable variation among Luvisols is due to variations in parent material (e.g. lime content) and environmental conditions (wetness, erosion).

<u>Use</u>: because of their moderate stage of weathering and high base saturation, Luvisols with a good drainage status are suitable for a wide range of agricultural uses.

REGIONAL DISTRIBUTION OF LUVISOLS

Luvisols cover some 500 to 600 million hectares worldwide. They are for the greater part located in temperate regions such as the west-central USSR, the USA and central Europe, but they occur also in warmer environments, e.g. the Mediterranean region and southern Australia. Figure 1 gives an indication of the major concentrations of Luvisols.

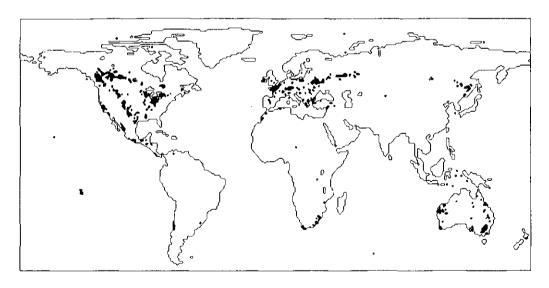


Fig. 1. Luvisols worldwide.

GENESIS OF LUVISOLS

The dominant characteristic of Luvisols is their argic B-horizon, formed by translocation of clay from the surface soil to the depth of illuviation. The process knows three essential phases:

- (1) mobilization of clay in the surface soil.
- (2) transport of clay to the accumulation horizon.
- (3) immobilization of transported clay.

Normally, clay is not present in a soil as individual particles but is contained in aggregates that consist wholly of clay or of a mixture of clay and other mineral and/or organic soil material. Mass transport of soil material along cracks and pores, common in cracking soils in regions with alternating wet and dry periods, does not necessarily enrich the subsoil horizons with clay. For an argic B-horizon to form, the (coagulated) clay must disperse in the horizon of eluviation before it can be transported to the depth of accumulation by percolating water.

Mobilization of clay can only take place if the thickness of the electric 'double layer', i.e. the shell around the individual clay particles that is influenced by the negatively charged sides of the clay plates, is increased. If the double layers are wide, the bonds between the negatively charged sides and the positive charges at the edges of the clay particles can no longer hold individual clay particles together in aggregates. The strength of aggregation is determined by:

- (1) the ionic strength of the soil solution,
- (2) the composition of the ions adsorbed at the exchange complex, and
- (3) the specific charge characteristics of the clay in the soil.

At high electrolyte concentrations in the soil solution, the double layer is compressed regardless of the nature of the adsorbed ions so that the clays remain flocculated. A decrease in ion concentration, e.g. as a result of dilution by rainfall, can result in dispersion of the clay and collapse of the aggregates. If the exchange complex is dominated by polyvalent ions, the double layer may remain narrow even at low electrolyte concentrations and consequently the aggregates may remain intact. Because concentrations of the various ions in solution and the charge characteristics of the clay are normally influenced by the pH of the soil, dispersion of clays is, to some extent, a pH-dependent process.

At pH(H₂0, 1:1) values below 5, the aluminium concentration of the soil solution is normally sufficiently high to keep the clay flocculated (Al³⁺ is preferentially adsorbed over divalent and monovalent ions in the soil solution). Between pH 5.5 and 7, exchangeable aluminium contents are negligible and the concentrations of divalent ions low, so that the clay can disperse. At still higher pH values, divalent bases will normally keep the clay flocculated unless there is a strong dominance of Na⁴-ions in the soil solution. If the pH exceeds 9, calcium and magnesium ions precipitate from the solution as carbonates. They are replenished from the exchange complex which becomes increasingly occupied by monovalent ions.

244 Luvisols

Certain organic compounds, especially polyphenols, stimulate the mobilization of clay by neutralizing the positive charges at the edges of the clay minerals. As iron-saturated organic complexes are insoluble, this process might be of little importance in Fe-rich Luvisols in the subtropics.

<u>Transport</u> of peptized clay particles requires percolation through wide (>20 um) pores and voids. Clay translocation is particularly prominent in soils which crack during the dry season but become wet during occasional downpours.

Smectite clays disperse more easily than non-swelling clays; they are a common constituent of Luvisols.

<u>Precipitation</u> of clay particles takes place at some depth in the soil as a result of flocculation or as a result of (mechanical) filtration of the clay suspension by fine capillary pores.

Flocculation can be initiated by an increase in the electrolyte concentration of the soil solution or by an increase in the content of divalent cations (e.g. in a $CaCO_3$ -rich subsurface horizon).

Filtration occurs where a clay suspension percolates through a relatively dry soil mass; it forces the clay plates against the faces of peds or against the walls of (bio)pores where skins of strongly oriented clay ('cutans') are formed (see Figure 2). With time, the cutans may wholly or partly disappear through homogenization of the soil by the soil fauna, or the cutans may be destroyed mechanically in soils with a high content of swelling clays. This explains why there is often less oriented clay in the argic B-horizon than one would expect on the basis of a budget analysis of the clay profile. There could also be more illuviated clay than expected, viz. if (part of) the eluviated surface soil is lost through erosion.

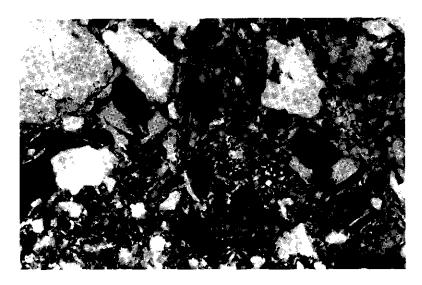


Fig. 2. Cutans ('channel ferri-argillans') in the argic B-horizon of a Luvisol in a Late Weichselian alluvial deposit in The Netherlands (cross polarized light; 1 cm = $100~\mu m$). Source: Miedema, 1987.

CHARACTERISTICS OF LUVISOLS

Luvisols have a brown to dark brown A-horizon over a (greyish) brown to strong brown or red Bt-horizon. In subtropical Luvisols in particular, a calcic horizon may be present or pockets of soft powdery lime occur in and below the reddish brown Bt-horizon. Soil colours are less reddish in Luvisols in cool regions than in warmer climates. In wet environments, the surface soil may become depleted of clay and free iron oxides to the extent that a greyish eluviation horizon forms under a dark but shallow A-horizon.

Mineralogical characteristics

Towards the dry end of their zone, Luvisols have increasing contents of swelling and shrinking clays and pressure faces and parallelpiped structure elements become more and more prominent. Luvisols are moderately weathered soils; they contain less Al-, Fe- and Ti-oxides than their tropical counterparts, the Lixisols, and have an $\mathrm{SiO}_2/\mathrm{Al}_2\mathrm{O}_3$ ratio of more than 2.0.

Hydrological characteristics

Most Luvisols are well drained but shallow groundwater may induce gleyic soil properties in and below the Bt-horizon in Luvisols in depression areas. Stagnic properties develop where a dense argic B-horizon obstructs downward percolation and the surface soil becomes saturated with water for extended periods of time.

Physical characteristics

Luvisols have favourable physical properties; they have granular or crumb surface soils that are porous and well aerated. The 'available' moisture storage capacity is around 15 to 20 volume percent. The illuviation horizon has stable blocky structures. Luvisols with a high silt content may be sensitive to erosion.

Chemical characteristics

Although the chemical properties of Luvisols vary with the parent material and pedogenetic history, these soils are generally fertile and have high mineral reserves. The (decalcified) surface soils contain a few percent organic matter with a C/N ratio of 10 to 15 and are slightly acid in reaction; the subsurface soils have a neutral reaction and may contain some lime.

MANAGEMENT AND USE OF LUVISOLS

With the possible exception of Albic soil units, Luvisols are fertile soils suitable for a wide range of agricultural uses. Structure deterioration may occur in Luvisols with a high silt content if the soils are tilled in wet condition and/or with heavy machinery. Luvisols on steep slopes

246 Luvisols

require erosion control measures.

Luvisols in the temperate zone are widely grown to small grains, sugar beet, fodder, etc. or, in sloping land, they are used for orchards and/or grazing. In the mediterranean region, where Chromic, Calcic and Vertic Luvisols are common in colluvial deposits of limestone weathering, the lower slopes are often sown to wheat and sugar beets whereas the (eroded) upper slopes are in use for extensive grazing or tree crops.

PODZOLUVISOLS

Soils having an argic* B-horizon showing an irregular or broken upper boundary resulting from deep tonguing of the E-horizon 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 with iron and having redder hues and stronger chromas than the interiors; lacking a mollic* A-horizon.

Key to Podzoluvisol (PD) Soil Units

Podzoluvisols having permafrost^a within 200 cm of the surface.

Gelic Podzoluvisols (PDi)

Other Podzoluvisols showing gleyic[®] properties within 100 cm of the surface.

Gleyic Podzoluvisols (PDg)

Other Podzoluvisols showing stagnic^a properties within 50 cm of the surface.

Stagnic Podzoluvisols (PDs)

Other Podzoluvisols having a base saturation (by $1M\ NH_4OAc$ at pH 7.0) of less than 50 percent in at least a part of the B-horizon within 125 cm of the surface.

Dystric Podzoluvisols (PDd)

Other Podzoluvisols.

Eutric Podzoluvisols (PDe)

SUMMARY DESCRIPTION OF PODZOLUVISOLS

<u>Connotation</u>: soils having the bleached eluviation horizon of Podzols and the illuviation horizon of Luvisols; from <u>Podzols</u> and <u>Luvisols</u>.

<u>Parent material</u>: mostly unconsolidated glacial till, or materials of glaciolacustrine or fluvial origin, or of eolian origin (loess).

<u>Environment</u>: flat to undulating plains under boreal taiga, coniferous forest or mixed forest. The climate is temperate to boreal with cold winters and relatively short and cool summers, and an average annual precipitation sum of 500 to 1000 mm. Precipitation is evenly distributed over the year or, in the continental part of the Podzoluvisol belt, it has a peak in early summer.

^{*} Diagnostic horizon; see Annex 1 for full definition.

a Diagnostic property; see Annex 2 for full definition.

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<u>Profile development</u>: mostly AEBtC profiles with a dark but thin ochric A-horizon over an albic E-horizon that tongues into a brown argic B-horizon. Stagnic soil properties are common in boreal Podzoluvisols.

<u>Use</u>: acidity, low nutrient status, tillage and drainage problems, the short growing season and severe frost are serious limitations of Podzoluvisols. Most of these soils are under forest; livestock farming ranks second and arable cropping plays a minor role. In the USSR, the share of arable cropping increases towards the south and west of the Podzoluvisol belt, especially on Eutric Podzoluvisols.

REGIONAL DISTRIBUTION OF PODZOLUVISOLS

Podzoluvisols cover an estimated 262 million hectares, mainly concentrated in a broad belt extending from Poland and western Russia eastward into central Siberia. Small occurrences are found outside this belt, e.g. in Germany, Belgium and the USA (Michigan). Their absence from Canada, with similar climatic conditions as in Russia and Siberia, is explained by less developed tonguing in Canadian soils (Albic Luvisols). Figure 1 indicates the worldwide occurrence of Podzoluvisols.

