

6

Classification of Soils

6.1 Introduction: General Characteristics of Soil Orders & Global Distributions

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The purpose of Chapter 6 is to present a discussion of the distribution, attributes, pedogenesis, classification, use and management of the 12 Soil Orders recognized by *Soil Taxonomy* (Soil Survey Staff, 1998). In this chapter, soil orders have been arrayed in a sequence to consider the organic soils first, followed next by soils which are weakly expressed or lack subsoil diagnostic horizons, and closing with soils exhibiting the greatest evidence of pedogenic weathering and/or accumulation of sesquioxides and translocated phyllosilicates. Only general statements can be made at this categorical level, but the information content becomes more precise as one descends to lower taxa (Sections 6.2–6.13). Because *Soil Taxonomy* is a dynamic classification system, several of the orders are under revision or have been revised recently; hence, subject matter included herein is subject to change. This is particularly true of the Histosols, Mollisols, Alfisols, Inceptisols, and Gelisols.

Because of space limitations, data sets verifying the physical, chemical, mineralogical and biological attributes of given orders are abridged. However, access to complete soil characterization databases, engineering properties, pedon descriptions, soil interpretations, and other information for different soil taxa are available online at the following address: <http://www.statlab.iastate.edu/soils/soildiv/>. For example, through the Soil Survey Laboratory, National Soil Survey Center, Lincoln, Nebraska, analytical data for more than 20,000 pedons in the United States and 1,100 pedons from other countries have been archived. Standard morphological pedon descriptions are available for about 15,000 of these pedons. Further, these databases include listings of published soil surveys in the United States, Puerto Rico, Virgin Islands, Trust Territories and some foreign countries, areal extent and classification placements of soil series, ongoing soil survey investigations, including soil-related studies on climatic change, and field and laboratory methodologies.

Fig. 6.1 illustrates the global distribution of soil orders at 1:1 million scale and Table 6.1 provides the areal extent of suborders associated with these orders. Tables 6.2 and 6.3 present proportional percentages of soil orders occurring in different soil temperature (Table 6.2) and soil moisture (Table 6.3) regimes as defined in *Soil Taxonomy* (Soil Survey Staff, 1998). Broad geographical distribution patterns (Fig. 6.1) reflect regional conditions of climate, vegetation, geology, and topography that function interactively over time. Soils found at any point on the land surface are considered products of multiple sets of pedogenic processes with state factors serving as controls (Section E, Chapter 4). Generally, soils are developed along polygenetic pathways, on dynamically evolving landforms under the influence of paleoclimates, in nonuniform parent materials. Hence, rarely can bio-, litho-, climo-,

