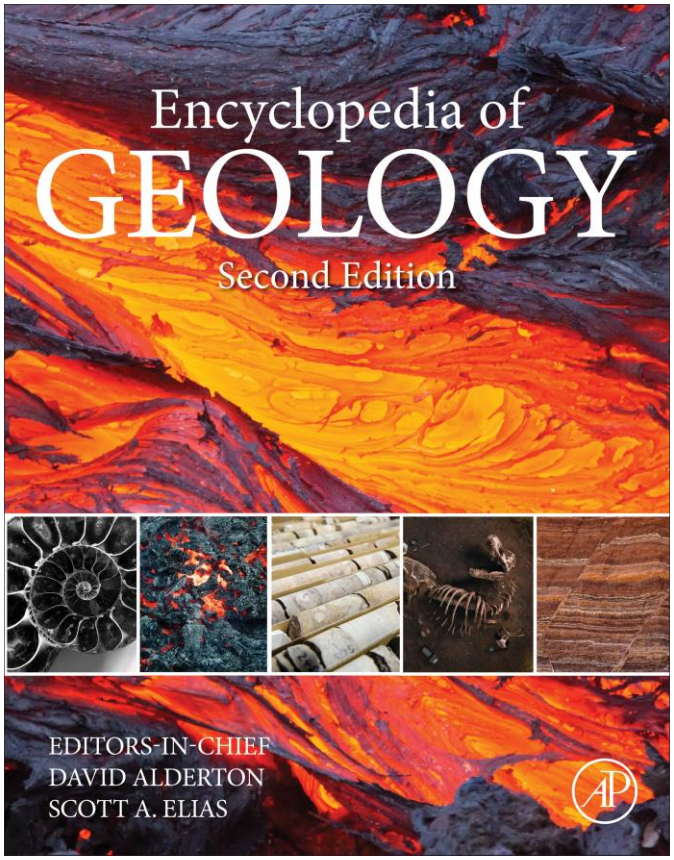

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Soil, Soil Processes, and Paleosols

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Glossary

Alfisol A fertile forested soil with subsurface enrichment of clay.
Andisol A volcanic-ash soil.
Andisolization A soil-forming process that creates low-density non-crystalline fertile soil from volcanic ash.
Anthropic Epipedon A soil surface modified by human use.
Anthrosolization A soil-forming process involving modification by human activities.
Argillan Clay skin, a kind of planar feature in a soil or cutan formed of clay.
Argillic Horizon A subsurface horizon of soil enriched in clay.
Argilluviation A soil-forming process that involves creating clay and washing it into a subsurface clayey horizon.
Aridisol A soil of arid regions, usually containing carbonate nodules.
Biocycling The recycling of nutrient elements by biota.
Birnessite A non-crystalline mixture of iron and manganese oxides.
Burial Decomposition An early diagenetic modification of a paleosol in which buried organic matter is decayed microbially.
Burial Gleization An early diagenetic modification of a paleosol in which buried organic matter fuels microbial chemical reduction of iron oxides and oxyhydrates to ferrous clays, siderite or pyrite.
Climofunction A mathematical relationship between a soil feature and a measure of climate.
Climosequence A set of soils formed under similar vegetation, topographic setting, parent material and time, but varied climate.
Concretion A segregation of materials in a soil, harder or more cemented than the matrix, with prominent internal concentric banding, for example iron-manganese concretion.

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Cutan A planar feature within a soil formed by enrichment, bleaching, coating or other alteration, for example a clay skin (argillan).
Entisol A very weakly developed soil.
Ferran Ferruginized surface, a kind of planar feature in a soil (cutan) formed by chemical oxidation.
Ferrallitization A soil-forming process involving intense weathering that removes most elements other than iron and aluminium.
Gelisol A soil of permafrost regions, usually containing ground ice
Geosol A mappable land surface of paleosols, a soil stratigraphic unit in the North American Code of Stratigraphic Nomenclature.
Gibbsite An aluminium hydroxide mineral ($Al(OH)_3$).
Gilgai A soil microrelief consisting of ridges or mounds alternating with furrows or pits.
Gleization A soil-forming process involving chemical reduction of the soil due to waterlogging.
Halite A salt mineral (NaCl).
Hematite An iron oxide mineral (Fe_2O_3).
Imogolite A colloidal weathering product of volcanic-ash soils.
Inceptisol A weakly developed soil.
Lessivage A soil-forming process that creates clay and washes it into a subsurface clayey horizon.
Lixiviation A soil-forming process that involves leaching nutrient cations from the soil.
Melanization A soil-forming process that involves darkening the soil with organic matter.
Mollic Epipedon A humic fertile crumb-textured soil surface typical of grassland soils.
Mollisol A grassland soil with a humic fertile crumb-textured surface.
Mukkara A soil structure consisting of festooned and slickensided cracks between uplifted parts of the soil; the subsurface structures below gilgai microrelief.
Natric Horizon A subsurface horizon of soil enriched in sodium-clay.
Nodule A segregation of materials in a soil, harder or more cemented than the matrix, with massive internal fabric, for example caliche nodule.
Oxisol A deeply weathered soil of tropical humid regions.
Paludization A soil-forming process involving peat accumulation in waterlogged soils.
Paleosol A soil of a landscape of the past: a past surficial region of a planet or similar body altered in place by biological, chemical or physical processes, or a combination of these
Ped A natural aggregate of soil; stable lumps or clods of soil between roots, burrows, cracks and other planes of weakness.
Pedoderm A mappable land surface of paleosols, a soil stratigraphic unit in the Australian Code of Stratigraphic Nomenclature.
Pedolith Soil sediment: a sedimentary rock dominated by clasts with the internal microfabrics of soils.
Pedotype A kind of paleosol; an ancient equivalent of soil series of the United States Soil Conservation Service.
Perched Water Table Level of water ponded in a soil by an impermeable subsurface layer.
Placic Horizon Fe- and Mn-stained bands and nodules in soils.
Plaggen Epipedon A plowed surface horizon of soils.
Podzol A sandy soil with a bleached near-surface horizon.
Podzolization A soil-forming process in which acid leaching creates a bleached sandy upper horizon and an iron- or organic-rich subsurface horizon.
Sepic Plasmic Fabric Birefringence microfabric appearance of the fine-grained part of a soil or paleosol in petrographic thin sections viewed under crossed nicols of wisps or streaks of highly oriented and highly birefringent clay in a less organized dark matrix.
Siderite An iron carbonate mineral ($FeCO_3$).
Solonization A soil-forming process that creates soda-rich clays and domed columnar peds in arid regions.
Spodosol A sandy clay-poor soil with an organic- or Fe-rich subsurface horizon.

Soils

Introduction

Soil is defined by the Soil Science Society of America as the natural unconsolidated mineral or organic material on the immediate surface of the Earth produced by biological, chemical, or physical weathering. Soil has been subjected to, and shows effects of, genetic and environmental factors of: climate (including water and temperature effects), and macro- and microorganisms, conditioned by relief, acting on parent material over a period of time. A product-soil differs from the material from which it is derived in many physical, chemical, biological, and morphological properties and characteristics.

There are many soil-forming processes, which in varying combinations create the large array of soils forming at the surface of the Earth. The study of soils is facilitated by the observation that soil-forming processes are slow and seldom go to completion. The parent materials of soils are modified over thousands of years by physical, chemical, and biological influences. However, few of these processes can be observed directly. Podzolization is one of the few soil- forming processes sufficiently rapid to be recreated in the laboratory. Soil-forming processes that operate over thousands of years are studied using a space-for- time strategy (i.e. studying soils of differing ages that are subject to the same soil-forming regime). A set of soils of different ages with comparable climates, vegetation, topographical positions, and parent materials is called a chronosequence (Fig. 1). Mathematical relationships between the development of particular soil features and time are called chronofunctions (Jenny, 1941), and include the increased clayeyness produced by the soil-forming process of lessivage, i.e. leaching (Fig. 2). While specifying the rate and progress of soil formation, chronofunctions can also be used to infer the ages of landscapes from undated soils by comparison with dated soils. Such estimates of soil age can be important in the study of the neotectonic deformation of landscapes and their suitability for long-term infrastructure such as dams and nuclear power plants. Soil fertility also varies with soil age, and chronofunctions can guide agricultural use and rehabilitation of soils (Richter and Markewitz, 2001).

