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SOIL SCIENCE

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Content

Preface

The amount of knowledge in the area of soil science, similarly like in other scientific fields, is continually increasing. It is more complex and many uncomprehended phenomena are still being explained. This is one of the principal reasons for the innovation of university textbooks, scripts, and study manuals. At the Slovak University of Agriculture the textbooks of soil science for the students are published at the regular intervals. However, as a result of the significant internationalization of study, the requirement arose to provide the interested persons, predominantly the students, with the basic study literature in the English language. Therefore, the university textbook "Soil Science" was created, which is the outcome of the common effort of the experts in the area of soil science and the teachers from the Centre of Languages SUA in Nitra. The textbook is written with the purpose to give comprehensive information sources from this subject. The publication will meet the needs of the students in their study but also it will have a wider application for both professional and non-professional public by containing the information utilizable in the practice. Firstly, the reader will become acquainted with the basic facts and elementary terms, and also the process of soil development and genesis. Next, the textbook introduces the particular soil functions and their chemical and physical properties. Finally, the attention is targeted at the soil fund, mostly in the Slovak Republic, and its protection. It is necessary to protect soil because its condition reflects the culture and maturity of each nation. Indeed, our respect for the soil is the benchmark of respect for ourselves wherever we live in the world.

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Authors

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Soil as a natural resource

Soil is a biologically active, structured porous medium, whose development takes place on the surface of our planet. Many scientists describe soil as the Earth cover that requires attention because it is characterized by enormous heterogeneity. Soil science deals with the study of soil as a natural source on the Earth’s surface. It involves mapping, classification of existing soils, the study of soil-forming processes, and the study of their physical, chemical, and biological properties. Soil science is a relatively young science that was established by the Russian scientist V. V. Dokučajev at the end of the 19th century.

When we look briefly at the historical (brief) development of the soil definition, we can find out that it has always reflected the level of the current knowledge and expectations in meeting the needs of society.

In the period of the substantial study of nature, a very simple definition of soil was sufficient “**as the upper layer of the Earth able to bear a plant cover**“. Later, under Dokučajev´s influence, the definitions were preferred which emphasized predominantly the essence of soil origin. Dokučajev (1880) defined soil as follows: “**The surface layer of rocks which are transformed by the activity of water, air, and organisms. It is the autonomous natural-historical unit with the constantly changing properties under the impact of rocks, climate, plant and animal organisms, relief, altitude, and age.**“

In the first half of the 20th century, the productive ability was being asserted considerably and soil was defined simply as “**the loose surface of the Earth able to provide yield**“. In the course of the past 50 years, the enormous development of all scientific fields was observed, which was obviously reflected also in the view of soil. The definitions began to stress the soil itself and they did not emphasize only one product – yield. The new definitions originated which described the soil as a natural formation, consisting of individual phases. The definitions of soil started claiming the requirements or recommendations for its protection and regulation of its utilization. The definition of soil is becoming the basis of state policies and numerous tools and measures for its protection. The recommendation of the Council of EU R(92)8 on soil protection (1992) defines soil as follows: “**It is an integral part of the Earth ecosystems situated between**

the surface and parent rock. It is divided into horizons of the specific physical, chemical and biological properties and the different functions. The soil concept also includes the porous sedimentary rocks and other permeable materials along with contained water. Such defined soil can achieve considerable depths, therefore, in some contexts the term soil can be understood also as an area.“

Soil is a unique and irreplaceable natural resource that serves to meet human needs. It is evident that as a result of the acquisition of still more amount of new and more precise knowledge of soil, there will occur new efforts to specify its definition. Within the standard education of soil science (Institute of Agronomical Sciences, Faculty of Agrobiology and Food Resources, the Slovak University of Agriculture in Nitra) the preferred definition of soil is: **Soil is a natural formation that is being developed as a result of the difficult and complex impact of the external (exogenous) factors on the parent rock (endogenous factor) and is characterized by fertility.**

1.1 Genesis and development of soil

The pedosphere is the Earth cover where the soil is being formed and developed. It is created at the borderline of the impact of atmosphere, biosphere, lithosphere, and hydrosphere (Fig. 1.1). Soil is a porous body, i.e. the spaces (pores/voids) occur between the solid particles, which are filled by air and water. The solid particles contain the mineral (lithosphere) and organic, living, and abiotic parts (biosphere). Soil is a substantial prerequisite for life in terrestrial ecosystems and affects the balance of energy, hydrologic cycle, nutrients, and productivity of ecosystems (Fig. 1.2). In soil the balanced inputs and outputs are enacted, and also the exchange of gases (O_2, CO_2, water, methane, nitrogen oxides) with atmosphere. The soluble parts, such as nitrates and soluble organic carbon are leached out from soils. Under the impact of many physical, chemical and biological processes taking place in the soil, the unequal division of material occurs there and the particular horizons of soil profile are being formed. In addition, the decomposition and erosive processes lead to the unequal placement of material in the landscape.

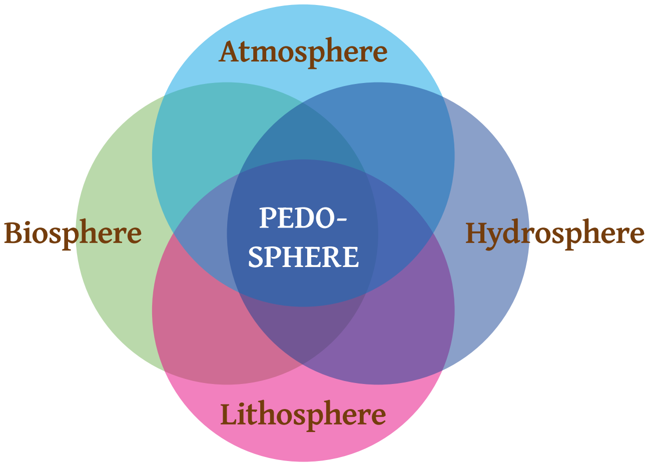


Figure 1.1: Relations between the individual spheres (Juma, 1999)

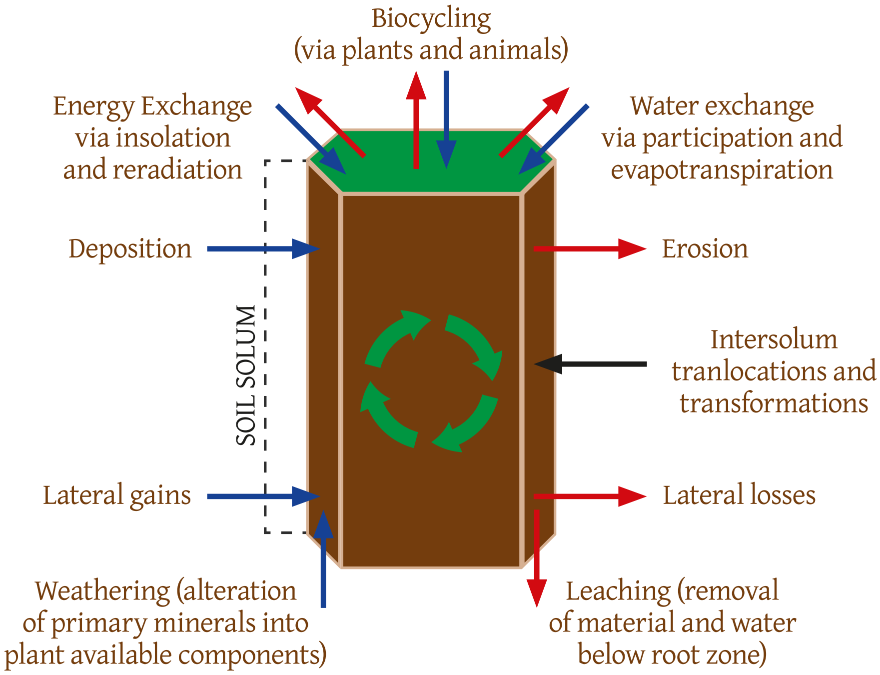


Figure 1.2: Schematic representation of the solum of a soil pedon as open system (Buol et al., 1989)

As it is indicated in Figure 1.3 soil is a result of the soil-forming process where the biological and geological cycle of substances plays a significant role.

The biological cycle of substances underlies the processes of the biological accumulation of organic and mineral compounds in the soil surface layers and it formation of a horizon.

The geological cycle of substances, which is characterized by the processes of weathering and migration of mineral and organic compounds (geological eluviation), causes the depletion of soil surface layers, **segmentation of soil profile**, or its disintegration.

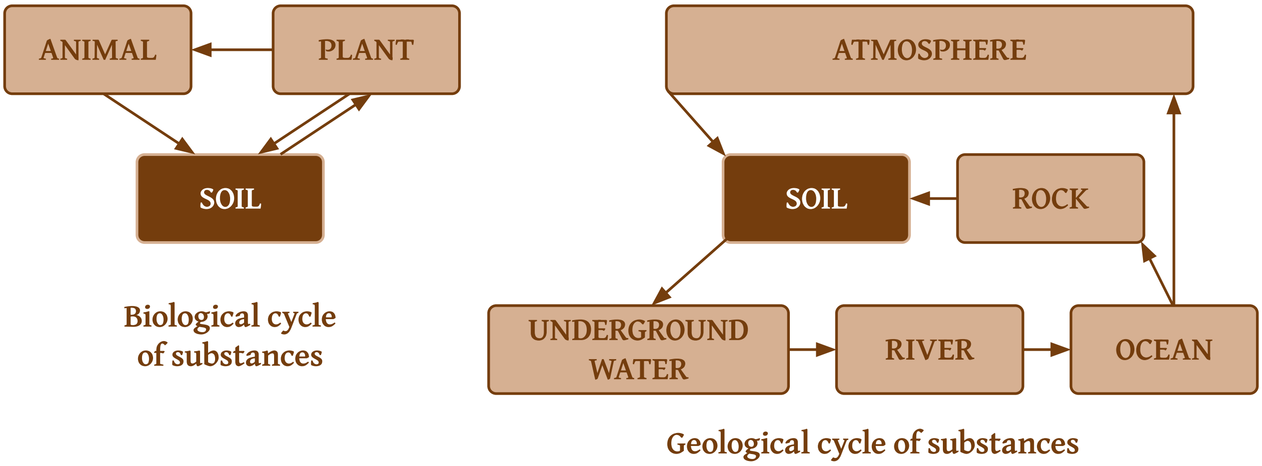


Figure 1.3: Biological and geological cycle of substances in nature (Zaujec et al., 2009)

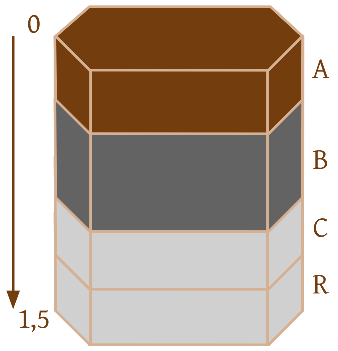


Figure 1.4: Soil profile (Šimanský, 2013)

The basic soil unit is a soil pedon. The soil pedon is a 3-dimensional body in the landscape area (Fig. 1.4) and it can be divided into four main components: air, water, mineral, and organic part. Air and water constitute a total of 50% of the pedon volume, while the living and abiotic organisms create approx. 5% and the rest is the mineral content. In particular layers, the soil horizons are formed by the arrangement of soil pores/voids and elementary particles of the mineral part or soil aggregates. Based on the horizons the soil scientists are able to identify and classify soils. The soil profile can have from one to four horizons which are located under the soil surface A, B, C, and the bedrock referred to as the R horizon. The soil profile can also contain organic horizons, which occur predominantly on forest soils with the different levels of decomposition of organic matter on the soil surface.

In the soil, many physical, chemical, and biological processes take place which are driven by the changes of different energies. For example, the soil erosion from the top of a hillslope to a lower place of the slope in the landscape is a reflection of the changes in kinetic energy of the falling rain drops on the soil surface. The daily thermal regime in the soil is a reflection of the absorbed solar radiation by the soil surface. The solar energy in the soil is being changed into a thermal one, whereby the soil temperature is increased during the day. The fluctuation of temperature has also a significant impact on the speed of weathering and soil-forming processes, and also biological activity. All the indicated examples point out that soil is a reactor where the energy is created and transferred.

1.2 Factors and conditions of soil-forming process

The formation and genesis of soils is a long-lasting process. During this development, which is often called pedogenesis, the original thin soil layer increases gradually the thickness of the individual layers or horizons. Soil formation includes two broad processes. The first one is the formation of parent material and, the second one is the evaluation of soil layers. There occurs the differentiation of the color, texture, and structure, and the soil profile is created.

a) **The factors of soil-forming process** participate in the soil formation *materially* (rocks and vegetation) or *energetically* (energy).

b) **The conditions of soil-forming process** are those soil-forming agents which do not participate in the soil-forming process either materially or energetically, but they have an impact on the functioning of the other directly affecting factors. Therefore, their effect is indirect.

As a result of the enormous diversity of the factors themselves and conditions of the soil-forming process, soil scientists defined the different soil types of the Earth.

1.2.1 Factors of soil-forming process

Bedrock, parent material of soils

The bedrock and parent materials as a soil-forming factor affect materially (particularly also energetically) and create the essential part of soil solid mass. Their mineralogical, and chemical composition affected the intensity of the soil-forming process and soil development. Rocks on the Earth's surface disintegrate, i.e. they change their structure and composition. The most significant factor of weathering is the climate, predominantly the temperature fluctuation and changes in humidity. In the process of chemical weathering, the chemical composition of rocks is changed as a result of water activity in humid and warm areas.

Generally, in mineral soils, the parent rock provides more than 90% of the total volume of the soil solid phase and as a consequence of the different compositions of the particular minerals and rocks, it also determines the future physical and chemical properties of soils. The bedrock and parent materials can have an impact on the origin of soil via weathering in situ (in this case the mineralogical composition of the soil is the result of the minerals and rocks themselves at that given place) or the sediments which were displaced by wind, water and glacier and they can overlap the original parent materials.

The impact of soil-forming substrate, or parent rock, can be illustrated in the following examples. The soil-forming process under forest on 1 granite (acid rock) and 2 basalts (alkaline rock). Under the identical conditions of the soil-forming process, podzolization begins much earlier on granites than on basalts. Explanation: During weathering of basalts the secondary minerals are being created with a higher presence of the alkali cations that inhibit the podzolization that take place in acid environment. The result is the origination of Cambisols, however without the considerable differentiation of soil profile. If the soil-forming substrate is the rock with high content of carbonates, calcium, or magnesium (calcites, dolomites, dolomitic limestones) then the organic sloughing and acid products of decomposition of the organic matter are neutralized.

Apart from the direct impact, the parent material has also an indirect effect on the soil-forming process. A relief depends on the properties of rocks and minerals, i.e. the character of vegetation or soil depth itself. Water and the thermal regime reflect the properties of minerals and rocks.

Climate

Climate has both **material and energic** impacts. The climate influences directly the formation and development of soil through precipitation, temperature, and changes in relative humidity or wind speed. It means that the climate has an impact mostly on the thermal and water regime of soil. The intensity of all reactions and processes in soil depends on it. The changes in soil temperature are closely related to the changes in humidity. The alternation of cycles with higher and lower temperatures in the soil medium results in acceleration but also inhibition of the transformation processes of the organic matter, and it also interferes with the weathering itself of minerals and rocks in the soil. For example, under colder conditions more intensive physical weathering occurs (i.e. the disintegration of bigger rocks into smaller spalls) than a chemical or biological one, mainly via ice-cover formation and defrosting of water in soils. As a result of low temperature, water changes its state from liquid to solid, which can lead to the bonding of the soil particles with the frozen water. On the contrary, under warmer conditions more intensive chemical and biological processes are running in soils.

The more humid climate allows the formation of a washable and periodically washable type of water regime and it supports the motion of substances in the direction down from the top. The excess moisture causes the soil to have a low content of water-soluble substances, acidification, chemical weathering, clay formation, and sesquioxides.

A more arid climate with the insufficiency of humidity promotes the formation of an impercolative even evaporative type of water regime. The processes of decomposition of minerals and chemical weathering are less intensive, and the movement of water-soluble substances down from the top is also limited. There the accumulation of the soluble substances occurs, whereby their capillary intake is connected with the water evaporation.

The indirect impact of climatic factors depends on the intensity of weathering and the composition of bedrock, or parent materials. The climate affects the formation of the relief. It influences vegetation through a man. Therefore as a final result, the climatic zones have the appropriate soil zones. In our country (Slovakia), where at a small part of the area the terrain is very rugged, we can observe apart from the latitude zonation also the vertical zonation. Under these conditions predominantly microclimatic areas originate which have a more distinctive diversity of soil types as a result.

Living organisms

The activity of living organisms in the soil-forming process involves:

1. The activity of different microorganisms that live in soil and participate actively in the processes of decomposition and synthesis of mineral and organic substances in the soil profile.
2. The soil-forming role of vegetation – substance exchange between soil and plants and at the same time the main source of soil organic matter.
3. The activity of soil living organisms.

Thus, living organisms of plant and animal origin are an essential part of the soil-forming process. The organisms vitalize a rock and engage it in the complex processes of decomposition and synthesis, accumulation, and migration of substances. Living organisms play a significant role in the formation of qualitatively new properties of the soil-forming substrate, mainly soil fertility.

The plants participate in the formation and accumulation of organic matter. The individual plant communities differ by the total production of organic matter by the way of its deposition and chemical composition. The significant differences are observed e.g. between **the forest and grasslands communities**.

The dominant source of organic matter under **the forest communities** is the dead residues of woody plants that are deposited on the soil surface as **litter**. The underground biomass of woody plants participates in the mechanical and chemical disturbance of the mineral part of the soil, and in addition, the root mass after the death of the woody plants becomes a part of the organic matter of the soil. The forest cover has also a considerable anti-erosive effect. It is also involved in the limitation of evaporation from the soil. The more humid surface horizons and drier lower layers (abundant drainage of moisture by the tree roots) support the motion of water, and thus also the water-soluble substances down from the top.

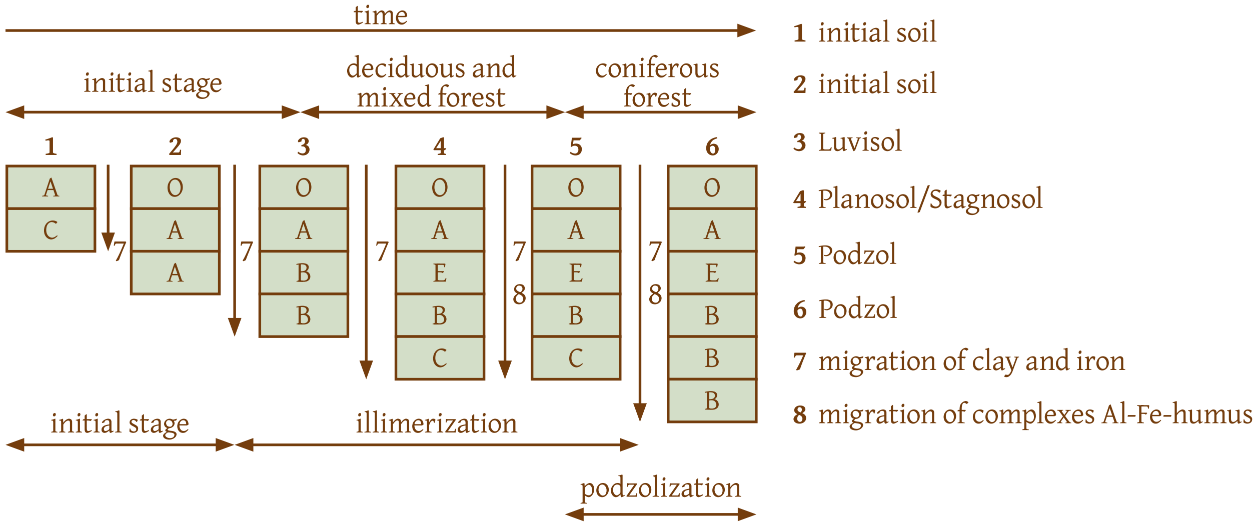


Figure 1.5: Change of soil units as a result of impact of different vegetation during time (Duschafour, 1991)

On the contrary, **the herbal communities** in comparison with the forest communities provide mostly the root mass for the formation of humic substances. This mass is deposited directly in the soil. Compared with the forest litter the dead residues of herbs are richer in easier degradable organic compositions, proteins, and ash. The herbal communities withdraw the most moisture from the surface layers, which are dried, and in this way, these communities facilitate the movement of moisture down from the top. The classical example of the effect of plant vegetation on the development of soil body from the initial stage through the illuvial stage to the podzolization was described by Duchafour (1991) – summarized in Figure 1.5.

Apart from vegetation also animals in soil are important in the soil-formation. They participate in the disintegration and change of plant residues by their transport into deeper parts of soil layers, as well as their mixture with soil particles. The optimal situation is when the soil contains a wide diversity of organisms from bacteria and fungi to arthropods and earthworms that participate in the decomposition of organic matter. The majority of these organisms rank among saprophytes (they consume the dead organic residues) and therefore, they are crucial for the nutrient cycle. They also take a significant part in the formation of optimal physical conditions in soils (e.g. formation and stabilization of soil aggregate by hyphae or slimes of bacteria).

The soil biota has an impact on:

- the nutrient cycle in soils,
- regulation of water movement and reserves in soils,
- regulation of transportation of soil itself and sediments,
- regulation of some harmful microorganisms, living organisms by other useful animals occurring in soil,
- detoxication of some xenobiotics and polluting substances,
- regulation of the composition of the atmosphere.

The volume and quality of biomass are essential for soil development (Bedrna and Jenčo, 2016). This biomass in soil mineralizes, humifies, or underlies peat formation. a part of organic matter, which comes from the dead edaphon and dead plant roots, remains directly in the soil body. However, the majority of the organic matter is accumulated on the soil surface.

Water

The increased water content can have a short-term impact on soil in the case of rains and floods, a temporary impact, e.g. in water reservoirs, or a permanent impact. Groundwater ranks among the directly affecting soil-forming factors but its impact become evident only under the conditions where its level is relatively high above the surface. The increased humidity is kept above the level of groundwater, which causes the formation of anaerobic conditions, supports the reductive processes, and inhibits the decomposition of organic substances; in that way, it facilitates their accumulation. The increased humidity in soil underlies the gley process that consists of the accumulation of clay and reduction of Fe and Mn. The groundwater with high mineral substances contents under the conditions of the evaporative water regime underlies also the accumulation of salts in the soil profile.

Human activity

A man affected the humic substances, water regime, and total chemism of soil processes by the change of the original vegetation. A human changed essentially the water and thermal regime, and the terms of decomposition of organic matter by **the soil tillage systems**. As a result of soil tillage the cultural layer – topsoil was formed in the soil. A man accelerates the mineralization of soil organic matter and humic substances in soils through intensive soil cultivation, which is reflected in the soil organic matter decline and associated deteriorative soil state, mainly its physical properties.

The fertilization by organic and mineral fertilizers has an impact on the soil properties and interferes into the natural cycle of substances. The utilization of high doses of physiologically acidic mineral fertilizers causes acidification, or the application of fertilizers containing monovalent cations (KCl, NH_4NO_3) leads to the peptization of soil colloid systems, and in that way the break down of soil aggregates.

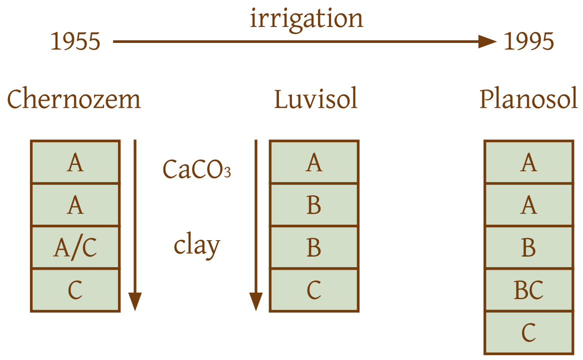


Figure 1.6: Impact of irrigation on soil development

On the contrary, the liming and application of manure, organic fertilizers, and amendments eliminate not only the acidification but also the intensity of processes of illimerization, or podzolization in soil. It is a well-known example of the genetic degradation of soil when in the original Chernozem developed on loess the genetic degradation, dealkalization, acidification and illimerization occurred as a result of the intensive irrigation of average annual dose of 150 mm for almost 40 years. Such significant analytic and even morphological changes occurred that resulted in the change of properties as well as the final soil type classification (Fig. 1.6). The human activities and intervention should always be targeted at the improvement of the soil properties.

1.2.2 Conditions of soil-forming process

Landscape relief

Landscape relief has an indirect impact on the soil-forming process, through directly affecting factors, predominantly microclimate, vegetation, and groundwater. Its impact has already been mentioned partially in relation to the influence of climate or bedrock or parent material. Landscape relief affects the soil-forming process by altitude and geomorphological forms of the landscape. Along with the rising altitude, the temperature is decreasing, and also humidity, the intensity of chemical weathering, leaching and eluviation are being increased. The altitude changes also the character of vegetative cover. The relief affects mainly the division of warmth and water, and also the mechanical transport of soil mass. These are mainly the slopes with different exposition and inclination, which influence differently on the degree of heating, humidification, and soil erosion process. In the flat and slope depressions the specific conditions are being created of the soil-forming process. As a rule, there occurs the accumulation of fine-grained particles, organic matters, the increased humidity, or the level of groundwater.

1.3 Partial soil-forming processes

In the soil-forming process there chemical and biological reactions, physical actions, and energic changes take place. Depending on the particular conditions the elementary processes are associated into smaller complexes or groups – **partial soil-forming processes**. The process of soil formation consists of the chain of the partial soil-forming processes and is called **the essential soil-forming process**. In each essential soil-forming process, which participates in the formation and development of a certain soil at a particular territory, the corresponding partial processes assert in the different quality and intensity. From the aspect of human life, the soil-forming processes are slow, however, in comparison with the geological processes they are faster.

Humification – is one of the most significant partial processes, its outcome is the formation and accumulation of humus in the surface part of the soil. During the process of humification, the organic substance changes its chemical composition and acquires new properties. From a chemical viewpoint, the products of humification constitute a very heterogeneous material, in which the high-molecular nitrogenous compounds with a high presence of aromatic components prevail. The result of humification is the formation of the **A-horizon.**

Eluviation – is another partial soil-forming process where the relocation of change products (soluble or dispersed) occurs by the impact of water down from the top. Most frequently the clay particles, Fe_2O_3, Al_2O_3, SiO_2, humic substances, $CaCO_3$, but also other salts are liable to eluviation. This process involves the mobilization and translocation of the moving parts, which leads to differences in the texture of the soil profile. The horizon created by the process of eluviation is called the eluvial horizon. Translocation depends on the relative mobility of the elements and the depth of penetration. According to the intensity of decomposition, dispersion, and dissolution of substances and their relocation we can distinguish:

1. **Leaching** – depletion of easily soluble salts from the surface horizons of the soil profile.
2. **Degradation** – is related to the depletion of easily and also more difficult soluble salts, such as e.g. $CaCO_3$. In the process of degradation, a part of changeable calcium is displaced by hydrogen in the sorption complex. At the same time, the shift of low-molecular organic compounds can happen.
3. **Illimerization** – the relocation (translocation) of mineral and organic colloids without their more substantial destruction. The translocation of colloids occurs in the period of higher humidity (higher hydration of colloids, a decrease of concentration of coagulating ions, mainly Ca_2^+ in the soil solution). The translocation of colloids leads to the formation of a depleted – **E-horizon.** The decrease of humidity and increased concentration of soil solution causes the immobilization of colloids, and thus, the creation of **the luvic B horizon** (partial process – illuviation).
4. **The formation of solonetz** – is associated with the peptization of soil colloids by the impact of the changeable Na^+ and with the translocation of peptized colloids. At a certain depth, the coagulation begins with the effect of the increased content of soluble salts. The translocation of colloids forms a thin and light humic-eluvial horizon (eventually eluvial horizon) and as a result of their accumulation the dark even black saline **B horizon** is created at a certain depth.
5. **Solodization** – the relocation of the peptized colloids where occurs the acidification of alkaline sois and the partial decomposition of clay minerals. In the process of solodization, the changeable Na+ is displaced by hydrogen.
6. **Podzolidation** – the process of peptization of humic substances and release Al and Fe from minerals (destruction of minerals), where the considerably depleted E horizon is formed by the migration of material via the gravitational waters and the consequent immobilization in a certain depth of soil profile under the strongly acidic reaction ($pH_{H_2O} < 5.0$). Under this horizon there significantly **enriched podzolic B horizon** is being formed (by sesquioxides – Al_2O_3, Fe_2O_3 or by sesquioxides and humic substances).
7. **Laterization** – is associated with the relocation of silicic acid (desilication) under the conditions of high humidity, temperature, and intensive biological activity (subtropic and tropic zones).

Illuviation – accumulation of eluviated substances in a deeper part of the soil profile. The enriched **B horizons** (luvic B horizon, podzolic B horizon and saline B horizon) are formed by the cumulation of the colloids, sesquioxides, and organic components.

Salinization – is the accumulation of water-soluble salts in the soil surface horizons. The process is pursued under the conditions with the evaporative type of water regime where the level of groundwater with higher salt conten is high below the surface.

Gleyification – is in progress under the level of groundwater (reductive conditions) or in the zone of its impact (changing of reductive and oxidative conditions). The characteristic demonstration of the process is:

- the reduction of Fe and Mn and their migration,
- specific formation and change of clay (particles < 0.001 mm; considerable hydration of clay materials), unfavorable microstructure (high bulk density),
- the limited mineralization of organic substances and significant peat formation as well as low-molecular organic substances,
- grey, grey-green and bluish color of the reduced compounds Fe and Mn.

Gleyification is in progress most intensively under the conditions of acid soil pH where the organic substances are the most active and cations Fe^{2+}, Mn^{2+} and Al^{3+} are in the ionic form. In the zone of changes of reductive and oxidative processes, the partial oxidation Fe^{2+} and Mn^{2+} (Fe^{3+}, Mn^{4+}) occurs and these elements are precipitated, which is demonstrated also in the change of color to rusty-red (Fe^{3+}) or black-blue (Mn^{4+}). Gley horizon is formed where the proportion of grey, grey-green and bluish color is > 50%.

Pseudogleyification is in progress under conditions where the level of inner drainage is considerably limited by the poorly permeable layer in the soil profile. In humid periods such a state causes the accumulation of rainwater in the soil profile. During periods of increased humidity the mobilization of reduced Fe and Mn and also compounds Al occurs by low-molecular organic substances, where the mobile complex compounds are created – chelate compounds. This process is demonstrated morphologically by the formation of bleached reduced localities (veins, spikes, blotches, tongues). The immobilization and formation of oxidated parts (black-brown – Mn^{4+}, rusty brown, orange-brown – Fe^{3+} concretions, coatings, blotches) occur after the soil has dried. In this way, the **marble-like B horizon** is formed.

In situ weathering – is a process where the alkaline cations (Ca^{2+},Mg^{2+},K^{+},Na^{+}), sesquioxides (Al_2O_3, Fe_2O_3) are released and the secondary minerals are formed. The intensity of the process depends on the mineralogical composition of rocks, acidity, and hydrothermal conditions of the environment. The strong demonstration of the process is browning, caused by the release of Fe_2O_3 from minerals and its dispersal on the surface of soil particles (predominantly under the colder more humid conditions the formation of clay is rather small). The low-molecular organic matters are neutralized R_2O_3 (Fe_2O_3 + Al_2O_3) with bivalent alkaline cations (Ca^{2+},Mg^{2+}), therefore they do not activate podzolization (the aggressiveness of organic acids is eliminated by their bond to clay minerals). The stated processes inhibit the migration of sesquioxides and clay in the soil profile and lead to the formation of **the cambic B horizon**.

1.4 Soil survey and mapping

Soil creates the living conditions for humankind and all living organisms. It constitutes the basic means of production for food security and raw materials for the life of human society and all living organisms on the Earth. The care and protection of agricultural soil is the essential prerequisite of its rational and sustainable utilization for the production of plant biomass, and also for the security of the protective and regulative soil functions in the agricultural landscape. Without the knowledge of soil, it is not possible its rational usage and prevention of possible consequences of its improper use. The knowledge of soils begins there where they are located – directly in terrain.

The detailed soil description and study of the terrain are targeted at several well-known goals:

- classify soils into soil categories,
- acquire the appropriate basis of knowledge of the genesis and development of soils,
- understand all physical, chemical, and biological processes in soils,
- know the classification of soil units (soil groups, types, subtypes) – soil mapping.

Evaluation of soils directly in terrain involves the following stages:

- preparatory work,
- reconnaissance of terrain and demarcation of the sounding network,
- digging of soil pits, description of the soil profile,
- soil sampling,
- the field evaluation of the selected soil categories at the area of interest and compilation of the outline of the soil map.

Soil investigation may be carried out on various levels of knowledge, research capacity, and proficiency. Scientists commonly apply an advanced methodology for soil resources inventory, including the professional terminology for landscape and soil description, data acquisition and processing, soil classification and mapping, and soil and land evaluation. a simplified version of methodological procedures suitable for site and soil description, and soil classification with specific examples can be found in the publication: Guidelines for Soil Description and Classification – Central and Eastern Europen Students´ Version (2018).

1.5 Agrochemical soil testing

It is essential to state that in Slovakia apart from the classical soil and hydrological surveys also so called Agrochemical Soil Testing is pursued (AST). The monitoring of the agrochemical soil parameters is a vital part of the modern agriculture. The regular detection of soil chemism is perceived as the significant indicator of the soil care. The results of AST allow the effective regulation of the soil chemism. Based on the tests it is possible to execute the optimization of the habitat conditions of plants through soil liming and rational fertilization. These tests have been carried out since 2000 in six-year cycles and their objective is to determine the supply of soils by nutrients, available phosphorus, potassium and mobile magnesium. Apart from these data also the soil pH is being detected. According to it the need of liming is set (Gáborík and Prístavka, 2013).

1.6 Soil monitoring in Slovakia (sub-monitoring system – SOIL)

The sub-monitoring system – Soil is a part of the Slovak Environmental Monitoring. The monitoring of agricultural and forest soils is mainly aimed at monitoring the development of those properties in space and time which are crucial for soil fertility, the ecological functions of soils, and, with the same degree of importance, their contamination by hazardous substances, in terms of the possible input of these substances from soils into the food chain.

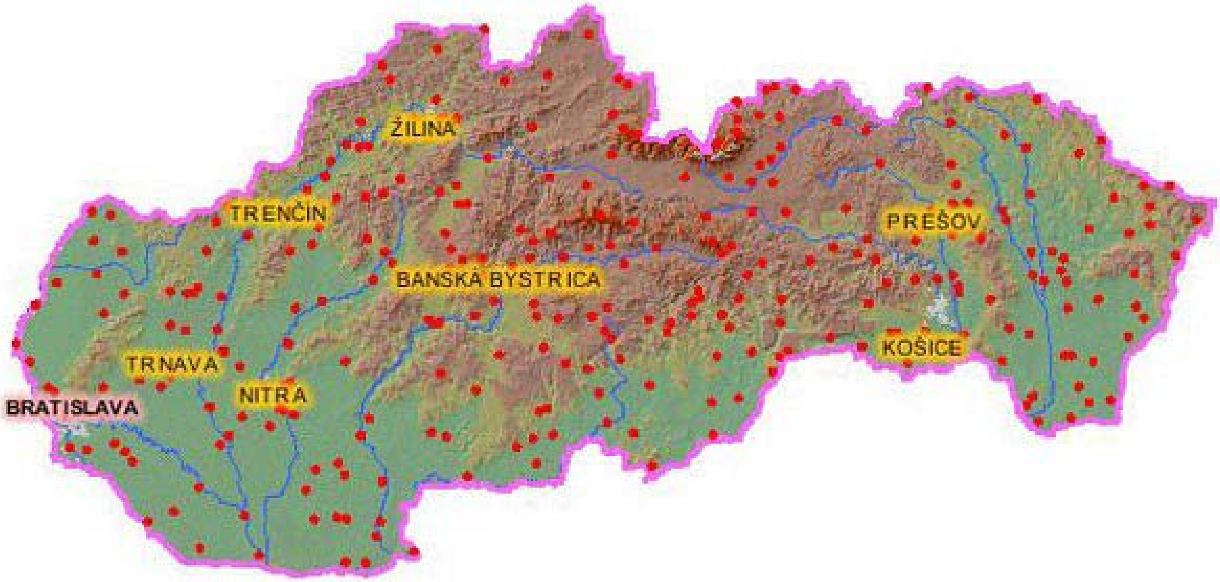


Figure 1.7: Partial monitoring system – Soil (VUPOP, 2005)

Table 1.1: Area, characteristics, and purpose of partial monitoring system – Soil (VUPOP, 2005)

Monitoring area	Monitored characteristic	Purpose
Soil contamination	Inorganic and organic contaminants	Current state and development of pollution
Acidification, salinisation and alkalinisation of soils	Soil reaction, sorption complex and exchange cations	Development trends of soil acidification and alkalinization, identification of soil vulnerability indicators
Monitoring of available nutrients	P, K, Mg	Evaluating the evolution of soil nutrient supply since 1990, especially in relation to changed levels of fertilization
Monitoring of soil organic matter and humic substances	Cox, humic and fulvic acids	An assessment of the importance of humus development in soil fertility

continuation of table 1.1

Monitoring area	Monitored characteristic	Purpose
Soil compaction	Physical properties of soils	The degree and development of compaction of arable soils
Soil erosion	Grain size, soil reaction, humus content, available nutrients P and K	Assessment of current trends in soil degradation
Desertification	Soil solution composition, sorption complex, water level and groundwater mineralisation	Assessment of properties under global climate change

To be able to monitor significant environmental impacts on soils throughout the country, the system is based on a network of monitoring sites where pedological probes are placed. These constitute all the most frequent soil representatives, all methods of their use, and their occurrence in the main climatic regions and within all the areas of the main types of contamination. The monitoring network has 429 monitoring sites, out of which 313 sites on agricultural soils and soils above the forest line, 111 sites on forest soils, and 5 sites above the forest line (Fig. 1.6). All monitoring sites are geodetically surveyed and documented. It is possible to use not only areal but also soil spatial data.

2

Soil functions

The utilization of soil productive function by a man, i.e. the soil ability to produce crops, has already had a long history because it has existed from the beginning of the Neolithic (earlier Stone Age). In Slovakia the agricultural usage of soil was documented 6 thousand years ago. So far a man has perceived all this era that soil has only this productive function. The existence of non-productive functions, i.e. those which do not participate in the biomass production, were aware only in the 60th and 70th of the last century. In addition, Blum (1990) defined the following soil functions:

- biomass production,
- filtering, and buffering capacity,
- environment for organisms, gene reserve,
- area for the development of infrastructure,
- resource of non-renewable raw materials,
- resource of archaelogical and paleontological discoveries.

Later the soil functions were presented also by the Council of Europe R(92)8 on soil protection (1992) as follows:

- biomass production as the basis of human and animal life on the Earth,
- filtration, neutralization, and transformation of substances in nature as a part of the functional and regulative mechanisms of nature,
- preservation of ecological and genetic potential of living organisms in nature (biodiversity of species),
- the spatial base for human economic activities (agriculture, forestry, industry, transportation, construction industry, tourism, etc.) and social security of inhabitants (employment, nutrition, incomes),
- reserve and source of raw materials (water, clay, sand, rocks, minerals),
- cultural heritage of countries and the Earth, including paleontological and archaeological artifacts.

Thus, the soil functions can be divided into **productive and non-productive** (Vilček, 2011).

The productive function – is the soil‘s ability to provide plants with the conditions for their growth and development. This assessment according to the cultivated **phytomass – yield** is characteristic of the agricultural soils. Therefore, the production function is perceived in the context of fertility. Hraško et al. (2010) claimed that the suitable term is “soil production capacity“ for the quantitative evaluation of soil fertility. The authors recommend the term “soil production potential“ to forecast the following development of the soil production capacity. This term defines the optimal possible level of the soil production capacity in a particular area and anticipated period, which becomes evident by the optimal production of a particular crop, or culture without the serious violation of the balance of factors and biological stability of the environment.

It is obvious that the significance of soil for society results from its functions. Although so far the attention of a man has been concentrated immediately on the production function used in agriculture and forestry, the professional view of soil has recently been considerably extended by so-called soil **non-productive functions**. Demo et al. (1998) assign the non-productive functions: filtering function, buffering function, transformation function, accumulation function, transportation function, soil as a biological habitat and gene reserve, sanitation function, soil as a historical medium, soil as a source of energy and raw materials, soil as a space for the human activity.

Bujnovský (2009) divided the soil functions into two basic groups: **ecological and socioeconomic functions**.

From the aspect of stabilization of landscape environment and human living conditions and health the soil **ecological functions** are:

Productive function – constitutes the soil capacity to insure the plant requirements of nutrients, water, air, and growth of root system, and thus to create the conditions for the crop production, production of utility phytomass, plant phytomass, and animal biomass, which seem to be useless for a man, as well as the restoration of energy (soil organic matter).

Filtering function – involves the water accumulation and transportation, and also the soil capacity to capture different substances entering into the soil. The accumulation of water in soil has an impact on both the regulation of the water cycle in the landscape and the soil production capacity. The filtration, or immobilization of substances in soil, inhibits the undesirable transportation of contaminating matters into the ground waters, or into the food chain.

Transformation function – includes the change of substances by physical, chemical, and biological processes in soil.

Buffering function – consists of the soil‘s capacity to inhibit the impact of chemical substances on the changes of parameters, or soil properties. Besides, soil, more precisely soil water and organic matter is able to mitigate the thermal fluctuation of soil, or the ground layer of air.

Carbon reservoir – this function is important from the view of mitigation of development, or consequences of climatic change.

Biodiversity reservoir – this function creates the environment for living organisms, and thus also gene reserve. Generally, the deterioration of soil properties leads to the biodiversity reduction. As a rule, the soil biological degradation is the consequence of the physical and chemical degradation.

The socioeconomic soil functions are:

- **the physical environment for people and human activities** (housing construction, development of industry, construction of landfills, sport and recreational activities),
- **the source of raw materials,**
- **preservation of geological and archaeological heritage.**

3

Soil mineralogy

From the aspect of origination and composition, soil mineralogy is determined not only by the mineralogical but also by chemical and texture soil composition. The earth‘s crust is formed by rocks. These rocks are classified according to the way of origination, mineralogical composition, occurrence, geological age, and texture. According to the origination the rocks are divided as follows: eruptive, sedimentary, and metamorphic rocks. They are liable to the continuous weathering processes that activate the changes in the structure, composition, and texture of rocks and minerals, as it is indicated in Figure 3.1.

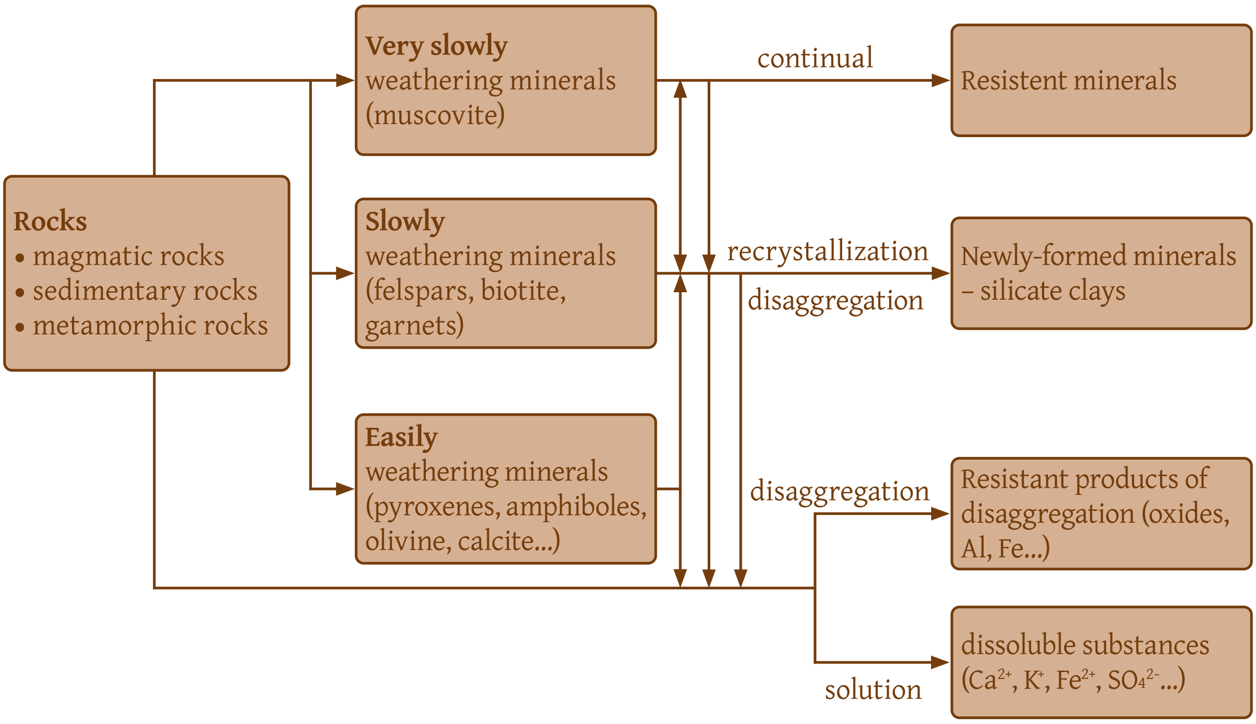


Figure 3.1: Scheme of rock weathering processes (Zaujec et al., 2009)

3.1 Weathering

Mechanical weathering

Mechanical (otherwise indicated also physical) weathering is a mechanical decomposition of solid parts of a rock into the fragments (clusters) of smaller size fractions. In fact, the mineralogical (also chemical) composition is not changed. Thus, only the surface area of the rock is being changed. These exposed – fresh surfaces of the crumbling rocks result in further processes of weathering.

One of the important factors of mechanical weathering is **insolation**, i.e. the change of intensity of the solar radiation and the related changes in temperature during day and night. Sometimes this change can achieve the fluctuation of temperature by tens of degrees (under tropic conditions it can be even 50 °C). The different thermal conduction and convection of the particular mineral grains in a rock activate the tension in the rock, or its volume changes, which causes the creation of cracks, and breaches of different sizes and distribution during the cyclic repetition of the changes in temperature. The rock decay is also accelerated by the heterogeneity of minerals in the rock, e.g. the dark minerals tend to absorb more heat (solar radiation) than the light ones.

Another factor is the **freezing and thawing of water** (water in the solid phase increases its volume by about 9%). During the periodical freezing and thawing of water, the breaches and pores are consequently enlarged, the new areas are exposed (pores/voids) and water penetrates deeper into the rock layers. The destructive effects of water become evident, especially in the high altitudes where the temperature varies around the water freezing point – i.e. below 0 °C. The created sharp-edged cuttings from the defrosted rocks are accumulated in the form of stony debris (creating the colluvia) at the foothills and slopes of the mountain ridges (Hovorka, 1999).

Chemical weathering

In the processes of chemical weathering, the primary rock-forming minerals are mostly decomposed. As a rule, the minerals with less chemical resistance against water and solutions are changed chemically first. Out of the released elements (ions) the new more stable compounds are formed (they are denoted also as the secondary minerals) (Dávidová, 1999). Depending on the surrounding conditions and petrological composition of rocks the final product can be also hydroxides, clay minerals, colloidal silicic acid, carbonates, chlorides, sulfates, and other acid salts. The minerals, that have a high level of mechanical and chemical resistance (difficult to degrade), such as quartz, tourmaline, zircon, rutile, magnetite, and others are accumulated and remain in geests.

The effect of the essential factors of chemical weathering – water, partial gas pressure O_2, CO_2, SO_2, inorganic and organic acids, humous matters, and the activity of microorganisms depends on the pressure, temperature, and pH – concentration of hydrogen ions in the environment. The most significant processes of chemical weathering are **dissolution, hydration, oxidation-reduction processes, processes of hydrolysis, and carbonatization** (Zaujec et al., 2009).

Dissolution is considered to be one of the basic forms of chemical weathering. The easily soluble minerals are nitrates, chlorides, and sulfates that dissociate fast in the aqueous medium. From the viewpoint of rock and mineral solubility in the aqueous medium, the carbonates underlie easier than silicates to the solubilizing processes. As the water in the natural environment cannot be assessed to be "pure", it comprises several positive and negative ions – cations and anions, dissolved gases CO_2, O_2, and others, we do not consider this process to be just physical but also a physicochemical one.

From the agricultural aspect, one of the most significant processes is **the dissolution of carbonates**, mainly calcium carbonate ($CaCO_3$ in the form of calcite, aragonite, etc.). This reaction is accelerated by the impact of CO_2 in water.

$$CaCO_3 + H_2O + CO_2 \rightarrow Ca(HCO_3)_2$$

Hydratation – is the process where the anhydrous minerals become hydrous, e.g. anhydrite becomes gypsum, and hematite is changed to goethite.

$$CaSO_4 + 2H_2O \rightarrow CaSO_4 \times 2H_2O$$
$$Fe_2O_3 + H_2O \rightarrow Fe_2O_3 \times H_2O$$

In the processes of hydration also the changes of mineral (rock) volume occur. The hydratation is almost always associated also with the processes of oxidation, e.g. iron in biotite, in siderite, in pyroxenes, in olivine, in amphiboles, etc. In the process of oxidation the changes of oxidation number occur from Fe^{2+} to Fe^{3+}, whereas the hydrated forms e.g. of iron in limonite (the mixture of oxides and hydroxides Fe).

Hydrolysis – ranks among the most important processes of chemical weathering when the salts are decomposed by dissociated water (H^+ and OH^-). The cations of hydrogen substitute (emboss) preferably other cations similar in valence from the structural grid, e.g. Na^+, K^+, Ca^{2+}, Mg^{2+}, and others, and remain in their place. In the presence of OH-groups the released cations, e.g. Na^+, K^+, Ca^{2+}, Mg^{2+}, and others can be bound with anions OH^- and hydroxides $Ca(OH)_2$, KOH, and others are being formed. Hydrolytic decomposition occurs most frequently with minerals – silicates.

$$\underset{\text{orthoclase}}{K(AlSi_3O_8)} + 3H_2O \rightarrow 2KOH + \underset{\text{kaolinite}}{Al_2(Si_2O_5)(OH)_4} + 4SiO_2$$

$$\underset{\text{anorthite (Ca-plagioclase)}}{Ca(Al_2Si_2O_8)} + 6H_2O \rightarrow 2Ca(OH)_2 + \underset{\text{kaolinite}}{Al_2(Si_2O_5)(OH)_4} + 3SiO_2$$

During the processes of hydrolysis also the process of hydration can be freely in progress. The typical examples are the formed clay minerals that contain adhesive water in the form of OH^- or H_2O groups in their structure. Overall, the formed minerals (reaction products) depend on the chemical and mineralogical composition of the original rock and other factors in the environment.

The degree of water dissociation is adequate to temperature, duration, and intensity of impact, the content of present substances, etc.

Oxidation – is a process where air oxygen is absorbed by rain – atmospheric water and oxidates the minerals with the present iron (in the form Fe^{2+}), manganese in the form of (Mn^{2+}), etc. In the process of oxidation, the change in the valence of elements occurs (oxidation state of an element), as well as the ionic radius, the position of the element the crystal structure, and others. The oxidation state of iron in the minerals of magmatic rocks (biotite, amphiboles, etc.) is Fe^{2+}.

On the earth's surface (in atmospheric conditions) the dominating bivalent form is changed into the trivalent form. This process is accompanied also by the change of color of the individual minerals. Fe^{2+} in the magmatic minerals cause greenish, green-blue, bluish even grey colors. The oxidated forms Fe^{3+} have dominant red, yellowish, rusty even brown colors. a similar situation is with manganese. The compounds Mn^{2+} have pink even or grey-purple colors, and after the process of oxidation they are becoming brown even black (Čurlík, 2011).

The model example of oxidation in the natural environment is the change of magnetite into hematite and limonite (mixture Fe oxides and hydroxides).

$$2(FeO \times Fe_2O_3) + O + nH_2O \rightarrow 2\ FeO + O + 2Fe_2O_3 + nH_2O \rightarrow 3\ Fe_2O_3 \times nH_2O$$

In the oxygen-free compounds (minerals) the typical oxidation takes place, e.g. sulfides – pyrites which is changed first into ferrous sulphate and after the following oxidation into ferric sulfate and sulfuric acid.

$$FeS_2 + 7\ O + H_2O \rightarrow FeSO_4 + H_2SO_4$$

Reduction – (reduction processes) is mostly in progress with the insufficiency of oxygen in the environment, e.g. in heavy soaked soils where Fe^{3+} and Mn^{3+} are reduced, which is characterized by the blue-green, or light green color of minerals. The processes of reduction occur often also with the presence of organic matter (residues) that use the present oxygen during their decomposition (peats, moorlands, etc.).

$$Fe_2O_3 + H_2S \rightarrow FeO + FeS_2 + 2H_2O$$

Carbonatization – is the process that results in the formation of carbonates. In the processes of carbonatization, the presence of CO_2 is needed that absorb the rain waters (from the atmosphere), or it is acquired by the decomposition processes or by the active function of microorganisms in soils, marshlands, oceans, etc. The cations of alkali metals, mobilized e.g. by the process of hydrolysis, react with CO_2.

$$KOH + CO_2 \rightarrow K_2CO_3 + H_2O$$
$$Ca(OH)_2 + CO_2 \rightarrow CaCO3 + H_2O$$

Biological weathering

The biological weathering is mostly evident in the areas with the appropriate conditions that support the active functioning of microorganisms. They can have an impact to a certain extent on the rock destruction (minerals) both in the chemical and mechanical way (Kolenčík et al., 2014). The first impulse for the beginning of weathering processes in mountainous areas is the colonization of the rock surface by mosses and lichens. The process of decomposition of rocks and minerals by microorganisms is called **biodeterioration** (Kolenčík et al., 2014). The effect of microorganisms excludes the participation of organic acids (citric acid, oxalic acid, and also carboxylic acids), which destroy the rock surface. Similarly, also the organic residues in soil (humic acid and formation of organo-mineral complexes), plant root systems, animal excrements, and dead plant, and animal bodies can facilitate the decomposition of rocks (Zaujec et al. 2009).

