

The Development of Soils and Weathering Profile

Introduction

Weathering processes and soil development have been classically considered as two aspects of a single phenomenon which is the transformation of rocks under climatic conditions (meteorization). White (1995) gave a clear definition of what is classically considered as a soil profile: “*The soil profile is generally considered to make up the upper part of a much thicker weathering profile. A more inclusive definition of a soil profile is the vertical section that includes all the layers that have been pedologically altered during chemical weathering. Soil horizons defined within such profiles clearly recognize the vertical transition in the degree of weathering from the intensively altered A and E horizons through the moderately weathered B, to the relatively unaltered C horizons to bedrock (Buol et al. 1989). Such an approach encompasses the entire span of chemical weathering occurring at the earth’s surface including both the distribution of primary silicates as well as secondary clay and oxyhydroxide minerals*”. This definition opens a fundamental question: do soil and weathering belong to a same alteration phenomenon at the Earth’s surface, the soil being the most altered level? If not, are soil formation and rock weathering different Earth’s surface phenomena? The idea that they are different phenomena has been proposed by Ildefonse et al. 1978). It is supported by Taylor and Eggleton (2001) who proposed that: “*The distinction from regolith is, in general subtle, but the relationship with the surface is distinctive. To amplify the difference it may help to suggest that soils develop horizons from a regolith parent, while regolith develops from in situ or transported weathered rock material to form zones*”.

Until now, it has not been possible to find an unambiguous definition of soil and weathered rocks in spite (or because) of the huge literature dedicated to the study of alteration. This is essentially due to natural and Gulliver factors. The first factor is imposed by natural environments, either in tropical or temperate climatic zones, in which the soils appear superficially to be in a continuum between clay-rich materials and with weathered rocks. For this reason weathering has been considered to be concomitant with the soil development. The Gulliver factor results from the choice of the side by which the egg is eaten! Indeed, pedologists study the soil downward: the A horizons are at the very surface of solid Earth while geologists begin their investigation upward from the unaltered rocks. Even if this looks like a student joke, it is certainly one of the major reasons why the two scientific communities have had so little collaboration in the past.

We can try to reduce this secular gap if soils and rocks are considered from a physical and chemical point of view. Starting from the simplest observations, it becomes easy to distinguish two environments which have contrasting properties:

- the *soil* is limited at the top by the interface with atmosphere. This interface plays a specific role because of the presence of living organisms (plants, bacteria, animals). It is the organic matter reservoir at the Earth's surface. A soil is organized in horizons whose limits are more or less parallel to the atmosphere interface. Soils involve plant and plant debris interaction with silicate and oxide materials.
- *rocks* are weathered along fluid pathways (fractures, joints, faults, ...). Weathered rocks are not necessarily covered by a soil. They do not contain large quantities of organic matter or living organisms. The alteration process occurs between rain-water and rock.

The long history of continent alteration on Earth since the Archean era shows that weathering has affected rocks for a long time before life appeared. Weathering (interaction between slightly acid rainwater and silicate minerals) can transform rocks independently of soil formation. Conversely, after each volcanic eruption, plants and micro-organisms are able to colonize fresh ash or lapilli deposits. This is the first step of the development of a soil on a rock which has not been previously altered. Thus, soils and weathered rocks may be considered as independent open systems. Soils imply plant and animal activity while altered rocks imply interaction between atmospherically equilibrated water and silicate materials.

3.1

Physical Description of Soils and Weathering Profiles

Soil develops downward from the rock-atmosphere interface. With increasing time, the soil material is progressively differentiated into entities of similar properties which are more or less parallel to the interface topography, i.e. the horizons. On the contrary, weathering does not depend much on the geometry of the rock-atmosphere interface but rather on that of the fluid pathways which crosscut the rock. Friable zones are frequently observed beneath unaltered ones. Weathering and soil formation have been considered to be concomitant because soil develops faster on permeable rocks than on hard ones. In fact, there is an interface between the base of the soils and the rocks whatever their alteration degree. With increasing time, a soil deepens while weathering progresses downward into rocks along the fluid pathways. In both cases, any given point in the soil or the rock experiences the propagation of the interactions between fluids (with or without the chemical control of organic matter) and silicates. Consequently, any given point is submitted to changes in local chemical conditions.

3.1.1

The Development of Weathering Profiles

The definition of rock facies or zones can be considered from the bottom upwards. The analysis which follows treats the alteration process and development of alteration zones from the bottom upwards. Initially one must define the zones concerning the alteration of coherent, high temperature silicate materials which can form clay minerals: rocks.

Previous Definitions of Rock Facies or Zones

During weathering the fabric of the parent rock changes progressively from the unaltered to the most clay-rich zones or facies. The definition of the term regolith was given by Taylor and Eggleton (2001, p. 157–159). These authors consider that the regolith is composed of saprock + saprolite where saprock is the altered coherent rock (structure conservative) and the saprolite: is the altered soft rock (also structure conservative). Above the saprolite, the altered soft rock which is a non-conservative zone is assigned as the pedolith. These definitions are a bit confusing at least for the weathering profiles developed under temperate climates, for two reasons:

- the pedolith could be considered as being concomitantly formed with the soil. This is not the case,
- the limits between zones where the rock structure is conserved and that where it is lost is not sharp. Most often non-conservative zones are observed inside the saprock itself. Conversely, saprock or unaltered rock pieces are conserved in the saprolite.

Saprock and saprolite form the residual regolith (see regolith definition in Taylor and Eggleton 2001, p. 1–3). In many publications, the term alterite is used as an equivalent of residual regolith.

New Definitions for Altered Rock Materials

Considering these observations we propose other definitions for altered rock facies based on the physical properties that are easily observable in the field: conservation or destruction of the primary rock structure and physical resistance (coherent, friable, soft, clay enriched). Weathering profiles developed under temperate climate exhibit two altered rock types: the coherent rock which may be sampled as self-sustained pieces conserving original rock texture and the friable rock which when sampled becomes gravel. According to these observations, one can distinguish three facies using the following features:

- *the altered coherent rock*: the original structure of the rock is perfectly maintained, although some primary minerals show pockets of alteration in which their debris are mixed with newly-formed clays,
- *the saprock* (friable) in which the original rock structure is conserved: the structure of the rock is maintained although its mechanical strength is very reduced. The porosity ranges from 10 to 20%. It results from the dissolution of primary minerals and the formation of new fractures generated by local variations in the stress state.
- *the saprolite* (friable) in which the original structure is erased: it forms more or less continuous zones above and within the saprock. In tropical climates, saprolite is the most developed facies of the weathering mantle (Nahon 1991; Tardy 1993). The mechanical strength and Young's modulus are very low and correspond to those of unconsolidated rocks. However, in specific cases, the upper levels of some laterite profiles are strongly indurated by the precipitation of Fe-oxyhydroxides. Their mechanical resistance is higher than the uncemented saprolite (duricrusts or cuirasses).

Propagation of the Weathering Mineral Reactions

The primary minerals are progressively replaced all along the weathering process by a microporous material composed of fine parent crystal debris and newly formed minerals such as clays and oxy-hydroxides. The microporous material is assigned as *the primary plasma*. As expected, the quantity of the primary plasma in an altered gabbro (volume approximated by point counting on petrographic thin sections) increases linearly with decreasing amounts of primary minerals (Fig. 3.1a). However, the relation indicates that only 75% of plasma is formed when the primary minerals have totally disappeared from the rock. That means that 25% of the parent components have been dissolved. This is coherent with the measured porosity.

The continuous functions (linear or logarithmic) depicted in Fig. 3.1a support a progressive alteration process in spite of the contrasted aspect of the different zones

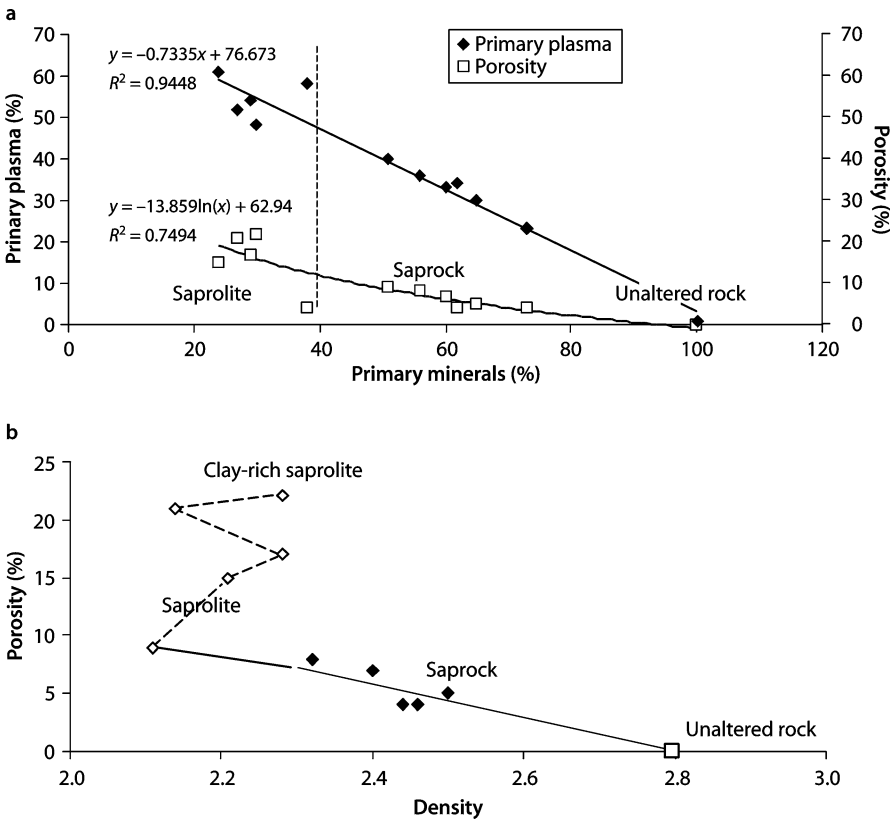


Fig. 3.1. Variation of rock components during gabbro weathering (amounts are expressed as % surface by point counting); **a** the primary plasma formation is a linear function of the disparition of the primary minerals. At 0% primary minerals, the primary plasma represents about 75% of the solids; **b** macroscopic structure changes in the saprolite correspond to an increase of density. Data from Ildefonse (1978)

in the alteration profile. The fabric of the parent gabbro is conserved at all observation scales (from macroscopic to microscopic) in the altered coherent rock and the saprock. It is conserved at the macroscopic scale but erased from place to place at the microscopic scale in the saprolite. At least, it is totally erased at all scales in the clay-rich saprolite. Again, the clay-rich saprolite facies is not considered here as the transition to soil through a B horizon. On the contrary, it is a step in the weathering process as shown by Fig. 3.1a. The density decreases progressively from unaltered rock to saprolite, but increases to clay-rich saprolite (Fig. 3.1b). This reversal is related to the macroscopic structure changes.