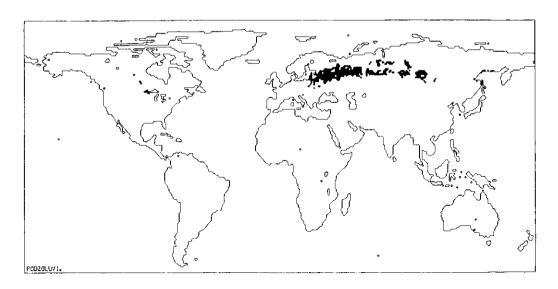


Fig. 1. Podzoluvisols worldwide.

GENESIS OF PODZOLUVISOLS

The genesis of Podzoluvisols has elements of podzolization, discussed in some detail in the chapter on Podzols, and of argilluviation (treated in the chapter on Luvisols).

Podzoluvisols occur in regions with a harsh climate which explains why there is little biological activity in their surface horizons. Leaching of metal-humus complexes from the litter layer is not offset by biological homogenization so that a distinct eluvial E-horizon could develop over an illuvial B-horizon. The sudden change in texture from the eluviation horizon to the illuviation horizon affects the internal drainage of Podzoluvisols. Periodic saturation of the surface soil and reduction of iron compounds (enhanced by dissolved organic compounds) cause strong bleaching of the eluvial horizon. The E-horizon extends into the underlying argic B-horizon along root channels and cracks (the characteristic 'tonguing'). Podzoluvisols are closely related with Albic Luvisols whose genetic history and morphological appearance are very similar to those of Podzoluvisols except that the E-horizon doesnot extend into the argic B-horizon.

Periodic saturation with water will, in first instance, cause iron compounds to concentrate in mottles or concretions of iron (hydr)oxides in a matrix that is increasingly depleted of iron. After many cycles, the surface soil loses virtually all its iron, either to the underlying illuviation horizon or to the deeper subsoil. See Figure 2.

Alternatively, it is possible that iron accumulations in the surface soil grow in size due to hysteresis between the precipitation of iron compounds in the oxidative phase and their (re)dissolution when reduced conditions prevail. This latter mechanism produces the discrete nodules, "the exteriors of which are enriched or weakly cemented or indurated with iron", that are mentioned in the definition of Podzoluvisols.

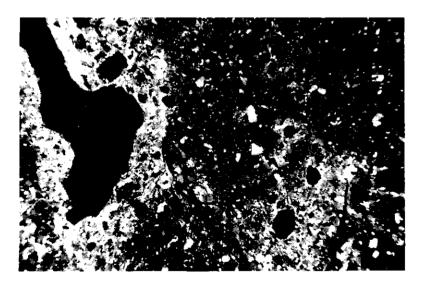


Fig. 2. Iron depletion zones along voids in a Stagnic Podzoluvisol in the Netherlands (cross polarized light; 1 cm = 500 μ m). Source: Miedema, 1987.

The recurrent saturation and leaching of the E-horizon foster acidification and loss of exchangeable bases. Ultimately, the leaching of clay

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and sesquioxides from the E-horizon may become so pronounced that only a sandy surface soil remains in which a secondary Podzol may eventually form. Low organic matter and iron contents explain the generally low structure stability of the leached surface soil; the E-horizon is normally compacted because of its low resistance to mechanical stress.

Alternating wetting and drying affects also the composition of the clay assemblage. In the extreme case, aluminium interlayering and clay decomposition lead to the development of acid, seasonally wet Planosols.

CHARACTERISTICS OF PODZOLUVISOLS

In the natural state, Podzoluvisols occur almost exclusively under a forest vegetation. A raw litter layer tops a dark but thin A(h)-horizon over a clearly bleached E-horizon which extends into a brown Bt-horizon. Both the E-horizon and the B-horizon are dense (see Figure 3).



Fig. 3. A Podzoluvisol from the Netherlands. Note the tonguing into the B-horizon. Source: Miedema, 1987.

Hydrological characteristics

The configuration of an E-horizon over a Bt-horizon is in itself evidence of downward water flow through the soil during at least part of the year. However, as the illuviation horizon grows, it becomes more and more an obstacle to percolation and may eventually cause regular stagnation of surface water. The diagnostic features of a Podzoluvisol (iron depletion and tonguing or, alternatively, iron nodules in the E-horizon) may or may not be strong enough to qualify as stagnic properties. Permafrost affects the hydrological characteristics profoundly.

Physical characteristics

Most Podzoluvisols are medium-textured; areas of coarse and/or finely textured Podzoluvisols occur in glaciolacustrine parent materials. The texture of the E-horizon is often very sandy. The low biological activity explains why most Podzoluvisols are somewhat compacted and why the eluvial horizon has often a platy structure. The low organic matter content of the surface soil and the high susceptibility to structure deterioration make it imperative that tillage be done in the proper moisture range. Rooting and uptake of water may be hindered by the dense Bt-horizon and/or by permafrost, either directly, or indirectly by poor internal drainage and inadequate aeration.

Chemical characteristics

The Ah-horizon of Podzoluvisols can contain between 1 and 10 percent organic carbon; the E-horizon contains rarely more than 1 percent organic C and a similar amount is present in the Bt-horizon. Podzoluvisols are moderately to strongly acid with pH(1M KCl) values from less than 4 to 5.5 or slightly higher. The CEC amounts to some 10 to 20 cmol(+)/kg, exclusive of the contribution by organic matter. The base saturation percentage ranges widely; values as low as 10 percent have been reported for Dystric Podzoluvisols with high contents of exchangeable aluminium and Al-interlayered clays, whereas Eutric Podzoluvisols with little Al interlayering have a base saturation between 60 and 90 percent.

The distinction between Eutric and Dystric Podzoluvisols is based on the base saturation of the argic B-horizon; the eluvial horizon is always very low in bases.

MANAGEMENT AND USE OF PODZOLUVISOLS

The major limitations of Podzoluvisols are posed by their acidity, low nutrient levels, tillage and drainage problems and by the climate, with its short growing season and severe frosts. The Podzoluvisols of the northern taiga are almost exclusively under forest; small areas are used as pasture land or hay fields. In the southern taiga zone, less than 10 percent of the non-forested area is used for agricultural production. Livestock farming is the main agricultural land use on Dystric Podzoluvisols (dairy production and cattle rearing); arable cropping (cereals, potatoes, sugar beets, forage maize) plays a minor role. In the USSR, the

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share of arable farming increases in southern and western directions, especially on Eutric Podzoluvisols.

With careful tillage, liming and application of fertilizers, Eutric Podzoluvisols produce 25-30 tons of fresh potatoes per hectare, 2-5 tons of winter wheat and 5-10 tons of dry herbage.

PLANOSOLS

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, exclusive of a natric* or spodic* B-horizon.

Key to Planosol (PL) Soil Units

Planosols having permafrost^a within 200 cm of the surface.

Gelic Planosols (PLi)

Other Planosols having a mollic * A-horizon or a eutric * histic * H-horizon.

Mollic Planosols (PLm)

Other Planosols having an umbric A-horizon or a dystric histic H-horizon.

Umbric Planosols (PLu)

Other Planosols having a base saturation (by 1M $\rm NH_4OAc$ at pH 7.0) of less than 50 percent in at least a part of the slowly permeable horizon within 125 cm of the surface.

Dystric Planosols (PLd)

Other Planosols.

Eutric Planosols (PLe)

* Diagnostic horizon; see Annex 1 for full definition.

a Diagnostic property; see Annex 2 for full definition.

SUMMARY DESCRIPTION OF PLANOSOLS

Connotation: soils with an eluvial horizon abruptly over a dense subsoil, typically in seasonally waterlogged flat lands; from L. planus, flat.

Parent material: mostly clayey alluvial and colluvial deposits.

<u>Environment</u>: level (depressed) areas in flat to gently undulating terrain, mainly in sub-tropical and temperate, semi-arid and subhumid regions with a climax vegetation of light forest or grasses.

<u>Profile development</u>: mostly AEBC profiles. Destruction and/or removal of clay has created a bleached and relatively coarse-textured surface soil that is abruptly underlain by a clayey subsurface horizon. Impeded downward movement of water is responsible for alternating oxidizing and reducing conditions in the upper part of the profile.

<u>Use</u>: the main limiting factor is the impervious clayey subsurface horizon. Flooding is common in the rainy season and severe drought stress occurs in

254 Planosols

the dry season. Planosols in regions with a warm summer season are mostly used for wet rice cultivation; elsewhere, for dry crops or as range land.

REGIONAL DISTRIBUTION OF PLANOSOLS

The main Planosol areas are shown in Figure 1. Their total expanse is estimated at approximately 150 million hectares worldwide, with the most important occurrences in South America (67 million hectares) and Australia (49 million hectares).

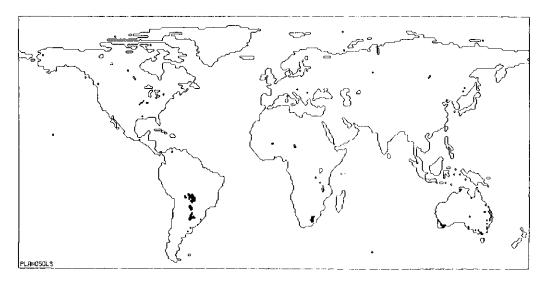


Fig. 1. Planosols worldwide.

The largest areas of Planosols are in subtropical and temperate climates with a clear alternation of wet and dry seasons, e.g. in southern Brasil, Paraguay, Argentina, the Sahelian zone, South Africa, the United States, and in eastern Asia and Australia. Planosols occur predominantly in flat lands with incised drainage channels 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 somewhat higher along the contours of a long and relatively gentle slope.

GENESIS OF PLANOSOLS

Planosols are soils with one or more upper horizons with a relatively low clay content, abruptly overlying a deeper horizon with considerably more clay. This abrupt textural change can have several causes:

- (1) It may be caused by <u>geogenetic processes</u> such as sedimentation of sandy over clayey layers, creep or sheet wash of lighter textured over clayey material, colluvial deposition of sandy over clayey material and selective erosion whereby the finest fraction is removed from the surface layers, and/or
- (2) It could be caused by <u>physical pedogenetic processes</u> viz. the eluviation-illuviation of clay in soil material with a low structure stability, and/or
- (3) <u>Chemical pedogenetic processes</u> may be involved, notably a process called 'ferrolysis', an oxidation-reduction sequence driven by chemical energy derived from bacterial decomposition of soil organic matter (Brinkman, 1979).