Soil-forming processes vary not only with time but also with parent materials, topographical relief, vegetation, and climate. For example, the fragments of volcanic glass in certain kinds of air-fall tuff are distinct from the minerals of most soils, and they bestow high fertility and low bulk density on some volcanic soils (the process of andisolization). Water- logging in low-lying parts of the landscape prevents the rusting of iron minerals and imparts a gray-green color to the soil (the process of gleization). Leachates from highly acidic vegetation, such as pine forest, create soils in which clays are destroyed but quartz and hematite accumulate (the process of podzolization). Finally, climate is also an important factor in soil-forming processes, encouraging deeper and more thorough weathering in wetter and warmer climates (Fig. 3).

The study of soil-forming processes has informed both soil taxonomy (Table 1) and soil-profile terminology (Table 2). The following outlines of soil-forming processes are presented in the order in which they would be encountered from warm wetlands to cold arid lands.

Soil Forming Processes

Gleization

Gleying, from the Russian word *gley*, meaning "mucky mass", is a process that produces conditions of saturation, flooding, or ponding long enough during the growing season to develop anaerobic conditions in the upper part of a soil column and, consequently, maintain unoxidized minerals in soils. The term is derived from a Russian name for the gray clay of swamps and bogs. In combination with microbial activity in the soil that causes a depletion of oxygen, such anaerobiosis promotes the biochemical accumulation of organic matter and the reduction of iron and similar elements. Waterlogged peat-covered stagnant groundwaters allow the preservation of ferrous iron in clay minerals, such as gray smectite, carbonates, such as the siderite of freshwater bogs, and sulfides, such as the pyrite of mangrove swamps and salt marshes. In normally drained soils these minerals oxidize to produce red and brown clays, hydroxides such as goethite, and oxides such as hematite (Table 3). Goethite and hematite also form within gleyed soils when a short-term depression of the water table allows the atmospheric penetration of oxygen. Despite these red nodules and concretions, the dominant color of gleyed soils is bluish or greenish gray (Fig. 4).

Paludization

Paludization is accumulation of undecayed plant debris as peat in the waterlogged surface layer (O horizon of Table 2) of Histosols (Table 1) resulting from soil flooding that is tolerated by swamp trees but not by most soil decomposers. This process requires a balance between plant production and decomposition. If ponding is intermittent and the soil is moderately oxidized, usually because of a low subsidence rate, then fungal and other decay prevents the accumulation of plant debris. If, by contrast, ponding is too deep or prolonged, because of high subsidence rates, then soil stagnation kills the roots of woody plants, thus cutting off the supply of vegetation for further peat accumulation. As swamp forests die from anoxia at the roots, peaty soils become overwhelmed by lakes, bayous, or lagoons. The rate of subsidence and accumulation of woody peats is generally between 0.5 mm and 1 mm/yr because of constraints on the growth rate of woody plants in low-fertility peaty substrates and the depth of penetration of air and decomposers within woody peats. Herbaceous plants and mosses are less constrained in their growth rates and form domed peats that rise well above the water table. Peat accumulation in both cases involves addition from the top, in the same way as sediment accumulation, and thus differs from soil-forming processes that modify pre-existing materials. The progress of paludization leads to progressively thicker peat (Fig. 1).

Podzolization

Podzol in its original Russian means 'under ash' and refers to the light-colored quartz-rich (E) horizon immediately beneath the humus. The soil type Spodosol (Table 1), includes the red, brown, or black (Bs) horizon below the light colored near-surface layer. This striking differentiation between white near-surface and dark subsurface horizons is created by podzolization, which effectively leaches iron and organic matter from the upper horizons and reprecipitates them in a lower horizon (Lundstrom et al., 2000). The resulting effect is as striking as the chromatographic separation of organic compounds, and podzolization is one of the few soil-forming processes that is sufficiently rapid to have been recreated under controlled laboratory conditions. The process is particularly helped by highly acidic soil solutions, having a pH <4 in well- drained soils of humid climates under acid-generating litter such as

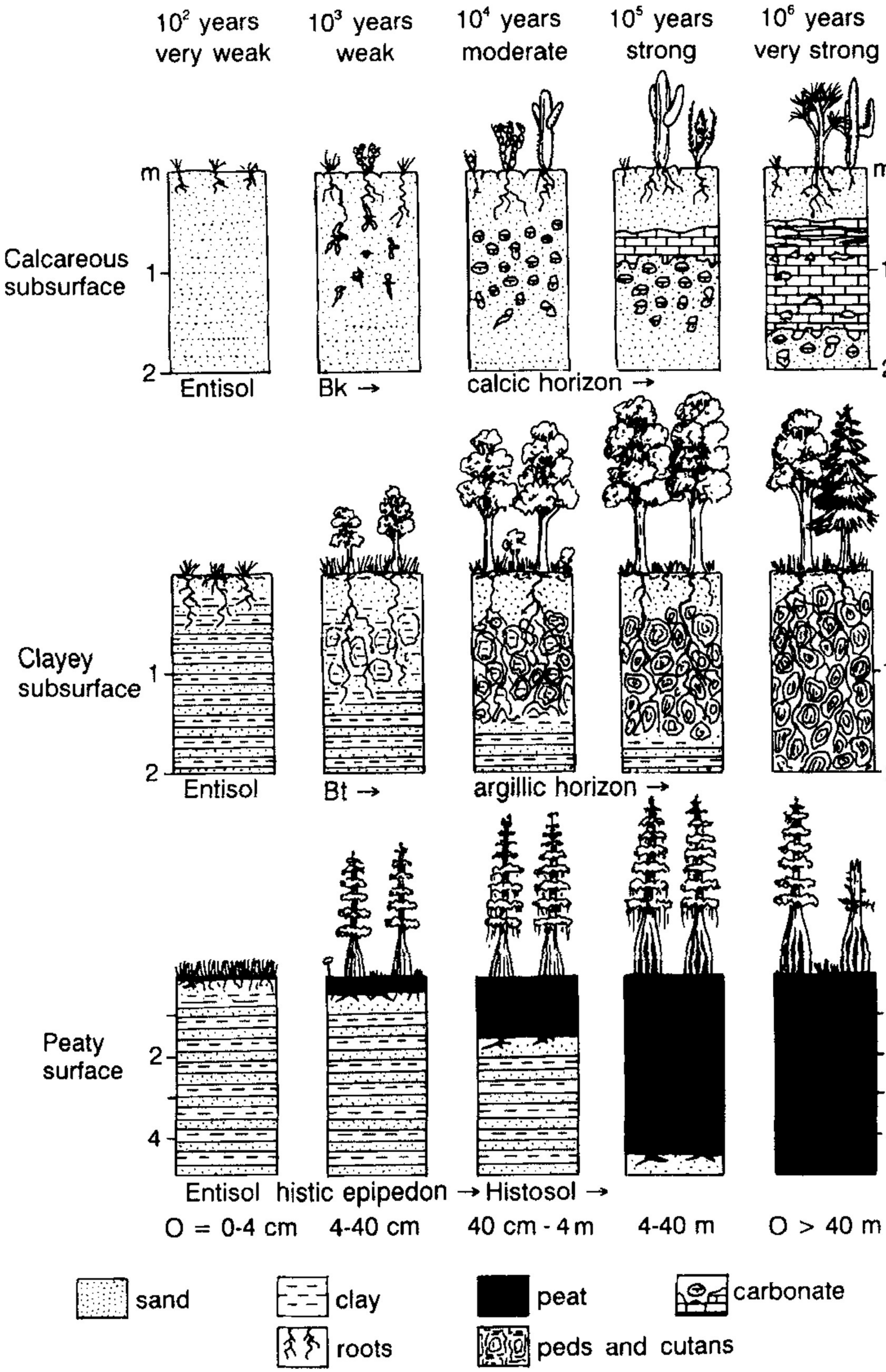

Fig. 1 Soil development stages involving progressive calcification (top), leaching (middle), and paludization (bottom). Reproduced from Retallack GJ (2001) *Soils of the Past*. Oxford: Blackwell with permission.