It could seem strange that density and porosity increase concomitantly in the clay-rich saprolite. In fact, the density of the newly formed clay material is higher than that of the primary plasma because of compaction and closure of voids. The new material is assigned as *the secondary plasma*. The rock structure collapses under its own weight as its mechanical resistance is considerably lowered by the formation of the primary plasma (Righi and Meunier 1995). This physical process is totally independent of pedological ones.

Because alteration increases the porosity, the density decreases concomitantly with the amount of remaining primary minerals in the rock structure conservative facies. Figure 3.2 shows that the relation is roughly linear and has more or less the same slope for a gabbro (Ildefonse 1978) and a granite (Meunier 1977). In spite of the fact that pyroxene and plagioclase are more sensitive to meteoric alteration than K-feldspar and micas, the gabbro and the granite experiment the same type of mineral reaction: high-temperature silicates \rightarrow clays + oxyhydroxides. For a given density, the gabbro is more altered than the granite. In other words, during alteration, the gabbro produces more clay minerals than granite. The difference is due to the presence of quartz in granite rocks. Quartz does not contribute to the formation of clay minerals.

This description is in fact brief and will be considered in greater detail in Chap. 4.

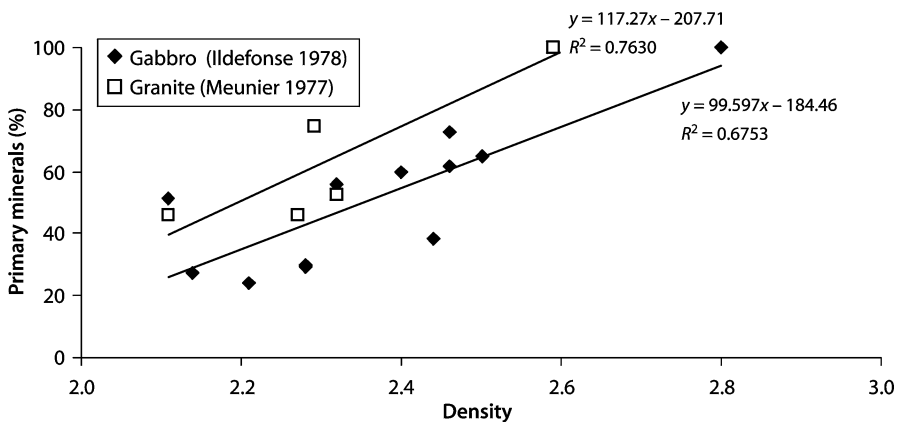


Fig. 3.2. Variation of the amount of primary minerals (% surface by point counting) with density for two magmatic rocks: gabbro (data from Ildefonse 1978) and granite (data from Meunier 1977)

3.1.2

The Development of Soils

One of the key points to study in the development of soils is to reconsider the definition of the C horizon. The C horizon is classically included in all soil profile descriptions as a transition with the substratum. Its lower limit is never indicated. This is a consequence of the current definition of the C horizon.

What Is the C Horizon?

The definition of C horizon is still debated (see the Geoderma Special Issue, May 2005). According to the Keys to Soil Taxonomy (1998), the present day definition is: “*Horizons or layers, excluding strongly cemented and harder bedrock, that are little affected by pedogenic processes and lack the properties of O, A, E, or B horizons. Most are mineral layers. The material of C layers may be either like or unlike the material from which the solum has presumably formed. The C horizon may have been modified, even if there is no evidence of pedogenesis. Included as C layers are sediment, saprolite, bedrock, and other geologic materials that are moderately cemented or less cemented. The excavation difficulty in these materials commonly is low or moderate. Some soils form in material that is already highly weathered, and if such material does not meet the requirements for A, E, or B horizons, it is designated by the letter C. Changes that are not considered pedogenic are those not related to the overlying horizons. Some layers that have accumulations of silica, carbonates, gypsum, or more soluble salts are included in C horizons, even if cemented. If a cemented layer formed through pedogenic processes, however, it is considered a B horizon*”.

The B horizon is the zone of illuviation, or translocation. It is formed by an accumulation of clays, Fe-oxyhydroxides, humic substances or carbonates derived from soil horizons above.

The above definition and the description of the C horizon are not perfectly clear and a bit contradictory. Some parts have to be carefully re-examined:

- what does “*little affected by pedogenic processes*” mean? Is the lack of requirements for O, A, E or B horizons an appropriate definition?
- why are the effects of deep roots of trees or plants which crosscut the soil and penetrate inside the substratum far away from A, E or B horizons ignored? Are they considered to be in contact with the bedrock without any altered layer?

The definition of the C horizon in the french taxonomy (Référentiel Pédologique, INRA 1995, p. 44–45) is a bit more explicit. The C horizon differs from the A, E or B because it does not have a generalized pedogenic structure. The original lithologic structure is the dominant feature. However, the C horizon may present some dispersed pedogenic features such as illuviation (carbonate, sulfate, salts or clay deposits) or lateral eluviation.

The definition of the C horizon is made difficult because of the static analysis of soils. If considered as a dynamic process, the formation of the C horizon can be better understood. Particularly, dynamics implies that the C horizon would have an interface with the substratum even if difficult to locate precisely in the profile. This inter-

face necessarily migrates downwards as the soil develops with time. Then, the C horizon may be seen as a transition between the A, E or B horizons and the substratum whatever its alteration stage. The pedogenic features progressively disappear toward the interface. This is the meaning of “*little affected by pedogenic process*”. What are the characteristics of the interface?

The Soil-Altered Rock Interface

The soil may exhibit a sharp interface when developed on hard rocks but this is not a rule. The transition between the C horizon and the saprock or the saprolite may be progressive suggesting that soil formation and weathering are a single phenomenon. This is supported by Taylor and Eggleton (2001) who claimed that “*The changes that transform the saprolite are largely the same as those that form soil. In some profiles this region where the parent fabric begins to be lost can be viewed as the base of the B horizon*”.

The depth at which the soil-rock interface is found in a given landscape varies from place to place. It may be located at only a few millimeters depth on hard rocks (boulders) or at several tens of centimeters in friable materials. The soil-rock interface depends on different parameters such as the permeability of the rock, the position in the topography where the alteration occurs and the time during which the alteration conditions remain stable enough for the soil to develop. Consider for instance a cambisol to luvisol soil sequence in which the eluvial and accumulation horizons are progressively differentiated (Fig. 3.3). The pedological horizon limits remain approximately parallel to the surface. Most often, soil descriptions do not give a lower limit of the C horizon. This habit intuitively comes from the idea that the soil is the most altered stage of a weathering profile. This is not correct and must be reconsidered because a soil develops from the atmosphere interface. Consequently, the C horizon would have a lower limit (sharp or progressive) which is assigned as the soil-rock interface.

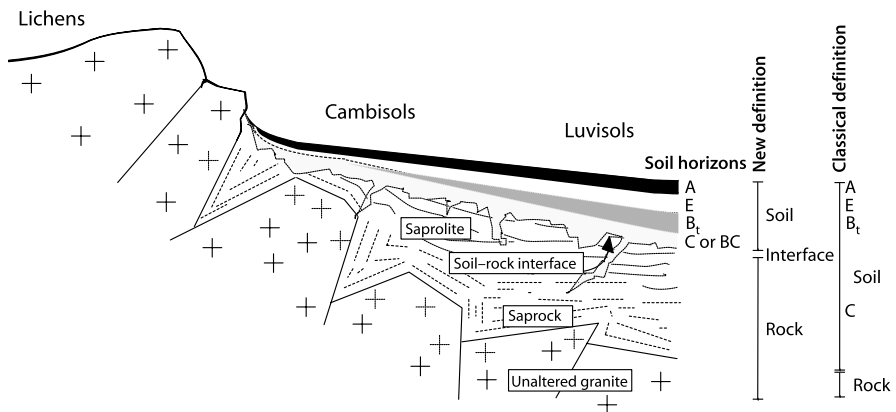


Fig. 3.3. Schematic representation of the position of the soil-rock interface in a granitic landscape. A: organic horizon, E: eluvial horizon present in acid soils, B_t: texturally observable accumulation horizon; C or B-C: transition horizon where the hydraulic conductivity is the lowest through the profile

Buol and Weed (1991) proposed the interface as a BC transition horizon in which the hydraulic conductivity is the lowest in the profile. Using thin section observations, they related the low conductivity to the absence of continuous pores and to the formation of thick cutans of clay translocated from the overlying soil. The soil-saprolite boundary may be identified through hydrology parameters such as the water saturation and the matric potential, i.e. tension forces (White et al. 2002).

Even if the location of the soil-rock interface is difficult to observe at a given place at the scale of the landscape, the C or BC horizon is seen to crosscut either saprolite, saprock or hard rock. Consequently, the soil-rock interface can be roughly located as shown in Fig. 3.3. Whatever the difficulties encountered to locate the soil-rock interface, we must keep in mind that the C or BC horizon is not the most altered stage of a weathered rock profile.

The C horizon systematically exists even in young soils in which the B horizon is absent or very thin. It may be reduced to a few millimeters forming a weathered skin on the hard surfaces of the substratum. On the contrary, in old thick soils formed under forest, the tree deep roots invade the substratum and locally induce an alteration. This altered zone does not have any of the characteristics of the A, E or B horizons. It should be considered as a C horizon exhibiting a sharp interface with the substratum (Fig. 3.4).