Ferrolysis is thought to proceed as follows:

In the absence of oxygen (e.g. in water-saturated soils with reducing organic matter), ferric oxides and hydroxides are reduced to Fe^{2+} -compounds which go into solution:

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

The Fe^{2+} -ions replace adsorbed basic cations and aluminium at the exchange complex; the replaced ions are partly leached out (together with some of the Fe^{2+}). When air re-enters the soil in a subsequent dry period, the exchangeable Fe^{2+} is oxidized again to Fe^{3+} -iron. This produces two H^{4-} -ions for each Fe^{2+} -ion oxidized:

$$4 \text{ Fe}^{2+} + O_2 + 10 \text{ H}_2O = 4 \text{ Fe}(OH)_3 + 8 \text{ H}^+$$

Part of the hydrogen ions replace aluminium or basic cations \underline{in} the clay structure; the cations become exchangeable. During the next wet season, Fe²⁺-ions are again formed from iron (hydr)oxides and a new cycle starts.

During the reduction phase, H^* -ions are consumed so that the pH rises. Once it has risen to about pH 5 to 5.5, $A1^{3+}$ -ions and OH^{-1} ons polymerize to hydroxy-Al-polymers with ring structures. The polynuclear Al-polymers are thought to remain 'trapped' in the interlayer spaces of 2:1 lattice clays. Other exchangeable cations, including Fe^{2+} , can then be built into the Al-polymers. The polymers in the interlayer spaces do not easily desintegrate and this causes hysteresis between wet and dry cycle processes and gradual alteration of the clay. Eventually, the clay has a lower cation exchange capacity and swells and shrinks less than before.

The abrupt change in clay content and, in some Planosols, in the nature of the clay, can only develop and persist if there is little homogenization of the soil. Examples are known of established Planosols that were later transformed to Phaeozems because of intense soil homogenization by termites.

CHARACTERISTICS OF PLANOSOLS

A typical horizon sequence of Planosols consists of an ochric or umbric A-horizon over an albic E-horizon directly on top of an argic B-horizon. In very wet locations there may even be a dystric histic H-horizon whereas in more arid regions the surface soil is commonly almost devoid of organic matter and cracked, and the soil qualifies as a 'yermic phase'. The albic E-horizon is greyish and has a sandy or loamy texture and a weak structure of low stability.

The most prominent feature of Planosols is the marked increase in clay content on passing from the upper to the middle horizon. The middle horizon may be a slowly permeable argic B-horizon or a heavy clay layer. It is commonly mottled and has a coarse angular blocky, prismatic or massive structure. Clay coatings may be present, but more often the change in texture appears to be due to strong weathering in situ in combination with clay destruction in the topsoil.

Mineralogical characteristics

Clay destruction through ferrolysis leads to a lower activity of the clay fraction (cation exchange capacity) in the topsoil and to a lower moisture retention capacity.

Hydrological characteristics

Planosols are subject to waterlogging in the wet season because of stagnation of (rain) water on level land with a slowly permeable subsoil layer at shallow depth (stagnic soil properties).

Physical characteristics

The surface soil horizons have a low structural stability; especially the silty ones become hard as concrete in the dry season and turn to a heavy mud with a very low bearing capacity when waterlogged in the wet season. The sandy topsoil is hard when dry but not cemented. The poor structure stability of the topsoil, the compactness of the subsoil and the abrupt transition from topsoil to subsoil are all detrimental to (the rooting possibilities of) crops.

Chemical characteristics

Planosols with signs of advanced ferrolysis are chemically degraded. The surface soil has lost (much of) its clay; the exchange capacity has become very low and bases are largely leached out.

Biological characteristics

The natural vegetation of areas with Planosols is mostly herbaceous. Where trees grow, it concerns species with extensive, shallow root systems that are capable of withstanding severe drought and strong winds. In line

with the poor physical and chemical properties of Planosols, the soil fauna is not very diverse and population densities are low.

MANAGEMENT AND USE OF PLANOSOLS

Planosols are used for various kinds of land use, albeit at a lower intensity than most other soils under the same climate. Large areas are used for extensive grazing. See Figure 2. Virgin Planosols are under a sparse grass and sedge vegetation or under grass with scattered bushes, trees or forest. The wood production of trees on Planosols is normally less than half of that on other soils with the same parent material.

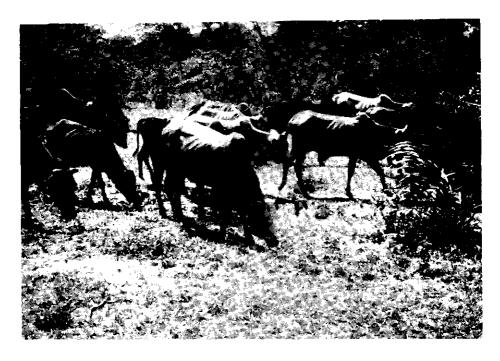


Fig. 2. Planosols are widely used for extensive grazing. Photograph by courtesy of ISRIC, Wageningen.

Planosols in the temperate zone are mainly in grass or they are used for arable crops such as wheat or sugarbeet. The yields are comparatively low as a consequence of the unattractive physical and chemical soil conditions. Root development is hampered by oxygen deficiency in wet periods, by the high density of the subsurface soil and by toxic levels of aluminium. The low hydraulic conductivity of the subsurface soil makes narrow drain spacings necessary. Experiments in Europe, spanning a number of years after

258 Planosols

drainage and deep loosening of the subsoil, failed to show conclusive results.

Planosols in southeast Asia are normally planted to a single crop of paddy rice, produced on bunded fields that are inundated in 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 supplemental irrigation. Fertilizers are needed for good rice yields. The soils should be allowed to dry out at least once a year to prevent or minimize micro-element deficiencies or toxicities associated with prolonged soil reduction. Some Planosols require application of more than just macronutrients and their low fertility level may prove difficult to correct. Where the temperature permits paddy rice cultivation, this is probably superior to any other kind of land use.

In climates with long dry periods and short infrequent wet spells, supplemental irrigation of grassland in the dry season may hold promise. Strongly developed Planosols with a silty or sandy surface soil are perhaps best left in their natural state.

PODZOLS

Soils having a spodic* B-horizon.

Key to Podzol (PZ) Soil Units

Podzols having permafrost^a within 200 cm of the surface.

Gelic Podzols (PZi)

Other Podzols showing gleyic^a properties within 100 cm of the surface.

Gleyic Podzols (PZg)

Other Podzols having a B-horizon in which a subhorizon contains dispersed organic matter and lacks sufficient free iron to turn redder on ignition.

Carbic Podzols (PZc)

Other Podzols in which the ratio of percentage of free iron to percentage of carbon is 6 or more in all subhorizons of the B-horizon.

Ferric Podzols (PZf)

Other Podzols lacking or having only a thin (2 cm or less) and discontinuous albic E-horizon; lacking a subhorizon within the B-horizon which is visibly more enriched with carbon.

Cambic Podzols (PZb)

Other Podzols.

Haplic Podzols (PZh)

SUMMARY DESCRIPTION OF PODZOLS

<u>Connotation</u>: soils with a subsurface horizon that has the appearance of ash due to strong bleaching by aggressive organic acids; from R. <u>pod</u>, under, and <u>zola</u>, ash.

<u>Parent material</u>: unconsolidated siliceous rock weathering, including glacial till, and alluvial and eolian deposits of quartzitic sands.

<u>Environment</u>: sites with poorly decomposed organic matter and downward percolation of fulvic acids. In the northern hemisphere in level to hilly land under heather and coniferous forest; in the humid tropics under light forest.

<u>Profile development</u>: mostly AhEBhsC profiles. Complexes of Al, Fe and organic acids migrate from the surface soil to the B-horizon with percolating rain water. The Fe- and Al-humus complexes precipitate in an

^{*} Diagnostic horizon; see Annex 1 for full definition.

a Diagnostic property; see Annex 2 for full definition.

260 Podzols

illuvial spodic B-horizon; the leached subsurface soil remains behind as a bleached albic E-horizon.

<u>Use</u>: severe acidity, low chemical fertility and unfavourable physical properties make most Podzols unsuitable for arable cropping. They have some potential for forestry and extensive grazing.

REGIONAL DISTRIBUTION OF PODZOLS

Podzols cover some 478 million hectares worldwide, almost exclusively in the temperate and boreal regions of the northern hemisphere (see Figure 1). Besides these zonal Podzols, there are smaller occurrences of intrazonal Podzols, both in the temperate zone and in the tropics.

Tropical Podzols total not more than 10 million hectares, mainly in residual sandstone weathering in perhumid regions and in old, alluvial quartz sands. The exact distribution of tropical Podzols is not known; important occurrences are along the Rio Negro and in the Guianas in South America, in the Malesian region (Kalimantan, Sumatra, Irian), and in northern Australia. They seem rather uncommon in Africa.

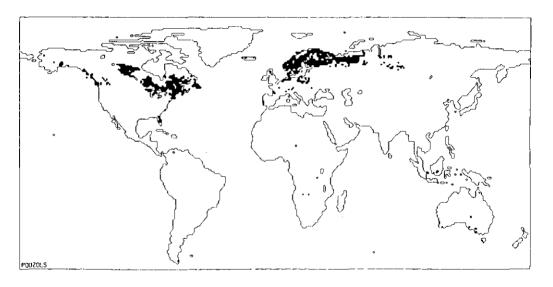


Fig. 1. Podzols worldwide.

GENESIS OF PODZOLS

Podzolization (the formation of a spodic B-horizon) is actually a succession of processes, including

(1) 'cheluviation', the movement of soluble metal-humus complexes (chelates) out of the surface layer(s) to greater depth, and (2) 'chilluviation', the subsequent accumulation of Al- and Fe-chelates in a spodic B-horizon. (Soluble organic compounds can move to still deeper horizons.)

Cheluviation

Soluble organic substances produced by microbial (mainly fungal) attack on plant litter, form complexes with ${\rm Al}^{3+}$ - and ${\rm Fe}^{3+}$ -ions and move downward with the soil solution. There is little known of the rates at which such reactions proceed in natural soil systems; they seem to depend on the nature and concentration of the organic compounds in solution, the pH, the mineral surface area that is susceptible to attack, and the nature of the minerals.

Fulvic acids are the dominant complexing organic compounds; their carboxylic and phenolic groups act as 'claws' (Gr. chela, hence the term 'chelation') which preferably grab polyvalent metal ions such as ${\rm Al}^{3+}$ and ${\rm Fe}^{3+}$. Chelation continues as long as the fulvic acids are not saturated with metal ions.

Fulvic acids are comparatively abundant where low temperatures, low chemical soil fertility and/or periodic water saturation retard the microbial decomposition of organic matter. There is little faunal activity under such unfavourable conditions, so that most soil organic matter (and the highest concentration of fulvic acids) remains in the surface soil.

Fulvic acids are soluble in water. However, as more and more reaction sites become occupied by ${\rm Al}^{3+}$ and ${\rm Fe}^{3+}$, the solubility product decreases; fully saturated fulvic acids are practically insoluble in water. If the soil supplies ${\rm Al}^{3+}$ and or ${\rm Fe}^{3+}$ at low rates compared with the rate of fulvic acid production, fulvic acids can migrate over considerable distances, even over hundreds of metres. Normally, however, the bulk of all fulvic acids is saturated with ${\rm Al}^{3+}$ and/or ${\rm Fe}^{3+}$ after migration over only a few decimetres so that an illuvial B-horizon (spodic B-horizon) forms in the profile. In sandy soils that are very low in ${\rm Fe}^{3+}$ and ${\rm Al}^{3+}$, the spodic B-horizon may occur at a depth of several metres or, in the extreme case, a spodic B-horizon may not form at all and (the bulk of) the fulvic acid is discharged in effluent 'black water'.

