Chapter 5
Introduction to Soils

5.1 Introduction

The term 'soil' has been derived from the Latin word '*Solum*', which means floor. Soil, according to pedologists, is a natural body of Earth material, different from all other natural bodies, possessing remarkable life-giving qualities. Dokuchaev (1900) viewed soil as a natural body composed of mineral and organic constituents, having a definite genesis and a distinct nature of its own. Jenny (1941) viewed soil as a naturally occurring body that has been evolved as a result of combined influence of climate and organisms, acting on parent material, as conditioned by relief over a period of time. It is amply evident that all soils share a number of characteristics, and have three-phase open systems (solid, liquid and gas) to which substances may be added or lost. All soils have profiles, some with more distinct horizons or layers than the others. Furthermore, soils have been considered as ecosystems with some of the main processes as biological. Roots inhabit soils, gain nourishment, remove water, encourage microbial growth and influence their environment by sloughing off the tissues and exuding chemicals. Dead leaves, roots and stem return organic matter for eventual decay, releasing inorganic nutrients for use again. Other essential biological processes include the transformations of nitrogen, phosphorus, and sulfur compounds and the mobilization of iron (Singer and Munns 1996). Land and soil are often used as synonyms. This is not true. Land, by definition, includes not only soil but also all the living organisms, the air and water bodies within or on it and the rocks below. Soil is, therefore, part of the land, and has comparatively a narrowly defined concept.

Soil is developed from parent rocks. Initially, weathering of rocks takes place which leads to the development of weathered material. The weathered material, thus developed, may remain in situ or it may be transported by fluvial, aeolian, gravitational or glacial activities. Soils are subsequently developed as a result of combined influence of climate and organisms, acting on parent material, as conditioned

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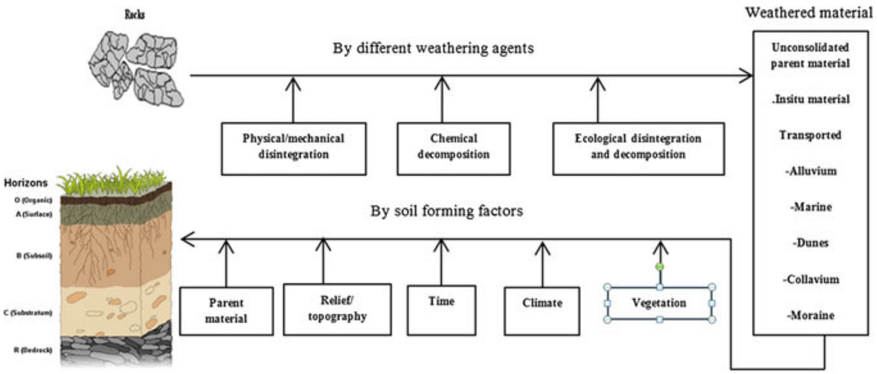


Fig. 5.1 Schematic diagram of various stages of soil formation

by relief over a period of time. If soils are developed in situ, they inherit the characteristic features of the underlying parent rock. A schematic diagram of various stages of soil formation is portrayed in Fig. 5.1.

5.2 Soil Formation

As mentioned in the previous section, soil are evolved from the parent materials under the influence of climate and flora and fauna conditioned by terrain's topography. Various stages of soil formation are discussed hereunder:

5.2.1 Weathering

Weathering is basically a combination of decomposition and synthesis. Rocks are first broken down into smaller pieces and finally into the individual constituents minerals. Simultaneously, rock fragments and minerals therein are altered by weathering forces and are transformed into new minerals—either by minor alterations or by complete chemical changes. The minerals which are synthesized and decomposed could be grouped into two categories, viz. (i) the silicate clays, and (ii) the very resistant end products including iron and aluminium oxides. During weathering two basic processes namely, disintegration and decomposition operate. Forces responsible for disintegration bring about a decrease in size of rocks and minerals without appreciably affecting their composition. By decomposition, however, definite chemical changes take place. Soluble materials are released, and new minerals are synthesized or are left as resistant end products. Mechanical and chemical processes that lead to disintegration and decomposition are listed below.

5.2.1.1 Mechanical Weathering

Variations of temperature, especially if sudden or wide, greatly influence the disintegration of rocks. During the day, rocks get heated and at night often heating much below the temperature of the air. This warming and cooling is particularly effective as a disintegrating agent. Rocks are aggregates of minerals which differ in their coefficients of expansion upon heating. With every temperature change, therefore, differential stresses which are set up eventually produce cracks and rifts, thus encouraging mechanical breakdown. Due to slow heat conduction, the outer surface of a rock is often at a markedly different temperature than the inner and more protected portions. This differential heating and cooling tend to set up lateral stresses that may cause the surface layers to peel away from the parent mass. This phenomenon is termed as exfoliation and at times is markedly accelerated by the freezing of occluded water. The presence of water, if freezing occurs, greatly increases the mechanical effects. The force developed by the freezing of water is equivalent to about 150 tonnes per square feet, an almost irresistible pressure. It widens cracks in huge boulders and dislodges mineral grains from smaller fragments.

Rainwater beats down upon the land and then travels towards ocean, continually shifting, sorting and reworking unconsolidated materials of all kinds. When loaded with such sediments, water has a tremendous cutting power as is amply demonstrated by the gorges, ravines and valleys the country over. The rounding of sand grains on an ocean beach is further evidence of the abrasion which accompanies water movement.

Ice is an erosive and transporting agency of tremendous capacity. The abrasive action of glaciers as they move under their own weight disintegrates rocks and minerals alike. Not only do glaciers affect the underlying solid rock, but also they grind and mix unconsolidated materials which have been picked up as they move over the countryside. Even though they are not so extensive at the present time, glaciers in ages past have been responsible for the transportation and deposition of parent materials over millions of hectares.

Wind has always been an important carrying agent and, when armed with fine debris, exerts an abrasive action also. Dust storms of almost continental extent have occurred in the past, with the result that tonnes of material have been filched from one section and transferred to another. As dust is transferred and deposited, abrasion of one particle against another occurs.

Plants such as mosses and lichens grow upon exposed rock. Dead mosses and lichens decompose and release organic acids which in turn help in disintegration of rocks and minerals. Higher plants sometimes exert a prying effect on rocks which results in some disintegration. This is most noticeable in the case of tree roots in rocky sections. Such influences, as well as those exerted by animals, are, however, of little importance in producing parent material when compared to the drastic physical effects of water, ice, wind and temperature changes.

5.2.1.2 Chemical Weathering

The chemical weathering encompasses the processes like hydrolysis, hydration, carbonation and other acidic processes, oxidation and solution. A brief outline of the chemical weathering is provided hereunder:

Hydrolysis: Hydrolysis is a decomposition reaction. Taking microcline ($KAlSi_3O_8$) as an example, the change in the mineral due to hydrolysis may be indicated as follows:

$$\begin{array}{ll} KA1Si_3O_8 + HOH & HA1Si_3O_8 + KOH \\ 2HA1Si_3O_8 + 8HOH & A1_2O_3 \cdot 3H_2O + 6H_2SiO_3 \end{array}$$

Hydration: Hydration involves the rigid attachment of H and OH ions to the compound target. In most cases, these ions become an integral part of the mineral crystal lattice. For instance, as micas become hydrated, some H and OH move in between the plate-like layers. In so doing, they tend to expand the crystal and make it more porous, thus hastening other decomposition processes.

The development of limonite from hematite is an example.

$$\begin{array}{ll} 2Fe_2O_3 + 3H_2O & 2Fe_2O_3 \cdot 3H_2O \\ \text{Hematite (red)} & \text{Limonite (yellow)} \end{array}$$

When the products of hydration dry out due to varying weather conditions, dehydration may occur. Thus, limonite may readily be changed to hematite, with a noticeable change in colour.

Carbonation and other Acidic Processes: The hydrogen ions present in percolating water hasten the disintegration and decomposition of mineral matter. For example, carbonic acid when it comes in contact with calcite, it results in formation of soluble bicarbonates. The chemical reaction is illustrated below.

$$\begin{array}{ll} CaCO_3 + H_2CO_3 & Ca\ (HCO_3)_2 \\ \text{Calcite} & \text{Soluble bicarbonate} \end{array}$$

In most humid region, soils and other acids much stronger than carbonic are also present. They include very dilute inorganic acids such as HNO_3 and H_2SO_4 as well as some organic acids. An example of the reaction of H ions with soil minerals is that of an acid clay with a feldspar such as anorthite.

$$\begin{array}{ll} CaA1Si_2O_3 + H & H_2A1Si_2O_8 \\ \text{Anorthite} & \text{Acid silicate} \end{array}$$

Oxidation: Oxidation occurs in well-aerated rocks where oxygen supply is high. The most important oxidation reaction is that of ferrous to ferric iron.

$$Fe^{++} \quad Fe^{+++} \quad + e^{-}$$

where e^{-} = election transfer

The change in size and charge of Fe^{++} as it is converted to Fe^{+++} form cause mineral structure to break apart.

Solution: The solvent action of water and the ions it carries as it moves through and around rocks and minerals furthers the weathering process. The release of soluble potassium by hydrolysis of orthoclase is an example. In each case, ions of the alkali metals (Na, K, etc.) or the alkaline earths (Ca, Mg, etc.) are represented as being solubilized rather readily.

5.2.2 Models of Soil Formation

There are several models of soil formations proposed by several pedologists. With advancement of the understanding of the process of soil evolution, improvements in the models have been made. Detailed discussion on soil formation model is given by Schaetzl and Anderson (2005). Dokuchaev (1883) the father of soil science, was the first person to show that soils usually form a pattern in the landscape and established that they develop as a result of the interplay of soil-forming factors, viz. parent material, climate and organisms and time which he put forward in the form of an equation:

$$\mathrm{P} = f(\mathrm{k}, \Phi, \mathrm{g}, v) \tag{5.1}$$

where P is any soil (*pochva*) or soil property k is the climate (*klimat*); Φ is organism (*organism*), g is subsoil (*gornaya poroda*), and ν is age (*vosrast*). Later he added relief as a fifth factor (Nikiforoff 1949). Wilde (1946) expanded on Dokuchaev's model and equated the soil as integral of three, somewhat reorganized soil-forming factors,

$$\mathrm{S} = \int (g, e, b)\mathrm{dt}. \tag{5.2}$$

where g is geologic parent material, e is environmental influences, b is the biological activity and t is time.

Jenny (1941) attempted to further quantify Dokuchaev's five factors model. Jenny's model represented a final *settling in* on five factors. He developed the following model, often referred to as *clorpt* model.

$$S = \int (cl, o, r, p, t\ldots) \tag{5.3}$$

where S is the soil or a soil property, *cl* is the climate factor, o is the organism factor, r is the relief factor, *p* is the parent material factor and *t* is the time factor, and the string of dots represent others, unspecified factors that may be important locally but may not be universally, such as inputs of aeolian dust, sulphate deposition in acid rain (Phillips 1999) or fire (noted that, to ascertain the role or impact a soil-forming factor on Ulrey and Graham 1993). Jenny (1941) pedogenesis, several soils may be examined in which factor is allowed to vary, while all the remaining factors are held constant. Thus he obtained the following set of equations to describe the possible scenarios and called them 'functions':

$$S = \int (\text{climate})o, r, p, t\ldots \quad \text{climofunction} \tag{5.4}$$

$$S = \int (\text{organism})cl, r, p, t\ldots \quad \text{biofunction orfloral function} \tag{5.5}$$

$$S = \int (\text{relief})cl, o, p, t\ldots \quad \text{topofunction} \tag{5.6}$$

$$S = \int (\text{parent material})cl, o, r, t\ldots \quad \text{lithofunction} \tag{5.7}$$

$$S = \int (\text{time})cl, o, r, p\ldots \quad \text{Chronofunction} \tag{5.8}$$

Based on the role of in the soil formation, soil farming factors have been grouped into two categories, viz. active factors and passive factors. Parent material, relief or topography and time are passive factors whereas climate and organisms are active factors (flora fauna) of soil formation. The role of these factors in soil formation is described hereunder.

5.2.2.1 Passive Factors

The passive soil-forming factors are those which represent the source of soil-forming mass and conditions affecting it. These provide a base on which the active soil-forming factors work or act for the development of soil. Included in this category are: parent material, relief or topography and time.

Parent Material: Parent material is that mass such as alluvium, colluvium, aeolian/loess, glacial till, etc., from which the soil has formed. Jenny (1941) defines parent material as the initial stage of the soil system. The parent material determines, within broad limits, such physical properties of a soil as texture, structure and water holding capacity. It may affect the downward movement of water for the profile development. The soils developed on sand or sandstone will have deeper and sandy profiles than those developed on granite, alluvium or loess. Some parent materials, such as sandstone and shales, develop poor agricultural soils, while those developed on alluvium and limestone usually form good agricultural soils.

Different parent materials affect profile development and produce different soils, especially in the initial stages. For example, acid igneous rocks like granite, rhyolite produce light-textured podzolic soils; basic igneous rocks like basalt, alluvium or colluvium derived from limestone or basalt produce fine-textured cracking clay soils (Vertisols), and basic alluvium or aeolian materials produce fine- to coarse-textured soils (Entisols or Inceptisols).

Relief: The relief and topography sometimes are used synonymously. The topography refers to the differences in elevation of the land surface on a board scale. They denote the configuration of the land surface. Relief may be described in terms of relative relief, drainage spacing and slope angle. The significance of topography, as a soil-forming factor, is more noticeable locally as it influences the climate and vegetation of an area. It also affects soil formation in many ways. For instance, the thickness of the soil profile is often determined by the nature of its position on the landscape. The soils on flat topography tend to be thick, but as the slope increases, so does the erosion hazard resulting in thin, stony/gravelly soils.

The soils on steep slopes are generally shallow, stony and have weakly developed profiles with less distinct horizonation due to (i) Accelerated erosion which removes surface material before it has the time to develop. (ii) Reduced percolation of water through soil because of surface run-off, and (iii) Lack of water for growth of plants which are responsible for checking erosion and promote soil formation.

Topography affects soil formation by modulating temperature and vegetative growth through slope exposures (aspect). The southern exposures (facing the Sun) are warmer and subject to marked fluctuations in the temperature and moisture. The northern exposures, on the other hand, are cooler and more humid. The eastern and western exposures occupy intermediate position in this respect.

Time: Soil formation is a very slow process requiring thousands of years to develop a mature pedon/soil profile. The period taken by a given soil from the stage of weathered rock (i.e. regolith) up to the stage of maturity is considered as time. The matured soils refer to soils with fully developed horizons (A, B, C), discussed later I this chapter. In soil formation nature works slowly. It has been reported that it takes hundreds of years to develop an inch of soil.

5.2.2.2 Active Soil-Forming Factors

The active soil-forming factors are those which supply energy that acts on the parent material resulting in soil formation. These factors are climate and vegetation.

Climate: Climate affects the soil formation both directly as well as indirectly. Directly, climate affects the soil formation by supplying water and heat to react with parent material. Indirectly, it determines the fauna and flora activities which furnish a source of energy in the form of organic matter. Organic matter acts a source for the formation of organic acids which reacts with the minerals and salts are released. Leaching and percolation of water through the soil are the two important processes in soil formation which determine soil profile characteristics. The percolation is dependent on intensity of rain, texture of the mineral material, slope and land, temperature and vegetation. On steep slopes, precipitation affects profile development by causing erosion and preventing soil development, thus resulting in thin soil cover. The deposition of the erosion products at the foot of hills and/or in piedmont plains discussed under landforms section, further interrupts with the normal profile development.