The Evolution of the Physicochemical Conditions During the Formation of Soil Horizons

Soils are shown (see Sects. 2.3.2 and 3.2.2) to need hundreds to thousands of years to be formed. This means that the sequence of soils shown in Fig. 3.4 results from a time-

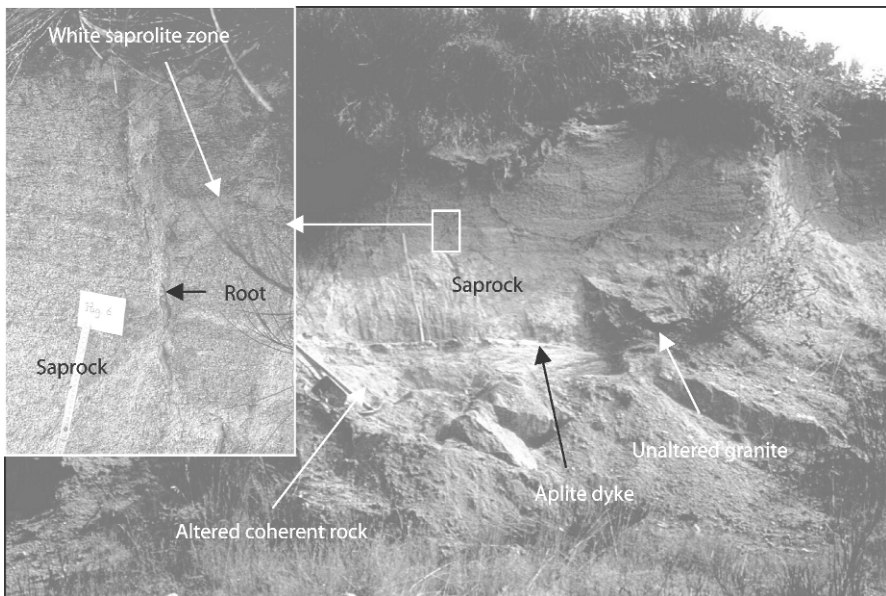


Fig. 3.4. Bleached zones around deep roots forming a C horizon inside saprock-saprolite formed on a granitic rock

dependent evolution. As an example of soil evolution, we shall consider the acid soil sequence. With time, the A, E and B horizons are formed. They are clearly identifiable in the most evolved stage (ulvisol). One can reconstruct the evolution during time as shown in Fig. 3.5a:

- t_0 : the saprolite is covered by lichens which trigger the first soil dissolution processes. The A and C horizons are millimeter thick,
- t_1 : vegetation grows on undifferentiated soil (cambisols). Chemical interactions between organic matter and silicates produce clay minerals while dissolution increases,
- t_2 : clay minerals are translocated from the A horizon. During the development of acid soils, they are dissolved. Then, an eluvial (E) and an accumulation (B_t) horizons appear. At this stage (young luvisol), the dissolution of silicate is intense,
- t_3 : clay translocation and silicate dissolution are active. The soil has attained its maximum development before degradation begins (old luvisol).

Because of translocation and dissolution, the amount of clay varies with depth. It is the lowest in the E horizon where eluviation is intense and the highest in the B_t horizon where particles sediment from the suspensions. Figure 3.5b shows how dissolution, translocation and accumulation affect the distribution of clay minerals in the soil profile during its development. Several consequences can be deduced:

- There is more clay in the A than the E horizon. The recycling of silica, potassium and other components through the vegetation sufficiently modifies the chemical conditions to stabilize the clay minerals in the upper part of the A horizon (see Sect. 8.5).
- As translocation progresses downward, it necessarily superposes on clay accumulation zones previously formed.

Considering the x , y and z points located at different depths in the profile, they will experience contrasted chemical conditions during the progressive deepening of the soil (Fig. 3.5c). At t_0 , the chemical conditions for all points are that of the rock or the C horizon. Then, point x will experience the chemical conditions imposed by the chemistry in the organic A horizon during all the soil evolution. Point y will experience first the B_t conditions and then the those in the E horizon where clays are actively dissolved. Point z remains in the C conditions from t_0 to t_2 , then those of B_t at t_3 .

The three active horizons in soil zones are A, where clays are abundant but also eluviated downward (i.e. physically transported in flowing solutions) E, when present, where clays are actively dissolved by acid conditions engendered by organic acids and B where the eluviated clays are accumulated.

One should remember that soils can be formed on other substrates than hard, high temperature rocks. Soils can develop on poorly indurated sediments, gravel or sand terraces, on salt marsh clay-rich sediments, and so forth. The processes and alteration zones are the same as those found for the alteration of hard, indurated rocks. However, their expression (seen in the field) is more difficult to observe and determine. For the most part, the A horizon is clearly present, identified by a darker color due to humic matter. The E horizon, when present, is indicated by a very light, sand-rich layer. However the C-alterite boundary is more difficult to identify.

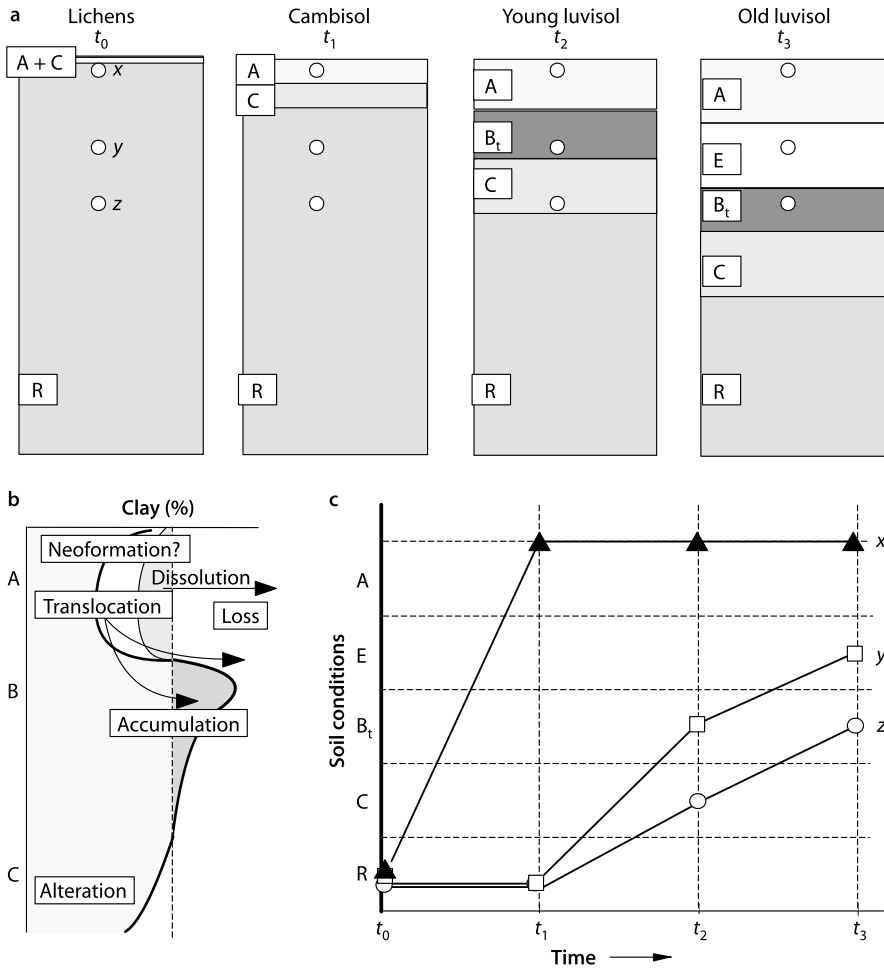


Fig. 3.5. A schematic representation of the development of a soil with increasing time from t_0 to t_3 . The effects are considered at different depths (x , y and z); **a** the progressive differentiation of A, E and B horizons in a luvisol; **b** dissolution and translocation during the soil development; **c** modification of the physicochemical conditions for points x , y and z during the soil development

3.1.3

Conclusion in a YES or NO Question Series

Are rock weathering and formation of soil independent? YES they are independent. In the Earth’s early history alteration was active before life appeared.

Do weathered rocks and soils form closed independent systems? NO, they form open systems which exchange minerals and chemical components.

Do alteration and pedogenesis contribute to reduce the rock heterogeneity? YES both transform the structure and the composition of rocks to approach the ideal equilibrium with surface conditions. As soil develops on altered rock by biological activity this process is intensified. The homogenization increases toward the atmosphere interface.

Are there limits to rock alteration and soil formation? YES even if altered profile may be several order of magnitude thicker than soils, both are limited in their development because of internal and external reasons:

- internal reasons: weakening of the driving force with increasing distance to the interface; chemical potential gradient and biological activity for altered rocks and soils respectively.
- external reasons: no systems at the Earth's surface may offer stable conditions for a long period of time because of climate change and/or, rejuvenation by erosion.

What are the rates of soil and alterite formation? The rate of soil formation is independent of that of weathering. Both are controlled by global (climatic) and local conditions (topography or position relative to the water table). Soils are formed during hundreds to thousand year periods; weathered mantles are formed during hundred thousands to million year periods.

Because classical definitions are sometimes ambiguous, we shall use here new ones (Table 3.1). When little information is present which would allow a determination of C-saprolite categories, we will use the term alterite to signify a material dominated by clays and oxides formed under conditions of water/rock interaction.

3.2

Dynamics of the Alteration Process under Temperate Conditions: An Investigation Comparing Soil and Rock Alteration in Profiles

3.2.1

Alteration in Temperate Climates

In the preceding section we have considered the physical description of alteration profiles, that proposed classically by pedologists and a new structural concept. The lower portions of alteration are dominated by rock water interaction where the pathways of water circulation determine the zones where alteration takes place. The result is an irregular spatial distribution of alteration zones where new clays minerals are found. The collapse of the old rock structure brings about a more homogeneous physical distribution of new clay minerals and the initiation of a horizon structure. A horizon, in the pedological concept, is the expression of general homogeneity on a horizontal scale.