Biomechanical impact – is mostly typical of hyphae of microscopic (symbiotic) fungi, plant roots that penetrate into the weakened rock zones (breaches) and by their further growth, they cause the exposure of new surfaces in rock, etc. The processes of weathering are accelerated also by the activity of animals that deepen the underground corridors, or burrows like earthworms, larvae of different insects, rodents, moles etc.

Impact of climatic conditions on rock weathering

In general, the intensity of weathering depends on climatic conditions. Therefore, the weathering types can be divided according to the climate:

- humid type of weathering – warm and humid climate,
- arid type of weathering – dry climate dominates,
- arctic type of weathering – cold climate dominates.

3.2 Chemical composition of Earth crust

The earth‘s crust (lithosphere) consists of eight basic chemical elements. In fact, only oxygen and silicon constitute 75%. On the other hand, many elements which occur in less quantity in soil have a significant impact on plant and animal growth. It is also important to mention that the division of the individual elements in the soil is different. For example, in some places a high concentration of phosphorus in the soil is present. On the other hand, there are locations where the insufficiency of this element exists for plant growth and development. The presence of the particular elements in the soil is indicated in Fig. 3.2

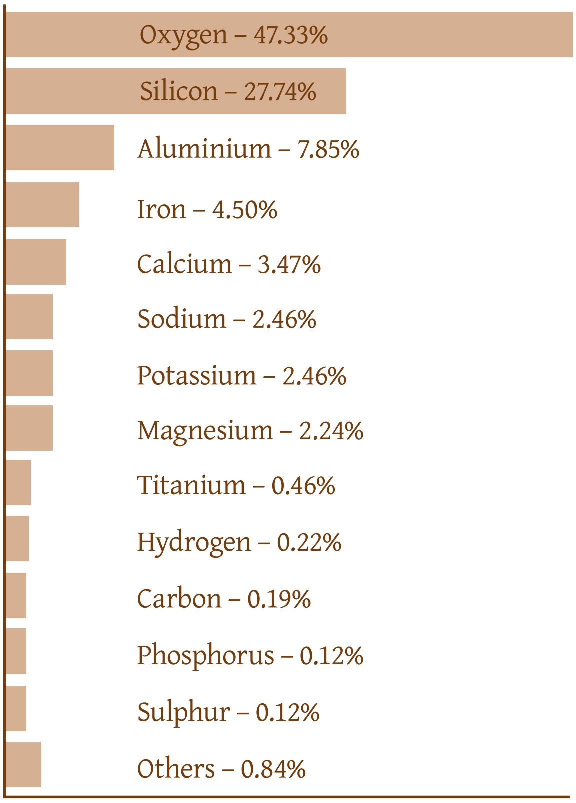


Figure 3.2: Composition of the solid crust of the earth to a depth of 10 miles (Millar et al., 1962)

3.3 Basic groups of rocks

In principle, soil originates from the surface parent rocks along with the complex synergism of many soil-forming factors. Many soils in Slovakia are affected by the parent rocks to a different extent and are also influenced by cultivation, which has an impact on their fertility.

In fact, rocks are a heterogeneous inorganic natural matter of different chemical compositions, composed of one or several minerals.

The rocks are divided according to their origin:

1. **magmatic:**
    - abyssal (granite, diorite, gabbro),
    - effusive (porphyry, andesite, basalt, tuffs)
    - rock vein (quartz, pegmatite),
2. **metamorphic:**
    - schists (gneiss, wart, phyllite),
    - amphibolic schists (amphibolite, serpentinite, granular limestone),
    - silicic rocks (chert, firestone),
3. **sedimentary:**
    - consolidated (sandstones, farewell rocks, clay shales, calcites)
    - softs (gravel, sand, loess, marl, clay).

The primary source of all rocks on the earths surface are the magmatic rocks. The essential part of these rocks consists of the modest number of minerals:

- **feldspars 59.5%,**
- **pyroxenes 16.8%,**
- **quartz 12%,**
- **micas 3.8%,**
- total more than 92%.

As quartz is resistant to chemical weathering, feldspars, pyroxenes, and micas underlie to the chemical changes. In the process of chemical weathering, the nutrients are released and the secondary minerals originate which have a crucial impact on the formation of physical, physicochemical, and chemical soil properties.

These minerals dominate in the magmatic rocks. According to the content of SiO_2 the magmatic rocks are divided into:

- **acid** – they contain more than 65% SiO_2. In the process of weathering, they form coarsely dispersed substrates with clay proportion <30% and the low content of alkaline cations (granites, siliceous diorites, dacites, liparites).
- **neutral** – they contain 52 – 65% SiO_2. In the process of weathering they create the substrate of more admissible texture, mineralogical and chemical composition (syenites, granodiorites, andesites, trachytes, clinkstones),
- **alkaline** – they contain 45 – 52% SiO_2, dark colored. As a result of a high content of alkaline elements they create the substrates for the formation of more fertile soils (gabbros, basalts, diabases, melaphyres),
- **ultrabasic** – the content of SiO_2 is below 42% and they are dark colored eridotites – dunites).

According to the level of weathering the rocks are divided into:

- **easily weathering**
    - from loose sediments calcic loess, loess loams, sandy and gravelly sediments,
    - from soild sediments clay rocks, schists, sandstones, pudding stones associated mainly with calcic or ferric, or clayey binder,

  - from schists relatively easily weathering are gneisses, wart, phyllites, amphibolic schists, serpentinites,
  - from the magmatic rocks gabbro, melaphyre, basalt, diorite, syenite and trachyte,
- **hard weathering:**
  - some clay schists and some acid eruptive rocks, such as siliceous trachyte, porphyry and granites.

3.4 Primary and secondary minerals

The primary minerals occur in soil in the form of residues of magmatic rocks. They constitute the principal reservoir of mineral nutrients for plants. These minerals have a slight impact on the physicochemical processes in soil because they form considerably big grains (they are the essential part of sand and dust), which have a small specific surface.

From the agronomical viewpoint, the value of minerals in the soil is determined according to the content and type of the available nutrients for plants. In the process of weathering, the minerals release the biogenic elements that are partially transferred into the soluble forms and they are partly bound to the soil colloids.

According to the quantity of minerals rich in the essential biogenic elements **the mineral power of soil** is assessed.

The mineral power of soil – constitutes the ratio of the weathered mineral proportion in soil with the content of biogenic elements to the total mineral proportion of soil. Based on the mineral power of soil Novák (1954) classified soils into particular categories (Table 3.1).

Table 3.1: Classification of soils according to their mineral power (Novák, 1954)

0 – 10	infertile soils
10 – 20	mineral weak soils
20 – 30	soils with medium mineral power
30 – 40	mineral strong soils
> 40	mineral very strong soils

The most frequently occurring minerals of the soil mineralogy are silicates, oxides, carbonates, etc. Depending on the genesis, i.e. the way of origination, these minerals are divided into **primary and secondary ones**.

3.4.1 Primary minerals

Quartz – α-SiO_2 is the most frequently occurring mineral of the continental earth crust, with a stable temperature to 573 °C. The quartz crystals (or their varieties) are greatly clustered in druses. It is characterized by the shape of a hexagonal prism that ends with the areas of a hexagonal pyramid. It can occur abundantly in the form of compact and granular aggregates. Quartz has different colors depending on its admixtures, most often grey and white. According to Mohs´ scale of hardness, its hardness is 7. It occurs most frequently in acid magmatic rocks, clastic sedimentary rocks, and metamorphites.

Magnetite – $Fe_2+Fe_3+2O_4$ is ranked among the most abundant iron ores. It contains approximately 72% Fe. Magnetite has a strong metal-black color and a black scratch. The hardness varies in the interval 5.5 – 6. It is strongly magnetic. Under the exogenic conditions (in the processes of the chemical weathering) it is relatively constant.

Hematite – α-Fe_2O_3 It contains approximately 70% Fe. The color of hematite is usually steel-grey, iron-black even reddish-brown, and it has a red scratch. The hardness is 5 – 6.

Goethite – α-$Fe_3+O(OH)$ is called according to the German poet (J. W. von Goethe). It originates mostly from weathering of minerals with a high content of iron. Goethite has the brown color and reddish-brown scratch. The hardness is 5. Its hydratation changes it into other iron oxides and iron hydroxides. In the sedimentary rocks and soils goethite is in the association along with a big group of iron oxides and iron hydroxides, where it fills the intergrained spaces as a putty.

Limonite – Its composition constitutes the mixture of hidden-crystal forms of aqueous oxides Fe, mainly geothite with the admixtures Al, Mn, Si, Ca, Cr, Ni, Zn, etc. It does not have any evident outer limitation (it is amorphic), it usually occurs in the form of masses and earthy heterogeneous plaques. Limonite is characterized by the brown, rusty even black-brown color. Its scratch is yellow, sometimes brown. The hardness is 4. Limonite is considered to be the final product of disintegration of Fe minerals. It is wide-spread in the nature. At the deposits it occurs along with Mn minerals in the form of putty and admixtures in all types of rocks.

Diaspore – AlO(OH) – contains approximately 45% of aluminium. The rhombic rock-crystals and microscopically stratified aggregates occur most frequently. Diaspore can be colorless, white even grey. Its scratch is white and hardness is 7. Diaspore originates usually under the exogenic conditions (in humid climate) as a product of so called lateric decomposition of aluminosilicates. It constitutes one of the most significant parts of laterites and bauxites.

Pyrolusite – $Mn^{4+}O_2$ – contains about 63% of manganese. It has the black color of scratch. The hardness in the interval 5 – 6. From the aspect of stability it ranks among the final (stabile) products of metamorphosis of the manganese minerals. The oxides and hydroxides in soils are almost always in the association with the oxides and hydroxides of iron (goethit, magnetite, etc.) (Čurlík, 2011).

Feldspars – aluminosilicates:

The feldspars constitute a very important part of magmatic rocks. The most significant feldspars are **potassium and sodium-calcium feldspars – plagioclases**. Orthoclase is the principal source of potassium and plagioclases constitute the important source of calcium and natrium in soil. The feldspars are a part of mainly coarser soil fractions:

- **(i) Potassium feldspar (orthoclase)** $K(AlSi_3O_8)$ contains approximately 11% of potassium. Orthoclase has most frequently slightly pink even red, grey, and white colors with glossy lustre. It is an essential part of granites, pegmatites, syenites and porphyries. In the metamorphic rocks, it occurs mostly as a part of gneisses. In the sediments, it participates in the composition of the clastic (fragmentary) sediments, predominantly sandstones, arkoses, and loesses,

- » **(ii) Sodium-calcium feldspars (plagioclases)** form more-less irregular grains in rocks. Sodium feldspar (albite) Na($AlSi_3O_8$) together with calcium feldspar anorthite Ca($Al_2Si_2O_8$) create a sodium-calcium variety of plagioclases. It has white even grey color, often with a slightly pink, bluish, or greenish shade. They show the white scratch and glossy lustre. The hardness is 6.0 – 6.5. They are characterized by good cleavability in two different directions. In rocks, plagioclases occur more often compared with potassium feldspars. Their frequent occurrence can be monitored with the magmatic and metamorphic rocks, as well as several clastic sediments (Hovorka, 1999). K-feldspars are more resistant to weathering than plagioclases. The alkaline and neutral plagioclases disintegrate mostly much easier than the agents of acid plagioclases. In this way, the alkaline plagioclases release Ca into the environment, which along with the water-soluble carbonated forms, e.g. HCO_3, can support the formation of $CaCO_3$. Thus, plagioclases can be considered to be one of the most significant sources of "carbon" in soils (Zaujec et al., 2009).

Micas – aluminosilicates – are very widespread minerals. They have a scaly form and excellent cleavability. The most widespread are:

- biotite – $K(Mg, Fe)_3[AlSi_3O_{10}][OH, F]_2$ – dark mica,
- muscovite – $KAl_2[AlSi_3O_{10}][OH, F]_2$ – light mica:
  - » **Biotite** – contains about 5 – 10% K and cca 1 – 18% Mg. It mostly forms leafy, table, and short-columnar crystals. Similarly to all micas, it is characterized by excellent cleavability. Biotite has a dark brown, dark green, and black color (depending on the content of Fe) and also pearly lustre. It has a white scratch and hardness 2.5 – 3. It occurs in the magmatic rocks, mostly in the neutral pegmatites and metamorphic rocks, predominantly in gneisses (Hovorka, 1999). In the process of decomposition, the rocks are transported to the weathered lamellae (gold color) into fluvial deposits, where they accumulate. Biotite is also an important source of nutrients in soils because it releases the necessary Mg and K for the plant root system (Slovík and Libant, 1996).
  - » **Muscovite** – contains approximately 10% K. It has a grey-white, colorless, and faint-yellow appearance. It has an excellent cleavability conditioned by the crystal structure, and also glossy even pearly lustre.

3.4.2 Secondary minerals

In soils and sediments, the minerals occur that originated during rock weathering. These are secondary minerals. The most widespread are opal, calcite, less frequently aragonite, dolomite, magnesite, clay minerals and others.

Opal – $SiO_2.nH_2O$ ranks among SiO_2 masses that contain a certain percentage of water. It has the different color, waxen, obscure, greasy even glossy lustre and splintery, conchoidal breakage.The hardness varies in the interval 5.5 – 6.5. Opal is built by the basic silicon-oxygen tetrahedrons. The origination of opals is bound to the secondary processes, i.e. pitching from hydrosol mainly of silicic acid, that originates during the gradual processes of weathering and transformation from the original silicates (Slavík et al., 1974). Opal creates mostly the putty and filling of air holes and cracks in rocks. Depending on the conditions opal can have the different varieties, e.g. noble opal, milk, wax, wood, common opal and others. All silicon monoxides resist considerably against the chemical processes of weathering. They underlie mostly the mechanical decomposition. Shards of different sizes cluster on the earth's surface and are the dominant part of soils and clastic sedimentary rocks.

Calcite – $CaCO_3$ its nomenclature is derived from the Latin word "calcium". It often contains a minor quantity of admixtures – bivalent cations of Fe, Mg, Mn, and others. It has glossy lustre and different colors – pink, white, reddish, brown, or luteolous. Calcite has a white scratch and very good cleavability. The hardness is 3. It differs from quartz, barite, and anhydrite by the reaction in HCl and the different hardness. It also differs from the similar aragonite by cleavability. Calcite is usually formed during the weathering of minerals enriched by Ca part, e.g. form plagioclases (Ca-feldspars). It can also originate from the simple crystallization of springs, and underground waters with the relatively higher content of component $Ca(HCO_3)_2$. From the aspect of rock-forming significance, calcite is the most widespread and most dominant carbonate in all. It constitutes the basic component of limestones and marbles, travertines, chalk, schists, the essential part of marls, calcium clays, sealing part of sandstones (Hovorka, 1999; Ružička et al., 2013). Its content in loesses is important. Calcite has significant importance in soils, where it has a positive impact not only on the chemical and physical soil condition but also on the soil solutions and soil water, it increases the buffering capacity of soils, etc. (Čurlík, 2011).

Aragonite – $CaCO_3$, its name is derived from Aragon, the area in Spain. It has white, faint-pink, and faint-yellow color. Aragonite has a white scratch, hardness is 3.5. The cleavability is missing. It reacts sharply in HCl. Its occurrence in nature is rare compared with calcite. It originates similarly to calcite from the solutions enriched by the component $Ca(HCO_3)_2$ at a temperature above 35 °C.

Dolomite – $CaMg(CO_3)_2$ is called according to the French mineralogist Dolomieu. It contains approximately 13% of Mg, and 22% Ca, along with different admixtures, like Mn, Fe, etc. It has a white color with brown and yellow shades and white scratch. The hardness is 3.5. Dolomite is cleavable very well, it reacts slightly or at all with diluted HCl. Its solubility is a bit lower than with calcite. Its genesis is bound to dolomitization, where Ca is changed, i.e. replaced by cation Mg (Slovík and Libant, 1996).

Magnesite – $MgCO_3$ the origin of the word comes from the Latin "magnesium", which means the chemical element magnesium. It contains approx. 29% Mg. Magnesium can be replaced also by other cations Ca, Mn and Fe, predominantly because of the common affinity, geochemical distribution, similar ionic radii, valence, etc. It has more frequently grey-white, grey, brown, or faint yellow color. The compact magnesite is usually white and its scratch is also white. The hardness is 4 – 4.5. Magnesite has good cleavability and it does not react to cold HCl. The opposite situation occurs during the application of hot HCl, where magnesite is decomposed.

Clay minerals

One of the most important groups of secondary minerals in soils are **clay minerals** which originated during the weathering of the primary minerals. More than a half of heretofore known (original) minerals is created by silicates. Clay minerals differ from the primary siliceous minerals by their chemical composition, stratiform structure and high dispersity, mostly at the level of the colloidal dimensions. They constitute the main part of clay fraction of soils. The essential structural elements of clay minerals are **silicon-oxygen tetrahedrons** – SiO_4^{4-} (they can be also AlO_4^{5-}, or FeO_4^{5-}) and **octahedrons** most often $Al_2(OH)_6$, $Mg_3(OH)_6$ a $Fe_2(OH)_6$ (Fig. 3.3).

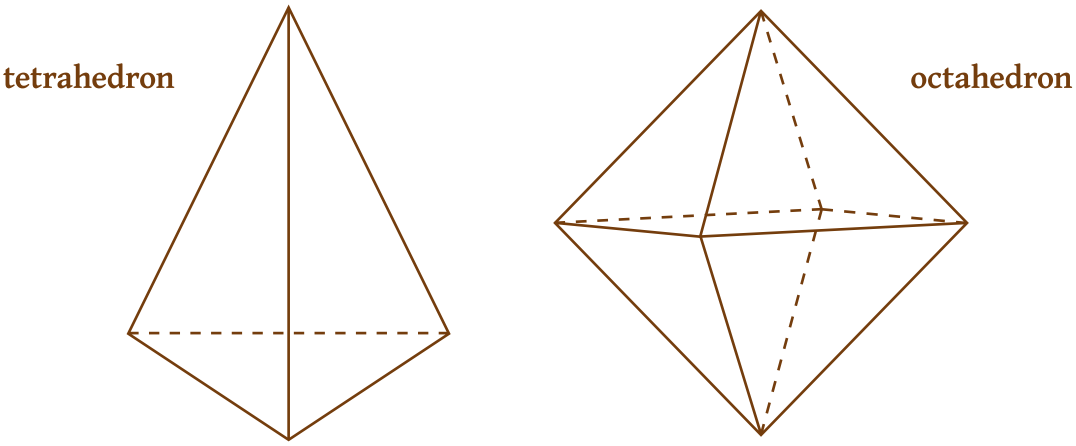


Figure: 3.3: Form of tetrahedron and octahedron (Kohnke, 1968)

Clay minerals are classified according to the endogenic arrangement:

1. **With a bilayer**, which has one layer created from tetrahedrons and one layer from octahedrons (otherwise they are referred to as the symbol 1:1, i.e. one layer of tetrahedrons and one layer of octahedrons). The small layers of tetrahedrons and octahedrons are placed opposite each other with regular repetition.
2. **With trilayer,** where the layer of octahedrons is surrounded by two layers of tetrahedrons (it has the conventional symbol 2:1). With the symmetric repetition of the packet – the trilayers are the layers of tetrahedrons always turned to each other. The mutual bonds between them are weak (the interaction at the level of Van der Wals forces), whereas the mutual distance of tetrahedrons can be changed (decrease and increase, e.g. depending on the water content).
3. **With the mixed structure,** which is formed as a result of weathering, the overgrowth of two different structures where one type of layer transforms earlier than the others. This type of structure is often referred to as the initial letters of the dominant clay minerals, or their changes, e.g. IK changing of illite and kaolinite, IM – illite and montmorillonite, etc.

Properties of clay minerals

Isomorphic substitution – the replacement of silicon with aluminum in the tetrahedron bonds and the substitution of aluminum by magnesium and iron in the octahedron bonds. The substitution of ions in octahedrons is more frequent, while the replacement in tetrahedrons is rare.

4. **Clay minerals swell during contact with water and they shrink after the water loss** – as a result of the ability to expand and contract the structural grid. This ability of the volume

changes is different with clay minerals. It depends on the endogenic structure, chemism, size, and form of particles and exchange ions. For example, kaolinite clays increase their volume by 5 – 60%, halloysite clays by 70%, and Ca-montmorillonite clays by 45 – 145%. The changes in volume are important for the physical condition of the soil, and the changes in bulk density are significant for the origination of the structural soil aggregates.

5. **Thanks to their specific surface clay minerals have a sorption capacity.** The change in the specific surface of clay minerals is shown in Table 3.2.

Table 3.2: Changes of properties and specific surface of clay minerals (Lal and Shukla, 2004)

Clay mineral	Cation exchange capacity (cmol.kg^{-1})	Anion exchange capacity (cmol.kg^{-1})	Change of volume (cmol(+) m^{-2}.10^{-3})	Specific surface (m^2.g^{-1})
Kaolinite	3 – 15	–	60 – 75	5 – 20
Illite	10 – 40	–	10 – 20	100 – 200
Vermiculite	100 – 150	5 – 10	30 – 33	300 – 500
Montmorillonite	80 – 150	–	11 – 19	700 – 800
Allophane	20 – 30	10 – 20	–	>600

6. **The sorbed ions are being retained in the exchange form** – i.e. they can be substituted, or displaced by other ions. The exchange ions are held at the external edges of structural layers, therefore, as a rule, the exchange reactions do not have an impact on the own structural composition. In some cases, the cations after sorption by clay minerals underlie sparely the next exchange. It is so-called non-exchange sorption. For instance, potassium sorbed by degraded illite can be fixed and underlie the regradation (improvement) of illite.
7. **Clay minerals form a big amount of organomineral compounds and complexes in soils.** The mineral and organic parts of soils are interconnected. During the mutual impact of humic acids and their salts on the clay minerals, organomineral compounds are created. The negatively charged surface of clay minerals can bind also the organic cations. The most durable complexes originate in the soils with a sufficient content of montmorillonite, humic acids, and calcium humates. The organomineral compounds influence positively the important soil properties – the formation of structure, sorption capacity, and preservation of organic substances against microbial decomposition.

Most important groups of clay minerals

1. The group of kaolinite (with bilayer)

Apart from kaolinite there belong also the minerals with the same stoichiometry $Al_2Si_2O_5(OH)_4$, but the different crystal structures, i.e. nakrite, halloysite, and dickite. The bilayers are associated directly via the mutual peaks of the tetrahedral and octahedral anions. The mutual ratio of the component SiO_2 to Al_2O_3 is mostly 2 : 1. The minerals of this group constitute the reaction product of subtropic, tropic, and humid weathering. In soils, they affect negatively because of their low sorption capacity.

Kaolinite $Al_2Si_2O_5(OH)_4$ – has the appearance of submicroscopic small pseudo-hexagonal columns or scales. It has white and even faint yellow color. The polymerization of tetrahedrons of kaolinite structure (Fig. 3.4) does not allow more significant expansion of the interlayer space, therefore the cations are bound scarcely and the sorption of the exchangeable base is considerably limited.

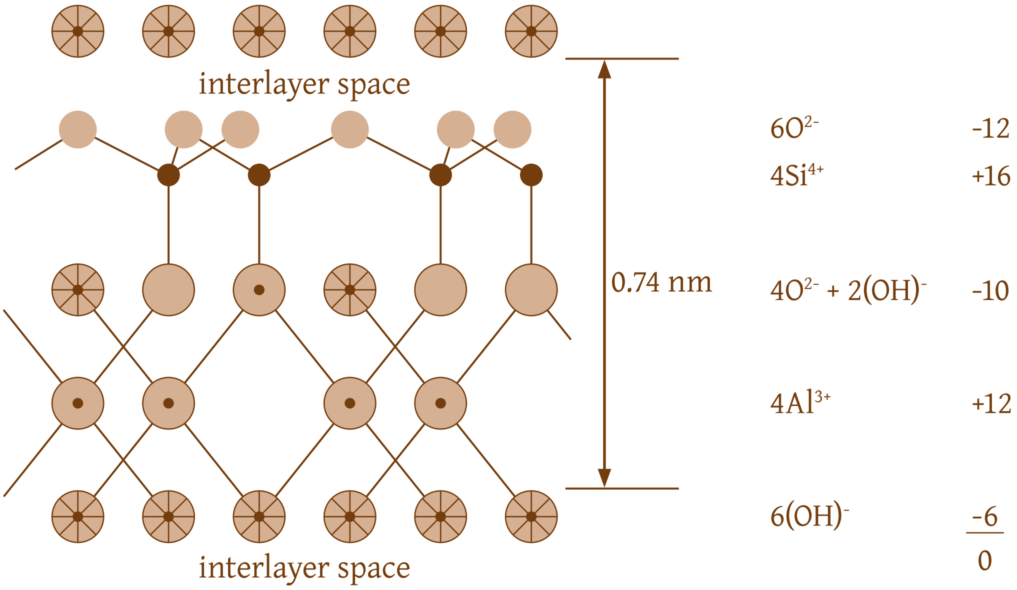


Figure 3.4: Structure of kaolinite (Singer and Munns, 1996)

Kaolinite is formed mostly in the acid medium (pH), where the determining factor is the chemical type of weathering from the primary aluminosilicates, predominantly micas, and feldspars. The processes of kaolinization affect acid magmatic rocks (liparites, granites, syenite, etc.), metamorphic also sedimentary rocks (gneisses and arkoses with a higher content of feldspars).

Halloysite – has usually the appearance of earthy aggregates of white, or grey-white color with a light greenish, yellowish, and brown shade. It has a greasy abrasion, it can be scraped by a nail. After drying the compact masses are formed which have the typical conchoidal breakage during decomposition. The genesis of halloysite is predetermined in the areas containing aluminosilicates of alkaline rocks and by the weathering processes occurring in the acid media. Its occurrence is frequent in soils enriched by organic acids and it occurs in the mineralogical association with kaolinite.

2. The group of montmorillonite (with the basic trilayer)

Apart from montmorillonite there belong also

- beidellite $(Na,Ca)_{0,3}Al_2(Si,Al)_4O_{10}(OH)_2\times nH_2O$,
- saponite $(Ca,Na_2)_{0,15}(Mg,Fe^{2+})_3(Si,Al)4O_{10}(OH)2\times 4H_2O$,
- notronite $Na_{0,3}Fe^{3+}{}_2(Si,Al)4O_{10}(OH)_2\times nH_2O$.

The mutual bonds between the layers of tetrahedrons and octahedrons are strong enough that they are not disintegrated under the standard conditions. Only the atoms of oxygen are free, which are localized in the tops of external tetrahedrons, therefore only weak bonds –

interactions (Van der Walls´) prevail between them. That leads to the increasing distances of the trilayer packets. The molecules of water enter into these interlayer spaces, which causes on one hand big volume changes of clays, on the other hand, the easy ion binding in these positions.

A good sorption capacity is directly conditioned also by the substitution of silicon in tetrahedrons and aluminum in octahedrons by the ions of lower valence. In the process of exchange of elements the charge is changed – a certain negative charge is formed, where easier sorbed (exchange) ions occur consequently. Out of the bivalent alkaline cations Ca, Mg, etc. are changed most frequently.

Montmorillonite $(Na,Ca)_{0.3}(Al,Mg)_2Si_4O_{10}(OH)_2 \times nH_2O$ – ranks among the group of aqueous clay minerals that contain Ca, Mg, Fe, Al (Na), thus we can define precisely its chemical composition. Its name is derived from the area " Montmorillon" in France. It has the appearance of microcrystalline aggregates of light green, white-grey, and light yellow color. The crystals can be followed only by the analyses in the electron-optical microscope. They often form mineral associations with other clay minerals (e.g. hydromicas and kaolinite). The crystals are intimately adherent to them and create so-called "mixtures". Montmorillonite is typical of its high exchange capacity, mostly of cations, and a big volume changes with the sufficient quantity of water – by swelling. The scheme of the montmorillonite structure is indicated in Figure 3.5.

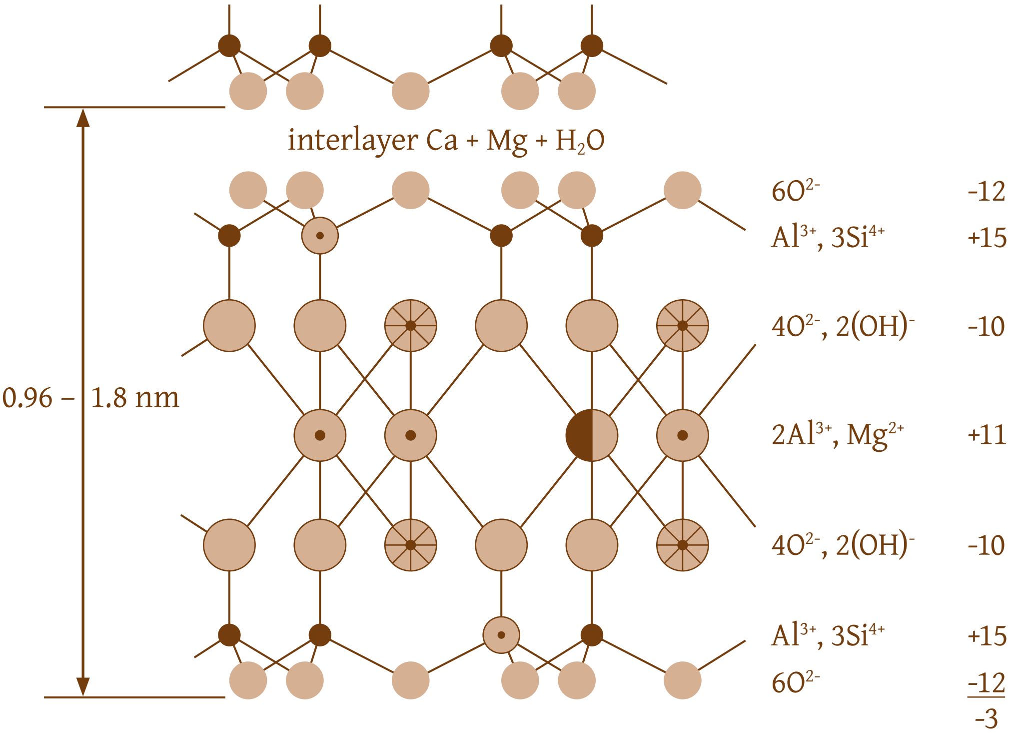


Figure 3.5: Structure of montmorillonite (Singer and Munns, 1996)

3. The group of hydromicas (internal trilayer structure)

We range clay micas into this group which represents a certain transition phase between micas and hydromicas – clay micas. They have an unstable chemical composition where mostly the content of water and cations is being changed. The monovalent cation of potassium can be substituted by bivalent cations of Mg and Ca.

Illite– aqueous aluminosilicate Al and K. It has the appearance of fine scales, mostly white or even light yellow color. In the presence of water, it is able to expand in the interval of 15 – 120%. It occurs in clays, loesses, and marls and it is a common part also in our soils (Slovík and Libant, 1996).

Vermiculite $(Mg,Fe^{2+},Al)_3(Si,Al)_4O_{10}(OH)_2 \times 4H_2O$ – is also aqueous aluminosilicate Fe, Mg. It has green or even brown-green color. It has a greater sorption capacity in comparison with montmorillonite.

Glauconite – is aqueous aluminosilicate Mg, K, and Fe of unstable changeable chemical composition. Its role in soils is significant mainly in connection with the easy release of potassium and the ability to bind and release cations. Therefore, glauconitic rocks are naturally usable as potassium fertilizers in the soil environment.

4

Soil organic matter

The organic matter of soil constitutes approximately 1 – 5% of the solid soil phase and it is an important segment in soil origination and development. It participates in the soil-forming processes by its impact on the mineral soil matter, which results in the formation of the pedosphere.

All processes occurring in soil are directly or indirectly associated with the vigorous activity of microorganisms and the effectiveness of humic substances. The living and dead components of the organic soil fraction affect each other. The organisms are the natural source of matters creating the nonliving organic mass which provides backwards the appropriate living conditions for organisms and the necessary source of nutrients and energy. In soils living and nonliving components are present more frequently in the A horizons/top soil (upper soil layers) where the most intensive biological activity occurs, associated with the formation and decomposition of organic matter with participation of the mineral soil fraction.

Table 4.1: Biomass of constituents of living component of organic mass (Brady and Weil, 1999)

Group of organisms	Average biomass (g m^{-2})
Plant roots	2,000
Microorganisms	400
Earthworms (*Lumbricidae*)	6
Animalcules (*Protozoa*)	5
Nematodes (*Nematoda*)	5
Enchytraeidae (*Enchytraeidae*)	3
Diplopods and chilopods (*Diplopoda a Chilopoda*)	1.1
Acarids (*Acarina*)	0.6
Isopods (*Isopoda*)	0.3
Collembolans (*Collembola*)	0.1

Living organisms

The living component of organic share of soil is created by plant roots and edaphone that represents a complex community of soil-dwelling organisms. Edaphone according to organism groups is divided into:

a) phytoedaphone – flora,
b) zooedaphone – fauna.

According to organism size:

1. Microedaphone – represented by all soil microorganisms of the plant (microflora) and animal (microfauna) origin. The main representants of microflora are actinomycetes, bacteria, cyanobacteria, algae, and fungi. Microfauna is represented by protozoa,
2. Macroedaphone – small vertebrates,
3. Mezoedaphone – includes larger worms, insects, larvae and adults, molluscs, and beetles.

According to how long or what part of life they spend in the soil, organisms forming edaphone can be divided into:

a) geobionts – they depend on the soil environment for their whole life,
b) geophiles – their development takes place in the soil, the rest of their live is outside it,
c) geoxens – they appear in the soil only by chance.

4.1 Soil organic matter

Soil organic matter is a complicated, heterogeneous, polydisperse set of organic substances of various origins with variable structure, degree of dispersing, and activity and thus in relation to the other components of soil matter and to living organisms. It is formed concurrently with other soil components in soil-forming process by the activity of soil-forming factors. We understand the term soil organic matter as a set of all dead residues which can be in a different level of change, it means that under this term, dead bodies of plants and animals are marked together with the original anatomical structure preserved. Soil organic matter affects a number of features and processes taking place in the soil.

4.1.1 Carbon cycle

There is a total of 790 Pg (Pg = 10^{15} g) of carbon (C) in the atmosphere, 620 Pg in vegetation, 3 – 4,000 Pg in soil, 5,000 Pg in fossil resources, 38,000 Pg in oceans, and 75,000,000 Pg in carbonates. The global carbon cycle is shown in Figure 4.1.

The numbers by arrows show amounts of C annually flowing (Pg/yr) by various processes between the pools. Note that the soil contains almost twice as much carbon as the vegetation and the atmosphere combined. Imbalances caused by human activities can be seen in the flow of carbon to the atmosphere from fossil fuel burning (7.5) and the fact that more carbon is leaving (62 + 0.5) than entering (60) the soil. These imbalances are only partially offset by increased absorption of carbon by oceans. The final result in that some 225 Pg/yr enters the atmosphere while only 215 Pg/yr of carbon is removed. It is easy to see why CO_2 levels in the atmosphere are rising.

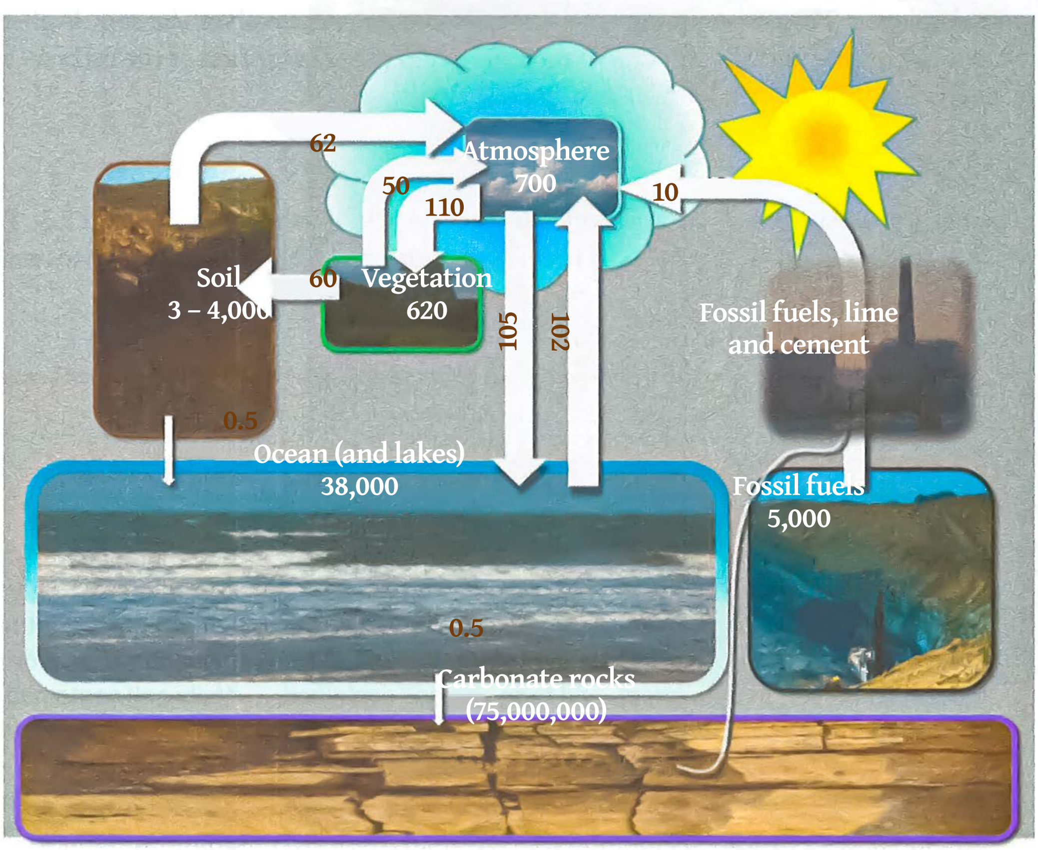


Figure 4.1: The global carbon cycle (Weil and Brady, 2017)

Sequestration of carbon in the soil

Organic carbon sequestration in soils is an effective strategy for mitigating global climate changes during the 21st century. In parallel with ensuring soil quality improvement, phytomass production and environmental quality improvement, the organic carbon sequestration strategy in soils on a global scale allows humanity to gain time for alternative energy sources from sources other than fossil fuels.

Methods of carbon sequestration in soils

Introduce technologies increasing soil organic matter content through:

- biodiversity – high biomass production in sowing procedures,
- catch crops cultivation,
- use of reduced tillage methods,
- use of rotating or limited grazing.

Change the dynamics of organic matter transformation in soils:

- create barriers to wind and water erosion by increasing the surface cover of the soil with plant residues,
- increase the production of more organic matters and nutrients by increasing production of crop residues and catch crops,
- reduce organic matter losses from soils by reducing tillage.

4.1.2 Sources of soil organic matter

We can divide sources of soil organic matters into:

1. **Plant residues:**
   - about-ground – dying parts of plants during vegetation, post-harvest crop residues, forest litter,
   - roots biomass – root hair dying during vegetation, dead root system, root exudates.
2. **Animal residues:**
   - metabolism products of macro fauna, mezo fauna and micro fauna,
   - dead animal.
3. **Microbial biomass:**
   - metabolism products of microorganisms during their life,
   - dead cells.

Dead plant residues are the most abundant and crucial source of organic matter which gets into the soil. The amount of litter and roots that penetrate into various depths of the soil profile varies considerably. It depends on the plant community and its soil-climatic conditions. The major types of plant communities that they decide on the formation of organic matter on land are **forest and grass communities**.

Forrest communities are typical for their great diversity and richness of the composition of plant formations located on several floors. The major part of the forest community is annual waste deposited on the surface in the form of forest litter. The amount of forest litter depends on the climatic zone, composition, age, density of trees, and the presence of grasses and mosses. In the case of forest communities, dying residues of the above-ground parts of trees are decisive for the formation of humus which is deposited on the soil surface in the form of litter consisting of needles, leaves, and twigs. Even though the root system penetrates into considerable depths still is, in comparison with the about-ground part a negligible source of soil organic matter. Annual forest litter in our conditions (Slovakia) represents 5 – 20 t ha^{-1} on average. Of course, the amount of fall produced depends on the climate zone.

Grass communities are becoming the major agent's organic matter accumulation wherever wood cannot grow and develop. Maximum production of grass communities is in conditions of humid meadows, prairies, savannas, and meadow steppes. Grass communities primarily provide root matter for humus production that is deposited directly in the soil. Organic matter storage produced by grass cover as well as annual gain and chemical composition is closely related to the type of community, its density, and geographically widespread. In comparison with forest litter, dead plant residues are richer in easily decomposable organic compounds – proteins and ash. Annual gain of dead root matter may represent on average 3 – 15 t ha^{-1} in a meter-high soil layer; however, the gain under annual and multiannual grass communities is different. As with forest communities and grass communities, the climate has a decisive influence on biomass production. Annually, grass communities produce a larger amount of underground biomass in comparison with aboveground biomass.

In agroecosystems, the primary source of organic matter is **crop and root biomass**, and the secondary is organic manure (farmyard manure, organic amendments, etc.).

Table 4.2: Total of organic matter in soil in various ecosystems (Schesinger, 1997)

Ecosystem	Average values of SOM (kg C m^{-2})	Area (ha 10^9)	Total content of C in soil (Gt)	Inventory of C in various ecosystems (%)
Tropical forest	10.4	24.5	255	17.5
Temperate zone forests	11.8	12	142	9.8
Boreal forests	14.9	12	179	12.2
Dwarf pine zone	6.9	8.5	59	4
Tropical savannah	3.7	15	56	3.8
Temperate zone meadows	19.2	9	173	11.9
Tundras	21.6	8	173	11.9
Desert bushes zones	5.6	18	101	6.9
Deserts (stone, ice)	0.1	24	3	0.2
Tillage	12.7	14	178	12.2
Swamps and marshes	68.6	2	137	9.4
Total		**147**	**1,456**	**100**

It is possible to regulate the amount and the way organic matter returns into the soil through crop rotation, thus it plays a very important role in carbon sequestration. This is not just about including such crops in the rotation process which leave more crop residues on the soil but it is also about the used crop harvesting and fertilization, the tillage system, the number of following years of perennial forage cultivation or inclusion of leguminous plants. However, it is not only the amount of post-harvest crop residues entering the soil that is important but also their elemental composition, often called the quality of post-harvest crop residues. In case nitrogen has a higher representation in the element composition of residues the C : N ratio will be narrower which results in higher decomposition intensity. Some post-harvest crop residues (grain straw, corn) are characterized by a wide C : N ratio. Considering their wide proportion in these resources its adjustment is important to avoid reduction of usable nitrogen in the soil during microbial decomposition of straw and bark. Application of nitrogen together with plowed of wheat or corn straw is important for regulation of transformation processes in the soil. In relation to the balance of soil organic matter it is also important to know the substance composition of post-harvest crop residues. For example, on one hand, the higher proportion of lignin in post-harvest crop residues, the slower decomposition but on the other hand, the higher amount of easily unstable organic matters in post-harvest crop residues, the faster decomposition.

The most important secondary source of soil organic matter is manure. **Farmyard manure** on average contains 24% of dry matter, minimum 18% of mineralizable organic substances as well as 0.4 – 0.6% of nitrogen and its quality is assessed by the content of nutrients in it. Among the other secondary sources belongs **liquid manure, green manure**. **Composts** can be organic fertilizers complements, in the production of which peat and various types of domestic and industrial waste are used. **Sludge** belongs to supplementary sources of organic matters which are rich in organic substances, macro and micro nutrients. These characteristics allow direct application of sludge into the agricultural soil. Many times sludges are contaminated with heavy metals, organic pollutants and pathogenic microorganisms. Thanks to considerable toxicity and minimal degradability of risk elements their content in sludge is the main factor that limits their application into soils.

Living organisms provide the least humus-forming matter, only 0.1 – 0.2 t ha^{-1}. The amount of **microbial biomass** is not big as well, in average it is 0.2 – 0.75 t ha^{-1}. In general microbial mass represents 2 – 5% of the total content of carbon in the soil.

4.1.3 Transformation processes of organic matter in soil

Transformation of organic residues is done in the soil with participation of microorganisms – mostly with their enzymes, oxygen, water and other factors. Residues of plants and organisms occurring in the soil or on its surface are decomposed by microorganisms and used as the source of nutrients and energy. During decomposition residues gradually lose their anatomic structure and the chemical substances they include are transformed to more mobile and simpler chemical compounds. One part of these interproducts of decomposition are mineralized by microorganisms, the other part is used by chemoorganotrophic (heterotrophic) microorganisms for synthesis of secondary proteins, saccharides, lipids – creating plasma of new microorganisms that are decomposing after death while a part of interproducts of decomposition is transformed into specific humic substances.

The most common processes for the transformation of dead residues are **mineralization and humification**. The substantial part of the organic residues mineralizes under favorable soil conditions; i.e. decomposes to H_2O, CO_2, NH_3, and simple salts. The mineralization coefficient is in the range of 0.5 – 0.85. Of course, its values depend on the soil and climatic conditions of a particular habitat.

Based on the soil humus content and climatic region (degree of mineralization), we can determine the amount of available nutrients from humus for plants. For example, in the conditions of southern Slovakia, about 1 – 1.5% of humus is mineralized annually and 100 – 120 $kg.ha^{-1}$ of inorganic nitrogen is released for plant nutrition. It is based on the total humus content in the A horizon. For example, with a humus content in the soil of 3% to the depth of 0.3 m and the soil weight of 4.5 mil kg ha^{-1} the total weight of humus is 135 t ha^{-1}. In a climatically unfavorable year, it mineralizes less than about 1%.; i.e. 1.35 t ha^{-1} per year. At a carbon content of 50% in humus and a C : N ratio of 10 : 1; 675 kg of C and 67.5 kg of nitrogen are released per hectare per year.

In the initial phases, the **mineralization** of organic residues is very fast if they contain a larger amount of more easily decomposable organic compounds. Over time, the intensity of decomposition and mineralization slows down due to the relative increase in harder-to-decompose organic compounds.

Humification is the second most widespread process of soil organic matter transformation, which takes place parallelly with the decomposition and mineralization of organic matter. It is a very complex process, including transformation of organic residues, mostly of a biochemical nature, leading to humus formation.

Biodegradation, as well as humification, cannot take place in adverse conditions where microbial activity is limited (adverse humidity, temperature, pH, nutrient deficiency). In conditions of moisture excess, **peat formation** is taking place and in case of lack of moisture, **carbonization** takes place.

The individual chemical components of organic residues have different decomposition and transformation in the soil. According to their decomposition resistance, we can compile them in the following order:

water-soluble saccharides < proteins < hemicelluloses < celluloses < lignin < tannins < fats < waxes and resins

The intensity of decomposition depends not only on the dead residues' chemical composition, the ratio between easily and hardly degradable compounds, and the C : N ratio but also on the soil environment conditions (soil composition and its physical, chemical, and technological features).

Impact of condictions on soil organic substances transformation

Brady and Weil (1999) identified as factors influencing the rate of decomposition:

1. **Environmental conditions:**
   - optimal pH – neutral,
   - optimal humidity – 60 – 80% of field water capacity (FWC), < 60% FWC – mineralization is more intensive due to aeration, > 80% – the activity of microorganisms decreases because aeration is reduced and therefore anaerobic conditions begin to prevail which cause slower decomposition,
   - optimal temperature – 25 – 35 °C – decomposition starts at t = 0 °C, at t = 50 °C CO_2 is released but it is connected with chemical decomposition,
   - microbial activity – the amount, composition, and intensive activity of microorganisms – depends mainly on environmental conditions and the composition of residues. Not only does the formation of new humic substances and the mineralization of fresh organic residues depend on microbial activity but also on the stabilization or mineralization of older humus stocks. Changes in humus content are the difference between newly formed humus and the decomposition of already formed humus substances. High biological activity causes intensive mineralization, and on the contrary weak biological activity prevents humus formation and accumulation. High clay content, moisture excess, and low temperature slow down mineralization.
2. **Quality of decomposing residues** – vegetable, microbial and animal organic residues:
   - physical state of residues: location and their availability and size:

   a) on the soil surface – slow decomposition, drying of residues, significant influence of temperature extremes, are out of reach of most soil organisms and microorganisms and mineralized elements are being lost faster from the soil surface,

b) in the soil – faster decomposition because closer contact of microorganisms and residues is ensured as well as higher moisture,
c) size – the smaller, the larger surface is available for microbial decomposition,

» chemical composition of residues:

a) C : N ratios in plant residues, microorganisms and soil,
b) ratio of labile and stable to decompose compounds.

4.2 Soil humus

Humus is a part of the organic matter of the soil. Humus is a complex, dynamic set of complex organic compounds formed during the decomposition and humification of organic matter in the soil. Many physical, physicochemical, chemical, and biological properties of soils are associated with the soil humus.

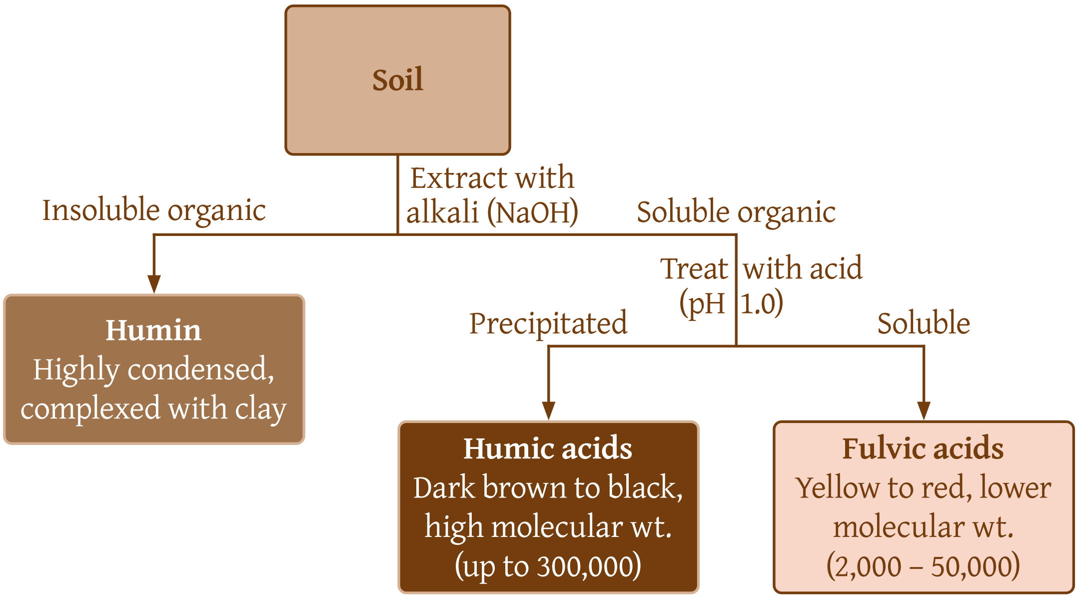


Figure 4.2: The traditional scheme for classifying so-called humic substances found to occur in highly alkaline extractions of soil (Weil and Brady, 2017)

Humic acids (HA) are amorphous matter, spherocolloids with a size of about 3 – 10 nm. The shape of the molecules is of great importance in the formation of the soil structure. It is also important in soil-forming processes that HA molecules have a porous structure which confirms their high water retention and sorption capacity. The generally accepted features of HA also include their solubility in weak alkaline solutions and the ability to precipitate with mineral acids.

Fulvic acids (FA) remain in the solution after precipitation of the humic acids (acids). The structure of the fulvic acid macromolecule is similar to humic acids with the difference that the aromatization of the core is less pronounced. Of the functional groups, carboxyl and phenolic groups predominate. Fulvic acids have a larger number of side chains as well as functional groups which determine their more hydrophilic character and helps fulvic acids to have better solubility, greater mobility, and aggressiveness towards soil minerals.

4.2.1 Functions of humus in soil

Humic substances and their complexes have a versatile effect on soil fertility. They have a significant effect on changes in the mineralogical, chemical, and also partly soil texture. Soil profile formation is closely connected with humus content and its composition. Humus (A) horizons are formed in favorable conditions of humification. Eluvialization and iluvialization are related to the activity of fulvic acids and low molecular compounds on the decomposition and migration of mineral components (processes of illimerization, podsolization, internal weathering, and transformation).

Humus in the soil has the following functions (Zaujec et al., 2009):

- participates in the creation and formation of physical and technological properties,
- participation in the biogenic elements cycle,
- providing energy and carbon for soil microorganisms,
- participation in ion exchange processes,
- impact on solubility and migration of elements in the natural environment,
- regulation of soil buffering capacity,
- the positive impact of a number of biologically active substances,
- regulation of soil redox potential,
- binding of pesticides and heavy metals, the biodegradability of pesticides,
- inhibiting development of certain plant pathogens,
- creation of structural aggregates,
- has a positive effect on the physical properties of soils.

4.2.2 Soil humus indicators

With intensive use of soils, the use of organic and mineral fertilizers the whole soil profile and their humus composition change. The quality of humus is most often assessed according to the carbon content of humic acids (C_{HA}) and fulvic acids (C_{FA}), or the C : N ratio.