There is no horizonation (horizon development) in the hyper-arid zones with little rainfall because there is no water or weak acid solution for percolation, with the result highly soluble salts, such as sodium and/or sulphate, concentrate in the *solum.* The *solum* (plural, sola) in soil science consists of the surface and subsoil layers that have undergone the same soil-forming conditions. The base of the solum is the relatively unweathered parent material. In terms of soil horizon designations, a solum consists of A, E and B horizons and their transitional horizons and some O horizons. Included are horizons with an accumulation of carbonates or more soluble salts if they are either within, or contiguous, to other genetic horizons and are at least partly produced in the same period of soil formation.

The amount of radiation reaching the surface and soil temperature are determined largely by daily (diurnal) and seasonal fluctuations. The diurnal cycle is more significant as during the day the heat moves downwards to the soil due to warming by incoming Sun radiation and upwards during the night as the surface cools at night. Keeping in view the above considerations, Soil Survey Staff (1975) has identified soil temperature as one of the criteria for classification of different categoric levels in Soil Taxonomy. In the cold humid climates, low temperatures retard chemical reaction in soils. Furthermore, the decomposition of organic matter is slow in cold humid zones.

Living Organisms: Accumulation of organic matter, profile mixing, nutrient cycling and structural stability are all made possible by the presence of organisms in the soil. And vegetative cover reduces natural erosion rates, thereby slowing down the rate of mineral surface removal. It is obvious that the nature and number of organisms growing in and on the soil play a vital role in the kind of soil that develops.

5.2.3 *Other Soil Formation Models*

In addition to Jenny's model, there are several other models of soil evolution. A few important ones are described below. Further details can be found in Schaetzl and Anderson (2005).

5.2.3.1 Simonson's Process Systems Based Model

Jenny's model of soil formation addresses the factors that are responsible for its formation. However, it does not directly address the process that actually formed the soil. Simonson's (Simonson 1978) model was entirely process based. In the process systems based model, soil genesis is viewed as consisting of two steps: (1) The accumulation of parent material, and (2) the differentiation of that parent material into horizons. Although not originally conceived as an equation, the model can be written as

$$S = f(a, r, t_1, t_2), \tag{5.9}$$

where S is the soil, a is the additions, r is the removals or losses, t_1 is transfer/translocations and t_2 is the transformations. He envisioned that losses and additions are to the soil as a whole while translocations are losses and additions i.e. movements, between horizons and within a single *pedon (the smallest volume that can be called 'a soil'. It has three dimensions. It extends down to the depth of plant roots or to the lower limit of genetic horizon. Its lateral cross section is hexagonal and ranges from 1 to 10 m*^{2} *depending on the variability in the horizons)* (Brady 1984). These four sets of processes occur simultaneously in all soils. Their balance and character govern the actual ultimate nature of the soil (Simonson 1978).

5.2.3.2 Runge's Energy Model

Runge (1973) developed a factorial model of soil development that is somewhat of a hybrid between Simonson's process-systems model and the state factor model of Jenny. Rung emphasized *two priority* factors from Jenny's model, climate and relief, which he felt were of most importance. He combined them into a single *intensity factor* that he defined as the amount of water available for leaching (*w*), which was governed by climate and topography. The model relies heavily on gravitational energy that drives infiltrating water and in turn causes horizon development. Besides, the relies (indirectly) on radiant solar energy for organic matter production. The has been widely known as the energy model (Smeck et al. 1983):

$$S = f\ (o.w, t), \tag{5.10}$$

where S is the soil, o is organic matter production, w is water available for leaching and t is time. Each of the two energy factors is condoned by a number of capacity factors. W is conditioned by such factors as duration and intensity of rainfall, run-off versus run-on, soil permeability, evapotranspirative demand, etc. Factor o is conditioned by nutrient (especially P), air and water availability, soil fertility, available seed sources, fire, etc.

5.2.3.3 Johnsons's Soil Thickness Model

Pedogenic models generally focus on the formation of the profile, the development of horizons, the loss or degradation of those horizons. Most models assume (for the ease of comprehension) that the soil surface is static and that parent material has already been there in place. This situation does not hold good for aggrading surfaces where soil gets buried (rapid aggradation) or upbuild (slow aggradation) by addition of loess, alluvium, tephra, etc. Soil thickness is an important soil geomorphic component, and additions to or removals from the soil surface are integral part of pedogenesis. Johnsons (1985) isolated soil thickness and set about examining the processes that affect it. It is an outgrowth of Simonson's process based model that focuses on additions and removals from the soil surface.

In the model, the thickness (T) of a mineral soil is viewed as a dynamic interplay involving processes of profile deepening (D), upbuilding (U) and removals (R):

$$T = D + U + R \tag{5.11}$$

Soil gets thinner when $D + U < R$, and they get thicker when $D + U > R$, $D > U - R$ or $U > D - R$. Deepening refers to the downward migration of the lower soil boundary, generally accomplished via leaching and weathering. Upbuilding refers to allochthonous surficial additions of minerals and organic matter. Removals refers mainly to losses of material from the surface through erosion and mass wasting, although subsurface removals by through—flow, leaching and biochemical processes are also included.

5.2.3.4 Johnson and Watson-Stegner's Soil Evolution Model

The thrust of Johnson and Watson-Stegner's (1987) model is that soils *evolve*, ebb and flow, rather than unidirectionally develop and progress from 'not soil' to some theoretical, steady state end point soil evolution model. The development of the model is influenced by the Russians concept of soil profile development especially Nikiforoff (1949) whose lesser model or concept of soil evolution was based on two assumptions; (1) that soil development is continuously affected by certain progressive processes, and (2) that these processes do not operate steadily. Each process begins, peaks and then fades over time. Each soil experiences successive waves of these beginnings, peaks and endings, each for a different type of process. Soil formation models discussed so far have laid emphasis on only progressive pedogenesis, although regressive processes that simplified or regress the soil were known but essentially ignored. A model that could see both sides was clearly needed. Such a model needed to address progressive and regressive soil development as well as soil thickness concepts. In view of this, Johnson and Watson-Stegner (1987) presented their soil evolution model to address this aspect as

$$S = \int (P.R), \tag{5.12}$$

where S is the soil or a soil property, P is progressive pedogenesis and R is regressive pedogenesis. The soil evolution model stresses that soils proceed along two *interacting genetic pathways* that reflect variable exogenic/endogenic processes, factors and conditions. Every soil has a progressive pathway along which the soil 'moves forward or develops' and a regressive pathway that typifies a reversion to an earlier or simpler form. Each pathway has three components, each of which consists of two opposing vectors or sets of processes (1) horizonation/hapladoization vectors or processes, (2) retardant or developmental upbuilding vectors or processes. The progressive pathway is composed of horizonation processes, developmental upbuilding and soil deepening or thickening. The regressive pathway includes hapladoization (simplification) processes, retardant upbuilding and soil thinning.

5.2.4 Soil-Forming Processes

Soil formation is a complex process, which includes several reactions and rearrangements of matter that simultaneously affect the soil. Processes of soil formation include (1) additions of organic and mineral materials to the soil as solids, liquids, and gases, (2) losses of these from the soil, (3) translocations of materials from one point to another within the soil and (4) transformation of mineral and organic substances within the soil (Simonson 1959). Some of the important soils forming processes are given hereunder:

Eluviation: Eluviation is the process of removal of constituents in suspension or solution by the percolating water from the upper to lower layers. The eluviation encompasses mobilization and translocation of mobile constituents resulting in textural differences within the profile.

Illuviation: The process of deposition of soil materials that have been removed from the eluvial horizon in the lower layer. The lower layers refers to horizon of gains having the property of stabilizing translocated clay materials, and is termed as illuviation. The horizons formed by this process are termed as illuvial horizons (B horizons, especially Bt).

Melanization and Leucinization: It refers to changes in colour value in soil, caused by addition or losses, respectively, in the content of organic matter (the common case), or by transformations from dark-coloured (melanized) to light-coloured (leucinization) organic compounds or vice versa.

Calcification: It is the process of precipitation and accumulation of calcium carbonate ($CaCO_3$) in some part of the profile. The accumulation of $CaCO_3$ may result in the development of a calcic horizon.

Decalcification: It is the reverse of calcification that is the process of removal of $CaCO_3$ of calcium ions from the soil by leaching.

Podzolization: Podzolization encompasses the downward migration of Al and Fe, together with organic matter, from the surface layers and their accumulation in the profile's deep layers forming the Bh and Bs horizons. Furthermore, the removal of the materials from the soil surface results in the development of an eluvial horizon on the surface with intense substance losses. (http://edafologia.ugr.es/miclogia/podzolw.htm) accessed on 02–04–2016.

*Laterizatio*n: The term laterite is derived from the word '*later*' meaning brick or tile and was originally applied to a group of high-clay Indian soils found in Malabar hills of Kerala, Tamil Nadu, Karnataka and Maharashtra. It refers specifically to a particular cemented horizon in certain soils which, when dried, become very hard, like a brick. Such soils (in tropies), when massively impregnated with sesquioxides (iron and aluminium oxides) to the extent of 70–80% of the total mass, are called laterites or latosols (Oxisols). The soil-forming process is called 'Laterization or Lotozation'. Laterization is the process that removes silica, instead of sesquioxides, from the upper layers and thereby leaving sesquioxides to concentrate in the solum.

Gleization: The term '*glei*' is of Russian origin which means blue, grey or green clay. The gleization is a process of soil formation resulting in the development of a glei (or gley horizon) in the lower part of the soil profile above the parent material due to poor drainage condition (lack of oxygen) and where waterlogged conditions prevail. The process is not particularly dependent on climate (high rainfall as in humid regions) but often on drainage conditions.

Salinization: Salinization is the process of accumulation of salts, such as sulphates, chlorides of calcium, magnesium, sodium and potassium, in soils in the form of a salty (salic) horizon. As a result of the accumulation of salts, solonchaks or saline soils develop. Soils are termed saline if the electrical conductivity of its saturation extract (ECe) exceeds 4 dSm^{-1}. Such soils develop under conditions of high and brackish ground water and where evaporation losses are much more than the precipitation. The ground water containing high salts moves in an upward direction by capillary action. The water on evaporation leaves the salts behind which accumulate at the surface or at some depth depending upon the capillary fringe.

Desalinization: It is the removal, by leaching, of excess soluble salts from horizons or soil profile by ponding water and improving the drainage conditions by installing artificial drainage network.

Solonization or Alkalization: The process involves the accumulation of sodium ion on the exchange complex of the clay, resulting in the formation of sodic soils (Solonetz).

Solodization or Dealkalization: The process refers to the removal of Na^+ from the exchange sites. This process involves dispersion of clay. Dispersion occurs when Na^+ ions become hydrated. Much of the dispersion can be eliminated if Ca^+ and/or Mg^{++} ions are concentrated in the water which is used to leach the solonetz (alkali soil), as these ions (Ca, Mg) can replace the Na^+ on the exchange complex, and the salts of sodium are leached out if drainage is improved.

Pedoturbation: Pedoturbation is the process of mixing of the soil within the profile. Mixing, to a certain extent, takes place in all soils. The most common types of pedoturbation are:

- Faunal pedoturbation: It is the mixing of soil by animals, such as ants, earthworms, moles, rodents and man himself.
- Floral pedoturbation: It is the mixing of soil by plants as in tree tipping that forms pits and mounds.
- Argillipedoturbation: It is the mixing of materials in the solum by the churning process caused by swell-shrink clays as is observed in deep black cotton soils of central India.

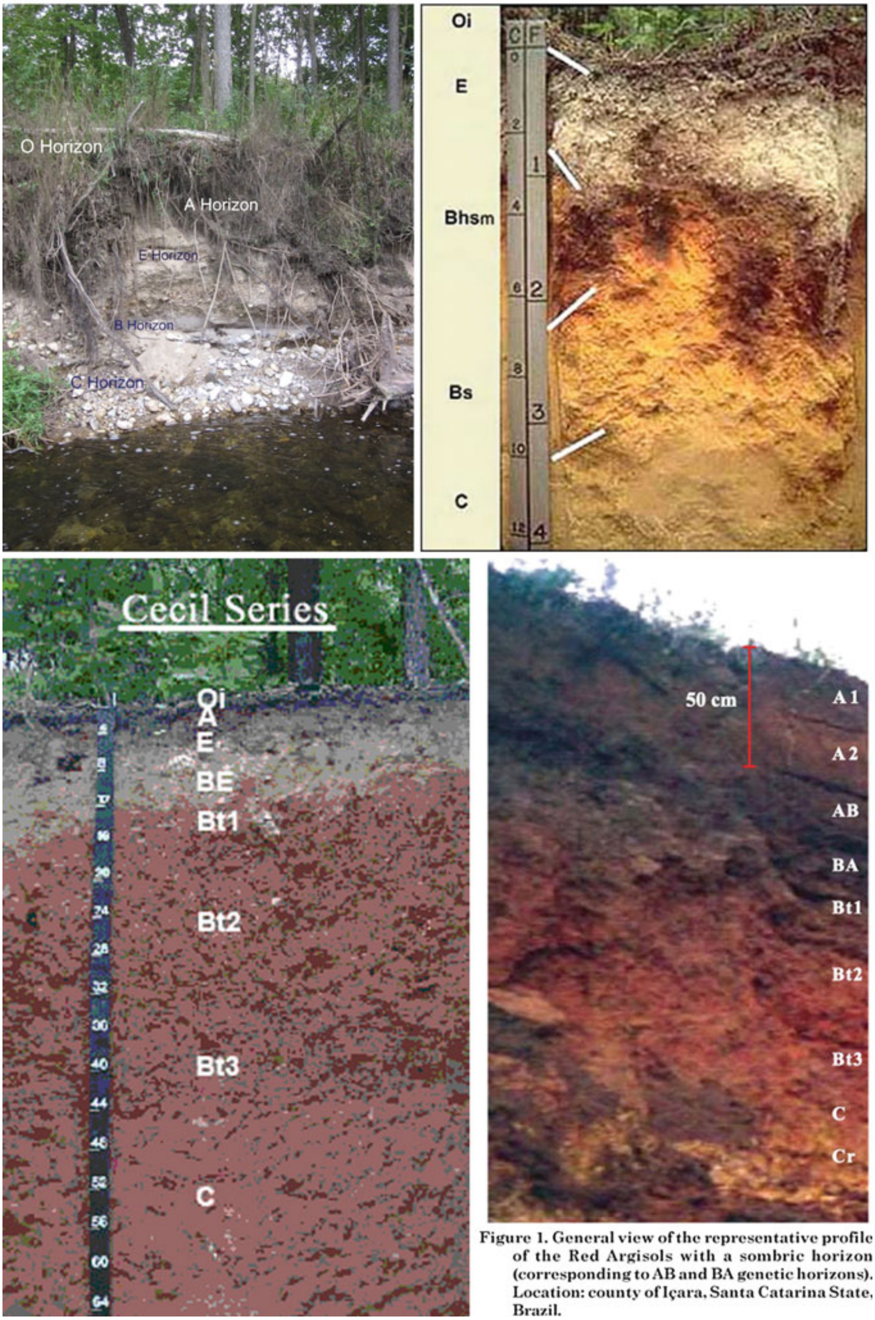
Humification: Humification is the process of transformation (i.e. decomposition) of raw organic matter into humus. It is an extremely complex process involving various organisms, such as bacteria, fungi, actinomycetes, earthworms and termites. The waxy pine needles after falling on the ground are attacked by waves of fungi breaking down complex plant compounds. First the simple compounds, such as sugars and starches, are attacked, followed by the proteins, cellulose, and finally very resistant compounds, such as tannins, are decomposed and the dark coloured substances, known as humus, are formed.