In fact, the occurrence of clays in the classical description of soil profiles is dominated by the top most level, that of plant/silicate interaction. Here clays play a great role in the physical and chemical properties of the altered matter. The A (organic) and

Table 3.1. Summary of the definitions used in the literature and in the present book. The differences come from distinct interpretations of the development of soils and weathering mantles: two parts of a single continuous system (classical) or two independent open systems (concept developed here)

		Classical definitions		Definitions used here
Weathering profile (water/rock interaction)	Zones or facies	<i>Residual regolith</i>		<i>Alterite</i>
	Unaltered rock	Bedrock (consolidated rocks)		Unaltered rock (hard and soft substratum: granite, limestone, mudstone, mud, sand, ...)
	Conserved rock fabric	Coherent	Saprock	Coherent altered rock
		Friable	Saprolite	Saprock
	Non-conserved rock fabric	Pedolith		Saprolite
Soil profile (pedological alteration)	Horizons	<i>Horizon characteristics</i>		<i>Horizon characteristics</i>
	C	Limited upwards by the interface with the B horizon (no clear limitation downwards)		Limited by two interfaces – upwards with the B horizon – downwards with the saprolite
	B	Accumulation		Accumulation
	E	Lixiviation – dissolution		Lixiviation – dissolution
	A	Organic		Organic

B (clay rich) horizons begin the profile. Under conditions of high acidity, a clay-poor horizon (E) develops between the A and B horizons by clay dissolution processes. Below these zones one finds, usually, a C zone of clay-rich altered rock with varying amounts of rock particles, unaltered or partially altered high temperature minerals and clays formed through rock/water interaction. We would like to investigate the clay mineralogy of alteration by comparing the A-B horizon clay minerals and the C horizon clay minerals as reported in the literature. The reason for doing this lies in the question, do plants have a significant impact on clay minerals? We have established that high temperature minerals form clay minerals through the exchange of hydrogen ions with alkali and alkaline earth ions in high temperature minerals. Further alteration is effected by oxidation processes of iron in silicate minerals which destabilize them electro-chemically allowing the formation of new, more stable minerals. Once formed, the low temperature minerals such as clay phases would be the stable minerals as expected. However, in most mineralogical studies of alteration profiles one finds a continuing change in minerals, clays included, up to and within the A horizon. Why do these changes occur?

The initial assumption of clay change in alteration profiles is that the clays interact with rainwater, as do the underlying rocks, to form less alkali-rich and less siliceous minerals. However, under conditions of temperate climates the clays persist and one finds the alkali silicic clay mineral illite which is often the dominant clay mineral. We will consider the clay facies under conditions of temperate climate alteration.

The adjective “temperate” is important here in that our analysis considers initially only places where rainfall is in reasonable abundance and a very large portion of the year is “frost free”. This excludes deserts and tropical rain forests. Plants are an important part of the surface horizon in the alteration profile. Here the action of aqueous fluids on rocks is important but in the weathering profile the upper portions are occupied by a three-phase system, silicates (solids), water and air. Such a system insures a high and regular growth of plants and an equilibrium between plants, rock alteration and soil development. If the climate is too cold, not enough water/rock interaction will occur and the impact of plants is relatively small. If the climate is too dry only sporadic patches of plant life are installed. The altered surface is dominated by water/rock interactions and physical disaggregation of rock materials. In tropical climates, the very high rainfall tends to dominate with the result that the clay formation and transformations are dominated by unsaturated solutions bringing about the end of alteration, oxide and hydroxyoxides (gibbsite, ferri-hydrates, kaolinite) with the elimination of the 2:1 minerals. Only a very thin layer of material at the surface is occupied by plant/silicate interaction. Temperate climates, not too hot, not too cold, not too much rainfall, lead to optimum conditions of plant growth and plant/soil development. It is not an accident that the temperate climate zones of continents have supported the major development of the human species, where the ability of the Earth to give sustenance (agricultural productivity per surface area) is greater than in other areas. Ideal conditions are not all that common. Not all of the soils of the world can be used to produce grains, those plants capable of sustaining both human and animal populations which lead to the inevitable development of large, concentrated populations.

The richness of the Earth for agricultural mankind resides in the presence of 2:1 clay minerals; illites, smectites, mixed layer minerals and to a lesser extent HI minerals. Such situations require that the plants can grow and prosper and that the soils can furnish the needed sustenance and structure for the effective growth of these plants. However, even though high altitude, mountain prairie soils in Colorado (USA) produce interlayer illite-smectite minerals, this is not enough to give a high agricultural productivity to the soil because of the climatic limits. The lower, plain areas next to these massifs are better adapted to produce crops and they also have a mixed layer mineral mineralogy (see Birkeland et al. 2003 for an overview of the clay mineralogy of these climatic zones).

In order to understand the relations between plant/silicate interactions which form the upper part of alteration profiles and the rock/water interactions of rock alteration we shall consider the differences in the clay minerals which represent the surface mineralogy in the two zone A-B and C divisions and we shall consider the rate of change of minerals in the two zones in order to establish whether or not there is continuity between the two portions of soil profiles.

3.2.2

Kinetics of Alteration Processes

The Effect of Climate

As we have seen in the preceding section, water/rock interaction is the alteration mechanism which forms clay minerals first. The mineral type formed is determined

by the host or unstable silicate mineral in the rock substrate. These minerals are subsequently mixed intimately through the action of compaction as the rock structure is lost when the materials move upward towards the surface as a result of the action of surface erosion. The surface, or interface between the plant biosphere (root zone and surface vegetal detrital layer) leads to, as we hope to demonstrate, a new chemical structure or system where clay minerals react not only to the impact of slightly acid water, as is the case for the rock/water interaction zone, but more importantly with the complex of root exudates and bacteriological modification of this material which forms the humus and dissolved organic material of surface soil media.

One must keep in mind that climate (the external forces acting upon alteration systems), is the prime agent of alteration and it is basically governed by two variables: temperature and rainfall. One can simplify this to the variables of temperature and the concentration of chemical elements in the altering solutions. The more the solution has a high concentration of dissolved elements, the slower the process of mineral destabilization. The solution is tending towards saturation with dissolved material. Hence at high rainfall levels, one will have more alteration and above all mineral dissolution because of dilute solutions. The pertinent relation of reaction is $k = 1 / (A)$ where A is the concentration of dissolved matter in solution (Ritchie 1966). The amount of reaction (k) is linear with time and concentration. However the relation with respect to temperature is $k = A \exp^{-E/RT}$. Here the reaction relation with temperature is exponential as temperature increases. Roughly one can visualize these two relations as in Fig. 3.6. Such relations have been demonstrated by observations relating clay content of top soils (humic horizon) and rainfall (Jenny 1994, p. 155).

The effect of climate is then complex, a linear function of rainfall and an exponential one with temperature. In general, there is much more water-rock alteration in hot, wet climates than in cold ones, wet or not. Of course if there is little water available, hot or cold, the chemical alteration is very small. As a result in latitudes approaching the poles or on mountains, alteration is weak. As one comes down off the polar caps or from the mountain tops two possibilities are evident, either higher alteration rate with high rainfall or little under desert conditions.

The formation of the alteration and soil structure takes some time to become established in nature. Some chrono studies indicate just how much.

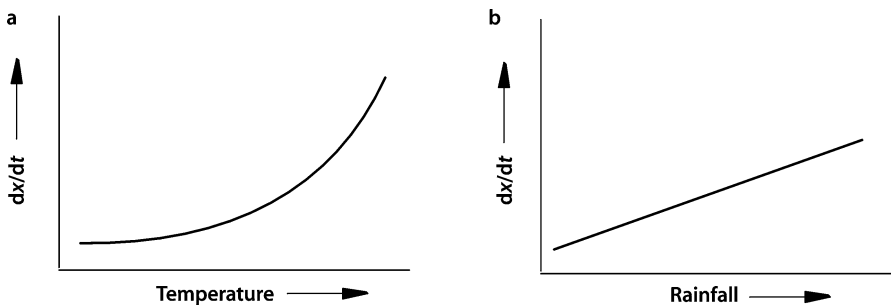


Fig. 3.6. Indication of probable relations between rainfall (a) and temperature (b) as they affect reaction rate (dx/dt) for the formation and alteration of silicate minerals at the surface

Initial Water/Rock Interaction

Establishing the rate of weathering is not an easy problem to solve. The most important point to resolve is the starting point, determined by unweathered material, and establishing the point in time at which weathering begins. Several approaches have been made.

First of all a very important experiment was carried out using crushed granite which was subjected to rainwater interaction for 30 years (reported in Millot 1964, p. 117–119). Investigation of the material after this period shows a reduction in grain size, with 2% entering the clay size category ($<2 \mu\text{m}$). However careful investigation shows that no clay minerals crystallized. Similar results were obtained in laboratory simulations. Physical breakdown of rock particles occurs (2% in the clay sized fraction produced in 30 years) but no new minerals were formed. This indicates that the water which flows freely through a rock system (macropore gravitational flow) did not become saturated with chemical elements that will form clay minerals but the near totality of the water/rock reaction is dissolution. Thus the first step necessary in soil clay mineral formation is obtaining water that approaches saturated with dissolved elements. This does not occur in macropore or free flowing water but in the capillary water. The first step in alteration is the dissolution of unstable minerals.

3.2.3

Kinetics of Soil Formation

Water-Rock Interaction

Colman (1986) uses measurements of weathering rinds (zones of partial mineral transformation) on boulders buried in glacial deposits to establish weathering rates. This is the initial rock/water interaction outside of the presence of plants and soils. Observations were made on boulders deposited by glacial action in moraine deposits. As can be expected, weathering rates decrease as the process proceeds. However the data gathered indicate very low rates of alteration, on the order of millimeters per 10^5 years. Whitehouse et al. (1986) indicate that weathering rinds on sandstone boulders in New Zealand where rainfall is abundant is on the order of millimeters per 10^4 years. Birkeland (1984, p. 94) indicates a rate of several millimeters per 10^3 years for alteration of a granite boulder in Colorado. Boulad et al. (1977) estimate that the advance of the alteration front into underlying granite in a soil in tropical Cameroon as being 70 mm per 10^3 years. The estimation is based upon disequilibrium dynamics of uranium in the altered and unaltered materials (Thomas 1994). These measurements indicate the penetration of the early stages of alteration into a heretofore unaltered rock which is unstable at Earth surface conditions. The values vary then from several millimeters per 1 000 years to millimeters per 100 000 years. The most rapid rates occur in hot, wet climates and the lowest in cold climates.

