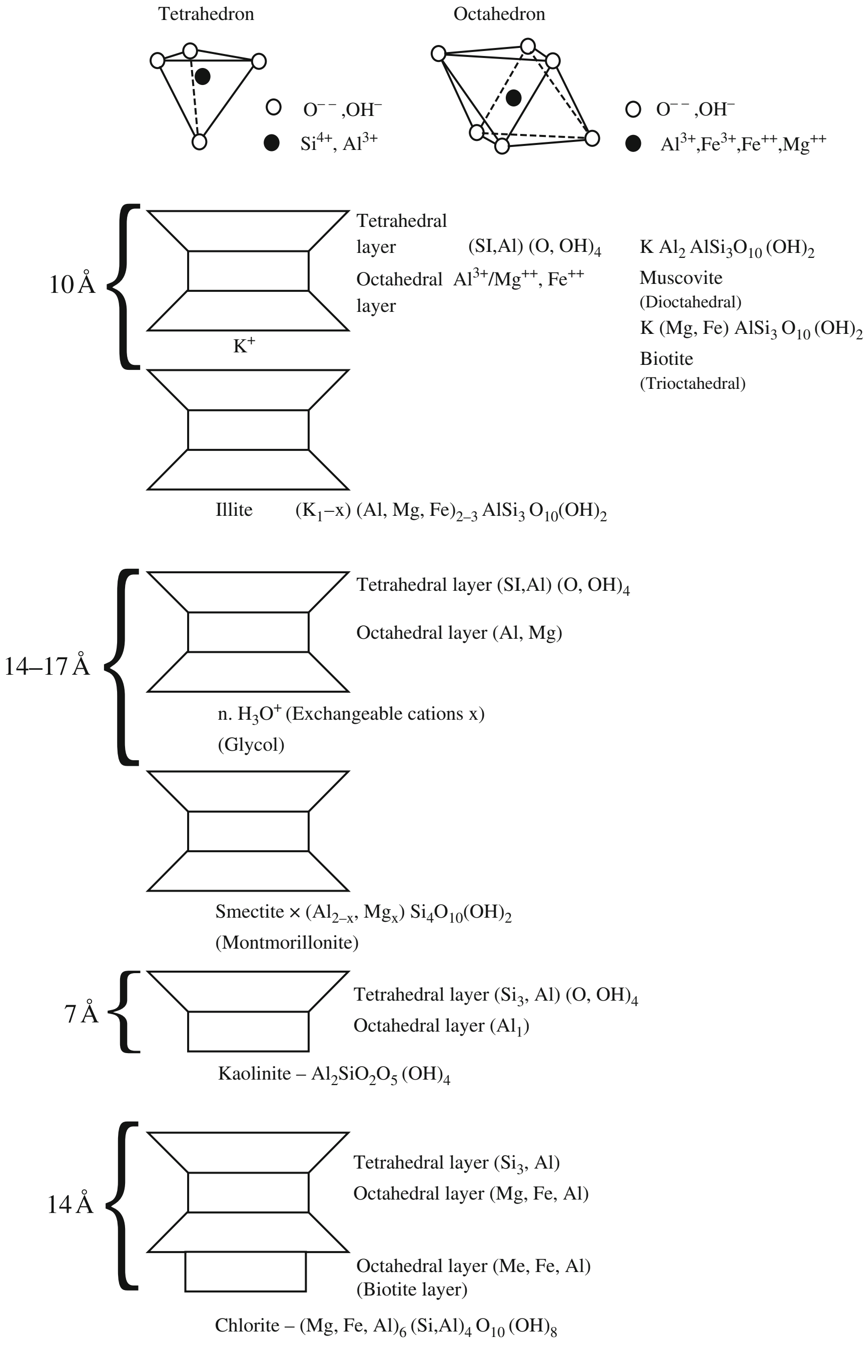

Fig. 3.8 Simplified illustration of the main groups of clay minerals. Their physical properties can be explained by their crystal structure. The chemical bonds between SiO_4^{4-} and O^{2-} in the tetrahedral structure are very strong. In the octahedral layers the bonds are weaker because Mg^{++} is surrounded (co-ordinated) with 6 oxygen. In the illite (mica) structure potassium is co-ordinated with 12 oxygen molecules, resulting in weak bonds that produce a strong cleavage

temperature metamorphic rocks, while iron-rich ones may form authigenically in sediments near the seafloor or at shallow depth.