We consider a **ratio between $C_{HA} : C_{FA}$** in humus higher than 1 to be favorable, a ratio lower than 1 to be of lower quality. The average C : N ratio in humic substances is 10 : 1. The narrower ratio is an expression of the higher quality humus and wider low quality humus.

Recently, the quality of humus in the soil is evaluated on the basis of labile components such as the content of labile carbon and nitrogen, respectively their mutual relations. It is on the basis of these indicators that we can assess changes in humus quality from the short-term point of view.

4.3 Influence of tillage on soil organic matter and humus

The aerobic conditions created by tillage support oxidative processes, and the development of microorganisms provide mineralization and humification. The relationship between mineralization and humification during regular cultivation depends on the amount of post-harvest residues incorporated into the soil, root exudates, and organic fertilizers. In a mellow, optimally aerated topsoil, favorable conditions are created for the mineralization and, through the development of plant roots and microorganisms also for the humification of organic residues and fertilizers. Most of the works testify to a favorable humus-forming process in conditions of optimal aeration. Reduced and increased oxygen content in the environment limits the formation and accumulation of humus and changes its quality composition.

Agricultural practice and long-term experiments with different agricultural techniques in different soil-climatic conditions prove that tillage without replenishing organic matter reserves leads to a significant reduction in humus content. There are great losses, especially in the first years of soil cultivation (20 – 50 years) while the humus content reaches the level of the reserves of the most stable humus component. In subsequent periods the humus supply does not change significantly and essentially depends directly on the amount of organic residues advancing into the soil and on the intensity of organic fertilization. For example, under the long-term experiments of the Timiriazev Agricultural Academy, the original humus content decreased by as much as 43% during the first 13 years of tillage and gradually by only 9% over the next 30 years. The decomposition of humus due to tillage in Chernozem is significantly slower and more even. In the conditions of Ukraine, in their Chernozem, over 100 years of tillage of these soils, the humus content decreased from 5.4% to 3.98%, while the rate of decomposition varied from period to period. In the first five years after plowing, the loss of humus was the highest and represented 14.6% of the original reserves.

In the following period, humus losses decreased and after 20 years they became even. A dynamic balance has been created between the creation of new humus and the mineralization of the old reserves. Rothamsted's classical field experiments also show a high rate of decomposition of the original humus reserves after plowing the ground and regular tillage. The humus regime of long-term cultivated soils depends on the original reserves and age of the organic matter of the soil, the amount of microorganisms, and the intensity of decomposition of plant residues.

The issue of basic agricultural technology influence, method, and the depth of tillage on the dynamics of humus is also given considerable attention in our country. Recently, special attention has been paid to comparing the classical and minimizing tillage systems on biological activity and the state of humus reserves. It is generally stated that deep plowing increases the biological activity of soils and promotes humification compared to shallow ones. However, the analysis of the state of the individual plow layer depths shows that after turning the comb to the bottom of the furrow, an improvement occurs in the lower part of the plow layer, while the bottom that has reached the surface is characterized by a lower humus content and weaker biological activity. The balance of the humus regime and biological activity in the whole plow layer usually occurs only after 3-4 years of thorough loosening and mixing of the soil mass. Non-traditional methods of soil preparation combined with minimization of mechanical interventions affect the whole complex of physical properties and the distribution of post-harvest residues.

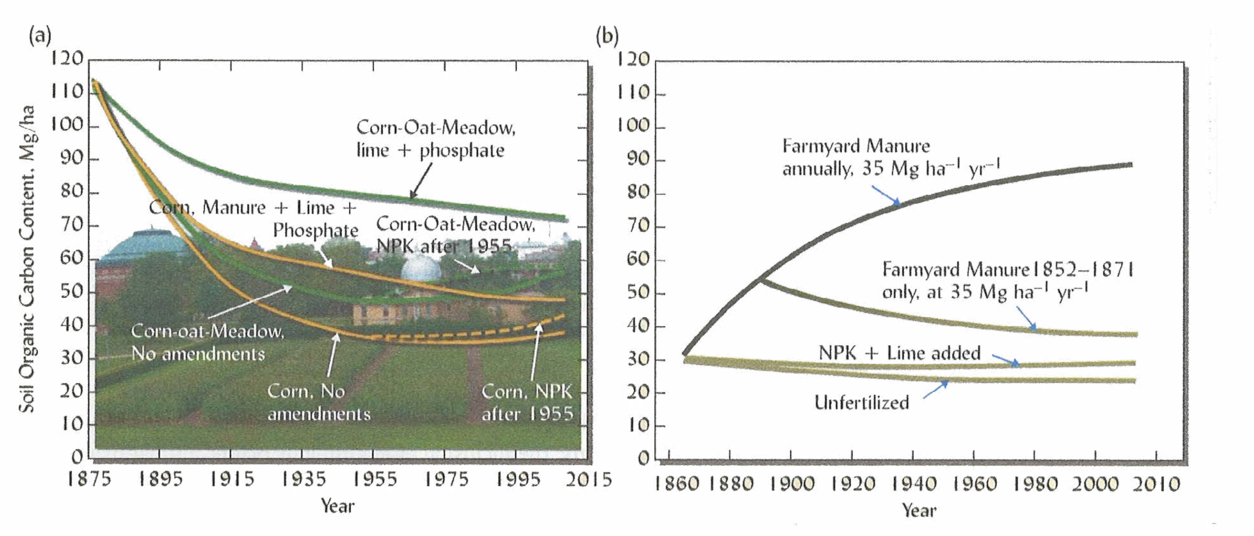
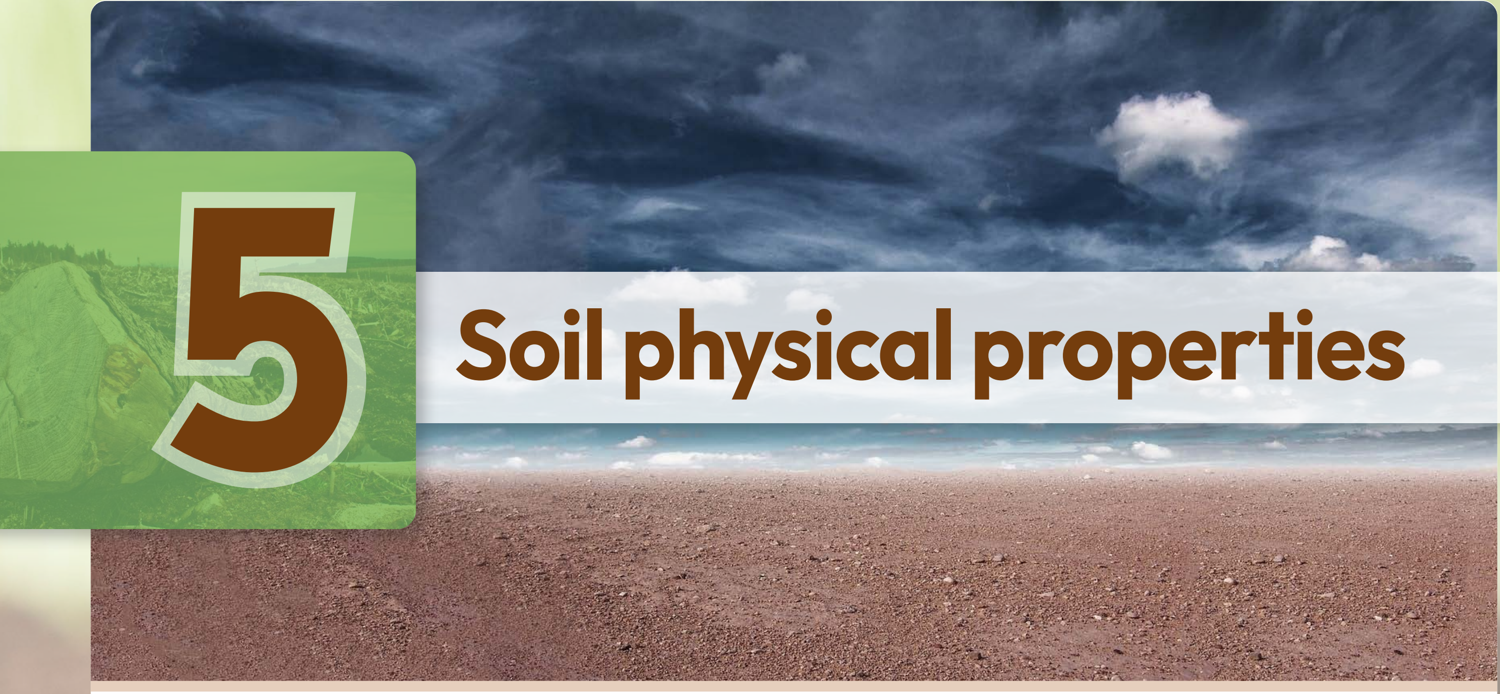


Figure 4.3: Soil organic carbon contents of selected treatments of (a) the Morrow plots at the University of Illinois, USA, and (b) the classical Hoosfieled experiment at Rothamstead Experimental Station in England (Weil and Brady, 2017)

The Morrow plots were established on virgin grassland soil in 1876 and so suffered rapid loss of organic carbon in the early years of experiment. The Rothamstead plots were estabilished on soil with a long history of previous cultivation. As a result, the soil at Rothamstead had reached an equilibrium level of organic carbon characteristic of the unfertilized small-grains (barley and wheat) cropping system traditionally practiced in the area. Note that large impact of farmyard manure application and the residual effect of manure some 135 years after the last application was made.



5

Soil physical properties

Soil physics is a relatively young branch of pedology that has been developing significantly since the middle of the last century. It deals with the study, measurement, and prediction of changes in the physical characteristics of soils (texture, soil structure, water retention, etc.) and processes (aeration, diffusion, etc.) in natural as well as man-influenced conditions. Soil physics involves the application of the principles of physics to understand the relationships between matter and energy.

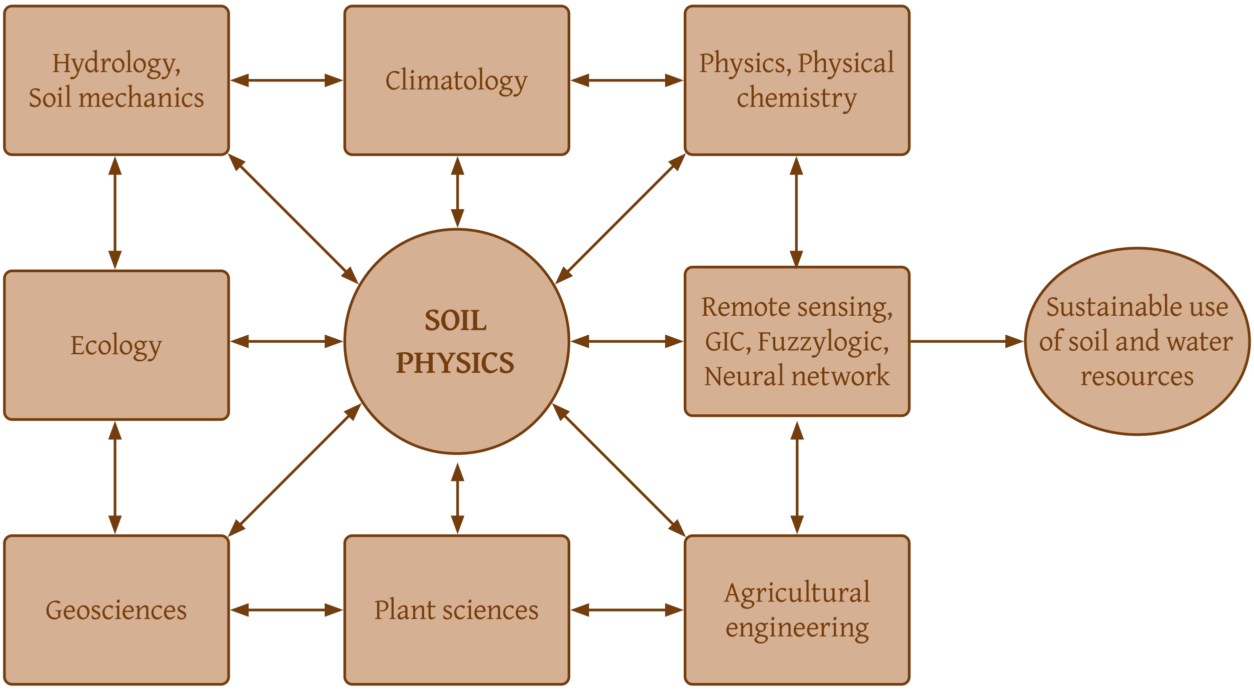


Figure 5.1: Interaction of soil physics with basic and applied sciences (Shukla, 2014)

Physical edaphology is actually a branch of soil physics and is therefore a science dealing with the application of soil physics into the agricultural environment. The subject of its study is monitoring the whole complex of physical properties and their relations to plant growth. Soil physics also significantly affects other scientific disciplines, such as ecology, geology, physical chemistry, and others (Fig. 5.1 on the previous page).

Soil physics plays an important role in:

1. Resource maintenance,
2. Improving resources and using them more efficiently,
3. Minimizing the risks associated with soil erosion and degradation, restoration and regeneration of degraded soils and ecosystems,
4. Increase of production by reducing soil-climate differences by finding and designing the most appropriate way of farming in a particular area.

In the context of environmental quality, the soil acts as a reservoir and filter of pollutants that are transforming, deactivating, denaturing or detoxifying, or retaining. Soil physical properties and processes in the soil therefore play the crucial role (Fig. 5.2).

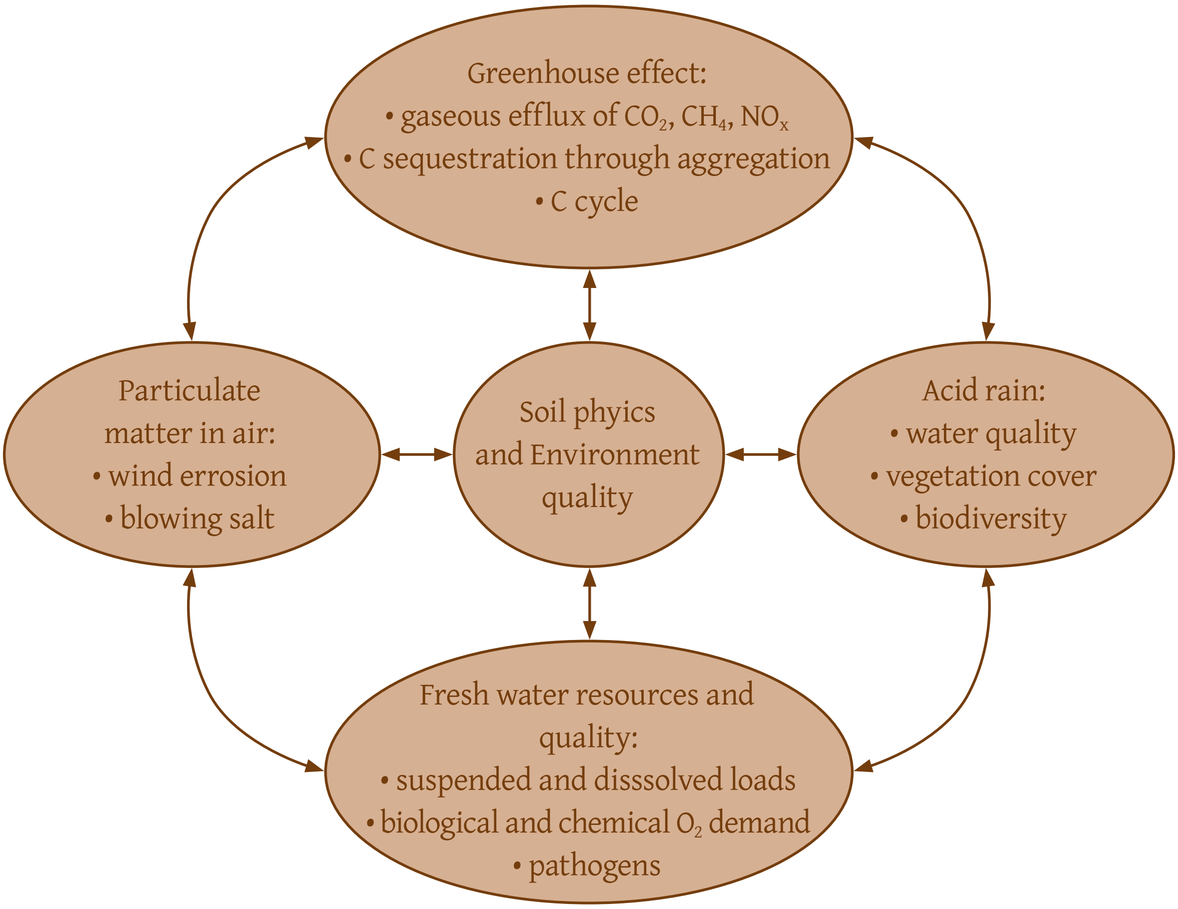


Figure 5.2: Applications of soil physics to environmental quality (Shukla, 2014)

Soil physics deals with interrelationships and changes in soil components and soil phases. Before we begin to describe the specific physical features for a better understanding, let us recall that soil consists of four components and three phases. Soil components are made up of mineral (inorganic) and organic matter, water and gases. The inorganic matter consists of primary and secondary minerals derived from the parent material. The organic matter comes from living and dead parts of plants and animals. The liquid component consists of dilute water solutions of inorganic and organic components. The gaseous component, including soil air, is a mixture of main gases (nitrogen and oxygen) and trace gases (CO_2, methane, nitrogen oxides). All four components of the soil are constantly changing in their relative mass, volume, space and energy states due to natural and anthropogenic factors. Thus, if we start from the spatial arrangement of the soil mass and take into account only the state of the substances that make up the soil, we can say that the volume of the soil sample intact represents a system of three phases, which we call as "V_{sm}". In this system, we must differentiate between the volume of the solid phase "V_s" with the mass ms and the total pore volume "V_p". The volume of the solid phase consists of the volume of the organic component "V_o" with the weight "m_o" and the volume of the mineral component "V_{in}" with the corresponding weight "m_{in}". A certain volume of water "V_w" with a weight of m_w and a volume of air "V_A" with a weight of m_A are represented in the total pore volume. It follows from the above statement that each of the mentioned phases occupies only a certain part of the total examined soil volume "V_{sm}" and it has a certain weight (Fig. 5.3).

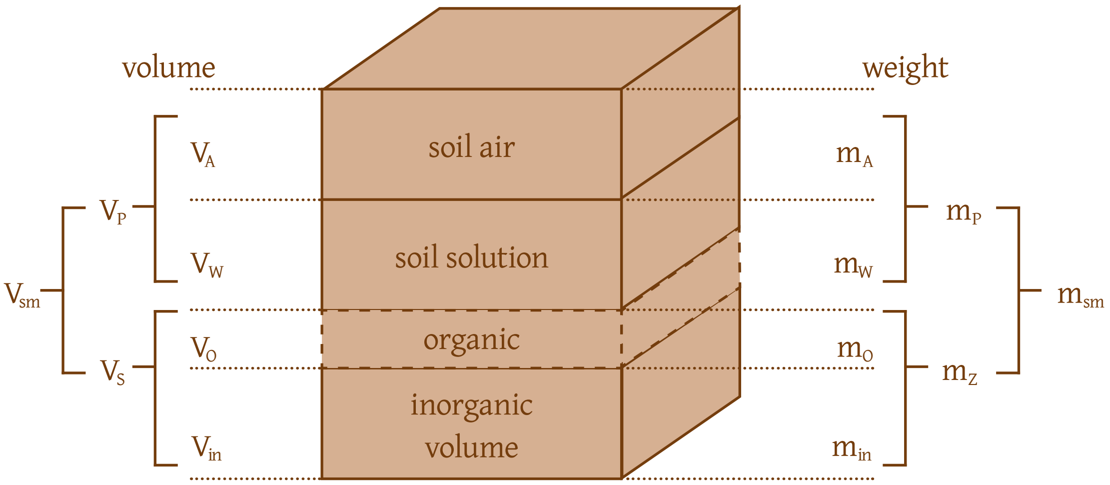


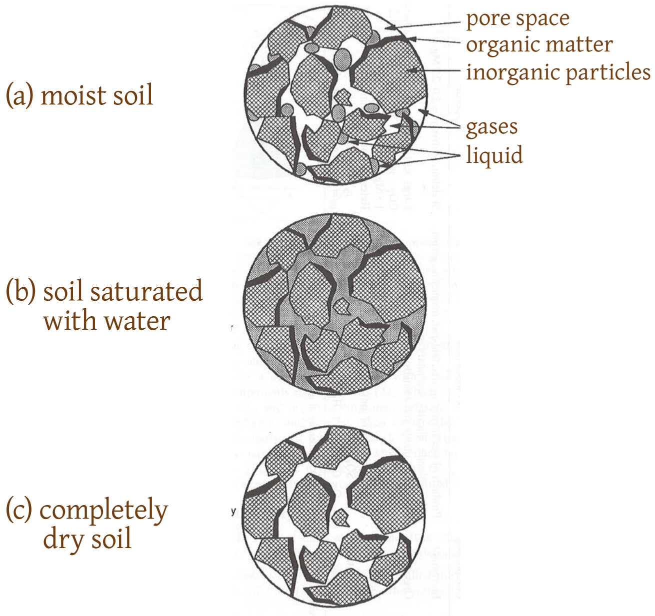
Figure 5.3: Relationships between soil components (Hanes et al., 1995)

Then the following applies:

$$V_{sm} = V_o + V_{in} + V_w + V_A = V_S + V_p$$
$$m_{sm} = m_o + m_{in} + m_w + m_A = m_s + m_p$$

Optimal conditions for the growth of the most cultivated plants are considered to be 50% of the solid phase of the total amount (organic and inorganic part), while the liquid and gaseous phases account for 25% each (Fig. 5.4A). However, we also know the crops such as rice, or those that require higher water or air content. In such cases, the representation of individual soil phases is different for their optimal growth and development. From the solid phase of the soil, it represents an organic content of about 1 – 5%. Immediately after rain or watering the soil, all pores are filled with water and the soil is saturated (Fig. 5.4B). When the soil dries out, the water from the pores evaporates into the atmosphere and air or gases remain in the pores (Fig. 5.4C).

Thus, the concept of physical properties of soil is a set of soil properties that are conditioned by the grain composition of soil particles, the spatial arrangement of the soil mass, and the relationships between solid particles, soil water, and air.



(a) moist soil

(b) soil saturated with water

(c) completely dry soil

Figure 5.4: Interaction among four components and tree phases for (a) moist, (b) water-saturated, and (c) completely dry soil (Lal and Shukla, 2004)

The amount and size of soil particles and their mutual arrangement are of great importance in the formation of the pores, their dimensions, and their method of distribution, in the infiltration and permeability of the liquid phase and air. Thermal properties are conditioned by the relationship of soil to external factors (heat, light, and wind). The physical properties of the soil significantly affect the course and speed of various physic-chemical and biological processes, as well as the development of soil microorganisms. Soil fertility also depends on a large extent on the state of physical properties, especially when providing plants with accessible nutrients, rainwater or irrigation water, heat, and air.

Table 5.1: Basic properties of individual components and phases of soil (Lal and Shukla, 2004)

Phase of soil	Components	Composition	Properties
Solid	mineral	weathering products, quartz, feldspar, magnetite, garnet, amphibole, silicates, secondary minerals	skeleton, matrix, $\rho s = 2.0 - 2.8\ t\ m^{-3}$
	organic	plant and animal remains, living organisms	large active surface, high activity, release of CO_2 into the atmosphere, $\rho s = 1.2 - 1.5\ t\ m^{-3}$
Liquid	soil solution	liquid ion solutions (e.g. Na, K, Ca, Mg, Cl, NO_3, PO_4, SO_4)	heterogeneity, dynamics, $\rho w = 1\ t\ m^{-3}$
Gaseous	soil air	N_2, O_2, CO_2, CH_4, C_2H_6, H_2S, N_2O, NO	$\rho a = 1 - 1.5\ t\ m^{-3}$, variability, dynamics

According to the functionality they play in the soil, we divide the physical properties into two groups:

1. Basic

2. Functional

The basic physical properties (primary) are closely connected with the qualitative properties of the soil mass and we include here the soil texture (particle-size distribution), soil structure, specific weight, bulk density, porosity, and soil color.

Functional physical properties (secondary) depend on the basics and are the result of the function of the soil as an environment inhabited by plants and animals. We include here all types of soil regimes, i.e. water, air, and heat, but also technological properties (physical-mechanical states).

The state of the physical features of soils is influenced by natural and anthropogenic factors. Among the natural ones, we include, for example, high content of clay particles, ilimerization, heavy rains, the pressure of the upper layers of the soil to the lower, and lack of organic matter and humus in the soil. The anthropogenic factors can be **biological, chemical, and mechanical**.

Biological – insufficient supply of organic matter to the soil, unsuitable crop structure, monocultural plant growing.

Chemical – unsuitable composition of industrial fertilizers containing monovalent cations conditioning colloid peptization (disruption of soil structural state), insufficient liming.

Mechanization – high specific pressure of machines on the ground, unsuitable composition and type of tools, unnecessary passes in the field, wheeled tractors driving in the channels, machining in unsuitable humidity, insufficient depth of tillage.

Maintaining favourable physical characteristics is of great importance for sustainable agriculture, as there is a constant decline in the area of agricultural land worldwide, which is also directly connected with population growth. It is estimated that from the current approximately 7 billion, the world's population will increase to 8 billion, and by 2050 it will be 9.4 billion. According to Lala (2000), there was 0.5 hectare of arable land per capita worldwide in 1950, 0.2 hectare in 2000 and it is further estimated that in 2050 it will be only 0.14 hectare and even in 2100 only 0.1 hectare. The total and permanent loss of soil must therefore be solved or compensated for through the improvement of its physical properties (e.g. subsoil disturbance, which ensures faster infiltration of rainwater and irrigation water into the soil and overall increase of water in the soil environment).

Many workplaces will therefore have to deal more intensively with issues such as:

- improving soil structure,
- more efficient use of nutrients through an integrated nutrient management system and strengthening of nutrient cycling mechanisms in the environment,
- maintaining and improving the quality of soil and water environment through soil protection technologies,
- more efficient use of water through its better retention in the soil,
- increasing production of individual crops.

The impact of individual physical features and processes on agriculture is summarized in Table 5.2.

Table 5.2: Physical features and processes and their impact on agriculture (Lal and Shukla, 2004)

Processes	Physical features
Compaction	Bulk density, porosity, grain size, soil structure
Erosion	Soil structure stability, particle size, infiltration, hydraulic conductivity
Water movement	Hydraulic conductivity, pore size
Aeration	Porosity, pore size, soil structure, concentration gradient, diffusion coefficient
Heat movement	Thermal conductivity, current water content

5.1 Soil texture

The speed and volume of many processes taking place in the soil are influenced by its soil texture (particle-size distribution). Therefore, the study of physical properties should begin with its determination. Soil texture actually expresses only the content of mechanical particles of mineral origin that have a variety of shapes and sizes, but also different mineralogical and chemical compositions.

The soil texture is actually the representation of individual grain fractions (sand, dust, and clay) in the soil sample in weight percent and serves to determine the basic soil property – soil class.

The mineral component of the solid phase of the soil is thus formed by particles of different sizes. Mineral soil particles of a defined size are divided into groups that we denote by grain fractions, and which, in addition to close dimensions, also have more or less identical some basic physical and physic-chemical properties. For example, the particle size range of 2 – 0.05 mm in terms of soil granular composition is professionally called sand, the category 0.05 – 0.002 mm silt, and the category <0.002 mm clay. As we have already mentioned, their size plays the most important role in the distribution of basic soil particles. Sand, silt, and clay particles are present in the soil in its mineral component. Some soils also contain gravel, stones, or solid neoplasms.

Soil texture is an important physical property of the soil that greatly affects formation and development of the soil and a range of soil properties. When the soil is formed, it affects the graininess of the substrate, its weather ability and the formation of the clay fraction. It decisively affects the physical, chemical and physic-chemical properties of the soil itself, its aggregation capacity, water and air permeability. It affects the overall porosity, pore size and the surface area. It affects adhesion, plasticity, plowing resistance, plant roots penetration, or fruit trees, etc.

Table 5.3: Selected characteristics of USDA granular fractions (Millar, 1962; Foth, 1990)

Fraction	Diameter (mm)	Number of particles per gram	Specific surface area in cm^2 per 1 g
Very coarse sand	2.00 – 1.00	90	11
Coarse sand	1.00 – 0.50	720	23
Medium sand	0.50 – 0.25	5 770	45
Fine sand	0.25 – 0.10	46 000	91
Very fine sand	0.10 – 0.05	722 000	227
Silt	0.05 – 0.002	5 776 000	454
Clay	<0.002	90 260 853 000	8 000 000

5.1.1 Particle-size distribution and their characteristics

The result of physical and chemical weathering of the parental rock is a wide range of mineral particles of different sizes. Based on the size of elementary soil particles, we divide and distinguish individual grain fractions which according to the size can be divided into four basic groups, namely: **gravel (skeleton), sand, silt, and clay**. The determination of the soil granular composition is based on the collected disturbed sample directly in the field which after drying is referred to as coarse soil which is divided into fine soil and skeleton. The size of 2 mm for the separation of the fine soil from the skeleton was chosen because it is the upper limit of the water capillary movement.

Skeleton

Soil particles larger than 2 mm are called skeleton. Due to its size, the skeleton does not bind any substances, does not create capillary pores, does not allow capillary movement of water, and has no direct share in the ongoing chemical processes and their dynamics in the soil. The shape of the skeleton is sharp-edged and primarily indicates possible deluvial deposits. Soils with a skeleton content > 10% by weight are skeletal.

Sand

The sand fraction is characterized by a particle range from 2 to 0.05 mm. Sand forms the coarse fraction of the soil and forms the basis of the soil profile. According to the American Classification System (USAD), sand can be divided into coarse, medium, and fine sand. Sand grains are mostly made of quartz, but they also contain fragments of feldspar, and mica but also trace amounts of heavy minerals such as zircon, tourmaline, and amphibole. The sand particles are thus formed of primary minerals and are solid and free of pores. The shape of the sand particles is sharp-edged, they are hard (hardness from 5 – 7 on the Mhos scale) and, with a high proportion in the soil, they participate in grinding the metal parts of machines. The sand particles are comparatively larger and thus have a smaller specific surface area compared to the same weight of dust and clay. Due to its small specific surface area, sand apart from $CaCO_3$ and $MgCO_3$ particles and some soluble components does not have any importance on the chemical and physical properties of soils. Small fractions of clay may be retained on the surface of the smaller sand fractions, thus increasing its activity which, however, is not caused by the sand but by the fact that the clay is currently present. The sand particles are inactive and lack the ability to join into larger aggregates. Sand increases the size of the spaces in the soil between the particles, thus helping to exchange the air and drain water. It can be added to soil to reduce the bulk density of heavy soils.

Table 5.4: Area of skeletal agricultural lands in Slovakia (Zaujec et al., 2009)

Soil skeleton	Skeleton (%)	Area (ha)	% of agricultural soil fund in Slovakia
Weakly skeletal	10 – 25	214 000	8.8
Medium skeletal	25 – 50	295 000	12.0
Strongly skeletal	over 50	214 000	8.7
Total		723 000	29.5

Silt

Silt represents one of the basic grain fractions in the soil. Its particles have the size of 0.05 to 0.002 mm.

Table 5.5: Representation of individual minerals in sand and silt (Zaujec et al. 2002)

Mineral	Degree of weatherability
Quartz	the most durable
Muscovite	↓
Orthoclass	
Biotite	
Albite	
Amphibole	
Augit	
Anorthite	
Olivine	the least durable

It consists of primary minerals and its particles are solid and non-porous. Larger silt particles (coarse silt) have similar properties as sand, and smaller silt particles (fine silt) approach the properties of clay. The mineralogical composition of the silt is similar to sand but the surface size is larger. Primary minerals are represented only in the sand and silt fraction. The coarser fractions of silt particles are very similar to fine sand in terms of specific surface area and therefore have very little effect on the chemical processes in the soil. On the other hand, finer silt has a sufficient active surface to allow a slight chemical activity. The silt particles, therefore, have a low sorption and retention capacity, so they are able to retain various compounds and water. Silt particles also have a tendency to stick together or stick to other particles and thus have an effect on aggregation and soil structure formation.

Clay

The particle size of clay is defined by dimensions smaller than 0.002 mm. Clay is the most reactive fraction of the soil – it has an electric charge on its surface. Due to its small overall size but on the other hand the high specific surface area of its particles, it is highly reactive which is also reflected in the high dynamics of volume changes and influencing a number of soil properties. The shape of the clay is plate-shaped or needle-like shaped. From the mineralogical point of view, clay particles include groups of minerals called aluminosilicates. They are secondary clay minerals containing fine particles of iron oxides (Fe_2O_3), aluminum oxides (Al_2O_3), carbonates ($CaCO_3$), and other salts. The properties of the clay fraction that significantly affect soil behavior are listed in Table 6.6 and e.g. include light hydration due to its high affinity to water, high swelling/shrinkage capacity due to the natural expansion of the clay lattice, high plasticity due to its ability to maintain shape, tack when wet and cracked due to shrinkage and sintering, high density of negative charge leading to the formation of the electrostatic bilayer, etc.

Table 5.6: Basic properties of basic grain fractions (Lal and Shukla, 2004, ∗ – in Fulajtár, 2006)

Properties	Particle size classes		
	sand	silt	clay
Size	2 – 0.02 mm 2 – 0-05 mm*	0.02 – 0.002 mm 0.05-0.002 mm*	<0.002 mm <0.002 but also <0.001 mm*
Shape	sharp-edged	slightly irregular	plate – like, needle-like
Tactile test	scratches, coarse-grained	smooth, floury	sticky
Plasticity	non-plastic	slightly plastic	plastic
Cohesion	incoherent	slightly coherent	coherent, gelatinous
Specific surface area	very low	medium	very high
Mineralogical origin	primary	primary	secondary clay
Secondary particles	none	several	in form of aggregates
Water retention capacity	none	medium	high
Hardness	5-5-7	5-5-7	-
Sorption capacity	missing	very low	high – very high

Clay, especially **colloidal clay**, is formed by the weathering of primary minerals or their synthesis. Chemical analyses of clay minerals have shown that they consist of a large extent of silicon, aluminum, and water with varying amounts of iron, calcium, magnesium, natrium, and potassium, and minor amounts of other elements. Clay and its stratification in the soil profile are used to assess the geological stratification of the soil profile and to assess the pedogenetic development of the soil. Significant changes in the clay content in the soil profile indicate a textural stratification of the soil profile. The pedogenetic stratification of clay has several forms. It allows you to assess the age and development of the soil and the so-called textural differentiation of two horizons caused by colloid translocation in the soil profile. **Textural differentiation of the soil profile** – manifests itself in the loss of clay in the impoverished, surface horizon and its increase in the following – enriched horizon. The degree of soil profile differentiation is therefore assessed by the clay content ratio between the two horizons. If the clay content ratio between the two horizons is in the range of 1.2 to 1.5 we speak of weak texture differentiation. If it is in the range from 1.5 – 2.0, it is a medium, and in> 2.0 a strong texture differentiation.

Colloid-dispersive particles

The size of these particles is <0.0001 mm and they occur in the colloidal clay fraction. They are represented by clay minerals, humic acids, sesquioxides and colloidal silicic acid. In addition to the outer surface, they also have an inner surface. Due to the high active surface, they are crucial factors in the high sorption capacity of the soil. The issue of soil colloids is elaborated in more details in Chapter 6 Soil chemistry.

Particle shape

The shape of soil particles varies and often depends on their size, parent rock and degree of weatherability. Coarse or large particles (sand, silt) are often square and sharp-edged. The angularity is reflected in the degree of weatherability. Very sharp-edged particles are poorly weathered and become rounder as the weathering process progresses. They are gradually abraded with water and wind. The clay particles have a plate-like or needle-like shape. The shape of the particles is determined by micrography, and is expressed by two indexes: roundness and sphericalness.

Specific surface of particles

The amount of soil properties (cation exchange capacity, retention, dynamics of chemical processes, plasticity, cohesion, strength, etc.) depends on the specific surface area of the individual particles (a). Therefore, knowledge of the specific surface area is very important for agriculture as well as for the industry. The specific surface area of particles can be expressed in three ways:

1. per unit weight (am),
2. the specific weight of the soil (av),
3. on the bulk density of the soil (ab).

$$1.\ am = As/Mz\ (\mathrm{m^2\ g^{-1}})$$
$$2.\ av = As/Vz\ (\mathrm{m^2\ m^{-3}})$$
$$3.\ ab = As/Vs\ (\mathrm{m^2\ m^{-3}})$$

As – total particle surface
Mz – soil weight
Vz – soil solids volume
Vs – total volume

The size of the active surface of the particles depends on their size and the shape of the particles themselves. The surface area increases logarithmically with decreasing particle size. Plate-shaped and needle-like shaped particles have a larger specific surface area than sharp-edged and spherical particles.

Particle surfaces can be calculated in the following ways:

Cubic particles

L-sized particles have a specific surface area of $6L^2$, a volume of L^3 and a weight of ρsL^3. Therefore, we calculate the specific surface area of the cubic particles as follows:

$$am = 6L^2/\rho sL^3 = 6/\rho sL$$
$$av = 6L^2/L^3 = 6/L$$

Spherical particles

The specific surface area of spherical particles is similar to cubic-shaped particles. For spherical particles of diameter D and specific gravity ρs we need to know: the total volume which is $\pi D^3/6$, the weight is $\pi D^3.\rho s/6$ and the total surface area is πD^2.

$$a_m = \pi D^2 / = \frac{(\pi D^3.\rho_s)}{6} = \frac{6}{(D.\rho_s)}$$

$$a_v = \pi D^2 / \frac{(\pi D^3)}{6} = 6/D$$

Plate-like shaped particles of equal length and width (L = b)

Clay particles have a plate-like shape and a large specific surface area in comparison to sand and dust. We can calculate it from the following relationship:

$$a_v = \frac{(2L^2 + 4ld)}{L^2 d} = \frac{2(L + 2d)}{Ld} = \mathrm{cm^2/cm^3}$$

d – particle thickness

we assume that d is negligible compared to l:

$$av = 2/d$$

Plate-like shaped particles with uneven length (L1 and L2) and thickness d

The total volume of the rectangular plate is $l1l2d$, the weight in the volume of $l_1l_2d\rho_s$ and the total specific surface area is $2(l_1l_2+dl_1+dl_2)$.

$$a_m = \frac{2(L_1L_2 + dl_1 + dl_2)}{\rho_s L_1 L_2 d} = \frac{2}{\rho_s d}(1 + \frac{d}{L_1} + \frac{d}{L_2})$$

$$a_v = \frac{2(L_1L_2 + dl_1 + dl_2)}{L_1 L_2 d} = \frac{2}{d}(1 + \frac{d}{L_1} + \frac{d}{L_2})$$

It is true that the smaller the particles in the soil, the larger the total specific surface area. This is demonstrated by the following example: particles with a known specific gravity (2 500 kg m^{-3}), in a spherical shape, with a diameter of 2 mm have a total active surface area of only 1.2 m^2 kg^{-1} while spherical particles with a diameter of 2 μm have a total the size of the active surface several times higher (1 200 m^2 kg^{-1}).

5.1.2 Ways of soil separation into textural classes

Before dividing the soil into individual grain fractions, we must ensure that each particle is separated. The amount of particles in the soil is not freely separated but in the form of certain clumps. By combining elementary particles with various sealants (organic matter, carbonates, aluminum, and iron oxides), soil aggregates are formed. Organic matter is one of the most important linking element that “holds” particles together. Thus, the first problem is the destruction and disruption of cementing (bonding) bonds. The most common and accepted procedure is the oxidation of organic matter with hydrogen peroxide. If the aggregates are cemented with carbonates, we must remove the carbonates from the soil (most often with HCl). The same is true for aluminum or iron oxides which also bind and hold the particles in the aggregates. Simply, if we want to redistribute the soil into individual grain fractions, we have to break the cementing bonds that hold the particles in the aggregates, so we have to carry out the dispersion (decomposition of the aggregates into basic particles). This will ensure that the individual grain fractions are made up only of mineral particles and the neoplasms (iron carbonates and aluminum oxides) will not affect the soil granularity. A variety of methods are used to separate soil particles into granular fractions. The most commonly used method by which we can reliably classify the soil into individual grain fractions is sedimentation. They are based on the rate of particle fall in water which depends on the particle size and the properties of the liquid.

The rate of particle fall is calculated as follows:

$$v = \frac{2}{9} g.r^2 \frac{\rho_s - \rho_l}{\eta}$$

To determine the sedimentation time “T”, in which the individual particles with radius “r” pass the sedimentation path “h” are fitted to the previous formula for v = h / T, then

$$t = \frac{9.h.\eta}{2r^2.g(\rho_s - \rho_l)}$$

$$r = \sqrt{\frac{9.h.\eta}{2.t.g.(\rho_s - \rho_l)}}$$

v – particle fall rate / sedimentation rate (m s^{-1})
g – gravitational acceleration (9.81 m s^{-2})
r – particle radius (m)
ρ_s – specific weight of particles (kg m^{-3})
ρ_l – specific weight of liquid (kg m^{-3})
η – dynamic viscosity of liquid (N s m^{-2}) for water at 20 °C (0.001004 N s m^{-2})
T – particle fall time (sedimentation time) (s)
h – particle trajectory (sedimentation trajectory) (m)
θ – final speed (m s-1)

Table 5.7: Methods for determination particle-size distribution (Lal and Shukla, 2004)

Fraction methods	Particle range
Sieving	100 – 0.05
Sedimentation	2 – < 0.002
Optical microscopy	1 – 0.001
Gravitational sedimentation	0.1 – 0.0005
Permeability	0.1 – 0.0001
Gas absorption	0.1 – 0.0001
Electron microscopy	0.005 – 0.00001
Washing	0.05 – 0.005
Centrifugal sedimentation	0.01 – 0.00005
Nephelometry	0.005 – 0.00005

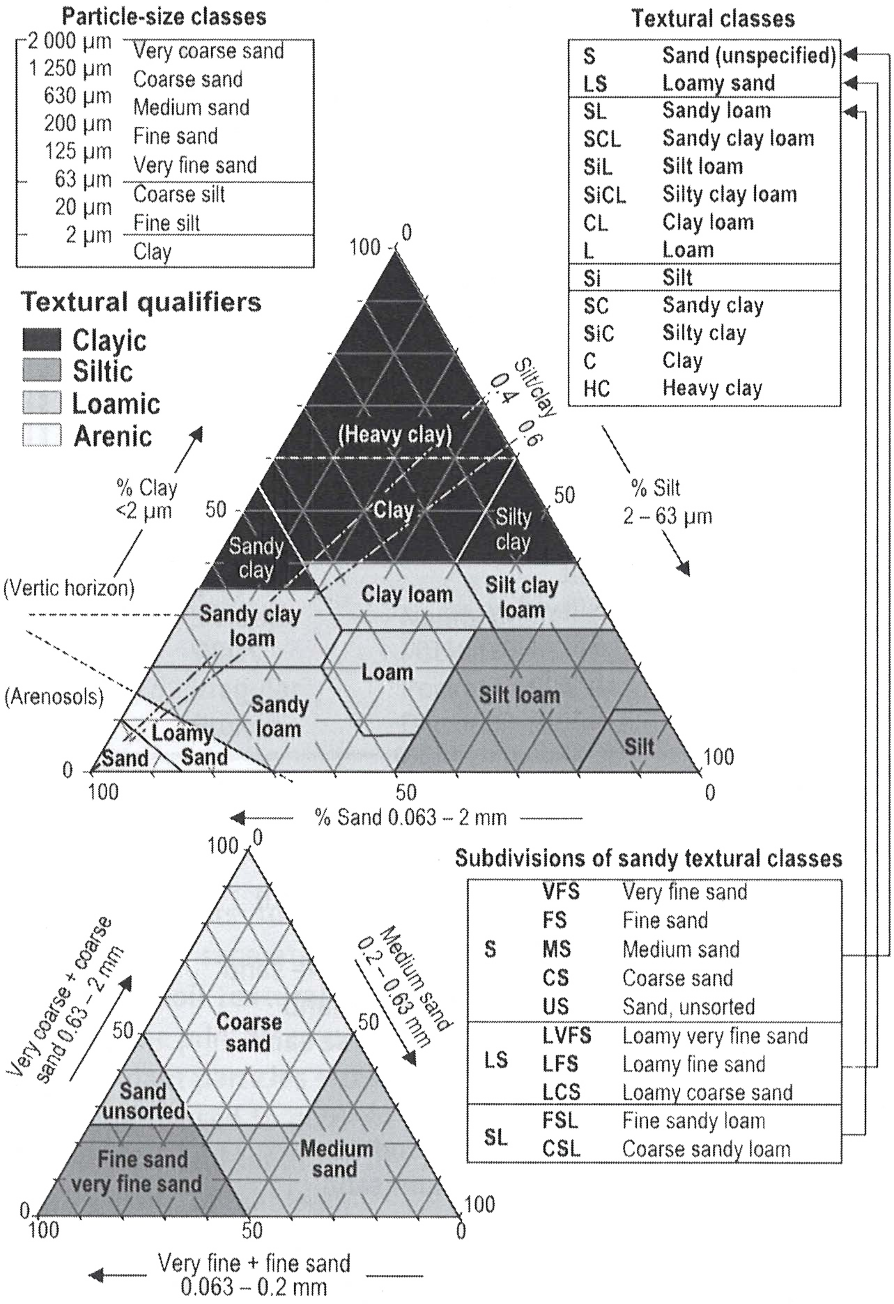
Classification of soils according to grain size

The results of soil texture are important in terms of:

1. Genetic – the results of grain analysis in the soil profile are the indicator of translocation of colloids in the profile. The content of the fraction < 0.002 mm and the ratio of its contents in the eluvial (depleted) and illuvial (enriched) horizon are evaluated and on the basis of this so-called textural differentiation of soil horizons. The results obtained also indicate the homogeneity or heterogeneity of the substrate on which the soil was formed.

2. Agronomic – based on the grain analysis various amelioration measures can be implemented and realized, such as drainage, irrigation, amelioration of sandy soils, oppression of light soils, and lightening of heavy soils. The results of the grain size analysis are of key importance for plant fertilization and liming but also for the tillage itself, i.e. determination of complexity and difficulty of machining.

Knowledge of soil texture, as already mentioned, is of key importance in terms of plant fertilization. Therefore, if we know the results of soil texture analysis and mineral fertilizer (forms of individual nutrients), we know when we will be able to apply it to the soil so that it can be used as efficiently as the currently growing plants. For example, nitrogenous fertilizers with nitrate nitrogen (NO_3^-) are characterized by the fact that they are not retained by the sorption complex of the soil.



Particle-size classes

2 000 µm	Very coarse sand
1 250 µm	Coarse sand
630 µm	Medium sand
200 µm	Fine sand
125 µm	Very fine sand
63 µm	Coarse silt
20 µm	Fine silt
2 µm	Clay

Textural classes

S	Sand (unspecified)
LS	Loamy sand
SL	Sandy loam
SCL	Sandy clay loam
SiL	Silt loam
SiCL	Silty clay loam
CL	Clay loam
L	Loam
Si	Silt
SC	Sandy clay
SiC	Silty clay
C	Clay
HC	Heavy clay

Subdivisions of sandy textural classes

S	VFS	Very fine sand
	FS	Fine sand
	MS	Medium sand
	CS	Coarse sand
	US	Sand, unsorted
LS	LVFS	Loamy very fine sand
	LFS	Loamy fine sand
	LCS	Loamy coarse sand
SL	FSL	Fine sandy loam
	CSL	Coarse sandy loam

Figure 5.5: Relation of constituents of fine earth by size, defining textural classes and subclasses (Świtoniak, et al., 2018)

A fertilizer with such a nitrogen content is dissolved in the soil solution and is available to plants or to microorganisms inhabiting the soil environment. Unaccepted nitrogen is leached from the soil. It is relatively better with a nitrogen fertilizer containing ammonia nitrogen (NH_4^+) because it is able to remain on the sorption complex of soils. It is clear from the above that the application of nitrogen fertilizers will be more suitable in spring, even in divided doses (mainly NO_3^- forms) for its more efficient use. The situation with phosphorous fertilizers in terms of soil texture is different. This is related to the relatively low migration ability of phosphorus in the soil profile. Overall, therefore, its losses are very small. The soil texture will not have a decisive effect on the application of phosphorous fertilizers. Phosphorus fertilizers are most often applied in autumn. Also in autumn, it is recommended to apply and incorporate chlorine-containing potassium fertilizers (chloride forms) on soils with good sorption capacity. Potassium cation is bound to the sorption complex of soils. On the other hand, the chloride anion can be leached out of the soil during winter which is of great importance for plants sensitive to higher chlorine content (hops, vines, vegetables). The results of soil analyses for the content of accessible nutrients by the Mehlich 3 method (P, K, Ca, Mg) expressed in mg kg^{-1} of soil are also evaluated in the supply category tables (low – satisfactory – good – high – very high content) in relation to soil texture, i.e. whether the soil is a higher contents of sand, silt, or clay.

5.2 Soil structure

The arrangement of soil particles in the soil profile is also called "soil structure". The arrangement is a dynamic complex that is still not well understood and known. That is why Jacks (1963) said that **"by combining the mineral and organic components an organo-mineral complex is formed which is extremely important for the cycle of life but is less known than photosynthesis."** Many advances in clay mineralogy, colloid chemistry, and other scientific disciplines have led to a better understanding of the genesis, soil structure, and its management. However, even with the current amount of knowledge in this area, we cannot give unequivocally satisfactory answers, and that despite the fact that the soil structure is the most important physical property.

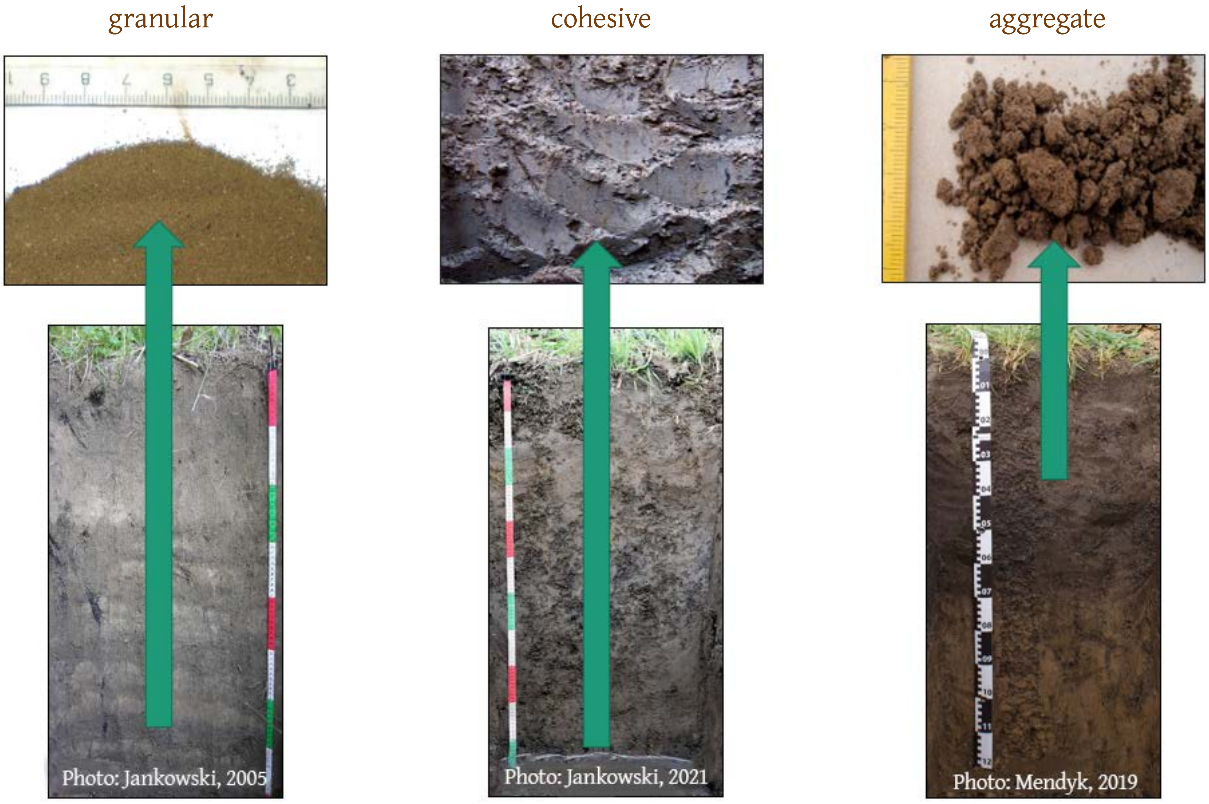
One of the reasons for the soil structure complexity is the range of scales in its evaluation. Soil structure formation processes take place in a range of several Å (Angström = 10^{-10}m) to several cm. Another cause of complexity is the natural dynamics of the soil structure. Structure changes fluctuate in time and space and are characterized by the high dynamics of their changes. As a result, it is very difficult to define soil structure, although the literature is cluttered with numerous and frequent definitions. Several terms are used to easily express and identify structural units including structural shape, aggregate, ped, etc. So far, the soil structure speaks of the shape, size, and arrangement of solid particles and free spaces, the continuity of pores and spaces, their ability to retain liquids and organic and inorganic substances, and their ability to support root growth and development.

5.2.1 Basic division of soil structure

1. Elementary (granular) structure – aggregates are not present and visible. There are no cohesive bonds between the individual elementary particles. It is possible to meet with such structure on soils with a dominant representation of sand.