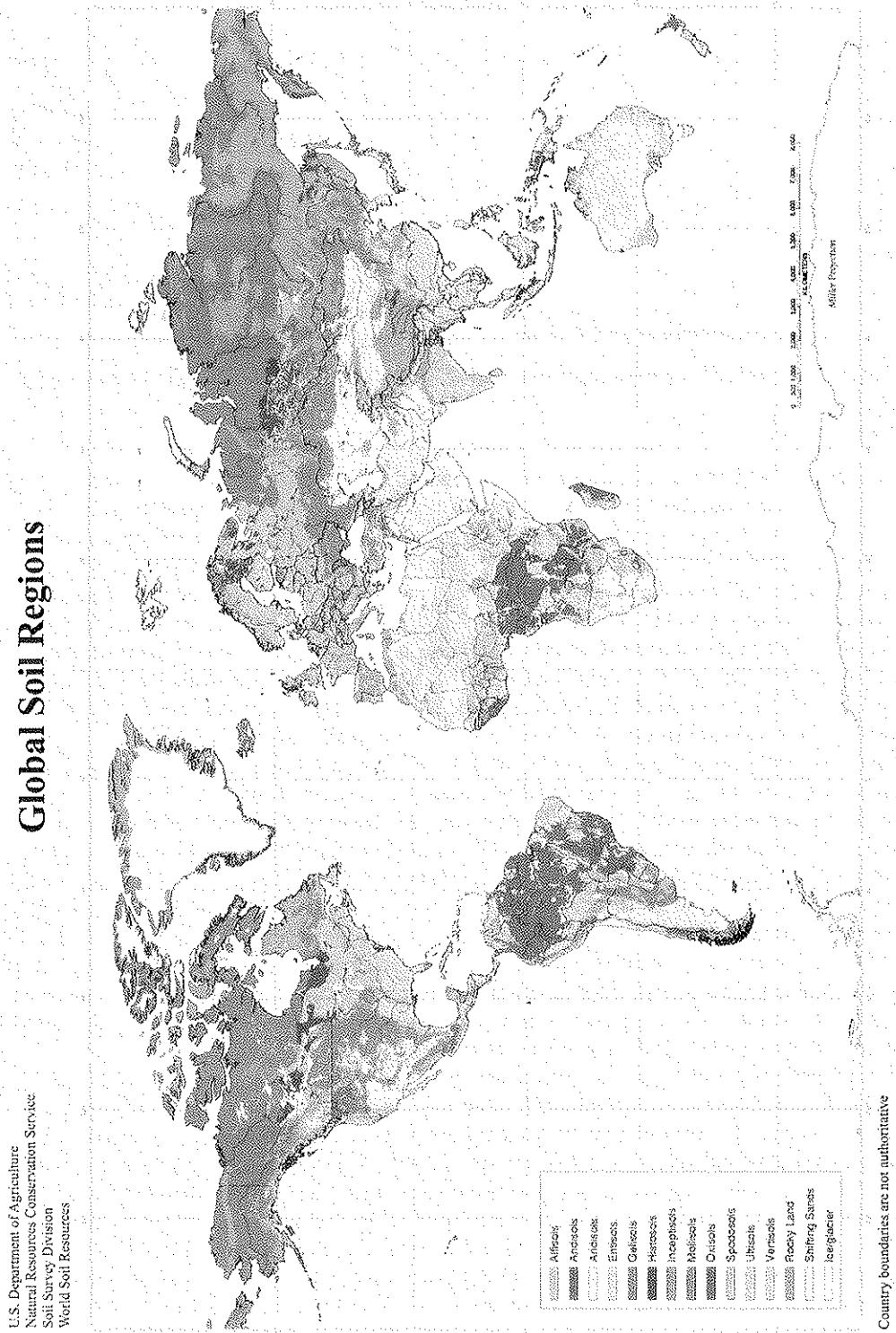


Fig. 6.1 Illustration of global distribution of soil orders at 1:1 million scale. See color insert following page E-204.
[Courtesy of the USDA-NRCS, Soil Survey Division, World Soil Resources, 1998]

Table 6.1 Global areas and percentages of suborders and miscellaneous land units based on ice-free land area [Courtesy of the USDA-NRCS, Soil Survey Division, World Soil Resources, 1998]

Order	Suborder	Area km ² (x 10 ³)	Proportion %
Alfisols	Aqualfs	836	0.7
	Cryalfs	2,518	1.9
	Ustalfs	5,664	4.3
	Xeralfs	897	0.7
	Udalfs	2,706	0.2
	Subtotal	12,621	9.6
Andisols	Cryands	255	0.2
	Torrands	2	« 0.1
	Xerands	32	« 0.1
	Vitrands	281	0.2
	Ustands	63	0.1
	Udands	279	0.2
	Subtotal	912	0.7
Aridisols	Cryids	943	0.7
	Salids	890	0.7
	Gypsid	683	0.5
	Argids	5,408	4.1
	Calcids	4,873	3.7
	Cambids	2,931	2.3
	Subtotal	15,728	12.0
Entisols	Aquepts	116	0.1
	Psamments	4,428	3.4
	Fluvents	2,860	2.2
	Orthents	13,733	10.5
	Subtotal	21,137	16.2
Gelisols	Histels	1,013	0.8
	Turbels	6,333	4.8
	Haplels	3,914	3.0
	Subtotal	11,260	8.6
Histosols	Folists	« 1	« 0.1
	Fibrists	197	0.1
	Hemists	988	0.8
	Saprists	341	0.3
	Subtotal	1,527	1.2
Inceptisols	Aquepts	3,199	2.5
	Anthrepts	< 1	« 0.1
	Cryepts	457	0.3
	Ustepts	4,241	3.2
	Xerepts	685	0.5
	Undepts	4,247	3.3
	Subtotal	12,830	9.8
Mollisols	Albolls	28	« 0.1
	Aquolls	118	0.1
	Rendolls	266	0.2
	Xerolls	924	0.7
	Cryolls	1,164	0.9
	Ustolls	5,245	4.0
	Udolls	1,261	1.0
	Subtotal	9,006	6.9