Chamosite is an iron-rich chlorite mineral which often forms close to the sediment surface in reducing conditions. Together with siderite, chamosite is

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an important mineral in sedimentary ironstones which, especially in England, earlier were used as iron ore.

Clay minerals have a number of properties which distinguish them from most other minerals. Because of their very large specific surface area they have a great capacity for adsorbing ions, which is increased by the fact that clay minerals have negatively charged edges due to broken bonds. In water with a low electrolyte content, clay minerals will therefore repel each other. If cations are added, clay minerals will then accumulate a layer of positive ions (a double layer), and repulsion between negatively charged clay minerals declines as the strength of the electrolyte increases. Van der Waal's forces will therefore cause flocculation more easily in saltwater, where repulsion due to the negative charge is reduced. This is why clays transported by rivers flocculate into larger particles which sink more rapidly to the seafloor.

A colloid solution of clay in water is called a *sol*, which can be regarded as a Newtonian fluid. Flocculated clay is called a *gel* and has thixotropic properties. This means that the shear strength decreases with increasing deformation so that it changes from a gel to a sol which consists of dispersed colloidal particles in water (hydrosol). After a time without deformation, it regains its strength. This is typical for smectitic clays.

Sediments which increase their volume when they are deformed, are called dilatant. When the original packing of the grains is destroyed, the new packing may be less effective and the volume then increases. Walking on a beach the deformation of the sand causes increased porosity so that water is sucked into the sand at the surface.

Norwegian clays deposited in the sea when the ice sheet retreated about 10,000 years ago, consist of crushed rock fragments and have approximately the same composition as the parent rock. During deposition these clays acquired a markedly porous structure, with the minerals stacked like a card house. Saltwater helped to hold this structure together because it neutralised the negative charges. When the clays are elevated above the sea by isostatic recovery of the land, they are exposed to meteoric water. Even if the clay has a low permeability, freshwater will slowly seep through it and remove the saline water. This reduces the strength of the clay's structure and hence its stability and the clay becomes "quick". This is caused by overpressure and weak effective stress between the grains and hence little friction. The card house structure may then collapse, releasing the interstitial water to produce a low viscosity clay slurry that will flow even down very gentle gradients. The addition of salt (NaCl or KCl) binds the clay particles so that the shear strength is increased, and this is employed when the ground has to be stabilised for buildings and construction.

3.6 Weathering

The composition of clastic rocks in sedimentary sequences depends to a large extent on the supply of sediments from source areas undergoing weathering and erosion. The physical properties of sedimentary rocks are controlled by the primary sediment composition and changes during burial (diagenesis). We shall here briefly look at the processes producing sediments.

Mechanical weathering is the physical breakdown of rocks into smaller pieces which can then be transported as clastic sediment.

Chemical weathering involves the dissolution of minerals and rocks and precipitation of new minerals which are more stable at low temperatures and high water contents. Parts of the parent rock will then be carried away in aqueous solution by the groundwater and rivers into the ocean.

Erosion is the combined result of the disintegration of rock and the removal of the products.

As we shall see, biological processes are not only important in connection with chemical weathering but also with respect to mechanical weathering. Both mechanical and chemical weathering, which are essentially land surface processes, are due to the chemical instability of rocks that were formed or modified under other conditions (at greater depths, higher pressures or temperatures, or in different chemical environments). They are no longer stable when exposed to the atmosphere, water and biological activity.