5.3 Soil Profile

Through the interactions of soil-forming processes, the soil constituents are reorganized into visibly, chemically and/or physically distinct layers, referred to as horizons. There are five major soil horizons: O, A, E, B and C. R is used to denote bedrock. A few profiles are shown in Fig. 5.2a and b and a conceptual framework of the arrangement of horizons in a profile is presented hereunder:

O Horizon: An O horizon rich in organic matter. Two main scenarios result in the formation of an O horizon: saturated, anaerobic (lack of oxygen) conditions (wetlands) or high production of leaf litter in forests. Anaerobic conditions slow the decomposition process and allow organic material to accumulate. An O horizon can have various stages of decomposed organic matter: highly decomposed, sapric; moderately decomposed, hemic; and minimally decomposed, fibric. In a fibric O layer, plant matter is recognizable (e.g. it is possible to identify a leaf). Sapric material is broken down into much finer matter and is unrecognizable as a plant part. Hemic is in between sapric and fibric, with some barely recognizable plant

(a)


O Horizon
A Horizon
E Horizon
B Horizon
C Horizon
Oi
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50 cm
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Figure 1. General view of the representative profile of the Red Argisols with a sombric horizon (corresponding to AB and BA genetic horizons). Location: county of Içara, Santa Catarina State, Brazil.

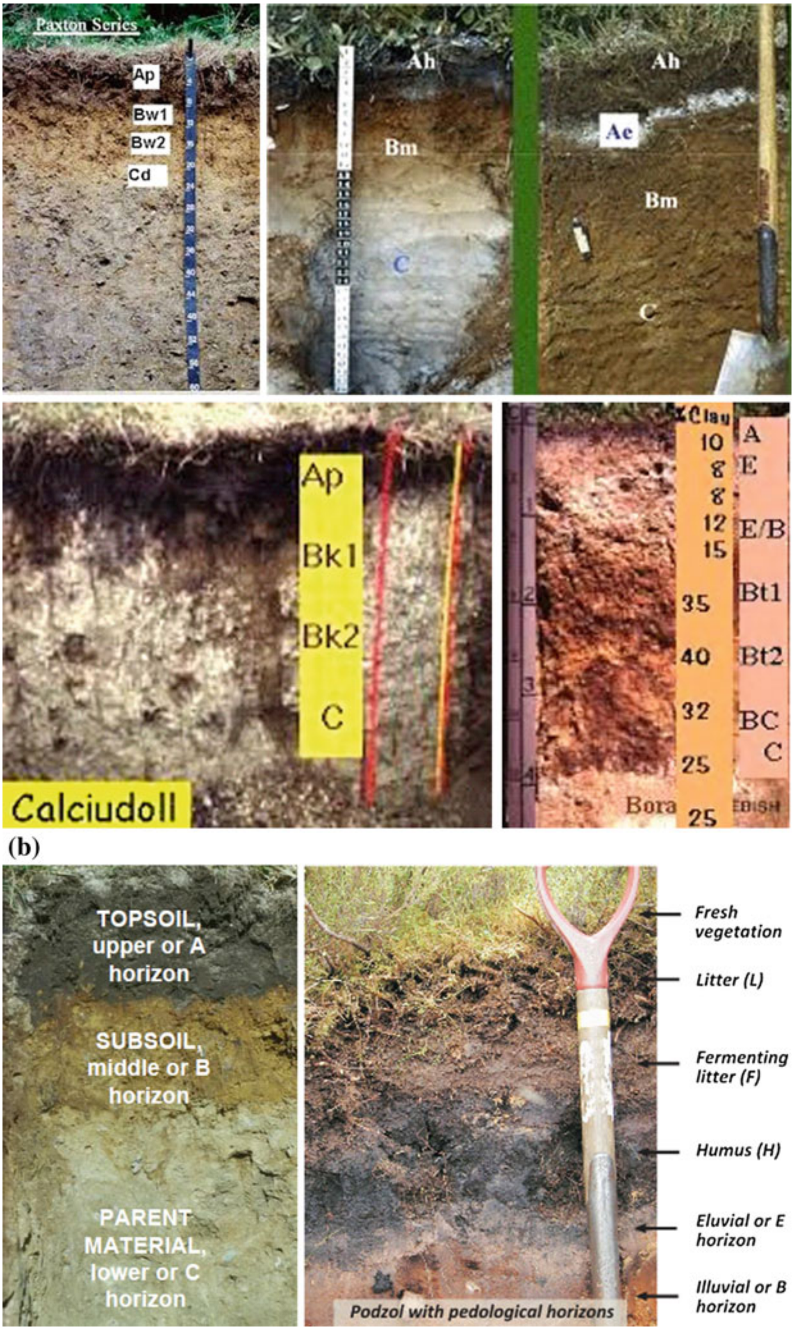


Fig. 5.2 (continued)

◂**Fig. 5.2** **a** Some typical profiles showing horizons (http://passel.unl.edu/pages/informationmodule.php?idinformationmodule=1130447032&topicorder=10&maxto=14 for Cecil soil series) 27-07-2016. There is no set order for these horizons within a soil. Some soil profiles have an *A–C* combination; some have an *O–E–B*, an *O–A–B*, or just an *O*. Some profiles may have all the horizons, *O–A–E–B–C–R*. And some profiles may have multiple varieties of one horizon, such as an *A–B–E–B*. There are some generalized concepts of how soil layers develop with time; these are expressed below, but due to the variability of natural processes over geologic time, **b** A diagram showing simplified soil horizons (*left*) and horizons in a podzol (*right*) (http://www.hutton.ac.uk/sites/default/files/files/education/soil-poster-introduction.pdf) Accessed on 27-07-2016

material present. It is possible to have multiple O horizons stacked upon one another exhibiting different decomposition stages. Because of their organic content, these horizons are typically black or dark brown in colour. The dominant processes of the O horizon are *additions* of organic matter, and *transformations* from fibric to sapric.

A Horizon: An A horizon is a *mineral horizon*. This horizon always forms at the surface. Natural events, such as flooding, volcanic eruptions, landslides and dust deposition can bury an A horizon so that it is no longer found at the surface. A buried A horizon is a clear indication that soil and landscape processes have changed some time in the past. Compared to other mineral horizons (E, B, or C) in the soil profile, they are rich in organic matter, giving them a darker colour. The A horizon, over time, is also a zone of loss—clays and easily dissolved compounds being leached out—and A horizons are typically more coarse (less clay) compared to underlying horizons (with the exception of an E horizon). *Additions* and *losses* are the dominant processes of A horizons.

- **E Horizon**: The E horizon appears lighter in colour than an associated A horizon (above) or B horizon (below). An E horizon has a lower clay content than an underlying B horizon, and often has a lower clay content than an overlying A horizon, if an A is present. E horizons are more common in forested areas because forests are in regions with higher precipitation and forest litter is acidic. However, landscape hydrology, such as perched water tables, can result in the formation of an E horizon in the lower precipitation grasslands, as seen in the profile below. The dominant processes of an E horizon are *losses*.
- **B Horizon**: A B horizon is typically a mineral subsurface horizon and is a zone of accumulation, called *illuviation*. Materials that commonly accumulate are clay, soluble salts, and/or iron. Minerals in the B horizon may be undergoing transformations such as chemical alteration of clay structure. In human modified (anthropogenic) landscapes, processes such as erosion can sometimes strip away overlying horizons and leave a B horizon at the surface. The dominant processes in a B horizon are *transformations and additions*.
- **C Horizon**: A C horizon consists of parent material, such as glacial till or lake sediments that have little to no alteration due to the soil-forming processes. Low intensity processes, such as movement of soluble salts or oxidation and reduction of iron may occur. There are no dominant processes in the C horizon; minimal additions and losses of highly soluble material (e.g., salts) may occur.

- **R Horizon**: An R layer is bedrock. When a soil has direct contact with bedrock, especially close to the soil surface, the bedrock becomes a variable when developing land use management plans and its presence is noted in the soil profile description.

5.3.1 Soil Physical Properties

Physical properties include soil colour, texture, structure, consistency, soil water bulk density.

5.3.1.1 Soil Colour

The first impression we have when looking at bare soil is of colour. Soil colour and other properties including texture, structure and consistence are used to distinguish and identify soil horizons (layers) and to group soils according to the soil classification system called *Soil Taxonomy*. Soil colour by Munsell notation is one of the many standard methods used to describe soils for soil survey. In colourimetry, the Munsell colour system is a colour space that specifies colours based on three colour dimensions: hue, value (lightness) and chroma (colour purity or colourfulness). Developed by Professor Albert H. Munsell in the first decade of the twentieth century and adopted by the USDA as the official colour system for soil research in the 1930s (http://en.wikipedia.org/wiki/Munsell_color_system).

The Munsell System allows for direct comparison of soils anywhere in the world. The system has three components: hue (a specific colour), value (lightness and darkness) and chroma (colour intensity) that are arranged in books of colour. A plate from the soil colour chart is appended as Fig. 5.3. A soil colour notation as observed in the Munsell soil colour chart, for example, is 10R5/3 corresponding to week red soil colour. Here, 10R represents hue, the hue the integers 5 and 6 represent value and chroma, respectively. Soil is held next to the chips to find a visual match and assigned the corresponding Munsell notation. Soil colour by Munsell notation is one of the many standard methods used to describe soils for soil survey. In colourimetry, the Munsell colour system is a colour space that specifies colours based on three dimensions: hue, value (lightness) and chroma (colour purity or colourfulness). It was created by Professor Albert H. Munsell in the first decade of the twentieth century and adopted by the USDA as the official colour system for soil research in the 1930s. Source: http://en.wikipedia.org/wiki/Munsell_color_system.

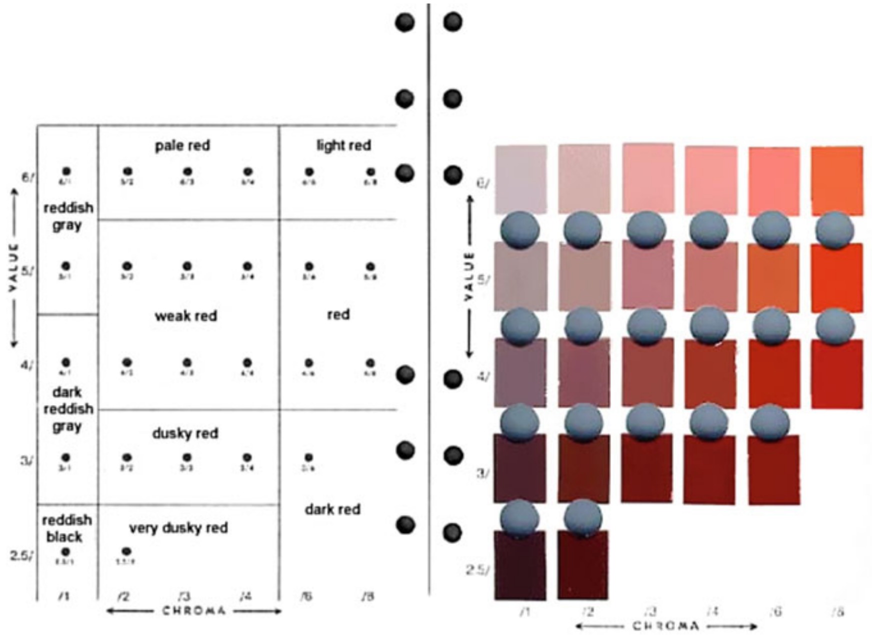


Fig. 5.3 A plate from Munsell soil colour chart showing the organization of Hue, Value and Chroma

5.3.1.2 Soil Texture

Soil texture is one of the physical properties that is used to describe the relative proportion of different sizes of mineral particles in a soil. Particles are grouped according to their size into what are called soil separates. These separates are typically named clay, silt and sand. Soil texture classification is based on the fractions of soil separates present in a soil. The soil texture triangle is a diagram often used to figure out soil textures (Fig. 5.4). In the United States, the smallest particles are *clay* particles and are classified by the United States Department of Agriculture as having diameters of less than 0.002 mm. The next smallest particles are *silt* particles and have diameters between 0.002 and 0.05 mm. The largest particles are *sand* particles and are larger than 0.05 mm in diameter. Furthermore, large sand particles can be described as *coarse*, intermediate as *medium*, and the smaller as *fine*. Other countries have their own particle size classifications. Classifications are typically named for the primary constituent particle size or a combination of the most abundant particles sizes, e.g. ‘sandy clay’ or ‘silty clay’. Another term, loam, is used to describe a roughly equal concentration of sand, silt and clay, and lends to the naming of even more classes, e.g. ‘clay loam’ or ‘silt loam’.

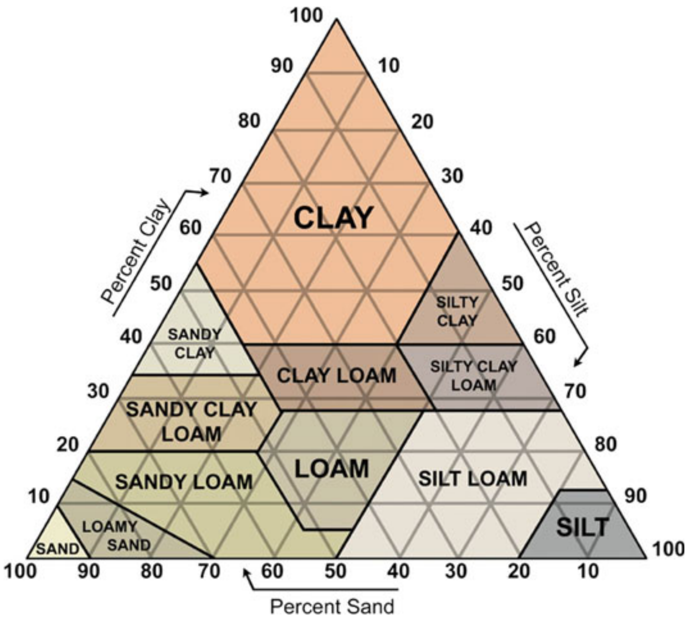


Fig. 5.4 Aternary diagram of soil texture (http://en.wikipedia.org/wiki/Soil_texture)

Another classification of soil texture, the International system, was first proposed by Atterberg (1905), and was based on his studies in southern Sweden. Atterberg chose 20 μm for the upper limit of silt fraction because particles smaller than that size were not visible to the naked eye. Commission One of the International Society of Soil Science (ISSS) recommended its use at the first International Congress of Soil Science in Washington in 1927. Australia adopted this system and according to Marshall (1947) its equal logarithmic intervals are an attractive feature worth maintaining. The USDA adopted its own system in 1938, and the FAO used the USDA system in the FAO-UNESCO world soil map and recommended its use.

5.3.1.3 Soil Water

Movement of water into the soil is controlled by gravity, capillary action and soil porosity. Within the soil system, the storage of water is influenced by several different forces. The strongest force is the molecular force of elements and compounds found on the surface of soil minerals. The water retained by this force is called hygroscopic water and it consists of the water held within 0.0002 mm of the surface of soil particles. The maximum limit of this water around a soil particle is known as the hygroscopic coefficient. Hygroscopic water is essentially non-mobile and can only be removed from the soil through heating. Matric force holds soil water from 0.0002 to 0.06 mm from the surface of soil particles. This force is due to two processes: soil particle surface molecular attraction (adhesion and absorption) to water and the cohesion that water molecules have to each other. This force declines in strength with distance from the soil particle. The force becomes nonexistent past 0.06 mm. Capillary action moves this water from areas where the matric force is low to areas where it is high. Because this water is primarily moved by capillary action, scientists commonly refer to it as capillary water. Plants can use most of this water by way of capillary action until the soil wilting point is reached. Water in excess of capillary and hygroscopic water is called gravitational water. Gravitational water is found beyond 0.06 mm from the surface of soil particles and it moves freely under the effect of gravity. When gravitational water has drained away the amount of water that remains is called the soil's field capacity (http://bettersoils.soilwater.com.au/module2/2_1.html).