2. Cohesive structure (compact) structure – no soil aggregates or cracks are present. The whole mass is uniform and cohesive. Such a structure can be encountered in the glue reduction horizons of hydromorphic soils.

3. Aggregate (pedal) structure – represents the soil structure with naturally developed aggregates. Smaller soil particles are arranged in certain clusters and forms.


granular

cohesive

aggregate

Photo: Jankowski, 2005

Photo: Jankowski, 2021

Photo: Mendyk, 2019

Figure 5.6: Basic types of soil structure A) granular, B) cohesive, C) aggregate (Šimanský, 2022)

5.2.2 Division of soil structure by its origin

1. Natural structure – originated by natural soil development. It is a characteristic genetic trait of soil, its individual horizons, and an important criterion in determining soil types. E.g. molar humus horizons of the Chernozem have a small-clumpy structure while the illuvial horizons of the brown earth often have a prismatic structure. Naturally developed aggregates are strong and relatively stable.

2. Artificial (anthropogenic) structure – created by a long-term human cultivation activity (regular cultivation, fertilization, liming). In our conditions, such an artificial structure applies to the topsoil. Through long-term interventions in this soil layer, naturally stable aggregates are transformed into so-called pseudo-aggregates which are in the form of grains, nuts, and large lumps.

Complete description of soil structure include: 1. the type – which notes the shape and arrangement of peds; 2. the class – which indicates ped size; 3. the grade – the distinctiveness of peds. A table of the types and classes of soil structure is in Table 5.9.

Table 5.8: Complete description of soil structure (Foth, 1990)

Type (Shape and Arrangement of Peds)						
Class	Platelike with one dimension (the vertical) limited and greatly less than the other two; arranged around a horizontal plane; faces mostly horizontal	Prismlike with two dimensions (the horizontal) limited and considerably less than the vertical; arranged around a vertical line; vertical faces well defined; vertical angular		Blocklike; polyhedronlike, or spheroidal, with three dimesnions of the same order of magnitude, arranged around a point		
				Blocklike; blocks or polyhedrons having plane or curved surfaces that are cast of the molds formed by the faces of the surrounding peds		Spheroids of polyhedrons having plane or curved surfaces which have slight or no accomodation to the faces of surrounding peds
		With rounded caps	With rounded caps	Faces flattened; most vertices sharply angular	Mixed rounded and flattened faces with many rounded vertices	
	Platy	**Prismatic**	**Columnar**	**(Angular) Blocky**	**Subangular Blocky**	**Granular**
Very fine or very thin	Very thin platy; <1 mm	Very fine prismatic; <10 mm	Very fine columnar; <10 mm	Very fine angular blocky; <5 mm	Very fine subangular blocky; <5 mm	Very fine granular; <1 mm
Fine or thin	Thin platy; 1 – 2 mm	Fine prismatic; 10 – 20 mm	Fine columnar; 10 – 20 mm	Fine angular blocky; 5 – 10 mm	Fine subangular blocky; 5 – 10 mm	Fine granular; 1 – 2 mm
Medium	Medium platy; 2 – 5 mm	Medium prismatic; 20 – 50 mm	Medium columnar; 20 – 50 mm	Medium angular blocky; 10 – 20 mm	Medium subangular blocky; 10 – 20 mm	Medium granular; 2 – 5 mm
Coarse or thick	Thick platy; 5 – 10 mm	Coarse prismatic; 50 – 100 mm	Coarse columnar; 50 – 100 mm	Coarse angular blocky; 20 – 50 mm	Coarse subangular blocky; 20 – 50 mm	Coarse granular; 5 – 10 mm
Very coarse or very thick	Very thick platy; >10 mm	Very Coarse prismatic; >100 mm	Very Coarse columnar; >100 mm	Very Coarse angular blocky; >50 mm	Very Coarse subangular blocky; >50 mm	Very Coarse granular; >10 mm

Terms for grading soil structures are as follows:

Structureless – no observable aggregation or no definite and orderly arrangement of natural lines of weakness. Massive, if coherent, as in clays. Single grined, if noncoherent, as in sands.

Weak – poorly formed indistinctive peds, barely observable in place.

Moderate – well-formed distinctive peds, moderately durable and evident, but not distinct in undisturbed soil.

Strong – durable peds that are quite are evident in undisturbed soil, adhere weakly to one another, and become separeted when the soil is disturbed (Foth, 1990).

The basic unit of the soil structure is the **soil aggregate – ped**. Aggregates are components that were created by combining mineral particles with organic and inorganic substances. Aggregates that resist the water action, which is very important in terms of their stability, are called water-stable aggregates.

Figure 5.9: Soil aggregate, basis of soil structure (Šimanský et al., 2018)

Forces causing particles to bond into soil aggregates

Aggregation is a process that consists of joining and cementing particles into aggregates. Particle bonding is caused by:

- internal and between molecular forces,
- Van der Waals forces,
- electrostatic forces,
- gravitational forces.

5.2.3 Aggregation mechanisms

Bronick and Lal (2005) summarized the mechanisms of aggregate formation as follows:

Soil aggregates are formed in stages – by various mechanisms.

1. Hierarchical theory of aggregate formation states that micro-aggregates are joined and macro-aggregates are formed from them. Micro-aggregates were formed from the organic molecules that formed a bond with clay and a polyvalent cation to form more complex particles which were combined with other complex particles to form a macro-aggregate. The scheme is below:

MICRO-AGGREGATES → MACRO-AGGREGATES

organic molecules (OM) → clay (Cl) → polyvalent cations (P) = composite particles (Cl – P – OM)

composite particles (Cl – P – OM) + composite particles (Cl – P – OM) = macro-aggregates [(Cl – P – OM) x] y

2. The roots and hyphae of microscopic fungi release organic substances that act as the glue connecting particles into wholes. The particles are rearranged during mixing until the soil is stabilized due to alternating wetting and drying cycles. Bacterial micro-aggregates are formed because bacterial colonies and their exudates form a polysaccharide capsule around which the clay particles are arranged and drawn into it due to the alternating soil wetting and drying cycle. Clay flakes form a protective coating against the bacterial colony, thereby inhibiting the decomposition of internal soil organic matter.

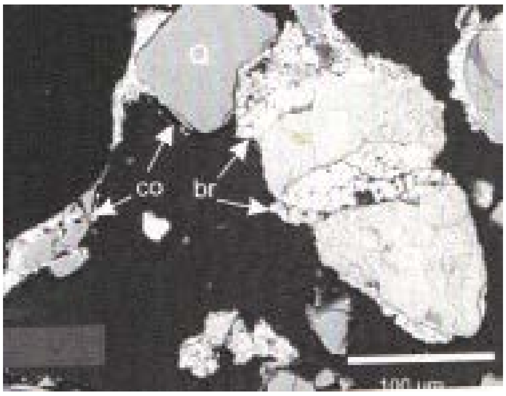


Figure 5.8: Cationic bridges connecting particles into aggregates (Jankowski, 2013)

3. Coagulation of hydroxides, hydrated oxides, phosphates, and carbonates promotes aggregation. Cations such as Fe_3^+, Al_3^+ and Ca_2^+ stimulate the coagulation of substances that bind the components of the primary particles. Cations form bridges between clay and soil organic matter particles which results in the formation of aggregates (Fig. 5.8).
4. Aggregate formation by a combination of these processes (Fig. 5.9).

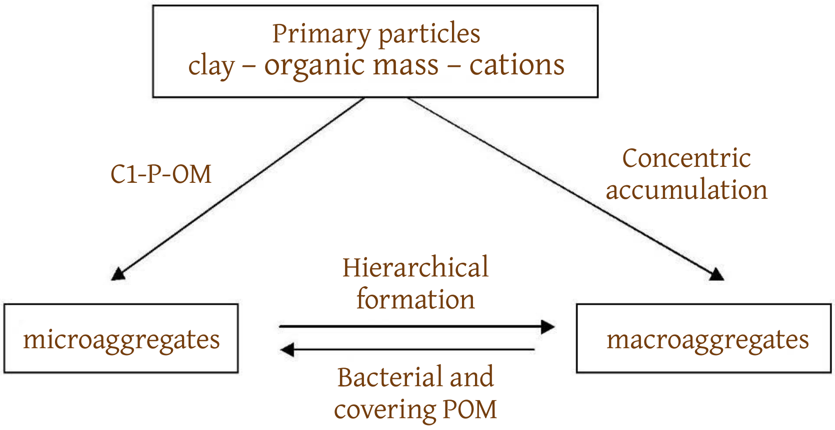


Figure 5.9: Mechanisms of aggregate formation (Bronick and Lal, 2005)

5.2.4 Shape of soil aggregates

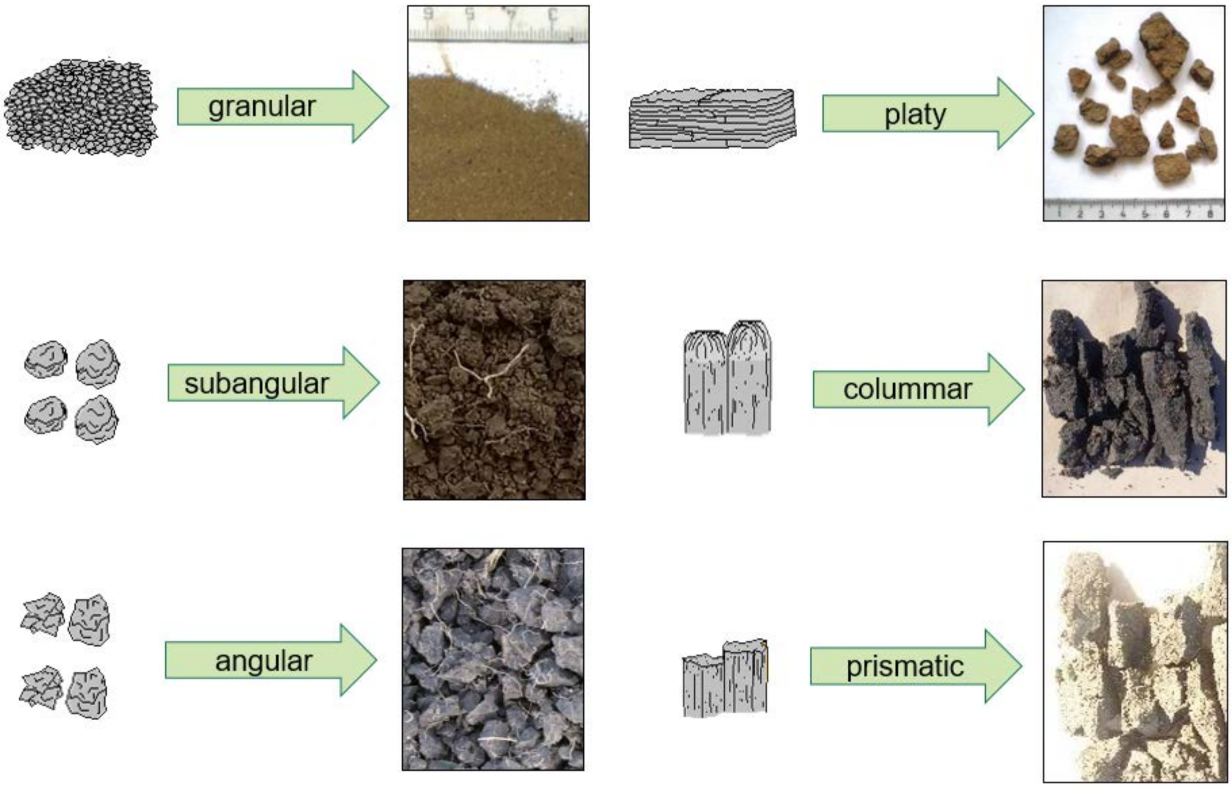


Figure 5.10: Shape of soil aggregates (Šimanský, 2022)

5.2.5 Size of aggregates

The basic unit of the soil structure, as was already mentioned, is the soil aggregate – ped. Primary peds are relatively stable aggregates that are no longer naturally divided into smaller soil units. On the contrary, they can be combined into larger units – aggregates of higher order.

In Slovakia as well as abroad, aggregates are divided into micro and macro aggregates according to the size, while the limit separating micro and macro aggregates is 0.25 mm. In our country (Slovakia), Fulajtár (2006) divides soil aggregates according to the size into three groups:

1. **Micro-aggregates** <0.25 mm,
2. **Macro-aggregates** 0.25 – 10 mm,
3. **Mega-aggregates**> 10 mm.

The aggregates with the size of between 0.5 and 3 mm are considered to be the most agronomically valuable. The division of aggregates according to the size is usually done by sieving through a set of sieves whereby the individual sizes of structural aggregates are obtained.

5.2.6 Aggregate stability

Aggregate stability is the measure of the vulnerability of soil aggregates to external destructive forces and is formed by the association between the mineral content and the components of organic matter. Aggregate stability is difficult to quantify and subsequently interpret because it is influenced by the whole complex of external and internal factors.

Internal factors influencing soil aggregates stability

Soil pH – the stability of the aggregates in acidic soils is ensured due to the higher accumulation of organic substances. On the other hand, in soils with a higher pH, the dissolution of humic substances occurs which can lead to a decline in soil structure stability. The stability of aggregates in alkaline soils is generally ensured by carbonates.

Na$^+$ – with a high content of Na$^+$, the dispersion of clay occurs, so its high content in the soil has a negative effect on the soil structure.

K$^+$ – its effect on the stability of the aggregates is controversial. For example, Quirk and Schofield (1955) described its negative action as identical to the Na+ cation. On the other hand, Levy and Torrento (1995) reported its positive effect that they explain through the lower hydration energy absorbed by potassium which is 72% lower compared to natrium. The larger the hydration shell, the lower its coagulation ability.

Ca^{2+}, Mg^{2+} – if these cations (Ca^{2+}, Mg^{2+}) are present in the soil, the stability of the aggregates increases due to the bonds formed with clay particles and organic carbon. In general, calcium is reported to be more effective than magnesium in binding clay particles and organic colloids that is related to their hydration shells (Mg^{2+} has a smaller hydration shell than Ca^{2+}).

Soil texture – higher content of the clay fraction = higher aggregation and stability → the clay itself – formation of micro aggregates that are stabilized by organic matter. Clay particles are an important cementing agent. However, the decisive factor is not the amount of clay but the type of clay mineral. Due to the physico-chemical properties of clays, the group of smectic clays is more effective for aggregation than other groups of clays, mainly due to its large specific surface and high exchangeable cation capacity. High sand content = lower aggregation and stability → organic matter is of decisive importance.

Figure 5.11: Effect of soil texture on soil structure formation

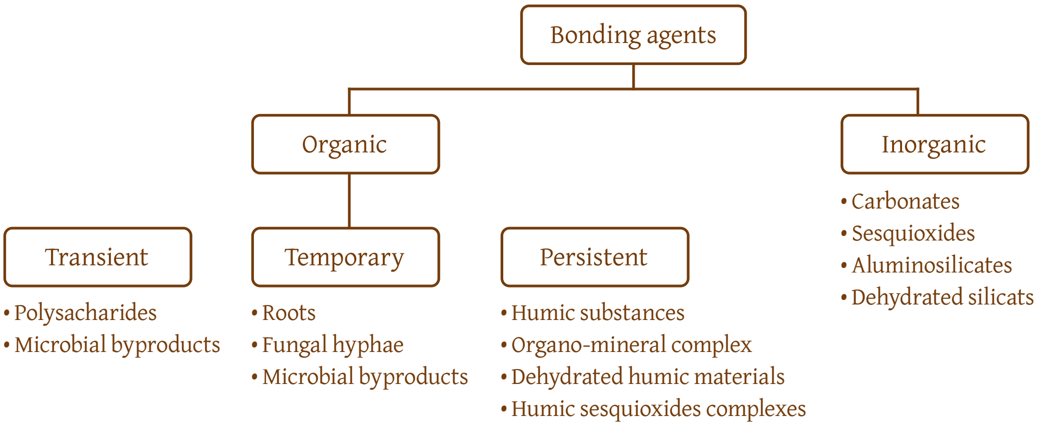


Figure 5.12: Different types of binding agents (Harris et al., 1966)

Soil organic matter – serves as a cementing substance that binds mineral particles together. Figure 5.12 shows the distribution of organic substances according to the length of action to connect particles into units. Soil organic matter also reduces the aggregates wettability and thus affects their mechanical strength. The stability of soil aggregates also depends on the total amount of soil organic matter. The more it is, the more the average size of the aggregates increases in direct proportion, which in turn leads to a higher stability of the aggregates.

Oxides and sesquioxides of iron and aluminium – They have a demonstrable effect on the stability of micro aggregates in particular. Their stabilizing effect consists in: iron and aluminium in solution have a flocculating effect, sesquioxides participate in the bonds between clay particles and organic molecules, sesquioxides can precipitate as gels on the surface of clay particles.

External factors influencing soil aggregates stability

Alternation of wetting and drying cycles of the soil – when the soil is significantly moistened, it becomes a more viscous mass. The connection forces between the particles of this material increase with the increase in humidity up to a certain limit, and then begin to decrease again. The size of the particles, the content of mineral and organic colloids in the soil and the content of bound cations are very important. Capillary forces arise in the soil mass when the soil moves from an overmoistened state to a lower humidity. The more colloidal particles the soil contains and the simpler the granular composition is, the greater its ability is to create cracks. The stability of the aggregates is also influenced by the soil moistening speed. Aggregates that will be moistened slowly will maintain their structure, while on the other hand, if they are quickly moistened, their destruction may occur (Fig. 5.13).

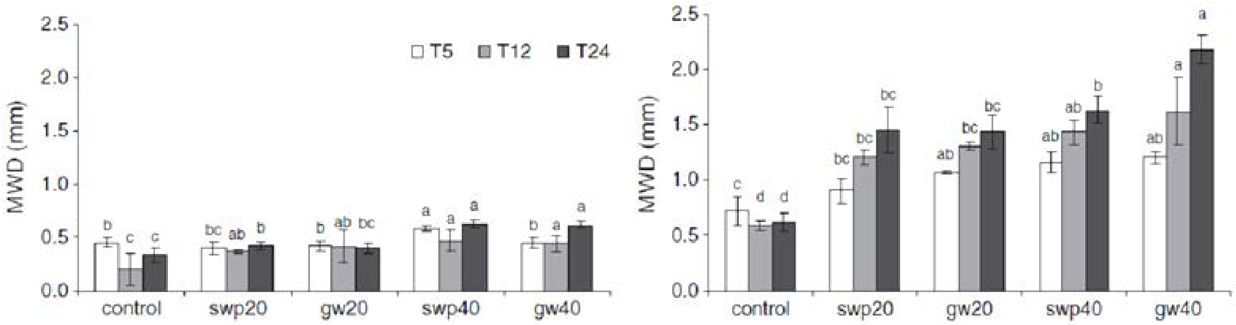


Figure 5.13: Effect of wetting rate on the aggregates stability at different levels of fertilization; *swp* – compost from municipal waste, *gw* – green manure, *20, 40* – rates in t ha^{-1}, *T5, T12, T24* – manurity of compost (in weeks) (Grosbellet, et al., 2011)

Alternation of thaw and freeze cycles – as the soil thaws, it becomes less cohesive because it breaks down into aggregates under less pressure. This is also related to the lower ploughing resistance during spring ploughing of frozen soils because cracks have been formed in them. Friability during freezing depends on the sand content. With a higher sand content, the friability of the soil decreases when the soil freezes. Also, with a low water content, the structure-forming effect of frost is very weak. Repeated freezing and thawing of soil increases fragmentation into aggregates. Fine cracks appear already during the first frosts and are filled with water during thawing and become larger during further freezing. The stability of aggregates is also influenced by the amount of water in the soil and the duration of its freezing. Aggregate stability decreases linearly with an increase in the volume of frozen water in the soil. At low temperatures of -1 to -2 °C in the soil, water freezes only in the coarsest pores, and at temperatures of -4 to -5 °C also in finer pores.

Biotic factors – soil biota plays an important role in aggregation processes and the formation of soil structure. Plant roots play an important role because they penetrate the soil and create a dense tangle. More massive roots create pressure on the soil during their growth while the gaps and pores between the aggregates are getting smaller until finally, the soil is compact also in the spaces between the plant's roots so that cracks are formed in the soil mass. The pressure that is developed during the growth of the roots in the soil affects the soil mass which crumbles and breaks up into aggregates. Root hairs are involved in connecting the aggregates and when they die, they enrich the soil with organic matter. The soil organic matter acts as a glue that binds the mineral particles together. Plant roots participate in drying the area around the roots which leads to the orientation of clay particles parallelly to the axis of the plant roots and the retention of soil particles. They also support the expansion and development of microbial communities in the rhizosphere. They increase the concentration of ions in the soil solution and release polyvalent cations. Vegetation participates in the stabilization of aggregates with root tangles, while soil microorganisms produce secretions (especially bacterial mucus serves as a sealant) and fungal hyphae (surround the surface and penetrate the aggregates, thereby strengthening them). Bacteria participate in the formation of micro aggregates while the activity of fungi is significant in the formation of the macro aggregates. After the death of plants and soil organisms, they become an important component of humus-forming material. Animal organisms participate in the formation of the structure, similar to the roots of plants, by crushing and loosening the soil mass. Earthworms affect the soil structure the most, and through their digestive system, they increase the share of agronomically valuable structure (Fig. 5.14).

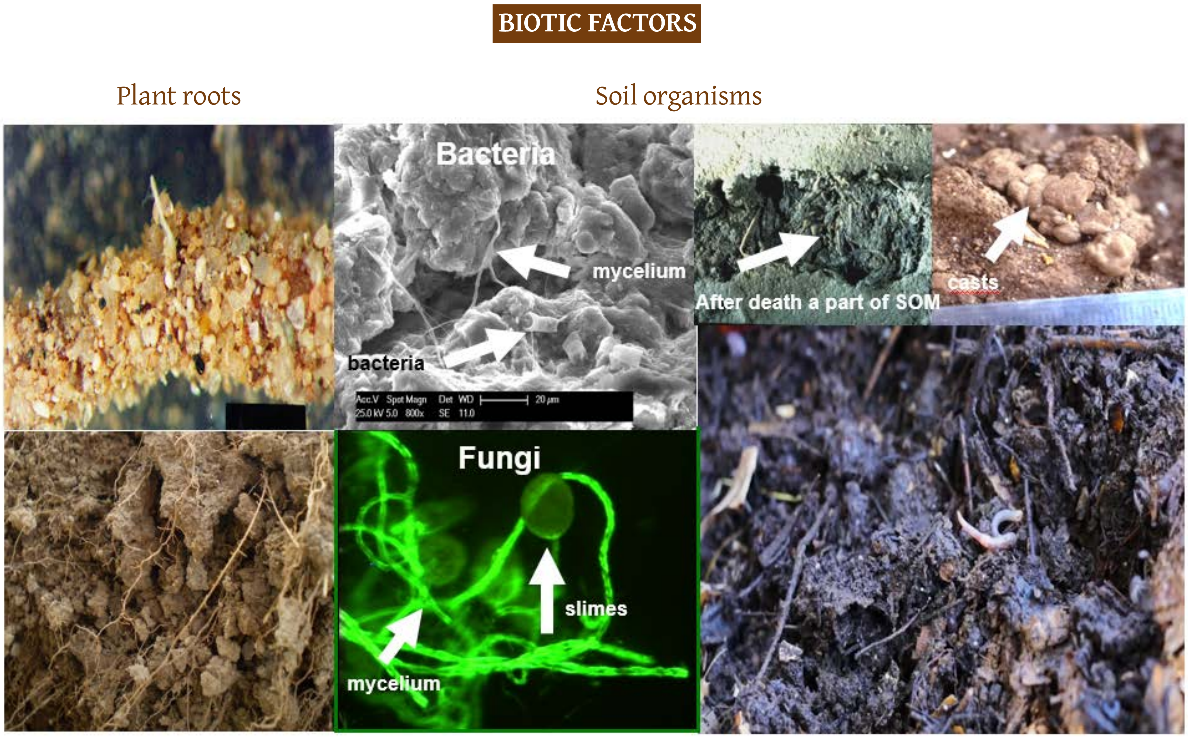
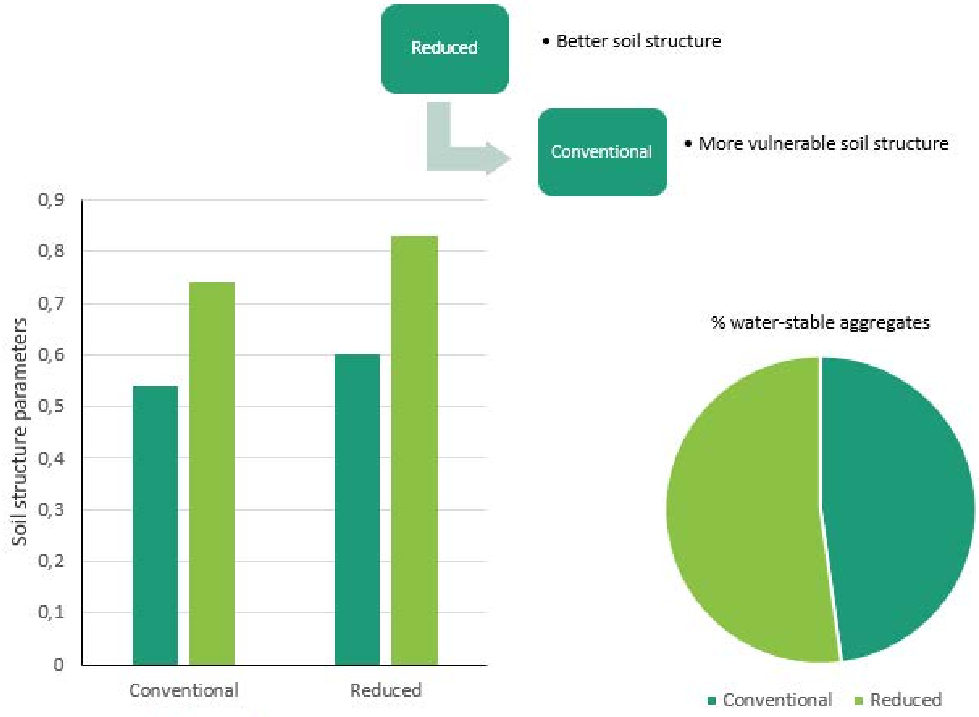


Figure 5.14: Effect of biotic factors on soil aggregates (Šimanský, 2022)

The action of mechanisms for soil tillage – every machine and working tool affects the structure of the soil specifically, so choosing the right tool is very important when processing soil. However, the main factor remains the soil moisture during processing and tillage.

What are the differences in soil structure depending on the tillage system?



In reduced tillage system: stability, +12%; MWD, 11%; %WSA +9%

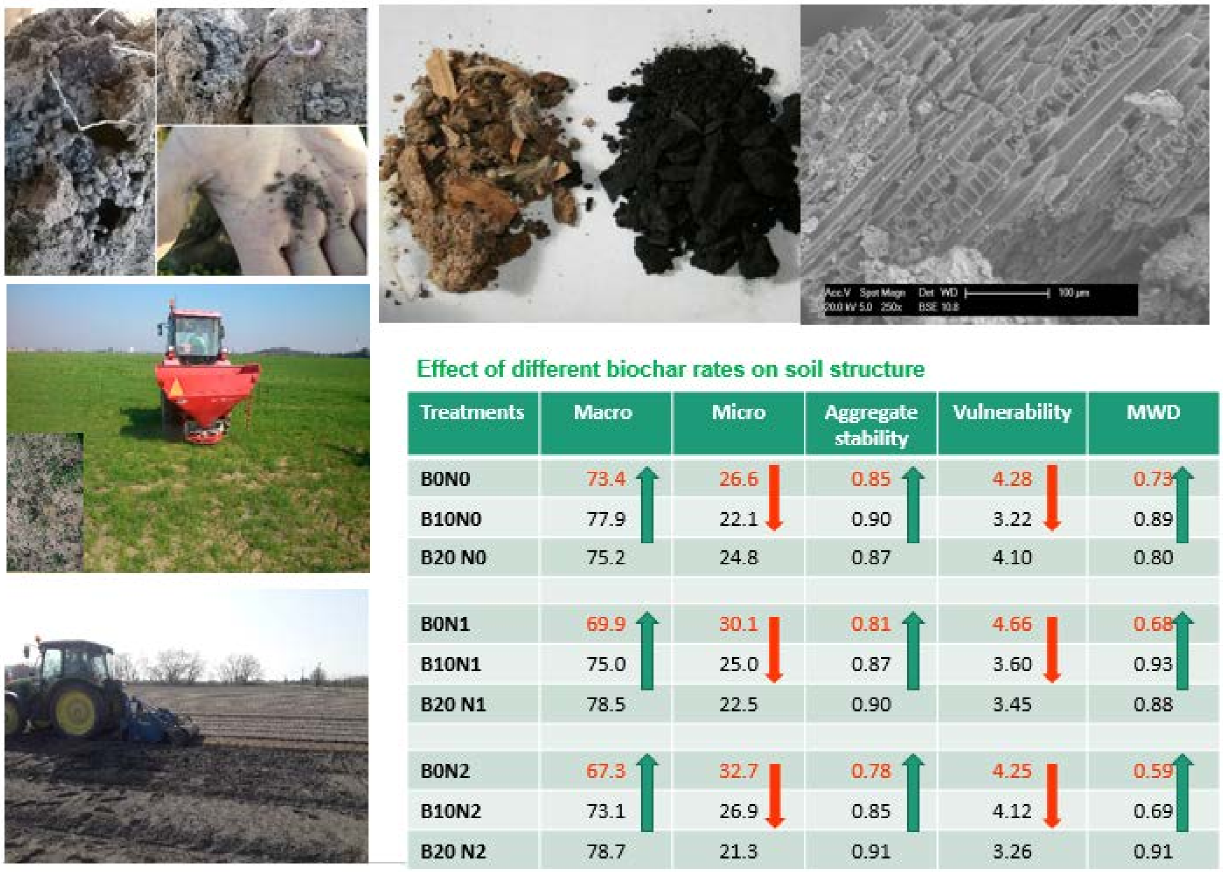
Figure 5.15: Effect of different tillage system on soil structure (Šimanský et al., 2008)

Soil management practices and soil use – the stability of soil aggregates is significantly the result of anthropogenic interventions in the soil through:

1. **Soil tillage** – depends on the method and tillage intensity. Intensive machining = lower stability of aggregates, minimization machining = higher aggregates stability (Fig. 5.15).
2. **Irrigation** – more intensive irrigation = flooding of aggregates = lower stability.
3. **Application of farmyard manures, organic substances, biochars, organic amendments etc.** – increasing organic substances = higher aggregate stability (Fig. 5.16).
4. **Plant residues and crop rotation** – post-harvest plant residues on the surface dampen the destructive effect of precipitation, and are the source of organic substances = higher aggregates stability, e.g. post-harvest residues of perennial fodder are a high source of organic matter.
5. **Application of mineral fertilizers** – higher content of mineral fertilizers accelerates the mineralization of organic substances, accelerates ion exchange and the soil is more susceptible to clay dispersion = lower aggregates stability, on the other hand, through mineral fertilizers, there is a higher content of nutrients in the soil which is reflected in the formation of above-ground, but also underground biomass = higher aggregates stability.

6. **Soil conditioners** – serve to stabilize the soil structure. Application of gypsum into the soil leads to aggregation improvement. For improvement or stabilization of the already favorable structural state in the soil environment, various synthetic organic polymers or soil conditioners (stabilizers) are used. In this way, they differ from the natural organic matter that not only stabilizes the structure but also forms it with the participation of microorganisms. The practical application of structural stabilizers is limited due to their high costs. It is used to solve special problems, such as the short-term reduction of the erosion danger on the slopes of technical works unless the growth of vegetation is established.

BIOCHAR → Organic amendment for improving soil structure?



Effect of different biochar rates on soil structure

Treatments	Macro	Micro	Aggregate stability	Vulnerability	MWD
B0N0	73.4 ↑	26.6 ↓	0.85 ↑	4.28 ↓	0.73 ↑
B10N0	77.9	22.1	0.90	3.22	0.89
B20 N0	75.2	24.8	0.87	4.10	0.80
B0N1	69.9 ↑	30.1 ↓	0.81 ↑	4.66 ↓	0.68 ↑
B10N1	75.0	25.0	0.87	3.60	0.93
B20 N1	78.5	22.5	0.90	3.45	0.88
B0N2	67.3 ↑	32.7 ↓	0.78 ↑	4.25 ↓	0.59 ↑
B10N2	73.1	26.9	0.85	4.12	0.69
B20 N2	78.7	21.3	0.91	3.26	0.91

Figure 5.16: Effect of biochar on soil structure (Šimanský et al., 2018a)

5.2.7 Evaluation of soil structure

The most common method of evaluating soil structure is based on the ratios of aggregate fractions obtained by sieving through a set of drought sieves (Revut, 1972).

$$K = \frac{A}{B}$$

K – structure coefficient

A – weight of dry-sieved aggregates from 0.25-7.0 mm

B – weight of the sum of dry-sieved aggregates <0.25 mm and >7 mm

Higher values of the coefficient of structure indicate more favorable soil structure.

The mean weight diameter (MWD_d) can be calculated from the proportional occurrence of individual fractions of structural macroaggregates (obtained by dry-sieving: using a set of sieves):

$$MWD_d = \sum_{i=1}^{n} X_i W_i$$

X_i – mean diameter of each size fraction (mm)
W_i – the proportion of the total sample weight occurring in the corresponding size fraction
n – the number of size fractions

The weighted average of water-stable macroaggregates (MWD_W) can be calculated from the proportional occurrence of individual fractions of water-stable macroaggregates (obtained by sieving using a set of sieves in distilled water)

$$MWD_W = \sum_{i=1}^{n} X_i\, WSA$$

X_i – mean diameter of each size fraction (mm),
WSA – content of water-stable aggregates
n – the number of size fractions

Nowadays, a coefficient of vulnerability (Kv), which determines how many times the initial average size of aggregates is reduced due to the degradation mechanisms, is used to evaluate the soil structure. The values of Kv of soils with optimal and stable structures are equal or are about to reach number 1. The coefficient of vulnerability (Kv) can be calculated based on the relation:

$$K_V = \frac{MWD_d}{MWD_W}$$

MWD_d – the mean weight diameter of aggregates for dry sieving (mm)
MWD_W – the mean weight diameter of WSA (mm)

The index of the stability of water-stable aggregates (Sw) can be calculated based on the soil texture and proportional occurrence of water-stable aggregates. If the values of Sw are higher than 1, then the soil structure is more stable.

$$S_W = \frac{WSA - 0.09\ \text{sand}}{\text{silt} + \text{clay}}$$

WSA – content of water-stable aggregates

Soil organic matter plays an important role in the formation and stabilization of aggregates; therefore, Pieri (1991) suggested the assessment of aggregation by the so-called critical content of the soil organic matter (St):

$$S_t = \frac{SOM}{(\text{clay} + \text{silt})}$$

SOM – soil organic matter

Visible features of soil structure decline are summarized in Figure 5.17.

Figure 5.17: Visible features of soil structure decline (Šimanský, 2022)

5.3 Specific weight of soil

The specific weight of soil is one of the most important basic physical properties. The specific weight of soil is the weight of the solid dry soil phase, i.e. weight of 1 m^3 of dry soil (with no pores/voids) expressed by t m^{-3}. The solid soil phase is the soil with no pores and dry soil is the soil that was dried out at 105 °C. The specific weight of soil can be also defined as the ratio of the weight of the solid soil phase to the weight of the same water volume at 4 °C. It can be expressed in g cm^{-3}, Mg m^{-3}, t.m^{-3} and calculated according to the relation:

$$\rho_s = \frac{m_s}{V_s}$$

ρ_s – specific weight of soil (t m^{-3})
m_s – soil weight (t)
V_s – volume of soil with no pores/voids (m^3)

Values of ρs depend on the specific gravity of mineral and organic constituents in soil. The proportion of the above mentioned constituents in the solid soil phase can be different and therefore, their final values in soils can differ as well. Foth (1990) claims that if feldspar and quartz, the most common minerals that can be found in soils, prevail in soils, the values of specific weight of soil tend to be between 2.56 – 2.76 t m^{-3}, or 2.65 t m^{-3}. Shukla (2014) maintains that the values of specific weight of soil are rising concerning the texture of minerals as follows: sand < silt < clay. Fulajtár (2006) states that ρ_s in soils of Slovakia is usually higher than the average value 2.65 t m^{-3}. In A horizons, the value of the ρ_s fluctuates between 2.65 – 2.7 t m^{-3}, but in parent materials or bedrock, it is higher from 2.7 – 2.8 t m^{-3}. It is logical because less soil organic matter is found in most soils in parent materials and on the contrary, there is more soil organic matter than mineral particles in A-horizons. The high content of soil organic matter causes a decrease of the ρ_s value under 2.5 t m^{-3} and even under 2 t m^{-3} in Histosols.

Table 5.10: Specific weight of selected soil components (Šimanský et al., 2018)

Mireral	Specific weight
Biotite	2.7-3.3
Calcite	2.72-2.94
Dolomite	2.86
Gibbsite	2.38-2.42
Hematite	5.26
Illite	2.60-2.90
Kaolinite	2.61-2.68
Montmorillonite	2.0-3.0
Muscovite	2.77-2.88
Orthoclase	2.55-2.63
Pyrite	5.02
Quartz	2.65
Mica	2.58-2.83
Vermiculite	2.3
Other components	**Specific weight**
Organic matter	1.0-1.4
Water	1.0
Air	1.0×10^{-3}

ρ_s changes only a little. Its values can be reduced more often in the A horizon (topsoil) as a result of the higher content of soil organic matter and humus but also due to the application of organic fertilizers (peat, compost, crop residues, manure, biochar, biochar substrates, and other organic amendments). The values rise as a result of the reduction of the humic substance content in soils. Higher values are also a consequence of intensive cultivation or continuous use of higher doses of mineral fertilizers. Values of ρs can be increased by erosion or topsoil deepening. As a result, the higher mineral proportion from the lower horizons can be found in the topsoil.

5.4 Bulk density

Soil bulk density is the weight of the undisturbed volume unit of dry soil. It is expressed in g cm^{-3}, Mg m^{-3} or t m^{-3}. When it is compared to the specific weight, it is lower because bulk density contains also the weight of individual pores. Values of bulk density depend on the soil texture (especially on the content of clay which is able to increase its volume due to the water absorption but on the other hand, its volume decreases as a result of soil drying), spacial arrangement of soil particles (soil structure) and content of water in the soil. Foth (1990) states that soils with a granular structure or massive soil structure have values of bulk density from 1.6 to 1.7 t m^{-3}. On the other hand, the values of bulk density of soils with properly developed aggregates, i.e. aggregates with properly developed soil structure, are lower (1.3 t m^{-3}). Fulajtár (2006) maintains that unstructured soils with prevailing elementary particles settle naturally to the values of 1.8 – 2 t m^{-3}, but the microstructured soils with prevailing microaggregates settle at values of 1.5 – 1.6 t m^{-3} and macrostructured soils at values of 1.1 – 1.2 t m^{-3}. Values of bulk density in A horizons of mineral soils fluctuate from 1 – 1.8 t m^{-3}; in organic soils, they are lower than 1 t m^{-3}; and in Histosol, they fall at 0.2 t m^{-3}. We distinguish:

1. Bulk density which is not reduced (ρ_w) refers to the weight per unit volume of soil (m^3) with the actual content of water and air. It can be calculated as follows:

$$\rho_w = \frac{m_t}{V}$$

m_t – soil volume at average moisture (t)
V – soil volume

Wet bulk density is a very dynamic property. It expresses the actual soil weight which often changes depending on the soil moisture.

2. Reduced bulk density (ρ_d) refers to the weight per unit volume of soil after drying the soil sample at 105 °C. It can be calculated as follows:

$$\rho_d = \frac{m_s}{V}$$

m_s – volume of dry soil (dried at 105 °C)
V – soil volume

In the soil profile, bulk density tends to increase with depth. Some pedogenetic processes (translocation, accumulation, internal soil weathering, etc.) and anthropogenic factors (technological compaction, deep soil loosening and knifing) support this tendency but on the other hand, they also destroy it. Bulk density without giving information about the soil horizon or the layer is always related to the surface horizon.

In space and at a particular time, the value of bulk density is not stable. In space, it is varying depending on the level of settling or compaction, content of humic substances, soil texture, and development of the plant root system. This spacial variability is eliminated by determining the average value of at least from three to five repetitions. The only determination for broader conclusions is not satisfactory.

One of the ways how to obtain a relatively stable value of bulk density is to determine the so-called equilibrium bulk density. It is the state when the soil, after spring thaw and processing, becomes compact gradually and reaches the state when this process is brought to an end and bulk density does not change any more. This state occurs in soils with spring crops approximately in the second half of June and in soils with root crops in July. In soils with winter crops, the equilibrium state should be observed in early spring after the process of offshooting and evapotranspiration of the higher spring moisture (April – May). The equilibrium state of bulk density defines soils which become compact naturally, and it does not change a lot.

Bulk density is important because it directly influences several physical properties of soil as well as some biological and chemical properties. The change of bulk density immediately causes the change of these properties, for example an increase of bulk density increases the proportion of the solid soil phase, hardness, soil density, penetrometric resistance. These changes reduce total porosity, air capacity, air and water permeability, but on the other hand, they deteriorate conditions for plant rooting and development of the root system. In general, it is claimed that an increase in bulk density of the soil density deteriorates the physical state of soil and vice versa. The values of bulk density are influenced by humans a lot.

When assessing particular values of bulk density from the viewpoint of the total assessment of the soil's physical state as well as from the point of view of cultivated crops, we recognize three categories of bulk density: optimal bulk density, permissible bulk density, and critical bulk density.

Optimal bulk density creates optimal physical conditions in soils and it results in higher yields. Its values depend on the plant species and soil texture. It ranges from 1.1 to 1.3 t m^{-3} in loamy soils, but it is higher, from 1.1 – 1.4 t m^{-3}, in sandy soils. The seedbed of cereals should be hardened into the bulk density of 1.3 t m^{-3} and the surface soil should be from 0.85 – 1 t m^{-3}. For potatoes planted in ridges, values from 1.1 – 1.2 t m^{-3} are required.

Between **optimal and critical bulk density**, there is a specific interval of values. Within them, most crops have average but also higher yields. This bulk density is called **permissible bulk density**.

Table 5.11: Optimal, permissible and critical bulk density for the crops (Fulajtár, 2006)

Crop	Bulk density (g cm^{-3})					
	1.1	1.2	1.3	1.4	1.5	1.6
Wheat	-	Optimal	Optimal	Permissible	Critical	Critical
Barley	-	Optimal	Optimal	Optimal	Permissible	Critical
Maize	-	Optimal	Optimal	Permissible	Permissible	Permissible
Potatoes	Optimal	Optimal	Optimal	Permissible / Critical	Critical	Critical
Sugar beet	-	Optimal	Optimal	Permissible	Critical	Critical

Critical bulk density is the bulk density that deteriorates the physical state of soil to such an extent that it limits the growth of plants and roots, and it also contributes to lower yields. In our climatological conditions, this state occurs at 1.5 t m^{-3}. Critical values are observed in Gleysol, and Planosols, in illuvial horizons, and in layers of the plow pan with unsuitable texture and labile structure. It has been proved that if the bulk density is higher than 1.94 t m^{-3}, no plants are grown there.

5.5 Porosity

The term ´porosity´ is defined as the total content of pores and voids (Fig. 5.18) in soil. Pores and voids are important because they are filled with air and water, and play an important role in the soil production capability. The number of pores in the given volume of soil affects the soil mass significantly. Porosity is an important soil property because of its ability to supply soils with water and support the nutrient cycle as well as the growth and development of plant roots.

Figure 5.18: Voids (pores) in the soil (Šimanský, 2013)

Porosity is an important soil property not only from the agronomical point of view but also from the soil-conserving and pedogenetic point of view because water is held in the pores of the soil, and the pores are responsible for the movement of water and migration of substances in the soil as well as for the processes of pedogenesis. Porosity determines the content and composition of air in the soil because it significantly determines the exchange of gases between the soil medium and atmosphere.

Porosity together with the soil structure is the main indicator of the spatial arrangement of the matter of soil and it refers to the fact that the soil does not have a compact structure but a pore structure. In pores, physical, physical and chemical, chemical, and biological processes take place. A low volume of pores or their unsuitable size can slow down and also prevent the root system from developing (for example: above the hardened illuvial horizon with low porosity and with a high abundance of small pores, there is the main proportion of the root biomass and roots are not able to penetrate deeper through the hardened horizon).

Total porosity depends on the organic and mineral proportion of soil. The high content of organic matter in soil is responsible for greater porosity. The total number of pores in soils with a lower content of silt and clay is low despite the fact that individual voids are large. Soils with high contents of silt and clay have a high relative volume of pores and if the aggregate ability of these soils is good, their total volume of pores can equal to or be higher than the volume of solid particles. It is caused by the fact that pores do not exist only among particles but also among aggregates.

Total porosity of soil (P) is the total volume of all pores and voids that occur among solid particles expressed (%) from the volume of soil. It is not determined directly, but it can be calculated from the specific weight and bulk density as follows:

$$P = \frac{\rho_s - \rho_d}{\rho_s}$$

ρ_s – specific weight

ρ_d – reduced bulk density

If we take into account that mineral soils have an average absolute specific gravity 2.65 t m^{-3} and an average bulk density 1.3 t m^{-3}, then the average total porosity in these soils is approximately 50%.

Pores are divided into (Fulajtár, 2006):

- **inter-aggregate pores** (intergranular pores) – situated between individual aggregates,
- **intra-agreggate pores** – inside aggregates,
- **trans-aggregate pores** – they get through more aggregates.

When total porosity is assessed, it is important to be familiar with the arrangement of pores in the matter of the soil. Therefore, the total volume of pores is the sum of inter-aggregate pores and intra-aggregate pores. **Good porosity of the topsoil (P>50%) should have 1/3 of inter-aggregate pores and 2/3 of intra-aggregate pores.** The highest intra-aggregate porosity was determined in Chernozem with small and cloddy structural aggregates. It is stated that the proportion of coarse, medium, and fine pores should be the same. When the porosity is assessed, we assess the number, size, shape, and continuity of pores.

Moreover, pores can be divided into two categories on the basis of their size and function (Table 5.12)

Table 5.12: Classification of pores according to their size and function (Lal and Shukla, 2004)

	Size of cylindrical pores (µm)	Function
Conductive pores	> 50	Movement of air and draining excess water
Storage pores	0.5-50	Retention of water against the force of gravity and release
Residual pores	0.5-0.005	Retention and diffusion of ions in solutions
Connected pores	<0.005	Support of main forces between soil particles

In our conditions, the classification of pores according to energetic principles is used. On the basis of these principles, pores are classified into capillary pores, non-capillary pores and semi-capillary pores (Zaujec et al., 2009).

1. Capillary pores (Pc) – inside them, capillary forces act and therefore they are called the pores with tension. Retention and movement of water in capillary pores are affected by capillary forces. When water moves, they decrease the gravity action of the Earth and enable the movement of water by capillary action. In these pores, penetration of air into the soil and also its movement are reduced. Hair roots of plants, that are able to find necessary nutrients in the soil solution, penetrate into them. Most chemical, physical and chemical, and biochemical reactions take place in them. The volume of capillary pores equals to the retention water capacity (Θ_{RWC}).

$$Pk = \Theta_{RWC}$$

The size of capillary pores is very important for the existence of living organisms in soils. According to the accessibility for microedaphons and hair roots, capillary pores are divided as follows (Fulajtár, 2006):

1. < 1 µm – uninhabitable for bacteria,
2. 1 – 10 µm – inhabitable for bacteria, inaccessible to hair roots,
3. 10 – 20 µm – accessible to hair roots, inaccessible to protozoa and algae,
4. > 20 µm – accessible to all microorganisms.

2. Non-capillary pores (Pnc) enable the downward movement of water because of the forces of gravity. These pores are the pores almost with no tension and they enable the exchange of air between the soil and atmosphere. The size of non-capillary pores is more than 0.2 mm. It can be calculated if the thirty-minute moisture ($\Theta_{30'}$) is subtracted from the total porosity.

$$Pnc = P - \Theta_{30}{}'$$

3. Semi-capillary pores (Ps) are a temporary category of pores. Within them, capillary forces and forces of gravity act. The volume of semi-capillary pores is calculated as follows:

$$Ps = \Theta_{30'} - \Theta_{RWC}$$

Θ_{RWC} – water retention capacity

$\Theta_{30'}$ – thirty-minute moisture

5.5.1 Shape of pores and connectivity

The shape of pores can be vesicular (spherical or elliptical) or tubular (cylindrical or elongated). Some pores have an irregular shape, especially those in gravelly soils. Connectivity and interconnection of pores are also very important with respect to the movement of liquids and transport processes in the soil as well as with respect to root growth. Vertical connectivity through a horizon is crucial to the transport of water and the exchange of air with the atmosphere. The vertical connectivity can be: low < 10 mm, medium 10 – 100 mm a high > 100 mm.

5.5.2 Origin and formation of porosity

A classification system of porosity can be also based on the origin or genesis of soil pores, as stated by Lal a Shukla (2004).

Macropores (> 75 µm) or conductive pores are formed by the biotic activity, developmnet of contracted ruptures, ice melting, activity of soil animals and soil cultivation. The content of the soil organic matter and clay minerals plays an important role in creation and stabilisation of macropores. Macropores are influenced a lot by an anthropogenic activity. Macropores are present mainly among structural aggregates. They are often called as non-capillary pores because of quick infiltration of water into pores. They are also important because of the exchange of gases between the soil and atmosphere.

Mesopores (75 – 30 µm) or retentive pores are important because of the plant growth. Mesopores are formed by the creation of micro-ruptures when the soil shrinks, by the changes of freezing and thawing, destruction and blocking of macropores at sedimentation or precipitation, as well as by the development of root hairs and hyphae of fungi. These pores are influenced by the soil texture, the content of organic matter, and clay minerals, and are only slightly influenced by soil cultivation and management. They are important for the transport of water in the soil and the exchange of air in structural aggregates.

Micropores (30 – 5 µm) are created by the process of soil matrix shrinkage and destruction of mesopores. Micropores are influenced by the soil cultivation and management only a little, and they are biologically inactive. These pores are always filled by water; they are inaccessible to microorganisms and they can be strategically important for carbon sequestration.

5.5.3 Evaluation of porosity

Total porosity tends to be lower in sandy soils than in loam and clay soils. On the other hand, it is higher in topsoils and A horizons than in subsoils and deeper horizons. In light-textured soils, the values of porosity of A horizons are between 36-50% with a large proportion of coarse pores. In heavy-textured soils, porosity is between 41-60% with a large proportion of fine capillary pores. The average values of porosity in A horizons of medium-textured soils are between 41-55% and occurrence of size categories of pores is more proportional. Porosity decreases unevenly towards the lower layers and in the lower parts of the profile, it amounts to 30-40%.

Low porosity of sandy soils is caused by the fact that the soil structure is not developed. Pores are mainly situated between individual grains of sand. Maximum values of porosity of loam soils are a result of optimum conditions for the formation of the soil structure. On the other hand, heavy clay soils are compact, and inside them, the value of interaggregate porosity decreases and total porosity is lower than in loam soils.

Due to the fact that soil porosity is dependent on the soil structure, it is possible to assess the soil structure using the values of porosity. It is easier to identify the reduced bulk density and, on its basis, to assess the soil structure (Tables 5.13 and 5.14).

Table 5.13: Soil classification according to porosity (Kosil, 1978)

Indication of state	Porosity(%)	
Consolidation and mellowness	Loamy and clay soils	Sandy soils
Topsoil		
very mellow	> 65	> 65
mellow	65 – 55	65 – 50
slightly consolidated	55 – 45	50 – 40
consolidated	< 45	< 40
Subsoil		
very mellow	> 57	> 50
mellow	57 – 46	65 – 50
consolidated	46 – 35	43 – 35
over consolidated	< 35	< 35

Table 5.14: Assessment of the structure of the A horizon (Kutílek, 1978)

Structure of the A horizon	bulk density (t m^{-3})	porosity (%)
excellent	1.2	>54
good	1.2 – 1.4	46 – 54
unsuitable	1.4 – 1.6	39 – 46
non-structural soil	1.6 – 1.8	31 – 39

5.6 Packing density

The indicator "packing density" (PD) is used to assess the soil compaction which is closely connected with the above-mentioned physical properties of soil (bulk density, soil structure, soil texture, porosity). This parameter is influenced mainly by the soil bulk density and content of clay particles. The assessment of the potential vulnerability of soil to compaction is advantageous because the emphasis is put on the content of clay which plays an important role in the determination of soil susceptibility to compaction but it also enables the prediction of soil vulnerability as the genetic compaction as well as technogenic compaction. It can be calculated as follows:

$$PD = \rho d + (0.009 \times \mathit{content\ of\ clay})$$

ρd – reduced bulk density

the content of clay means the content of particles <0.002 mm

Table 5.15: Criteria for evaluating packing density (Novák and Válla, 2002)

Textural classes	Packing density (g cm-3)		
	<1.40	1.40 – 1.75	>1.75
	Vulnerability		
Sandy soil	low	low	medium
Loamy sand	low	low-medium	medium
Sandy loam	low	medium	medium
Loam soil	low	medium	medium-high
Clay loam	medium	high	hardened
Clay soil	medium	high	hardened

5.7 Soil color

When the soil profile is described, the first thing which attracts almost everybody is the soil color which is the most important morphological feature of soil indicating a lot about the soil genesis and individual horizons. It is also an indicator of climatic soil groups. Soils in humid, mild, and cold zones are predominantly grey. Black and yellow colors originate in tropical and subtropical regions. In mild warm regions, the color of soil is red as a result of soil development on the limestone. In steppe and hydric conditions, the soil color is black. In general, soils developed on alkaline rocks are darker than soils developed on acidic rocks. Color provides a lot of information about soil, for example, the darker the soil is, the more productive it is. It is a result of the higher content of organic matter. The soil color directly affects thermal conditions in soils which can further influence moisture conditions. Black and dark soils absorb more heat than pale soils. Therefore, dark soils can have temperatures from 7 to 8 °C higher than pale soils. Consequently, dark soil evaporation is higher than pale soil evaporation. Under the same conditions, dark soils tend to dry out quicker than pale soils.