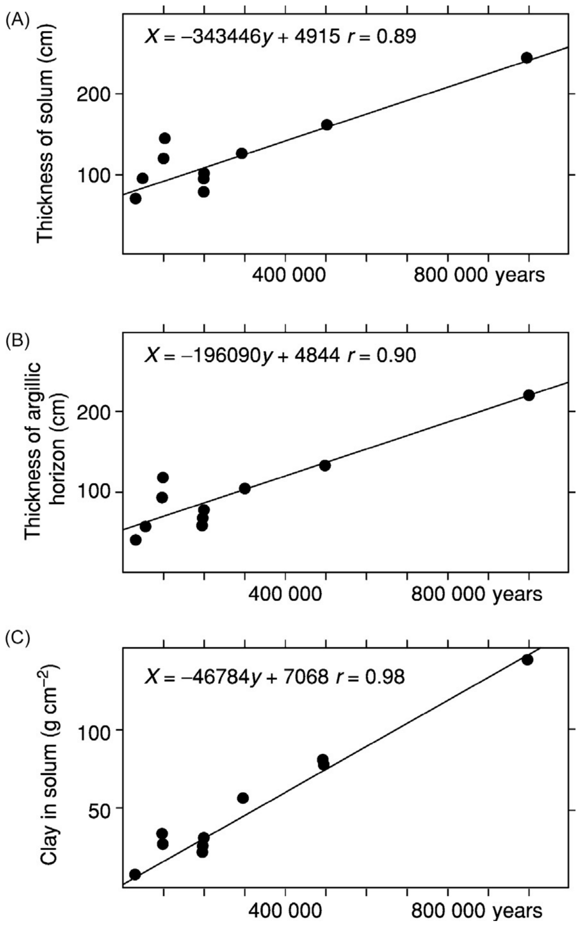

Fig. 2 Chronofunctions for the progress of lessivage in soils of the Coastal Plain and Piedmont of southeastern USA over time: (A) solum thickness; (B) thickness of the argillic horizon; and (C) the amount of clay in the solum. The solum is the A and B horizons; the argillic horizon is the Bt horizon; and the total profile is the A, B, and C horizons as defined in Table 2. Reproduced from Retallack GJ (2001) *Soils of the Past*. Oxford: Blackwell with permission using data from Markewich HW, Pavich MJ, and Buell GR (1990) Contrasting soils and landscapes of the Piedmont and Coastal Plain, eastern United States. *Geomorphology* 3: 417–447.

that of conifer forest (Fig. 3). Under highly acidic conditions clay minerals are destroyed, so Podzols and Spodosols usually have a sandy texture.

Ferrallitization

The term ferrallitization is derived from iron (Fe) and aluminium (Al), which become enriched in minerals such as hematite, kaolinite, and gibbsite during intense weathering of well-drained tropical soils such as Oxisols (Fig. 3). Much of the loss of major cations (Ca^{2+}, Mg^{2+}, Na^{+}, K^{+}) by hydrolysis requires carbonic acid derived from the carbon dioxide of soil respiration, yet the soil pH remains above a pH of 4, so that clays are not destroyed. Mitigation of acidity and deep oxidation of these soils may in part be due to the activity of termites and tropical trees, as ferrallitization is primarily found in soils under tropical rainforest. The broad-leaved trees of tropical rainforests produce less acidic litter than conifers and other plants, and litter decomposition rates are high on humid and warm forest floors. Furthermore, ferrallitic soils commonly contain abundant microscopic (125–750 mm) spherical to ovoid pellets of oxidized clay, like the fecal and oral pellets of termites. Some ferrallitic soils appear to have passed through the guts of termites many times. Termites are unique in having extremely alkaline midguts (with a pH of 11–12.5).

Biocycling

Biocycling includes a variety of processes in which nutrient elements are exchanged by soil biota without reincorporation into soil minerals. In tropical soils such as Oxisols (Table 1) this is a very efficient process in which the decay of leaves and wood is orchestrated by waves of bacteria, fungi, ants, and termites, which excrete and die to feed a copious network of epiphytes and tree roots (Eisenbeis and Wichard, 1987). Effective biocycling explains the spectacular luxuriance of tropical rainforest ecosystems despite their extremely nutrient-depleted and humus-poor mineral soils (Oxisols) (Sanford, 1987). Comparable mechanisms operate in swamp forests growing in peat (Histosols), which also experience severe mineral-nutrient limitations. These mineral

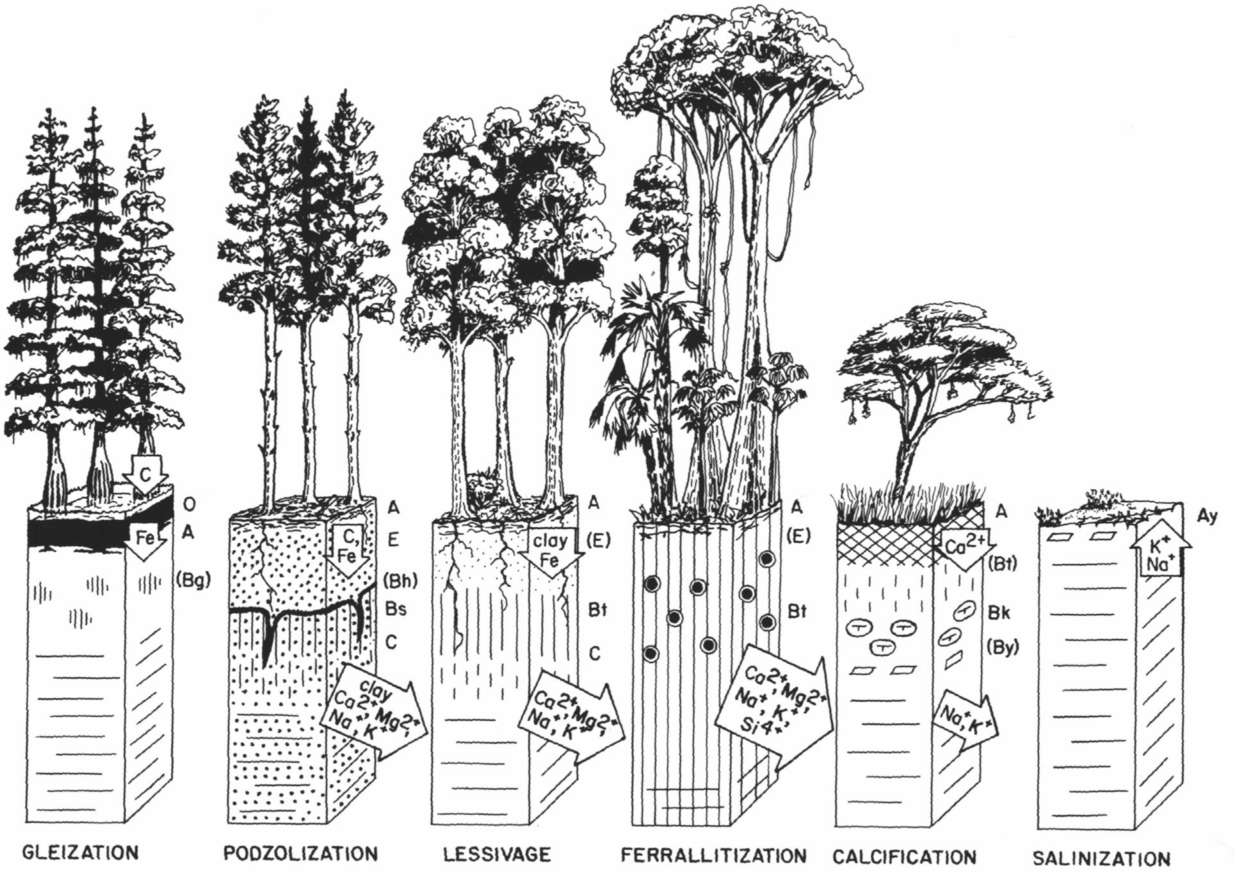

Fig. 3 Selected common soil-forming processes arranged along a climatic gradient. The ecosystems depicted are (from left to right): bald cypress swamp, spruce forest, oak forest, tropical rain forest, *Acacia* savannah, and saltbush scrub. Horizon nomenclature is described in Table 1, and the large arrows indicate the movement of key soil components. Reproduced from Retallack GJ (2001) *Soils of the Past*. Oxford: Blackwell with permission.