Table 6.1 (cont.)

Order	Suborder	Area km ² (x 10 ³)	Proportion %
Oxisols	Aquox	320	0.2
	Torroxo	31	« 0.1
	Ustox	3,096	2.4
	Perox	1,162	0.9
	Udox	5,201	4.0
	Subtotal	9,810	7.5
Spodosols	Aquods	169	0.1
	Cryods	2,460	1.9
	Humods	58	< 0.1
	Orthods	667	0.5
	Subtotal	3,354	2.5
Ultisols	Aquults	1,281	1.0
	Humults	344	0.3
	Udults	5,540	4.2
	Ustults	3,870	3.0
	Xerults	19	« 0.1
	Subtotal	11,054	8.5
Vertisols	Aquerts	5	« 0.1
	Cryerts	15	« 0.1
	Xererts	99	0.1
	Torrerts	889	0.7
	Usterts	1,768	1.3
	Uderts	384	0.3
	Subtotal	3,160	2.4
Miscellaneous	Shifting	13,076	10.0
	Other	5,322	4.0
	Subtotal	18,398	14.0
Total ice-free land area		130,797	100.0

chrono- and toposequence functions be developed with scientific rigor. Clearly, the distribution of soil orders in which local conditions strongly govern pedogenic dynamics over short-range distances is not well illustrated in (Fig. 6.1). In particular, such comments are germane to distributions of Histosols, Spodosols, Vertisols and Gelisols. In these cases, the distributions presented in subsections of this chapter may differ somewhat from those presented herein (Fig. 6.1). This uncertainty will also impact the accuracy of global distributions and areal data given for Histosols, Andisols, and Inceptisols found interspersed with Gelisols.

The human impact on soils is recognized in some taxa of Entisols and Inceptisols, but, in general, *Soil Taxonomy* does not adequately handle the anthropogenic effects on soils. For example, Mollisols can be transformed to Alfisols upon severe erosion; paddy soils can undergo secondary salinity and waterlogging under long-term rice culture; soils in urban/industrial environments can be markedly modified by landfills, landfarming, earth movement and heavy metal contamination; and drastically disturbed soils are common in regions where precious metals, rock aggregate and fossil fuels have been mined. The International Committee on Anthropic Soils (ICOMANTH) where ANTH stands for anthropogenic, is currently addressing this matter. For information on this activity refer to the ICOMANTH web-site address: <http://wwwscas.cit.cornell.edu/icomanth>.

Soil Taxonomy currently does not accommodate systematic cataloging of paleosols formed under remarkably different paleoenvironments. With increasing geomorphic age, properties of soils commonly reflect a welding of contemporaneous and paleoenvironments. Especially when paleosols

Table 6.2 Extent of soil orders found in different temperature regions [Modified from data supplied through the courtesy of the USDA-NRCS, Soil Survey Division, World Soil Resources, 1998]

Soil Order	Soil temperature regimes*			
	Tropical	Temperate	Boreal	Tundra
		%		
Alfisols	38	39	23	0
Andisols	49	23	28	0
Aridisols	12	74	14	« 1
Entisols	28	68	4	0
Gelisols	0	0	0	100
Histosols	21	8	71	0
Inceptisols	47	42	11	0
Mollisols	4	50	46	0
Oxisols	98	2	0	0
Spodosols	« 1	18	82	0
Ultisols	69	31	« 1	0
Vertisols	47	52	1	0

*Soil temperature regimes defined by *Soil Taxonomy* (Soil Survey Staff, 1998, <http://www.statlab.iastate.edu/soils/soildiv>)

Tropical: isomesic, isothermic and isohyperthermic (MAST > 8 °C in which the difference in mean summer and mean winter temperatures is < 6 °C at 50 cm soil depth)

Temperate: mesic, thermic, and hyperthermic (MAST is > 8 °C at 50 cm soil depth).