3.6.1 *Mechanical Weathering*

Igneous and metamorphic rocks which have been formed at many km depth at higher temperatures and pressures are not stable when exposed at the surface.

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When uplifted and unloaded the rocks expand, mostly in the vertical direction, producing horizontal fractures (*sheeting*) parallel to the land surface. This is because as the rocks near the surface are unloaded by the reduction in overlying rock, they can expand vertically but not horizontally. In this way the vertical stresses become less than the horizontal ones, and joints develop normal to the lowest stresses. We see this most clearly in granites, which are homogeneous, while expansion in metamorphic and sedimentary rocks occurs along bedding surfaces, along tectonically weak zones with crushing, or along fractures that were formed at great depth. Joints opened by stress release in turn provide pathways for groundwater to circulate, increasing the surface area of rock exposed to chemical weathering.

In areas that experience freeze-thaw cycles, frost weathering becomes very important. When water freezes in cracks in rock, it expands by 9% and can generate very high stresses, further widening cracks near the surface. The surfaces of exposed rocks are also subjected to daily temperature fluctuations which cause greater expansion of the outer layers relative to the rest of the rock. Desert regions in particular experience very wide daily temperature ranges, though the importance of this process for mechanical weathering has been questioned. The roots of plants and moss can also contribute to mechanical weathering as they grow into fractures, take up water and expand.

3.6.2 Biological Weathering

Rocks are a source of nutrients for plants, and plants are capable of dissolving and breaking down the major rock-forming minerals. Moss, which consists of algae and fungi living in symbiosis, produces organic compounds that can slowly dissolve silicate minerals. Even in the earliest stages of weathering, we see that fungus hyphae penetrate into microscopic cracks.

Plant roots produce CO_2 which helps to lower the pH and dissolve minerals such as feldspar and mica, thus freeing an important plant nutrient, potassium. Plants also produce humic acids, which likewise strongly influence the solubility of silicate minerals, and also affect the stability of clay minerals. The production of humic acids is perhaps the major factor influencing the rate of weathering. In heavily vegetated areas, such as rain forest in the tropics, the weathering rate is exceptionally high because so much humic acid is produced.

Bacteria and fungi, which are found in almost all soil types, are active in breaking down minerals. Animals also contribute to weathering, and certain marine organisms such as mussels are able to bore into solid rock (see Chap. 8). Microbiology has become a key area of research in the quest to understand how minerals are dissolved and precipitated.

3.6.3 Chemical Weathering

There is no sharp demarcation between biological and chemical weathering, because we find biological activity in almost all soils and rocks near the surface. The chemical environment in water at the surface of the earth is very much affected by local biological activity, and in most cases it is biological processes that cause weathering to continue after rainwater has been neutralised through reaction with minerals. We will therefore use the term "weathering" here for both chemical and biological processes.

3.6.4 Weathering Profiles (Soil Profiles)

Both chemical and biological weathering are to a large extent controlled by climate. The crucial factor is the ratio between precipitation and evaporation in an area. In areas where precipitation far exceeds evaporation, *podsol profiles* develop in which there is a net transport of ions down through the soil profile as minerals are dissolved. In other words, we get weathering due to the fact that rainwater is slightly acidic (on account of its CO_2 and H_2SO_4 content) and contains oxygen. Rainwater is initially undersaturated with respect to all minerals. Some minerals are only very slightly soluble, others more soluble in this slightly acidic, oxidising water. Dissolved ions are transported down to the water table, but ferrous iron liberated from iron-bearing minerals will be oxidised and precipitated as ferric iron ($Fe(OH)_3$). Vegetation at the top of the soil profile produces CO_2 from roots and organic compounds, particularly humic acid, which will increase the solubility of silicate minerals. Similarly, aluminium derived from