Chilluviation

Normally, water-soluble organic substances react with weathering products while percolating downward until the ratio of (Al+Fe) to organic carbon becomes sufficiently high that the complex precipitates. Alternatively, saturation may be brought about by evaporation of the solvent, or low-ratio complexes may be absorbed by high-ratio complexes that are already precipitated. The result is an accumulation horizon in all cases.

Evaporation of soil moisture after a single shower causes precipitation of Al- and Fe-fulvates at the depth to which water penetrated the soil. Partial decomposition of the precipitated complexes increases the degree of Al- and Fe-saturation of the remaining fulvic acids, which prevents that

these dissolve again during the next rain. The humus 'fibres' in well drained Podzols are probably formed in this way.

The accumulation process is reversible to some extent. If uncomplexed humic substances are added at a faster rate than Al^{3+} - and Fe^{3+} -ions are released by weathering minerals, the Al,Fe-humus complexes will acquire an ever lower metal-to-organic ratio and become increasingly soluble. Ultimately, an entire spodic B-horizon can be moved to a greater depth. Strongly podzolized soils with a very thick albic E-horizon occur in the humid tropics. The illuvial B-horizon of such soils is very deep ('Giant Podzols'), or may even be absent altogether.

Podzolization versus ferralitization

From the point of view of soil formation, opposite processes take place in Podzols, where Fe- and Al-oxides dissolve and iron and aluminium are leached out, and in Ferralsols where Fe- and Al-oxides remain stable and increase in content through relative accumulation. It has long been a point of academic debate whether the Podzol or the Ferralsol represents the ultimate in (weathering and) soil formation in the humid tropics.

One suggestion was that there is only one fundamental process, that of podzolization, and that red tropical soils are essentially soils in an early stage of podzolization, but it has also been said that podzolization could follow ferralitization as the soil becomes progressively more acid.

The bleached horizons of well developed Podzols are formed when virtually all weatherable constituents (including non-quartz silica and such Aland Fe-compounds as may be present) are removed by intense leaching, leaving behind only a skeleton of quartz sand grains. Strongly leached Ferralsols, however, although very low in cations and with a pH of 4.0 or less, show no tendency to develop an eluvial E-horizon; the ferric iron is present in highly stable forms.

The situation can perhaps be summed up as follows: In the wet tropics, soil formation will produce a Ferralsol in well drained clayey parent materials that are rich in iron; a Podzol will result in imperfectly drained, coarse-textured and quartz-rich materials which receive organic matter that decomposes slowly under conditions of oligotrophy.

CHARACTERISTICS OF PODZOLS

Not all soils with evidence of podzolization qualify as Podzols. The FAO-Unesco classification system stipulates that a Podzol must have a spodic B-horizon, with identifiable 'active amorfic materials' of organic matter and aluminium.

NOTE THAT although there are two characteristic horizons in a Podzol, viz. an eluviation and an illuviation horizon, only the illuviation horizon is a classification criterion. By convention, a soil with a thin Ah-horizon over an eluviation horizon that extends to more than 125 cm depth, is not classified as a Podzol even if a marked illuviation horizon occurs deeper than 125 cm. Despite a history of strong podzolization, such deeply eluviated soils are no Podzols anymore; they have become Albic Arenosols.

A typical zonal Podzol has a pale grey, strongly leached E-horizon between a dark surface horizon with organic matter, and a brown to very dark brown spodic B-horizon. Most Podzols have a surface litter layer of 1 to 5 cm thick, loose and spongy, and grading into an Ah-horizon with partly humified organic matter. In the litter layer in particular, most of the individual plant fragments are still recognizable and live roots may be beset with mycorrhizae. The Ah-horizon consists of a dark grey mixture of organic matter and mineral material (mainly quartz) with many of the grains showing black isotropic coatings. The underlying bleached E-horizon has a characteristic single grain structure; the structure of the brown to black Bhs-horizon varies from loose through firm subangular blocky to very hard and massive.

At the drier end of the climatic range for zonal Podzols, the illuviation horizon has commonly a high chroma signifying accumulation of iron and aluminium oxides. In more humid regions, the Bhs-horizon is darker and has a higher content of translocated organic matter. Where the parent material consists of (almost) pure quartz, a Bh-horizon may form in which the illuviated material is entirely organic. There may be an indurated layer or Cx-horizon (fragipan) in the lower solum. Figure 2 shows the typical AhEBhsC configuration of a mature Podzol.



Fig. 2. A Podzol in full splendour. Photograph: the Marbut collection.

The profile of a typical intrazonal tropical lowland Podzol has a surface layer of poorly decomposed ('raw'), acid humus with a high C/N-ratio. This O-horizon is underlain by a humus-stained A-horizon over a light grey to white eluvial E-horizon of sand texture that varies in thickness from 20 cm to more than 2 metres (Giant Podzols). The Bh-horizon is commonly dark brown and irregular in depth. In places, there are mottles or soft concretions of secondary iron and aluminium oxides, and/or there is slightly more clay in the illuvial horizon than higher in the profile. Brightly coloured B(h)s-horizons with sesquioxide accumulation as occur in the temperate zone, are less common in the tropics.

Mineralogical characteristics

The mineralogy of Podzols is somewhat variable but is nearly always marked by a high quartz content. In cool, humid climates where leaching is intense, the parent material may originally have been of intermediate or even basic composition. A bleached E-horizon is hardly noticeable then; there is only a thick, dark B-horizon of intricately mixed organic and mineral material.

The highest concentration of free $\mathrm{Fe}_2\mathrm{O}_3$ occurs in the middle of the illuviation horizon; the lowest contents are in the upper portion of the eluviation horizon. There is normally more aluminium than iron present in the B-horizon but since ferric oxides have strong colours, the iron compounds are more conspicuous.

Hydrological characteristics

Most Podzols formed under conditions of free movement of water through the upper and middle portions of the soil; where the illuviation horizon is dense, or an indurated layer (fragipan) has formed, water movement through the subsoil may be impaired. Although Podzols are associated with regions having an annual precipitation surplus, their low water holding capacity may induce drought stress in dry periods.

Physical characteristics

The clay content is normally low or very low; it seldom exceeds 10 percent by weight in the bleached eluviation horizon but could be slightly higher in the illuvial horizon. The sandy texture facilitates tillage.

Chemical characteristics

The organic matter profile of Podzols shows two areas of concentration, viz. a major one at the surface and a minor one in the Bh(s)-horizon. The C/N-ratio has a characteristic pattern with values of 20 to 50 in the surface horizon, decreasing to 10 to 15 in the bleached horizon and then increasing again to 15 to 25 in the illuviation horizon. The contents of 'total' iron and aluminium are highest in the spodic B-horizon; Al-humus complexes tend to accumulate somewhat deeper in the illuviation horizon than Fe-humus compounds. See Figure 3.

The chemical fertility of Podzols is low as a consequence of their high degree of leaching and their low CEC. The latter is largely a function of the properties and distribution of soil organic matter. Plant nutrients are concentrated in the surface horizon(s) where cycling elements are released upon decomposition of organic debris. The surface horizons are normally acid with $pH(H_2O,\ 1:1)$ values between 3.5 and 4.5; the pH-value increases with depth to a maximum of about 5.5 in the deep subsoil.

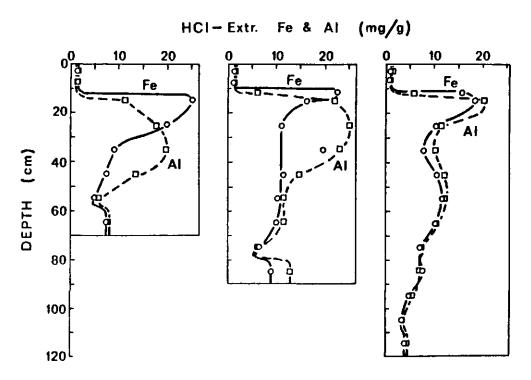


Fig. 3. Three iron and aluminium profiles of Podzols under deciduous forest in New Hampshire, USA. Note that the distributions of free iron and free aluminium diverge in two of the three B-horizons (Wood, 1980).

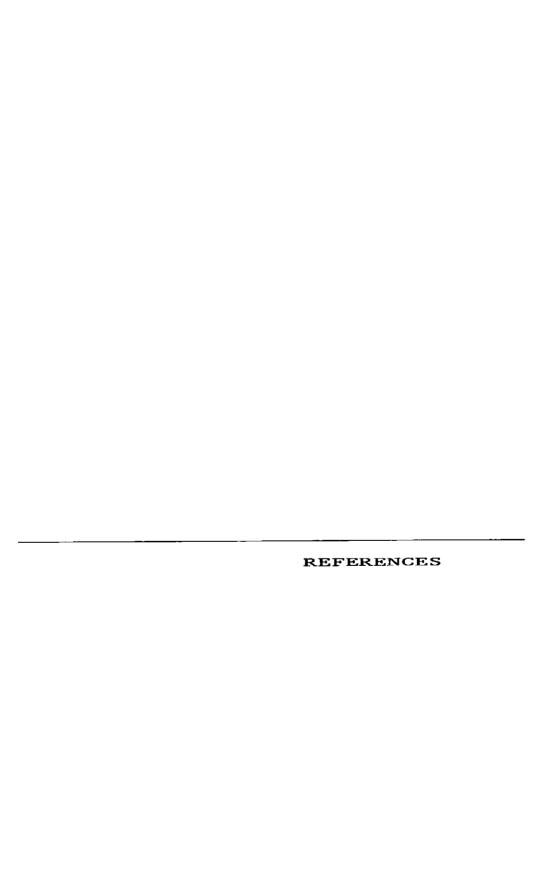
Biological characteristics

Larger soil animals such as earthworms are scarce in Podzols; decomposition of organic matter and surface soil homogenization are slow and are mainly done by fungi, small arthropods and insects.

MANAGEMENT AND USE OF PODZOLS

The low nutrient status, low level of available moisture and low pH make Podzols unattractive soils for agricultural production. Intrazonal Podzols are more frequently reclaimed for arable farming than zonal Podzols, particularly those in temperate climates. <u>Deep ploughing</u> to improve the moisture storage capacity of the soil and/or to eliminate a dense B-horizon or fragipan, and <u>liming</u> are the main ameliorative measures applied. Podzols need full fertilization with micro-nutrients in addition to N, P and K. In places, stoniness and/or high sensitivity to erosion are further limitations to arable farming.

Most zonal Podzols are under forest or are used for low volume grazing. Tropical Podzols are normally under a light forest that recovers only slowly after cutting/burning. By and large, mature Podzols are best left under their natural (climax) vegetation.