Boreal: frigid, isofrigid, and cryic (MAST is 0-8 °C at 50 cm soil depth).

Tundra: pergelic (MAST < 0 °C at 50 cm soil depth)

are well preserved, they are valuable proxies of biological and physicochemical evolution of the Earth. This is an area for *Soil Taxonomy* development that will likely gain greater attention in the future. With multiple uses of soil surveys for connectivity between the land surface and geological substrata (e.g., for environmental, hydrological, archaeological and biogeoscience synergisms), there are increasing driving forces to observe soils beyond the 2 m depth limit currently in vogue. This will likely enhance the interests in revising *Soil Taxonomy* to better accommodate recognition and cataloging of paleosols as important morphogentic markers of paleoenvironments (Section E, chapter 1).

In perusing the global distributions of soil orders (Fig. 6.1), and aerial extents given in Tables 6.1–6.3, several noteworthy relationships become apparent. These are summarized in a cursory overview in the following paragraphs. The organic-rich kingdom of soils [Histosols (Chapter 6.2)] comprises about 1.2 % of the Earth's ice-free land surface (Table 6.1). They are generally in cool, humid, high latitude, boreal regions (or in microhabitats) that favor the balance of organic matter accumulation over decomposition (Tables 6.2 and 6.3). This is because of saturated and reduced conditions, biological inhibitors, and/or cool, short summer periods with low evapotranspiration rates. Many of the Histosols that could be illustrated on this general map are now considered Gelisols or Andisols with the introduction of these two new soil orders. Histosols are important wildlife and wetland habitats. They also serve extensively for cranberry, citrus, vegetable, rice, and sugarcane production. Major constraints are subsidence, nutrient deficiencies and wind erosion. Effective watertable management is critical to minimize subsidence. These land resources are also critically limiting to construction activities because of their low soil strength and load bearing capacities.

Entisols (Chapter 6.3) comprise about 16.2% of the Earth's ice-free land surface (Fig. 6.1). Over two thirds of the Entisols are in the temperate region and the remainder mostly in the tropics (Table 6.2). Sixty percent of the Entisols have aridic soil moisture regimes with ustic and udic, the next most extensive (Table 6.3). These are mineral soils which do not have expression of diagnostic subsurface horizons within a specified depth of the soil surface, generally 1–2 m. Subsoil materials of these soils

Table 6.3 Extent of soil orders found in different soil moisture regimes [Modified from data supplied through the courtesy of the USDA-NRCS, Soil Survey Division, World Soil Resources, 1998]

Soil Order	Soil moisture regimes*			
	Aridic	Xeric	Ustic	Udic**
			%	
Alfisols	0	8	56	36
Andisols	4	3	30	63
Aridisols	95	1	3	1
Entisols	60	4	22	14
Gelisols	N/A [†]	N/A	N/A	N/A
Histosols	8	1	23	68
Inceptisols	0	6	42	52
Mollisols	0	13	66 [‡]	22
Oxisols	« 1	0	35	65
Spodosols	« 1	1	9	90
Ultisols	0	« 1	42	58
Vertisols	28	3	56	13

* Soil moisture regimes defined by *Soil Taxonomy* (Soil Survey Staff, 1998, <http://www.statlab.iastate.edu/soils/soildiv>)

** Udic soil moisture regime includes perudic (continuously moist) and aquic conditions (saturated, reduced and with redoximorphic features)

[†] N/A - not appropriate for the Gelisol order

[‡] A few percentage of Mollisols placed in Ustic soil moisture regimes may in fact occur under Aridic soil moisture regimes, especially in steppe zones of Eurasia that border on Ustic-Aridic soil moisture regimes.

reflect the nature of the geological materials from which they were derived. These weakly expressed soils are commonly found on geomorphic surfaces which are unstable because of frequent flooding, erosion/truncation, or human impact (drastically disturbed lands). They are formed also in coarse-textured, resistant mineral parent materials (e.g., quartzose sands) that are subject to little pedogenic development over time. Entisols are common along flood plains of river and stream valley systems, sand dunes in desert regions, on high gradient mountainous terrain, and associated with recently mined or disturbed lands.