The soil color is the basic physical property that is influenced by the content of mineral and organic constituents of soil. As for mineral constituents, compounds of iron, limestone, feldspar, quartz, clay, and manganese are the main components of the matter of soil; on the other hand, humus is the main constituent of the organic part of the soil. Humus is the main color component mainly in surface soil layers. It is responsible for the greyish brown, brown, and black colors of soils. Soils with a higher content of humus are darker in comparison to soils with a lower content of organic matter and humus.

As for the color, compounds of iron, that are present almost in all soils, differ depending on the iron valence. Trivalent iron compounds, that are more widespread compounds in soils, are responsible for the red, rusty-brown color of soil but soils can also be yellow depending on the degree of hydration. In some soils, the color of iron compounds changes because of moisture, for example, the moist red color is yellow-brown after drying out. Bivalent iron compounds are responsible for soils with a greenish and bluish tone of color. They are present in subsoils of not much aerated wet erratic soils. On the other hand, quartz is responsible for the whitish and greyish hue of soils, for example, greyish-fawn, whitened, leached subsurface eluvial horizons of Luvisol and Podzol. Limestone is white, grey, and sometimes olive green. Feldspar has different colours, but the red one prevails. Clay is grey, white or red. Compounds of manganese are responsible for dark or purple red colour of soil.

When we want to describe the soil profile, the soil color is determined in the state of natural soil moisture. Colors are determined according to the Munsell color tables (created by A. H. Munsellom in the USA in 1905 – systematic classification of colors). When colors are determined using the Munsell tables, basic color (hue), value and chroma are detected.

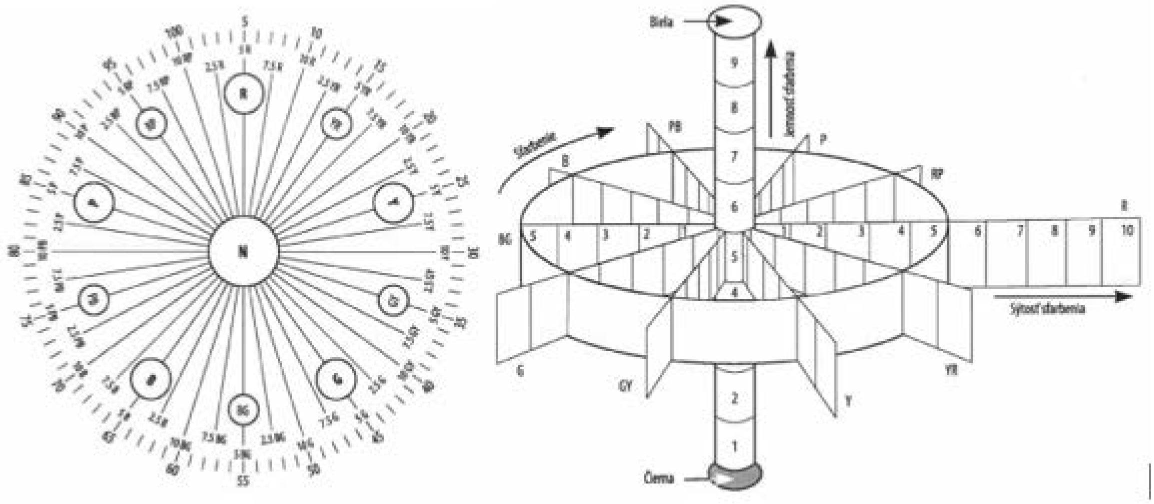


Figure 5.19: Principle of soil color assessment using Munsell color tables (Mocek et al., 2016)

Basic color – hue (Hue) – is the degree of the color composition of soils that is visible to an observer. The system is based on five basic colors: red (R-red), yellow (Y-yellow), green (G-green), blue (B-blue), and purple (P-purple). There are also five intermediate hues (YR, GY, BG, PB, RP). Altogether, there are ten hue sectors spaced around the spectral circle. Each sector is further divided into ten parts, but in tables, only their one-quarter wavelengths expressed by a number before each color are stated (2.5 YR, 5 YR, 7.5 YR, 10 YR). A soil scientist can indicate the value between two cards of colors (8 YR, 9 YR).

Value – (Value, light value) – refers to the relative darkness or lightness of the color within the scale 0-10 in relation to the neutral grey scale. The type of difference, that occurs when a black color is added to a white one, is assessed. Extreme values (0 – absolute black, 10 – absolute white) are not mentioned in the tables. Value is the degree of light that enters the eye at normal daylight (5 – grey). Brighter colors are from 5 to 10 and darker colors are from 5 to 0.

Chroma (chroma, intensity of colors) represents the degree of purity of color on a scale of 0 – 20. It indicates the degree of saturation of the neutral grey color by the spectral color. When the intensity of the soil color chroma is assessed, the scale 0 – 8 is used. Value 8 represents the most intensive hue, value 0 expresses a scale of neutral grey colors. Figure 5.19 illustrates the color spectrum of soils, value, and chroma (basic color).

5.8 Water and soil water regime

Water – is a chemical substance composed of hydrogen and oxygen. It is one of the most important requirements for life on the Earth. Under normal temperature and pressure, it is a colorless and transparent liquid with no odor or taste. It is the most widespread substance on the Earth and also the fundamental constituent of the biosphere, the main agent responsible for the transport of nutrients as well as for their intake and excretion. For plants, the total annual amount of water together with its occurrence and division during a vegetation period is essential together with respect to their growth phases.

The water molecule radius is 1.38 Å; in the centre of the molecule, there is an oxygen core. Atoms of hydrogen are bound to oxygen in the shape of V and they form an angle of 105° (Fig. 5.20). It causes that the water molecule is dipolar with positive and negative charges. It results in formation of electric fields with molecular neighbours. Water molecules are bound together by hydrogen bonds (Fig. 5.20). These bonds are weaker than covalent bonds. 18 cm^3 (18 g) of water is composed of 6.02×10^{23} molecules of water. Therefore, 1 cm^3 of water contains 3.3×10^{22} individual water molecules.

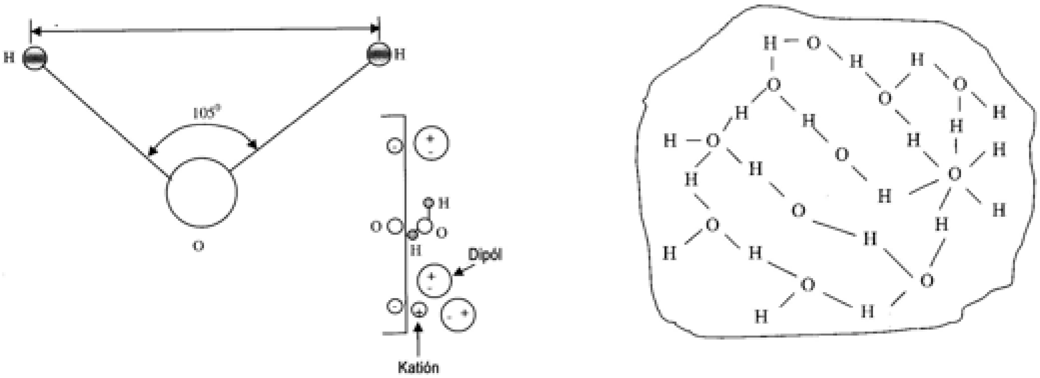


Figure 5.20: Water molecule and hydrogen bonds between molecules (Lal and Shukla, 2004)

5.8.1 Water in soil

The hydrosphere influences the pedosphere a lot and it is comprised of the water in all oceans, rivers, and lakes, groundwater, and soil water. Although there is a lot of water on the Earth and 70% of the Earth´s surface is water-covered, drinking water is very precious. Soil is the main source of fresh water whose volume is 50 times higher than the volume of water in rivers and streams.

Table 5.16: Sources of water (Juma, 1999)

Water sources	Volume (km^3)	% of the total volume
Oceans	1 370 000 000	97,2
Fresh water in lakes	125 000	<0.1
Saline water in lakes	104 000	<0.1
Rivers and streams	1 300	<0.1
Glaciers		
Polar ice caps a glaciers	29 200 000	2.2
Lithosphere		
Soil water	67 000	<0.1
Ground water	8 350 000	0.6
Atmosphere	13 000	<0.1

There are three main categories of water in soil: **gravitational, capillary, and water absorbing**. Soil water is interconnected with soil particles and aggregates. It is in capillary pores, but it also flows from macropores, therefore it is important to classify the soil water and determine its main functions. Classification of soil water is determined by relations of water molecules with mineral and organic surfaces, cations, and anions in the soil.

Soil water potential

Movement and retention of water, its intake, and translocation in plants, as well as the loss of water into the atmosphere, are controlled by energy gradients. Forms of energy that cause it are **potential, kinetic, and electric forms.** Differences in levels of energy determine the direction and speed of water movement in soils and plants. An energy difference between pure water and soil water is determined by the soil water potential. It is comprised of **matrix potential, osmotic potential, and gravitational potential.**

Matrix potential – determines the movement of soil water and availability of water for plants. It is a result of adhesive and cohesive forces and capillarity. When water molecules are bound to solid surfaces, free water energy is reduced. The value of the matrix potential is always negative and it also significantly influences the retention and movement of water in the soil.

Osmotic potential – if soil water contains salt or other soluble substances, the mutual attraction of molecules of water by these substances reduces the potential energy of water. The rate of energy reduction is the so-called osmotic potential. Its value is always negative. It does not influence the movement of water in the soil, but it affects the intake of water by the roots of plants. For example, in salt-land, the value of the osmotic potential is lower than the osmotic potential of the intracellular solution of plant roots. Therefore, plants are not able to take water from the soil. If the plant roots are about to take water from the moist soil, they have to overcome an interaction of all forces at the same time.

Gravitational potential – gravitation plays an important role in the elimination of too much water from the upper part of the root zone by heavy precipitation or irrigation that is, in general, positive. The formula for calculation of the gravitational potential is:

$$\Psi_g = g.h$$

g – gravitational acceleration
h – the maximum water level over the recommended rise

Soil moisture – is the current water content in the soil. It is expressed in % to the weight of dry soil (w), or in volume percentage (Θ). The value of soil moisture is changing over the course of time. Its changes depend on the weather conditions, desuction of plant roots, agrotechnical operations, depth of soil, soil texture, and soil structure. Soil moisture "*w*" is mainly determined in a disturbed soil sample by the gravimetrical method.

Soil moisture expressed in mass percentage refers to the ratio of the weight of water to dry soil and it can be calculated as follows:

$$w = \frac{(m_{kv} - m_{kp}) - m_s}{m_s}$$

m_{kv} – weight of a steel cylinder with the current moisture
m_{kp} – weight of an empty steel cylinder
m_s – weight of dry soil

Soil moisture expressed in volume percentage refers to the ratio of the volume of water to the volume of soil (undisturbed soil) and it can be calculated as follows:

$$\Theta = w.\rho_d \ (\%)$$

w – soil moisture expressed in mass percentage
ρ_d – reduced bulk density

It is essential to be familiar with the regime of soil moisture because of irrigation management. If we want to use the soil water the most efficiently, it is not enough to observe only the amount of water in the soil. It is also important to take into account the use of water and energy by plants in order to absorb water from the soil at some soil moisture, as well as the energy used to hold water in the soil, and the amount of water necessary for optimal moistening of soil in relation to requirements of individual plants.

The most important feature connected with the water-holding capacity of soil is the ability of soil to supply the root systems of plants or organisms that live in the soil by bound water. As the force, which is used to bind water, changes a lot and reaches several thousands of Joules (J) or Pascals (P) at low soil moisture, Schofield was the first one who used the term **absorbing or moisture-retention capacity of soil** and symbol "pF" (pF curve) that is a logarithm of power needed for the suction of water from soil.

The so-called "**retention curve**" is used to characterise the retention force of soil. It is used to express the relation between the soil moisture and its ability to retain water by force. The value pF refers to the average rate of energy that is used by roots of plants to absorb the necessary amount of water from soil at the determined moisture. The retention curve pF characterises the intake of soil water, retention properties of soil and its capillary conductivity.

If we know the values of soil moisture and pF, we are able to answer the question which soil will be drier or moister, and determine the amount of water in % that will be available and used by the root systems of plants, i. e. we will be able to determine the plant **available water contenty** (AWC), which is defined as a difference between determined **field capacity (FC)** and wilting point (WP).

Table 5.17: Soil texture and availability of soil water (Zaujec et al., 2009)

Types of soil	State of soil moisture in %		
	FC	WP	AWC
Clay soil	20	17	3
Loam soil	15	10	5
Sandy loam soil	12	5	7
Sandy soil	10	2	8

Energy categories of soil water

All the water that is present in soil from the lowest to the highest moisture is influenced by the complex of effective forces of different nature, direction and intensity that affect properties of water as well as its movability and use by plants.

When soil moisture changes, one force from the complex of the forces prevails and the activity of other forces is minimal. It is evident that the circumstance will affect the nature of the water influenced by the force, the water holding capacity of the soil, and the movement of water in the soil. From the theoretical and practical point of view, it is advantageous to use these energy relations of water in the soil as criteria for classification and division of the total volume of water in soil into individual energy categories.

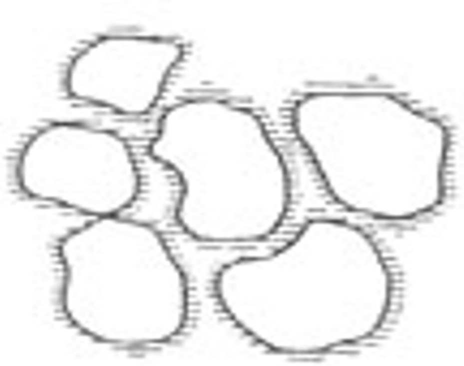
These energy categories are characterized by the interaction of attractive forces, capillary forces, hydrostatic forces, osmotic forces, and pressure forces. It means that energy categories of soil water are determined by intervals of moisture. Within them, the water´s energy relation to the soil is of the same nature, i.e. for a certain category, the domination of forces of one category over other forces of the complex is typical. Nowadays, according to some authors, the most important forces for soil water are **adsorptive forces, capillary forces, and gravitational forces**.

Adsorptive soil water

The surface of the dry soil particles is able to bind molecules of water by using its attractive forces. These forces are called adsorptive forces and the term includes forces of different properties with similar external manifestations.

Adsorptive forces and processes

Atoms and ions of the solid soil phase are bound by the forces of chemical and physical nature that are compensated inside the solid phase but on the surface, they are partially unsaturated. Thus, the force field is created. It attracts molecules or ions of other states that are in contact with the solid soil phase (water, water vapor, gases). It results in adsorption which is created by different forces. If adsorption, which is conditioned by physical forces, is created, it is the so-called physical adsorption. If it is formed by chemical forces, it is the so-called chemical adsorption.



Adsorptive water (Fig. 5.21) is formed by the interaction of electrostatic forces and molecular Van der Waals forces. When molecules of water are adsorbed, on the surface of negatively charged soil particles, dipole molecules of water are influenced by a positive charge. Therefore, when the monomolecular layer is created, the surface does not lose its polar character

Figure 5.21: Adsorptive water (Šimanský, 2013)

When adsorptive forces are not active anymore, the orientation effect of attraction of water molecules stops its activity and molecules are disorderly arranged into free space. Later, the water is influenced by capillary forces and creates the so-called capillary water. The properties of adsorptive water differ from other categories by different density, lower freezing point, missing solubility and ability to move by diffusion of water vapour instead of liquid state.

The water absorbing process of adsorption is presented by the amount of water that has been created by adsorption on the surface of solid particles and it depends on the air moisture of the environment around the sample, size of soil particles, and features of their surface. The value of adsorption is rising together with the rising effect of relative air moisture as well as with the effect of specific surfaces of soil particles. If the conditions of the environment are steady (constant moisture and temperature), the amount of adsorbed water is directly proportional to the soil dispersity. The amount of adsorbed water depends a lot on the composition of the clay fraction, humus content, and quality as well as on the exchangeable cations in the adsorbing complex of soil. If the conditions are steady, the equilibrium state between the air moisture and the amount of adsorbed water in the soil sample occurs.

An adsorption isotherm, which represents a curve created on the basis of obtained values of sample moisture and the environment in the coordinate system, is the graphical interpretation of adsorption (Fig. 5.22). The figure shows that the mutual position of adsorption and desorption isotherms indicates that at the same relative moisture of the environment, the equilibrium moisture of the sample is lower at adsorption than at desorption. Hysteresis occurs during these mutually opposing processes.

The causes of hysteresis can be different, for example, different properties of dry and wet adsorbents, adsorption of air on the surface of the dry sample that makes the adsorption of water more difficult as well as late leakage of closed water clusters, etc. It is important, when relations of equilibrium states of sample moisture and the environment are determined, to mention how they were obtained. The usual way how to determine it is by using water vapour. Adsorption of water vapour in soil has got the character of added substance.

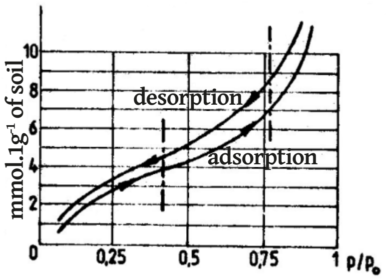


Figure 5.22: Hysteresis of adsorption and desorption izotherm of grain-size fraction <0.002 mm (Sisák et al., 1990)

The influence of clay minerals and exchangeable cations on water adsorption

The amount of water vapor adsorption by clay minerals of individual groups depends mainly on the size of their specific surface and the sequence is: montmorillonite > illite > kaolinite. Exchangeable cations do not suppress the effect of the specific surface on adsorption.

They substantially influence the adsorption isotherms of montmorillonite. On the other hand, the adsorption isotherms of illite and kaolinite are not influenced so much. Univalent cations in comparison with bivalent cations reduce adsorption and a smaller number of molecular layers is created.

In soil, **hygroscopic water** moves in the form of vapor, and the intensity of its movement depends on temperature changes. When the temperature is low, the movement is slower. With the rising temperature, the content of water hygroscopicity in the soil is decreasing. Soil layers in different depths have different temperatures. Therefore, hygroscopic water moves from warmer places to colder ones where it condenses in the form of drops, i.e. it is transformed from the gaseous state into the liquid one. This movement of hygroscopic water occurs mainly in autumn when the lower layers of the soil profile have a higher temperature than the surface layers and as a consequence of it, water vapor can move into the top layers. Because of the day-time and night-time temperature changes, water vapor condenses in the soil as well as on the soil surface. This is the way how dew and white frost are created. Hygroscopic water accounts for the main proportion of moisture in the air-dry soil.

Lentocapillary water – is the second form of the adsorption category of soil water. On the surfaces of solid particles, it creates thin films of liquid water that bind it by their surface attraction of molecules (adsorption). Film water is created from the excess amount of hygroscopic water and it represents the temporary form of water that is strongly bound by soil particles and free capillary water. In contrast to hygroscopic water which can move only in the form of water vapor, film water moves slowly as the liquid water. Therefore, it is also referred to as moving or lentocapillary water. Since it is not under gravitation, it can move in all directions because of the influence of molecular forces and it moves from soil particles with the thicker film to particles with the thinner film of water. Fixation of film water depends on graininess; larger soil particles hold a thicker film of water. In spite of the difficult movability and lower force of fixation, film water in not an accessible source of water for plants.

Capillary soil water – when molecules of water in soil are out of the influence of adsorptive forces, the change in the complex of interactive forces occurs. Adsorptive forces are gradually replaced by capillary forces whose formation and existence are connected with unusual ratios of molecular forces on the borderline of solid, liquid, and gaseous phases (Fig. 5.23).

If the weight of water is too low, the influence of Earth´s gravitational forces is, in comparison to the capillary forces, insignificant and capillary forces prevail, then the water under the influence of these forces is called capillary water. An energy category of capillary water is determined by the interval of moisture in which, from the whole complex of interactive forces, capillary forces mainly prevail.

Figure 5.23: Capillary water (Šimanský, 2013)

They are the result of the combination of cohesive forces between molecules of water (surface tension and surface pressure) with adhesive forces between molecules of water and molecules of other substances. They become evident by the curvature of water into menisci and therefore, they are sometimes called meniscus forces. The granulometric composition of soil significantly influences the velocity and height of capillary action. Capillary water rises the most slowly (by capillary action) in soils with the predominant content of clay particles, but the capillary rise is the highest.

Capillary-linked water – is the water retained by capillary pores in the surface soil horizons for a longer period of time after irrigation by atmospheric precipitation or by irrigation. This water bears no hydraulic or capillary relation to the groundwater. The capillary-linked water is kept inside the capillaries by the central meniscus forces. Between the horizon of the capillary-linked water and the groundwater, there is an interbed of different weights with relatively low moisture. The interbed is called a dead soil horizon (Fig. 5.24).

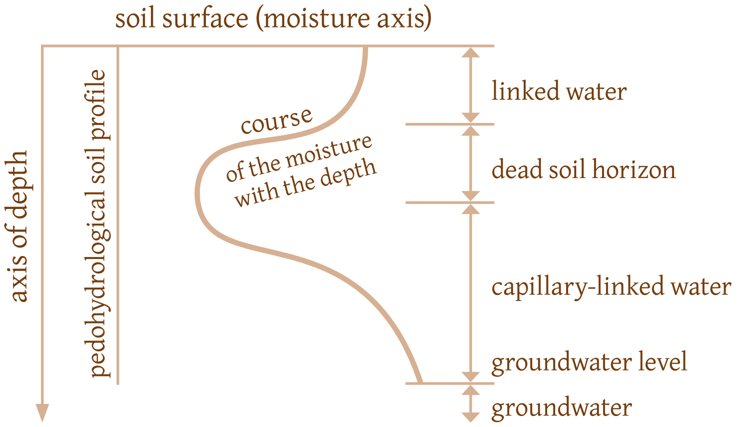


Figure 5.24: Formation of soil water (Sisák et al., 1990)

The performance and volume of capillary-linked water depend on the soil graininess and structure. The low movability is typical for capillary-linked water. It is well known that the lower the degree of structure evolution is, the higher the number of pores filled by capillary-linked water is. Capillary-linked water is a result of the soil´s ability to keep a certain amount of water against the activity of gravitation forces.

Capillary-linked water vanishes by vapor because capillarity moves it to surface layers into the atmosphere. This loss is not substituted by the capillary action from the ground-water level because the joining of the capillary-linked water and the ground-water level is interrupted. Capillary-linked water in soil after irrigation is available for the needs of plants. Water is absorbed by the roots of plants in the whole root zone. Moreover, water is brought up into capillaries from the areas with higher soil moisture as well as from the soil layer which is irrigated, and then, it moves below the root zone. Fine roots of plants are also able to find water. The soil with no vegetation loses water by evaporation. As a result, the amount of water on the soil surface is lowered and subsequently, it is replaced by capillary action from the deeper layers. Then, the moisture availability declines equally in the whole layer of the capillary-linked water. As soon as the state of moisture near the lento-capillary point is reached, the capillary movement is decreased dramatically. Another decrease of moisture occurs gradually in layers towards the soil depth. Evaporation from soil is also caused by gradual transportation of water pores into the soil surface.

The properties of topsoil, mainly its structure, influence significantly the proportion of linked water in soil. A well-developed and stabilized crumb structure of soil determines the sufficient amount of non-capillary pores between aggregates, increases filtration, and the zone of capillary-linked water is made deeper. Methods and intensity of irrigation cause that water only flows through non-capillary pores in the structural layer and when it gets into aggregates, it is gradually saturated. The surface layer with a good structure decreases the water losses by evaporation significantly. Subsequently, the connection between capillary pores is broken, it exists only on the links between aggregates, and as a result, the capillary movement of water towards the evaporating surface is reduced.

Capillary rising water – it occurs in capillary pores above the level of groundwater when capillary water moves from the surface into the soil and is linked with capillary water at the groud-water level. The level of the capillary rise depends on the soil texture and soil structure.

1. **Closed capillary water** – is the water with capillaries filled by water with no air bubbles (e.g. above the groundwater level),
2. **Open capillary water** – is the water in which parts filled by water and those with air bubbles exist together.

The characteristics of the capillary category of soil water indicate that its properties are not homogeneous. It does not move in its liquid state at lower moisture. When the total soil moisture rises, the soil water in its liquid state becomes moveable.

The proportion of capillary-linked water in the homogeneous soil profile is influenced by the number of capillary, semi-capillary and non-capillary pores.

Non-capillary pores – they enable the inflow of water into the soil and its penetration into the depth. When they prevail, the soil moisture in the surface layers does not rise and the speed of water seeping through non-capillary pores is so high that the soil does not become moist enough and water disappears towards the depth as well as away from plant roots.

Capillary pores – the flow of water is slowed down and as a result, the soil is made moist easily. Capillary pores are important because they are able to retain the soil water. When capillary pores prevail, the infiltration of water is getting worse, the intake of water by soil is not sufficient, the soil becomes moist towards the lower depth, rainfall is not used sufficiently, the overland flow is rising and the risk of erosion occurs.

Semi-capillary pores – they enable the sufficient retention of water resulting from the influence of those capillary forces whose activity is against the direction of gravity. Instability of soil water is caused by the high number of semi-capillary pores. **Semi-cappilary linked water** is the transition to another energy category – the so-called gravitational category.

Gravitational water

A gravitational force is the one which dominates among other forces which are active in this energy category. When the soil moisture rises, gravitational forces prevail over capillary forces and therefore, it is the main and determining attribute of this energy category of soil water (Fig. 5.25). The use of the gravitational force is dependent on the properties of soil as well as on the quality and number of pores.

Gravitational water flows freely **in the direction of the action of the gravitational force in non-capillary pores**. The precipitation and surface water are a source of water in soil. The movement of water is dependent on the number and size of non-capillary pores as well as on permeability.

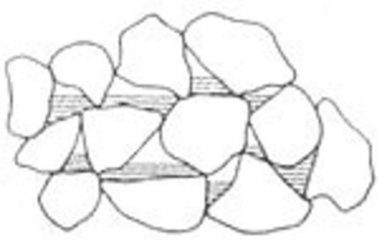


Figure 5.25: Gravitational water (Šimanský, 2013)

The occurrence of gravitational water in the soil profile after precipitation and irrigation is temporary and short-term. Water drains quickly due to the action of gravitational forces but on the other hand, it drains more slowly in semi-capillary pores. It is not a significant source of water for plants because of its short-term occurrence in the vegetation profile. It is important for soil because it takes away soluble and dispergated substances from the surface layers into the deeper layers of soil. If it hits the insoluble layers, it stagnates and becomes a source of the creation of forms of **gravitational water** as well as **groundwater.**

If the soil is moistened too much by precipitation or irrigation water, all the pores in the surface layer are saturated, gravitational water drains off non-capillary pores and creates space for further infiltration. Water which flows quickly (percolates) in the profile is called **(percolating) gravitational water.** If there are any mechanical barriers, gravitational water is changed into the form of **supported gravitational water**. Supported gravitational water is able to create either a coherent or incoherent layer. If the coherent layer achieves the sufficient depth, it creates the groundwater level in the pit. The **groundwater level** is the level corresponding with the permanent level of water in the pit. In the groundwater level, the total potential of soil water is zero. It means that the level of the groundwater is the equipotential surface which is sometimes irregular. Capillary forces act above this surface and cause the so-called capillary action. But, the water which is in the lower part of the capillary zone is the gravitational water.

Groundwater – is an important soil-forming process. Its depth influences the intensity of its action in the soil-forming process. The depth of groundwater is the vertical distance from the groundwater level to the horizontal layer. Groundwater is formed by the accumulation of liquid water in impermeable water-bearing layers in different depths and under the terrain surface. The depth and the amount of accumulated ground water depend on the geological structure of the area, relief, climate, and plant cover.

Water that is retained in impermeable horizons in the soil profile is the surface groundwater. In geological layers, the so-called subsurface groundwater is formed. Through the process of capillary action, groundwater gets above the level and creates the so-called capillary boundary. If the water-bearing layer is horizontal, groundwater stagnates, and the so-called inactive groundwater is created. If the layer has declivity, groundwater can have the subsurface flow and it is called flowing water.

Groundwater is formed from gravitational water which percolates by infiltration (mainly in areas with more humid climate and excess rainwater) and condensation of water vapor (mainly in dry areas with inadequate precipitation). In some areas where wet and dry seasons change, the groundwater is formed by infiltration of gravitational water and condensation of water vapor.

The amount of groundwater and its level change constantly. It depends on the amount of precipitation, soil permeability, evaporation, desuction, and groundwater flow. A significant decrease of groundwater level is caused by vast forest cover which can take up a lot of water especially from deeper layers. The groundwater reaches the maximum height in spring after melting of snow.

Groundwater which stagnates in the soil profile can influence soil and plants negatively. It can cause the formation of anaerobic and reducing conditions as well as gleization, and reduce the development of root systems of plants. It can serve as an additional source of water only when it is in the favourable depth and is not mineralized a lot. Kopecký (in Fulajtár, 2006) mentions the following values of the critical depth of groundwater for some crops: meadow grassland 0.5 – 0.6 m; field crops 1 – 1.2 m; orchards 1.4 – 1.6 m and hop gardens 1.8 – 2 m.

Water adsorption capacity (Θ_{WAC}, w_{WAC}) – is the maximum volume of water in the soil which is bound by adsorptive forces. It depends on the soil texture, clay minerals, content of cations, number and types of minerals, and quality of humus. It is the boundary between the category of adsorptive water and capillary water.

Water retention capacity (Θ_{WRC}, w_{WRC}) – is the stabilized state of soil moisture after 24 hours of flowing. It is in accordance with the volume of water in the soil which is under the influence of capillary forces, i.e. in capillary pores. Therefore, it is possible to claim that this value is equal to the volume of capillary pores. Its value should be expressed as 2/3 of the total porosity, and it also equals the approximate field water capacity.

$$W_{WRC} = \frac{(m_{kp24} - m_{kp}) - m_s}{m_s} \times 100\ (\%)$$

m_{kp24} –weight of steel cylinder after 24 hours of suction

m_{kp} –weight of empty steel cylinder

m_s –weight of dry soil

conversion into volume percentage

$$\Theta_{WRC} = w_{WRC} \times \rho_d \ (\%)$$

w_{RC} –water retention capacity in mass percentage
ρ_d – reduced bulk density

Saturated soil (Θ_s, w_S) – it characterizes the soil moisture when pores and interstices are fully filled up by water. This value equals porosity.

Wilting moisture – is the soil moisture, when wilting of certain species becomes evident at a certain development stage and under particular atmospheric conditions. Within the humidity of wilting, the humidity of the beginning of wilting and humidity of persistent wilting is determined.

Wilting point – (Θ_{WP}, w_{WP}) is the soil moisture when plants are permanently supplied insufficiently by soil water, therefore they fade and die. Soil water is bound by greater forces and it is inaccessible to plants. With the increased soil moisture, the plant growth will not continue anymore.

Refill point – (Θ_{RP}, w_{RP}) also lento-capillary point – LB is moisture at which the movability of soil capillary water decreases, and thus also its exploitability by plants. There occurs the discontinuity of relation of the capillary water column, which is present only between the contacts of soil particles in the finest capillaries.

Field capacity – (Θ_{FC}, w_{FC}) is the maximal quantity of capillary suspended water, which after the complete saturation soil is able to retain in capillaries in the whole profile for a longer time with the exclusion of the impact of precipitation, evaporation, and capillary inflow of groundwater.

Capillary water absorption (Θ_{CWA}, w_{CWA}) – characterizes the maximal permeation of soil pores along with the capillary saturation of soil by water within 24 hours.

$$W_{CWA} = \frac{(m_{kn24} - m_{kp}) - m_s}{m_s} \times 100 \ (\%)$$

m_{kn24} – weight of steel cylinder after 24 hours of absorption
m_{kp} – weight of empty steel cylinder
m_s – weight of dry soil

convertion to volume percentage

$$\Theta_{CWA} = w_{CWA} \cdot \rho_d \ (\%)$$

w_{KN} – capillary absorbability in weight percents
ρ_d – bulk density reduced

Maximal capillary water capacity (Θ_{MCWC}, w_{MCWC}) – the soil ability to retain the maximal quantity of water suspended in capillary pores for the purpose of vegetation after 2 hours of effluxion after the completion of saturation.

$$W_{MCWC} = \frac{(m_{kp2} - m_{kp}) - m_s}{m_s} \times 100\ (\%)$$

m_{kp2} – weight of steel cylinder after 24 hours of exhaustion
m_{kp} – weight of empty steel cylinder
m_s – weight of dry soil

convertion to volume percentage

$$\Theta_{MCWC} = w_{MCWC} \cdot \rho_d\ (\%)$$

w_{MCWC} – maximal capillary water capacity in weight percentage
ρ_d – bulk density reduced

5.8.2 Soil water regime

The water regime is the whole complex of phenomena related to the hydrologic cycle between the atmosphere, hydrosphere, and the top layer of the lithosphere. The water regime comprises the water balance of the soil. It involves water intake, retention, and leakage during a certain period in a geographical area. It consists of the moisture regime and average longtime water balance. In the process of characterizing the soil water regime, the water conditions are detected in the soil profile and in the subsoil to the depth of their hydrologic relation. The impact of the amelioration measures is taken into consideration in the reclaimed soils.

All components are being evaluated for the same balance period and their values are expressed v mm of the water column, similarly to precipitation. The soil water balancing is carried out by the balance equation:

$$\boldsymbol{Z_z + S + P_{pv} + P_{pz} + K = E + T + O_{pv} + O_{pz} + Z_k}$$

Z_z – water supply in soil at beginning of balance period ,
S – total precipitation (including irrigation),
P_{pv} – surface water inflow from foreign territory,
P_{pz} – underground inlow (including capillary rise from surface of underground water),
K – condensation of water vapor,
E – evaporation (soil vapor),
T – transpiration (plant vapor),
O_{pv} – surface runoff (precipitation rate prevails over infiltration rate),
O_{pz} –underground runoff,
Z_k – water supply in soil at the end of balance period.

The adjustment of the water regime is manifested by the change in air, thermal, microbiological, and nutritive soil regimes. The principal measure for the adjustment of the water regime of the waterlogged soils is drainage, with air-dry soil it is irrigation. At the same time, it is necessary to use also other agrotechnical principles.

5.9 Warmth and thermal soil regime

The movement of heat from the inside of the Earth is broken by the loss of heat from the surface by radiation. For example, in the soil depth of 0.9 – 1.2 m at the equator, the temperature is the same as in the depth of 15 m in New York (Shukla, 2014). Soil is not heated from the decomposition of organic matter and the heat, which is released in this way in the form of gases, is not measurable. The most significant source of heat is the sun which allows plant growth and development. The regular radiation of the sun is influenced by the atmospheric shell of the Earth. Without this impact, the sun heat could be intensive during the day and at night all processes would be destroyed rapidly by frost.

Along with the falling temperature, the life processes are being decelerated and even stopped. The growth and development of the majority of cultivated crops are very slow at the temperature of about 4 °C and with the rising temperature in the interval 21 – 32 °C the life processes are increased (Fulajtár, 2006). The chemical processes and activity of microorganisms, which make the nutrients available into accessible forms, are also affected just by the temperature.

The soil temperature is the factor needed for the existence of life of plants, animals, and predominantly soil organisms. It has an impact on evaporation, soil moisture, and air movement and it is also the condition of many physical, chemical, and physicochemical reactions.

Soil acts as a transformer because it changes the sun's energy into thermal energy. At the same time, it is the accumulator of this energy and regulator of the thermal regime of the ground layers of the atmosphere. The temperature value is affected by the quantity of accepted energy, its bonding, and thermal conductivity.

The heat adsorption by the soil surface depends on the height position, relief, slope exposure, vegetation, soil structure, porosity, moisture, evaporation, thermal capacity, and conductivity of the individual soil components. The vegetation and also unprocessed post-harvest residues on the soil surface act insulating because they limit the direct heating and heat radiation from soil. The dark and lumpy surface absobes more heat than the light and smooth surface.

Soil thermal capacity – expresses the ability of soil to accept and retain heat. It depends on the thermal capacity of the solid, liquid, and gaseous soil components. The water content has the most significant impact on the soil temperature, as it has the highest specific heat, which means the heat quantity needed for heating of 1 g by 1 °C. Water has a specific heat of 4.19 J, the solid particles are only 2.09 – 2.51 J (Zaujec et al., 2009), and air has the lowest value of the specific heat. The higher the water content in the soil, the higher its thermal capacity. Heavy and moist soils are cold, they are heated slowly but cooled also slowly. The temperature increases by the evaporation of water. The slight warmth of moist soils detains considerably spring work and limits microbial activity. The result is the accumulation of more acid raw humus.

Soil thermal conductivity – is the ability of the soil components to conduct the heat from the surface to the deeper layers. It is measured by the heat quantity transilient through the area 1 $mm^2.s^{-1}$ at a temperature gradient of 1 °C into the depth of 10 mm. Water has the relatively highest conductivity in soil (30-times higher than air) and air has the lowest conductivity

(Fulajtár, 2006). The conductivity of solid particles depends on their size and origin. The mineral particles have better conductivity than humus.

A higher thermal capacity and weaker soil conductivity compared with the atmosphere cause the accepted warmth penetrates into a relatively small depth, whereas the temperature development in soil lags behind the atmosphere. The slow penetration of heat into deeper layers leads to the delay of both maximal and minimal values. The retardation is evident in the daytime and night fluctuations of temperature. In our soils, the daytime fluctuations of temperature reach the depth of 0.75 – 1 m, monthly about 5 m, and annually approximately 6 – 20 m. Soil freezing becomes evident mostly into the depth of 1 m, and the high temperatures penetrate into the depth of 1.5 m. Therefore, the drainage pipes should be laid deeper than 1 m so that they are not damaged by frost and the water pipelines should be placed deeper than 1.5 m not to heat water in summer (Bedrna et al., 1989; Zaujec et al., 2009).

The temperature fluctuations in the different depths of soil have a considerable impact on water vapor and its condensation. Water vapor rising from warmer bottom layers to cooler surface layers condensate and form soil dew. Water vapor descending from warmer surface layers to bottom cooler layers can create underground water after condensation.

Soil as a heat accumulator retains temperature in autumn and it is warmer than the atmosphere. On the contrary, in spring soil is heated more slowly and it is colder than the atmosphere. In autumn and winter, the soil releases gradually the heat accumulated in the summer period, whereby it warms up the ground layers of the atmosphere and fosters the development of winter crops. In spring the soil cools the atmosphere and thus prevents premature vegetation. Soil is the coldest in the deeper layers in March.

The soil temperature is measured in the depths of 0.05, 0.2, 0.5, and 1 m by the earth thermometers, or more suitable temperature recorders, electrical thermometers, or semiconductor thermometers, which are used for specific purposes. The development of temperature is monitored most frequently along with the evapotranspiration in the lysimeter experiments.

5.9.1 Regulation of temperature in soil

The methods of regulation of soil temperature are one of the most important tasks of agriculture. The most effective way, to have an impact on the soil temperature regime, is mulching or leaving the crop residues on the soil surface (Fig. 5.26). The following materials are being used: straw, foliage, peat, sand, different colors, the latex of foil, and even stones in some vineyards in France (White, 2009). Mulching changes the conditions of exchange of warmth and air between soil and the ground layer of the atmosphere. It has also a positive effect on plant growth and development.

In practice, **loosening and solidification** of topsoil are used. **Soil loosening** leads to increased porosity and decreased thermal conductivity, and the temperature in the topsoil is increased. **Soil solidification** decreases soil porosity and temperature. The positive effect on the increase of soil temperature has also bedding, where the soil surface area is being increased. **Draining of soil** accelerates its heating by 2 – 3 °C. The decline of moisture decreases also the thermal conductivity.

Figure 5.26: Bedding and leaving crop residues on soil surface (Šimanský, 2013)

5.10 Air and soil air regime

Soil air, the soil gas phase, takes up the free space in soil, which is not filled with water, or it is closed in the form of bubbles in the capillary pores. Soil air along with water is a component of soil. It gets into the soil from the atmosphere as a result of the continual changes in air temperature and barometric pressure. The intake of air in soil underlies its porosity, moisture, and water movement.

The important factor is the soil permeability of air, i.e. the quantity of air that passes by the area of 1 cm^2 and depth of 1 mm in the time unit. The air quantity in the soil is changed along with the increased pores, lightweight by organic matters, sand, and loosening. On the other hand, it is decreased by soil lying down, compaction, and mechanical clashes of big rain drops. The air content in soil can be dropped even to zero. Soil shrinkage, roots, and microedaphone increase the porosity, thus, the air content rises in the soil.

The exchange of soil gases with the atmosphere (Fig. 5.27) occurs by diffusion, whereas the air temperature and barometric pressure play an important role. Rainwater penetrates into the soil, it either displaces gases or absorbs them. The soil permeability for gases decreases with the growing depth and rising moisture. However, it increases with the growing porosity. The compact surface layer has a significant impact on the air regime, e.g. soil crust, which decelerates the permeability and soil aeration by diffusion.

The composition of soil air differs from the atmospheric air by lower content of oxygen, higher content of CO_2, and high content of water vapor. Nitrogen content in soil air is approximately the same as in the atmosphere. In soil air the different gases containing nitrogen, sulfur and methane occur.

Oxygen – is the essential source of breathing of plant roots, aerobic microorganisms, and soil fauna, that intake it from the soil air. This continual consumption of oxygen in the soil causes its content in soil air to be lower compared with the atmosphere. Oxygen is a significant factor in the oxidation of organic and mineral substances in soil. Oxygen in soil air occurs most frequently in the interval of 10 – 21%. In the surface aerated layer its content is in the interval 18 – 21%, it declines gradually deeper in profile and it achieves very low values in the long-term waterlogged soils. The source of oxygen in soil is also rainwater, containing soluble oxygen. E.g. 10 mm of precipitation per 1 ha, i.e. 100 m^3 of water, contains cca 4.34 kg of soluble oxygen at the temperature of 20 °C. It is equal to 3,000 l of pure oxygen that gets into the soil under atmospheric pressure (Singe, 1996). Therefore, rainwater has a considerably more positive impact on plant growth than irrigation.

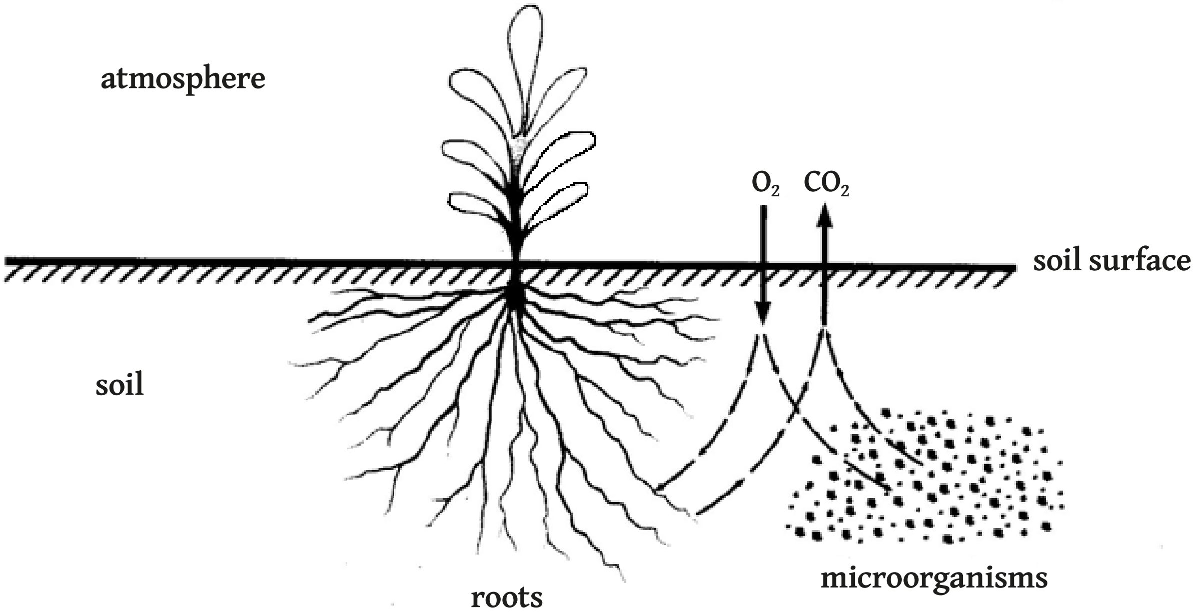


Figure 5.27: Exchange of gases between soil medium and atmosphere (Fulajtár, 2006)

The normal growth of the majority of plants occurs at the concentration of O_2 in soil air above 10%. A lower concentration of O_2 causes the retardation of root growth and development, a low quantity of intaken nutrients, and lower yields. When the oxygen content **decreases below 2%**, the root growth stops. As a rule, well-aerated soils have more stable soil structure than slightly aerated soils. The oxygen insufficiency entails the reductive conditions, when methane, hydrosulphide, ammonia, free nitrogen and ferreous compounds are formed, which are toxic for plants. The excessive content of oxygen is also adverse, it accelerates the mineralization of soil organic matter and fast soil drying.

Carbon oxide – occurs in the soil air on average 0.3%, which is about 10-times more than in the atmosphere. However, the content of CO_2 in soil fluctuates in a broader interval, in the top layer 0.1 – 1%, in lower subsoils 0.3 – 5%. In the grass-covered soils around 1.5% of CO_2 is present in the soil air and about 0.5% in the tilled soils (Fulajtár, 2006). The content of CO_2 in soil fluctuates depending on the season and temperature of the atmosphere. It is the indicator of the intensity of soil biological activity. The source of CO_2 in the soil is its exhalation by plant roots and microorganisms. It is also created by the decomposition of organic matter under aerobic conditions. Carbon oxide with water forms weak carbonic acid that dissolves the mineral compounds in soil, releases nutrients, dissolves and makes them available for plants. Its higher content in the soil air enriches the ground air with CO_2 during its exchange with the atmospheric air, where the plants can find more favorable conditions for its assimilation.

However, the increased content of CO_2 in the soil is harmful to plants. It derogates and even damages the root system. The reason for its high content is the weak exchange of air for the atmospheric air, whereby CO_2 in the soil is accumulated and its content increases to 10 – 15%. This process occurs mostly in unstructured and compact soils with a closed soil surface (soil crust).

Water vapor – soil air compared with atmospheric air has high relative humidity. The relative humidity constitutes often 100%. Therefore, the soil is moisturized with the temperature drop, and dew or hoarfrost is formed.

Nitrogen – nitrogen in the soil air is identical to its content in the atmosphere, i.e. its content in soil is 78% (Kohnke, 1968). It occurs in the form of N_2 and NH_3 in the soil air. It is important mainly as the biochemical agent because the nitrogen-fixing bacteria bond N_2 and nitrifying bacteria oxidate NH_3.

5.10.1 Air content in soil

Based on the known values of moisture and porosity in the volume percentage, we can calculate:

Current air content (C_{AM}), which represents the air volume at the time of sampling of the intact sample with the current moisture. We calculate it in the following way:

$$C_{AM} = P - \Theta\ (\%)$$

P – total porosity

Θ – current water content in soil in volume%

Minimal air capacity (C_{AC}), which means the volume of pores filled with air after the stabilization of saturated soil sample to the maximal capillary water capacity. From the practical viewpoint, it is the lowest quantity of air in soil that is needed for the biological processes of plants and microorganisms. We can calculate:

$$C_{AC} = P - \Theta_{MCWC}$$

P – total porosity

Θ_{MCWC} – maximal capillary water capacity in volume%

5.11 Technological soil properties

5.11.1 Soil consistency

The soil consistency is evident as a set of physical strengths (adhesive, cohesive) of soil in a certain interval of soil moisture. Thus, it involves the properties of soil material that are induced on one hand by bonding of soil particles mutually between each other, referred to as cohesiveness –

soil cohesion, on the other hand, sticking – soil adhesion to foreign objects. The soil consistency comprises the following attributes: friability, cultivation, plasticity, resistance to compaction.

5.11.2 Soil cohesion

Soil cohesion is an important technological property of soil, that is evident in the process of soil cultivation. It affects the root penetration and conditions the changes in the soil structure, especially in wet soil. It is the demonstration of the internal portative forces of soil particles that cause the soil particles and structural aggregates to hold together, and that the soil resists again its disconnection and deformation during the intrusion of foreign objects. With a low water content, the water covers are formed around the contact areas of soil particles. The water covers are associated with these particles. Therefore, cohesion is the function of the number and thickness of water covers around the particles. As the number of covers depends on the number of particles, the portative forces in soil with colloidal clay are much higher than in sandy soils. The lamelliform particles create higher cohesion than spherical ones. The highest cohesion occurs in soils with high content of colloidal particles of mineral and organic origin. On the other hand, the lowest cohesion is in light soils with thicker grains of sand, which are poor in colloids.

The soil cohesion is evident if the dry fine earth is moisturized by water and mixed carefully. In this way, the structural aggregates are formed, sometimes of big diameters. After drying these aggregates are disintegrated because they were bound by water. However, if the particles were agglutinated by the bonding solutions, the structural aggregates are formed, which do not disintegrate and are resisting also to the disturbing effects of water. The cohesive soils require increased energy during the crumbling of dry topsoils. The cohesion can be improved by liming and fertilizing by the organic matter.

Mutual relations exist between the soil cohesion, bulk density, compactness, and cohesion of aggregates. For example, higher values of bulk density lead to higher soil cohesion, which is reflected by the increase of the penetrometer resistance. White (2009) claims that the ideal values of soil cohesion should be in the interval 1 – 2 Mpa, that are the values when the plant roots penetrate easily into the soil medium.

5.11.3 Soil strength

The water content is important in clay soils. The drier soil is, the more cohesive it is, thus the cohesion is transformed into strength or soil hardness. Soil strength originates in the increased internal friction between the soil particles. It is demonstrated as the soil resistance to the mechanical effects which try to disintegrate bigger soil aggregates into small particles. The soil strength is detected in a laboratory by measuring the power needed for the crush of the structural elements with a diameter of 50 mm and it is expressed in kg. The following strength values are given for the particular soils: sandy soils 1.2 kg, loamy sand 2.1 kg, loamy soil 3.4 kg, clay loam 4.6 kg, and clays 6.3 kg.

5.11.4 Soil plasticity

The increase of film water between the soil particles separates them, the portative forces are decreased, and thus also the cohesion. With certain moisture the soil becomes plastic, and it can be formed into some consistent forms. Further water acts as a lubricant, it releases gradually the soil particles and allows their movement.

Therefore, soil plasticity is demonstrated along with the gradual rise of soil moisture and the changes in the soil consistency from friable to moderate even plastic. Thus, plasticity is defined as the soil ability to change its form without the creation of cracks. Soil plasticity depends on the content of clay in the soil. Sandy and coarse-grained soils are not plastic. Plastic soils can be formed only when wet and it is not possible when drying. The tension needed for the creation of the specific degree of deformation is proportionally as big as the cohesive forces, which hold the soil particles together. The cohesive forces depend on the clay properties and the degree of soil moisture or the thickness of the water cover.

5.11.5 Penetrometric resistance

The penetrometer resistance develops during the dissociation of soil particles by the penetrometer needle when it penetrates into the soil. The penetrometer resistance or also the index of the cone resistance is the empirical measure of soil strength. It is expressed in kg.cm^{-2}.

The factors, which influence significantly the penetration resistance, can be divided into the soil: (moisture, soil density, porosity, and strength) and the properties of the penetrometer: (base diameter, apex angle, surface cone roughness, length, and rod diameter). In order to measure the penetrometric resistance, the penetrometers are used. The analysis of the result of penetration measurements is very important for the acquisition of the correct data, allowing to understand the complex processes in the soil medium and express them simply physically. The critical values of the penetrometric resistance are indicated in Table 5.18.

Table 5.18: Critical values of soil physical properties identifying strong compaction in soil profile (Fulajtár, 2006)

Parameter	Soil type					
	p	hp	ph	h	Ih-iv	i
Key parameters						
Bulk density	>1.70	>1.60	>1.55	>1.45	>1.40	>1.35
Porosity	<38	<40	<42	<45	<47	<48
Penetrometric resistance MPa	6	5.5	4.5-5	3.7-4.2	3.2-3.8	2.8-3.2
Soil moisture	10	12	13-15	18-16	24-20	28-24
Min. air capacity	<10	<10	<10	<10	<10	<10
Auxiliary parameters						
Field capacity				>35	>35	>35
Plasticity index				>25	>25	>25
Clay content					>30	>30

5.11.6 Plowing resistance

The plowing resistance occurs during cutting, elevating, crumbling, and turning of the furrow dump. The plowing resistance (Pr) depends on the specific soil resistance (K), which constitutes the ratio between the quantity of energy needed for cutting and turning of soil slice and on the friction of the work surface of the tool, next on the depth (a) and width of cut (b). It is calculated:

$$Pr = K.a.b$$

In fact, the need for traction and capacity is being determined. It is significant for the utilization of towing devices and plows, so-called the output at hanging hook, i.e. the product of traction force and gradual speed. The value of the plowing resistance correlates closely with the soil texture, the content of humus, the soil type, and the depth of plowing.