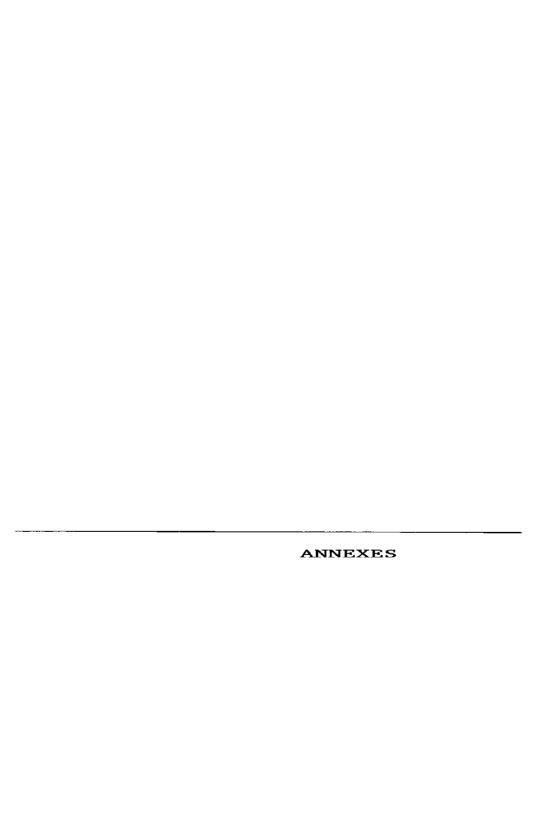
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NOTES





DEFINITIONS OF DIAGNOSTIC HORIZONS

(Source: the Revised Legend of the Soil Map of the World, 1988, pp 20-27)

Soil horizons that combine a set of features which are used for soil identification are called 'diagnostic horizons'. Since the characteristics of soil horizons are produced by soil forming processes, the use of diagnostic horizons for separating soils relates the classification to general principles of soil genesis. Objectivity is secured, however, in that the processes themselves are not used as criteria but only their effects, expressed in terms of quantitatively defined morphometric properties that have identification value.

Some of the characteristics used to classify soils do not (always) occur in connection with horizons. They are 'diagnostic properties'. The definitions of diagnostic properties are presented in Annex 2.

The <u>histic H-horizon</u> is an H-horizon which is more than 20 cm but less than 40 cm thick. It can be more than 40 cm but less than 60 cm thick if it consists for 75 percent or more, by volume, of sphagnum fibres or has a bulk density when moist of less than $0.1~\text{Mg/m}^3$.

A surface horizon of organic material less than 25 cm thick qualifies as a histic H-horizon if, after having been mixed to a depth of 25 cm, it has 16 percent or more organic carbon if the mineral fraction contains more than 60 percent clay, or 8 percent or more organic carbon if the mineral fraction contains no clay, or intermediate contents of organic carbon for intermediate contents of clay. The same criteria apply to a plough layer which is 25 cm or more thick.

The mollic A-horizon is an A-horizon which, after the surface 18 cm are mixed as in ploughing, has the following properties:

- the soil structure is sufficiently strong that the horizon is not both massive and hard or very hard when dry. Very coarse prisms larger than 30 cm in diameter are included in the meaning of 'massive' if there is no secondary structure within the prisms.
- 2. both broken and crushed samples have colours with a chroma of less than 3.5 when moist, and a value darker than 3.5 when moist and 5.5 when dry; the colour value is at least one unit darker than that of the C-horizon (both moist and dry). If a C-horizon is not present, comparison should be made with the horizon immediately underlying the A-horizon. If there is more than 40 percent finely divided lime, the limits of colour value 'dry' are waived; the colour value 'moist' should be 5 or less.
- 3. the base saturation (by 1M NH_4OAc at pH 7.0) is 50 percent or more.
- 4. the organic carbon content is at least 0.6 percent throughout the thickness of mixed soil, as specified below. The organic carbon content is at least 2.5 percent if the colour requirements are waived because of finely divided lime. The upper limit of organic carbon content of the mollic A-horizon is the lower limit of the histic H-horizon.

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- 5. the thickness is 10 cm or more if resting directly on hard rock, on a petrocalcic horizon, a petrogypsic horizon, or a duripan; the thickness of the A-horizon must be at least 18 cm and more than one third of the thickness of the solum where the solum is less than 75 cm thick, and must be more than 25 cm where the solum is more than 75 cm thick. The measurement of the thickness of a mollic A-horizon includes transition horizons in which the characteristics of the A-horizon are dominant.
- 6. The content of P_2O_5 soluble in 1 percent citric acid is less than 250 mg/kg soil. This restriction is made to eliminate plough layers of very old arable soils or kitchen middens. The restriction does not apply, however, if the amount of P_2O_5 soluble in citric acid increases below the A-horizon or when the soil contains phosphate nodules, as may be the case in highly phosphatic parent materials.

The <u>fimic A-horizon</u> is a man-made surface layer 50 cm or more thick which has been produced by long continued manuring with earthy admixtures. The fimic A-horizon commonly contains artefacts such as bits of brick and pottery throughout its depth. If the fimic A-horizon meets the requirements of the mollic or umbric A-horizon, it is distinguished from it by a P_2O_5 content (by 1 percent citric acid) which is higher than 250 mg/kg soil. (Both the analytical method and the limiting amount of P_2O_5 are likely to be changed in the future).

The <u>umbric A-horizon</u> is comparable to the mollic A-horizon in its requirements concerning colour, organic carbon and phosphorus content, consistency, structure and thickness. The umbric A-horizon, however, has a base saturation (by 1M NH_2OAc at pH 7.0) of less than 50 percent.

The <u>ochric A-horizon</u> is an A-horizon that is too light in colour, has too high a chroma, too little organic carbon, or is too thin to be mollic or umbric, or is both hard and massive when dry. Finely stratified materials, e.g. surface layers of fresh alluvial deposits, do not qualify as an ochric A-horizon.

The <u>albic E-horizon</u> is a horizon from which clay and free iron oxides have been removed, or in which the oxides have been segregated to the extent that the colour of the horizon is determined by the colour of the primary sand and silt particles rather than by coatings on these particles. An albic E-horizon has a colour value 'moist' of 4 or more, or a value 'dry' of 5 or more, or both. If the value 'dry' is 7 or more, or the value 'moist' is 6 or more, the chroma is 3 or less either 'dry' or 'moist'. If the value 'dry' is 5 or 6, or if the value 'moist' is 4 or 5, the chroma is closer to 2 than to 3 either 'dry' or 'moist'. If the parent materials have a hue of 5YR or redder, a chroma 'moist' of 3 is permitted in the albic E-horizon where the chroma is due to the colour of uncoated silt or sand grains.

The <u>argic B-horizon</u> is a subsurface horizon which has a distinctly higher clay content than the overlying horizon. The textural differentiation may be due to an illuvial accumulation of clay or to destruction of clay in the surface horizon, or to selective surface erosion of clay, or to biological activity, or to a combination of two or more of these different processes. Sedimentation of surface materials which are coarser than the subsurface horizon may enhance a pedogenic textural differentiation. However, a mere lithological discontinuity, such as may occur in alluvial deposits, does not qualify as an argic B-horizon. If an argic B-horizon is formed by clay illuviation, clay skins may occur on ped surfaces, in fissures, in pores and in channels.

An argic B-horizon must meet the following requirements:

- 1. have a texture that is sandy loam or finer and have at least 8 percent clay in the fine earth fraction.
- 2. lack the set of properties which characterize the ferralic B-horizon.
- 3. contain more total clay than an overlying coarser-textured horizon (exclusive of differences which result from a lithological discontinuity only):
 - (a) if the overlying horizon has less than 15 percent total clay in the fine earth fraction, the argic B-horizon must contain at least 3 percent more clay;
 - (b) if the overlying horizon has 15 percent or more and less than 40 percent total clay in the fine earth fraction, the ratio of clay in the argic B-horizon to that in the overlying horizon must be 1.2 or more:
 - (c) if the overlying horizon has 40 percent or more total clay in the fine earth fraction, the argic B-horizon must contain at least 8 percent more clay.
- 4. if at least some part of the argic B-horizon shows clay skins on at least 1 percent of both horizontal and vertical ped surfaces and in the pores, or shows oriented clays in at least 1 percent of the cross section, the increase in clay content must be reached within a vertical distance of 30 cm. If the requirement for the presence of clay skins is not met, the increase in clay content between the argic B-horizon and the overlying horizon must be reached within a vertical distance of 15 cm.
- 5. the argic B-horizon must be at least one tenth of the thickness of the sum of all overlying horizons. If the argic B-horizon is entirely composed of lamellae, the lamellae must have a combined thickness of at least 15 cm.
- 6. the coarser-textured horizon overlying the argic B-horizon must be at least 18 cm thick after mixing, or 5 cm if the textural transition to the argic B-horizon is abrupt. If a soil shows a lithological discontinuity or if only a plough layer overlies the argic B-horizon, at least some part of the horizon needs to show clay skins on at least 1 percent of both horizontal and vertical surfaces and in the pores, or show oriented clays in at least 1 percent of the cross section.
- lack the structure and sodium saturation characteristics of the natric B-horizon.

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The definition of the argic B-horizon recognizes distinct textural differentiation as a diagnostic feature, even if clay skins cannot be identified. On the other hand, an accumulation layer marked by a low CEC of the clay, a low content of dispersible clay, and a low silt-clay ratio, as may occur in Ferralsols, is excluded from the argic B-horizon.

The <u>natric B-horizon</u> has the properties 1 through 6 of the argic B-horizon as described above. In addition it has:

- 1. a columnar or prismatic structure in some part of the B-horizon, or a blocky structure with tongues of an eluvial horizon in which there are uncoated silt or sand grains extending more than 2.5 cm into the horizon
- 2. a saturation with exchangeable sodium of more than 15 percent within the upper 40 cm of the horizon; or more exchangeable magnesium plus sodium than calcium plus exchange acidity (at pH 8.2) within the upper 40 cm of the horizon if the saturation with exchangeable sodium is more than 15 percent in some subhorizon within 200 cm of the surface.

The <u>cambic B-horizon</u> is an altered horizon lacking properties that meet the requirements of an argic, natric or spodic B-horizon; lacking the dark colours, organic matter content and structure of the histic H-horizon, or the mollic and umbric A-horizons; having the following properties:

- has a texture of sandy loam or finer and has at least 8 percent clay in the fine earth fraction.
- is at least 15 cm thick with its base at least 25 cm below the soil surface.
- 3. soil structure is at least moderately developed or rock structure is absent in at least half the volume of the horizon.
- 4. has a cation exchange capacity of more than 16 cmol(+)/kg clay, or has a content of 10 percent or more weatherable minerals in the 50-200 μm fraction.
- 5. shows evidence of alteration in one of the following forms:
 - (a) stronger chroma, redder hue, or higher clay content than the underlying horizon;
 - (b) evidence of removal of carbonates.
 - (c) if carbonates are absent in the parent material and in the dust that falls on the soil, the required evidence of alteration is satisfied by the presence of soil structure and the absence of rock structure in more than 50 percent of the horizon.
- 6. shows no cementation, induration, or brittle consistence when moist.