Soil Development (Water/Rock and Plant/Silicate Interaction)

Direct observations of soil formation are reported by Evans and Cameron (1979) who indicate that the rate of weathering resulting in the thickness of solum formation (the

sum of weathered material both new minerals and small grain sized particles due to physical weathering) occurs at a value of near 10 mm per 10^3 years on moraine material of Baffin Island. Observations are made on initially denuded rock surfaces which were left open to atmospheric effects with glacial retreat. The soil material develops, of course, in the presence of plants, mostly grasses (Fig. 3.7). Thus soils develop more rapidly than does an alteration front in a rock, given similar climatic conditions.

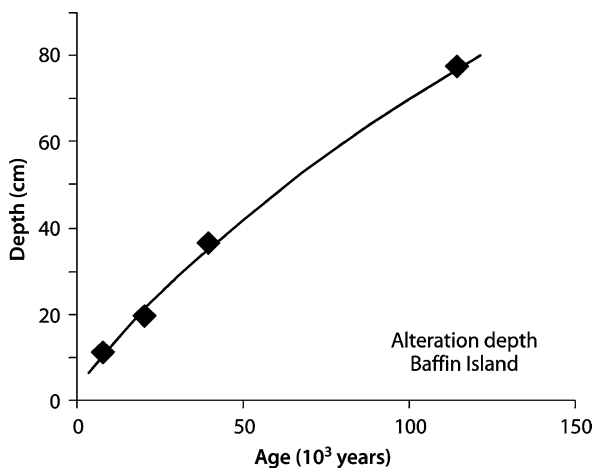
Another method of establishing weathering rate is to calculate chemical loss in a zone undergoing alteration within a mature profile context. Such measurements indicate rate of loss by both erosion but mostly by dissolution of high temperature minerals. Calculations of transport using dissolved matter in streams released from profiles 3 meters thick in Rhodesia (Owens and Watson 1979) suggest weathering rates of granite to be about 8 millimeters per 10^3 years. Dethier (1986) calculates that the loss from the saprolite and bedrock of volcanic rocks in the Pacific Northwest USA is on the order of 30 mm per 10^3 years.

The conversion of rock to clay, i.e. total reaction, can be measured as percent clay in the A horizon material. Righi et al. (1999) found a value of near 12% after 7 000 year weathering under alpine weathering conditions. From the shape of the curve of clay formation as a function of time, it appears that the rate slows greatly after 3 000 years (Fig. 3.8).

Thus total conversion to clays will be almost impossible under geologic conditions of erosion in such an area. However alteration of tephra in different parts of New Zealand, described by Lowe (1986), indicates a continuing change to clay of up to 85% conversion after nearly 750 000 years (Fig. 3.9).

Under these conditions the rock to clay conversion takes nearly a million years, a geologic number. Jahn quoted by Star (1994, p. 118) indicates that weathering of volcano-clastic materials under a semi arid climate may take 4 Ma to reach 80% conversion to clays. Hence total conversion of a rock to clay takes a significant period of time, somewhere between 1 and 4 Ma. Accomplishment of such a conversion necessitates that the soil be little eroded and the profile stable for long periods of time to be entirely transformed by weathering process.

Fig. 3.7. Depth of solum formation observed on Baffin Island as a function of soil formation time (data from Evans and Cameron 1979)



The above estimations of weathering rate are direct observations of processes under various climates (arctic to tropical) where the process is observed very locally (weathering rinds) or as an overall process in developing solum material. Jenny (1980, p. 242) has proposed a similar schematic description of weathering rates for the different parts of soil profiles under a variety of climates.

Considering these various studies which use different methods of observation to determine weathering processes under different climates, we can assume that the initial rate of development of clays in the first stages of weathering (water/rock interaction) is variable and rather slow when disconnected from plant activity, millimeters of alteration in hundreds of thousands of years. Climate and rock type can vary the absolute values greatly. Given these estimations, it would take between 6 000 to

Fig. 3.8. Proportion of clay in alpine soils as a function of time (data from Righi et al. 1999)

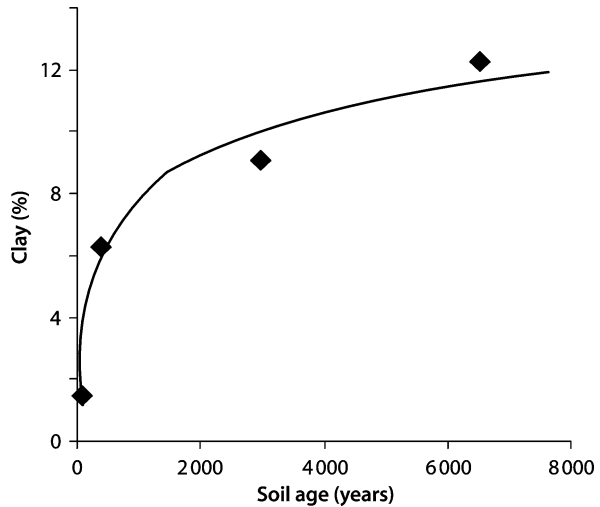
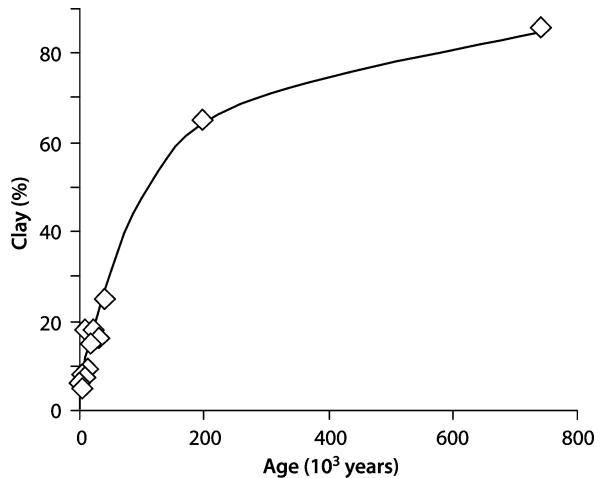


Fig. 3.9. Proportion of clay in the upper portion in tephra soils as a function of time (data from Lowe 1986)



4 000 000 years to form 40 cm of altered material, clay and disaggregated rock, by water rock interaction depending upon the climate and other factors such as the intensity of erosion by physical transport, and plant activity. Layers of solum as thin as 40 cm are those of rather young profiles which are usually not well differentiated into A, B and C structures. One would expect to encounter the faster rates estimated in tropical climates and those slower in temperate to mountain climate alteration processes.

The formation of clay minerals, rock + water to clay, takes on the order of 1 Ma under conditions of moderate climates.

Establishing Clay Mineral Facies in Soils (Plant/Silicate Interaction)

There are some estimations of the time necessary to establish an equilibrium clay mineralogy in the upper portions (B, A horizons) of a soil. This information is extremely important as a basis in understanding the kinetics of soil mineral formation after rock alteration. Bain et al. (1993) give data comparing loss of Ca (which can be interpreted as the weathering of primary minerals) from the A horizons in Scottish soils and cation exchange capacity (formation of clays). The curves of measured values against age of the soil indicate stabilization after 40 000 years. Amounts of clays are small, on the order of several percent. Righi et al. (1999) indicate that the formation of the stable smectite S/I mineral in the upper horizons of a prairie podzol as well as gibbsite in the high altitude Swiss Alps is complete after about 3 000 years of weathering of glacial moraine material. Observations used to determine this time frame range from 80 year old moraine materials subjected to weathering to those of 6 500 year age.

Egli et al. (2004) demonstrate the formation of smectite S/I minerals in the upper zones of glacial moraine material in similar situations to those of Righi et al. (1999). Here the relative proportion of the different new minerals was compared in the upper portion of soils over an 11 500 year time span for two sites. The change in abundance of the new minerals begins to become very small after about 2 000–3 000 years of development, a conclusion very similar to that showed by Righi et al. (1999). The shape of the curve clearly indicates an exponential function, compatible with kinetic formulations. These two studies indicate that it takes near 3 000 years to establish clay mineral assemblages in a soil profile under alpine prairie conditions where the substrate is granitic moraine material. Mahaney and Halvorson (1986) indicate similar time scales for soils developed on a variety of rock substrata in the Wind River Mountains (WY).

Thus it appears that the stabilization of soil clay mineralogy occurs at a rather more accelerated rate than that of water/rock interaction, thousands of years compared to a million. However, one should note that all of the studies available are made under conditions of new soil formation at low temperatures, mountain climates. The temperature factor will be quite important. Soils forming at lower altitudes, and in warmer climates will develop faster.

Formation of Humic Layers

The humic layer is a direct expression of plant activity in the alteration profile.

Observations of color change in poldered sediments kept in prairie over different periods up to 850 years (Velde et al. 2003) indicate that after the maximum period observed, the humic zone continues to develop only slowly under prairie in sediments

(Fig. 3.10). It is probable that the development of the humic horizon will stabilize at near 1 000 years. This is towards the upper limit indicated by Birkeland (1984, p. 225). Data from Egli et al. (2002) indicate that both N and C content in alpine prairie soils reaches a more or less stable content after 4 000 years development (Fig. 3.11). It is clear that the effect of temperature is important in establishing the stable humic layer chemistry, varying the time necessary to reach equilibrium by a factor of four.