Table 1 Outline of soil taxonomy (Soil Survey Staff, 2000).

Order	*Description*
Entisol	Very weakly developed soil with surface rooting and litter (A horizon) over weathered (C horizon) sediment with relict bedding or weathered igneous or metamorphic rock with relict crystals
Inceptisol	Weakly developed soil with surface rooting and litter (A horizon) over somewhat weathered (Bw horizon) clayey (Bt horizon) or calcareous (Bk horizon) subsurface
Andisol	Soil composed of volcanic ash with low bulk density and high fertility
Histosol	Peat (O horizon) over rooted gray clay (A horizon)
Spodosol	Quartz-rich clay-poor soil with bleached subsurface (E horizon) above a red-black iron–alumina–organic cemented zone (Bs horizon)
Vertisol	Very clayey profile with common swelling clay (smectite), laterally variable thickness of surface (A horizon), and strongly slickensided subsurface (Bt horizon)
Mollisol	Grassland soil with thick crumb-textured carbon-rich surface (A horizon)
Gelisol	Permafrost soil with frost heave and other periglacial features
Aridisol	Desert soil with a shallow subsurface accumulation of pedogenic carbonate (Bk horizon) and soluble salts (By horizon)
Alfisol	Fertile forest soil with clay-enriched subsurface (Bt horizon) and high amounts of Mg, Ca, Na, and K
Ultisol	Infertile forest soil with clay-enriched subsurface (Bt horizon) and low amounts of Mg, Ca, Na, and K
Oxisol	Deeply weathered tropical soil, often highly ferruginous and aluminous, but with very low amounts of Mg, Ca, Na, and K

For technical limits of soil orders see Soil Survey Staff (2000).

Table 2 Standard designations for soil horizon description.

Designation	*Description*
O	Surface accumulation of peaty organic matter
A	Surface horizon of mixed organic and mineral material
E	Subsurface horizon rich in weather-resistant minerals, e g. quartz
Bt	Subsurface horizon enriched in washed-in clay
Bs	Subsurface horizon enriched in organic matter, or iron or aluminium oxides
Bk	Subsurface horizon enriched in pedogenic carbonate
Bn	Subsurface horizon with domed columnar structure and sodium-clays
By	Subsurface horizon enriched in salts such as gypsum and halite
Bo	Subsurface horizon deeply depleted of Ca, Mg, Na, and K
Bw	Subsurface horizon mildly oxidized and little weathered
C	Mildly weathered transitional horizon between soil and substrate
R	Unweathered bedrock

For technical limits of soil orders see Soil Survey Staff (2000).

Table 3 Common types chemical reactions that occur during weathering (Bohn et al., 1985).

Reaction	*Example*
Hydrolysis	$2NaAlSi_3O_8 + 2CO_2 + 11H_2O \rightarrow Al_2Si_2O_5(OH)_4 + 2Na^+ + 2HCO_3^- + 4H_4SiO_4$ albite + carbon dioxide + water → kaolinite + sodium ion + bicarbonate ion + dissolved silica
Oxidation	$2Fe^{3+} + 4HCO_3^- + ½O_2 \rightarrow Fe_2O_3 + 4CO_2 + 2H_2O$ ferrous ion + bicarbonate ion + oxygen → hematite + carbon dioxide + water
Dehydration	$2FeOOH \rightarrow Fe_2O_3 + H_2O$ goethite → hematite + water
Dissolution	$CaCO_3 + CO_2 + H_2O \rightarrow Ca^{2+} + 2HCO_3^-$ calcite + carbon dioxide + water → calcium ion + bicarbonate ion

nutrients include the major cations (Ca^{2+}, Mg^{2+}, Na^+, K^+), but these are seldom as limiting as nitrogen, which is derived largely from the microbial recombination of atmospheric nitrogen, or phosphorous, which is derived largely from the weathering of apatite. Biocycling of nitrogen is especially important during the early development of soils such as Entisols and Inceptisols, which are developed over decades or centuries. Biocycling of phosphorus becomes increasingly important in very old soils such as oxisols and ultisols, which are depleted in apatite over thousands or millions of years.

Lessivage

Lessivage or argilluviation is the process of clay accumulation within a subsurface (Bt or argillic) soil horizon (Figs. 1–3). This is a common and widespread soil-forming process in the forested soils of humid climates, particularly Alfisols and Ultisols (Fig. 5). The clay is primarily derived from a hydrolytic weathering reaction in which clays remain as a residuum and dissolved cations are removed in groundwater during the incongruent dissolution of feldspars and other minerals by carbonic acid (Table 3).

Driving the reaction are abundant rainfall and high soil respiration rates fuelled by high primary productivity. Clay forms rinds around mineral grains of the sedimentary, igneous, or metamorphic parent material, but is also washed down cracks in the soil created by desiccation, roots, and burrows. This washed-in or illuvial clay has a very distinctive banded appearance, which is obvious in petrographic thin sections. The clay is not washed any lower than the water table, where percolating rainwater ponds. Clay is less common near the surface of the soil, where unweathered grains are added by wind and water, and grains are leached of clay by plant acids. The net effect is a subsurface clayey horizon that becomes more clayey over time (Figs. 1–3).

Lixiviation

Lixiviation is a process of leaching of major cations (Ca^{2+}, Mg^{2+}, Na^+, K^+) from soil minerals and their loss from the soil in groundwater. Lixiviation is a component of ferrallitization, podzolization, and lessivage, and represents the progress of the hydrolysis chemical reaction, in which hydrogen ions (H^+) of a weak acid (usually carbonic acid – H_2CO_3) displace cations into solution and thus convert primary minerals such as feldspars into soil minerals such as clays (Table 3). The term lixiviation is primarily used to describe the beginnings of this process in soils such as Entisols and Inceptisols that have developed over only decades or centuries. Such young soils have not yet acquired the distinctive deeply weathered and oxidized horizons produced by ferrallitization in Oxisols, the distinctive leached (E) and enriched (Bs) horizons produced by podzolization in Spodosols, or the distinctive clay-enriched subsurface (Bt) horizons produced by lessivage in Alfisols and Ultisols.