Andisols (Chapter 6.4) comprise < 0.7% of the Earth's ice-free land surface (Table 6.1), but these soils are so unique physically, chemically and mineralogically that a separate soil order was established for them. They are found along the tectonically active Pacific Ring of Fire, Central Atlantic Ridge, North Atlantic rift, the Caribbean, and the Mediterranean regions where volcanic or pyroclastic deposits are common (Fig. 6.1). About half of them are located in tropical regions and the remainder split between temperate and boreal climates (Table 6.2). About two thirds are in udic soil moisture regimes and one third in ustic regions (Table 6.3). These weakly developed, fertile soils are texturally undifferentiated and characterized by short-range order (amorphous) aluminosilicates that have not been translocated from upper to lower horizons. Andisol land resources are used extensively for crop production in developing countries employing slash/burn traditional agricultural practices.

Inceptisols (Chapter 6.5) comprise about 9.8% of the Earth's ice-free land surface (Fig. 6.1). These soils occur indiscriminately globally because they lack a sharply defined central concept. Over 90% occur in tropical and temperate climates (Table 6.2) under ustic and udic soil moisture regimes (Table 6.3). Inceptisols serve to make Soil Taxonomy fully bifurcated ("junk basket" category). Soils classified as Inceptisols generally contain light colored surface horizons and weakly expressed diagnostic subsoil horizons (cambic), but under certain circumstances, may contain more strongly developed subsoil diagnostic horizons (e.g., petrocalcic, petrogypsic, duripan, and fragipan). Where these soils reflect youthfulness, they occur in high gradient mountainous regions, along major river

systems as terraces and deltaic/fluviial plains, and as soils developed from carbonate rich bedrocks or sediments in Mediterranean environments. They also occur interspersed among more strongly developed Alfisols, Ultisols and Oxisols in tropical and subtropical regions.

Gelisols, the newest of the soil orders (Chapter 6.6), comprise about 8.6% of the Earth's ice free land surface (Fig. 6.1). These are soils of the high latitude polar regions underlain by permafrost (materials wet/dry that remain below 0 °C for 2 consecutive years) at depths of 1–2 m depending on cryoturbation (frost churning) activity (Table 6.2). Cryoturbation in Gelisols is driven by the physical volume change from water to ice and subsequent moisture migration along thermal, hydrostatic, chemical, and electrical gradients. Cryopedogenic processes include freezing/thawing, cryoturbation, frost heaving, cryogenic sorting, thermal cracking, and ice segregation. Remarkable spatial variability occurs in soil properties over distances of a few meters or less giving rise to pattern ground in the form of stone stripes, stone circles, high center polygons, low center polygons and barren frost boils. Construction activities on these land resources require management of thermal balances so the insulation properties of the surface are not disturbed. If the organic rich surface layers are compacted, compressed or removed, substrata ice wedges will melt causing the surface to become a highly irregular thermokarst topography. Because Gelisols are the most recent soil order established, and because the depth and presence (or absence) of permafrost is locally controlled, the precise global distribution of Gelisols as presented in Fig. 6.1 is preliminary and approximate. Likewise, the areal extent of Gelisol suborders presented in Table 6.1 is still under development and refinement.