5.3.1.4 Soil Structure

Structure refers to the arrangement of soil particles. Soil structure is the product of processes that aggregate, cement and compact or unconsolidated soil material. In essence, soil structure is a physical condition that is distinct from that of the initial material from which it formed, and can be related to processes of soil formation. Soil structure has been classified based on the grade, form and size of particles. The grade describes the distinctiveness of the peds (differential between cohesion within peds and adhesion between peds). It relates to the degree of aggregation or the development of soil structure. In the field a classification of grade is based on a finger test (durability of peds) or a crushing of a soil sample. The form is classified on the basis of the shape of peds, such as spheroidal, platy, blocky or prismatic. A granular or crumb structure is often found in A horizons, a platy structure in E horizons, and a blocky, prismatic or columnar structure in Bt horizons. Massive or single-grain structure occurs in very young soils, which are in an initial stage of soil development. Another example where massive or single-grain structure can be identified is on reconstruction sites. There may two or more structural arrangements occur in a given profile. The size of the particles have to be recorded as well, which is dependent on the form of the peds. The types of soil structure are described hereunder:

Granular—roughly spherical, like grape nuts. Usually 1–10 mm in diameter. Most common in A horizons, where plant roots, microorganisms, and sticky products of organic matter decomposition bind soil grains into granular aggregates.

Platy—flat peds that lie horizontally in the soil. Platy structure can be found in A, B and C horizons. It commonly occurs in an A horizon as the result of compaction.

Blocky—roughly cube-shaped, with more or less flat surfaces. If edges and corners remain sharp, we call it angular blocky. If they are rounded, they are called subangular blocky. Sizes commonly range from 5 to 50 mm across. Blocky structures are typical of B horizons, especially those with a high clay content. They form by repeated expansion and contraction of clay mineral.

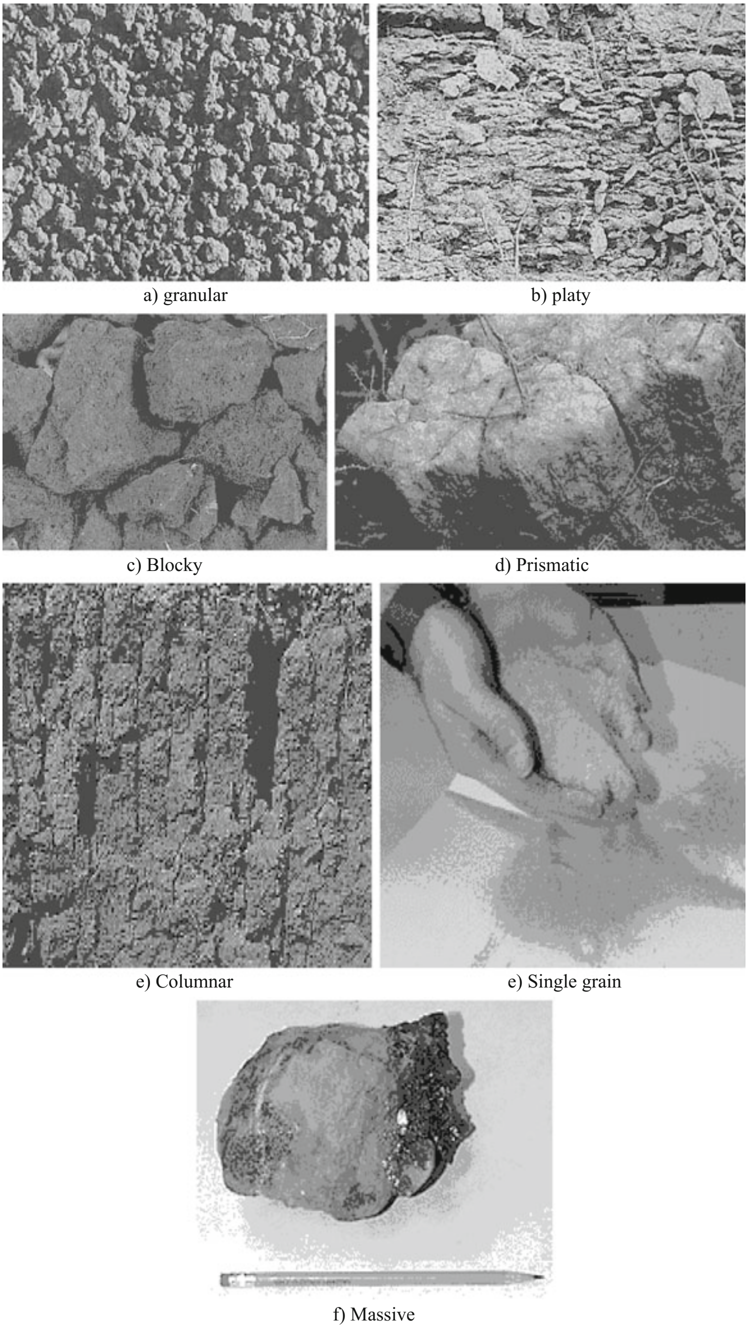


Fig. 5.5 **a** granular **b** platy **c** Blocky **d** Prismatic **e** Columnar **f** Single grain **g** Massive. Reference: http://en.wikipedia.org/wiki/Soil_structure

Prismatic—larger, vertically elongated blocks, often with five sides. Sizes are commonly 10–100 mm across. Prismatic structures commonly occur in fragipans.

Columnar—the units are similar to prisms and are bounded by flat or slightly rounded vertical faces. The tops of columns, in contrast to those of prisms, are very distinct and normally rounded.

Massive—compact, coherent soil not separated into peds of any kind. Massive structures in clayey soils usually have very small pores, slow permeability, and poor aeration.

Single grain—in some very sandy soils, every grain acts independently, and there is no binding agent to hold the grains together into peds. Permeability is rapid, but fertility and water holding capacity are low. During field check while carrying out soil survey after excavation of soil profile, soil structure is one of the soil properties that is recorded. Type, grade and sizes of soil structure are recorded. Table 5 describes the type, grade and sizes of soil texture (Tables 5.1, 5.2 and 5.3).

5.3.1.5 Soil Temperature

Soil temperature plays an important role in many processes, which take place in the soil such as chemical reactions and biological interactions. Soil temperature varies in response to exchange processes that take place primarily through the soil surface. These effects are propagated into the soil profile by transport processes and are influenced by thermal properties of soils, viz. specific heat capacity, thermal conductivity and thermal diffusivity.

The simplest mathematical representation of the fluctuating thermal regime in a soil profile is to assume that at all depths in the soil the temperature oscillates as a pure harmonic (sinusoidal) function of time around an average value (Hillel 1980). He observed that at each succeeding depth, the peak temperature is dampened and shifted progressively in time. The degree of damping increases with depth and is related to the thermal properties of the soil and the frequency of the temperature fluctuation.

Table 5.1 Classification of soil structure based on their development

Grade	Abbreviation	Description
Structureless	0	No observable aggregation or no orderly arrangement of natural lines of weakness
Weak	1	Poorly formed indistinct peds
Moderate	2	Well-formed distinct peds, moderately durable and evident, but not distinct in undisturbed soil
Strong	3	Durable peds that are quite evident in undisplaced soil, adhere weakly to one another, withstand displacement, and become separated when soil is disturbed

Table 5.2 Classification of soil structure based on their shapes

Form	Abbreviation	Description
Granular	Gr	Relatively nonporous, spheroidal peds, not fitted to adjoining peds
Crumb	Cr	Relatively porous, spheroidal peds, not fitted to adjoining peds
Platy	Pl	Peds are plate-like. The particles are arranged about a horizontal plane with limited vertical development. Plates often overlap and impair permeability
Blocky	Bk	Block-like peds bounded by other peds whose sharp angular faces form the cast for the ped. The peds often break into smaller blocky peds
Angular blocky	Abk	Block-like peds bounded by other peds whose sharp angular faces form the cast for the ped
Subangular blocky	Sbk	Block-like peds bounded by other peds whose rounded subangular faces form the cast for the ped
Prismatic	Pr	Column-like peds without rounded caps. Other prismatic caps form the cast for the ped. Some prismatic peds break into smaller blocky peds. In these peds, the horizontal development is limited when compared with the vertical
Columnar	cpr	Column-like peds with rounded caps bounded laterally by other peds that form the cast for the peds. In these peds, the horizontal development is limited when compared with the vertical
Single grain	sg	Particles show little or no tendency to adhere to other particles. Often associated with very coarse particles
Massive	m	A massive structure show little or no tendency to break apart under light pressure into smaller units. Often associated with very fine-textured soils.

The highest peak at 42° is the temperature at 1 cm, the second highest peak is the temperature at 10 cm and the lowest amplitude is the temperature at 25 cm below the soil surface. This data clearly show how damping increases with depth.

5.3.1.6 Soil Consistency

Soil consistence refers to the ease with which an individual ped can be crushed by the fingers. Soil consistence, and its description, depends on soil moisture content. Soil consistence is very important from soil tilth point of view for field operations. Terms commonly used to describe consistence are:

Table 5.3 Classification of soil structure based on their sizes

Size	Angular and subangular blocky structure [mm] diameter	Granular and crumb structure [mm] diameter	Platy structure [mm] width	Prismatic and columnar structure [mm] diameter
Very fine	<5	<1	<1 (very thin)	<10
Fine	5–10	1–2	1–2 (thin)	10–20
Medium	10–20	2–5	2–5	20–50
Coarse	20–50	5–10	5–10 (thick)	50–100
Very coarse	>50	>10	>10 (very thick)	>100

http://www.cartage.org.lb/en/themes/sciences/Earthscience/Geology/Soils/SoilMorphology/SoilMorphology/SoilStructure/SoilStructure.htm

Moist soil:

- loose—noncoherent when dry or moist; does not hold together in a mass
- friable—when moist, crushed easily under gentle pressure between thumb and forefinger and can be pressed together into a lump
- firm—when moist crushed under moderate pressure between thumb and forefinger, but resistance is distinctly noticeable

Wet soil:

- plastic—when wet, readily deformed by moderate pressure but can be pressed into a lump; will form a 'wire' when rolled between thumb and forefinger
- sticky—when wet, adheres to other material and tends to stretch somewhat and pull apart rather than to pull free from other material

Dry Soil:

- soft—when dry, breaks into powder or individual grains under very slight pressure
- hard—when dry, moderately resistant to pressure; can be broken with difficulty between thumb and forefinger

5.3.1.7 Bulk Density

Bulk density is the proportion of the weight of a soil relative to its volume. It is expressed as a unit of weight per volume, and is commonly measured in units of grams per cubic centimetres (g/cc). Bulk density is an indicator of the amount of pore space available within individual soil horizons, as it is inversely proportional to pore space:

Pore space = 1 − bulk density/particle density.

The weight per unit volume of the solid portion of soil is called particle density. Particle density is also termed as true density. The average particle density of

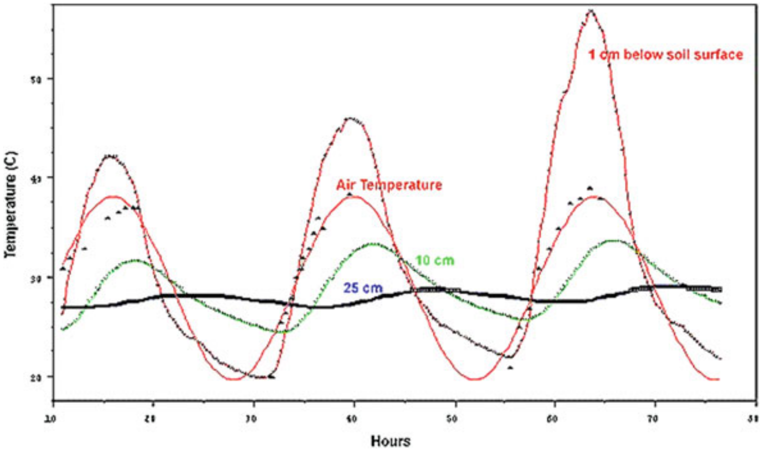


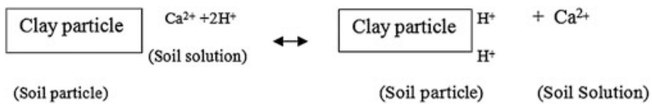
Fig. 5.6 Plots of temperature verses time were fitted with a sinusoidal function for depths of 1, 10 and 25 cm

mineral soil material is 2.65 g/cc, which approximates the density of quartz. Conversely, the average particle density of organic soil material is 1.25 g/cc. Generally particle density of normal soils is 2.65 g/cm^3. It differs from bulk density because the volume used does not include pore spaces. Particle density represents the average density of all the minerals composing the soil. For most soils, this value is very near **2.65 g/cm^3** because quartz has a density of 2.65 g/cm^3 and quartz is usually the dominant mineral. The particle density is higher if large amount of heavy minerals such as magnetite; limonite and hematite are present in the soil. With increase in organic matter of the soil the particle density decreases.

5.3.2 Soil Chemical Properties

5.3.2.1 Cation Exchange Capacity (CEC)

The ‘cation exchange capacity’, or ‘CEC’, of a soil is a measurement of the magnitude of the negative charge per unit weight of soil, or the amount of cations a particular sample of soil can hold in an exchangeable form. The greater the clay and organic matter content, the greater the CEC should be, although different types of clay minerals and organic matter can vary in CEC. The adsorbed cations are subject to replacement by other cations in a rapid, reversible process called ‘cation exchange’.



Cations leaving the exchange sites enter the soil solution, where they can be taken up by plants, react with other soil constituents, or be carried away with drainage water. It plays an important role in wastewater treatment in soils. Sandy soils with a low CEC are generally unsuited for septic systems since they have little adsorptive ability and there is potential for groundwater.

5.3.2.2 Soil Reaction (pH)

By definition, 'pH' is a measure of the active hydrogen ion (H+) concentration. It is an indication of the acidity or alkalinity of a soil, and also known as 'soil reaction'. The pH scale ranges from 0 to 14, with values below 7.0 acidic, and values above 7.0 alkaline. A pH value of 7 is considered neutral, where H+ and OH− are equal, both at a concentration of 10^{-7} mol/litre. A pH of 4.0 is ten times more acidic than a pH of 5.0. The most important effect of pH in the soil is on ion solubility, which in turn affects microbial and plant growth. A pH range of 6.0–6.8 is ideal for most crops because it coincides with optimum solubility of the most important plant nutrients. Some minor elements (e.g., iron) and most heavy metals are more soluble at lower pH. This makes pH management important in controlling movement of heavy metals and potential groundwater contamination in soil (Brady 1990).

Source: http://www.nrcs.usda.gov/wps/portal/nrcs/detail/nj/home/?cid=nrcs141p2_018993.

5.4 Soil Composition

Any given soil mixture is made up of: solid, liquid and gas. A typical soil may consist of about 50% pore space, with spatially and temporally variant proportions of gas and liquid.

5.4.1 Solid Phase

This phase contains organic and inorganic components in a complicated and generic mixture of primary and secondary minerals, organic components and salts. The solid phase consists of three main particle size fractions—s and (2–0.2 mm), silt (0.2–0.002 mm) and clay $\ll$0.002 mm)-which together govern two major soil properties: texture and structure. Soil texture and structure play a major role in soil behaviour and influence some major soil characteristics, such as drainage, Dixon and Weed (1989) fertility, moisture and erosion. The inorganic portion of the solid phase consists of soil minerals, which are generally categorized as either primary or secondary minerals. Primary minerals are components derived directly from weathering of parent materials that were formed under much higher temperatures and pressures than are found at the Earth's surface. Secondary minerals are formed by geochemical weathering processes of the primary minerals.

In general, the dominant primary minerals are quartz, feldspar, orthoclase, and plagioclase. Some layer silicate minerals, such as mica and chlorite, and ferro-magnesian silicates, such as amphibole, peroxide and olivine, also exist in soils.