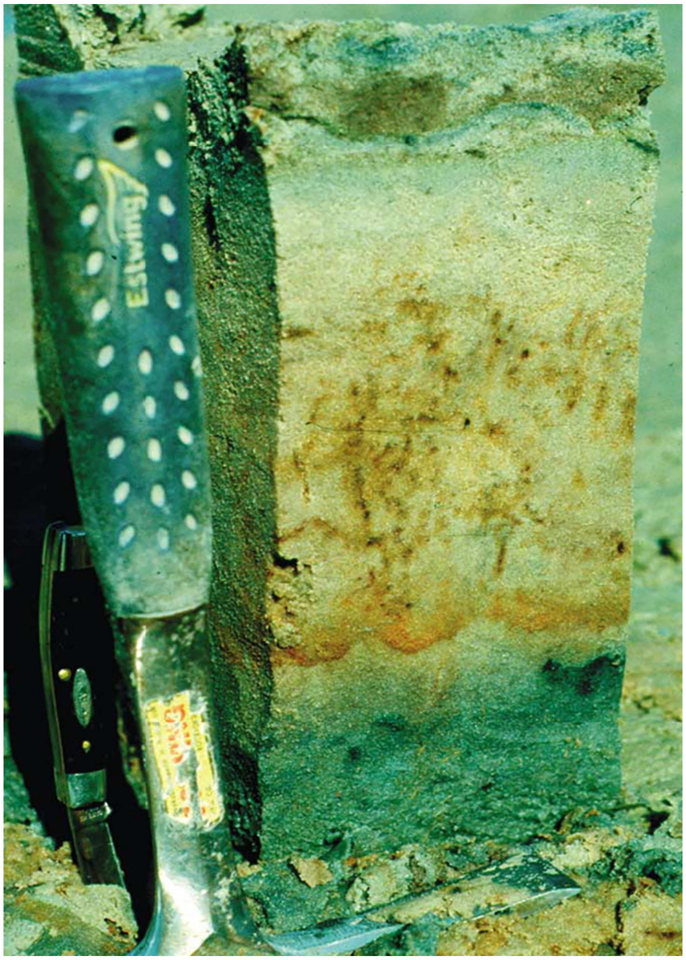

Fig. 4 Red and brown mottles of goethite in the upper part of the profile and dark stains of pyrite formed by gleization in the lower part of the profile of a gleyed Inceptisol, excavated as a soil column from a salt marsh on Sapelo Island, Georgia, USA. Hammer handle is 25 cm long.

Melanization

Melanization is a process of soil darkening due to the addition of soil organic matter. The process is best known in Mollisols, the fertile dark crumb-textured soils of grasslands (Fig. 6). In these soils melanization is largely a product of the activities of grasses and earthworms. Earthworms produce fecal pellets rich in organic matter and nutrients such as carbonate. Earthworms also produce slime, which facilitates their passage through the soil. Root exudates from grasses are also added to soil crumbs. Many soils have dark humic near-surface horizons, but a peculiarity of grassland soils is that dark organic fertile crumb- textured soil extends to the base of the rooting zone, which can be more than a meter deep in soils under tall-grass prairie. Melanization also occurs in swamp and marsh soils (gleyed Inceptisols and Entisols), where the decay of humus is suppressed by poor oxidation and waterlogging. Unlike the alkaline crumb- textured melanized surface of grassland soils, the melanized surface of wetland soils is nutrient-poor, acidic, and has a massive to laminated fabric. Melanization is not usually applied to the precipitation of amorphous Fe–Mn oxides (birnessite) in gleyed soils, which can also produce dark soil. The creation of these Fe–Mn-stained (placic) horizons is a process of gleization rather than melanization.

Andisolization

Andisolization is the formation of fertile mineralogically amorphous low-density horizons within soils of volcanic ash (Andisols). Many volcanic ashes are composed largely of small angular fragments (shards) of volcanic glass. Unlike soil minerals such as feldspar, volcanic glass weathers, not to crystalline minerals such as clay, but to non-crystalline compounds such as imogolite. The loosely packed angular shards and colloidal weathering products create a soil of unusually low bulk density (1.0–1.5 g/cm^3, compared with 2.5–3.0 g/cm^3 for most common minerals and rocks). Furthermore, these colloidal compounds contain plant-nutrient cations, and particularly phosphorous, in a form that is more readily available to plants than those of other kinds of soils dominated by crystalline minerals such as apatite. Andisolization is not sustainable for more than a few thousand years unless there are renewed inputs of volcanic glass, because glass and other colloids (such as imogolite) weather eventually to oxides and clay minerals.

Fig. 5 Light-brown near-surface (E) and dark-brown subsurface (Bt) horizons of an Alfisol produced by lessivage near Killini, Greece. Hammer handle upper right is 25 cm long.

Fig. 6 Dark organic-rich surface (mollic epipedon) of a Mollisol formed by melanization near Joliet, IL, USA. The shovel handle is 15 cm wide at the top.

Vertization

Vertization is the physical overturning and mixing of soil by means of the shrink–swell behavior of clays. It occurs mostly in Vertisols but also in Entisols, Inceptisols, Mollisols, and Alfisols. It is especially characteristic of soils rich in swelling clays (smectites), which swell when wet and shrink when dry. Also characteristic is a climate with a pronounced seasonal contrast in precipitation. During the wet season the clays swell and buckle under the pressure of their inflation. During the dry season they open up in a system of cracks, which are then partly filled by wall collapse. This fill exacerbates the buckling in the next wet season so that the soil develops ridges or mounds with intervening furrows or pits, called gilgai microtopography. In a soil pit, the cracks of mounded areas divide areas of festooned slickensides under the furrows and pits in a distinctive arrangement called mukkara structure (Fig. 7). Vertization is mainly a phenomenon of semiarid to sub-humid regions (Paton et al., 1995). Soils of arid regions are generally not sufficiently clayey, whereas soils of humid regions are generally too deeply weathered to contain abundant smectite and are also stabilized by massive plant and animal communities.

Anthrosolization

Anthrosolization is the alteration of soil by human use, such as buildings, roads, cesspits, garbage dumps, terracing, and plowing. Archeological ruins and artifacts are important clues to prior occupation of a site, but many sites also contain impressive amounts of mollusk shells and mammal and fish bones. A distinctive soil structure of subsoil pockets of laminated clay between large soil clods is produced by moldboard plows. The primitive or ard plow also tends to disrupt the natural crumb structure to a fixed depth (plow line). Phosphorus content is an indicator of human use. Many soils have trace amounts of phosphorus (10–20 ppm by weight) but occupation floors and long-used garden soils and middens have large amounts of phosphorus (1000–2000 ppm). Anthrosolization is locally common worldwide in cities and fields, both ancient and new, but is scattered and local in deserts, polar regions, and high mountains.

Calcification

Calcification is the accumulation of calcium and magnesium carbonates in the subsurface (Bk) horizons of soils (Figs. 1 and 3). The carbonate is usually obvious, appearing as soft white filaments, hard white nodules, and thick white benches within the soil. Calcification is largely a soil-forming process of dry climatic regions, where evaporation exceeds precipitation (McFadden et al., 1991). It is characteristic of Aridisols but is also found in some Mollisols, Andisols, Vertisols, Inceptisols, and Alfisols. The source of the carbonate is the soil respiration of roots, soil animals, and micro-organisms. Calcification requires soil respiration at levels greater than those in hyperarid soils, where halite and gypsum formed by salinization prevail, and less than those in humid soils, where lessivage prevails. The source of the cations of calcium and magnesium, which create the soil minerals calcite and dolomite respectively, is the weathering of soil minerals by hydrolysis (Table 3). Some of these cations originate from feldspars and other minerals of the parent material, but dry regions of calcification have open vegetation and are often dusty, so that carbonate and feldspar dust is an important source of cations. Dissolved cations from hydrolytic weathering are commonly lost downstream in the ground- water in humid regions, but in arid lands the water table is commonly much deeper than the soil profiles, which are seldom

Fig. 7 Deep clay-lined cracks (plan view to left and cross sections to right) produced by vertization in a Vertisol, near Savner, Maharastra, India. Geology hammer for scale is 25 cm long.

wet much beyond the depth of rooting. The subsurface zone of groundwater evaporation and absorption is where the wisps of soil carbonate form, then coalesce into nodules and, eventually, thick layers.