Comparison of Reaction Rates in the Different Parts of an Alteration Profile

In the above summary of several studies, it appears that the formation of the different parts of an alteration-soil profile occur at different rates. The initial stages of rock weathering (altered rock zone) have rates of formation of millimeters per 10^3 years and hence 10^{4-5} years are needed to form a minimal solum under intense weathering

Fig. 3.10. Observed thickness of humic layer in soils developed on clay-rich sediment (data from Velde et al. 2003)

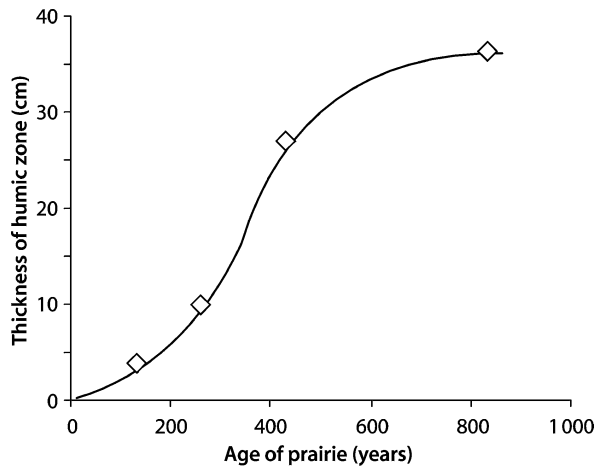
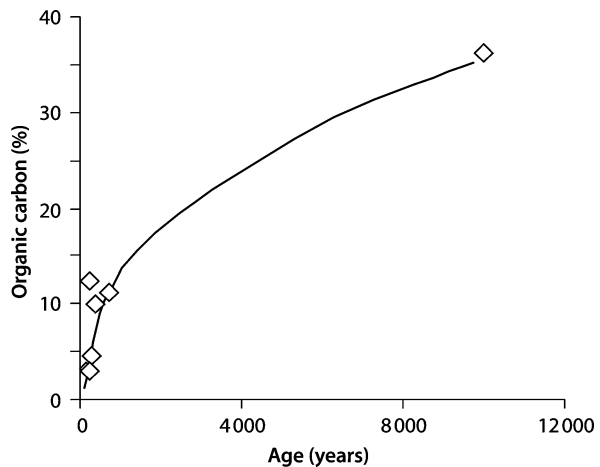


Fig. 3.11. Portion of organic carbon in surface soils in Alpine environment as a function of time (data from Egli et al. 2002)



conditions. Formation of stable clay mineral suites in the C-B horizons take near 3×10^3 years to reach a steady state. The formation of the humic layer takes less than 10^3 years. These numbers are only indicative in that the examples are taken from very different climatic sites, mostly in mountain areas. The time scale of mineral formation and mineral interactions in weathering necessary is given in Table 3.2.

The above hierarchy of reaction rate is important in further considerations of soil profile development and mineral equilibrium. Initially, it is clear that the water-rock interaction is the slowest process and thus will determine the extent of solum development, i.e. the amount of material converted to minerals (essentially clays) stable at the Earth's surface. Next the clay species and assemblages appear to develop in the soil (A-B) zone at a faster rate. Most rapid is the development of the humic, plant-soil interface. It seems clear that the development of an alteration profile is not one single, inter-connected sequence. The several zones react at very different rates. The slowest is at the bottom and the most rapid is related to plant/silicate interactions.

3.3 The Inter-Relation of the Dynamics of the Alteration-Soil Profile Sequence

3.3.1

Overview of Soil and Weathering Mineralogy

In Chap. 2 one finds that rocks alter to clays under specific conditions of interaction between aqueous solutions and unstable minerals. The sequence of rock to altered rock to saprock to saprolite is one of progressive change in the density of the material, signifying the dissolution and transport of material in solution out of the immediate alteration zones. New silicate minerals occur as a result of the interaction of slightly acidic aqueous solutions with the minerals present which were formed at higher temperatures than those of the surface. This process leads to the loss of divalent and monovalent ions, such as K, Na, Ca, Mg. The end result is the concentration of the major elements Al, Si and Fe. The balance of mono and divalent ions determines the fertility of the alteration products. Most common agricultural crops need some mono and divalent ion sustenance. This is of course especially true in temperate climates where plants derived from grasses are the major source of agricultural benefit. It is clear that trees can thrive more easily on substrates containing less 2:1 phyllosilicate material. They are planted, in temperate climates, on soils containing less 2:1 silicate material, and especially that of the smectite-illite type.

In the initial stages of alteration, excepting desert environments, plants install themselves on rocks. Here we have the interaction of plants (and the organic matter associated with them such as leaf litter, roots, and root exudates) in direct contact with the unstable silicate material. In almost all situations the first step in alteration is the formation of the soil (plant/silicate) interaction zone. The chemical action of plant material dominates in the soil or fine material of the A horizon. However, the initial stages of alteration in rocks are dominated by microsite reactions within the coherent rock itself, where the presence of unstable minerals dominates the mineral reactions. The production of 2:1 minerals dominates in such reactions. However, one finds the illite, smectite, vermiculite 2:1 minerals most often accompanied by the 1:1 mineral kaolinite.

Table 3.2. Time scale of mineral formation and mineral interactions in soils

Horizon	Action	Time (years)
C	Water/rock interaction	1 000 000
B	Clay mineral facies in A-B horizons	3 000
A	Stable formation of the humic horizon	1 000

Depending upon the initial rock composition one can find other phases, talc, chlorite, zeolite among others, which can at times be important in quantity but which rapidly become unstable as the altered material becomes saprock and saprolite. Rocks containing metamorphic and diagenetic phyllosilicates such as chlorite, biotite and muscovite change to a lesser degree because the minerals present are less in disequilibrium with the surface chemical forces than higher temperature phases such as feldspars or pyroxenes. They are already phyllosilicates. The relative instability of minerals at the Earth surface reads in the same sense as Bowen's reaction series for minerals at high temperature. Sedimentary rocks show less of a tendency to change in that the material is even closer in mineral nature and thermodynamic stability to soil clays, i.e. fine-grained illite, mixed layer minerals and chlorite, kaolinite assemblages. The difference between the free energy of these minerals and that of the new or eventual clay minerals under surface conditions is less than that of say an olivine grain or a plagioclase and a clay mineral. Nevertheless chlorite, muscovite, diagenetic illite, and diagenetic illite/smectite minerals are not stable in the alteration environment. This is witnessed by the fact that soil clays change mineralogy with burial near 50–80 °C to become a full smectite mineral (Meunier and Velde 2004, p. 82). Surface clay minerals have a specific stability field of temperature and are modified by prolonged heating above about 50 °C.

The microsystems in the old rock structure are gradually effaced through the collapse of the old mineral structure pseudomorphing the initial rock mineralogy. The saprolite structure becomes more prevalent in the middle zones of an alteration profile. In the saprolite zone the clay minerals of surface origin become more intimately mixed and lose the chemical specificity of their microsite configuration. This structure is destined to progress downward with time into the rock substrate. In situations of low erosion rate such alteration can be traced to depths of tens of meters or more (see Millot 1964). Thus under conditions of normal or rather high rainfall, the alteration of rocks proceeds downwards with time. Saprolite and saprock are extended downwards.

If there is uplift, erosion becomes important and the surface layer of an alteration profile is taken away at various rates depending upon slope or the difference between the flowing river structure and the surface of the profile. Here the surface approaches the rock, or the rock "moves upward" depending upon the relativistic point of view one has. Under high rates of erosion, rock bodies appear at the surface. On average, where granular, macrocrystalline rocks are concerned, the rock alteration interface is quite irregular. In the case of sedimentary rock of low grade metamorphism, the interface is less irregular. This interface determines the richness of the soil. If it is too near the surface the farmer says that it is rocky, it is too far below the soil of the Earth he farms it is too "heavy" i.e. rich in clay.

3.3.2

The Mineralogy of Soil Horizons

The C Horizon Mineralogy

This is the top of the water/rock interaction zone. In pedology the altered rock (saprock) and more fine-grained layers (saprolite) have often been combined into one horizon for consideration in a nomenclatural scheme: C horizon (see Sect. 3.1.2). However the differences are great between the two concerning structure and water flow. The characteristics of the C horizon are based upon grain size distribution (stages of rock alteration since alteration materials are normally more fine grained than the source rocks) and dispersed pedologic features (presence of scarce clays or organics deposits from upper horizons; alteration around deep roots). One must remember that soils can develop on all sorts of geologic material; sand dune, sediments, sedimentary rocks and eruptive rocks for example, and hence the starting materials can be in the form of a rock, a transported sediment or soil. Hence the transformation from source material to soil is diverse and varied. However, considering a solid, metamorphic, sedimentary or crystalline rock, one can describe a succession of stages in alteration progression.

In macrocrystalline rocks, the effects of microsystem alteration are primordial. They are less evident in rocks with smaller initial crystallites but nevertheless the microsystem reactional scheme is operative. The most striking effect in the C horizon is the loss of density. Much material is dissolved and transported out of the initial system. This can be seen in the conductivity or permeability curves in soil profiles on alteration profiles (see Chap. 4).

The C horizon is dominated by minerals stable under conditions of interaction of surface (free flowing) aqueous solution and unstable high temperature minerals.

The B Horizon Mineralogy

In ideal, "average" conditions, one has a reasonable saprock and saprolite zone, called the C horizon by most pedologists, and a plant interaction horizon, the A horizon, at the interface with the atmosphere. In between these two zones is the B horizon where a portion of the clay material in the upper A horizon is concentrated through downward water flow until it is stopped by lowered pore space. This is the zone of accumulation where clays and some organic matter are accumulated. However if this is a zone of clay accumulation, not all clay mineral material present comes from above. The B horizon is a transit zone where the C horizon clays will move upward to reach the surface through erosion. They are mixed with clays from this upper, erodable A horizon.

As far as clay minerals are concerned, the B horizon is a zone of mixture of surface (soil) and subsurface (C horizon, saprolite) clay minerals.

The E Horizon Mineralogy

This zone, present in acid soils, is dominated by the development of a clay-free system where alumino-silicates and oxides are largely absent due to dissolution processes.

The A Horizon Mineralogy

Clays here are dominantly the result of plant/clay interaction. This is governed by the chemistry engendered by plant detritus decomposition and transformation as well as root exudation material and its decomposition.

The clay minerals in the A horizon where plants are established reflect the new chemistry and chemical equilibria engendered by plant activity. This portion of the alteration profile is by far the most complex, where plants (roots and litter) are modified by bacteria, fungi, burrowing animals and other living organisms. The variety of chemical products is multiplied greatly compared to the rock/water regime. Much work has been done on the interactions of the living or organic material but relatively little has been done on the organic-clay complexes. Thus we ask more questions than can be answered reasonably in this domain. This is not due to a lack of scientific effort, the agronomic community has been active for more than a hundred and fifty years to be sure. The system is however extremely complex, variable and the objects to be observed are normally of extremely small size. Recent improvement in instruments have allowed immense progress but these have been available on a routine basis only for several decades and we are only beginning to make significant strides in understanding the processes operative in the A soil horizon.