The secondary minerals in soils—most of them, often termed clay minerals, are aluminosilicates, such as montmorillonite, illite, vermiculite, sepiolite, kaolinite and gibbsite. The type of clay minerals is strongly dependent on the weathering stage of the soil and can be a significant indicator of the environmental conditions under which the soil was formed. Other secondary minerals in soils are aluminium and iron oxides and hydroxides, carbonates (calcite and dolomite), sulphates (gypsum) and phosphate (apatite). Most of these minerals are relatively insoluble in water and maintain equilibrium with a water solution. Soluble salts such as halite may also be found in soil, but they are mobile in water. Clay minerals are most likely found in the fine-sized particles of the soil (clay fraction) and are characterized by relatively high specific surface areas (50–800 $m^2\ g^{-1}$). The primary minerals and other non-clay minerals are usually found in both the sand and silt portions and consist of relatively small specific surface areas $\ll 1\ m^2\ g^{-1}$).

In addition to the inorganic components in the solid phase, organic components also exist. Although the organic matter content in mineral soils does not exceed 15% (usually less), it plays a major role in soil chemical and physical behaviour (Schnitzer and Khan 1978). Organic matter is composed of decaying tissues from vegetation and micro- and macro-faunal bodies. Organic matter in soil can be found in various stages of degradation, from coarse dead to complex fine components called humus (The surface horizons of a soil profile typically contain more organic matter than do the subsurface horizons).

5.4.2 Liquid and Gaseous Phases

These phases in soils are complementary to the solid phase and occupy about 50% of the soil's total volume. The liquid consists of water and dissolved ions in various amounts. The water molecules either fill the entire pore volume in the soil ('saturated'), occupy only a portion of the pore volume ('wet'), or are absorbed on the surface area ('dry'). The composition if the soil's gaseous phase is normally very similar to the composition of the atmosphere, with the exception that the concentration of oxygen and carbon dioxide varies depending on the biochemical activity at the root zone.

5.5 Physical Setting

Since while deriving information on soils either through conventional approach or by using remote sensing data, the delineation of rock types (parent materials) and landform is the first step that is followed preceding soil mapping. An overview of parent material and landform is provided in the following sections. For detailed discussions on different types of parent material, readers may refer Brady (1980). Brady 1980. Nature and Properties of Soils. McMillan publishing co. Inc. and for landforms (Thornburry 1978; Gupta 1991, 2003).

5.6 Parent Materials

Parent material is the one wherefrom the soil is developed and, in most cases, is of geological origin. It could be organic parent material or inorganic. The upper part of the *regolith* is designated as parent material of soils. A study of weathering and of parent materials provides a basis for understanding the soil composition and is necessary introduction to soil formation and classification. In the context of the functional factorial analysis of the role of individual soil-forming factors, Jenny (1941) defined a litho-sequence as a set of soils with property differences due solely to differences in parent material, with all other factors constant. Expressed mathematically, the function is shown as:

$$\mathrm{S} = f(pm)cl,\ o,\ r,\ p, t \tag{5.7}$$

Although such a sequence or array of soils is difficult to recognize and establish in the field because of the problems in establishing that all soils in the set have property differences due solely to parent material differences without some effects from environmental differences or from differences in local landscape position. Several sets of soils have been defined as approaching this condition such that we can analyse the effects of parent material compositional differences, primarily on young and relatively simple landscape, such as in recently glaciated region. For example, while carrying out a study in Manitoba, Canada where soils are formed in glacial sediments of Mankato age (late Wisconsin, the last period of glaciations) and which varied in particle size, calcium carbonate content and mineralogical composition, Ehrlich et al. (1955) observed that the composition of parent materials had a profound effect on the type of profile developed. Further, it was observed that these differences control the soil properties to the extent that the soils are placed in different orders of Soil Taxonomy (Buols et al. 1980).

Early approaches to soil survey and classification relied heavily on interpretations of a soil's parent material, largely because soils were thought of as disintegrated rock mixed with decaying organic matter (Simonson 1952, 1959). The references of 'granitic soils', 'loessic soils' were common. With the advancements

Fig. 5.7 Examples of soil formations on continuous, hard bedrock in Tunisia (*left* GT) and transported sediments in East Africa (*right* EM), in this case, of fluvial origin. Where consolidated parent material lies close to the surface, soil depth is generally shallow and horizon development is weak. Unconsolidated sediments can completely mask the characteristics of the underlying bedrock. (http://eusoils.jrc.ec.europa.eu/library/maps/Africa_Atlas/download/14.pdf)

of our understanding about soils, there was a realization of the fact that soils developed from the same parent material may have some property/ies in common, but still can vary considerably spatially and temporarily (Phillips 2001).

Two groups of inorganic parent materials are recognized: (a) sedentary—formed in place, also known as residual, and (b) transported, which may be subdivided according to the agencies of transportation and deposition, namely *colluvium* for materials transported by gravity, and *alluvium/marine/lacustrine* for materials transported by water. Whereas materials transported by ice are known as glacial (till, moraine), those transported by wind are termed as aeolian.

5.6.1 Residual Parent Material

All loose consolidated materials that overlie bedrock are referred as *regolith*. Regolith may be either residual (formed in situ) or it may have been transported there by water, wind or gravity. Residual parent material develops in places from the underlying rock below and is rarely transported to another site. The residual regolith is termed as *residuum*. Often it is not possible to know with certainty that the residuum has not been transported at some point in the past. The mineralogy, texture, porosity, base status, etc, of the rock largely determine the nature of its residuum (Plaster and Sherwood 1971) In many instances, the relationship is intuitive and obvious. For example, sand stone will produce sandy, porous residuum; shale leads to clayey residuum. Quartzite is so difficult to weather that it produces little residuum and soils on quartzite tend to be shallow. Limestone weathers so completely that its residuum mainly forms as insoluble components that had been within the parent rock, such as clays. Basic rocks like basalt, diorite and gabbro weather to a dark, clay-rich residuum that is high (initially) in pH. Soils formed from this type of residuum tend to be more fertile and 'balanced' with respect to cations than are the soils from acidic rocks.

Granite and granitic gneiss are common coarse grained, acid crystalline rocks. Other example includes tonalite, quartz monzanite and granodiorite. These rocks form deep under the Earth crust, where slow cooling allows for the formation/development of large mineral crystals. Because of their high quartz content, acid igneous rocks tend to weather primarily by physical means, to sandy and gravelly residuum, often with low base status and poor nutrient reserves (Eswaran and Bin 1978). Typically, it has experienced long and often intense weathering. In a warm, humid climate, it is likely to be thoroughly oxidized and well leached. It is generally comparatively low in calcium because this constituent has been leached out. Red and yellow colours are characteristic when weathering has been intense as in hot humid areas.

Soils formed from the residuum from base-rich crystalline rocks tend to be dark, clayey and base rich. The low quartz content of these rocks is the reason for the general lack of sand in the saprolite and the soil profile (Schaetzl and Anderson 2005). Saprolite is in situ isovolumetrically and geochemically weathered rock that still retains the some of the original rock structure, such as strata, veins or dikes (Aleva 1983). It contains both primary minerals and their weathering products and is usually soft enough that it can be penetrated with a sharp shovel blade or knife. (Hurst 1977). In cooler and especially drier climate, residual weathering is much less intense and the oxidation and hydration of the minerals may by hardly noticeable. Also, the calcium content is higher and the colours of the debris subdued. Residual materials are of wide distribution on all Continents. A great variety of soils occur in these regions.

5.6.2 Colluvial Debris

Colluvial debris is made up of the fragments of rock detached from the heights above and carried down the slopes mostly by gravity. Frost action has a major role to play in the development of such deposits. Talus slopes, cliff detritus and similar heterogeneous materials are good examples. Avalanches are made up largely of such accumulations. Parent material developed from colluvial material is usually coarse and stony, since physical rather than chemical weathering has been dominant. At the base of slopes, the chemical weathering has been dominant resulting in medium- to fine-textured materials such as loess, medium- to fine-textured deep soils develop. However, soils developed from colluvial materials are generally not ideal for cultivation due to their unfavourable physical and chemical characteristics.

5.6.3 Alluvial Stream Deposits

The alluvial deposits, known generally as alluvium, are the material transported by fluvial activities, and vary in areal extent depending on the quantum of run off the river/steam carries and the topography of the terrain. size of there are three general classes of alluvial deposits: (a) flood plains (fluvial deposits), (b) alluvial fans, and (c) deltas.

5.6.3.1 Fluvial Deposits

A stream on a gently inclined bed usually begins to swing from side to side in variable curves, depositing alluvial material on the inside of the curves and cutting on the opposite banks. This results in *oxbows* and *lagoons*, which are ideal for the further deposition of alluvial matter and development of swamps. This state of meander naturally increases the probability of overflow at high water, a time when the stream is carrying much suspended matter. Part of this sediment is deposited over the flooded areas, the coarser near the channel, building up *natural levees*, and finer farther away in the *lagoons and slake water*. Thus, there are two distinct types of deposits—meander deposits and flood plain deposits. As might be expected, floodplain deposits are variable, ranging texturally from gravel and sands to silt and clay.

If there is a change in grade, a stream may cut down through its already well-formed alluvial deposits, leaving *terraces* above the floodplain on one or both sides. Often two or more terraces of different heights may be detected along some valleys, marking a time when the stream was at these elevations.

5.6.3.2 Alluvial Fans

Where streams descend from uplands, a sudden change in gradient sometimes occurs as the stream emerges at the lower level. A deposition of sediment is thereby forced, giving rise to alluvial fans. Fan material generally is gravelly and stony, somewhat porous, and well drained. Alluvial fan debris is found over wide areas in arid and semi-arid regions. The soils developed there from, when irrigated and properly handled, often prove very productive. In humid regions especially in certain glaciated sections, such deposit also occurs in large enough areas to be of considerable agricultural importance.

5.6.3.3 Delta Deposits

Much of the finest sediment carried by streams/rivers is not deposited in the floodplain but is discharged into the body of water to which the stream is tributary.

Unless there is sufficient current and wave action, some of the suspended material accumulates, forming a delta. Such delta deposits are by no means universal, being found at the mouths of only a small proportion of the rivers of the world. A delta often is a continuation of a floodplain, its front so to speak, and is not only clayey in nature but is likely to be swampy as well. The deltas of the Nile, Po and the Ganges rivers are good examples of these conditions.

5.6.4 Marine Sediments

Much of the sediment carried away by stream/river action is eventually deposited in the oceans, seas and gulfs, the coarser fragments near the shore, the finer particles at a distance. Also considerable debris is torn from the shoreline by the pounding of the waves and the undertow of the tides. If there have been changes in the shore line, the alternation of beds will show no regular sequences and considerable variations in topography, depth and texture. These deposits have been extensively raised above sea level along the Atlantic and Gulf coasts of the United States and elsewhere, and have given origin to large areas of valuable soils.

5.6.5 Glacial Till and Associated Deposits

The materials deposited directly by the melting ice are commonly called *glacial till*. Till is a mixture of rock debris of great diversity, especially with respect to size of particles. Boulder clay, so common in glaciated regions, is typical of the physical heterogeneity of the material. Glacial till is found mostly as irregular deposits called moraines. There are various kinds of moraines. A terminal, or end, moraine consists of a ridge like accumulation of glacial debris pushed forward by the leading glacial snout and dumped at the outermost edge.

The ground moraine, a thinner and more level deposition laid down as the ice front retreated rapidly, is of much more importance. It has the widest extent of all glacial deposits and usually possesses a favourable agricultural topography. Associated with the moraine in certain places are such special features as kames (conical hills or short ridges of sand and gravel deposited by ice), eskers (long narrow ridges of coarse gravel deposited by ice-walled streams coming from the glacier) and drumlins (cigar-shaped hills composed of till and oriented parallel to the direction of ice movement).

A *glacial till* is unsorted sediment deposited directly by glacial ice. Glacial till is a soft rock identified by large angular rock fragments on the surface and within the soil. Because of their huge mass, ice sheets flow outward as if they were huge piles of peanut butter.

5.6.6 Outwash Plains

The outwash plain is formed by streams heavily laden with glacial sediment. This sediment is usually assorted and therefore variable in texture. Such deposits are particularly important in valleys and on plains, where the glacial waters were able to flow away freely. These valley fills are common in the United States, both north and south of the terminal moraine.

5.6.7 Glacial Lake (Lacustrine) Deposits

In many cases, the ice front came to a standstill where there was no such ready escape for the water, and ponding occurred as a result of damming action of the ice. Often very large lakes were formed that existed for many years. Important amongst them are those south of Great Lakes in New York, Ohio, Indiana, Michigan and in the Red River Valley.

The glacial deposits in these glacial lakes range from coarse delta materials near the shore to find silts and clay in the deeper and stiller waters. As a consequence, the soils developed from these lake sediments area most heterogeneous. Even so, large areas of fertile fine-textured soils have developed from these materials. Because of climate differences, weathering has been variable, and profile contrasts are great. Extending westward from New England along the Great Lakes to the broad expanse of the Red River Valley, these deposits have produced some of the most important soils of the northern states. In the intermountain regions of the United States, they have given rise to agriculturally important soils, especially when irrigated.

5.6.8 Glacial Aeolian Deposits

During the glaciations, much fine material was carried far away the front of ice sheets by streams that found their source within the glaciers. This sediment was deposited over wide areas by the overloaded rivers. When added to the great stretches of unconsolidated till in the glaciated regions in the residual material devoid of vegetation on the Great Plains, these sediments presented unusually favourable conditions for wind erosion in dry weather.

5.6.9 Loess

This loess, wind-blown material, was deposited in the uplands, the thickest deposits being found where the Valleys were widest. The fine material high in silt sized particles covered existing soils and parent materials, both original and glacial in origin.

5.7 Landforms

Like parent material, landform/relief is one of the five factors of soil formation. In fact, the catena concept studying soils along a slope is one of the simplest yet most elegant ways to discern spatial interrelationships between soils and topography. A catena is transact of soils from top of the base of a hill, perpendicular (or nearly so) to the contour lines. Its name comes from the Latin *catena,* chain.

The dynamic interrelationship between physiography and soils is utilized while deriving information on soils from aerial photographs/digital space-borne images with stereo capability. The guiding principle has been that soils are the product of the same natural processes and conditions that sculpture the land they dwell in. However, this does not imply that any given physiographic unit will contain a single class of soils; but that the soils within the physiographic unit normally vary within a certain range. While the identification of specific soil series and types can be established only with extensive field work, mapping of certain phases like slope, erosion, stoniness and drainage can be accomplished with a minimum of field checking (Leuder 1959).

The various landforms are described in detail in standard works on geomorphology (e.g. Bloom 1978; Thornbury 1978; Short and Blair 1986). As landforms are directly observed on remote sensing data products, it is important that the image interpreter must have a sound knowledge of geomorphological principles and processes. Geomorphological applications of remote sensing, particularly aerial photography have been reviewed by Tator (1960), Miller and Miller (1961), Ray (1965), Verstappen (1983) and Von Bandat (1983). A detailed account of how orbital remote sensing is useful in deriving information on landforms is given by Gupta (1991, 2003). An introduction to major landforms is presented hereunder.

5.7.1 Tectonic Landforms

Tectonic landforms may be defined as structural landforms of regional extent. W.M. Davis in 1899 considered that structure, processes and time constitute the three most significant factors shaping the morphology of a land. Of the three, structure, i.e. the deformation pattern, has the most profound control. In almost all cases, the structure of the rock has an intrinsic influence on landforms due to selective differential erosion and denudation along structurally weaker zones. Everett et al. (1986) provide numerous examples.

5.7.2 Volcanic Landforms

Volcanic landforms are primarily constructional, and result from extrusion of magma along either vent centres or fractures on Earth's surface. Central-type neo-volcanic eruptions are confined to plate boundaries, most being concentrated on the convergent margins around the Pacific Ocean. They result in landforms such as conical mountains. Fissure-type eruptions create sheets of flows forming plateaus. Basaltic weathered surfaces are frequently marked by black cotton soil; and high-density dendritic, trellis and rectangular drainage patterns. Short (1986) has reviewed and classified the various volcanic landforms types.