5.11.7 Soil compaction

The soil compaction is the process when the elementary particles come near to each other, and thus they affect significantly the soil properties. Nowadays, soil compaction constitutes a very serious problem in Slovakia and all developed countries with mechanized agriculture. As a result of compaction, the productive, as well as unproductive soil functions, are decreased considerably. The reasons for compaction can be **natural and anthropogenic:**

- **natural** – high content of clay particles, clay minerals, ilimerization, heavy rains, the pressure of the upper soil layers on the lower ones, lack of organic matter and humus in the soil, waterlogging,
- **anthropogenic –** these can have biological, chemical, and mechanical nature:
  - **biological** – insufficient uptake of organic matter into the soil, inappropriate structure of crops, monoculture, poor breeding technologies of beef cattle in terms of manure production,
  - **chemical** – unsuitable composition of industrial fertilizers (the content of monovalent cations (Na – peptization of colloids), insufficient liming,
  - **mechanical** – high specific pressure of machines on the soil, unsuitable composition, and type of tools, unnecessary driving over the field, driving of wheeled tractors in the furrow, working in unsuitable moisture, insufficient plowing depth.

Proposal of regulatory soil protection measures from the results of soil monitoring in the Slovak Republic

Technical:

1. Reduction of pressure on the soil (reducing the load of mobile agricultural machinery, double assemblies, low-pressure tires, tracked agricultural machinery mechanisms, axle load – distribution of load on several axles, drawn agricultural machines should be preferred over carried ones) – limit of 6 t per axle or 150 kPa, or 8 t per tandem, higher pressure reaches a depth of 0.4 m.
2. Application of immobile agricultural machinery.

Organizational:

1. Entering the field only when the soil has adequate moisture (exclusion of crossings within the excessive soil moisture) – limit: 80% of SWC (soil water capacity) for medium and heavy soils, 90% of SWC for sandy soils.
2. Exclusion or at least limitation of the transport on the field (separation of field and road transport).
3. Limiting of crossing the field by aggregating operations.
4. Application of agricultural machinery with controlled crossings across the field (track intermediate rows).
5. Optimal use of the power of agricultural machines (slippage occurs when overloaded, otherwise there is usually an unnecessary load on the soil by the towing agricultural machines).
6. Optimal speed (at low speed, higher compression effect occurs, at high speed there is its multiplication by vibrations).

Agrotechnical:

1. Suitable crop rotations (sowing structure, compliance with the principles of crop rotation, sufficient representation of structure-forming crops).
2. Increasing soil resistance to compaction (sufficient organic fertilization, green fertilization, rational liming).

Soil protective:

1. Deep mechanical loosening (including preparation of the land – selection of a suitable preceding crop that is taken out of the land in time, choice of a suitable loosening tool, compliance with technological discipline at suitable soil moisture of 25-30%),
2. Subsequent fertilization measures stabilize the effect of deep loosening:
   - subsequent crossings can be delayed until spring, the shortest interval is 3 weeks,
   - some soil processing operations can be omitted thanks to deep loosening,
   - differentiated adjustment of the crop rotation with the choice of deeply rooted structure-forming crops,
   - differentiated fertilization (liming as needed, choice of non-acidic and non-peptizing fertilizers, profile application of fertilizers as needed, supporting a biological revival of deeper parts of the soil profile).

5.11.8 Plow pan

A plow pan is created on weakly structured soils by systematic pressure of tilling tools to the same depth as with the use of plows with a blunt, worn share. The result of such plowing is the formation of a strongly compacted, impermeable soil layer, several centimeters thick, immediately below the topsoil. The compaction of the plow pan (layer) is increased by the presence of high content of clay particles leached from the topsoil. The plow pan worsens the porosity of the topsoil, aeration, water regime, as well as penetration of the root system into the soil. We can limit the formation of the plow pan by good plowing with chiselling and alternating the plowing depth.

The degree of the plow pan development is identified by the difference in porosity in the plow pan layer and in the following soil layer. If the difference is 0-3%, the layer is poorly developed, if it is 3-6%, it is moderately developed, and if it is 6-9%, it is strongly developed.

5.11.9 Soil permeability and density

As the seasons change, the soils are gradually getting compact. The result is soil permeability, which expresses the density of the natural deposition of mechanical particles.

Soil moisture, its soil texture and structure as well as the season and the vegetation largely determine the value of soil permeability. It is therefore a true technological value, where several soil components are manifested together. For example, humus and calcium promote the improvement of the structure, the mellowness of the soil is improved, and thus the soil premeability is reduced. On the other hand, Fe, Mn, Al, Na, and clay colloids increase soil permeability. The mentioned elements are sometimes more intense than the soil texture itself. This influence is manifested differently in the permeability of the same soil types. For example, medium-grained (loamy) soils without Ca with low humus, but with the above-mentioned elements, have a greater tendency to density. Concerning natural factors, long-lasting heavy rain affects soil permeability. Due to the influence of higher rainfall with a tendency to waterlogging, the compaction of topsoil increases considerably, especially on soils with greater cohesion. Therefore, it is necessary to loosen the soil in spring and during the growing period concerning the protection against year-round gradual compaction of the soil, which is higher in a more humid climate compared to an arid one.

5.11.10 Bearing capacity of the soil

The term "bearing capacity of the soil", can be understood as the load of the soil, at which the soil begins to push out due to the applied pressure. The pressure that exceeds the bearing limit will permanently push the soil to the sides and the object will sink into the soil. The bearing capacity depends on the soil texture, cohesion, and, above all, the soil moisture. Therefore, moist soils have a significantly lower bearing capacity than dry soils."

When the load is applied, the soil is first compacted, but since this compaction acts not only in the vertical but also in the horizontal direction, the soil moves sideways. If the load continues to increase, the soil is compacted under pressure until a kind of "wedge" is formed, which is prevented from further compaction by the neighboring soil with its passive resistance. With further loading up to the limit state of bearing capacity, the slip surfaces are formed, the entire mass of the soil is pushed out and the machine gets stuck and sinks into the soil.

5.11.11 Soil adhesion (stickiness)

Soil adhesion is its ability to stick to foreign bodies. It is determined by the weight of the load needed to pull the soil sample adhered by the given surface onto the metal plate (g.cm^{-2}). Likewise, soil cohesion, the degree of soil stickiness depends on soil texture, moisture, soil structure, and exchangeable cations. Soil texture affects adhesion primarily by the size of the active surface. Therefore, clay soils have the highest stickiness (400 g m^{-2}) and sandy soils have low stickiness (30-50 g.m^{-2}). Dry soil does not stick. When increasing the moisture up to a certain limit, the stickiness increases, after exceeding this maximum, it drops again quickly. The soil structure and cations of calcium reduce the stickiness. On the other hand, cations of sodium and alkaline salts increase the stickiness.

Stickiness together with cohesion is a very important technological property of soil. They are decisive factors in assessing the condition of the soil for its cultivation. For example, by plowing wet soil, a thick layer (0.1 – 0.15 m) of soil sticks to the plow. The plowing is of poor quality because the plow does not turn the topsoil, it only pushes and compacts it. By plowing heavy soils within an increased moisture, non-disintegrating mounds are formed, which harden after drying and are difficult to process. Clay soils stick more to wooden objects, while lighter and humus soils stick more to metal objects. Poor-quality plowing results in damaging the structure, porosity, and other technical properties of the soil and its regimes, which leads to reduced production of cultivated plants.

5.11.12 Soil viscosity

As soil moisture increases, its consistency changes from plastic to sticky up even viscous. If it is already viscous, the soil spreads under pressure, and spreading is relatively dependent on the applied pressure on the soil. If the soil is plastic, a certain amount of force must be applied before the soil begins to spread. Behavior within the spreading of soil is expressed by Bingham's equation:

$$V = k.\mu(F - F')$$

V – volume of flowing soil
μ – mobility coefficient
F – applied force
F' – the force necessary to overcome cohesive forces or F' is zero and the volume of flowing is proportional to the force (V and F)
k – coefficient of viscosity of liquids

5.11.13 Swelling and shrinkage of soil

Soil particles are covered with water on the surface (hydrated) and thus separate from each other, while the soil increases its volume – it swells. By reducing the soil moisture due to the temperature, water covers on soil particles become thinner and soil particles become closer together reducing the volume of the soil, and causing the soil to shrink.

The swelling and shrinkage of the soil are affected by soil texture. Significant volume changes take place with increased content of clay and colloidal particles. The alternation of soil swelling and shrinking processes is very beneficial for the formation of soil aggregates, when a favorable spatial arrangement of the soil matter is created, especially in loamy soils.

The swelling of the soil itself is mainly an unfavorable process because the total porosity is reduced, especially the non-capillary one, and thus the permeability for water and air is reduced. The intensity of the soil swelling increases with the content of the clay fraction. Concerning clay minerals, montmorillonite swells the most, beidellite, vermiculite, halloysite, illite, and kaolinite less. Exchangeable cations also affect the swelling process. In general, monovalent cations promote swelling more than divalent cations.

The shrinkage of soil is expressed in % to the swollen volume of the soil. It is mainly present in dry periods. Cracks up to 0.15 m wide and 1 – 3 m deep are often formed. Shrinkage has both positive and negative effects.

Table 5.19: The volume changes of soils due to swelling (Fulajtár, 2006)

Class of soil	Maximum volume change in%		Type of clay mineral	Maximum volume change in %
Loamy sand	<10		Montmorillonite	>100
Sandy loam	5-20		Beidellite	30-300
Loamy	15-25		Illite	10-40
Clay loam	20-30		Kaolinite	<20
Clayey	25-30			
Clay	25-45			

Positive – in clay soils, clods break up when the soil dries; thereby, permeability for water and air is increasing, especially in the plow pan.

Negative – damage to plant roots and drying of soils to considerable depths.

5.11.14 Thixotropy of soil

Thixotropy of soil is a spontaneous, isothermal, reversible process in which a colloidal sol turns into a gel over time. By mixing, the gel is mechanically disrupted. Subsequently, a sol is formed, which transforms into a gel again at rest. When soil swells, thixotropy manifests itself in such a way that the volume of wet soil decreases after thorough mixing and, after being left at rest, it swells again without receiving water.

Thixotropy is most evident in soils with Na-montmorillonite, and less so in soils with Ca-montmorillonite and illite. Thixotropy, like swelling, has a great influence on changes in volume weight of the soil, porosity, structurality and water capacity.

The rest period of thixotropic soils can be interrupted by changes in moisture, development of the root system, tillage etc., which more or less do not disturb the thixotropic grid arrangement and thus the volume of the given soil and its physical properties. Therefore, in these soils, a single determination of physical properties is not sufficient for fixing of their average values.

5.11.15 Soil crust

Soil crust is formed mainly on the surface of heavier soils (higher clay content) after rapid drying of over-moistened and structurally disturbed soil. Its formation is explained by the fact that more colloids and salts remain near the topsoil surface after the evaporation of water. Among them, mainly sodium salts, less potassium salts, have an effect on the formation of soil crust. On the other hand, calcium and magnesium salts have no effect on the formation of soil crust. The more colloidal particles or sodium salts and less organic matter the soil contains, the more vulnerable it is to crust formation. Certainly, water has a decisive influence. The more intensive the rainfall (torrential) or irrigation is, the more susceptible the soil is to crust formation.

Soil crust is not formed on structural soils with a high proportion of humus as easily as on non-structural soils, where it usually appears after every rain. Therefore, e.g. Chernozems are less prone to crust formation than, for example, Luvisols. The grain composition itself also affects the formation of soil crust. The soil crust is not formed in soils that contain more than 60% of sand. On the other hand, the soil crust is formed in soils containing 25% of clay and silt.

Mechanical soil processing has a significant influence on the formation of soil crust. The formation of soil crust increases when dispersing soil aggregates or smearing the soil. The moisture during soil cultivation is important for its creation as well.

The soil crust has an adverse effect on the germination and growth of plants because they have to overcome mechanical obstacles during their germination. This weakens them and thus reduces their yield. There are several technical ways to eliminate the soil crust formation (Fig. 5.28). Naturally, they depend on the type of soil crust that is being formed.

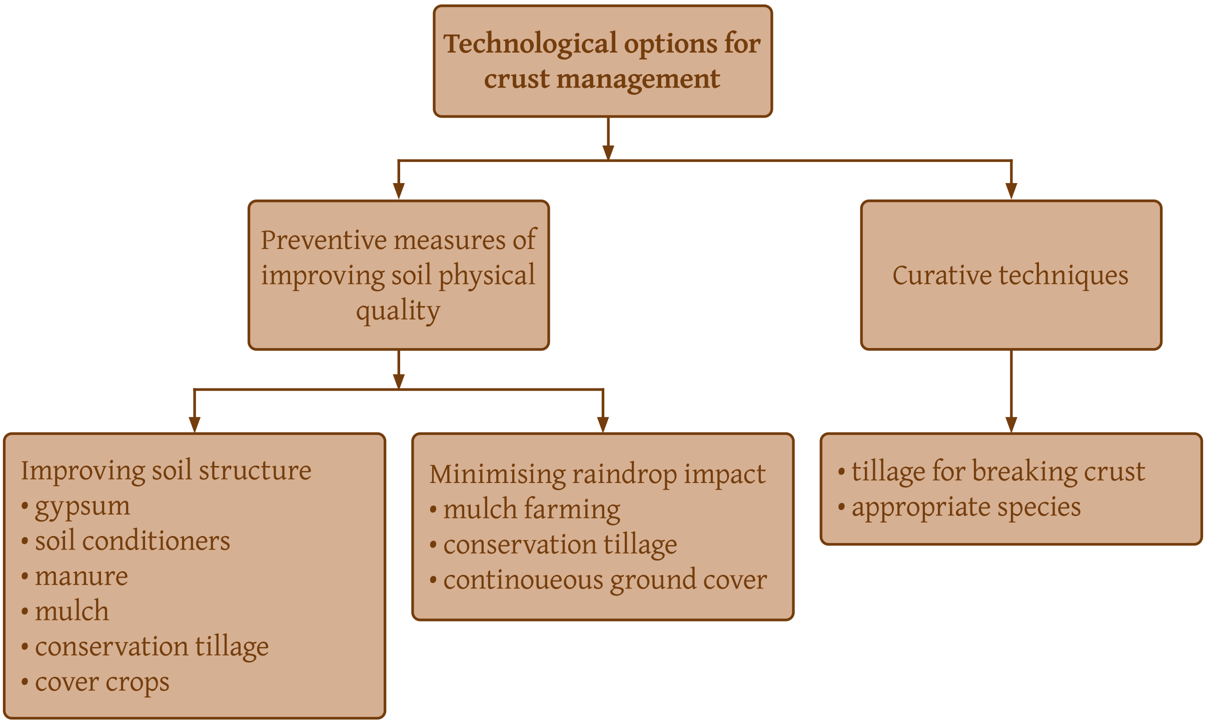


Figure 5.28: Soil and crop management opinions for reducing soil crust (Lal and Shukla, 2004)

Types of soil crusts

Chemical crust – is formed due to the crystallization of salts (naturally present in soils, or contained in industrial fertilizers) on the surface of the soil in arid and semi-arid conditions (Fig. 5.29 A).

Biological crust – is formed by the growth of algae and cyanobacteria on the surface of the soil. Slowly flowing water into the soil is the environment in which algae and cyanobacteria grow and develop. The water flows into the soil, and they remain and form the soil crust. Such soil crust is strongly hygroscopic and greatly reduces the rate of water infiltration into the soil (Fig. 5.29 B, C, D).

According to the FAO (1979), the soil crust can be expressed by the index of soil crust formation (Ic), which is based on the soil texture and soil organic matter content.

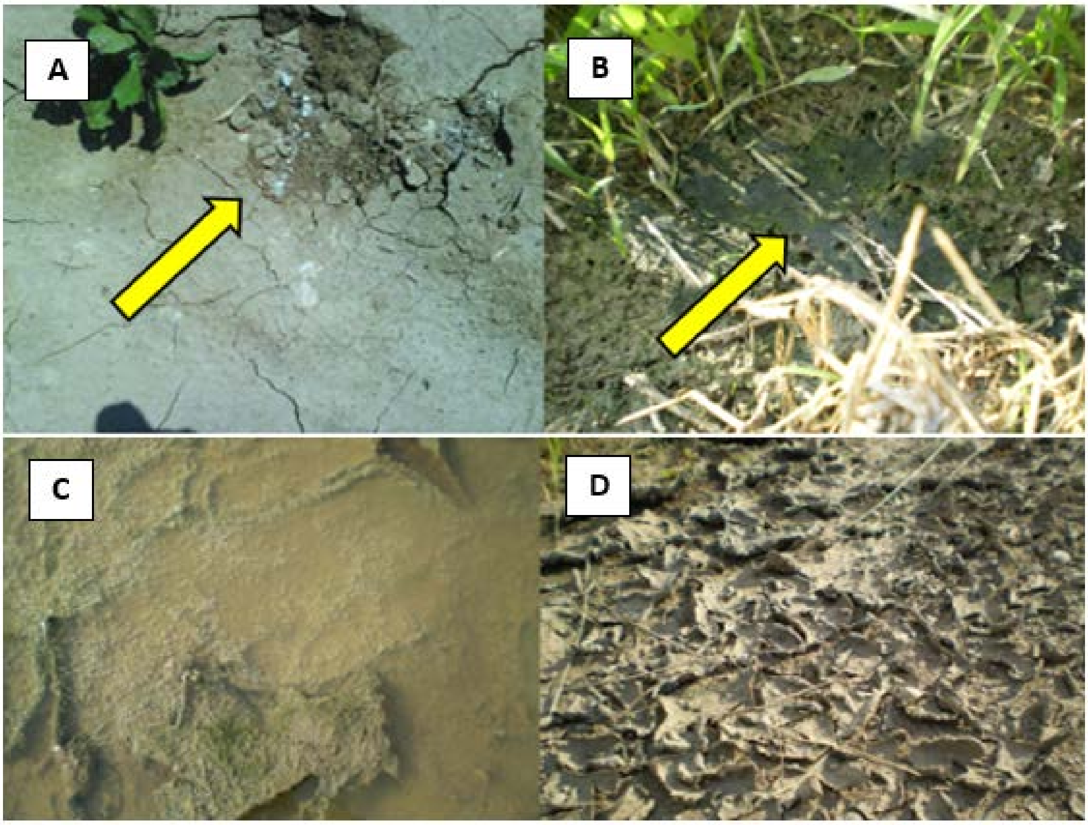


Figure 5.29: A) chemical and B, C, D) biological crust (Šimanský, 2010)

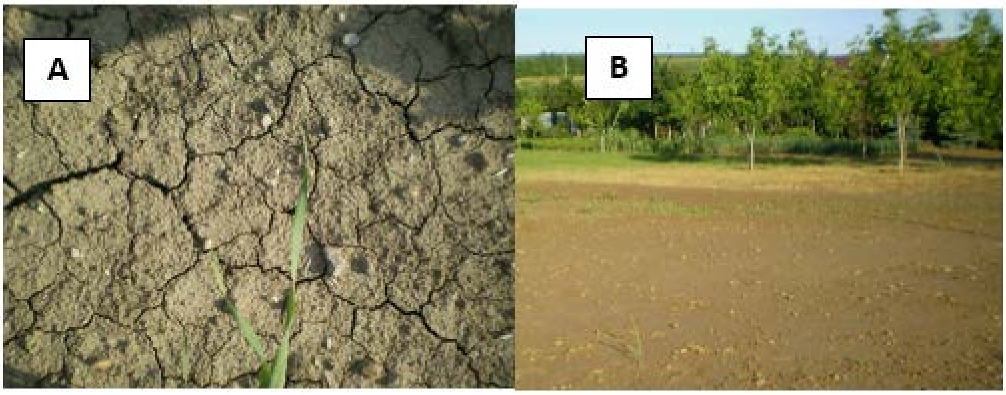


Figure 5.30: Physical crust A) structural and B) sedimentional (Šimanský, 2010)

Physical crust – is formed due to changes in the structure of the soil and can be: structural or sedimentional.

Structural crust – is formed due to the disintegration of aggregates caused by raindrops and the physicochemical disintegration of clay minerals. The top layer of the structural crust has low permeability and is 1-3 mm thick. Soils with a high content of exchangeable sodium are very susceptible to crust formation (Fig. 5.30 A).

Sedimentional crust – is formed by transporting and depositing fine particles within the surface flow. It tends to be coarser compared to the structural crust and is formed where the water brings with it granular fine particles that settle. The kinetic energy of raindrops does not affect their formation (Fig. 5.30 B).

$$l_C = \frac{S_j + 0{,}75.S_c}{C_l + (10.S_{OM})}$$

S_f – fine silt
S_c – coarse silt
C_l – clay
S_{OM} – soil organic matter

5.11.16 Soil lumpiness

Lumps are formed during plowing if it is carried out under unsuitable moisture conditions (Fig. 5.31). The tendency to lumpiness is influenced by the soil texture, soil cohesion, and the content of humus and calcium. In addition to soil effects, the lumpiness is affected by weather conditions and inappropriate tillage operations. Heavier carbonate-free soils with a low humus content have a high tendency to lumpiness. In terms of granularity, light soils do not form lumps, clay soils form about 20% of smaller lumps, loamy-clay soils form about 40%, and clay weak-humus soils up to 70% of large lumps.

When plowing is done correctly, the soil should be properly comminuted. The amount of lumps over 50 mm should be only about 10% for loamy soils. For clay soils, the amount of lumps over 100 mm should be only up to 20%. In order to increase the friability of heavy soils, it is necessary to increase the content of organic matter in the soil. Favorable soil moisture levels should be used for plowing or subtle tillage when the soil crumbles best. Moreover, we must not forget that the soil must not be let dry out before and after plowing.

Figure 5.31: Soil lumpiness (Šimanský, 2013)

5.11.17 Lumps disintegration due to intensive tillage of dry soil

It is the disintegration of the soil structure into elementary particles, which is caused by frequent harrowing and dragging of the dried-up soil surface. An important factor influencing lumps disintegration is soil texture and stability of soil aggregates, mineralogical and petrographic composition, humus content, and lack of water. Concerning individual granularity fractions, the content of silt and sand is the most important because they are not sufficiently bound with humic substances and other components in the soil. So, it happens that the dispersed silt falls through the rest of the structural topsoil and accumulates in its lower part. And when wet and dry conditions alternate, it can form a less permeable layer.

Sandy soils (79.6%), loamy-sand soils (44.4%), loamy soils (24.2%) as well as clay loam (12.9%), and clayey soils (7.6%) are characterized by the highest lumps disintegration.

5.11.18 Soil maturity

It is the optimal soil condition that allows the most suitable cultivation. The soil is not smeared, nor clumped, nor sprayed, but crumbles into structural elements. The basic condition for maturity is a stable, fine-grained structure, which is able to maintain optimal moisture for a long time. There are two types of soil maturity: physical and biological.

Physical maturity – is defined by the favorable structure and moisture of the soil, which allow tillage, in which the soil is crumbled into structural aggregates. During the growing season, physical maturity is maintained by the plant cover, which protects the surface from unproductive evaporation. After the harvest, shallow stubble tillage protects the soil from drying out.

Biological maturity – is defined by the intensive activity of soil microorganisms and plants and the release of nutrients for plants. It depends on soil regimes and the content and quality of soil organic matter.

6

Chemical properties of soil

The chemical composition of soil is expressed by its percentage of individual elements. However, this does not mean that all the elements found in the soil are accessible to plants. The interrelationships between the soil environment and the plant are complex and constantly studied. Great attention is paid to the accessibility of particular nutrients for plants, the study of ion exchange, and to the composition of the soil sorption and colloidal system.

The solid phase of soil contains practically all 95 or more chemical elements that have been discovered so far. Numerous analyses of individual elements found in the Earth's crust show that only a few of them make up the main share in the soil, while the others are found in minimal representation. The representation of individual elements in the soil is significantly influenced by the course of the soil-forming process. An increase in the concentration of some elements takes place during the weathering of minerals and rocks. On the other hand, their decrease in the soil environment occurs when they are intaken by plants or animals.

Chemical composition of soil

Most plants receive nutrients from the soil environment, but there are also those that are able to receive e.g. nitrogen from the atmosphere. Inorganic elements found in the soil come from the weathering of rocks and minerals, and their representation depends on the percentage content in original rocks and minerals. Their content in the soil depends on the degree of weathering of rocks and minerals and on the losses caused by their leaching from the soil. Soils that contain more organic matter are more productive than soils with low organic matter content. Moreover, soils that contain a relatively high content of calcium, especially in an easily soluble form, are usually more productive. On the other hand, high content of calcium carbonate can also have a negative effect.

Composition of soil types

If we talk about the chemical composition of soils, unless otherwise stated, we always mean the surface layer, or, the topsoil layer of the soil. This is a part of the soil from which the plants obtain a significant part of their nutrients, which can be regulated by man through tillage using fertilizers, lime application, incorporation of post-harvest residues, or by means of various methods of soil management and land use. The amount of different components found in the soil can be expressed as their percentage share or converted to mass units in a certain volume. The composition can be expressed in chemical elements such as Ca, P, and K or as element oxides such as CaO, P_2O_5, and K_2O. Individual soils were formed on diverse soil-forming substrates, such as various fluvial sediments that were accumulated along river courses (e.g. eolitic sediments that were blown by the wind, or soils that were formed on volcanic ash, different rocks such as granite or limestone). Their development was significantly influenced by the effects of climatic factors, flora, and fauna, simply, by different factors and conditions of the soil-forming process, which was also reflected in the different chemical compositions of soils. In almost all mineral soils, SiO_2 is present in the largest percentage. In deeper layers of soils, we can find a higher representation of Al_2O_3 and Fe_2O_3. The next (fourth) most widespread element (compound) is CaO or K_2O. Soils are generally lower in MgO than CaO and significantly lower than Na_2O and K_2O. The representation of Al_2O_3, Fe_2O_3, CaO, MgO, K_2O, and Na_2O in soils is much more variable than the representation of SiO_2. Higher content of CaO and MgO is found in a more arid climate, where the soils are drier compared to a more humid climate. As already mentioned, the chemical composition of soils also affects the content of soil organic matter.

Table 6.1: Chemical composition in individual soil horizons (Millar et. al., 1962)

Horizon	Representation in %						
	SiO_2	Al_2O_3	Fe_2O_3	CaO	MgO	K_2O	P_2O_5
A	93.0	2.9	0.8	0.13	0.05	0.05	0.04
C	75.5	11.4	4.4	0.03	0.04	0.22	0.06
A	79.3	10.9	1.0	0.33	0.06	5.15	0.26
C	70.5	17.5	1.0	0.24	0.18	6.22	0.32

Composition of soil classes

Since the soils also differ from each other in their granulometric composition, we can expect that they will also have different chemical compositions. Concerning granularity, soils that contain larger particles present a higher proportion of mineral substances resistant to weather conditions (e.g. quartz) compared to soils containing finer fractions (dust, clay). The presence of compounds such as P_2O_5, K_2O, CaO, and MgO is significantly higher in finer-grained soils than in texturally coarser-grained ones. Sandy soils often have a higher content of quartz and a lower content of other elements, especially nitrogen and phosphorus. In general, soils with a considerable presence of silt or clay are richer in nutrients than soils with a significant representation of sand or containing a higher representation of the skeleton.

Comparison of the chemical composition of soils within the soil profile

Since each soil type developed as a result of the soil-forming process on different parent substrates, we can logically expect a different chemical composition depending on the soil horizon. Usually there is a difference between the upper horizon (A) and the parent materials (C), or bedrock (R), but significant differences can often be observed between horizons B and C, resp. all diagnostic horizons.

6.1 Soil colloids

Soil colloids are particles that range from 1 nm to 1 µm. They represent very complex systems that are characterized by high dynamic properties depending on environmental conditions (soil reaction, humidity, temperature, soil solution concentration).

Concerning their origin, they can be divided into: **1. mineral, 2. organic.**

1. Mineral colloids

Colloids of aluminosilicate and silicate origin are formed during weathering by dispersion, thermal disintegration (crumbling) and splitting of minerals. The formation of colloids of the mentioned origin is accelerated by the reactions of hydrolysis, hydration and oxidation of minerals.

Colloids of hydrated oxides are formed by the association of hydroxide molecules released during weathering. In the association, polymerization, condensation and partial dehydration reactions are applied.

Mineral colloids are represented in the soil by a significantly larger proportion compared to organic ones. The group of mineral colloids includes clay minerals (mainly aluminosilicates), finely dispersed primary minerals (silicates, aluminosilicates), polymeric colloidal silicic acid and hydrated polymeric oxides of aluminum, iron and manganese.

2. Organic colloids

Organic colloids are formed by microbial activity as products of enzymatic hydrolysis of plant and animal residues and macromolecular products of microbial synthesis. In the soil, they are mainly represented by humic substances (humic acids, fulvic acids, humins) and to a lesser extent by proteins, lignin substances, and polyuronides. Organic aggregates with the properties of colloids are formed by association, polymerization and polycondensation.

Mineral and organic colloids are usually not separated in the soil, but form a very complex organo-mineral colloidal complex. Essentially, a layer of hydrated sesquioxides (Fe_2O_3, Al_2O_3) combined with organic colloids is adsorbed on the surface of aluminosilicates.

6.1.1 Properties of soil colloids

A characteristic feature of colloidal systems is a large specific surface area. As the particle size decreases, the specific surface area increases. Due to their large surface area, colloidal particles exhibit special physical and chemical properties. Contact surface layers are formed at the interface of two phases (solid and liquid), which are rich in energy (surface energy) and therefore enable diverse physical and chemical reactions. Adsorption of soluble substances and gases takes place on these active surfaces. Soil colloidal particles are attracted to one another (cohesion). Colloidal phenomena of an electrical nature arise.

Surfaces of soil colloids can be:

- **polar** – if they create an electric field without external influences,
- **non-polar** – if an electric field is created on the surface only after the adsorption of polar ions, molecules or by other external influence.

The surface of soil colloids is heterogeneous. This means that places with polar and non-polar character alternate. The parts of the surface with polar properties represent active centers that are most used in adsorption. Colloidal particles of expansive three-layered clay minerals and humic substances have both the outer and inner surfaces, which is manifested by an increased adsorption ability.

One of the most important properties of soil colloids is their ability to dissociate ions into the solution and adsorb ions from the solution instead. Due to the dissociation of ions, soil colloids acquire electrokinetic properties, which are very important in the adsorption of ions in the soil. According to the nature of dissociation and adsorption, we divide soil colloids into:

- **acidoids** – they have a negative electric charge and are thus negatively charged; they release hydrogen cations and adsorb cations from the soil solution during dissociation,
- **basoids** – they have a positive electric charge and they release hydroxide anions and adsorb anions from the soil solution during dissociation,
- **ampholytoids** – they are similar to amphoteric compounds and can change the electric charge. Concerning dissociation, they release either hydrogen cations or hydroxide anions and adsorb cations or anions depending on the reaction of the dispersion medium.

With an excess of H^+ ions (acidic environment), they behave like basoids, that is, positively charged particles capable of adsorbing anions are formed.

$$Al(OH)_3 \leftrightarrow Al(OH)_2^+ + OH^-$$

In alkaline soils, with an excess of OH^- ions, negatively charged particles capable of adsorbing cations are formed.

$$Al(OH)_3 \leftrightarrow Al(OH)_2^- + H^+$$

In the soil, the colloidal fraction is mainly formed by clay minerals and humic substances, which have the character of acidoids. Polymeric hydrated sesquioxides are represented by a smaller proportion and have the properties of ampholytoids, which behave as basoids in our (predominantly acidic) soils. The complex structure of the organomineral colloidal complex causes that its predominantly acidoid character is not clearly manifested.

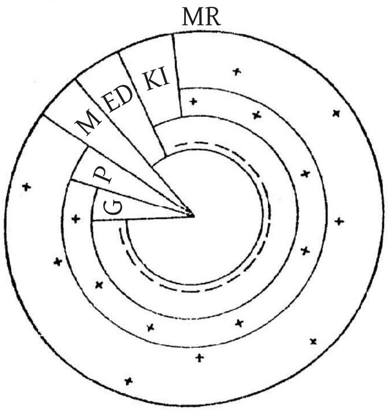
According to the relationship to water and the stability of the system, soil colloids are divided into:

1. **Hydrophobic** – they are not able to retain water, they are not very stable and easily precipitate with electrolytes or due to dehydration,
2. **Hydrophilic** – they are strongly hydrated and are more stable in the medium.

Hydrophobic colloids primarily include hydrated sesquioxides, and hydrophilic colloids include humic substances, proteins, and starch. Clay minerals behave as weakly hydrated colloids in relation to water.

6.1.2 Structure of soil colloids

Soil colloids form a system consisting of a dispersed phase (colloidal particles) and a dispersion medium (soil solution). Figure 6.1. shows a simple diagram of the structure of a colloidal micelle. In the center of the solid part, there is a core made up of aluminosilicates, iron oxides, silicic acid, humic acids, and proteins, i.e. the core is made up of either crystalline or amorphous substances.



MR – soil solution
KI – compensatory ions
ED – electrical double layer
M – micelle
P – particle
G – granules

Figure 6.1: Structure of colloidal micelle (Gorbunov, 1957)

On the surface of the core, there is a tightly bound layer of ions, which determines the potential of the colloidal micelle. In both cases, these are negatively charged acidoids. The core with the charging layer forms a colloidal granule. The electric charge of the granule is neutralized by the adsorption of oppositely charged ions and thus a layer of compensatory ions is formed on the surface of the granule. The layer of compensatory ions with the charging layer form an electrical double layer, which determines the nature of the electrokinetic potential. The layer of compensatory ions is divided into two layers – the immobile one, which closely adheres to the charging layer, and the diffuse layer, i.e. towards the external soil solution.

6.1.3 Coagulation and peptization of soil colloids

Soil colloidal systems can be in the state of a colloidal solution, colloidal sol, or in the state of a colloidal gel. Colloidal solutions and sols pass through coagulation into a gel state, and conversely, the colloidal gel changes into colloidal solutions and sols through peptization.

The transformation of a sol into gel and vice versa can be reversible or irreversible. The vast majority of soil colloids are characterized by reversibility, which is of great importance in terms of restoring soil structure and fertility.

An important property of soil colloids is their stability, i.e. the ability to remain in the state of sol and resist coagulation. From the point of view of soil fertility, this is an undesirable feature. The soil structure improves by losing the stability of the colloidal system in the soil.

The stability of the colloidal system is conditioned by the high value of electrokinetic potential and the considerable thickness of the hydration shell. Exchangeable cations have the most significant effect on the value of the electrokinetic potential and thus also on the change of the colloidal system (acidoids). Their influence depends on the ionic properties, i.e. valence, radius, atomic weight, and thickness of the hydration shell (Fig. 6.2). The higher the valence, radius, and mass of the ion, and the smaller its hydration shell is, the greater is its coagulation efficiency.

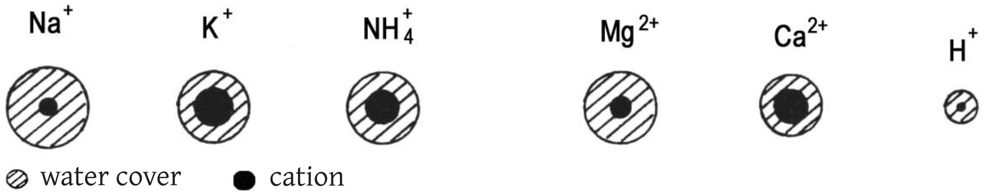


Figure 6.2: Size of cations (Fecenko and Ložek, 2000)

At the same potency, the influence of exchangeable cations depends on their position in the lyotropic series:

$$Cs^+ > Rb^+ > K^+ > Na^+ > Li^+$$

$$Ba^{2+} > Sr^{2+} > Ca^{2+} > Mg^{2+}$$

Various salts also affect the colloidal system in soil. The influence of calcium and sodium salts is particularly important:

1. **Calcium salts** ($CaCO_3$, $Ca(HCO_3)_2$, $CaSO_4$) effectively induce coagulation of colloids,
2. **Sodium salts** ($NaCl$, Na_2CO_3, $NaHCO_3$, Na_2SO_4) cause peptization of colloids.

6.1.4 Sorption capacity of soil

The ability of the soil to bind ions and molecules of various substances from the solution is called the sorption capacity of the soil. The formation of this very important property of the soil is conditioned by the soil colloidal complex – colloids of mineral and organic origin. The sorption capacity of soils is the second most important process that takes place on the Earth after photosynthesis. Sorption manifests itself positively (retention of water and nutrients), but also negatively (retention of harmful substances). Main mechanisms of soil sorption are:

1. Mechanical sorption

The principle of mechanical sorption consists of the mechanical retention of fine particles in soil pores. It depends on the soil texture and aggregate composition of soils. Clay soil and finely aggregated soils are characterized by higher mechanical sorption than sandy and coarsely aggregated soils.

2. Physical sorption

It is related to surface phenomena at the phase interface of the colloidal system. It is conditioned by the free surface energy that is formed at the interface between solid parts of the soil and the soil solution. It manifests itself by increasing (positive adsorption) or decreasing (negative adsorption) the concentration of molecules on the surface of the solid phase and their decrease or increase in the soil solution.

3. Physical-chemical (exchange) sorption

It is based on the ion exchange between the soil colloidal complex and the soil solution, which takes place in equivalent proportions. The energy of cation exchange depends on their valence, atomic weight, and radius size. We consider this adsorption to be the most important, and therefore we pay the most attention to it.

Cation exchange cpacity

Most colloids have a negative charge, which conditions the adsorption of cations in the soil. The heterogeneity of the colloidal system in the soil makes the process of cation exchange complex and cannot be described in a simple way. Nevertheless, cation exchange takes place on the basis of certain principles.

During the exchange adsorption, the cations of the soil solution penetrate into the compensation layer of the colloidal system, from which they force out an equivalent amount of other cations into the soil solution:

$$[PKK]^{Na^+}_{Na^+} + soil\ solution = Ca \leftrightarrow [PKK] = Ca + soil\ solution^{Na^+}_{Na^+}$$

Principles of cation exchange

Diffusion of cations caused by their thermal movement, and electrostatic attractive forces cause their constant movement from the solution into the ionic bilayer in the phase interface and in the opposite direction. The number of ions moving in both directions is equivalent.

The rate of exchange on the outer surface of colloids is high (it takes several minutes) and on the inner surface, it takes several hours to days.

The exchange and energy of adsorption are influenced by the properties of cations – valency ($Na^+ < Ca^{2+} < Al^{3+}$), radius (cations with a smaller radius have a greater electric charge density, therefore, they are able to bind a larger amount of water, which weakens their adsorption capacity – $Na^+ < K^+$; $Mg^{2+} < Ca^{2+}$). The cation exchange capacity is also influenced by the concentration and volume of the solution. The properties of adsorbents influence the cation exchange (adsorption) as well. Different adsorbents have different adsorption capacities and adsorb cations differently, which is related to their chemical composition and structure. The exchange capacity of some clay minerals is presented in Table 6.2.

Table 6.2: Cation exchange capacity of some clay minerals (Scheffer and Schachtschabel, 1970)

Clay mineral	Cation exchange capacity (mmol kg^{-1})
Kaolinite	30 – 150
Halloysite	50 – 100
Montmorillonite	800 – 1 200
Vermiculite	1 000 – 2 000
Glauconite	50 – 400
Illite	200 – 500
Chlorite	100 – 400
Allophane	up to 1 000

4. Chemical adsorption

Chemical adsorption is the soil‘s ability to retain nutrients as a result of chemical reactions that take place in the soil. During these chemical reactions, compounds that are soluble in water are formed into compounds that are slightly soluble or insoluble in water. Monovalent anions NO_3^- and Cl^- and monovalent cations K^+, NH_4^+ do not form insoluble compounds and therefore cannot be bound by chemical adsorption. The divalent anions CO_3^{2-} and SO_4^{2-} can form hard-to-dissolve compounds with the divalent cations Ca^{2+} and Mg^{2+}.

That is why these anions can be bound in the soil by chemical adsorption. Chemical adsorption of anions of phosphate fertilizers is of great practical importance. In the reactions between $H_2PO_4^-$ and HPO_4^{2-} anions with divalent Ca^{2+} and Mg^{2+} cations on neutral and alkaline soils and with trivalent Al^{3+} and Fe^{3+} cations on acidic soils, slightly soluble to insoluble compounds are formed.

The reaction between the active component of superphosphate $Ca(H_2PO_4)_2$ in neutral to slightly alkaline soils containing $CaCO_3$ and $Ca(HCO_3)_2$ takes place according to the following scheme:

$$Ca(H_2PO_4)_2 + Ca(HCO_3)_2 \rightarrow 2CaHPO_4 + 2H_2CO_3$$
$$Ca(H_2PO_4)_2 + 2Ca(HCO_3)_2 \rightarrow Ca_3(PO_4)_2 + 4H_2CO_3$$

Slightly soluble to insoluble phosphorus compounds can also be formed in soils with a pH value of around 7, which do not contain carbonates. Exchange reactions with the calcium of the diffuse layer of soil colloids take place here:

$$[PKK]Ca^{2+} + Ca(H_2PO_4)_2 \rightarrow [\mathrm{PKK}]^{H^+}_{H^+} + 2CaHPO_4$$

In alkaline soils, first $CaHPO_4$ and less soluble CaHPO are formed from water-soluble phosphates. Further, $2H_2O$ and $Ca_4H(PO_4)_3$ are formed. Then, $3H_2O$ and probably after years apatite, e.g. $Ca_5(PO_4)_3OH$ are formed. These transformations from amorphous to crystalline forms are very slow. Less soluble calcium phosphates are also made accessible during local reactions in the soil with free acids, root secretions, and soil microorganisms. For example, $Ca_3(PO_4)_2$ can react with nitric acid, which is produced in the nitrification process, and water-soluble $Ca(H_2PO_4)_2$ is formed.

Insoluble phosphates are formed in acidic soils during reactions between active Fe, Al, and Mn.

$$Fe(OH)_3 + H_3PO_4 \rightarrow FePO_4 + 3H_2O$$

Chemical adsorption in acidic soils can be limited by liming and the use of organic fertilizers.

5. Biological sorption

Biological sorption can be understood as binding of nutrients in the soil in a biological way by means of soil microorganisms and plants. Biological sorption is selective because organisms absorb those elements they need for their life. Soil microorganisms participate very intensively in the transformation of nutrients in the soil, both in their binding into organic compounds and in their release into mineral, plant-accessible forms. At the same time, a part of nutrients accumulates in the microbial plasma. The symbiosis of nodule-forming bacteria with plants takes place on the roots of leguminous plants. Nodule-forming bacteria have the ability to biologically bind elemental nitrogen in the air. The ability to biologically fix atmospheric nitrogen is presented by some free-living bacteria in the soil, e.g. *Clostridium, Azotobacter*, or by bacteria living in the rhizosphere of plants (*Azosprillum*). Biological sorption is important for the balance of nutrients, their immobilization in the soil and thus also for the nutrition and fertilization of plants. We can have a positive relationship, but also a negative impact on the cultivated crop. After the application of organic matter with a wide C : N ratio into the soil and within a lower nitrogen content in the soil, we can observe a negative effect of biological sorption on the cultivated crop. This is because microorganisms use nitrogen from fertilizers at the expense of the crop, which is manifested by a decrease in yield. The binding of mineral nitrogen by microorganisms leads to the so-called nitrogen depression.

Exchangeable cations in soil

The sorption capacity and composition of exchangeable cations in the soil reflect the course of the soil-forming process and, at the same time, characterize the properties of the soil. The representation of individual cations in soils is a reflection of the action of climatic factors.

Ca^{2+}, Mg^{2+}, H^+ and Al^{3+} belong to the main exchangeable cations in soils of humid temperate climates. Potassium, sodium, iron, manganese and other elements are present in smaller quantities. Ca^{2+} is the most represented in arable soils (more than 50% of the sum of exchangeable cations). In forest soils, the exchangeable calcium content is significantly lower than in arable soils. Sometimes it reaches <10% of the sum of exchangeable cations. H^+ and Al^{3+} ions predominate in these soils.

In soils of an arid climate, in addition to Ca^{2+} and Mg^{2+} , the main exchangeable cations also include Na+, which can be dominant in some soils. If the content of Na^+ is >5% of the sum of exchangeable cations, it indicates the process of soil solonization, which is manifested by an increase in the content of exchangeable sodium in the sorption complex (solonetz contains >20% of Na+ from the sorption capacity).

6.1.5 Effect of exchangeable cations on soil properties and plants

Exchangeable cations of the colloid complex of soils significantly influence not only chemical processes and biological conditions, but also the physical condition and technological properties of the soil.

Soils with a predominance of exchangeable Ca^{2+} and Mg^{2+} have optimal chemical dynamics and a neutral soil reaction. Humic substances are saturated with Ca^{2+} and a water-resistant fine-grained structure is formed. Soils have a high buffering capacity. They resist changes in pH also with the content of calcium carbonate and adsorbed divalent base cations.

Soils with a predominance of exchangeable H^+ and Al^{3+} are sorptionally unsaturated. Strong leaching by rainwater, which is also intensified by organic acids, causes a loss of base cations and an increase in the adsorption of H^+ and Al^{3+} ions. The humus of these soils is unsaturated, the reaction is acidic and the structure is not very stable. At an extremely acidic pH, Al^{3+}, Mn^{2+} and Fe^{2+} ions are significantly activated, which have a toxic effect on plants at a higher concentration in the soil solution.

Soils with a high content of exchangeable Na^+ have also unfavorable properties. The adsorbed Na^+ cation causes intense and permanent peptization of colloids, which adversely affects the state of the physical properties of the soil. The reaction of the soil is alkaline a large amount of exchangeable Mg^{2+} also increases soil alkalinity, which adversely affects the mobility and accessibility of some nutrients in the soil (P, N).

The importance of exchangeable cations for plants is that they affect the overall chemistry, biological activity, and physical condition of the soil, and that they are sorbed by plants in varying amounts.

Sorption properties of the soil are characterized by:

1. **Cation exchange capacity (CEC)**
   - It is the amount of cations in mmol (p^+) that 1 kg of soil can bind. Sandy soils have a total sorption capacity of 20-100, loamy 200-300, clayey 400-500 and soil with higher content of organic matter around 1,500 mmol (p^+) kg^{-1} of the soil.
2. **Sum of exchangeable base cations (SBC)**
   - It indicates the amount of base cations (Ca^{2+}, Mg^{2+}, K^+, Na^+) in mmol (p^+) kg^{-1} of the soil.
3. **Hydrolytic acidity (HA)**
   - It indicates the amount of H^+ and Al^{3+} ions in mmol (p^+) kg^{-1} of the soil, which are displaced from the soil colloidal complex by hydrolytically alkaline salt.
4. **Base saturation (Bs)**
   - It expresses the share of exchangeable base cations in % of the total sorption capacity. Chernozems have a degree of saturation above 95%, Luvisols soils above 75%, Albiluvisols, Planosols, Stagnisols 40-60%, Cambisols 30-70%, and Podzols below 30%. If the value of Bs>75%, we consider such soil to be sorptionally saturated. Anthropogenic interventions in the soil environment can significantly affect changes in the indicators of soil sorption properties.

Fixation of cations

During the adsorption of cations, part of the ions is bound by soil colloids so tightly that they cannot be released by desorption. These non-exchangeable cations are difficult or inaccessible for plants. Those cations that enter the hexagonal cavities that are formed in the crystal lattice of layered aluminosilicates are subject to non-exchangeable adsorption. Hydration increases the size of ions, so fixation takes place when the soil dries. Cations with a smaller diameter (Na^+, Mg^{2+}, Ca^{2+}) are located outside the center, closer to the layers in the hexagonal cavities. Therefore, they are not subject to fixation.

The binding strength of cations in the soil is not absolutely constant because even irreversibly bound cations and cations of the crystal lattice of minerals can be gradually released during weathering and pass into an exchangeable or water-soluble form.

Anion exchange

Adsorption of anions in the soil is carried out by colloids with the properties of basoids. Its capacity is lower because there are far fewer positive charges than negative charges on the polar surface of adsorbents. In the adsorption of anions, the same principles and rules apply as in the adsorption of cations. It is conditioned by the properties of anions, soil colloids, and the reaction of the medium. According to the sorption capacity, we can place anions in the following series:

$$Cl^- = NO_3^- < SO_4^{2-} < PO_4^{3-} < OH^-$$

The higher the valency of the anion, the more intensively it is adsorbed on the surface of colloids. The exception is the anion OH^-, which exhibits the greatest adsorption activity. The adsorption of anions is significantly influenced by the composition and properties of colloids, and the content of hydrated sesquioxides, which have the properties of basoids. Changes in the pH of medium affect the electrokinetic potential of colloids. In an alkaline medium, the negative potential increases, and in an acidic medium, the positive potential increases. In an acidic medium, the adsorption of anions is higher than in an alkaline one.

From the mentioned series, it follows that Cl^- and NO^{3-} anions are practically not adsorbed even by chemical adsorption because they form easily soluble compounds and easily migrate in the soil.

Negative adsorption

When diluted salt solutions come into contact with the soil, unlike cations that are adsorbed by the soil, the concentration of anions in the solution increases. They are repelled by the surface of solid particles. It is, so-called, negative adsorption (cations are attracted by negative charges, anions are repelled). Another cause of negative adsorption of ions is the simultaneous adsorption of water molecules. Within an increased adsorption of water by the solid phase of the soil, the concentration of ions in the solution increases. This mechanism causes a negative adsorption of both types of ions within their equivalent adsorption.

Negatively adsorbed anions are easily leached from the soil, causing the loss of important nutrients (NO_3^-). Losses depend on the position of the anion in the series ($Cl^- = NO_3^- > SO_4^{2-}$) and on the cation saturation of the soil ($Na^+ > K^+ > Ca^{2+}$). High cation exchange capacity, alkaline reaction and a high degree of dilution of the soil solution (irrigation, rain) increase the negative adsorption of anions and thus also their loss from the soil.

6.2 Soil pH

Soil reaction is an important agrochemical property of soil that significantly affects plant growth and development. It is formed by internal factors, such as the chemism and texture of rocks from which the soil was formed, and by external factors, such as precipitation, plant root secretions, agricultural engineering, and the use of fertilizers.

The reaction of the soil depends on the concentration (activity) of H^+ and OH^- ions in the soil solution. It is usually expressed in the 14-item Sorensen scale from 0 to 14, and is represented by the symbol pH. pH (*pundus hydrogenii*) is the potential of H^+ ions with the corresponding value indicating the negative decimal logarithm of the hydrogen ion concentration. The concentration of hydrogen ions according to Sorensen can, therefore, be expressed as follows:

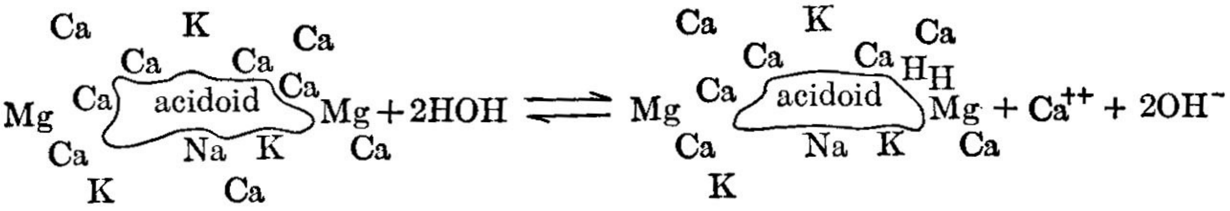
$$pH = - \log [H+]$$

One liter of decarbonized distilled water contains 0.00000001 or 10^{-7} g of H^+ and the same amount of OH^- ions. This means that in a neutral reaction, pH is pH = pOH = 7.

When there are more hydrogen cations than hydroxide anions in a solution, the solution has an acidic reaction. If hydroxide anions predominate over hydrogen ions in the solution, the solution has a basic reaction. Since the product of the concentration of these ions is always constant, any increase in the alkalinity of the solution means an increase in the concentration of hydroxide ions and a decrease in the concentration of hydrogen ions in the solution.

6.2.1 Soil acidity

Soil acidity is primarily influenced by the function of the colloidal fraction of the soil, but soluble acids in the soil solution also contribute to its formation to a lesser extent. It is estimated that the acidity of the soil medium influences the composition of the mineral content of the soil to 95%. Colloidal particles mostly behave as complexes of acids with radicals of very poor solubility in the surroundings, and there are mostly positive ions at different distances from the surface of the particle. These cations dissociate to a certain extent into the soil solution, and hydrogen ions are adsorbed from the soil solution onto the acidoid surface.



Hydroxide anions then cause an alkaline reaction in the soil solution. The hydrogen cations will undergo some degree of dissociation on the soil complex. However, they are still located close to the acidoid core, while the other cations may be dispersed further away and are not subject to such strong attraction. If hydrogen cations and hydroxide anions are equally represented around the core, the soil reaction is neutral. However, if the number of H^+ exceeds the number of OH^-, the soil reaction is acidic.

Soil acidity occurs in regions with sufficiently high rainfall, causing leaching of significant amounts of exchangeable base (Ca^{2+}, Mg^{2+}) and alkaline cations (Na^+, K^+) from the surface layers of the soil.

Small amounts of inorganic acids, such as HNO_3, HNO_2, H_2SO_4 and H_3PO_4, can be commonly found in soils. A significant concentration of H_2SO_4 can be found in soils containing FeS_2. If CO_2 is present in the soil, a slightly acidic reaction is created in water. Organic acids such as acetic, citric and oxalic acids, are usually found in the soil within the decomposition of fresh organic residues.Their action is short because acids of this nature in well aerated soils are decomposed by soil microorganisms. Many studies have pointed out that if the soil colloidal complex is saturated with divalent Ca^{2+} and Mg^{2+} cations, the soils have a weak alkaline reaction. As the soil reaction approaches neutral, there is very little exchangeable hydrogen in the soil.