Overall, in the A horizon bacterial and other actions change the organic matter to a large extent. The main action is oxidation of a large portion of the carbon and nitrogen. This is called mineralization. This effect is measured in a relative manner by observing the elemental ratio of nitrogen to carbon (C/N ratio). The ratio is not an absolute measure of carbon loss in that much nitrogen is lost through the production of gas also. However, one consistently sees a change in C/N ratio which indicates the humification of the soil organic matter (Stevenson 1986). Besides the C and N ratio, humification increases the amount of carbon relative to oxygen present in the organic matter as larger molecules are produced which contain fewer radicals. Humus is associated with soil fertility in agricultural terms. In the A horizon, soils tend to form stable aggregates, conducive to good plant growth. They promote capillary retention of water. Thus the soil horizon is or was the focus of research into the fertility of soils for agronomists.

An interesting point, made by Dethier (1986) considering mass balance studies of weathering is that the loss of material by dissolution from the A horizon is small compared to that from the zone of water-rock interaction. It would appear that the soil zone is one of conservation of material due to dissolution while that below is one of great loss.

The A horizon is the agricultural one. It determines the fertility of crop growth and the mechanical conditions under which the land can be utilized. Changes in agricultural practice can change the mineralogy and the soil structure, rendering it less amenable to crop growth or more amenable and more fertile. However the A horizon is the most susceptible to physical erosion, and hence the most fragile in the short term. It is this effect that lets the clays from the C horizon move upwards to the surface soil horizon.

3.3.3

Mineralogical and Chemical Differences between Alteration and Soil Zones

In Sect. 3.1 we have seen the physical and structural relations between the different parts of an alteration profile. Since our main objective is to describe the clay formed

in surface environments, it is useful to establish the differences between the lower C horizon and the upper A-B horizons which are influenced by plant/silicate interaction. Given that the soil horizon interface between plants and weathering clay minerals is one where new chemical forces (organic matter) come into action as compared to those of water/rock interaction at the bottom of the weathering profile, how can one establish this contrast?

In the plant/soil interface we have the interaction of mineral and organic material which reacts to form new minerals according to chemical potentials engendered by the activity of plants at the surface; roots and decomposition of detrital organic matter. This soil horizon designated as A by classical pedology is at times sub-divided into different categories. If the weathering of rocks to form clay-rich strata is well understood, or accepted as presented in the preceding chapters, the plant/clay interface which we call soil horizon in this discussion is less accepted as a specific system producing new clays or at least less considered as a special part of an alteration profile even though it was proposed as such by Duchaufour (1979). Most pedologists consider the rock to organic-rich surface zone as a continuum of inter-related reaction horizons. However, if one considers the structure of a soil profile, one sees an upper, organic-rich zone (A), a horizon of clay concentration due to physical displacement of fine particles (B) and a gradation of zones with decreasing clay content toward the rock (C).

The upper, A, horizon is not only one of high organic matter content but also the zone where much of the biological activity in a soil is concentrated. Plant roots are in high concentration. In general more than 50% of the root mass of plants is found in the topmost 20 cm of the profile (Schenk and Jackson 2002). The exudates of these roots are strongly present. Bacteria are especially active here also, where they transform leaf and other plant litter (entering the soil from the surface). Burrowing animals continually work and re-work this portion of the alteration profile. This horizon is one where mineral encounters living matter.

In considering the whole soil, Oh and Richter (2005) present data indicating variable behavior of different elements in deeply altered profiles, 4 to 5 meters thick. Phyllosilicate rock loses relatively little potassium but most of its Na, Ca and Mg is lost. Gneiss loses about one third of K, and almost all of the Na and Ca present. Diabase rock loses two thirds of its K, three quarters of Ca and Na, but almost all of the Mg present. By contrast, the typical profile of overall potassium content in intensely weathered tropic alteration is one of continual potassium loss (Mohr and Van Baren 1959, data p. 142). The same trend is seen for an alpine fir forest alteration profiles from low altitude forests (Egli et al. 2002, 2004). Depending on rock type, climate and plant regime the loss of alkali and alkaline earth ions is more or less great.

As we have seen in the preceding chapter the alteration reaction zone (water/rock interaction) is first governed by microsite reactions effected by descending, gravity-driven aqueous solutions. These processes are the slowest to proceed in the alteration process. At the lowest levels of the alteration profile the interactions are those driven by the instability of high temperature minerals, often anhydrous ones such as feldspars, and slightly acidic aqueous solution. In the upper reaches of such reaction zones one finds faster reactions which develop inter-stability of clay mineral species which were produced by the first reactions in micro-sites of deeper sub-surface origin. New phases occur and relative proportions of old minerals change. Normal surface erosion brings these clay zones, formed at depth, up to the surface and hence into the zone of interaction with

plants at the surface. The reaction times in the upper most horizon, A, are the most rapid and hence new equilibria determine the minerals which will be present.

3.4

What Are the Clay Mineral Assemblages?

3.4.1

Formation of Different Clay Mineral Phases in A Horizon

In order to establish the new chemistry affecting clay minerals in the soil, plant zone, one should look at published work on clays in alteration profiles in order to see whether or not the clay mineralogy does in fact change in the root, organic-rich part of the profile compared to that in the lower zones. To do this we must often “re-read” published data because it is rare that authors have considered this possibility. Thus we look at published X-ray diffraction diagrams to see whether or not changes in mineralogy occur at the interface of the high density root zone, the A horizon in most cases. Some examples of such interpretations are given below.

Smectitic, S/I minerals in Soil (A) Horizons

Fichter et al. (1998) indicate that there is a strong change in mineralogy in the clay-rich upper horizons in acid, forest soils developed on hydrothermally altered granite in the Vosges mountains, France. New expanding mixed layer clays occur rather abruptly in the upper horizons at the expense of smectite in lower zones.

Righi and Lorphelin (1987) present data for a toposequence of prairie soils developed in the Himalayas where clay minerals change rather abruptly in the upper B-A horizons compared to those present in deeper portions of the profiles. Mixed layer minerals become very important in a podzol but are lost in a brown acid soil in the A-B horizons. Gibbsite, found in the lower portion of the profile (aluminum hydroxide), is lost in the upper A horizon.

Olsson and Melkerud (1989) give XRD spectra which indicate the formation of a smectitic mineral in the E horizon of podzol soils (part of a complex A horizon) based on till in Southern Sweden. The lower portions of the profile show chlorite and vermiculite with minor amounts of illite. One can assume the formation of smectitic material from vermiculite in this upper, plant influenced horizon.

Ross et al. (1987) indicate that the mineralogy of the A horizon of soils developed from marine clay for 8 000 years near Ottawa (Canada) shows a loss of illite and a strong gain of smectitic mixed layer mineral compared to lower horizons.

Takahashi et al. (1993) give data which indicate a strong loss of illite from the clay assemblages in going from C and B to A horizons in soils on volcanic materials in California. The plant cover is conifer. Much of the silicate material has been transformed into hydroxyl-oxides and forms of kaolinite. However, it appears that illite is replaced by a HI or smectitic mineral in the A horizon.

Alpine profiles studied by Egli et al. (2004) indicate that vermiculitic material is lost at the expense of smectite in the AE or E (upper plant influenced) horizons of soils developed on rhyolite. The amount of smectite present appears to be related to climate differences.

Harris et al. (1987) indicate that the A and E horizon mineralogy of Ultic Haplaquods developed on Plio-Pleistocene sediments in Florida woodlands present a kaolinite/smectite mineralogy whereas the lower horizons show HIV mineralogies. These authors invoke mineral reactions to produce the smectite in the upper soil horizons.

April et al. (2004) indicate that in spodosols formed on crystalline rocks in the Adirondack mountains, smectite is prevalent in the A, E horizons while HI minerals are present below these horizons.

Buurman et al. (1976) present data which indicate that smectite forms at the expense of chlorite and HI minerals in calcareous micaschists forming cryorthod soils under alpine conditions. Mica seems to be stable for the most part. The vegetal cover is prairie and grazed conifer forest.

Tevelde et al. (1990) give data which indicate a loss of a hydroxy-interlayered mineral (vermiculite) and a gain of smectite in the E-O layer of a spruce forest on sandy soil in Norway.

Droste et al. (1962) indicate that smectitic S/I forms from illitic I/S minerals in soil horizons in alteration profiles developed on Wisconsin till and limestone in Indiana.

Illite, Illitic I/S Minerals in Soil (A) Horizons

An example is given by Kuzila and Lewis (1993) where loessal soils in Nebraska were studied. Here we see a smectite-rich interlayered mineral (S/I) of loessal parent material origin transformed to a more illitic I/S mineral in the upper horizon of a prairie soil.

Spiers et al. (1984) show data which indicate the strong shift in illite content of mixed layered minerals and the formation of vermiculite in the A horizon of Solenetz soils in Alberta (Canada). The zeolite clinoptilolite was identified in this horizon also.

Badraoui et al. (1987) show RXD spectra which indicate an increase in illite content of the S/I mineral and the appearance of illite in the A horizon of tall grass vertisol soils in Minnesota.

Tice et al. (1996) give data which indicate the formation of illite in the A horizon of a sandy loam based upon diorite rock in the San Gabriel mountains, California. The study reports results on a comparative study of the effect of either oak or pine forest planted on essentially scrub soils. In the A horizon of the oak forest, 41 years after planting, one finds a strong illite peak along with the initial HI mineral found in the original soil as well as in the lower portions of the profile. The pine forest showed no change in clay mineralogy between the initial state or at different points at depth.

Madsen and Nornberg (1995) also demonstrate that the type of vegetal cover can influence the clay mineralogy of the A horizon in soils derived from glaciofluvial material in Denmark. Grass cover favors the formation of illite while spruce shows a strong tendency to produce HI minerals. The lower portions of the soil profiles show I/S and S/I plus illite as the major clay minerals.

Boettinger and Southard (1995) present data which indicate a change of smectitic I/S minerals into more illitic I/S minerals in the upper horizons of sage and grass desert soils derived from granitic materials.

Vermiculite and Hydroxy-Interlayered (HI) Minerals in A Horizons

Hussain et al. (1984) show XRD spectra which indicate a change of a smectitic clay assemblage (S/I) to one with a strong vermiculite component in the A horizon of chestnut soils in Iraq. Parent material is unidentified. These soils are now farmed.