The <u>spodic B-horizon</u> is an illuviation horizon which meets one or more of the following requirements below a depth of 12.5 cm, or, if present, below an A-horizon or an E-horizon:

- 1. a subhorizon more than 2.5 cm thick that is continuously cemented by a combination of organic matter with iron or aluminium or with both.
- 2. a sandy or coarse-loamy texture with distinct dark pellets of coarse silt size or larger or with sand grains covered with cracked coatings which consist of organic matter and aluminium with or without iron.

- 3. one or more subhorizons in which:
 - (a) if there is 0.1 percent or more extractable iron, the ratio of iron plus aluminium extractable by pyrophosphate (at pH 10) to percent clay is 0.2 or more, or if there is less than 0.1 percent extractable iron, the ratio of aluminium plus carbon to clay is 0.2 or more; and
 - (b) the sum of pyrophosphate-extractable iron plus aluminium is half or more of the sum of dithionite-citrate-extractable iron plus aluminium; and
 - (c) the thickness is such that the index of accumulation of amorphous material (CEC at pH 8.2 minus one half the clay percentage multiplied by the thickness in centimetres) in the horizons that meet the preceding requirements is 65 or more.

The ferralic B-horizon is a horizon that:

- 1. has a texture that is sandy loam or finer and has at least 8 percent clay in the fine earth fraction.
- 2. is at least 30 cm thick.
- 3. has a cation exchange capacity of the fine earth fraction equal to or less than 16 cmol(+)/kg clay, or has an effective cation exchange capacity of the fine earth fraction equal to or less than 12 cmol(+) per kg clay (1M NH₂OAc-extractable bases plus 1M KCl-extractable Al).
- 4. has less than 10 percent weatherable minerals in the 50-200 μm fraction.
- 5. has less than 10 percent water-dispersible clay.
- 6. has a silt-clay ratio which is 0.2 or less.
- 7. does not have andic properties.
- 8. has less than 5 percent by volume showing rock structure.

The <u>calcic horizon</u> is a horizon of accumulation of calcium carbonate. The accumulation may be in the C-horizon, but it may also occur in a B- or in an A-horizon. The calcic horizon features secondary carbonate enrichment over a thickness of 15 cm or more, has a carbonate equivalent content of 15 percent or more and at least 5 percent greater than that of a deeper horizon. The latter requirement is expressed by volume if the secondary carbonates in the calcic horizon occur as pendants on pebbles, or as concretions or as soft powdery lime. If such a calcic horizon rests on materials with 40 percent or more calcium carbonate equivalent, the percentage of carbonates need not decrease with depth.

The <u>petrocalcic horizon</u> is a continuous cemented or indurated calcic horizon, cemented by calcium carbonate and in places by calcium and some magnesium carbonate. Accessory silica may be present. The petrocalcic horizon is continuously cemented to the extent that dry fragments do not slake in water and roots cannot enter. It is massive or platy, extremely hard when dry so that it cannot be penetrated by spade or auger, and very firm to extremely firm when moist.

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The gypsic horizon is a horizon of secondary calcium sulfate enrichment that is 15 cm or more thick, and has at least 5 percent more gypsum than the underlying C-horizon; the product of the thickness in centimetres and the percentage of gypsum is 150 or more. The percentage of gypsum (CaSO₄.2H₂O) is calculated as the product of gypsum content, expressed as cmol(+)/kg soil, and the equivalent weight of gypsum, 86, divided by 10³. Gypsum may accumulate uniformly throughout the matrix or as nests of crystals; in gravelly material gypsum may accumulate as pendants below coarse fragments.

The <u>petrogypsic horizon</u> is a gypsic horizon that is so cemented with gypsum that dry fragments do not slake in water and roots cannot enter.

The <u>sulfuric horizon</u> is a horizon at least 15 cm thick and characterized by a $pH(H_2O, 1:1)$ of less than 3.5 and jarosite mottles with a hue of 2.5Y or more and a chroma of 6 or more.

DEFINITIONS OF DIAGNOSTIC PROPERTIES

(Source: the Revised Legend of the Soil Map of the World, 1988, pp 28-34)

'Diagnostic properties' are features of horizons or of soil materials which, when used for soil classification, are quantitatively defined. They are defined as follows:

The term 'abrupt textural change' refers to a clay increase between two layers, which takes place over a distance of less than 5 cm, where the lower layer shows the following:

- 1. a clay content which is twice the amount of clay in the overlying layer if the latter has less than 20 percent clay.
- 2. an absolute clay increase of 20 percent or more over the amount of clay in the overlying layer if the latter has 20 percent clay or more.

The term <u>andic properties</u> applies to soil materials which meet one or more of the following three requirements:

- (a) acid oxalate extractable aluminium plus half of the acid oxalate extractable iron is 2.0 percent or more in the fine earth fraction; and
 - (b) bulk density of the fine earth, measured in the field moist state, is 0.9 Mg/m^3 or less; and
 - (c) phosphate retention is more than 85 percent.
- 2. (a) more than 60 percent by volume of the whole soil is volcanoclastic material coarser than 2 mm; and
 - (b) acid oxalate extractable aluminium plus half of the acid oxalate extractable iron is 0.40 percent or more in the fine earth fraction.
- 3. the 0.02 to 2.0 mm fraction is at least 30 percent of the fine earth fraction and meets one of the following:
 - (a) if the fine earth fraction has acid oxalate extractable aluminium plus half of the acid oxalate extractable iron of 0.40 percent or less, there is at least 30 percent volcanic glass in the 0.02 to 2.0 mm fraction; or
 - (b) if the fine earth fraction has acid oxalate extractable aluminium plus half of the acid oxalate extractable iron of 2.0 percent or more, there is at least 5 percent volcanic glass in the 0.02 to 2.0 mm fraction; or
 - (c) if the fine earth fraction has acid oxalate extractable aluminium plus half of the acid oxalate extractable iron of between 0.40 percent and 2.0 percent, there is a proportional content of volcanic glass in the 0.02 to 2.0 mm fraction (between 5 and 30 percent).

The term 'calcareous' applies to soil materials which show strong effervescence with 10 percent HCl in most of the fine earth, or which contain more than 2 percent calcium carbonate equivalent.

The term 'continuous hard rock' applies to underlying material which is sufficiently coherent and hard when moist to make hand digging with a spade impracticable. The material is continuous except for few cracks produced in place and (a) without significant displacement of the pieces, and (b) not horizontally distant an average 10 cm or more. Continuous hard rock does not include hard subsurface horizons such as a petrocalcic or a petrogypsic horizon, or a duripan or a petroferric layer (See Annex 3).

The adjective ' $\frac{dystric}{d}$ ' refers to a base saturation percentage (by 1M NH_{α}OAc at pH 7.0) of less than 50 percent.

The adjective 'eutric' refers to a base saturation percentage (by $1M NH_2OAc$ at pH 7.0) of 50 percent or more.

The term 'ferralic properties' is used in connection with Cambisols and Arenosols which have a cation exchange capacity (by 1M NH_4OAc at pH 7.0) of less than 24 cmol(+)/kg clay or less than 4 cmol(+)/kg soil in at least some subhorizon of the cambic B-horizon or the horizon immediately underlying the A-horizon.

The term '<u>ferric properties</u>' is used in connection with Luvisols, Alisols, Lixisols and Acrisols showing one or more of the following:

- 1. many coarse mottles with hues redder than 7.5YR or chromas more than 5, or both;
- discrete nodules up to 2 cm in diameter, the exteriors of the nodules being enriched and weakly cemented or indurated with iron and having redder hues or stronger chromas than the interiors.

The term 'fluvic properties' refers to fluviatile, marine and lacustrine sediments which, unless empoldered, receive fresh materials at regular intervals, and have one or both of the following properties:

- an organic carbon content that decreases irregularly with depth or that remains above 0.02 percent to a depth of 125 cm. Thin strata of sand may have less organic carbon if the finer sediments below, exclusive of buried A-horizons, meet the requirement.
- stratification in at least 25 percent of the soil volume within 125 cm of the surface.

The term 'geric properties' refers to soil materials which have either:

- 1.5 cmol(+)/kg soil or less of extractable bases (Ca, Mg, K, Na) plus unbuffered 1M KCl extractable aluminium and a pH(1M KCl) of 5.0 or more, or
- 2. a delta pH (pH KCl minus pH H₂O) of +0.1 or more.

The term 'gleyic properties' refers to soil materials which are saturated with groundwater at some period of the year or throughout the year, in most years, and which evidence of reduction processes or of reduction and segregation by iron:

- 1. reduction conditions for part of the year or throughout the year, in most years, evidenced by one or more of the following:
 - (a) a value of rH equal to or less than 19. (rH = Eh(mV)/29 + 2pH).
 - (b) the appearance of a solid dark blue colour on the freshly broken surface of a field-wet soil sample, after spraying it with a solution in water of 1 percent potassium ferric cyanide.
 - (c) the appearance of a strong red colour on a freshly broken surface of a field-wet soil sample, after spraying it with a 0.2 percent α, α dipyridyl solution in 10 percent acetic acid.
- 2. saturation by groundwater evidenced by the following:
 - (a) reduction conditions as defined under 1. for a part of the year or throughout the year.
 - (b) groundwater standing in a deep unlined bore hole at such a depth that the capillary fringe reaches the soil surface; the water in the bore hole is stagnant and remains coloured when dye is added to it.
 - (c) white to black (N), or blue to green (GY, BG, G or B) colours in more than 95 percent of the soil matrix; high chroma oxidized mottles, if present, occur on ped faces or in root channels or animal burrows.

NOTE THAT for soils in which the content of iron oxides is very low, or in which iron oxides are present in such large quantities or are inert and so well crystallized that they remain brown or red even in reduced conditions, the above colour requirements do not apply.

The term 'gypsiferous' applies to soil material which contains 5 percent or more gypsum.

The term 'interfingering' refers to penetrations of an albic E-horizon into an underlying argic or natric B-horizon along ped faces, primarily vertical faces. The penetrations are not wide enough to constitute 'tonguing', but form continuous coatings of clean silt or sand, more than 1 mm thick, on the vertical ped faces ('skeletans'). A total thickness of more than 2 mm is required if each ped has a coating of more than 1 mm. Because quartz is such a common constituent of soils, the skeletans are usually white when dry and light grey when moist, but their colour is determined by the colour of the sand or silt. The skeletans constitute more than 15 percent of the volume of any subhorizon in which interfingering is

recognized. They are also thick enough to be obvious, by their colour, even when moist. Thinner skeletans that must be dry to be seen as a whitish powdering on a ped are not included in the meaning of 'interfingering'.

The term 'nitic properties' applies to soil material that has 30 percent or more clay, and has a moderately strong or strong angular blocky structure which falls easily apart into flat edged ('polyhedric' or 'nutty') elements with shiny ped faces that are either thin clay coatings or pressure faces. Soil materials with nitic properties have more than 0.2 percent acid oxalate extractable iron which is at least 5 percent of the dithionate-citrate-bicarbonate extractable iron.