5.7.3 Fluvial Landforms

Running water is one of the most prominent agents of landforms sculpturing, whose effects could be seen almost everywhere. Huge quantities of sediments or rock material are removed, transported from one place to another and dumped by rivers, thus modifying the land surface configuration (Baker 1986). The fluvial landscape comprises valleys, channel ways and drainage networks. The landforms associated with fluvial erosion are gorges, canyons, V-shaped valleys, steep hill slopes, waterfalls, pediments, etc. A *canyon* is a deep ravine between pairs of escarpments or cliffs and is the most often carved landscape by the erosive activity of a river. A *pediment* is a very gently sloping (0.5°–7°) inclined bedrock surface. It typically slopes down from the base of a steeper escarpment but may continue to exist after the mountain has eroded away. Typical depositional landforms include fans, cones, alluvial plains, flood plains, natural levees, river terraces, meanders scars, channel fills, point bars, back swamps and deltas. A *levee* is an elongated naturally occurring ridge which regulates water levels. It is usually earthen and often parallel to the course of a river in its floodplain.

A *terrace* is a step-like landform consists of and a flat or gently sloping geomorphic surface, called a tread, that is typically bounded one side by a steeper ascending slope, which is called a 'riser' or 'scarp'. A meander scar, occasionally meander scarp, is formed by the remnants of a meandering water channel. They are characterized by a crescentic cut in a bluff or valley wall, produced by a meandering stream. A *point bar* is a depositional feature made of alluvium that accumulates on the inside bend of streams and rivers below the slip-off slope. They are crescent-shaped and located on the inside of a stream bend. A *backswamp* is the section of a floodplain where deposits of fine silts and clays settle after a flood. Backswamps usually lie behind a stream's natural levees. Depending upon dimensions involved, the landforms can be identified on satellite images. Stereoscopic analysis of remote sensing data is of great help in studies.

5.7.4 Coastal and Deltaic Landforms

The oceans cover a major part of the Earth and surround the continents. A coastline is the boundary between land and ocean. In a general sense, the coast refers to a zone of indefinite width on both sides of the coastline. Coastal landforms are those which are influenced and are controlled by proximity to the sea. Several types of coastal erosional landforms, such as cliffs, terraces, benches, shelves, caves, islands etc., and depositional landforms, such as beaches, spits, bars, tidal, flats and deltas, can be identified on aerial photographs and satellite images, depending upon the dimensions involved and scale provided by the sensor. Selected examples are given by Bloom (1986) and Coleman et al. (1986).

5.7.5 Aeolian Landforms

Landforms in deserts developed by the erosion, transportation and deposition activities of the e wind action. The various landforms can be distinguished on the basis of shape, topography and pattern (see e.g. Walker 1986). Erosional landforms include yardung, blowouts, desert pavemnt desert varnish, loess deposits. Yardang is usually elongated in the direction of the prevailing winds and is nearly always carved from relatively weak material. Blowouts are common in areas of sand accumulation where they form small basins on or within dunes and other types of sand accumulation. In the process of removal of sand and other small-sized particles by deflation, there is a sorting of materials according to sized with the coarser materials left behind These concentration of pebbles and boulders have been designated by the general name of lag deposits. *Desert pavement* and *desert armour* are terms often applied to them. The areas covered with large-sized rocks are called *hamadas*.

True dunes have been defined as a heap of sand whose existence is independent of either ground form or fixed wind obstruction (Bagnold 1933). Barchan or crescentic dunes—A barchans is crescentic–shaped dune with the tips extending to the leeward, making this side concave in plan and the winward side convex. Sief or longitudinal dunes are parallel to the prevailing wind. Transverse dunes, including common barchans, are nearly always free of vegetation. The tips of a transverse dune extend to the leeward (Hack 1941). Parabolic dunes have been defined as 'long scoop-shaped hallows or parabolas, of sand with points tapering to windward'(Hack 1941). Loess—the term is applied to wind-blown silt which commonly are buff-coloured, non-indurated, calcareous, permeable particularly in the vertical direction and consists of angular to subangular particles of quartz, feldspar, calcite, dolomite and other minerals held together with montmorillonite binder sand sheets or more commonly called sand drift is applied to a sand area marked by an extremely flat surface and absence of any topographic relief other than ripples.

Loess deposits are homogenous non-stratified and unconsolidated wind-blown silt. They are susceptible to gullying and may develop pinnate and dendritic drainage patterns. Dry loess slopes are able to stand erect and form steep topography. Aeolian deposition leads to sand sheets, various types of dunes such as crescent dunes, linear dunes, star dunes, parabolic dunes, and complex dunes and ripples. Other landforms in deserts could be due to fluvial activity, such as fans, dry river channels and lakes. Desert lakes (playas) are generally salty, shallow and temporary, and constitute sources of mineral wealth such as salts formed by evaporation.

5.7.6 Glacial Landforms

Glaciers are stream-like features of ice and snow, which move down slopes under the action of gravity. Glaciers occur at high altitudes and latitudes, and about 10% of the Earth's land surface is covered with glacial ice. The areal extent of glaciers is difficult to measure by field methods, and remote sensing data images provide information of much practical utility in this regard (Gupta 2003). Further, multispectral data can help delineate different zones in a glacier (Hall and Martinee 1985; Williams 1986).

Typical erosional landforms of glacial origin are broad U-shaped valleys, hanging valleys, fords, cirques and glacial troughs. A cirque (French, from the Latin word circus) is a theatre-like valley formed by glacial erosion. Alternative names for this landform are corrie (from Scottish Gaelic coire meaning a pot or cauldron) and cwm (Welsh for 'valley', pronounced coom). Glacial troughs, or glaciated valleys, are long, U-shaped valleys that were carved out by glaciers that have since receded or disappeared. Troughs tend to have flat valley floors and steep, straight sides. The huge moving masses of ice and snow erode and pick up vast quantities of fragmental material and transport these varying distances before deposition. The glacial deposit is typically heterogeneous, consisting of huge blocks to fine silt or rock flour, and is called till matrix. The depositional landforms include moraines, drumlins, till, glacial drift, etc. Below the snow line (line of perpetual snow), the ice melts and gives rise to streams. Drumlins are elongated, teardrop-shaped hills of rock, sand and gravel that formed under moving glacier ice. They can be up to 2 km long. In geology, drift is the name for all material of glacial origin found anywhere on land or at sea, including sediment and large rocks (glacial erratic). Glacial origin refers to erosion, transportation and deposition by glaciers.

In this region, up to a certain distance downstream the landforms have characteristics with both fluvial and glacial properties, and they are called fluvioglacial. Typical fluvioglacial landforms include outwash plains, eskers, fans and deltas and glacial lacustrine features. Broadly, glacial landforms produce gently rolling or hummocky topography with a deranged or kettle-hole drainage pattern. Images exhibit a mottled pattern due to varying soil moisture and the presence of a large number of ponds and lakes.

5.8 Soil Classification

Unlike plants and animals, which can be identified as separate entities, the world's soil cover is a continuum. Its components occur in temporal and/or spatial successions. Soil classification addresses the grouping of soils with a similar range of properties (chemical, physical and biological) into units that can be geo-referenced and mapped. The many soil classification schemes developed over the years reflect different views held on concepts of soil formation and mirror differences of opinion about the criteria to be used for classification. In addition, emphasis shifted away from the genetic approach, which often contained an element of conjecture, to the use of soil *properties* as differentiating criteria. By and large, consensus evolved as to the major soil bodies which needed to be distinguished in broad level soil classification.

A large number of countries have developed their own soil classification systems, e.g. US (Soil Survey Staff 1975), Russian (Shishov et al. 2001), France (Baize and Girard 1990), Australia (Isbell 1996). and Brazil (Embrapa Solos 2006), each with their foci and structure peculiar to them. Whereas the Russian classification system lays emphasis on climate and ecological factors, the French classification system emphasizes pedogenic processes. The American classification system, on the other hand, uses quantifiable soil properties resulting from pedogenic processes as controlled by the factors of soil formation although differences in definitions and terminology remained.

The soil classification has been evolved through several stages with the progress of our comprehension about soil genesis and soil properties. Early soil classification systems, e.g. Russian, USDA (Baldwin et al. 1938) focused on the environment and the soil-forming factors to classify soils in zonal soils in which the pedogenesis was mainly determined by climate and vegetation and azonal and intrazonal in which pedogenesis was mainly determined by parent material and time of development. The difference between azonal and intrazonal soils was made on the basis of soil profile development. Subsequent development focused on the processes occurring in the soil itself (such as ferallitization, salinization, leaching and accumulation, etc.). These processes were roughly characterized by soil properties. A good example of the latter approach is the French classification system (CPCS 1967). Modern soil classification started with the publication of the 7th Approximation of the USDA Soil Taxonomy (Soil Survey Staff 1975), where precisely defined and quantified soil properties as such, or in combination, were used to define 'diagnostic soil horizons'. Postmodern soil classification approaches make extensive use of statistics and fuzziness and include numerical soil classification systems (http://www.fao.org/soils-portal/soil-survey/soil-classification/en/). Accessed on 25-12-2015.

5.8.1 USDA Soil Classification System

Soil Taxonomy, the new U.S. system of soil classification, is an attempt at a comprehensive classification of soils. Initiated by the Soil Conservation Service of the United State Department of Agriculture in late fifties, the system went through a series of approximations of which the 7th Approximation was published in 1960 (Soil Survey Staff 1960). After substantial revisions, it was later published in 1975 as a book titled '*Soil Taxonomy*': *A Basic System of Soil Classification for Making and Interpreting Soil Surveys'* (Soil Survey Staff 1975). Like most of the taxonomic systems, Soil Taxonomy is a multi-categoric system. Each category is an aggregate of taxa, defined at about same level of abstraction, with the smallest number of classes in the highest category and the largest number in the lower category. In order of decreasing rank, these categories are: order, suborder, great group, subgroup, family and series.

Based upon the presence or absence of a variety of combinations of diagnostic horizons and soil properties, order, suborder and great group have been differentiated. The diagnostic horizons such as mollic epipedon or spodic and oxic horizons are used as differentae (distinguishing/differentiating criteria) at the order level. Soil moisture regimes and extreme chemical or mineralogical properties, such as the presence of large amount of allophane are examples of criteria used for differentiating suborders. Properties that appear to be superimposed on the diagnostic features of the orders and suborders, such as various kinds of pans or the presence of plinthite are used to differentiate great groups.

The next category in order of decreasing rank is subgroups which are subdivisions of great groups, representing either the central concepts of the category, intergrades to other groups, or the extragrades that have additional aberrant properties. Families and series are distinguished on the basis of properties selected to create taxa that are successively more homogenous for practical uses of soils. Thus, families are intended to provide classes having relative homogeneity in properties important to plant growth. Series are subdivisions of families intended to give the greatest homogeneity of properties in the rooting zone, consistent with the occurrence of mappable areas at scales of detailed soil survey. For further details readers may refer Soil Taxonomy (Soil Survey Staff 1975) and 1st–12th editions of Keys to Soil Taxonomy (Soil Survey Staff 1983, 1985, 1987, 1990, 1992, 1994, 1996, 1998, 2003, 2006, 2010, 2014).

5.8.1.1 Structure of Soil Taxonomy

Soil Taxonomy has six categories and includes all the currently recognized soil series of the United States as well as soils in other parts of the world that have been sufficiently described and characterized. The six categories are discussed hereunder:

Orders: In the first edition of Soil Taxonomy (Soil Survey Staff 1975) only 10 soils orders, namely Entisols, Inceptisols, Histosols, Alfisols, Mollisols, Aridisols, Ultisols, Spodosols, Oxisols and Vertisols were recognized. Andisols, the 11th order was added in 1990 in the 11th edition of Soil Taxonomy (Soil Survey Staff 1990) whereas the twelfth order Gelisols was introduced in 1998 (Soil Survey Staff 1998). Entisols and Inceptisols exhibit minimum degree of development of horizons. Vertisols, Aridisols, Histosols, Mollisols, Spodosols, Alfisols and Ultisols represent differences in the dominant kinds of genetic horizons. Oxisols represent a combination of both the kind and degree of weathering and soil formation. Halomorphicand hydromorphic soils are not classified in separate orders but are distributed according to other characteristics thought to be more important in a comprehensive scheme.

Suborders: This is the second highest category in the hierarchical system of Soil Taxonomy. The differentae/differentiating characteristics vary, but most tend to emphasize similar moisture and temperature regimes, with closely associated natural vegetation. For example, in case of Vertisols four subgroups, namely, Aquerts Uderts, Usterts, Torrerts and Xererts have been made based on soil moisture regime. These suborders have aquic, udic, ustic, torrid and xeric moisture regime.

Great Groups: In this category the major emphasis is on the kind and arrangement of diagnostic horizons, except in Entisols, which have no distinctive horizons. For example, within suborder Aquerts8 Great groups, namely.

Sulfaquerts, Salaquerts, Duraquerts, Natraquerts, Calciaquerts, Dystraquerts, Epiaquertsand, Endoaquerts have been made. The great groups have been further divided into subgroups.

Subgroups: The following three kinds of subgroups are defined;

Typic subgroup: There is a typical or central concept for each group. This is the *typic* subgroup.

Intergrade; subgroups are transitional to other orders, subgroups or great groups.

Extragrade subgroups have properties that are not representative of the great group but that do not indicate transition to any other known kind of soil. For example, within Dystraquerts great group 8 subgroups, namely Sulfaqueptic *Dystraquerts*, Aridic *Dystraquerts*, Ustic, Aeric, Leptic, Entic, Chromic and Typic subgroups have been identified.

Families: Families are a group of soils within each subgroup that have similar chemical and physical properties that affect their responses to management and manipulation for use. Families are defined by a number of properties, the most common of which are;

- Particle size distribution in the horizons of major biological activities below plough depth (the 'family control section').
- Mineralogy of the same horizon that is considered in naming particle size classes; and
- Soil temperature regime.

Other characteristics, such as soil depth, content of polysulphides, and the like are applied if they are important in the particular subgroup.

Series: It is the lowest category in the system. The differentae are mainly the same properties used to define classes in higher categories, but with much narrower ranges. Soil series, like families, are used mainly for practical purposes, and the taxa in both of these categories are closely related to interpretative applications of the system. The typical profiles—one representing each soil order—are appended as Plates 5.1, 5.2, 5.3, 5.4, 5.5, 5.6, 5.7, 5.8, 5.9, 5.10, 5.11, 5.12.