Pevear et al. (1984) show data indicating that HI mineral (HI expandable) is present in glacial marine drift soils in the Pacific Northwest at the surface (A and upper B horizon) at the expense of smectite in lower portions of the profiles.

Ransom et al. (1988) present data suggesting that smectite (S/I) is lost to the advantage of HI in the upper B horizon of an Aqualaf soil in Southwestern Ohio. This is especially important in the coarse clay fraction ($<2-0.2 \mu\text{m}$). The clays collapse upon potassium saturation suggesting that the HI is an HI of smectitic character with exchangeable interlayer alumina.

Graham et al. (1990) indicate that weathering profiles on micaceous colluvium and serpentine show a change of smectite to HI in the A horizon of Argixeroll soils developed under sparse shrub growth in northern California.

Soller and Owens (1991) give X-ray data which indicate a loss of mixed layer (I/S and S/I) minerals from the B horizon of a sandy soil in Maryland USA and an increase in HI mineral in the upper horizon.

Millot and Camez (1963) indicate that HI mineral is the product of transformation of illite in sandstone rocks under forest in Alsacian alteration profiles.

In old soils formed under humid conditions in the Southeastern United States one finds gibbsite and kaolinite dominant at depth while HI minerals are dominant at the surface under forest soil conditions (Norfleet and Smith 1989; Bryant and Dixon 1964).

Summary of Possible Mineral Changes in A Horizons

The major mineral changes cited above are summarized in Table 3.3.

It is clear that almost all clay species appear in the A horizons compared to other minerals found below in the alteration clay materials in an alteration profile. That is, almost all types of minerals appear to be formed from other types if one considers that the evolution of clays is from the bottom of a profile upwards. Such a variety of changes can most likely be attributed to changes in soil chemistry (organic and inorganic) in the A horizon but which do not pertain in the lower zones of the alteration profile.

Given the wide variety of apparent mineral reactions observed in going from altered rock or parent material to upper horizon soil clay mineralogy, some the reverse of others, it appears that the change in chemical and mineral stability is not the same in all types of soil (A) horizons. The only variable possible, obvious at that, is the type of plant present which forms or influences the A horizon. If plants can influence the clay minerals present they must be creating a specific chemical environment typical of their species or group. This situation is indicated in the work of Madsen and Nornberg (1995) where they compare the clays developed under different plant regimes in sandy soils in Denmark. They find that the type of plant cover can determine the type of clay mineral developed from essentially the same subsoil. Spruce develops HI, grass illite and oak or heather effected little change in the clays.

Table 3.3. Summary of major mineral changes from C to upper horizons in soils

C horizon-alterite material	Soil clays
Illite	Smectitic mixed layer mineral
Illite	Vermiculite
HI plus gibbsite	I/S
HI	Illite
HI	Smectite
I/S and/or S/I	Illite or HI
Smectitic S/I	Illitic I/S and/or HI
Chlorite,HI	Smectite
Gibbsite, kaolinite	HI

The contrast between soil clay mineralogy (A horizon) and that of the lower B, C horizons in the profiles cited above indicates that the chemistry of the aqueous solutions (capillary) in soil and alteration profile is not the same. This reflects the difference between free or bulk water which flows freely downward in the profile due to gravitational forces and the capillary water which is more strongly held in the alteration material and which in fact moves upward in the profile. Depending upon the rate of rainfall, bulk water, flowing in pores and fractures is essentially in equilibrium with the atmosphere, having but little chemical exchange with the capillary water intimately held by the clay particles or it is in various stages of equilibrium with the clays in the soil zone. Data of Ranger (2002) support this view. In this work forest soils were investigated for pore water and capillary water. Significant differences in the concentration of dissolved elements can be seen as well as specifically different elemental ratios between the two water types. The chemistry of alteration and clay stability in the B-C horizons is that of the capillary water. Hence the alteration process in each portion of an alteration profile is governed by microsite chemical reaction and, in the case of the soil horizon, by the transformation of other non-mineral material which interacts with the silicate matter. Here the chemical forces which dominate are other than those of the alteration horizons.

3.4.2

General Schema of Alteration Zone and Soil Relations

Figure 3.12 shows the relations of rock, saprock, saprolite, clay accumulation and plant/clay interaction. These are considered in the context of the present definitions of the A, B and C horizons and the interface with the substratum.

In the lower most part of the alteration profile the water/rock interaction is initiated. Here clays are formed under essentially mineral chemical constraints. These processes are generally considered to be slow demanding several thousand years or several tens of thousands of years to come to chemical equilibrium by the transformation of rock to clays. The major process in this zone is that of hydration and chemical

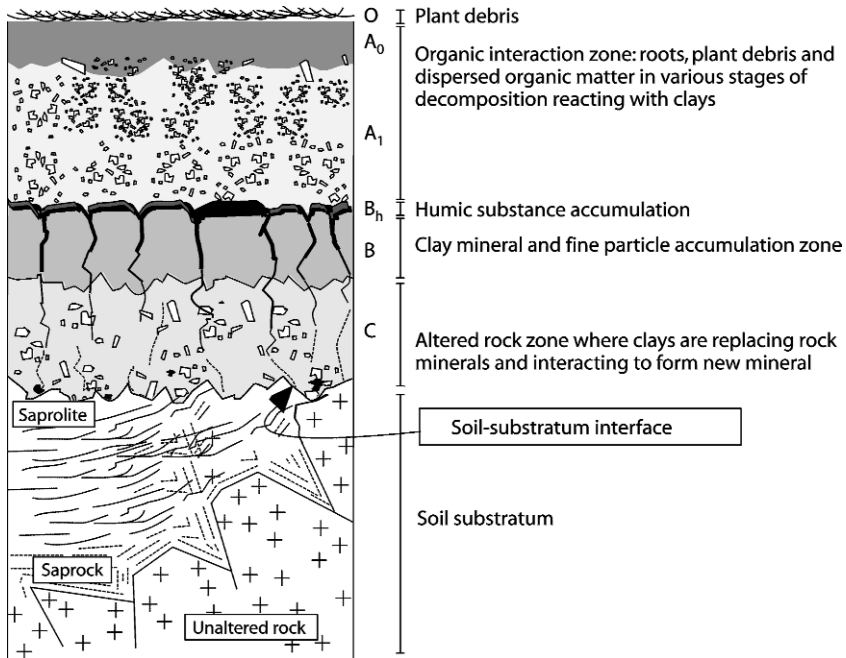


Fig. 3.12. Schematic representation of the components of a soil profile and different chemically active parts of the profile

dissolution. The initial source minerals are normally less hydrated (H_2O and OH molecules) than the soil clay minerals which form from them.

By contrast, the A horizon is one of relatively rapid change necessary to come to an equilibrium between plant chemical regime and that of the silicates. The starting material for reaction in this zone comes from the rock (altered or not) through erosion of the surface soil layer and from other clays which have been moved down in the soil profile by water flow gravitational movement and subsequently mixed with the C horizon clays, largely dominated by minerals formed by water/rock interaction but still influenced to a certain extent by the presence of organically derived chemical components (essentially roots).

The B horizon is one of mixing, the surface clays and the alteration clays finding themselves physically mixed and eventually brought to the surface. The B is the re-cycling horizon where illuvial transport of clays puts them in contact with new, water/rock interaction clays. This mixture is gradually cycled upward when the A horizon is taken away by erosion. The clays reaching the A soil horizon are a mixture of old, equilibrated clay minerals and new, unequilibrated minerals.

The most important aspect of this schematic representation is the reality of two different reaction rates, slow in the substrate and the rather rapid reaction in the A horizon. This is the plant interface. The energy of plant and living material is very great. Clearly the input of living organisms can strongly affect the stability and the kinetics of mineral reactions.

3.4.3

Overview of Alteration in the Soil Zone in Temperate Climates

The adjective “temperate” is important here in that our analysis considers only places where rainfall is in reasonable abundance and a large portion of the year is “frost free”. Here the action of aqueous fluids is important but in the weathering profile the upper portions are occupied by a three-phase system, silicates (solids), water and air. Such a system insures a high and regular growth of plants and an equilibrium with rock alteration and soil development. If the climate is too cold, not enough water/rock interaction can occur and the impact of plants is relatively small. It is not an accident that the temperate climate zones of continents have supported the major development of the human species, where the ability of the Earth to give sustenance (agricultural productivity per surface area) is greater than in other areas. Ideal conditions are not all that common. Not all of the soils of the world can be used to produce grains, those plants capable of sustaining both human and animal populations which lead to the inevitable development of large, concentrated populations.

The richness of the Earth for agricultural mankind resides in 2:1 clay minerals; illites, smectites and to a lesser extent HI minerals. Such situations require that the plants can grow and prosper and that the soils can furnish the needed sustenance and structure for the effective growth of these plants. Thus, even though high altitude, mountain prairie soils in Colorado (USA) produce interlayer I/S minerals, this is not enough to give a high agricultural productivity to the soil. The lower, plains areas next to these massifs are better adapted to produce crops and they also have a mixed layer mineral mineralogy (see Birkeland et al. 2003 for an overview of the clay mineralogy of these climatic zones).

Alteration in very dry climates is largely determined by different processes of rock fragmentation, due to differential thermal expansion of constituent minerals. It is mechanical without the formation of new clay minerals. At the other extreme, alteration under conditions of very high rainfall for long periods of time produce at a final stage hydrous oxides of iron and alumina, hematite-goethite and gibbsite. In both of these extreme soil conditions, plants are overwhelmed by climate, either too little water or too much.

The following chapter gives the processes of clay formation in the rock/water interaction zone.

Suggested Reading

- Birkeland PW (1984) *Soils and geomorphology*. Oxford University Press, 372 pp
Duchaufour P (1991) *Pédologie. Sol, végétation, environnement*, 3rd edn. Masson, Paris, 289 pp
Jenny H (1994) *Factors of soil formation*. Dover, New York, 281 pp
Meunier A (2005) *Clays*. Springer-Verlag, Berlin, 472 pp
Millot G (1964) *Géologie des argiles*. Masson and Cie, Paris, 499 pp (English translation 1970: *Geology of clays*. Springer-Verlag, New York, 429 pp)
Stevenson FJ (1986) *Cycles of soil: carbon, nitrogen, phosphorous, sulfur and micronutrients*. John Wiley and Sons, New York, pp 356