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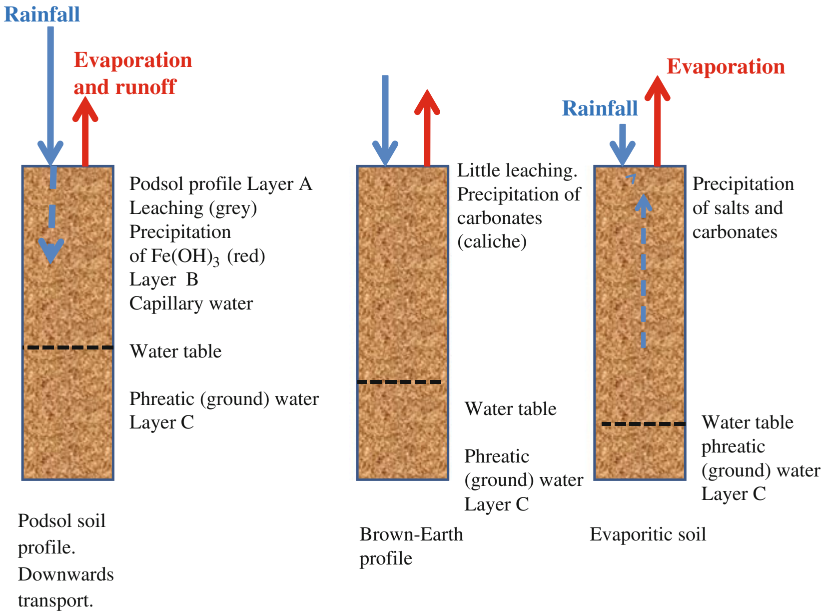
Fig. 3.9 Simplified representation of soil profiles as a function of rainfall (precipitation) and evaporation



a solution of feldspar and mica, for example, precipitates as $Al(OH)_3$ but is less noticeable because aluminium hydroxide is white. The uppermost part of the soil profile, where dissolution due to undersaturated rainwater and organic acids dominates, is called the A-horizon. Some of the dissolved salts and particularly iron hydroxide is precipitated in the B-horizon below (Fig. 3.9). These may develop into a layer of solid rock (hard-pan) cemented with iron and aluminium oxides and hydroxides.

Where precipitation is approximately equal to evaporation, there is less leaching within the soil profile. At a certain depth (about 0.5–1 m) carbonate will be precipitated and form an indurated layer (calcrete) which may be eroded and from conglomerates.

The organic content is greater in the B-horizon which is brown due to less oxidation of organic matter, hence the term *brown-earth profiles*. If evaporation is greater than precipitation there will be a net upward transport of porewater, causing dissolved salts from the groundwater to be precipitated high in the soil profile.

3.6.5 What Factors Control Weathering Rate and Products?

Because weathering is the most important sediment-producing process, we are interested in understanding how the rate of weathering is related to rock type, precipitation, temperature, vegetation, relief etc. We also try to establish correlations between weathering products, particularly clay minerals, and these factors. By studying sediments from older geological periods, we can learn something about weathering conditions at those times. Weathering products will also bear the stamp of the rocks undergoing weathering. The stability of a mineral during weathering is largely a function of the strength of the bonds holding the cations in the crystal lattice. Potassium (K^+) in mica is held by weak bonds (low ionic potential) which are responsible for the pronounced cleavage. In biotite, the Mg^{++} and Fe^{++} in the octahedral layer will also be weakly bonded. During weathering cations like K^+, Na^+, Ca^{++}, Mg^{++} and Fe^{++}can be attacked by protons (H^+) which will replace them and send them into solution. Chain silicates like hornblendes and pyroxenes will also be relatively unstable and rapidly weather. In feldspars the alkali ions are dissolved so that the whole mineral disintegrates. Stability is lowest in calcium-rich plagioclase, while pure albite (sodium feldspar), orthoclase and microcline (potassium feldspar) are more stable. The breakdown of these silicate minerals will primarily liberate alkali cations. Silicon and aluminium have very low solubility and form new silicate minerals, largely clay minerals, though some silicic acid (H_4SiO_4) goes into solution.

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