The term 'organic soil materials' refers to soil materials which are:

- saturated with water for long periods, or artificially drained, and, excluding live roots, have,
 - (a) 18 percent or more organic carbon if the mineral fraction is 60 percent or more clay.
 - (b) 12 percent or more organic carbon if the mineral fraction has no clay, or
 - (c) a proportionnal content of organic carbon between 12 and 18 percent if the clay content of the mineral fraction is between zero and 60 percent; or
- never saturated with water for more than a few days and have 20 percent or more organic carbon.

The term 'permafrost' refers to the condition that the temperature of a soil layer is perennially at or below 0 $^{\circ}$ C.

The term 'plinthite' refers to an iron-rich, humus-poor mixture of clay with quartz and other diluents. It commonly occurs as red mottles, usually in platy, polygonal or reticulate patterns, and changes irreversibly to a hardpan or to irregular aggregates on exposure to repeated wetting and drying. In a moist soil, plinthite is usually firm but it can be cut with a spade. When irreversibly hardened, the material is no longer considered plinthite but develops into a petroferric or skeletic phase.

The term 'salic properties' refers to an electric conductivity of the saturation extract of more than 15 dS/m at 25 $^{\circ}$ C at some time of the year within 30 cm of the surface, or of more than 4 dS/m within 30 cm of the surface if the pH(H₂O, 1:1) exceeds 8.5.

The term 'slickensides' refers to polished and grooved surfaces that are produced by one soil mass sliding past another. Some of them occur at the base of a slip surface where a mass of soil moves downward on a relatively steep slope. Slickensides are very common in swelling clays in which there are marked seasonal changes in moisture content.

The term 'smeary consistence' as used in connection with Andosols, refers to thixotropic soil material, i.e. material that changes under pressure, or by rubbing, from a plastic solid into a liquefied stage and back to the solid condition. In the liquefied stage, the material skids or smears between the fingers.

The term 'sodic properties' refers to a saturation of the exchange complex by exchangeable sodium of 15 percent or more, or by exchangeable sodium plus magnesium of 50 percent or more.

The term 'soft powdery lime' refers to translocated authigenic lime, soft enough to be cut readily with a finger nail, precipitated in situ from the soil solution rather than inherited from a soil parent material. It should be present in a significant accumulation. Pseudomycelia which come and go with changing moisture conditions, are not considered as soft powdery lime in the present definition.

The term 'stagnic properties' refers to soil materials which are saturated with surface water at some period of the year or throughout the year, in most years, and which show evidence of reduction processes or of reduction and segregation of iron:

- 1. reduction conditions for part of the year or throughout the year, in most years, evidenced by one or more of the following:
 - (a) a value of rH equal to or less than 19. (rH = Eh(mV)/29 + 2pH).
 - (b) the appearance of a solid dark blue colour on the freshly broken surface of a field-wet soil sample, after spraying it with a solution in water of 1 percent potassium ferric cyanide.
 - (c) the appearance of a strong red colour on a freshly broken surface of a field-wet soil sample, after spraying it with a 0.2 percent α, α dipyridyl solution in 10 percent acetic acid.
- 2. if mottling is present, a dominant moist chroma of 2 or less on the surface of peds and mottles of higher chroma occurring within the peds or a dominant moist chroma of 2 or less in the soil matrix and mottles of higher chroma or iron-manganese concretions, or both, occurring within the soil material.
- 3. if no mottling is present, a dominant moist chroma of 1 or less on the surfaces of peds or in the soil matrix.
- 4. the dominant moist chroma on the surfaces of peds and of the soil matrix increasing with depth.

NOTE THAT for soils in which the content of iron oxides is very low, or in which iron oxides are present in such large quantities or are inert and so well crystallized that they remain brown or red even in reduced conditions, the above colour requirements do not apply.

Soils with stagnic properties key out either at the first level, as Planosols or Plinthosols, or at the second level as 'stagnic' units. Soils which

are subject to flooding or show reduction as a result of irrigation are marked by the 'inundic' and 'anthraquic' phases respectively. See Annex 3.

The term 'strongly humic' refers to soil material with more than 1.4 percent organic carbon (in g per 100 g fine earth) as a weighted average over a depth of 100 cm from the surface. This rule assumes a bulk density of 1.5 Mg/m^3 .

The term 'sulfidic materials' refers to waterlogged mineral or organic soil materials containing 0.75 percent or more sulfur (dry weight), mostly in the form of sulfides, and having less than three times as much calcium carbonate equivalent as sulfur. Sulfidic materials differ from the sulfuric horizon in their reduced condition, their pH, and the absence of jarosite mottles with a hue of 2.5Y or more or a chroma of 6 or more.

The term 'tonguing' is connotative of the penetration of an albic E-horizon into an argic B-horizon along vertical ped surfaces, if peds are present. Tongues must have greater depth than width, have horizontal dimensions of 5 mm or more in clay, silty clay and sandy clay argic B-horizons, 10 mm or more in moderately fine argic B-horizons, and 15 mm or more in coarse (silt loam, loam and sandy loam) argic B-horizons. The tongues must occupy more than 15 percent of the mass of the upper part of the argic B-horizon. With Chernozems, the term 'tonguing' refers to penetrations of the A-horizon into an underlying cambic B-horizon or into a C-horizon. The penetrations must have greater depth than width, and must occupy more than 15 percent of the mass of the upper part of the horizon in which they occur.

The term 'vertic properties' is used in connection with clayey soils which at some period in most years show one or more of the following: cracks, slickensides, wedge-shaped or parallepiped structural aggregates that are not in a combination, or are not sufficiently expressed for the soil to qualify as a Vertisol.

The term 'weatherable minerals' refers to minerals that are unstable in a humid climate relative to other minerals such as quartz and 1:1 lattice clays, and that, when weathering occurs, liberate plant nutrients and iron and/or aluminium. They include:

- clay minerals, including all 2:1 lattice clays except aluminium interlayered chlorite. Sepiolite, talc and glauconite are also included in the meaning of this group of weatherable clay minerals although they are not always of clay size.
- sand and silt-sized minerals, feldspars, feldspathoids, ferromagnesian minerals, glasses, micas, and zeolites.

DEFINITIONS OF PHASES

(Source: the Revised Legend of the Soil Map of the World, 1988, pp 60-63)

Phases indicate surface or subsurface features of the land which are not necessarily related to soil formation and (may) cut across the boundaries of soil map units. The features concerned may result from or form a constraint to land use.

An 'anthraquic phase' marks soils showing stagnic properties within 50 cm of the surface due to surface waterlogging associated with long continued irrigation, particularly of rice.

A 'duripan phase' marks soils having a subsurface horizon that is cemented by silica so that dry fragments do not slake during prolonged soaking in water or in hydrochloric acid. The upper level of the duripan must occur within 100 cm of the soil surface.

must occur within 100 cm of the soil surface.
Duripans vary in the degree of cementation by silica and they commonly contain accessory elements, mainly iron oxides and calcium carbonate. As a result, duripans vary in appearance but all of them have a very firm or extremely firm moist consistency, and they are always brittle, even after prolonged wetting.

A 'fragipan phase' marks soils having a loamy (exceptionally a sandy) subsurface horizon which has a high bulk density relative to the horizons above it, and is hard or very hard and seemingly cemented when dry, and weakly to moderately weakly brittle when moist. When pressure is applied, peds or clods tend to rupture suddenly rather than to undergo slow deformation. Dry fragments slake or fracture when placed in water. The upper level of the fragipan must occur within 100 cm of the soil surface. A fragipan is low in organic matter, slowly or very slowly permeable and often shows bleached fracture planes that are faces of coarse or very coarse polyhedrons or prisms. Clay skins may occur as patches or discontinuous streaks both on the faces and in the interiors of the prisms. A fragipan commonly, but not necessarily, underlies a B-horizon. It may be from 15 to 200 cm thick with commonly an abrupt or clear upper boundary, whereas the lower boundary is mostly gradual or diffuse.

The 'gelundic phase' marks soils showing formation of polygons on their surface due to frost heaving.

The 'gilgai phase' marks soils showing gilgai, the typical microrelief of clayey soils (mainly Vertisols) that have a high coefficient of expansion with distinct seasonal changes in moisture content. The gilgai micro-

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relief consists of either a succession of enclosed microbasins and microknolls in nearly level areas, or of microvalleys and microridges that run up and down the slope. The height of the microridges commonly ranges from a few cm to 100 cm. Rarely does the height attain 200 cm.

The 'inundic phase' marks soils with standing or flowing water present on the soil surface for more than 10 days in the growing period.

The 'lithic phase' marks soils with continuous hard rock occurring within 50 cm of the surface.

The 'petroferric phase' refers to the occurrence of a continuous layer of indurated material, in which iron is an important cement and in which organic matter is absent, or present only in traces. The indurated layer must either be continuous or, if it is fractured, the average lateral distance between fractures must be 10 cm or more. The petroferric layer differs from a thin iron pan and from an indurated spodic B-horizon in containing little or no organic matter. The upper part of the petroferric layer must occur within 100 cm of the soil surface.

The 'phreatic phase' refers to the occurrence of a groundwater table within 5 metres from the surface, the presence of which is not reflected in the morphology of the soil. Therefore, the phreatic phase is not shown, for instance, with Fluvisols or Gleysols. Its presence is important especially in arid areas where, with irrigation, special attention should be paid to effective water use and drainage in order to avoid salinization as a result of rising groundwater.

The 'placic phase' refers to the presence of a thin iron pan, a black to dark reddish layer cemented by iron, or by iron and manganese, or by an iron-organic matter complex, the thickness of which ranges generally from 2 mm to 10 mm. In spots it may be as thin as 1 mm or as thick as 40 mm, but this is rare. It may, but not necessarily, be associated with stratification in parent materials. It runs roughly parallel to the soil surface, commonly within the upper 50 cm of the soil, and has a pronounced wavy or convolute form. It normally occurs as a single pan, not as multiple sheets underlying one another, but in places it may be bifurcated. It must occur within 100 cm of the soil surface and is a barrier to water and roots.

The 'rudic phase' marks areas where the presence of gravel, stones, boulders or rock outcrops in the surface layers or at the surface makes the use of mechanized agricultural equipment impracticable. Hand tools can normally be used and also simple mechanical equipment if other conditions are particularly favourable.

The 'salic phase' marks soils which, in some horizon within 100 cm of the surface, show electric conductivity values of the saturation extract higher than 4 dS/m at 25 °C. The salic phase is not shown on soil maps for Solonchaks because their definition implies a high salt content. Salinity in a soil may show seasonal variations or may fluctuate as a result of irrigation practices.

The 'skeletic phase' refers to soil materials which consist for 40 percent or more, by volume, of coarse fragments of oxidic concretions or ironstone, or of other hard materials. This concretionary layer has a thickness of at least 25 cm and its upper part occurs within 50 cm of the soil surface. The difference from the petroferric phase is that the concretionary layer of the skeletic phase is not continuously cemented.

The 'sodic phase' marks soils which have more than 6 percent saturation with exchangeable sodium at least in some subhorizons within 100 cm of the surface. The sodic phase is not shown for soils which have a natric B-horizon or which have sodic properties since a high percentage of sodium saturation is already implied in their definition.