Forms of soil acidity

1. Active acidity

The acidity of the soil solution is conditioned by the predominance of free hydrogen ion concentration over hydroxide ions. Free hydrogen ions can be maintained in the soil solution only when there is a lack of alkaline salts in the soil and H+ ions predominate in the sorption complex.

In addition to fertilizers, the acidification of the soil solution can also be influenced by CO_2 produced during biological processes in the soil. CO_2 conditions the formation of H_2CO_3 in the soil solution, which dissociates into H^+ and HCO_3^- ions. This increases the concentration of H^+ ions in the soil solution and causes its acidification. 5 mg of CO_2 in 1 liter of water reduce the pH to 5.2, and when 150 mg are dissolved in 1 liter of water, the pH is reduced to 4.5. However, the reaction of the soil solution in the soil does not change significantly under the influence of CO_2.

Active acidity in the soil solution consists of carbonic acid, water-soluble organic acids and hydrolytically acidic salts.

2. Potential acidity

Potential acidity is caused by hydrogen ions released from the sorption complex, which increase the concentration of H+ ions in the soil solution. In this way, its acidity is increased as well.

a) Exchangeable acidity

If sorptionally unsaturated soil is treated with a neutral salt solution, e.g. KCl, salt cations are exchanged for hydrogen ions of the sorption complex:

$$[PKK]^{H+} + KCl \rightarrow [PKK]^{K+} + H^+ + Cl^-$$

Hydrogen ions released from the sorption complex increase the concentration of hydrogen ions in the soil solution, and the acidity of the soil solution increases by, so-called, exchangeable acidity.

When the solution acts on sorptionally unsaturated soil, an exchange reaction can occur between K+ ions and Al3+ ions of the sorption complex:

$$[PKK]^{Al^{3+}} + 3KCl \rightarrow [PKK]^{K^{+}\,\substack{K^{+}\\K^{+}}} + AlCl_3$$

$AlCl_3$ in the soil solution undergoes hydrolysis forming the weak base $Al(OH)_3$ and the strong acid HCl:

$$AlCl_3 + 3H_2O \rightarrow Al(OH)_3 + 3H^+ + 3Cl^-$$

Exchangeable acidity is present mainly on sorptionally unsaturated soils when physiologically acidic fertilizers are used. Therefore, exchangeable acidity is also a starting point for determining the need for liming. Physiologically alkaline fertilizers should preferably be used on soils with exchangeable acidity.

b) Hydrolytic acidity

This acidity represents the amount of hydrogen ions that can be displaced from the sorption complex of the soil by the action of the solution of hydrolytic alkaline salt on the soil. Such salt is sodium or calcium acetate, which hydrolytically splits in aqueous solution to form weakly dissociated acetic acid and strongly dissociated base:

$$CH_3COONa + H_2O \rightarrow CH_3COOH + NaOH \rightarrow CH_3COOH + Na^+ + OH^-$$

The solution has a pH of around 8.5. The alkaline reaction of the solution of this salt is the main cause of the more complete displacement of bound hydrogen ions in the soil sorption complex. When the sodium acetate solution interacts with the soil sorption complex, sodium ions displace hydrogen ions from the sorption complex into the solution, where they bind with hydroxide ions and form a slightly dissociated water molecule:

$$[PKK]^{H^+}_{H^+} + CH_3COOH + Na^+ + OH^- \leftrightarrow [PKK]^{Na^+} + CH_3COOH + H_2O$$

Hydrolytic acidity depends on the degree of saturation of the sorption complex with bases. The value of hydrolytic acidity is expressed in mmol of hydrogen per 1 kg of soil. Due to the fact that hydrolytic acidity represents a less mobile part of hydrogen ions, it is less harmful to plants than exchangeable acidity. It is mainly used in the assessment of the sorption complex because hydrolytic acidity is an important indicator of the sorption properties of soils.

6.2.2 Soil alkalinity

Soil alkalinity is conditioned by the presence of alkaline salts, which easily hydrolyze and allow the formation of bases. Soil reaction values ($pH > 7$) are associated with the presence of carbonates in the soil.

Similar to soil acidity, we can talk about active and potential alkalinity.

1. Active alkalinity

Active alkalinity is caused by dissolved alkaline salts in the soil solution (Na_2CO_3, $NaHCO_3$).

$$Na_2CO_3 + 2H_2O \leftrightarrow H_2CO_3 + 2NaOH$$

2. Potential alkalinity

Potential alkalinity is caused by exchangeable Na^+ cations bound by the soil colloidal complex.

$$[PKK]^{Na^+}_{Na^+} + H_2CO_3 \leftrightarrow [PKK]^{H^+}_{H^+}\ Na_2CO_3$$

Active alkalinity is expressed in pH values, while potential alkalinity is expressed in mmol (Na^+) kg^{-1} of soil.

6.2.3 Buffering capacity of soils

Concerning normal plant growth, not only the pH value of the soil solution at a given moment is important, but also the dynamics of this value. Biological and chemical processes occurring in the soil can cause rapid and continuous changes in the reaction of the soil solution. However, soils do not respond to the mentioned processes and to the use of physiologically acidic and alkaline fertilizers with the same changes in the reaction of the soil solution. Some soils have the ability to prevent a change in soil reaction. This ability is known as the buffering capacity of soils. The buffering capacity of soils can be understood as the ability of soils to prevent changes in the soil reaction. This ability depends on the presence of certain substances in the solid or liquid phase of the soil, which are called buffers. In the soil solution, it is primarily the presence of weak acids and their salts. In the solid phase of the soil, the buffering capacity depends on the amount and composition of cations in the soil complex as well as on the content of $CaCO_3$ and organic substances.

6.2.4 pH values in soils

The pH values in soil profiles also differ depending on the degree of soil development, i.e. its values are directly influenced by the factors and conditions of the soil-forming process. Different pH values can be also found in individual diagnostic horizons. Liming, gypsing, and the application of ammonia fertilizers and fertilizers with elemental sulfur significantly influence the soil pH in the upper layers of the soil and later also in the entire soil profile.

The reaction of the soil solution of various soils varies in the range of pH 3.5 – 9 and above. Very low pH values occur in acidic marshy soils. Podzols have a reaction of the soil solution in the range of pH 4-6. The active reaction of Chernozems varies in the range of pH 5.8-7.8. Soils containing magnesium and sodium carbonates can have a pH of 8-9 or higher.

The soil pH varies between 5 and 7 in temperate climates and humid regions. Calcareous soils are found in more arid regions and are dominated by Ca^{2+}, Mg^{2+}, K^+ and Na^+ cations. Soil pH values in such soils range from 7 to 8.5. The precipitation of calcium carbonate occurs at pH 8.5, and therefore, soils with its presence do not exceed this value. Saline soils have a high concentration of exchangeable sodium. The pH values in such soils range from 8.5 to 10.5. In forest soils, the leaching of base cations from the soil profile is more intense and Al_3^+, $Al(OH)_2^+$ and H^+ cations begin to dominate it. The dominant cations and anions in acid sulfate soils are H^+ and SO_4^{2-}.

In acidic soils, pH is closely related to the degree of saturation of the sorption complex. However, this cannot be applied to all soils because hydrogen does not ionize equally easily with all mineral and organic colloids.

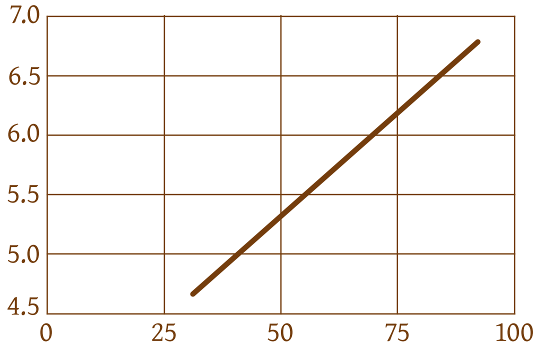


Figure 6.3: The average relationship between pH and percentage of base saturation (Millar et al., 1962)

The relationship between pH and the degree of saturation of the sorption complex is presented in Figure 6.3. The sorption complex is approximately 40% saturated at pH 5.

Table 6.3: Dominant cations depending on the range of soil pH and soil diversity (Šimanský, 2013)

Soil	Soil pH	Cations
Saline soils	8.5-10.5	Na^+
Calcareous soils	7-8.5	Ca^{2+}, Mg^{2+}, K^+ , Na^+
Arable soils	5-7	Ca^{2+}, Mg^{2+}, K^+ , $Al(OH)_2$, H^+
Forest soils	3.5-5.8	Al_3^+, $Al(OH)_2$, H^+
Acid sulphate soils	2-4	H_2SO_4

The relationship between pH and the degree of saturation is linear, i.e. for each increase of one pH unit, the degree of saturation of the sorption complex increases by 30%. The sorption complex is saturated to 85% at pH 6.5.

6.2.5 Plant requirements for soil pH

Table 6.4: Soil pH requirements of selected cultivated crops (Masaryk et al., 1980)

Crop	Optimal range	Crop	Optimal range
Plants more demanding on exchangeable pH with a lower limit of 6-6.5		Plants with lower requirements for exchangeable pH with a lower limit of 5-5.5	
Alfalfa	7.0-8.0	Vetch	5.7-6.5
Sugar beet	6.2-7.5	Potatoes	5.0-7.0
Rapeseed oil	6.8-7.5	Flax	5.5-6.8
Winter wheat	6.3-7.6	Oat	5.0-7.5
Spring barley	6.2-7.5	Rye	5.5-7.5
Maize	6.0-7.0	Carrot	5.5-7.0
Pea	6.0-7.0	Tabacco	4.8-8.2
Fruit trees	6.2-8.0	Meadow grasses	5.3-6.2

The soil can have an acidic, neutral or alkaline reaction. The soil pH is of great importance for growing plants and farming methods because it significantly affects the entire complex of chemical, physico-chemical, physical and biological properties of soils. The lower limit for our cultivated plants is pH 4 and the upper limit is 8. Exceeding these limits for some plants usually causes a complete crop failure. The pH interval, in which the soil reaction is not presented as a factor limiting the yields, has a different optimum range for crops. Some plants tolerate or even require slightly acidic soils, up to almost pH 5, others require neutral soils and do not thrive on acidic soils. On the other hand, some plants require alkaline soils and do not thrive on acidic soils either. Most cultivated plants prefer pH in the range of 6-7, which corresponds to the requirements for the release of nutrients from soil reserves and their maintenance in accessible forms. Individual plants can produce crops relatively well even with a wide range of pH values, which means that the plant can adapt to environmental conditions to a large extent.

A higher concentration of hydrogen ions affects the dynamics of nutrients in the soil and their release into accessible forms for plants. The optimal soil pH for the release and uptake of nutrients is around 6.5. Hydrogen ions at lower concentrations, mainly due to their high mobility and reactivity, have a favorable effect on the release of nutrients into the soil solution, especially during exchange reactions. At higher concentrations, i.e. at low soil pH values, the chemism of substances in the soil is significantly affected, and many of their chemical and physical-chemical transformations have a secondary effect on the availability of nutrients for plants (e.g. binding of phosphorus).

6.2.6 Liming and gypsing of soils

We adjust the soil pH of acidic soils by liming. According to the degree of acidity and the need for liming, we distinguish:

1. **Maintenance liming** – periodically repeated liming with doses of lime fertilizers that are necessary to maintain the soil reaction at an agronomically desirable level. It is carried out if soils have a carbonate content lower than 0.3% and the soil reaction is in the range of pH 6.0-6.7.
2. **Ameliorative liming** – liming of very acidic soils, where liming is the primary condition for increasing their fertility. Ameliorative doses of lime fertilizers to the soil are intended to modify unfavorable chemical and physical properties of soils.

The development of soil pH values on average for Slovakia indicates a gradual trend of acidification of our soils and a decrease in alkaline soils.

We adjust the soil reaction of alkaline soils by gypsing. During plastering, sodium carbonate is removed from the soil solution and sodium is displaced from the sorption complex by calcium to form a neutral salt – sodium sulfate:

$$Na_2CO_3 + CaSO_4 = CaCO_3 + Na_2SO_4$$

$$[PKK]^{Na^+}_{Na^+} + CaSO_4 \leftrightarrow [PKK]^{Ca^{2+}} + Na_2SO_4$$

Table 6.5: Representation of categories of soil pH in Slovakia in the XI. cycle of ATS (Kotvas, et al., 2007)

	Category of soil acidity (%)			
	acidic	slightly acidic	neutral	alkaline
Agriculture soil	21.2	36.9	28.1	13.8
Arable land	13.8	38.6	31.2	16.4
Hop fields	-	18.0	48.8	33.2
Vineyards	7.1	24.1	37.4	31.4
Orchards	10.1	27.5	31.7	30.7
Meadows and pastures	49.6	31.1	16.1	3.2

<table>
<tr><th colspan="11">A Summary of Soil Reaction, Lime Requirement, and Associated Plant-Nutrient Relationships</th></tr>
<tr><td>pH</td><td colspan="10">4.0 4.5 5.0 5.5 6.0 6.5 6.7 7.0 8.0 9.0 10.0 11.0</td></tr>
<tr><td>Acidity</td><td colspan="2">Very strongly acid</td><td>Strongly acid</td><td>Moderately acid</td><td>Slightly acid</td><td>Neutral</td><td>Weakly alkaline</td><td>Alkaline</td><td>Strongly alkaline</td><td>Excesive alkaline</td></tr>
<tr><td>Lime requirements</td><td colspan="2">Lime needed except for crops requiring acid soil</td><td colspan="2">Lime needed for all but acid-tolerant crops</td><td>Lime generally not required</td><td colspan="5">No lime needed</td></tr>
<tr><td>Occurence</td><td>Rare</td><td>Frequent</td><td colspan="2">Very common in plow soil in humid climates</td><td>Plow soils when limed</td><td colspan="3">Common in sub-humid and arid regions</td><td colspan="2">Black alkali occurs in limited areas in arid regions</td></tr>
<tr><td rowspan="6">Fertilitiy conditions</td><td colspan="3">Phosphates fixed</td><td colspan="3">Phosphates soluble</td><td colspan="4">Phosphate fixed</td></tr>
<tr><td colspan="6">Lime nad potash leach</td><td colspan="4">Boron, iron, manganese, and potash may be deficient</td></tr>
<tr><td colspan="3">Iron, aluminium, and manganese are soluble</td><td colspan="7">Lime present in optimum quantities</td></tr>
<tr><td colspan="10">Desirable bacterial activity</td></tr>
<tr><td colspan="7">Fungi thrive</td><td colspan="3">Exces boron my occur in limited areas</td></tr>
<tr><td colspan="4">Bacteria languish</td><td colspan="6">Nitrogen freely fixed</td></tr>
</table>

Figure 6.4: A summary of soil pH, lime requirement, and associated plant-nutrient relationships (Millar et al., 1962)

6.3 Carbonates in soil

Carbonates represent an important component of the mineral content of the soil. They significantly affect all soil properties. They participate in creating the buffering capacity of the soil. Ca^{2+} cation significantly affects the value of the electrokinetic potential of colloids, soil reaction, and the saturation of the sorption complex. It is also important in the formation of humus in the soil.

In soils, carbonates occur mainly in the form of $CaCO_3$, less in the form of $MgCO_3$. Other carbonates are found in larger quantities in saline and alkaline soils (especially Na_2CO_3, $NaHCO_3$). The leaching of Ca^{2+} and Mg^{2+} carbonates from the soil profile is a manifestation of soil degradation.

Carbonates in the soil can be of primary (parent rock, soil-forming substrate) or secondary (Ca – fertilizers) origin. If they are of primary origin, their content usually increases with the depth or occurs only in the lower part of the profile. Arable soils may contain $CaCO_3$ due to intensive liming. In such a case, calcium carbonate is generally present in higher quantities in the upper layers while the lower ones have a lower content or do not contain it at all.

6.4 Soil solution

Meteoric water that penetrates the soil profile and groundwater that also reaches the soil profile by upwelling contain soluble air components (CO_2, O_2), nitrates, ammonia, and soluble products of chemical weathering of rocks and biochemical decomposition of organic substances.

Soil water with dissolved components intensively participates in various chemical and biochemical processes in the soil. At the same time, it acts as a very good solvent of substances, further as a dispersion medium for the formation of colloidal systems, and also because it contains hydrogen ions, which are released during the dissociation of water molecules.

The water found in the soil, which contains various substances in a dispersed state, represents the liquid phase of the soil or soil solution. This part of the soil is the most mobile and active and is of great importance in terms of the soil-forming process as well as plant nutrition. The quantitative and qualitative characteristics of the soil solution are determined by its composition, concentration, and properties.

6.4.1 Composition and concentration of soil solution

The composition and concentration of the soil solution are the results of the whole series of processes that take place in the soil medium. The soil solution contains mineral, organic and organo-mineral substances in ionic, molecular, and colloidal forms. In addition, various gases are also present in the soil solution.

The vast majority of soils have a low concentration of salts in the soil solution. In saline soils, the concentration is high. The mineral component of the soil solution is represented by various salts, which are evaluated on the basis of the anionic and cationic composition.

1. Anionic composition of the soil solution

Anions of HCO_3^-, CO_3^{2-}, NO_3^-, SO_4^{2-}, Cl^-, $H_2PO_4^-$, and HPO_4^2 represent up to 90% of the soil solution. The content of HCO_3^- depends on the intensity of the processes of oxidation of organic matter, during which a significant amount of carbonic acid is formed. In soils that contain $CaCO_3$ or have been limed, a large amount of bicarbonates can be always found:

$$CaCO_3 + H_2CO_3 \rightarrow Ca(HCO_3)_2$$

Higher content of CO_3^{2-} anion is found in saline soils that contain Na_2CO_3. The amount of NO_3^- in the soil solution depends on the biological activity (nitrification) and the intensity of fertilization. The content of this ion in the solution is significantly affected by resorption by plants and microorganisms, leaching by meteoric water, denitrification, and others. Nitrite anion (NO_2^-), which is also a product of the biological activity, is practically not present in the solution (under normal conditions, it is quickly oxidized to NO_3^-). The content of Cl^- anion in non-saline soils is very low as most chlorides are well soluble in water and are leached from the soil. Higher content of chlorides in the soil solution can be detected immediately after the application of the chloride form of potassium fertilizers. Higher content of chlorides in the solution indicates soil salinization. The content of sulfate anion in the soil depends on the activity of sulfur bacteria (they oxidize H_2S, which is released during the rotting of organic residues to H_2SO_4). Sulfates also enter the soil through the application of superphosphate and the sulfate form of potassium fertilizers. The sulfate content is low (20-40 mg kg^{-1}) in non-saline soils. There are much more of them in saline soils. The content of phosphorus anions in the soil solution is very low, it does not exceed 1-2 mg of P_2O_5 per 1 liter. This condition is caused by the fact that phosphates very quickly undergo chemosorption in the soil. Therefore, their solubility is very low.

2. Cationic composition of the soil solution

The source of cations in the soil solution are silicates, aluminosilicates, ferrisilicates, carbonates, sulfates, mineral fertilizers, and exchangeable sorbed cations. There are various cations in the soil solution, the most important of which are Ca^{2+}, Mg^{2+}, Na^+, K^+, NH_4^+, H^+, Al^{3+}, Fe^{3+}, and Fe^{2+}. In non-saline soils, Ca^{2+} usually predominates over other cations. In saline soils, there is a larger amount of Na^+, but in our conditions, it does not exceed the calcium content. The amount of Mg^{2+} makes up approximately half to a fifth of the calcium content. In saline soils, the presence of magnesium can be significantly higher. Cations of alkaline metals are represented in the soil solution by a smaller proportion than cations of alkaline earth (there is more sodium than potassium). In unfavorable conditions for nitrification, NH_4^+ cation is present in the soil solution. Fe^{3+} and Al^{3+} cations are also present in the soil solution of very acidic soils. Fe^{2+} cation is present in marshy soils, primarily in the form of organo-mineral compounds. Acidic soils contain a large amount of H^+ cation. Concerning the organic compounds, the soil solution contains water-soluble substances of organic residues and products of their decomposition as well as the components of the vital activity of plants and microorganisms, and humic substances mostly in the form of organo-mineral compounds.

Properties of the soil solution

The reaction of the soil solution is conditioned by the concentration (activity) of hydrogen ions. It has a great influence on chemical, physico-chemical and biological processes in the soil as well as on the development of plants. An important property of the soil solution, which is closely related to the reaction, is buffering, i.e. ability to withstand pH changes. The carriers of the buffering capacity of the soil solution are primarily buffering systems. Oxidation-reduction reactions are of great importance in the soil. They are closely related to the activity of various organisms in the soil. The ratio of oxidizing and reducing substances determines the oxidation-reduction conditions of the soil solution. To characterize these conditions, the oxidation-reduction (redox) potential (mV) is determined.

The osmotic pressure of the soil solution is of great importance in plant nutrition. If the osmotic pressure of the soil solution is the same as the pressure of the cell sap or higher, the plant stops receiving water and dies. The value of the osmotic pressure depends on the concentration of the soil solution and the degree of dissociation of substances.

7

Soil nutrients

Several centuries ago, people already knew that animal excrements, ash, blood, and other substances have a stimulating effect on plant growth. It was only later discovered that the “nutritional effect“ is the result of elements found in these materials. So far, 15 elements are known that most plants must have for normal development, flowering and maturation (White, 2015).

These elements are divided into:

1. **Macronutrients** – which are needed in a relatively large concentration (>1,000 mg kg^{-1}).
2. **Micronutrients** – which are needed in small concentrations (<1,000 mg kg^{-1}).

Almost all cultivated plants receive most of the nutrients from the soil, which are found there in the form of ions (cations, anions) of individual elements if the compounds of the elements are dissolved (dissociated) in water. The most commonly accessible ionic forms of macro and micronutrients are listed in Table 7.1.

Table 7.1: The most commonly accepted ionic forms of macro and micronutrients (White, 2009)

Macronutrients	Ionic forms	Micronutrients	Ionic forms
Carbon	HCO_3^-, CO_3^{2-}, CO_2	**Iron**	Fe^{3+}, Fe^{2+}
Hydrogen	H^+	**Manganese**	Mn^{4+}, Mn^{2+}
Oxygen	O_2, H_2O and many anions such as OH^-, NO_3^-, SO_4^{2-}	**Zinc**	Zn^{2+}
Nitrogen	NH_4^+, NO_3^-	**Copper**	Cu^{2+}
Phosphorus	$H_2PO_4^-$, HPO_4^{2-}	**Boron**	BO_3^{3-}, HBO_3^{2-}
Sulfur	SO_4^{2-}	**Molybdenum**	MoO_4^{2-}
Calcium	Ca^{2+}		
Magnesium	Mg^{2+}		
Potassium	K^+		

Sources of elements in soil

During photosynthesis, the leaves and green parts of plants absorb carbon and oxygen from the atmosphere in the form of CO_2, which is important for the synthesis of carbohydrates. Oxygen and hydrogen enter the plant in the form of water, either from the atmosphere or from the soil. Nitrogen as (N_2) is present at 78% in the Earth's atmosphere and can enter the soil and plants through biological fixation. A small amount of mineral nitrogen enters the soil through precipitation or dry deposition. The most important source of elements in the soil is the mineral component formed by primary or secondary minerals. Macro and microelements are released into the soil during the weathering of rocks. Minerals and rocks are also a source of essential elements (e.g. chromium, selenium, iodine, but also cobalt) needed for healthy animal nutrition. On the other hand, there are also elements (arsenic, mercury, cadmium, lead, nickel) that can have a toxic effect if their high concentration accumulates in soil, plants, animals, and humans. So, it should not be forgotten that if their concentration exceeds the background values, they become potentially harmful and to a greater extent can reach the food chain. We draw attention to this in connection with the fact that some enter the soil consciously – through anthropogenic activity. For example, cadmium enters the soil with natural apatites as a source of P fertilization.

7.1 Macronutrients

Elements that are needed in higher concentrations for normal plant development are called **macronutrients** – usually, their contents in plants are more than 500 ppm. These include: nitrogen, phosphorus, potassium, calcium, magnesium and sulfur. The main roles of these macronutrients in the plant are summarized in Table 7.2. and the uptake of macronutrients to produce a crop of some plants is presented in Table 7.3.

Table 7.2: Basic macronutrients and their role in plants (Foth, 1990)

Element	Role in the plant
Nitrogen	An essential element of all proteins, chlorophyll, coenzymes and nucleic acids.
Phosphorus	An important element in energy transfer as part of ATP. An essential element of many proteins, coenzymes, nucleic acids and metabolic substrates.
Potassium	An important function in the regulatory mechanism of photosynthesis and protein synthesis.
Calcium	An important structural element of cell walls. It has a significant role in the structures and permeability of cell membranes.
Magnesium	A basic element of chlorophyll and an important activator of enzymatic reactions.
Sulfur	An important element of vegetable proteins.

Table 7.3: Average uptake of macronutrients by the harvest of selected plants (source: Foth, 1990[1], Vaněk et al., 2013[2])

Crop	Harvest (in ha^{-1})	N	P	K	Ca	Mg	S
Barley (straw)	1[1]	15[1]	3[1]	25[1]	8[1]	2[1]	4[1]
Wheat (straw)	1[1]	20[1]	3[1]	29[1]	6[1]	3[1]	5[1]
Cabbage	20[1]	130[1]	16[1]	108[1]	20[1]	8[1]	44[1]
Sugar beet	15[1]	60[1]	9[1]	42[1]	33[1]	24[1]	10[1]
Wheat (grain)	1[2]	22-26[2]	4.4-6.2[2]	16.6-21[2]	2.8-5.7[2]	1.2-3[2]	-
Maize (grain)	1[2]	22-26[2]	4.4-6.6[2]	21-33[2]	4.3-7.1[2]	4-6[2]	-
Sugar beet	1[2]	4-5[2]	0.7-0.9[2]	5-8.3[2]	0.7-1.4[2]	0.6-1.2[2]	-

Table 7.4: Contents of total nitrogen in relation to the production potential of soils (Bielek, 2014)

Soil type	Point value of the production potential of soils	% of total nitrogen
Regosols	0 – 20	0.20
Podzols	20 – 40	0.27
Gleysols	20 – 40	0.25
Rendzic Leptosols	30 – 50	0.20
Cambisols	40 – 60	0.17
Planosols / Stagnosols	50 – 70	0.17
Albiluvisols	60 – 80	0.16
Fluvisols	65 – 85	0.18
Luvisols	70 – 90	0.14
Chermozems	75 – 95	0.19
Mollic Fluvisols	80 – 100	0.20

7.1.1 Nitrogen

It is the most important component of proteins and nucleic acids. It is a part of practically all structures of living organisms and, through them, also of chemical and biochemical structures of the soil medium. Vanek et al. (2013) stated that the total nitrogen content in the soil is usually in the range of 0.1-0.2%. Concerning topsoil, it represents 3,000-6,000 kg of N in the topsoil per 1 ha. In the topsoil of high-quality and highly productive soils, there is a significantly higher content of nitrogen compared to low-production soils. The largest reserves of nitrogen are associated with Chernozems and mollic Fluvisols. On the other hand, relatively low contents of nitrogen are observed in brown soils, Luvisols, and Regosols. Significant differences in nitrogen content are observed within the soil profile. The highest content of nitrogen is concentrated in the topsoil layer or on the A horizon. As stated by Bielek (1998), there is almost 700,000 ha of soil in Slovakia with a low and very low content of total nitrogen, approx. 1.3 million ha with a medium content and 400 thousand ha with a high and very high content of total nitrogen. The contents of total nitrogen in relation to the production potential of soils are presented in Table 7.4. The assessment of total nitrogen in agricultural soils of Slovakia is introduced in Table 7.5.

Nitrogen cycle

Nitrogen is a highly mobile element that circulates between soil, atmosphere, and living organisms. This nitrogen cycle is caused by many factors and processes, some of which are physical, chemical, or biological in nature. a dynamic balance is maintained between the different forms of nitrogen.

Table7.5: Assessment of total nitrogen in agricultural soils of Slovakia (Bielek, 2014)

Category of nitrogen content	Nitrogen content in %
Very low	up to 0.12
Low	0.121 – 0.16
Medium	0.161 – 0.24
High	0.241 – 0.34
Very high	above 0.34

Nitrogen is fixed by various processes and released into the atmosphere by microbiological activity and chemical reactions. It is clear from the nitrogen cycle that the source of nitrogen for plants is nitrogen from industrial fertilizers, organic fertilizers, soil organic mater as well as the fixation of atmospheric nitrogen by symbiotic and non-symbiotic bacteria and electrical discharges in the atmosphere (Fig. 7.9).

As stated by Bielek (2014), we find mineral and organic nitrogenous compounds in the structure of total nitrogen. Since no parent rock contains nitrogen, all forms of nitrogen in the soil originate in the biological enrichment of the soil with nitrogen from the air. Nitrogen enters the soil from the atmosphere through fixation by microorganisms, fertilizers, and in the form of falls. Plant residues are also a source of nitrogen. Vanek et al. (2013) reported that the supply of nitrogen to the soil through biological fixation can be quite variable. For example, a high-quality alfalfa stand is able to fix 200-250 kg of N ha^{-1} year^{-1}, while annual legumes fix approx. 40-80 kg of N ha^{-1} year^{-1}. Soil microorganisms influence slightly a biological fixation (approx. 5 kg of N ha^{-1} year^{-1}). Concerning the current density of farm animals, the supply of N to the soil is at the level of approx. 20 kg of N ha^{-1} year^{-1}. In addition, the application of nitrogen in the form of mineral fertilizers has decreased significantly in the last decade.

Of course, there are significant differences between companies, but the average dose of N per 1 ha is about 70 kg per year. The supply of nitrogen to the soil is also ensured through precipitation, which represents approx. 15 – 20 kg of N ha^{-1} year^{-1}. A part of nitrogen enters the soil through post-harvest residues.

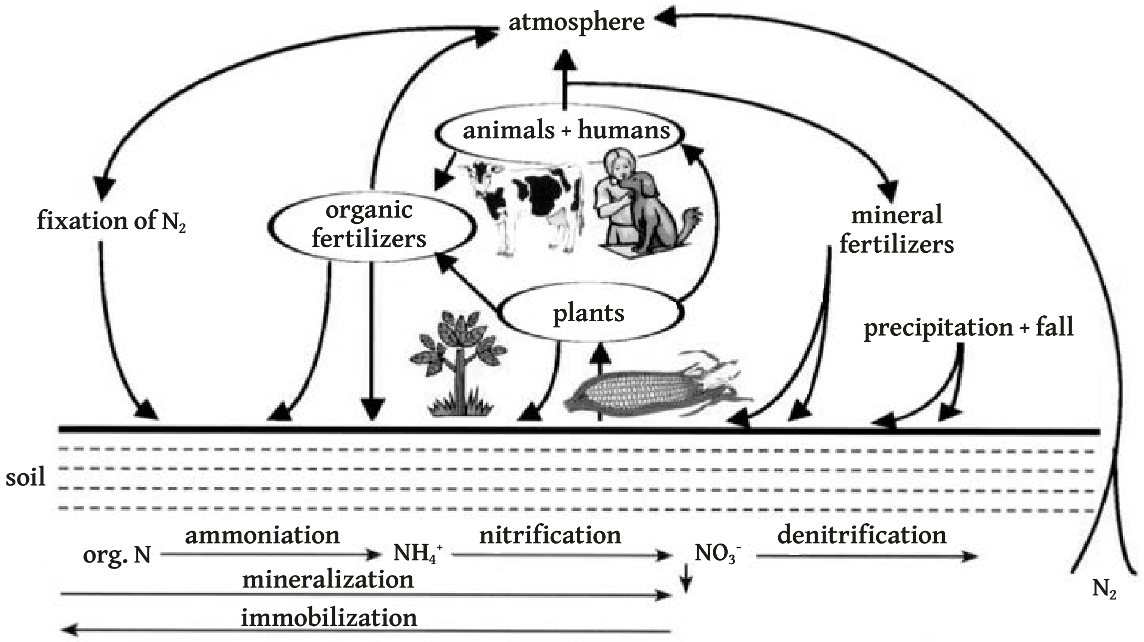
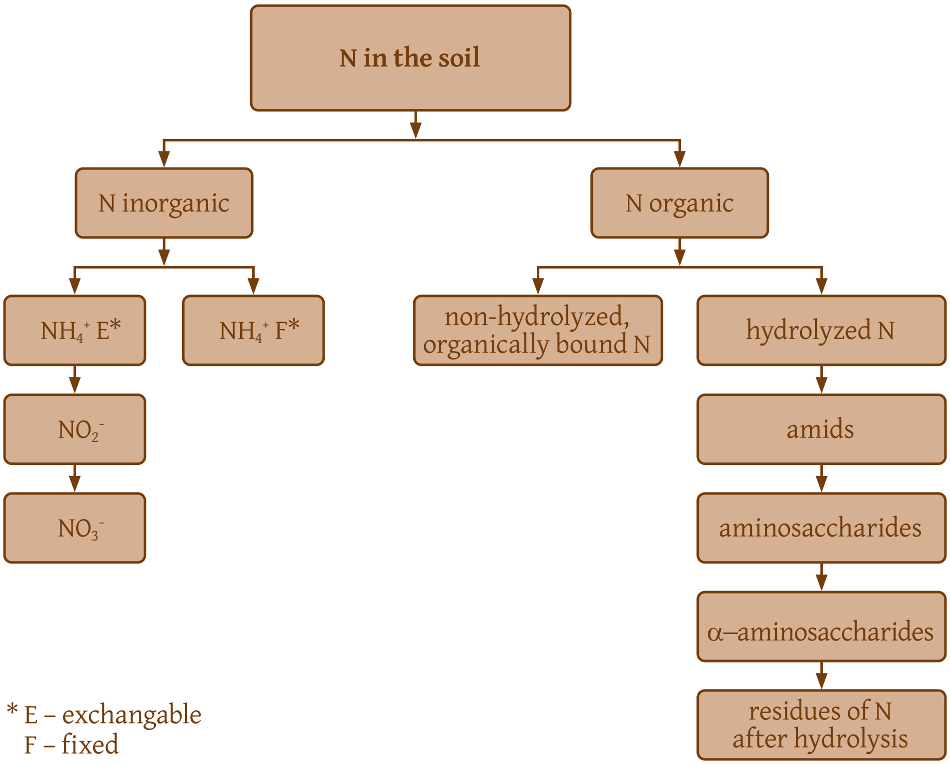


Figure 7.1: Nitrogen cycle in nature (Vaněk et. al., 2013)



* E – exchangable
F – fixed

Figure: 7.2: Forms and fractions of nitrogen in the soil (Fecenko and Ložek, 2000)

The content of total nitrogen is a relatively constant value since nitrogen is to a decisive extent incorporated into organic compounds, which are relatively difficult to decompose microbiologically and chemically. The total nitrogen content in the soil is the sum of organic and inorganic nitrogen. The individual forms and fractions of nitrogen in the soil are presented in Figure 7.2.

Organic nitrogen in soil

It forms a significant part of the total nitrogen in the soil and its share is 98-99%. According to the ability to undergo mineralization, we divide organic forms of nitrogen in the soil into hydrolyzable and nonhydrolyzable. **Nonhydrolyzable organic nitrogen** – it is difficult to decompose and represents the majority of nitrogen in the soil. **Hydrolyzable forms of nitrogen** – represent the most important precursors of mineral nitrogen in the soil, and according to their ability to undergo decomposition, they are divided into easily hydrolyzable and difficult to hydrolyzate. Higher contents of easily hydrolyzable nitrogen in the soil in the pre-spring period are an important argument for the readiness of the soil for mineralization of soil reserves of nitrogen to ensure the nitrogen nutrition of plants in the coming growing season. Hardly hydrolyzable forms of nitrogen in the soil are mainly represented by complex heterocyclic compounds and nitrogen forming bridges between benzene nuclei. Therefore, they are also less mineralizable and contribute little to the nitrogen nutrition of plants.

Nitrogen bound in organic compounds is relatively less accessible to cultivated plants. It becomes available in the soil environment through chemical, biological, and thermal mineralization (Kováčik, 2014).

Chemical mineralization – is the decomposition of organic substances by chemical compounds that are formed in the soil by the action of microorganisms, plants (root secretions), and substances entering the soil by anthropogenic activity (fertilizers, pesticides).

Thermal mineralization – is the mineralization of organic substances through their combustion (fire).

Biological mineralization – is a two-stage process of converting organic substances into ammonia.

- 1st stage (**aminization**):
  - organic matter $\rightarrow R - NH_2 + CO_2 + energy + other\ products$,
- 2nd stage (**ammonization**):
  - $R - NH_2 + HOH \rightarrow NH_3 + R - OH + energy$.

In the process of microbial mineralization of organic compounds, energy, carbon dioxide, and mineral substances, including by-products, are released while mineralization is often associated with synthesis and immobilization. The C : N ratio determines, which of the given processes will dominate in the soil (Kováčik et al., 2012). If it is higher than 30:1, immobilization prevails over mineralization. If it is less than 20:1, mineralization prevails over immobilization, and if the C : N ratio is in the range of 20-30:1, we are talking about a balance between mineralization and immobilization. This is a consequence of the ratio between the contents of C and N in the biomass of microbial plasma (45% C : 1.5% N = 30 : 1). Kovacik et al. (2012) stated that in the soils of Slovakia, such as Chernozems and Mollic Fluvisols, the C : N ratio ranges from 6-12:1, and in soils such as Luvisols and Albiluvisols, this ratio is wider, about 15-20:1.

Without the active action of the grower, approx. 20-25 kg of N ha^{-1} becomes available in poor Podzols per year through mineralization. In Luvisols and loamy soils, it amounts to approx. 50-75 kg of N ha^{-1} and in Chernozems and Mollic Fluvisols, approx. 75-160 kg of N ha^{-1}. If the grower is not sure that after plowing in the post-harvest residues their significant decomposition will occur by the time of sowing the next crop, it is advisable to apply various fertilizers with nitrogen content to reduce the C : N ratio (to a value of approx. 25:1). Kováčik (2014) points out that agricultural production practice often forgets the fact that the application of fertilizers not only supplies the soil with nutrients but also initiates more intensive mineralization of organic substance. Consequently, it is observed that after the supply of 1 kg of N in mineral fertilizers, the content of Nan (inorganic nitrogen) in the soil is higher not by the expected 1 kilogram, but by 1.2-2.25 kg. This phenomenon is called "**priming effect**" (extra nitrogen). The effect of applied nitrogen in agricultural fertilizers is even higher. 1 kg of N an releases up to 1.5 kg of Nan, which means that after applying 10 kg of Nan in farmyard manure, we can expect an inorganic nitrogen content of 25 kg.

Inorganic nitrogen in soil

It represents only a small part of the total nitrogen in the soil, about 1-2% (Fecenko and Ložek, 2000). However, even this small amount is subject to large seasonal changes. It consists of ammoniacal nitrogen (free and fixed in the grid of soil minerals), nitrate nitrogen, nitrite nitrogen, and nitrogen oxides as well as intermediate products of microbial processes such as hydroxylamine (HN_2OH) and nitramide ($N_2H_2O_2$). A special fraction of soil mineral nitrogen are its gaseous forms, especially its oxides.

From the point of view of plant nutrition with nitrogen, **ammonia** and **nitrates** are the most important. Since both forms are the product of biological processes in the soil, their contents depend on seasonal changes. Their highest contents are observed in the so-called **spring** and **autumn maximum**.

Ammonia is oxidized in the process of **nitrification** in biologically active soils. It is an oxidation process where ammonium nitrogen is gradually oxidized by autotrophic microorganisms to $N\text{-}NO_3^-$. The process takes place in two stages.

- 1st stage (**nitritation**): provided by nitrifying bacteria of the genera: *Nitrosomonas*, *Nitrosocystis*, *Nitrosospira*:
  - » $2NH_3 + 3O_2 \rightarrow 2NO_2^- + 2H^+ + 2H_2$,
- 2nd stage (**nitratation**): provided by nitrating bacteria of the genus: *Nitrobacter*:
  - » $2NO_2^- + O_2 \rightarrow NO_3^-$

Nitrification – is a process very sensitive to external conditions. At a temperature below 5 °C, it almost stops. It requires sufficient air in the soil and a weakly acidic to slightly alkaline medium.

Denitrification – by contrast, is a reduction process where nitrates are reduced to nitrogen oxides, even to elemental nitrogen in the presence of organic substances. In our conditions, denitrification caused by facultatively anaerobic microorganisms, which use the oxygen of nitrates during decomposition, prevails.

$$24HNO_3 + 5C_6H_{12}O_6 = 12H_2 + 30O_2 + 42H_2O + energy$$

The lack of oxygen in the soil, the presence of nitrates, and the abundance of readily degradable organic matter are the three conditions for denitrification, i.e. the reduction of nitrate in the presence of amides without the participation of micro-organisms, is less important.

Denitrification can result in relatively large nitrogen losses. During the nonvegetation period, when there is an increased risk of denitrification losses, as well as N-NO_3^- leaching from the topsoil, it is necessary to decrease the amount of N-NO_3^- in the soil. Nitrates are most intensively leached at the end of winter and in spring, when the soil has no vegetation cover. During this period, 70-90% of the nitrate nitrogen from the total amount of leached nutrients is carried by the leachate water (Lehotský, 1984). Average values of leached nitrate nitrogen range from 5 to 55 kg.ha^{-1}.year^{-1}. The total nitrogen balance shows that the nitrate form of nitrogen accounts for 90-97%, the ammonium form 0.5-3%, and the remainder is made up of other soluble nitrogen compounds (Fecenko and Ložek, 2000).

Ammonia volatation is the release of ammonia from the soil environment into the atmosphere. It occurs especially with the application of higher rates of nitrogen fertilizers – particularly with amide, ammoniac, and ammonium forms of nitrogen. Volatilisation can lead to losses of up to 25% of nitrogen from such fertilisers. Table 7.6 summarizes the total nitrogen losses on agricultural soils in the Slovak Republic.

Table 7.6: Average nitrogen losses from agricultural land in Slovakia (Kováčik, 2013)

Loss	Causes of losses	Kg ha^{-1} year^{-1} N
Inefficient	Denitrification	20 – 25
	Volatilization	10 – 15
	Leaching	approximately 10
	Erosion	5 – 10
Efficient	Harvest (farm crops/ horticultural crops)	approximately 85 – 110/100 – 140

A special form of mineral nitrogen in the soil is non-biologically fixed nitrogen. It is formed by the process of NH_4^+ ions entering and binding into the interlayer spaces of the crystal lattice of clay minerals. It is nitrogen that is inaccessible to plants, but can be displaced into the soil solution, for example, by additions of potassium. In addition, molecular ammonia (NH_3) can be fixed to soil organic matter and may represent a fixed form of nitrogen in the soil. It is a relatively strongly stabilized form of nitrogen resistant to availability to plants (Bielek, 1998).

7.1.2 Phosphorus

In Slovak soils, the total amount of phosphorus ranges from 0.02 – to 0.2% (Michalík and Hanáčková, 2011). Higher phosphorus content is generally found in soils with higher soil organic matter content, while sandy soils with low soil organic matter content have a low phosphorus supply.

Organic phosphorus in soil

It is an integral part of soil organic matter. The amount of phosphorus bound in this way usually accounts for 30-50% of the total phosphorus in soils (Vaněk et al., 2013). A substantial part of organic phosphorus is phytin (up to 50% of organic P), phospholipids, nucleic acids, nucleoproteins, and phosphorylated lipids, which occur in the root mass and enter the soil through post-harvest residues, manures, and other organic fertilizers as well as amendments. Much of the organic phosphorus in the soil is the result of biological sorption by soil microorganisms that immobilize phosphorus in their bodies. In addition to temperature, humidity and pH, the intensity of biological sorption is also determined by the C:P ratio (Kováčik, 2014). If this ratio exceeds 300:1, phosphorus immobilization occurs. At a ratio of 200 – 300:1, a balance between immobilization and mineralization occurs, and at a ratio closer than 200:1, phosphorus release occurs.

Inorganic phosphorus in soil

Phosphorus minerals such as apatites are the primary source of inorganic phosphorus in soil. Phosphoric acid salts are a secondary source of inorganic phosphorus in slightly acidic to alkaline soils.

Phosphorus cycle

The phosphorus cycle is shown in Figure 7.3. The weathering of phosphorus-rich minerals (especially apatite) releases dihydrogen phosphates ($H_2PO_4^-$), which are thus released into the soil solution. $H_2PO_4^-$ is further immobilized when roots and microorganisms take up and convert phosphorus into organic compounds. This results in a significant increase of phosphorus in soils in the form of organic phosphorus.

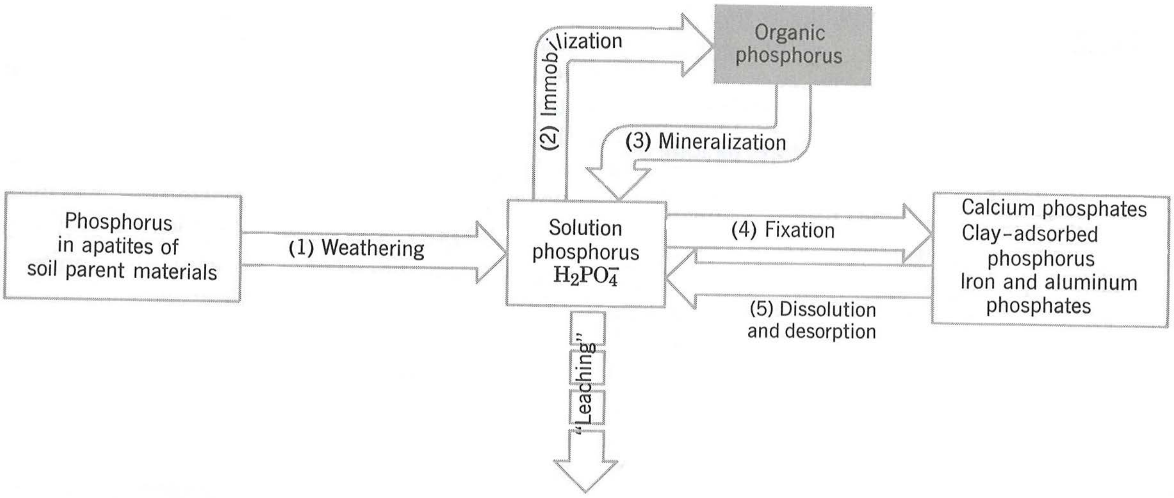


Figure 7.3: Phosphorus cycle (Foth, 1990)

Usually, 20-30% of the organic phosphorus in the topsoil of mineral soils is formed in this way. The organic phosphorus is mineralized by microorganisms and released back into the soil solution as $H_2PO_4^-$. Dihydrogen phosphates react very quickly with other ions in the soil solution, leading to their precipitation and adsorption to mineral colloids; they convert the phosphorus to an unavailable or solid form. As a result, most phosphate ions from organic phosphorus mineralization or mineral weathering can be converted to an unavailable form before plants have a chance to take up phosphorus and before it can be leached out of the soil. All these processes are very significantly affected by soil pH.

The mobility of phosphorus in soil is relatively small and may migrate within 20-30 cm from the point of the fertilizer application into the soil. This is one of the reasons why phosphorus losses from the soil are very small. In sandy soils, the losses are at the level of 3 – 5 kg ha^{-1} $year^{-1}$, in loamy soils, it is about 2 – 3 kg ha^{-1} $year^{-1}$ and in clay soils, the losses are well below 2 kg ha^{-1} $year^{-1}$ (Richter and Hlušek, 1994). The solubility and thus the availability of phosphorus in the soil is regulated by complex mechanisms.

Chemically mediated phosphorus sorption occurs in an acidic and alkaline soil environment, while the water-soluble dihydrogen phosphate anion $H_2PO_4^-$ precipitates with divalent or trivalent cations to less soluble or insoluble phosphates. Chemical phosphorus sorption on acidic soils takes the following reactions:

$$[PSK]^{H^+}_{H^+} + Ca(H_2PO_4)_2 \rightarrow [PSK]^{Ca^{2+}} + 2H_3PO_4$$

$$Fe(OH)_3 + H_3PO_4 \rightarrow FePO_4 + 3H_2O$$

$$Al(OH)_3 + H_3PO_4 \rightarrow AlPO_4 + 3H_2O$$

$$[PSK]^{Al^{3+}} + H_3PO_4 \rightarrow AlPO_4 + [PSK]^{H^+}_{H^+} H^+$$

This sorption can be avoided by liming the soil and sufficient fertilization with manures, organic fertilizers, and amendments. Chemical sorption in carbonate soils proceeds according to the following reactions:

$$2Ca(HCO_3)_2 + Ca(H_2PO_4)_2 \rightarrow Ca_3(PO_4)_2 + 4H_2CO_3$$

$$[PSK]^{Ca^{2+}} + 2CaHPO_4 \rightarrow [PSK]^{H^+}_{H^+} + Ca_3(PO_4)_2$$

$$[PSK]^{Ca^{2+}} + 2CaHPO_4 \rightarrow [PSK]^{H^+}_{H^+} + Ca_3(PO_4)_2$$

The water-soluble calcium dihydrogen phosphate reacts with calcium bicarbonate or calcium in the sorption complex to form tricalcium phosphate. This conversion of water-soluble phosphorus to insoluble phosphate is called **phosphorus retrogradation**. The chemical sorption of phosphorus is evaluated negatively. By changing the soil pH, a part of the phosphorus can be made available to plants. It is recommended to apply physiologically acid fertilizers to alkaline soils.

Physiochemically mediated phosphorus sorption is formed in a neutral and slightly acidic soil environment by binding phosphate anions on the surface of clay colloids, which show positive charges after OH^- group dissociation. The phosphate anions can be exchanged from the sorption complex into the soil solution for other anions, thus replenishing the phosphorus removed by plants from the soil solution.

Biologically mediated phosphorus sorption – is the action of soil microorganisms that fix phosphorus to their microbial biomass or to newly form organic compounds containing phosphorus. After the mineralization of the microbial biomass, phosphorus is released, converted to an inorganic form, and becomes available to plants.

7.1.3 Potassium

The average potassium concentration is approximately 2.6% in the earth's crust. In the topsoil layers, there is an average of 40,000 – 50,000 kg ha^{-1} of potassium, with up to about 95-99% in inorganic forms (Foth, 1990). The potassium content in the soils of Slovakia ranges from 0.8 – 3.2%, with the lowest content in sandy and peat soils (0.15 – 0.30%); while clay soils can contain up to about 4% or more K (Fecenko and Ložek). The main source of potassium for plants, apart from potassium mineral fertilizers, is the potassium released by weathering of minerals with a higher potassium content, e.g. from micas, feldspars, and clay minerals, especially of the 2:1 type – illite, glauconite. The rate of potassium release by weathering, however, depends not only on the potassium content but also on the mineral structure and lattice strength. Figure 7.4 shows the area, from which the K release occurs after the expansion of a 2:1 crystal lattice in vermiculite. For example, biotite with a lower potassium content releases potassium more rapidly than muscovite with a higher K content. Similarly, Ca – mica, weathers faster than K – mica.

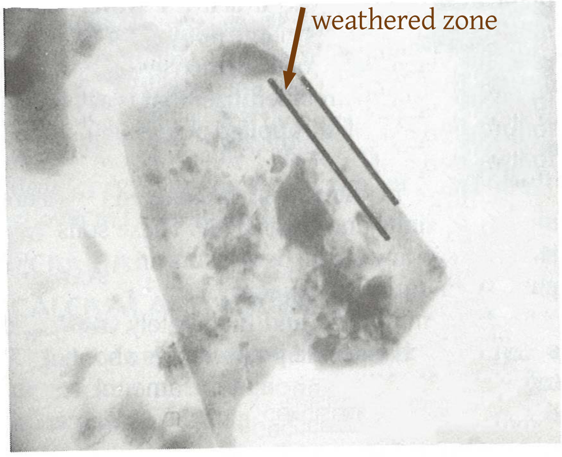


Figure 7.4: Weathering layer of vermiculite – vermiculite crystal lattice expands and K is released (Foth, 1990)

Figure 7.5 shows the potassium cycle and, as noted above, potassium in soil occurs mainly as primary and secondary silicates. There are three forms of potassium in soil: **non-exchangeable, exchangeable, and water-soluble** (Vaněk et al., 2013).

Non-exchangeable potassium – this form of potassium represents 95% of the total potassium in the soil (Kováčik, 2014). Non-exchangeable potassium is fixed in the primary minerals lattice, such as potassium aluminosilicates, especially mica (biotite, muscovite) and feldspar (orthoclase, sanidine, albite, anorthite). During their weathering, K is gradually released and, at the same time, secondary minerals (clay minerals) are formed. Potassium is a part of the lattice of these minerals. Scheffer and Schachtschabel (1970) reported the following K concentrations in clay minerals: illite contains 3-6%, gluaconite 3-7%, vermiculite 0-2%, and montmorillonite 0-0.5%.