Solonization

Solonization is a process which produces clays rich in sodium within the soils of dry climates (Aridisols), where the hydrolytic mobilization of major cations (Ca^{2+}, Mg^{2+}, Na^{+}, K^{+}) is weak. In humid climates, hydrolysis removes cations from soils by lixiation, but in dry climates the acidity created by soil respiration after rain storms is sufficient to remove cations from minerals such as feldspar without leaching them from the profile. Solonized soils commonly contain carbonate nodules of dolomite or low-magnesium calcite, formed by calcification, as well as salts of halite and gypsum, formed by salinization. Solonized soils have illitic clays rich in potassium and smectite clays rich in sodium, and the progress of solonization can be assessed by the pH (which is usually around 9–10), by chemical analysis, or by X-ray diffraction to determine the mineral composition. A field indicator of solonization is the presence of domed columnar peds which run through most of the subsurface (natric or Bn) of the soil (Fig. 8). The sodium-smectite clays of solonized soils have some shrink–swell capacity, meaning that they form prismatic cracks as the soil dries and swelling or domed tops to the prisms when the soil is wet. Solonization is common around desert playa lakes and salinas and in coastal soils affected by saltwater spray.

Solodization

Solodization is intermediate between solonization and lessivage, and creates profiles with acidic-to-neutral near-surface horizons but alkaline subsurface horizons dominated by sodium-smectite. Solodized soils have domed columnar clayey peds in a subsurface (Bn) horizon, but these are sharply truncated by a granular leached (E) horizon. Solodization occurs in desert soils (Aridisols) with better vegetative cover and a more humid climate than solonized soils.

Salinization

Salinization is the precipitation of salts in soils (Fig. 3) and is found mostly in desert soils (Aridisols). The most common salts are halite and gypsum, which can form either as clear crystals within soil cracks or as sand crystals that engulf the pre-existing soil matrix. Salts are easily dissolved by rain and so accumulate in regions where there is a marked excess of evaporation over precipitation, which is generally less than 300 mm per year. There is a strong relationship between mean annual precipitation and the depth of leaching of salts in soils. Salinized soils are sparsely vegetated or lack vegetation, and occur in playa lakes, sabkhas, and salinas. Although these are commonly regarded as depositional environments, they are soil environments as well.

Cryoturbation

Cryoturbation is the mixing of soils by the freezing and thawing of ground ice. The ice can form disseminated crystals, hair-like threads, thin bands, thick benches, or vertical cracks depending on the local climatic conditions (Washburn, 1980). Soil mixing

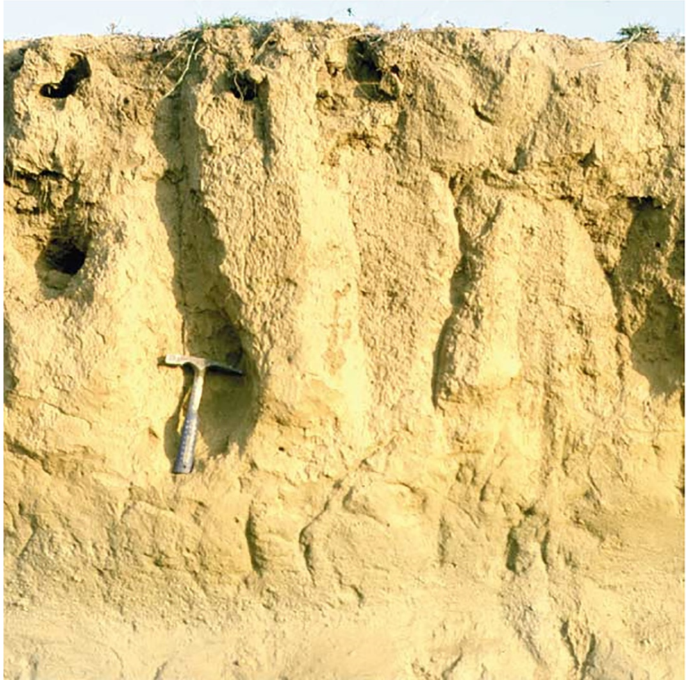

Fig. 8 Domed columnar peds produced by solonization in an inceptisol near Narok, Kenya. Hammer handle is 25 cm.

results from the expansion of water to ice during winter freezing and the relaxation of the deformation on summer melting. Ice-wedge polygons, for example, are wide polygonal cracks that are filled with ice in winter but can be filled with layered sediments in water during the summer in climates where the mean annual temperature is less than −4 °C. Sand-wedge polygons form in colder climates where the mean annual temperature is less than −12 °C; here, summer melting of ice is limited and sediment fills cracks between the ice and soil in a series of near vertical layers.

Conclusion

Soil-forming processes are varied and complex, and the changing understanding of them guides the classification, description, and management of soils (Bockheim and Gennadiyev, 2000). The processes are also of interest in simplifying the vast array of chemical reactions, biological processes, and physical effects that create soil. Some processes are more common and widespread than others. Lixiviation and its underlying hydrolysis chemical reaction is perhaps the most important weathering process on Earth, affecting geomorphology, sedimentation, ocean chemistry, and climate. Other processes are restricted to more specific climatic, biotic, geomorphological, geological, and temporal environments, but are no less important in their local environments.

Paleosols

Introduction

Paleosols are ancient soils, formed on landscapes of the past. Most paleosols have been buried in the sedimentary record, covered by flood debris, landslides, volcanic ash, or lava (Fig. 9). Some paleosols, however, are still at the land surface but are no longer forming in the same way that they did under different climates and vegetation in the past. Climate and vegetation change on a variety of time-scales, and the term relict paleosol for profiles still at the surface should be used only for such distinct soil materials as laterites among non-lateritic suites of soils (Fig. 10). Therefore, some surface soils are paleosols, although most paleosols are buried soils. Pedoderm and geosol are names for whole landscapes of paleosols, and are soil stratigraphical units, named and mapped in order to establish stratigraphic levels.

The terms pedotype and soil facies are more or less equivalent and are used to refer to individual paleosol types preserved within ancient buried landscapes. These terms are used to distinguish one type of paleosol from another in environmental interpretations of paleosols. Pedolith, or soil sediment, describes a sediment, as indicated by bedding and other sedimentary features, with distinctive soil clasts, such as ferruginous concretions. Pedoliths are uncommon in sedimentary sequences, because soils are readily eroded to their constituent mineral grains, which retain few distinctive soil microfabrics.

Recognition of Paleosols

Paleosols buried in sedimentary and volcaniclastic sequences can be difficult to distinguish from enclosing sediments, tuffs, or lavas and were not widely recognized until recently. Three features of paleosols in particular aid their identification: root traces, soil horizons, and soil structure. Soil is often defined as the medium of plant growth. Geological and engineering definitions of soil are broader, but fossilized roots and their traces are universally accepted as indicators of paleosols. Not all paleosol root traces are permineralized or compressed original matter: some are tortuous and made of clay, with discolored or mineralized alteration (Fig. 11). Both fossilized roots and root traces show the downward tapering and branching of roots. Soils also contain fossil

Fig. 9 The subtle color banding in these cliffs is the result of a sequence of 87 Eocene and Oligocene paleosols in a sequence of non-marine silty claystones exposed in the Pinnacles area of Badlands National Park, South Dakota, USA.

Fig. 10 The red rock exposures to the left on the beach are a lateritic paleosol of Middle Miocene age. Even though these horizons are at the surface, they are considered to be paleosols because soil horizons of this type are not currently forming in this area. The red rock in the background is a sequence of Early Triassic paleosols in Bald Hill Claystone, near Long Reef, New South Wales, Australia.