The 'takyric phase' applies to heavy textured soils which crack into polygonal elements when dry and form a platy or massive surface crust.

The 'yermic phase' applies to soils which have less than 0.6 percent organic carbon in the surface 18 cm when mixed, or less than 0.25 percent organic carbon if the texture is coarser than sandy loam (the organic carbon conditions are waived if the soil meets the requirements of the salic phase), and which show one or more of the following features connotative of arid conditions:

- 1. presence in the surface horizon of gravels or stones shaped by the wind or showing desert varnish (manganese oxide coating at the upper surface), or both. When the soil is not ploughed, these gravels or stones usually form a surface pavement; they may show calcium carbonate or gypsum accumulating immediately below the coarse material.
- 2. presence in the surface horizon of pitted and rounded quartz grains showing a matte surface; they constitute 10 percent or more of the sand fraction and have a diameter of 0.25 mm or more.
- 3. presence of 2 percent or more palygorskite in the clay fraction in at least some subhorizon within 50 cm of the surface.
- 4. surface cracks filled with in-blown sand or silt; when the soil is ploughed this characteristic may be obliterated. However, cracks may extend below the plough layer.
- 5. a platy surface horizon which frequently shows vesicular pores and which is generally indurated but not cemented.
- 6. accumulation of blown sand on a stable surface.

NOTES

SOIL HORIZON DESIGNATIONS

(Source: the Revised Legend of the Soil Map of the World, 1988, pp 85-90)

A soil is usually characterized by describing and defining the properties of its horizons. Abbreviated horizon designations, with a genetic connotation, are used for showing the relationships among horizons within a profile and for comparing horizons among soils. The symbols used to designate soil horizons are as follows:

Capital letters H, O, A, E, B, C and R indicate <u>master horizons</u>, or dominant kinds of departure from the assumed parent material. A combination of capital letters is used for <u>transitional horizons</u>.

Lower case letters are used as <u>suffixes</u> to qualify the master horizons in terms of the kind of departure from the assumed parent material. The lower case letters immediately follow the capital letter. Two lower case letters may be used to indicate two features which occur concurrently.

Arabic figures are used as <u>suffixes</u> to indicate vertical subdivision of a soil horizon. For A- and B-horizons the suffix figure is always preceded by a lower case letter suffix.

Arabic figures are used as prefixes to mark lithological discontinuities.

MASTER HORIZONS

<u>H-horizon</u>: organic horizon formed or forming from <u>accumulations of organic material</u> deposited on the surface, that is <u>saturated with water</u> for long periods, unless artificially drained, and contains 18 percent or more organic carbon if the mineral fraction contains more than 60 percent clay, 12 percent or more organic carbon if the mineral fraction contains no clay, or intermediate proportions of organic carbon for intermediate contents of clay.

H-horizons form at the surface of wet soils, either as thick cumulative layers in organic soils or as thin layers of peat or muck over mineral soils. Even when ploughed, the surface soil keeps a high content of organic matter following the mixing of peat with mineral material. The formation of the H-horizon is related to prolonged waterlogging unless soils are artificially drained. H-horizons may be buried below the surface.

<u>O-horizon</u>: organic horizon formed or forming from <u>accumulations of organic material</u> deposited on the surface, that is <u>not saturated with water</u> for more than a few days a year and contains 20 percent or more organic carbon.

O-horizons are the organic horizons that develop on top of mineral soils. The organic material in O-horizons is generally poorly decomposed and occurs under naturally well drained conditions.

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O-horizons do $\underline{\text{not}}$ include horizons formed by a decomposing root mat $\underline{\text{below}}$ the surface of the mineral soil. O-horizons may be buried below the surface.

<u>A-horizon</u>: mineral horizon formed or forming at or adjacent to the surface that either (a) shows an <u>accumulation of humified organic matter</u> intimately associated with the mineral fraction, or (b) has a <u>morphology acquired by soil formation</u> but lacks the properties of E- and B-horizons.

The organic matter in A-horizons is well decomposed. As a result, A-horizons are normally darker than the adjacent underlying horizons. In warm arid climates, where there is only slight or virtually no accumulation of organic matter, surface horizons may be less dark than adjacent underlying horizons. If the surface horizon has a morphology distinct from that of the assumed parent material and lacks features characteristic of E- and/or B-horizons, it is designated as an A-horizon on account of its surface location.

<u>E-horizon</u>: mineral horizon showing a <u>concentration of sand and silt</u> fractions high in resistant minerals, resulting from a loss of silicate clay, iron or aluminium or some combination of them.

E-horizons are usually eluvial horizons which underlie an H-, O- or A-horizon from which they differ by a lower content of organic matter and a lighter colour. From an underlying B-horizon, an E-horizon is commonly differentiated by colours of higher value or lower chroma, or by coarser texture, or both.

<u>B-horizon</u>: mineral horizon in which rock structure is obliterated or is only faintly evident, characterized by one or more of the following features:

- (a) an <u>illuvial concentration</u> of silicate clay, iron, aluminium, or organic matter, alone or in combinations;
- (b) a <u>residual concentration of sesquioxides</u> relative to source materials;
- (c) an <u>alteration of material</u> from its original condition to the extent that silicate clays are formed, oxides are liberated, or both, or granular, blocky, or prismatic structures are formed.

It is generally necessary to establish the relationship between overlying and underlying horizons before a B-horizon can be identified. Often, B-horizons need to be qualified by a suffix to have sufficient connotation in a profile description. Examples: a 'humus B-horizon' is designated as Bh, an 'iron B' as Bs, a 'textural B' as Bt, a 'colour B' as Bw.

 ${\underline{\tt NOTE\ THAT}}$ accumulations of carbonates, or gypsum or other more soluble salts do not by themselves distinguish a B-horizon.

 $\underline{\text{C-horizon}}$: mineral horizon (or layer) of unconsolidated material from which the solum is presumed to have formed and which does not show properties diagnostic of any other master horizons.

Traditionally, C has been used to designate parent material or rather the unconsolidated material underlying the solum that does not meet the requirements of the A, E, or B designations. Accumulations of carbonates, gypsum or other more soluble salts may be included in G-horizons.

<u>R-layer</u>: layer of <u>continuous indurated rock</u> that is sufficiently coherent when moist to make hand digging with a spade impracticable. The rock may contain cracks but these are too few and too small for significant root development. Gravelly and stony material which allows root development is considered a C-horizon.

TRANSITIONAL HORIZONS

Soil horizons in which the properties of two master horizons <u>merge</u> are indicated by the combination of <u>two capital letters</u> (for instance AE, EB, BE, BC, CB, AB, BA, AC and CA). The first letter marks the master horizon to which the transitional horizon is most similar.

Mixed horizons that consist of <u>intermingled parts</u> each of which are identifiable with different master horizons, are designated by two capital letters separated by a <u>diagonal stroke</u> (for instance A/B, B/C). The first letter marks the master horizon that dominates. Note that transitional horizons are no longer marked by suffix figures.

LETTER SUFFIXES

A small letter may be added to the capital letter to qualify the master horizon designation. Suffix letters can be combined to indicate properties which occur concurrently in the same master horizon (for example Ahz, Btg, Cck). No more than two suffixes should be used in combination. In transitional horizons, no use is made of suffixes which qualify only one of the capital letters but a suffix may be used if it applies to the transitional horizon as a whole (for example BCk, ABg).

The suffix letters used to qualify master horizons are the following:

- b. Buried or bisequal soil horizon (for example Btb).
- c. Accumulation in concretionary form; this suffix is commonly used in combination with another suffix which indicates the nature of the concretionary material (for example Bck, Ccs).
- g. Mottling reflecting variations in oxidation and reduction (for example Bg, Btg, Cg).

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- h. Accumulation of organic matter in mineral horizons (for example Bh). For the A-horizon, the h-suffix is applied only where there has been no disturbance or mixing from ploughing, pasturing or other activities of Man (h- and p-suffixes are thus mutually exclusive).
- i. Occurrence of permafrost.
- j. Occurrence of jarosite.
- k. Accumulation of calcium carbonate.
- m. Strongly cemented, consolidated, indurated; this suffix is commonly used in combination with another suffix which indicates the cementing material (for example Cmk, marking a petrocalcic horizon within a C-horizon, or Bms, marking an iron pan within a B-horizon).
- n. Accumulation of sodium (for example Btn).
- p. Disturbance by tillage practices (for example Ap).
- q. Accumulation of silica (for example Cmq, marking a silcrete layer in a C-horizon).
- r. Strong reduction as a result of groundwater influence (for example Cr).
- s. Accumulation of sesquioxides (for example Bs).
- t. Accumulation of clay (for example Bt).
- u. Unspecified; this suffix is used in connection with A- and B-horizons which are not qualified by another suffix but have to be subdivided vertically by figure suffixes (for example Aul, Au2, Bul, Bu2). The u-suffix is provided to avoid confusion with the former notations Al, A2, A3, B1, B2, B3 in which the figures had a genetic connotation. If no subdivision using figure suffixes is needed, the symbols A and B can be used without the u-suffix.
- w. Alteration in situ as reflected by clay content, colour or structure (for example Bw).
- x. Occurrence of a fragipan (for example Btx).
- y. Accumulation of gypsum (for example Cy).
- z. Accumulation of salts more soluble than gypsum (for example Az, Ahz).

Letter suffixes can be used to describe diagnostic horizons and features in a profile. For example, argic B-horizon: Bt; natric B-horizon: Btn; cambic B-horizon: Bw; spodic B-horizon: Bhs, Bh or Bs; ferralic B-horizon: Bws; calcic horizon: k; petrocalcic horizon: mk; gypsic horizon: y;

petrogypsic horizon: my; petroferric horizon: ms; plinthite: sq; fragipan: x; strongly reduced gleyic horizon: r; mottled layers: g.

NOTE THAT the use of a certain horizon designation in a profile description does not necessarily indicate the presence of a diagnostic horizon or property since the letter symbols merely reflect a qualitative estimate.

FIGURE SUFFIXES AND PREFIXES

Horizons designated by a single combination of letter symbols can be vertically subdivided by numbering each subdivision consecutively, starting from the top of the horizon (for example Bt1-Bt2-Bt3-Bt4). The suffix number always follows all of the letter symbols. The number sequence applies to one symbol only, so that the sequence is resumed in case of change of the symbol (for example Bt1-Bt2-Btx1-Btx2). A sequence is not interrupted, however, by a lithological discontinuity (for example Bt1-Bt2-Btx3).

Numbered subdivisions can also be applied to transitional horizons (for example AB1-AB2), in which case it is understood that the suffix applies to the entire horizon and not only to the last capital letter. Numbers are not used as suffixes of undifferentiated A or B symbols to avoid conflict with the previous notation system. If an otherwise unspecified A- or B-horizon must be subdivided, the u-suffix must be used.

To distinguish lithological discontinuities, an <u>arab number</u> is prefixed to the horizon designations concerned (for example A, B, 2C, 3C).

NOTES