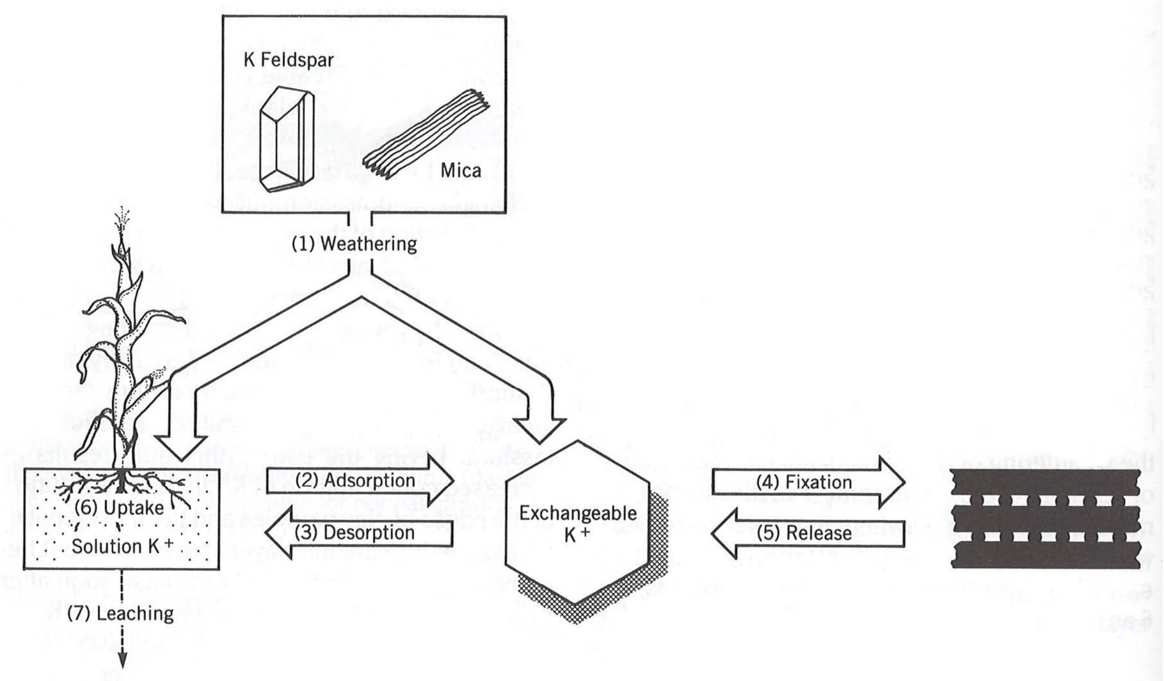


Figure 7.5: Potassium cycle in soil (Foth, 1990)

K, which is trapped between the layers of clay minerals – the so-called **fixed potassium**, is classified as a non-exchangeable form of potassium. After a certain period, it can convert to an exchangeable form. **Fixed potassium** represents that part of K which is located between the layers of clay minerals and is not replaceable. The same happens to clay minerals, where K is very tightly fixed. Potassium fixation in soils with a high concentration of clay minerals can reach high values and is higher in dry conditions; it is significantly affected by alternating periods of wet and dry conditions. After the wet period, the lattice is released, and so, part of the fixed K can be converted into exchangeable forms. Therefore, in spring, there is usually a higher amount of available potassium and increased plant uptake. On clay minerals, there are three differently located centres (Fig. 7.6). The places are on the outer surfaces (v – position), on the corners (r – position) and in the interlayer space (m – position). The potassium cations are fixed in the m – position.

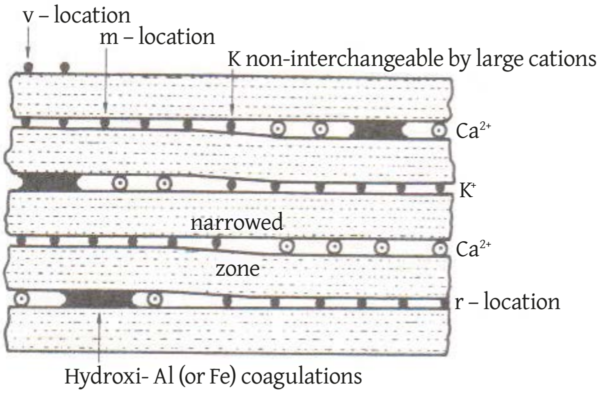


Figure 7.6: Model of expanded silicate layers with fixed and adsorbed cations in different places (Rich, 1968)

Organic fixed potassium in the soil represents only a very small share of the total potassium. Potassium present in microorganisms and organic residues is released and becomes available to plants only after their mineralization. This rate is about 20 – 50 kg ha^{-1} of potassium per year, depending on the amount of organic matter in the soil.

Exchangeable potassium is a K^+ cation that is adsorbed on the soil sorption complex as a result of physico-chemical sorption and can be easily exchanged by another cation. It accounts for 1-5% of the total potassium. The exchangeable potassium content in topsoil varies depending on the soil-forming substrate, degree of weathering, pH, water regime, and fertilization level. High values of exchangeable K (500 – 700 mg kg^{-1} K) are found in clay soils with a higher content of clay fraction, while low values of 30 – 50 mg kg^{-1} K are recorded in sandy soils. The share of exchangeable potassium in the total exchange capacity of the soil is 2-7%. Exchangeable potassium is readily available to plants; it does not leach from the soil rapidly, and it does not significantly increase salt concentrations in the soil solution.

Water-soluble potassium is the potassium found in the soil solution and is the best uptaken by plants. In our soils (Slovakia), it accounts for 0.2% of the total and 1-20% of the exchangeable potassium. It is mainly made up of easily soluble potassium salts such as KCl, KNO_3, K_2SO_4, and K_2CO_3. Water-soluble potassium is the most vulnerable form of potassium in terms of its loss from soils. Thus, annual potassium losses can be about 60 kg ha^{-1} K (Fecenko and Ložek, 2000).

Constant balance is established between the different forms of potassium in the soil, which is dynamic by nature (K content is continuously effected by plants, moisture, fertilization etc.) and a new balance is always established according to the state of the whole system.

$$K^+ \text{ in soil solution} \leftrightarrow K^+ \text{ exchangeable} \leftrightarrow K^+ \text{ non-exchangeable (fixed)}$$

7.1.4 Calcium

We have soils (Slovakia) with calcium contents ranging from 0.15 to 10%. Its total content in soils is most significantly affected by the parent material/rock. Soils formed on limestones, dolomites (Rendzic Leptosol), or eolithic sediments with a high content of $CaCO_3$ (Chernozem), contain higher levels of Ca. Calcium content in soils is also significantly affected by the water regime (in intensively washed soils such as Luvisol, or Podzol soils, its content may be deficient), but also by fertilization – liming (essentially due to acidification of the soil environment).

Calcium in the soil is represented by organic and inorganic fractions. The level of the organic fraction is very low, often less than 1%. Thus, calcium in the soil is found mainly in inorganic forms. They are formed by unavailable and available calcium.

Unavailable calcium – a substantial part of calcium in the soil is in the form of water-insoluble compounds (85-99%); it is a part of primary and secondary minerals, e.g. calcite (42% Ca), dolomite (21% Ca) or anorthite (10-13%).

Available calcium is represented by **exchangeable and water-soluble calcium.**

- **Exchangeable calcium** – varies in soils depending on soil type, carbonate content etc. In the sorption complex of fertile soils, its content should be 65% or more (Zaujec et al., 2009).

- **Water-soluble calcium** accounts for 1-5% of the exchangeable calcium content and generally ranges from 10 to 200 mg l^{-1}. Calcium is relatively easily released from calcium carbonate, depending on the CO_2 content of the soil.

$$CaCO_3 + H_2O + CO_2 \rightarrow Ca(HCO_3)_2$$

Inorganic and organic acids in the soil are neutralized by calcium, which eliminates the solubility and subsequent mobility of harmful substances in the soil and their transfer to plants. Calcium is an important factor influencing the buffering capacity in soils. Materials with a higher calcium content are used to reduce soil acidity and adjust soil pH. This was elaborated in more detail in Chapter 6 – "Soil chemistry". Calcium is an important element that significantly improves the formation and stabilization of soil aggregates. This has a positive effect on improving the physical properties of soils. Despite the fact that calcium is fixed to the soil colloidal complex or may precipitate secondarily, it leaches out of the soil. These losses of leached calcium can vary from 50 to 600 kg ha^{-1} $year^{-1}$ in the form of CaO (Fecenko and Ložek, 2000).

7.1.5 Magnesium

In Slovakia, soils contain an average of 0.4 – 0.6% magnesium, but if the soils developed on dolomites such as Rendzic Leptosol, the Mg content can be as high as 10%. Magnesium in the soil is found in *inorganic* (95-99%) and in *organic forms* (1-5%).

Organic fixed magnesium – is magnesium fixed by microorganisms from post-harvest residues and organic fertilizers. Mg does not enter available forms until it dies and decomposes (microorganisms). The concentration of organically fixed magnesium is low.

Inorganic magnesium – in terms of availability for plants, it is divided into *available* and *unavailable* forms.

Available magnesium – is represented by *exchangeable* and *water-soluble* Mg.

Exchangeable magnesium is the most important form of Mg in terms of plant nutrition (5–10% of the total Mg concentration in the soil). Because of physicochemical sorption, it is fixed to the soil sorption colloidal complex. The concentration of magnesium fixed to the soil colloidal complex as well as its rate to exchangeable potassium and other monovalent cations such as Na^+ and NH_4^+ are of the same importance. When the concentration of these cations in the soil colloidal complex is high, then, magnesium is poorly adsorbed onto the soil colloidal complex. In well-stocked soils of Mg, the Mg:K ratio should be 3:1, which creates conditions for a balanced uptake of both cations. In the sum of cations in the soil sorption complex, magnesium should be 10-15% of the sum of cations.

Water-soluble magnesium is magnesium released from inorganic parts of the soil or released during the decomposition of plant and animal residues in the form of sulfates, chlorides and carbonates, which are found in the soil solution. Its amount in the soil depends on the exchangeable magnesium content, amount of water in the soil, and concentrations of other cations and clay minerals. These factors, in the end, also affect Mg leaching and Mg losses from soils. Kováčik (2014) stated that losses in Mg concentration per year were about 13–26 kg ha^{-1} and sometimes even 50 – 80 kg ha^{-1}. Magnesium supply in our soils is on average very good because up to 96% of soils have a satisfactory to very high Mg content.

Unavailable magnesium (non-exchangeable) is a part of primary and secondary minerals, which are a potential source of Mg. Of the primary minerals, serpentine silicates (26% Mg), talc (19% Mg), and mica biotite (18% Mg) are the most significant. Of the secondary minerals, Mg is contained in montmorillonite.

Compared to calcium, magnesium has a positive effect on soil hydrological conditions by reducing water infiltration, slowing its capillary rise and increasing the swelling of soil colloidal particles.

7.1.6 Sulfur

Soils contain on average 0.01-0.5% of sulfur. Significant sulfur inputs to the soil from SO_2 emissions have been significantly reduced since 1990. This is a result of a number of environmental measures, in particular the desulphurization of thermal power stations. For example, in the Czech Republic, these inputs decreased from an average of 120 kg S ha^{-1} $year^{-1}$ to 15 kg S ha^{-1} $year^{-1}$ (Lošák, 2008). Soils rich in organic matter contain significantly higher amounts of sulfur compared to mineral soils.

Organically fixed sulfur accounts for 40 – 90% of the total S content in soils. Organic sulfur is found in plant and animal residues in the form of proteins, polypeptides, and amino acids.

Inorganically fixed sulfur accounts for 10 – 60% of the total S soil content. Inorganic sulfur is found in the form of sulfates, sulfides, and sulfane.

Plants intake sulfur in the sulfate form. Sulfate sulfur contributes 10-15% of the total sulfur content. The transformation of sulfur in the soil is divided into four basic phases: oxidation, reduction, incorporation of sulfur from organic compounds, and mineralization of organic sulfur compounds (Fig. 7.7). Plant and animal residues in the soil are decomposed by aerobic and anaerobic bacteria, actinomycetes and fungi into amino acids. During the decomposition of sulfur amino acids, hydrogen sulfide is released. Under anaerobic conditions in wet soils, H_2S can accumulate and then have a toxic effect on plants and soil microorganisms. Under aerobic conditions, H_2S is gradually oxidized to sulfates. Sulfur conversions in the soil are carried out by sulfurization, desulfurization and sulfur immobilization.

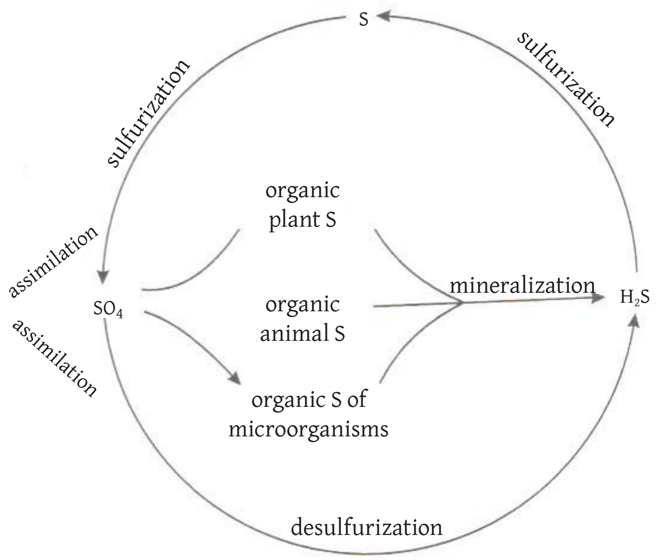


Figure 7.7: Sulfur cycle in soil (Fecenko and Ložek, 2000)

Sulfurization is a microbial process in which reduced forms of sulfur are gradually oxidized to sulfates by the action of sulfur bacteria.

$$SH^{-} \rightarrow S \rightarrow S_2O_3^{2-} \rightarrow S_4O_6^{2-} \rightarrow SO_4^{2-}$$

Desulphurization is a process of gradual reduction of sulfates, sulfites, and sulfates to H_2S by the action of desulphurization bacteria.

Immobilization – microorganisms use protoplasms, organic (sulfur amino acids) and inorganic (elemental sulfur, sulfides) sulfur compounds for synthesis. In microorganisms, sulfur is temporarily immobilized and, after their death, returns to the cycle.

Sulfur sorption and leaching – sulfur can be sorbed in the soil by physicochemical, chemical, and biological sorption. Physico-chemical sorption is more persistent on neutral soils and chemical sorption in acidic soils. Sulfur can leach from the soil in the form of sulfates. As a result of fertilization, the intensity of sulfur leaching from the soil also increases. 40 – 630 kg ha^{-1} of sulfur can be leached from the soil per year.

7.2 Micronutrients

Microelements present in soil: mainly iron, manganese, molybdenum, copper, and zinc; also aluminum, sodium, and chlorine. These elements are essential for both, life in the soil and the fertility of the cultivated plants. In the soil, they are present in low concentrations and when they are deficient, their application to the soil in the form of fertilizers is recommended. They can be harmful at high concentrations in the soil – soil pollution. The main roles of some micronutrients in the plant are summarized in Table 7.7.

Table 7.7: Basic micronutrients and their role in plants (Foth, 1990)

Element	Role in plant
Boron	Important in sugar translocation and carbohydrate metabolism
Iron	Important for chlorophyll synthesis and electron transfer enzymes
Manganese	Important in oxidation-reduction (redox)processes, O_2 production during photosynthesis
Copper	Respiratory catalyst, an important component of enzymes
Zinc	Regulates metabolic activities in enzymatic systems
Molybdenum	Required in nitrogen fixation

Boron – the average boron content in the soil is 30 mg kg^{-1}. Its content is mainly a result of the parent material/rock on which the soil was formed. Boron is found in tourmaline, a mineral with a boron content of 3-4%. In terms of availability, **water-soluble boron** plays an important role, accounting for 0.5-5% of the **total boron** in the soil. Boron deficiency occurs in soils with low boron content and in soils with low sorption capacity, in strongly acidic soils, after application of peat, after liming of the soil, and in areas with high rainfall.

Iron – the iron content in the soil may be up to 10%. It occurs in the form of oxides, silicates, and sulfates. Despite its high content in the soil, only a very small amount is available for plants. The presence of available forms of Fe (especially Fe^{2+}) decreases with increasing pH, while that of insoluble forms (especially Fe^{3+}) increases. Fe dissolves well in acidic and wet soils. Its availability is significantly lower in intensively cultivated and slightly alkaline soils. The uptake of Fe^{2+} is antagonistically affected by excess Cu^{2+}, Mn^{2+}, and Zn^{2+}, but also by higher phosphorus and calcium content.

Manganese – the manganese content in the soil is relatively high, 100 – 3,000 mg kg^{-1}. Its average content is approximately 450 mg kg^{-1}; up to 80% of the total content is in the exchangeable form. It occurs in the form of minerals (eg rhodonite), from which it is relatively easily released (during their weathering). In soil, manganese occurs as Mn^{2+}, Mn^{3+}, and Mn^{4+}. From the point of view of plant nutrition Mn^{2+} is of the greatest importance. Its increased concentration is recorded mainly in wet soils and due to intensive soil tillage, it oxidizes to unavailable forms. As the pH of the soil increases, the availability of Mn decreases, while increasing soil acidity results in the leaching of Mn from the soil.

Copper – the average Cu content in soil is 40 mg kg^{-1} (1 – 250 mg kg^{-1}). Both, the soil organic matter and soil texture have an effect on the Cu content of the soil. Only about 2-25% of the total Cu content in the soil is available for plants. Plants intake Cu as Cu^{2+} and, in the form of mineral and organic complexes. Cu forms poorly soluble compounds when the soil pH is low and there is enough soil organic matter (especially low molecular weight of organic substances). In general, Cu availability decreases with increasing soil pH. Cu^{2+} concentration is lower in slightly acidic to alkalescent soils.

Zinc – Zn content in soils ranges from 10 – 30 mg kg^{-1}. In the soil, Zn occurs in the form of sulfides (especially sphalerite). Zn availability is affected by both, soil pH and soil chelate concentration. a higher concentration of chelates increases the availability of ions and thus the uptake of zinc by plants. Zn availability decreases due to the increasing soil pH.

Molybdenum – the molybdenum content in soils ranges from 0.1 – 5 mg kg^{-1} and is therefore very low. Of this amount, approximately 0.1 – 10% is in an available form. The mobility of molybdenum is analogous to that of sulfate – it is very good. The mobility of Mo is influenced by soil pH – it increases with the increasing soil pH up to pH = 8.

8

Soil protection

8.1 Soil contamination

Soil acts as a natural sorbent, in which, in terms of soil contamination, harmful substances accumulate; subsequently they accumulate in plant and animal organisms. The activity and mobility of pollutants accumulating in the soil environment depend on many factors that determine their participation in the food chain. Factors influencing the mobility of pollutants:

- soil structure,
- sorption capacity of the soil,
- pH of the environment,
- the chemical properties of the pollutants,
- the nature of the subsurface water,
- the presence of chemically related ions.

Soil pollution is not immediately apparent, like air or water pollution. It is hidden and therefore the initial stages of soil pollution are difficult to control. Usually, it is noticed indirectly, i.e. by observing reduced or deteriorated quality of production. Some types of pollution can be eliminated by chemical and biological processes; other types of pollution can only be partially eliminated.

8.1.1 Contaminants in soil

Contaminants are heterogeneous mixtures of inorganic and organic substances. They are evaluated in terms of the source of their origin, their properties, and their effects.

According to the source of their origin, they are divided into:

1. **Anthropogenic** – they have entered the soil environment, as a result of anthropogenic human activity,
2. **Pedogenous** – they occur naturally in the soil environment from the parent rock.

According to their properties and effects, they are divided into:

- abiogenic elements (and their compounds) – Cd, Pb, Hg,
- microbiogenic elements and their compounds (active excess) – B, Co, Mn, Cu, Mo and Zn,
- macrobiogenic elements and their compounds (active excess) – N, P, K, Mg, S, Na, Ca,
- indicators of radioactive pollution – Cs, Sr, Zr, I, U,
- inorganic and organic pesticides,
- organic substances – phenols, polycyclic hydrocarbons, polychlorinated biphenyls, etc.,
- pathogenic organisms – viruses, bacteria, fungi, etc.

Anthropogenic sources of contaminants

Industrial activity causes widespread pollution by substances such as fly ash, sulfur dioxide, nitrogen oxides, chlorine, fluorine, carcinogenic hydrocarbons, heavy metal compounds, and the like. Emission sources are diverse, but generally include combustion processes in energy industries, production processes in industry and transport, waste incineration, services, etc.

Agricultural production – contaminants entering the soil in this way include:

- **chemical residues for plant protection** – (insecticides, herbicides, fungicides, nematicides, molluscicides, etc.). Although these substances save 25-35% of the crop, they are unknown to nature and are widespread even where they have not been applied (DDT, HCH, PCB, etc.). Some of them show considerable persistence and high accumulation capacity,
- **sewage sludge and other wastes** – are applied to the soil to enrich it with organic matter and nutrients. These sludges usually contain undesirable contaminants of heavy metals (Zn, Cr, Pb, Hg, Cd, Cu, Ni, etc.) and aromatic hydrocarbons, which limit their use in agricultural production,
- **application of excessive amounts of mineral or organic fertilizers** – excessive nitrates in the soil: acidify the soil, contaminate drinking water sources, and degrade production quality,
- **polychlorinated biphenyls (PCBs)** – have a very negative effect on the environment. They enter the soil and plants from oils, paints (in silage pits), some pesticides (as carriers), and other sources.

Characteristic properties of risk elements in soils

- **the toxic effect** of risk elements is not immediately obvious. A gradual increase in the content of risk elements leads to adverse changes in the soil, resulting in an overall decrease in soil fertility,
- **the chemical nature** of the element, its relationship to soil properties and its ability to dissolve and thus enter the plant is also of great importance,
- the **persistence** of risk elements in soils is longer than in the atmosphere and hydrosphere.
- **soil contamination itself**, especially in the surface layer, increases proportionally with the intensification and chemicalisation of industrial and agricultural production. However, the resulting damage increases exponentially,
- **soil detoxification** by technical means is very limited in comparison with the detoxication of the hydrosphere and atmosphere,
- **soil** is the strongest natural **buffer** of the elements. For this reason, they are most affected by soil properties in the natural cycle,

- **the self-cleaning capacity** of soil is lower than the self-cleaning capicity of air and water,
- given **the cumulative effects of heavy metals**, their assessment must be comprehensive and not isolated,
- in the case that heavy metals have carcinogenic effects, no dose can be considered safe,
- elements such as **As, Hg, Se, Sb** form **volatile** compounds which are subject to long-range airborne transmission,
- **soil contamination by heavy metals is local**, except for the use of industrial fertilizers and atmospheric gradients,
- **heavy metal toxicity is not a permanent condition**. The entry of heavy metals into the soil results in physico-chemical changes and the element is involved in the biogeochemical cycle of substances. Different substances are formed or disappear, which increase or decrease the toxicity of the element in the environment,
- we all know the short-term toxic effects of heavy metals. However, we know little about their long-term, slow accumulation, which is typical for soils,
- each element has a different physiological significance and thus a different degree of biotoxicity.

8.1.2 Reclamation of contaminated lands

Reclamation of contaminated soils is a long-term process and depends on a number of factors that determine it. In general, reclamation measures combine technical, chemical and biological interventions. When cultivating contaminated soils, minimal tillage should be avoided; deeper soil loosening is necessary to dilute contaminants.

Liming is necessary on acid soils; application of natural sorbents is needed on light-textured soils. Salts, bentonites, and zeolites, which bind toxic substances, are best suited. Important reclamation measures include regular fertilization with organic fertilizers. The content of heavy metals in the soil can be reduced biologically, too.

Measures to prevent contamination:

- stop their release in the form of air emissions, and thus in the form of immissions into the soil,
- stop the intrusion of foreign substances into the soil: in industrial fertilizers, in the form of used agrochemicals, or in the form of contaminated reclamation materials (sludges, composts, sediments from water reservoirs),
- for irrigation of plots, care should be taken to use hygienically safe water for irrigation.

8.2 Alleopathy and soil hygiene

Alleopathy of the soil is its contamination by the excretion of microorganisms and plant roots and various organic compounds during the decomposition of fallout and plant residues. **Allelopathy refers to a negative or positive effect of one plant on another plant. It also relates to the impact of microorganisms on plants, the impact of microorganisms on microorganisms, and the impact of plants on microorganisms.** The listed effect may be **beneficial – stimulating or unfavorable – inhibitory.**

8.3 Soil exhaustion and mitigation measures

Soil exhaustion is characterized by gradually decreasing yields due to the multi-year cultivation of monoculture plants. **"Exhausted soils"** do not respond to fertilization, loosening, irrigation, or sterilization and must rest for some time to restore their fertility**. Soil exhaustion** occurs when **there is a sharp decline in yields, as well as in the death of germinated and planted seedlings.**

Soil exhaustion is a complex degradation of the soil through poor rotations. The causes are:

1. Physical soil compaction (by irrigation, layering of dead plants, etc.),
2. Chemical (pesticide intoxication, nutrient depletion, change in soil pH),
3. Biological (alleopathy, infection, unbalanced soil diversity) nature.

The primary factors of soil exhaustion are biological ones. There are prophylactic and curative measures against soil fatigue.

Prevention – the purity and healthy seeds and seedlings. The purity of the environment is important in preventing pathogen-transmitted soil infections. Mixed crops and annual crop rotation prevent soil exhaustion. For better protection from plant pests, protective plants are positioned nearby.

Solutions to soil exhaustion – physical, chemical, and biological characteristics of fixing soil exhaustion.

1. **Physical** – loosening the soil, removing roots from the soil, washing the soil with water, substrate replacement, soil sterilization by steam, mechanical collection of soil pests, etc.
2. **Chemical** – fertilization, chemical disinfection and sterilization of the soil, soil liming, and the use of various melioration materials (peat, treated lignite Ekofert, coal dust) to modify the chemical properties of the soil.
3. **Biological** – in addition to the already mentioned cultivation approaches (mixed crops, crop rotation) – soil setting, application of high doses of compost and biological control of soil pathogens and diseases (ladybugs against lice, bacteriophages against pathogenic bacteria).

9

Soil fund

According to FAO (2012), the area of agricultural land in the world – on Earth – is about 4.9 billion hectares, which is 37.6% of the total land in the world (13.0 billion hectares). The area of agricultural land in developing countries (3.3 billion ha) exceeds the area of agricultural land of economically developed countries (1.6 billion ha) by almost 100%. The share of arable land in agricultural land, i.e. the degree of plowing, is about 28.2%. China, Australia, the USA, Brazil, and Russia have the highest agricultural land area and the highest arable land area. The highest share of arable land out of agricultural land is in Japan (93.7%), India (87.8%,) and Norway (81.8%). In the European Union, it is 57.6%.

There are 3.7 hectares of agricultural land per agricultural worker globally and 17.6 hectares in the European Union. Australia (894.3 ha), Canada (202.7 ha), and the USA (160.8 ha) have the highest agricultural land per active agricultural worker. India (0.7 ha) and China (1.0 ha) have the least. According to Eurostat statistics (2010), the European Union has 166 million ha of agricultural land and 177 million ha of forest land.

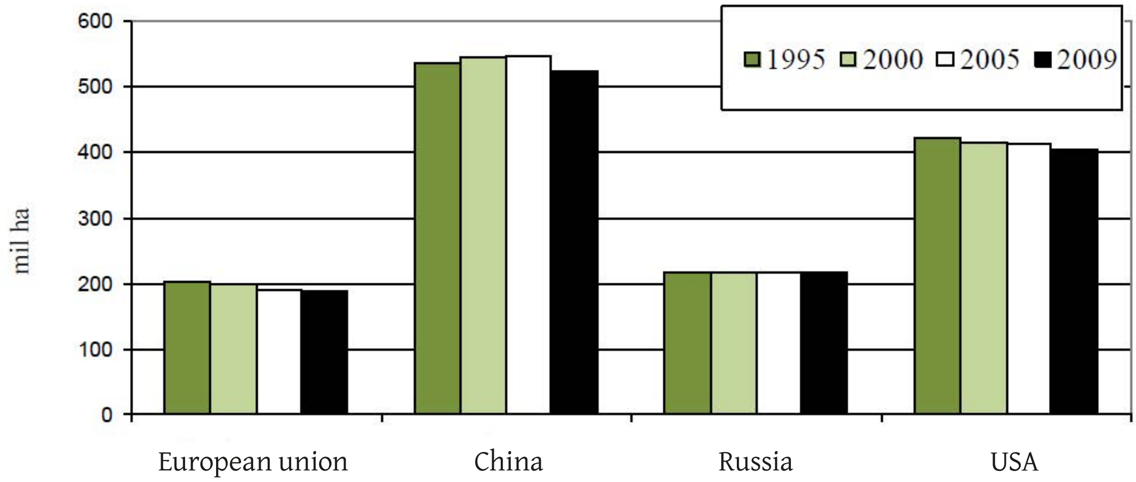


Figure 9.1: Evolution of agricultural land area (FAO, 2011)

Table 9.1: Agricultural land on Earth (FAO, 2012)

Country	Total land in million ha	Agricultural land in million ha	Arable land in million ha	Agricultural land/ total land in %	Arable land/ agricultural land in %	Number of ha per unit of field population	Number of ha per active field worker
World	13 003.5	4 889.3	1 378.4	37.6	28.2	1.9	3.7
EU	418.2	188.6	108.7	45.1	57.6	8.7	17.6
Argentina	273.7	140.4	30.9	48.5	22.0	45.3	9.9
Australia	768.2	408.7	46.9	54.3	11.5	454.1	894.3
Brazil	845.9	264.8	60.9	31.3	23.0	12.5	24.0
Canada	909.4	67.3	45.5	7.4	67.6	112.2	202.7
China	932.7	524.2	110.1	56.0	21.0	0.6	1.0
India	297.3	179.9	157.9	60.4	87.8	0.3	0.7
Japan	36.5	4.6	4.3	12.7	93.7	1.7	3.2
Mexico	194.4	102.8	23.7	52.7	23.1	5.1	13.0
New Zealand	26.3	11.5	0.5	43.2	4.1	38.3	61.7
Norway	30.5	1.0	0.8	3.4	81.8	5.0	11.5
Russia	1 637.7	216.2	121.2	13.2	56.1	18.8	34.6
South Africa	121.4	99.2	14.3	81.8	14.4	20.2	83.5
Switzerland	4.0	1.5	0.4	30.6	26.8	3.8	11.1
Turkey	77.0	38.9	21.3	50.8	54.7	2.7	4.8
USA	914.7	403.4	162.8	45.0	40.4	79.1	160.8

The evolution of agricultural land area in the world has changed since 1995. Since 2005, the total agricultural land area has been declining. This trend was reflected in almost all critical regions of the world, including Europe, the United States, and China. The evolution of agricultural land has been characterized by a decline in agricultural land in developed countries, where there is a higher agricultural production. In contrast, there is an increase in developing countries, with more extensive agricultural production.

The world population has grown from 2-10 million since the beginning of agricultural production 10,000-12,000 years ago, to 7 billion in 2010, and may stabilize at 10-12 billion by 2100. Globally, the fastest-growing populations are in developing countries, where natural resources are already under stress and where most hungry people live. In 2050, the Earth's population is estimated at 9 billion. According to Lal (2000), globally in 1950, there were 0.5 hectares of arable land per capita, in 2000 it was 0.2 hectares, and it is further estimated that in 2050 it will be only 0.14 hectares, and even in 2100 only 0.1 hectares. Approximately 6,350 m_2 of land is lost worldwide every second, so the trend of land loss is obvious. For example, annual land loss in Germany is 43,000 ha, 13 000 ha in the Netherlands, 12,000 ha in Austria, 3,000 ha in Switzerland, and in Slovakia, for example, it is about 1,000 ha.

Table 9.2: Evolution of the Earth's land mass and population (FAO, 2011)

Earth	1970	1975	1980	1985	1990	1995	2000	2010
Agricultural land (million ha)	4,603	46,70	4,720	4,815	4,910	4,972	4,974	4,974
Arable land (million ha)	1,303	1,313	1,332	1,372	1,382	1,375	1,364	1,364
Population (million inhabitants)	3,691	4,066	4,430	4,825	5,255	5,662	6,057	7,000
p.p./population (ha)	1.247	1.149	1.065	0.998	0.934	0.878	0.821	0.71
o.p./pop. (ha)	0.350	0.323	0.301	0.284	0.263	0.243	0.225	0.195

FAO projections indicate that food from land management (only 11% of the planet's land is suitable) will continue to be the most significant source of nutrition well into the third millennium. If we want to manage land both efficiently and ecologically, we need to know its characteristics, the causes of any low fertility, and how to remedy them. We need a comprehensive approach in order to make rational use of soil stock, as well as to increase soil fertility and productivity at a lower cost. The issue of soil protection is one of the most pressing environmental challenges facing mankind.

9.1 Soil in Slovakia

Wackernagel (1998) estimated that the area required for one inhabitant to live on our planet should have the following percentage structure, which for the Slovak Republic means (Zaujec et al., 2009):

- **55%** for energy crops	2 695 000 ha
- **5%** for housing and infrastructure	245 000 ha
- **15%** for food production	735 000 ha
- **11%** for grazing land	539 000 ha
- **14%** for forests	686 000 ha

Slovakia has an area of 49 036 km^2 = 4 903 600 ha and 5 400 000 inhabitants, so there is an area of 0.907 ha per inhabitant. As Vilček (2011) pointed out, a frequently asked question, especially in the context of food sufficiency for the human population, is how much agricultural land we actually need. From a food sufficiency perspective, it is relatively easy to answer. It is risky to determine the right threshold of need, especially when it comes to the non-productive functions of soils, without which life on Earth cannot exist. This question is particularly sensitive in a regional context. As part of the regional soil potential assessment system, as well as soil fertility, a social, political, economic, or social dimension – **the carrying capacity of the area** – is included in the detailed assessment approach.

Currently, our soils are capable of producing enough food to meet the needs of around 6.2 million people. In order to achieve food self-sufficiency in the whole of Slovakia, we need to have at least 1,367,853 ha, which represents about 56% of the currently registered agricultural land in Slovakia (Vilček, 2006).

Vilček (2011) states that the Slovak Republic has enough land to provide food self-sufficiency from its own resources, given the existing population. However, there are regions where this statement is not true. The principle of assessing and delimiting such areas is to determine the area of land per inhabitant. There are some differences in this parameter according to self-governing regions (Table 9.3).

Table 9.3: Land area in m^2 per capita in regions of the Slovak Republic (Vilček et al., 2007)

Region	Total area	Agricultural land	Arable land	Forest land	Water areas	Built-up areas
Bratislava region	3, 422	1, 585	1, 252	1, 255	93	257
Banská Bystrica region	14, 323	6, 341	2, 525	7, 006	120	501
Košice region	8, 795	4, 400	2, 661	3, 473	213	444
Nitra region	8, 922	6, 599	5, 720	1, 352	221	530
Prešov region	11, 313	4, 849	1, 887	5, 555	178	395
Trenčín region	7, 460	3, 086	1, 629	3, 655	105	383
Trnava region	7, 528	5, 323	4, 782	1, 186	265	497
Žilina region	9 ,825	3, 549	912	5, 482	185	361

In Slovakia, there is currently 0.44 ha of agricultural and 0.26 ha of arable land per capita. In the world, this ratio is 0.80 and 0.27 ha, respectively. Per capita, the largest amount of agricultural and arable land is in the Nitra region, the smallest amount of agricultural land is in the Bratislava region and the smallest amount of arable land is in the Žilina region. In the Bratislava region, there are only 257 m^2 of built-up areas per inhabitant, while in the Nitra region, it is as much as 530 m^2. According to calculations, a region with arable land per inhabitant less than 0.18 ha, or where the carrying capacity of the territory (expressed as the number of inhabitants per hectare) is greater than 5.5 in the case of arable land, is no longer able to feed its inhabitants and therefore becomes dependent on imports. There is already exhaustion of soil carrying capacity in the regions of Bratislava, Trenčín, and Žilina (Table 9.4). The strategy for protecting arable land from unwanted seizures should therefore be oriented towards these regions in particular. In the Nitra and Trnava regions, on the other hand, the production capacity of arable land is sufficient to ‘feed’ up to three times the current population.

Table 9.4: The population per ha, the remedies available, and the reserves relevant to food sovereignty in the regions (Vilček, 2011)

Region	Population per 1 ha			Alternative remedies		Reserve	
	Agricultural land	Arable land	Built-up area	Decrease in population (in thousands.)	Increase of arable land (in thousands. ha)	In number of population (in rhousands.)	In area of arable land (in thousands. ha)
Banskobystrický	1.6	4.0	20.0	-	-	257	46
Bratislavský	6.3	8.0	38.9	180	33	-	-
Košický	2.3	3.8	22.5	-	-	356	64
Nitriansky	1.5	1.7	18.9	-	-	1 525	277
Prešovský	2.1	5.3	25.3	-	-	30	5
Trenčiansky	3.2	6.1	26.1	63	11	-	-
Trnavský	1.9	2.1	20.1	-	-	898	163
Žilinský	2.8	11.0	27.7	345	63	-	-

Table 9.5: Relative expression of the production potential of agricultural soils in the Slovak Republic (Bielek, 2014)

Relative production value of land	Point value	% share of the ag. soil fund
High-yield	100 – 95	9.2
Highly productive	94 – 83	19.6
Productive	82 – 63	20.0
Medium production	62 – 51	7.9
Less productive	50 – 40	13.0
Low productive	39 – 30	13.5
Very low productive	29 – 20	9.6
Less suitable for agriculture	19 – 10	5.2
Not suitable for agriculture	below 10	2.0

While the Slovak Republic's land cover allows productive activities, it also plays a crucial role in nature. This fact emphasizes the importance of agricultural land in Slovakia not only because of their relatively small size and average quality, but also because there are no alternatives to their productive capacities. The current quality of Slovak soils used in agriculture is the result of both, their natural and anthropic characteristics. Using a numerical score, Džatko (2002) quantified the production potential of agricultural soils (evaluated soil-ecological units). For example, Mollic Fluvisols is rated with a maximum of 100 points, while very shallow soils on slopes above 25° have the lowest number of points (2). The average point value of the production potential of agricultural soils in Slovakia under the 100-point rating system is 54. In terms of the evaluation of individual regions, it was found that the Trnava region has the highest point value (76.9 points), followed by the regions of Nitra (75.5), Bratislava (72.4), Košice (54.2), Trenčín (45.7), Banská Bystrica (43.2), Prešov (36.3) and Žilina (29.9) (Vilček et al., 2007). In Table 9.5, soils are represented as a percentage of their production value.

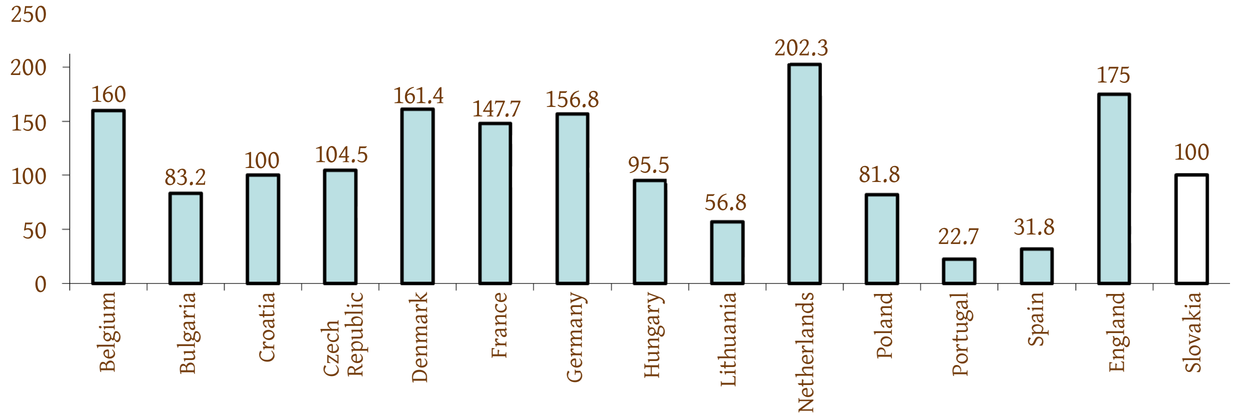


Figure 9.2: Comparison of soil production potential of the Slovak Republic with other European countries (Bielek, 2014)

A comparison of the soil production potential of the Slovak Republic with other European countries is shown in Figure 9.2. The Czech Republic, Croatia, and Hungary have roughly the same soil potential as Slovakia. The production potential is slightly lower in Bulgaria (app. by 17%) or Poland (app. by 18%). It is considerably lower in Portugal (by 77%), Spain (by 68%), and Lithuania (by 43%). Conversely, the Netherlands has a production potential that is 102% higher than Slovakia, England is 75% higher, Belgium, Denmark, and Germany are about 60% higher, and France is 48% higher (Vilček, 2011).

Real-life use of Slovakia's agricultural soils reflects the current state of the agrarian sector, as well as changing natural (especially climatic) and social conditions. Arable soils are used to about 80.3% of their production potential. Incorrect crop rotation, poor agrotechniques, high crop weediness, as well as inadequate fertilisation and liming of soils are to blame. In terms of utilizing the potential of permanent grassland, the situation is quite critical. The reduction in livestock numbers and the low investment in this type of land mean that only 57,6% of the potential of grassland (meadows and pastures) is being used.

9.2 Soil fund of the Slovak Republic

The Geodesy, Cartography and Cadastre Authority of the Slovak Republic registers the following special-purpose categories of the land fund – the so-called land types: arable land, hops, vineyards, gardens, orchards, meadows, pastures and their aggregate in the category of agricultural land, next, forest land, ponds, other water areas, built-up areas, and other areas. In Table 9.6, the sum of land areas is shown as the area of each special-purpose category or type of land, and the sum of their areas equals the total area of the state's territory – **the soil fund** (Table 9.6).

Table 9.6: Soil fund of the Slovak Republic in 2016 (Statistical Yearbook, 2016)

Type of land	Area (ha)
Agricultural land	2,397,041
Forest land	2,017,105
Water areas	95,250
Other lands	159,679
Non-agricultural and non-forest land	489,345
Total land area	4,903,491

The registration of the land fund according to individual types of land is used to monitor the development of the structure of the land fund of the state (e.g. monitoring the development of the area of arable land, agricultural land, the development of their area per capita). In addition, registering types of land provides the basis for the protection of land funds, the determination of land taxes, the distribution of subsidies for agricultural production as well as other ways of regulating agricultural production, landscape protection, and environmental protection.

In the data on the representation of the area of each land type, relevant information can be derived, including the share of arable land in the total area of agricultural land, the so-called **plowingercentage,** and **the area of arable land per capita**.

Table 9.7: Agricultural soil fund and its subdivision (Statistical Yearbook, 2016)

Type of land	Area (ha) in 2016
Arable land	1,411,294
Hops	511
Vineyards	26,359
Gardens	76,287
Orchards	16,565
Permanent grassland	858,601

Official data on the area of agricultural land and its registration by land type based on data from cadastral offices are provided by the Statistical Office of the Slovak Republic. Nevertheless, these data do not indicate how much land is actually used for agriculture. Therefore, the Ministry of Agriculture of the Slovak Republic has established a register of agricultural land based on user relations, the so-called **Land Parcel Identification System (LPIS).**

As of 1st January 2010, 2,086,068 ha (86.2%) of the total area of agricultural land registered in Slovakia (2,432,478 ha) was in the LPIS register. As much as 337,410 ha (13.8%) of agricultural land was not included in the system due to its neglected condition. These are mainly areas overgrown with scrub, trees, devastated, or otherwise unsuitable for agricultural use. Of the land registered in the LPIS, 1,971,489 ha are included in the system of direct payments for which the user receives a subsidy (of which 1,398,365 ha are arable land, 544,325 ha of permanent grassland and 29 799 ha of permanent crops). No direct payments are made for 114,580 ha due to unknown users.

With this registration, it is possible to determine the state of land use in the vast majority of Slovakia as well as the user of that land. Unfortunately, the real structure of land use is only known in areas (territories) that are reported to the system by applicants for state subsidies. This means that the land of owners or users who do not apply for subsidies is included in the system. However, neither the user nor the land use is known. These are mainly small agribusinesses, subsistence farmers, gardeners, etc.

9.2.1 Established trends in development of soils in the Slovak Republic

An important indicator for the national economy is the tracking of **the long-term development of the area** of each type of land in the agricultural soil fund, which shows the results of state land policy over a certain period of time. However, it can also be used to forecast future developments, either in the area of agricultural land or in the percentage of plowing.

It is evident from the data on the development of the agricultural soil fund and its qualitative structure that both the total area and arable land have decreased steadily. Indicators of the development of agricultural or arable land per capita have a similar trend – **a constant decline**.

Almost all of the phenomena that negatively impact soil quality originate from human economic activities. Agricultural soil degradation in Slovakia is mainly caused by water and wind erosion, subsurface soil compaction, acidification, and soil pollution:

- water erosion potentially threatens about 1,087, 839 ha of agricultural land, which represents about 46% of agricultural land,
- wind erosion is a less serious problem, threatening 202,429 ha of agricultural land, i.e. about 9%,
- subsurface soil compaction affects about 649,000 ha of agricultural land. 192,000 ha of agricultural land is actually compacted,
- acidification processes on agricultural land are intensive on an area of 425,000 ha,
- saline and highly alkaline soils cover an area of about 2,500 ha,
- over-polluted soils cover an area of approximately 30,000 ha,
- taking and removing land for economic purposes or housing development is the least significant cause of land degradation.

Changes can be expected in the development of our land area. In recent years, the average loss of land from the agricultural soil fund has been around 1,000 ha per year, which is about 3 ha per day. The Statistical Yearbook on the Soil Fund of the Slovak Republic (2016) shows that agricultural and arable land losses between 2015 and 2016 were 7,425 ha, or 934 ha. Agricultural land loss is expected to increase annually to a level of 2,000 ha or more. Of this, at least 50% will be our most productive soils on the plains. Table 9.8 shows the trends in the development of land area in the Slovak Republic. According to the data of the Statistical Office, there was a decrease of 388,000 ha of agricultural land and 300,000 ha of arable land in 2016 compared to 1950.

Table 9.8: Development of land area in the Slovak Republic in thousands of ha (Statistical Yearbook, 2016)

Year	Agricultural land	Arable land	Forest land
1950	2,785	1,711	1,723
1955	2,679	1,708	1,801
1960	2,754	1,761	1,785
1965	2,647	1,720	1,843
1970	2,628	1,683	1,850
1975	2,560	1,592	1,868
1980	2,477	1,516	1,912
1985	2,467	1,517	1,956
1990	2,448	1,509	1,989
1995	2,446	1,479	1,992
2000	2,442	1,460	2,000
2005	2,432	1,429	2,005
2016	2,397	1,411	2,017

On the other hand, an increase of 294,000 ha of forest land was recorded, which is consistent with the trend in developed countries, like the Slovak Republic. The general trend in land use around the world is towards a steady loss of forests and an increase in agricultural and built-up areas. The situation is different in developed countries, as Vilček (2011) pointed out. Over the last 60 years, there has been a gradual loss of agricultural land and an increase in forest and built-up areas.

9.2.2 Current state and development of soil degradation in the Slovak Republic (latest data from CMS – Soil)

Kobza et al. (2017) published results documenting the current status and evolution of land degradation within the Slovak Republic. The results come from CMS – Soil and focus on the assessment of soil erosion, soil compaction, assessment of organic matter content and quality, soil acidification, and soil contamination. The results indicate that the most significant negative changes in agricultural soil degradation processes are recorded in terms of soil erosion, soil acidification, and especially acidic soils.

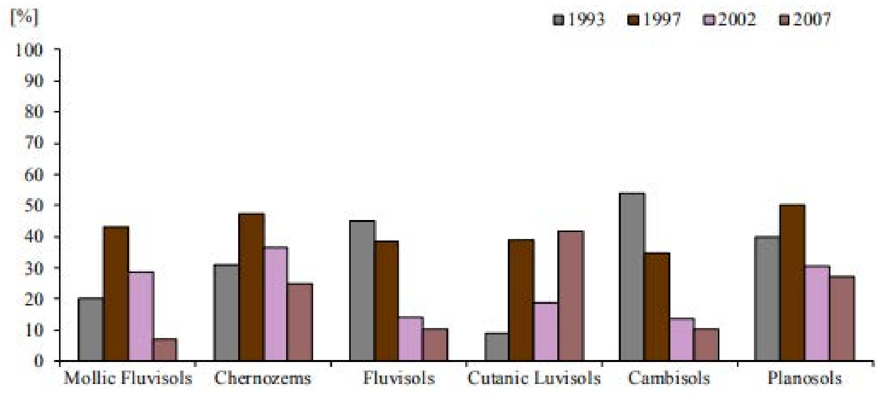


Figure 9.3: Evolution of soil compaction on selected soil types within the Slovak Republic (Kobza et al., 2017)

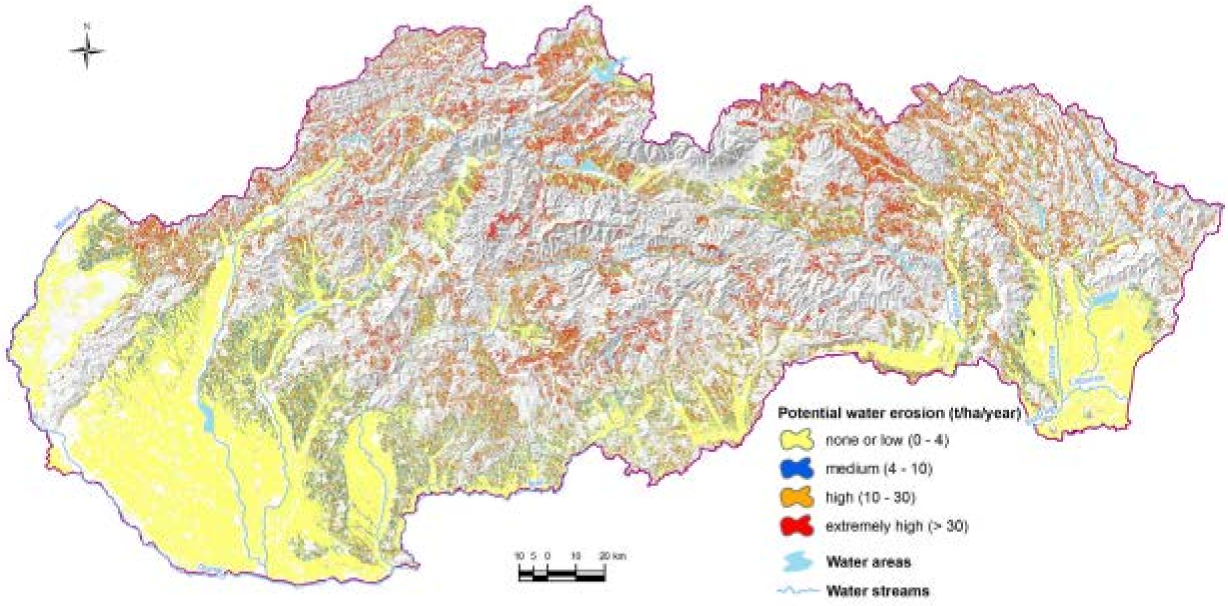


Figure 9.4: Potential vulnerability to water erosion in the Slovak Republic (Kobza et al., 2017)

Table 9.9: Current state and development of individual degradation processes within the Slovak Republic; ↑ increase (positive change), ↓ decrease (negative change), – without significant change (Kobza et al., 2017)

Soil types	Soil degradation processes				
	Erosion	Compaction	Acidification	Decline in organic matter	Contamination
Fluvisols	-	↑	↓	↑	↓
Phaoezems	-	↑	-	↑	-
Chernozems	-	↑	-	↑	-
Luvisols	-	↓+	-	↓	-
Albic Luvisols, Planosols	-	↑	↓	↓	-
Cambisols	↓	↑	↓	-	↓
Regosols	-	-	-	↓	-
Rendzic Leptosols	↓	-	-	↑	-
Podzols	-	-	↓	↑	-

The physical degradation was especially manifested in compacted and the eroded soils. It was calculated that about 39% of the agricultural land (mostly on Cambisols and Rendzic Leptosols) is potentially affected by soil erosion in Slovakia. The slight sensitivity to compaction was detected mostly on cultivated arable soils – texturally heavy Fluvisols, Chernozems, Luvisols, Albic Luvisols and Planosols and heavy Cambisols. Future attention must be paid to arable land with pH values in the slight acid range as well as soils with a low quantity and quality of organic matter. In order to reduce the acidity effect on plant growth and development, Ca amendments are commonly used to overcome the limitations of Al phytotoxicity.

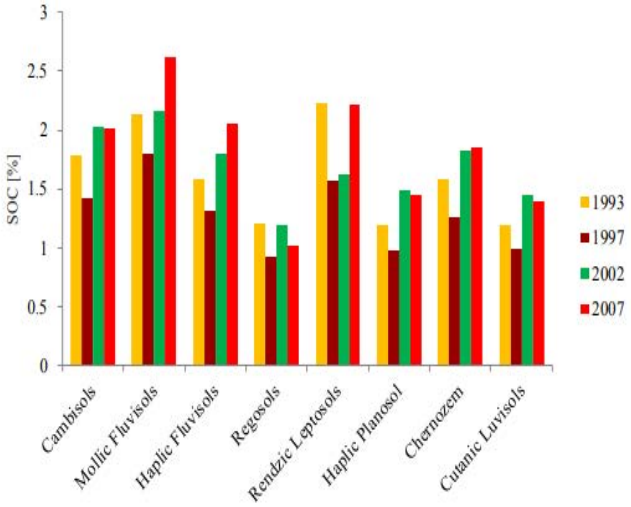


Figure 9.5: Development of organic matter content on selected soil types used as arable land in the Slovak Republic (Kobza et al., 2017)

During the last period, soil content has stabilized and increased slightly, especially in arable land, since soil monitoring began in Slovakia. A significant change in the concentration of risk elements was not determined during monitored period of 20 years. We found a slight increase in contaminants, especially in Fluvisols down rivers and in Cambisols influenced mostly by geochemical anomalies.

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