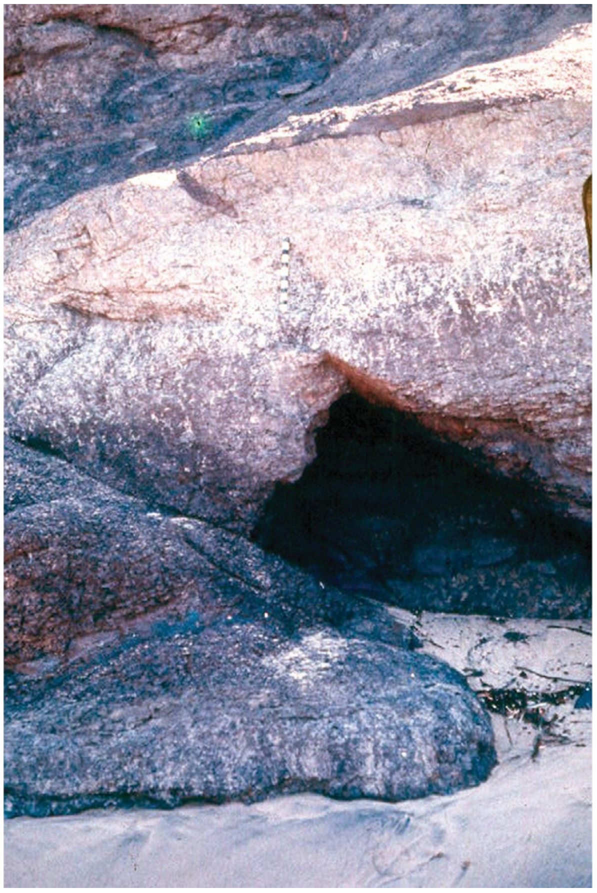

Fig. 11 The sharply truncated top and abundant drab-haloed root traces (A horizon) diminishing downwards into red claystone (Bt horizon) are soil horizons of a paleosol (Long Reef clay paleosol, Early Triassic, Bald Hill Claystone, near Long Reef, New South Wales, Australia).

burrows, but these are sparsely branched parallel-sided traces. The distinction between burrows and roots can be blurred in cases where soil animals feed on roots and where roots find an easier passage through the soft fill of burrows. For very old rocks, predating the Early Devonian evolution of roots, the criterion of root traces is of no use in identifying paleosols.

Paleosols also have recognizable soil horizons, which differ from most kinds of sedimentary bedding in their diffuse contacts downwards from the sharp upper truncation of the paleosol at the former land surface. Paleosol horizons, like soil horizons, are

Fig. 12 Two successive paleosols overlain sharply by volcanic grits show crumb-structured organic surfaces (A horizon) over calcareous-nodule- studded subsurfaces (Bk horizon). In the upper right corner is a comparable modern soil (Middle Miocene fossil quarry near Fort Ternan, Kenya).

seldom more than a meter thick and tend to follow one of a few set patterns. Subsurface layers enriched in clay are called Bt horizons in the shorthand of soil science (Fig. 11). Unlike a soil, in which clayeyness can be gauged by resistance to the shovel or plasticity between the fingers, clayeyness in paleosols that have been turned to rock by burial compaction must be evaluated by petrographic, X-ray, or geochemical techniques. Subsurface layers enriched in pedogenic micrite are called Bk horizons in the shorthand of soil science and are generally composed of hard calcareous nodules or benches in both soils and lithified paleosols (Fig. 12).

A final distinctive feature of paleosols is soil structure, which varies in its degree of expression and replaces sedimentary structures such as bedding planes and ripple marks, metamorphic structures such as schistosity and porphyroblasts, and igneous features such as crystal alignment and columnar jointing. Because they lack such familiar geological structures, they are commonly described as featureless, massive, hackly, or jointed. Paleosols, like soils, have distinctive systems of cracks and clods. The technical term for a natural soil clod is a ped, which can be crumb, granular, blocky, or columnar, among other shapes. Peds are bounded by open cracks in a soil and by surfaces that are modified by plastering over with clay, by rusting, or by other alterations.

These irregular altered surfaces are called cutans, and they are vital in recognizing soil peds in paleosols that have been lithified so that the original cracks are crushed. The rounded 3–4 mm ellipsoidal crumb peds of grassland soils and paleosols (Fig. 12) are quite distinct from the angular blocky peds of forest paleosols (Fig. 11). Common cutans in soils and paleosols include rusty alteration rinds (ferrans) and laminated coatings of washed-in clay (argillans). Cutans and other features of lithified paleosols are best studied in petrographic thin sections and by scanning electron microscopy, and electron microprobe analysis. Some petrographic fabrics, such as the sepic plasmic fabric or streaky birefringence of soil clays observed when viewed with crossed nicols under a petrographic microscope, are diagnostic of soils and paleosols.

Alteration of Soils After Burial

Paleosols are seldom exactly like soils because of alteration after burial or exposure to additional weathering, and this can compromise their interpretation and identification with modern soils. Paleosols, like sediments, can be altered by a wide array of burial processes: cementation with carbonate, hematite, or silica; compaction due to pressure or overburden; thermal maturation of organic matter; and metamorphic recrystallization and partial melting. These high-pressure and high-temperature alterations of paleosols are not as difficult to disentangle from processes of original soil formation as are three common early modifications: burial decomposition, burial reddening, and burial gleization.

Some soils are buried rapidly by chemically reducing swamps or thick lava flows, preserving most of their organic matter. In contrast, many paleosols are covered thinly by floodborne silt or colluvium, and their buried organic matter is then decomposed by aerobic bacteria and fungi deep within the newly-forming soil of the paleosol sequence. For this reason many paleosols have much less organic carbon (fractions of a weight percent) than comparable modern soils (usually 5–10% by weight of carbon at the surface). Thus paleosol A horizons are seldom as dark as soil surface horizons, and must be inferred from the abundance of roots rather than from color and carbon content.

Soils vary considerably in their degree of redness, but most paleosols are red to reddish brown from hematite (iron oxide) or occasionally yellowish brown from goethite (iron hydroxide). Soils become redder from the poles to the tropics, from moderately drained to well-drained sites, and with increasing time for development, as iron hydroxides are dehydrated to oxides. The dehydration of iron hydroxides continues with the burial of soils, so that red paleosols are not necessarily tropical, unusually well drained, or especially well developed.

In river-valley and coastal sedimentary sequences with abundant paleosols, formerly well-drained soils can find themselves subsiding below the water table with root traces and humus largely intact. Burial gleization is a process in which organic matter is

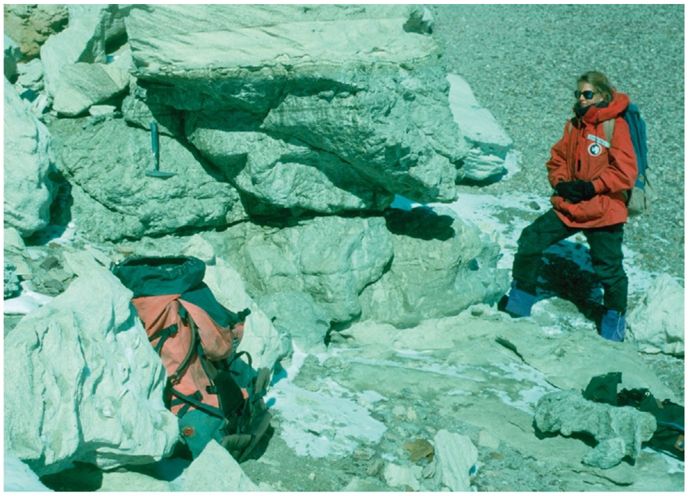

Fig. 13 An unusually warm paleoclimate is indicated by this paleosol, which is unusually thick, clayey, and deeply weathered for its paleolatitude of 70°S and is comparable to soils now forming no further south than 48°S (Early Triassic Feather Conglomerate, Allan Hills, Victoria Land, Antarctica).

used by microbes as a fuel for the chemical reduction of yellow and red iron oxides and hydroxides. Comparable processes of biologically induced chemical reduction are common in swamp soils, but superimposition of this process on the organic parts of formerly well-drained soils produces striking effects in some paleosols. The whole A horizon is turned gray, with gray haloes extending outwards from individual roots, which diminish in abundance down the profile (Fig. 3). Burial gleization is especially suspected when the lower parts of the profile are highly oxidized and have deeply penetrating roots, as in well-drained soils, and when there is no pronounced clayey layer that would 'perch' a water table within the soil.

The combined effect of burial decomposition, dehydration, and gleization can completely change the appearance of a soil. The gaudy gray-green Triassic paleosol shown in Fig. 13, for example, was probably modified by all three processes from an originally dark brown over reddish-brown forest soil.