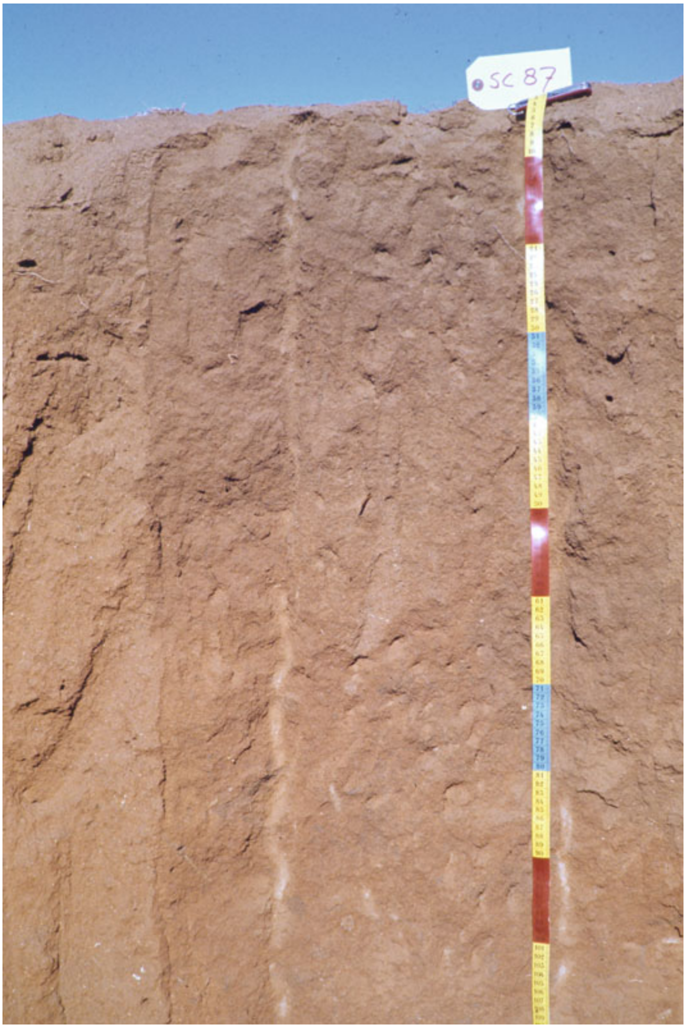


Plate 5.1 Siliceous hyperthermic, coarse-loamy Psammentic Haplustalf (Soil Survey Staff 1999) from Shoshong, Botswana (*Courtesy* ISRIC)

Plate 5.2 Acrudoxic Ultic Hapludand (Soil Survey Staff 1999) from Sumatra, Republic of Indonesia (*Courtesy* ISRIC)

Plate 5.3 Typic Natrargid (Soil Survey Staff 1999) from Garissa district, Kenya (*Courtesy* ISRIC)

Plate 5.4 Typic Torripsamment (USDA-NRCS 1999) from China, Xinjiang Autonomous region (*Courtesy* ISRIC)

Plate 5.5 Histosol (Soil Survey Staff 1999) from The Netherlands (*Courtesy* ISRIC)

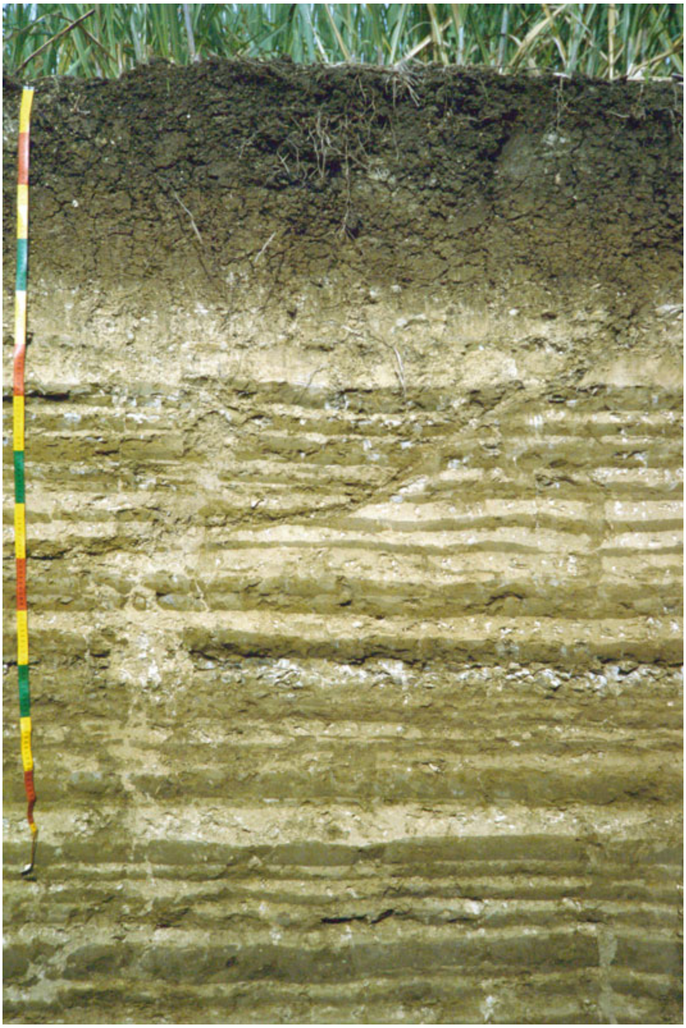


Plate 5.6 Clayey montmorillonitic (calcareous) isohyperthermic Eutropept (Soil Survey Staff, 1975) from Cuba, Provincia Santiago de Cuba (*Courtesy* ISRIC)

Plate 5.7 Mollisols from South Africa, Kwazulu Natal (*Courtesy* ISRIC)

Plate 5.8 Typic Kandiudox (Soil Survey Staff 1999) from Rwanda, Nkanga, 62 km south from Kigali (*Courtesy* ISRIC)

Plate 5.9 Ultic Haplorthod (Soil Survey Staff 1999) from Ireland, 16 km W of Athlete (*Courtesy* ISRIC)

Plate 5.10 Typic Palehumult (Soil Survey Staff 1999) from Kenya, Embu (*Courtesy* ISRIC)

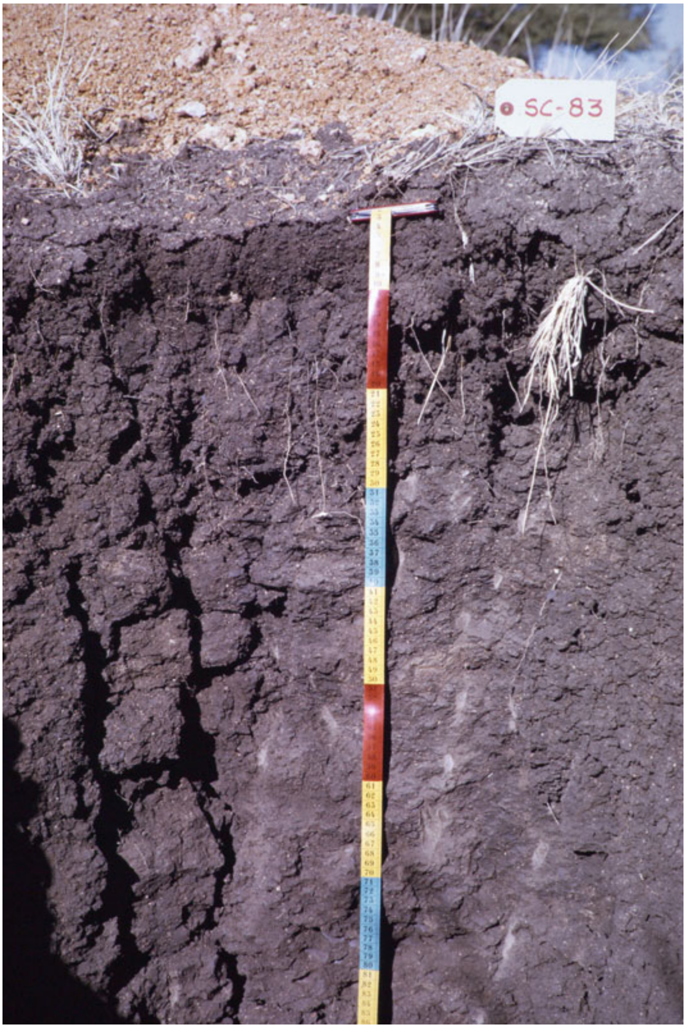


Plate 5.11 Montmorillonitic hyperthermic fine Aridic Haplustert (Soil Survey Staff 1999) from Botswana, Shoshong (*Courtesy* ISRIC)

Plate 5.12 Mixed cryic loamy nonacid Typic Aquiturbel (Soil Survey Staff 1999), from Alaska, USA (*Courtesy* ISRIC)

5.8.2 The FAO-UNESCO Soil Classification System

In order to prepare a soil map of the world, FAO developed a supranational legend in 1974 which has been used as an international soil classification system FAO-UNESCO (1974). The major objectives of the system were to: (1) Make a first appraisal of the world's soil resources, (2) Provide a scientific basis for the transfer of experience between areas with similar environments, (3) Promote the establishment of a generally acceptable soil classification and nomenclature and (4) Establish a common framework for more detailed investigations in developing areas. Many of the names used in that classification are known in many countries and do have similar meaning. The FAO/UNESCO legend is a very simple classification system with very broad units. It was the first truly international system, and most soils could be accommodated on the basis of their field descriptions. The legend of the Soil Map of the World is not meant to replace any of the national classification schemes but to serve as a common denominator' FAO-UNESCO (1974).

Initially, the legend to the SMW consisted of 26 ('first level') 'Major Soil Groupings' comprising a total of 106 ('second level') 'Soil In Units'. 1990, a 'Revised Legend' was published and a third hierarchical level of '*Soil Subunits'* was introduced to support soil inventory at larger scales. Soil subunits were not defined as such but guidelines for their identification and naming were given. De facto this converted the SMW map legend, with a finite number of entries, into an open-ended, globally applicable 'FAO-UNESCO Soil Classification System'. (http://www.fao.org/soils-portal/soil-survey/soil-classification/fao-legend/en/. Accessed on 24–12–2015). A major revision of the system was published in 1988. This system was finally replaced by the World Reference Base for Soil Resources in 1998.

The legend is described by its authors as a 'monocategorical classification,' but is presented as a two level hierarchical system of 26 first level classes ('soil units') and 106 second level classes with three kinds of textural phases, three slope phases, and twelve management phases. A great deal of generalization was required to correlate the diversity of classification systems and scales of mapping to one system. The map's scale for this system is 1:5,000,000 and is equally general in detail (1 cm^2 on the map equals 2500 km^2). Nevertheless, this system as well as others is useful to organize the diversity of soils and their characteristics into more manageable classes (Table 5.4).

Since the original publication in 1974, FAO (1988) has made revisions to their legend based on a better understanding of soil conditions. The modified legend applies only to new studies and updated GIS based maps. There are now 28 first level classes and 153 s level classes. A major change to the legend has been the removal of two first level classes that were defined by an aridic soil moisture regime, Yermosols and Xerosols. This change was based on one of FAO's general principles of their classification system, 'that no climatic criteria would be used to define the soil units.' The two classes were originally established because there

Table 5.4 Soil units of FAO soil map legend (FAO-UNESCO 1974)

J Fluvisols	Arenosols	Z Solanchaks	K Kastanozems
Je Eutric Fluvisols	Qc Cambic Arenosols	ZoOrthic Solanchaks	Kh Haplic Kastanozems
Jc Calcaric Fluvisols	Ql Luvic Arenosols	ZmMollic Solanchaks	Kc Calic Kastanozems
Jd Dystric Fluvisols	Qf Ferralic Arenosols	Zt Takyric Solanchaks	Kl Kastanozems
Jt Thionic Fluvisols	Qa Albic Arenosols	Zg Gleyic Solanchaks	
G Gleysols	R Rendzinas	S Solonetz	C Chernozems
Ge Eutric Gleysols		So Orthic Solonetz	Ch Haplic Chernozems
Gc Calcaric Gleysols		Sm Mollic Solonetz	Cc Calcic Chernozems
Gd Dystric Gleysols	U Rankers	Sg Gleyic Solonetz	Cl Luvic Chernozems
Gm Mollic Gleysols			Cgf Glossic Chernozems
Gh Humic Gleysols			
Gp Plinthite Gleysols			
Gg Gelic Gleysols			
Gg Gelic Gleysols			
R Regosols	T Andosols	Y Yermosols	H Phaeozems
Re Eutric Regosols	Yo Ochric Andosols	Yh Haplic Yermosols	Hh Haplic Phaeozems
RcCalcaric Regosols	Ym MollicAndosols	Yc Calcic Yermosols	Hc Calcaric Phaeozems
Rd Dystric Regosols	Yh Humic Andosols	Yg Gypsic Yermosols	Hl Luvic Phaeozems
Rg Gelic Regosols	YvVtric Andosols	Yl Luviv Yermosols	Hg Gleyic Phaeozems
		YtTakyric Yermosols	
V Vertisols	X Xerosols	G Grezems	I. Lithosols
Vp Pellic Vertisols	Xh Haplic Xerosols	Go Orthic Grezems	
Vc Chromic Vertisols	Xc Calcic Xerosols	Gg Gleyic Grezems	
	Xg Gypsic Xerosols		
	XL Luvic Xerosols		
C Cambisols	P Podzoluvisols	A Acrisols	H Histosols
Ce Eutric Cambisols	Pe Eutric Podzoluvisols	Ao Orthic Acrisols	He Eutric Histosols
Cd Dystric Cambisols	Pd Dystric Podzoluvisols	Af Ferric Acrisols	Hd Dystric Histosols
Ch HumicCambisols	Pg Gleyic Podzoluvisols	Ah Humic Acrisols	Hg Gelic Histosols
Cg Gleyic Cambisols		Ap Plinthic Acrisols	
Cg Gelic Cambisols		Ag Gleyic Acrisols	
Cx Calcic Cambisols	P Podzols		
Ck Chromic Cambisols	Po Orthic Podzols		N Nitosols
Cv Vertic Cambisols	Pl Leptic Podzols		Ne Eutric Nitosols
Cf Ferralic Cambisols	Pf Ferric Podzols		Nd Dystric Nitosols
	Ph Humic Podzols		Nh Humic Nitosols
	Pp Placic Podzols		
	Pg Gleyic Podzols		

(continued)

Table 5.4 (continued)

J Fluvisols	Arenosols	Z Solanchaks	K Kastanozems
L Luvisols	P Planosols	F Ferralsols	
Lo Orthic Luvisols	Pe Eutric Planosols	Fo Orthic Ferranosols	
Lc Chromic Luvisols	Pd Dystric Planosols	Fx Xenthic Ferranosols	
Lc Calcic Luvisols	Pm Mollic Planosols	Fr Rhodoc Ferranosols	
Lv Vertic Luvisols	Ph Humic Planosols	Fh Humic Ferranosols	
Lf Ferric Luvisols	Ps Solodic Planosols	Fa Acric Ferranosols	
La Albic Luvisols	Pg Gelic Planosols	Fp Plinthic ferranosols	
Lp Plinthic Luvisols			
Lg Gleyic Luvisols			

were no better separation criteria. Accumulation of calcium carbonate and gypsum are now used as additional separation criteria to deal with the aridic problem. Calcisols and Gypsisols classes were introduced for this purpose. These soils occur predominately under arid and semi-arid conditions (FAO 1988, p. 5–6).

5.8.3 *The World Reference Base (WRB)*

This international effort to provide a reference base that could incorporate the numerous soil classification systems into a system of names that would be universally recognized was started in 1980 and carried forward by working groups. The effort is an outgrowth of a two categorical system (Dudal 1968a, b) to define the map units of the FAO/UNESCO world soil map project (FAO 1988).

The World Reference Base (WRB) to soil resources is the international standard for soil classification system endorsed by the International Union of Soil Sciences (IUSS). It was developed by an international collaboration coordinated by the IUSS Working Group. It replaced the FAO/UNESCO legend for the soil map of the world as international standard. The WRB borrows heavily from modern soil classification concepts, including Soil Taxonomy, the legend for the FAO Soil Map of the World 1988, the Référentiel Pédologique and Russian concepts. As far as possible, diagnostic criteria match those of existing systems, so that correlation with national and previous international systems is as straightforward as possible. Although originally designed for general purpose soil correlation at world scale, WRB is increasingly used as a classification system. The Revised Legend of the FAO/UNESCO Soil Map of the World (FAO 1988) was used as a basis for the development of the WRB in order to take advantage of the international soil correlation that had already been conducted through this project and elsewhere. The first edition of the WRB, published in 1998, comprised 30 RSGs; the second edition published in 2006 and the current (third) edition both have 32 RSGs.

5.8.3.1 Historical Sketch

The development of WRB could addressed in two phases: From its beginnings to the second edition 2006 and during the period 2006–2014.