Paleosols and Paleoclimate

Many paleosols and soils bear clear marks of the climatic regime in which they formed. The Berkeley soil scientist Hans Jenny quantified the role of climate in soil formation by proposing a space-for-climate strategy (Jenny, 1941). What was needed was a carefully selected group of soils, or climosequence, that varied in climate of formation but were comparable in vegetation, parent material, topographical setting and time for formation. He noted that mean annual rainfall and the depth in the profile to calcareous nodules decline from St. Louis west to Colorado Springs, in the mid-western USA, but that temperatures and seasonality at these locations are comparable. Also common to all these soils is grassy vegetation on postglacial loess that is about 14,000–12,000 years old. From these soils he derived a climofunction or mathematical relationship between climate and soil features. A 2005 compilation of comparable data showed a clear relationship between the depth from the surface of the soil of carbonate nodules (D in cm) and the mean annual precipitation (P in mm) according to the formula:

$$P = 137.24 + 6.45D + 0.013D^2.$$

Such climofunctions can be used to interpret paleoclimate from the depth within paleosols of calcareous nodules (Fig. 12), once allowance is made for reduction in depth due to burial compaction.

Climatic inferences also can be made from ice deformation features, concretions, clay mineral compositions, bioturbation, and chemical analyses of paleosols. The thick clayey paleosol shown in Fig. 14 is riddled with large root traces of the kind found under forests and is very severely depleted in elemental plant nutrients such as Ca, Mg, Na, and K. Comparable modern soils are found at mid-latitudes, yet this paleosol formed during the Triassic at a paleolatitude of about 70° S. This paleoclimatic anomaly indicates pronounced global warming, in this case a postapocalyptic greenhouse effect following the largest mass extinction in the history of life at the Permian-Triassic boundary.

Paleosols and Ancient Ecosystems

Just as soils bear the imprint of the vegetation and other organisms they support, so many aspects of ancient ecosystems can be interpreted from paleosols. The paleosols shown in Fig. 12, for example, have a dark crumb-textured surface horizon with abundant fine (1–2 mm) roots, comparable to the modern grassland soil seen forming on the outcrop to the upper left. Forest soils, in contrast (Fig. 11), have large woody root traces, a blocky structure, and thick subsurface clayey horizons (Bt).

Fig. 14 Swamp forests of tree lycopsids (*Stigmaria ficoides*) grew in waterlogged soils, in which lack of oxygen forced the roots to form planar mats rather than reaching deeply into the soil (Carboniferous Lower Limestone Coal Group, Victoria Park, Glasgow, Scotland).

In some cases root traces in paleosols are identifiable, although the species *Stigmaria ficoides* (Fig. 14) is a form genus for roots of a variety of extinct tree lycopsids and not a precisely identified ancient plant. The tabular form of the roots of *Stigmaria* indicates a poorly drained soil, because roots do not photosynthesize, but rather respire using oxygen from soil air. Tabular, rather than deeply reaching, root traces (Fig. 11) are characteristic of swamp paleosols.

Some paleosols also contain fossil leaves, fruits, wood, stones, bones, and teeth. These are direct evidence of soil ecosystems. Unlike fossils in deposits of lakes and shallow seas, fossil assemblages in paleosols have the advantage of being near the place where the organisms lived. However, the preservation of fossils in paleosols is seldom as ideal as complete skeletons in river-channel deposits or compressed leaves in carbonaceous shales. The carbon and carbonate contents of paleosols can be used to evaluate the Eh and pH, respectively, of the paleosol preservational environments of the fossils.

Paleosols and Paleogeography

Just as soils vary from mountain summits to coastal swamps, so do paleosols give clues to their ancient topographical setting. Many paleosols within sedimentary sequences show clear relationships with deposits of paleochannels and levees, so that their depositional subenvironment can be inferred from context. Water tables are close to the ground surface in many sedimentary environments, and paleosols yield important information on their position relative to ancient water tables. Paleosols formed below the water table include peats and are gray with chemically reduced minerals such as pyrite and siderite. Burrows of crayfish and other aquatic organisms are locally common in waterlogged soils, but burrows of most rodents and beetles are not. Root traces also do not penetrate deeply into waterlogged soils or paleosols (Fig. 14). Deeply penetrating roots and burrows and red oxidized minerals of Fe or Al are common in formerly well-drained paleosols (Fig. 11). Paleosols may also reveal upland sedimentary environments such as alluvial and colluvial fans, glacial moraines, river terraces, and erosional gullies (Fig. 15).

Major geological unconformities often mark erosional landscapes of the past. Rocky cliffs and bedrock platforms are found along geological unconformities, but so are upland paleosols. For example, the hilly erosional landscape of Lewisian Gneiss in northern Scotland had 1 km of relief (Fig. 16).

Paleosols and Their Parent Materials

The parent material of a soil or paleosol is the substance from which it formed and can usually be inferred from the less-weathered lower parts of the profile. The parent material may be precisely known if the paleosol is on metamorphic or igneous rocks (Fig. 16), because pedogenic minerals are easily distinguished from igneous and metamorphic minerals. Parent material is more difficult to find in paleosols that are developed from sedimentary parent materials, especially if sedimentary facies reveal erosional relief (Fig. 15). In such settings, the sediment is derived from pre-existing soils, whose degree of weathering can be quite varied. The kinds of soils formed on sediment and rock also can be very different. If soil were a commercial product, economy would dictate manufacturing it from materials that are already similar in chemical composition and physical characteristics. Soils form more readily from sediments than from rocks. Perhaps the most distinctive of parent materials is volcanic ash, because it may consist of more volcanic glass than minerals. Volcanic glass weathers to noncrystalline amorphous substances such as imogolite, which confer high fertility from loosely bound phosphorus, potassium, and other plant nutrients. Such soils also have low bulk density and good moisture-retaining properties. Such soils around tropical volcanoes support intensive agriculture, despite the hazards of the nearby active volcano, because they are so much more fertile than surrounding soils. Comparable paleosols are commonly associated with volcanic arcs of the past (Fig. 9).

Fig. 15 A paleogully in a strongly developed sequence of paleosols (dark colored) is filled with alluvium including weakly developed paleosols (Late Triassic Chinle Formation, Petrified Forest National Park, Arizona, USA). The hill in the foreground is 11 m high. Photograph courtesy of Mary Kraus.

Fig. 16 The bleached pink paleosol formed on gneiss to the right (Sheigra paleosol) is thicker and more deeply weathered than the light green paleosol formed on amphibolite to the left (Staca paleosol). Both paleosols are overlain by red quartz sandstones of the Torridonian Group (Late Precambrian, near Sheigra, Scotland).

Paleosols and Their Times for Formation

Soils develop their profiles over time, although some soils, such as peats, also accumulate layer-by-layer in the manner of sediments. Each paleosol within a sedimentary or volcanic sequence represents a short break in sedimentary accumulation, or diastem, whose duration can be calculated from key features of the soil. The peats that become coal seams in the geological record, for example, cannot accumulate at rates of more than 1 mm/year because the roots will be suffocated by stagnant water. Nor can they accumulate at rates of less than 0.5 mm/year because aerobic decay will destroy the organic debris as fast as it accumulates. Thus, the durations of coal-bearing paleosols can be calculated from coal thickness, once compaction is taken into account. Calcareous soils and paleosols accumulate carbonate at first in wisps and filaments, and later in nodules, which become larger and larger (Fig. 12). The size of the nodules thus gives us an idea of the time over which they formed. The development of clayey subsurface horizons is comparable (Fig. 11) in that clay becomes more and more abundant over time (Markewich et al., 1990). The amount of washed-in clay can thus be a guide to the time over which paleosols formed.

From the times for paleosol formation and the thickness of rock for successive paleosols it is possible to calculate rates of sediment accumulation. In the badlands of South Dakota, for example, the clayey lower part of the section accumulated at a slower rate than the ashy and silty upper part of the section (Fig. 9). Variations in the rate of sediment accumulation can be used to address a variety of tectonic, volcanic, and sequence stratigraphic problems using paleosols.

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