From its beginnings to the second edition 2006

The World Reference Base (WRB) is based on the Legend (FAO-UNESCO 1974) and the (FAO 1988) of the Soil Map of the World (FAO-UNESCO 1971–1981). In 1980, the International Society of Soil Science (ISSS, since 2002 the International Union of Soil Sciences, IUSS) formed a Working Group 'International Reference Base for Soil Classification' for further elaboration of a science-based international soil classification system. This Working Group was renamed 'World Reference Base for Soil Resources' in 1992. The Working Group presented the first edition of the WRB in 1998 (FAO 1998) and the second edition in 2006 (IUSS Working Group WRB 2006). In 1998, the ISSS Council endorsed the WRB as its officially recommended terminology to name and classify soils.

From the second edition 2006 to the third edition 2014

The second edition of the WRB was presented at the 18th World Congress of Soil Science 2006 in Philadelphia, USA (book: IUSS Working Group WRB 2006 (ftp://fao.org/agl/agll/docs/wsrr103e.pdf). After publication, some errors and needs for improvement were identified, and an electronic update was published in 2007 http://www.fao.org/fileadmin/templates/nr/images/resources/pdf_documents/wrb2007_red.pdf detailed description of the WRB history before 2006 is given in the second edition of the WRB (IUSS Working Group WRB 2006).From the second edition 2006 to the third edition 2014(IUSS Working Group WRB 2015). The second edition of the WRB was presented at the 18th World Congress of Soil Science 2006 in Philadelphia, USA (book: IUSS Working Group WRB 2006; file: ftp://fao.org/agl/agll/docs/wsrr103e.pdf). After publication, some errors and needs for improvement were identified, and an electronic update was published in 2007 http://www.fao.org/fileadmin/templates/nr/images/resources/pdf_documents/wrb2007_red.pdf.

In 1998, the International Union of Soil Sciences (IUSS) officially adopted the *World Reference Base for Soil Resources* (WRB) as the Union's system for soil *correlation*. The structure, concepts and definitions of the WRB are strongly influenced by (the philosophy behind and experience gained with) the FAO-UNESCO Soil Classification System. At the time of its inception, the WRB proposed 30 '*Soil Reference Groups*' (Tables 5.5 and 5.6) accommodating more than 200 ('second level') *Soil Units*. To provide an overview of the reference system, the 30 Reference Soil Groups are aggregated in 10 'sets' composed as follows:

Table 5.5 Diagnostic horizons, properties and soil materials of World Reference Base and approximate Soil Taxonomy equivalent or description

S.no	Diagnostic horizons	Approximate Soil Taxonomy equivalent or (description)
1	Albic horizon	Albic materials
2.	Anthraquic horizon	(Puddled layer and plough pan)
3.	Anthric horizon	(Ap horizon)
4.	Argic horizon	Argillic horizon
5.	Calcic horizon	Calcic horizon
6.	Cambic horizon	Cambic horizon
7.	Cryic horizon	Permafrost
8.	Duric horizon	(10% or more silica cemented Durinodes)
9.	Ferralic horizon	Oxic and Kandic horizons
10.	Ferric horizon	(Coarse red mottles)
11.	Folic horizon	Folisticepipedon
12.	Fragic horizon	(Strong structure, restricts roots and water movement to cracks)
13.	Fulvic horizon	Andic soil properties
14.	Gypsic horizon	Gypsic horizon
15.	Histic horizon	Histicepipedon
16.	Hortic horizon	Anthropic epipedon
17.	Hydragric horizon	(Redox features resulting from wet cultivation)
18.	Irragric horizon	(Mineral surface horizon resulting from irrigation)
19.	Melanic horizon	Melanicepipedon
20.	Mollic horizon	Mollicepipedon
21.	Natric horizon	Natric horizon
22.	Nitric horizon	(Argillic or Kandic horizon with > 30% clay and shiny ped faces)
23.	Petrocalcic horizon	Petrocalcic horizon
24.	Petroduric horizon	Duripan
25.	Petrogypsic horizon	Petrogypsic horizon
26.	Petroplinthic horizon	Petroferric contact
27.	Plaggic horizon	Plaggenepipedon
28.	Plinthic horizon	Plinthite
29.	Salic horizon	Salic horizon
30.	Sombric horizon	Sombric horizon
31.	Spodic horizon	Spodic horizon
32.	Takyric horizon	(Clayey surface crust on arid soils periodically flooded)
33.	Terric horizon	(Mineral material applied by humans)
34.	Thionic horizon	Sulphuric horizon
35.	Umbric horizon	Umbricepipedon
36.	Vertic horizon	(30% or more clay and slickensides
37.	Voronic horizon	(Black,80% or more B.S.|CEC_7| earthworm-rich epipedons)
38.	Yermic horizon	(Surface layer of gravel, desert pavement)

(continued)

Table 5.5 (continued)

S.no	Diagnostic horizons	Approximate Soil Taxonomy equivalent or (description)
Diagnostic properties		
39.	Abrupt texturalchange	Abrupt textural change
40.	Albeluvic tonguing	Interfinering of Albic materials
41.	Andic properties	Andic soil properties
42.	Aridic properties	(Surface features resulting from wind)
43.	Continuous rock	Lithic contact
44.	Ferralic properties	(Apparent $CEC_7 < 24$c mol kg^{-1} clay)
45.	Lithological discontinuity	Lithologic discontinuity
46.	Reducing conditions	(Saturation and reduction of iron)
47.	Secondary carbonates	Identification secondary carbonates
48.	Stagnic colour pattern	(Colour pattern indicates saturation and reduction)
49.	Vertic properties	(Presence of slickensides and cracks open 1 cm or more)
50.	Virtic properties	Volcanic glass
Diagnostic materials		
51.	Artefacts	(Human manufactured material such as bricks, glass, pottery, etc.)
52.	Calcaric material	(Strongly effervesces in 1 M HCl)
53.	Colluvic material	(Sediments from human caused erosion)
54.	Fluvic material	(Recent fluviatile, marine and lacustrine sediments)
55.	Gypsiric material	(Contains 5% or more gypsum)
56.	Limnic material	Limnic materials
57.	Mineral material	Mineral soil material
58.	Organic material	Organic soil material
59.	Ornithogenic material	(Bird excrement)
60.	Sulfidic material	Sulfidic materials
61.	Technic hard rock	(Human-made hard material)
62.	Tephric material	Volcanic glass

Source (1) IUSS Working Group WRB (2006)
(2) Soil Survey Staff (2006)

Table 5.6 Simplified guide to the WRB Reference Soil Groups (RSGs) with suggested codes (FAO 2015)

	RSG	Code
1. **Soils with thick organic layers**	Histosols	HS
2. **Soils with strong human Influence**		
With long and intensive agricultural use:	Anthrosols	AT
Containing significant amounts of artefacts:	Technosols	IC

(continued)

Table 5.6 (continued)

	RSG	Code
3. **Soils with limitations to root growth**		
Permafrost-affected:	Cryosols	CR
Thin or with many coarse fragments:	Leptosols	LP
With a high content of exchangeable Na:	Solonetz	SN
Alternating wet–dry conditions, shrink–swell clays:	Vertisols	VR
High concentration of soluble salts:	Solonchaks	SC
4. **Soils distinguished by Fe/Al chemistry**		
Ground water-affected, underwater and in tidal areas:	Gleysols	GL
Allophanes or Al-humus complexes:	Andosols	AN
Subsoil accumulation of humus and/or oxides:	Podzols	PZ
Accumulation and redistribution of Fe:	Plinthosols	PT
Low-activity clay, P fixation, many Fe oxides, strongly structured:	Nitisols	NT
Dominance of kaolinite and oxides:	Ferralsols	FR
Stagnating water, abrupt textural difference:	Planosols	PL
Stagnating water, structural difference and/or moderate textural difference:	Stagnosols	ST
5. **Pronounced accumulation of organic matter in the mineral topsoil**		
Very dark topsoil, secondary carbonates:	Chernozems	CH
Dark, topsoil. secondary carbonates:	Kastanozems	KS
Dark topsoil. no secondary carbonates (unless very deep), high base Status:	Phaeozems	PH
Dark topsoil, tow base status:	Umbrisols	UM
6. **Accumulation of moderately soluble salts or non-saline substances**		
Accumulation of, and cementation by, secondary silica:	Durisols	DU
Accumulation of secondary gypsum:	Gypsisols	GY
Accumulation of secondary carbonates:	Calcisols	CL
7. **Soils with clay-enriched subsoil**		
Interfingering of coarser textured, lighter coloured material into a finer textured, stronger coloured layer:	Retisols	RT
Low-activity clays, low base status:	Acrisols	AC
Low-activity clays, high base status:	Lixisols	LX
High-activity clays, low base status:	Alisols	AL
High-activity clays, high base status:	Luvisols	LV
8. **Soils with little or no profile differentiation**		
Moderately developed:	Cambisols	CM
Sandy:	Arenosols	AR
Stratified fluviatile, marine and lacustrine sediments:	Fluvisols	FL
No significant profile development:	Regosols	RG

Source FAO (2015)

- First, a separation is made between *organic soils* and *mineral soils*; all organic soils are grouped in Set #1.
- The remaining (mineral) major soil groups are each allocated to one of nine sets on the basis of '*dominant identifiers*', i.e. those soil-forming factor(s) which most clearly conditioned soil formation.

SET #1 holds all soils with more than a defined quantity of 'organic soil materials'. These organic soils are brought together in only one Reference Soil Group: the Histosols.

SET #2 contains all *man-made soils*. These soils vary widely in properties and appearance and can occur in any environment but have in common that their properties are strongly affected by human intervention. They are aggregated to only one Reference Soil Group: the Anthrosols.

SET #3 includes mineral soils whose formation is conditioned by the particular properties of their *parent material*. The set includes three Reference Soil Groups: the Andosols of volcanic regions; the sandy Arenosols of desert areas, beach ridges, inland dunes, areas with highly weathered sandstone, etc., and the swelling and shrinking heavy clayey Vertisols of back swamps, river basins, lake bottoms, and other areas with a high content of expanding 2:1 lattice clays.

SET #4 accommodates mineral soils whose formation was markedly influenced by their *topographic/physiographic setting*. This set holds soils in low terrain positions associated with recurrent floods and/or prolonged wetness, but also soils in elevated or **accidented terrain**? where soil formation is hindered by low temperatures or erosion. The set holds four Reference Soil Groups: In low terrain positions: Young *alluvial* fluvisols, which show stratification or other evidence of recent sedimentation, and non-stratified gleysols in *waterlogged areas* that do not receive regular additions of sediment. In elevated and/or eroding areas: *Shallow* Leptosols over hard rock or highly calcareous material, and 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 holds soils that are only moderately developed on account of their *limited pedogenetic age* or because of *rejuvenation* of the soil material. Moderately developed soils occur in all environments, from sea level to the highlands, from the equator to the boreal regions, and under all kinds of vegetation. They do have not more in common than '*signs of beginning soil formation*' so that there is considerable diversity among the soils in this set. Yet, they all belong to only one Reference Soil Group: the Cambisols.

SET #6 accommodates the 'typical' red and yellow soils of *wettropical and subtropical regions*. High soil temperatures and (at times) ample moisture promote rock weathering and rapid decay of soil organic matter. The Reference Soil Groups in this set have in common that a long history of dissolution and transport of weathering products has produced deep and genetically mature soils (1) Plinthosols on old weathering surfaces; these soils are marked by the presence of a mixture of clay and quartz ('*plinthite*') that hardens irreversibly upon exposure to the open air, (2) deeply weathered Ferralsols that have a very *low-cation exchange capacity* and are virtually devoid of weatherable minerals, (3) Alisols with *high-cation exchange capacity* and *much exchangeable aluminium*, (4) deep Nitisols in relatively rich parent material and marked by *shiny, nutty structure elements*, (5) strongly leached, red and yellow Acrisols on acid parent rock, with a *clay accumulation horizon*, *low-cation exchange capacity* and *low base saturation*, and (6) Lixisols with a *low-cation exchange capacity* but *high base saturation percentage*.

SET #7 accommodates Reference Soil Groups in *arid and semi-arid regions*. Redistribution of calcium carbonate and gypsum is an important mechanism of horizon differentiation in soils in the dry zone. Soluble salts may accumulate at some depth or, in areas with shallow groundwater, near the soil surface. The Reference Soil Groups assembled in set #7 are:(1) Solonchaks with a high content of *soluble salts*, (2) Solonetz with a high percentage of *adsorbed sodium ions*, (3) Gypsisols with a horizon of *secondary gypsum enrichment*, (4) Durisols with a layer or nodules of soil material that is *cemented by silica*, and (5) Calcisols with *secondary carbonate enrichment*.

SET #8 holds soils that occur in the *steppe zone* between the dry climates and the humid temperate zone. This transition zone has a climax vegetation of ephemeral 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 three Reference Soil Groups:

- Chernozems with *deep, very dark surface soils* and *carbonate enrichment* in the subsoil,
- Kastanozems with *less deep, brownish surface soils and carbonate and/or gypsum accumulation* at some depth (these soils occur in the driest parts of the steppe zone), *and*
- Phaeozems, the dusky red soils of prairie regions with *high base saturation* but *no visible signs of secondary carbonate accumulation.*

SET #9 holds the brownish and greyish soils of *humid temperate regions*. The soils in this set show evidence of redistribution of clay and/or organic matter. The cool climate and 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 (coating) colours of soils of this set. Set #9 contains five Reference Soil Groups:

- acid Podzols with a *bleached eluviation horizon* over an *accumulation horizon* of organic matter with aluminium and/or iron,
- Planosols with a bleached topsoil over *dense, slowly permeable subsoil*,
- base-poor Albeluvisols with a *bleached eluviation horizon tonguing* into a *clay-enriched subsurface horizon*,
- base-rich Luvisols with a distinct *clay accumulation horizon*, and
- Umbrisols with a thick, *dark, acid surface horizon* that is rich in organic matter.

SET #10 holds the soils of *permafrost regions*. These soils show signs of '*cryoturbation*' (i.e. disturbance by freeze–thaw sequences and ice segregation) such as irregular or broken soil horizons and organic matter in the subsurface soil, often concentrated along the top of the permafrost table. Cryoturbation also results in oriented stones in the soil and sorted and non-sorted patterned ground features at the surface. All 'permafrost soils' are assembled in one Reference Soil Group: the Cryosols (Table 5.6). A simplified guide WRG soil units is appendages as Fig. 5.6.

5.9 Conclusions

Intimate knowledge of soils with regard to their genesis, morphology and physical and chemical properties is a prerequisite for undertaking any soil inventory programme. A great deal of literature is available on soils on above-mentioned aspects including the approaches for conducting soil surveys and soil classification. While there are general agreements amongst various schools of thoughts on almost all aspects of soils, the consensus building process is still in progress with respect to soil classification. This is because of the fact that many countries have developed systems to classify their soils, but the results often do not translate well between taxonomic systems. There are glaring anomalies in some of the national-level classification systems. For instance, in Soil Taxonomy, changes in the definition of '*iso*-soil temperature regimes, the questionable criteria for *spodic* horizon; the case of Mollisols with an *aridic* moisture regime which are classified as Ustolls; the problem of extremely acid, wet Vertisols,; the ambiguity of Pale great groups, the dilemma of classifying paddy soils, and the predicament of *kandic* horizon, are some of the issues that need further discussion and clarification (Beinroth and Eswaran 2003).

Attempts have been made through efforts such as the *FAO Legend for the Soil Map of the World*, and the *World Reference Base for Soil Resources* to address the need for a globally acceptable soil classification system. But so far, this goal has not been achieved. Concerted efforts, therefore, need to be made to develop such a soil classification system. Golden et al. (2010) have suggested the formation of working group under the aegies of International Union of Soil Science to develop a universal soil classification system.

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