Vertisols (Chapter 6.7) comprise about 2.4% of the Earth's ice-free land surface (Fig. 6.1). About half the Vertisols are found in tropical environments and about half in temperate (Table 6.2). Over 50% of the Vertisols occur in regions of ustic soil moisture regimes where seasonal desiccation is common. The remainder are found primarily under aridic, udic and xeric environments; here the oscillation from wet to dry soil conditions is less frequent (Table 6.3) and gilgai expression less extreme. These clayey, shrink/swell soils are highly diverse in physical, chemical, biological and mineralogical properties. They are concentrated in subtropical and tropical environments, but occur across broad climatic, vegetative, topographic, and geologic regions of Africa, India, Australia, China and North and South America (Fig. 6.1). Vertisols are active and subject to soil failure when swelling pressures exceed shear strength upon wetting a dry soil. Soil failure is usually expressed as slickensides occurring as diagonal, polished and grooved slip surfaces, wedge-shaped structural units, and microtopography in the form of gilgai. Commonly these soils have self-swallowing and self-mulching surface features where strongly aggregated surface materials infill vertical structural cracks. While inversion by turbation is commonly invoked as a major pedogenic processes, most movement in these soils is believed to be diagonally along inclined major slickenside fault planes. Major limitations for using these highly productive soils for agriculture, especially in developing countries, are their high energy requirements for tillage and narrow favorable soil moisture range for workability. Construction activities are constrained by their propensity for soil failure and high shrink/swell activity.

Mollisols (Chapter 6.8) comprise about 7% of Earth's ice-free land surface (Fig. 6.1). Essentially all the Mollisols are in temperate and boreal regions, approximately equally split (Table 6.2). Approximately two thirds of the Mollisols are in ustic soil moisture regimes (Ustolls); most of the remainder occur under udic regions (Udolls) and the rest are in climates with winter precipitation and summer periods of soil moisture deficit (Xerolls) (Table 6.3). A few Mollisols may occur under aridic soil moisture regimes in the southern steppe zones of Eurasia; however, clear definition of the extent of these soils is still wanting. Mollisols are the dark colored, high base status soils commonly found under prairie grass vegetation in mid latitudes of North and South America and Eurasia. The distribution pattern of these soils in Eurasia is in east-west trending belts between Alfisols to the north

and Aridisols and Inceptisols to the south. This reflects decreasing precipitation and increasing temperature gradients in traversing from north to south. In contrast, in North America the precipitation isohyets and isotherms are set normal to each other with precipitation decreasing from east to west and temperature increasing from north to south. These climatic controls result in Mollisols being bordered on the east by Alfisols and Ultisols, and to the west by Aridisols, Entisols and drier Alfisols. Parent materials also control Mollisol distribution; Mollisols are commonly associated with calcareous or high base status glacial drift deposits, limestones and loess. Mollisols (especially Udolls) are the "bread basket" soils of the Americas and Eurasia because of their high native fertility, favorable climate, and excellent physical and chemical properties for high crop production potentials.

Spodosols (Chapter 6.9) comprise about 4% of Earth's ice-free land area (Fig. 6.1). Over 80% of the Spodosols occur under boreal environments (Table 6.2) with udic soil moisture regimes (Table 6.3). Spodosols occur between Mollisols of mid to high latitudes and Alfisols in lower latitudes. These soils are very photogenic, and commonly, have prominent expression of eluvial E horizons over black to red spodic and placic horizons. The latter are formed from translocated illuvial organic-metal (Al and Fe) complexes. These soils have many properties in common with Andisols. The major difference is that Spodosols have formed for a sufficient period of time for the weathered organic-metal complexes to be translocated to subsoils. In Andisols little or no translocation of these complexes has taken place. Formation of Spodosols is favored by environments with abundant acid litter, coarse-textured, resistant parent materials, and strong leaching potentials. These soils are very strongly acid and extremely phosphorus deficient. Native vegetation reflects acid tolerant plants. They are little used for agricultural production because of the short, cool growing seasons, high management inputs, and low nutrient and water retention. Major land use is forests for lumbering and recreation. Pasture and forages are used for livestock production and dairy animals.

Aridisols (Chapter 6.10) comprise about 12.0% of the Earth's ice-free land area (Fig. 6.1). Aridisols are mostly in hot temperate deserts with aridic soil moisture regimes, dry coastal regions, or on rain-shadow plains leeward of high mountains (Tables 6.2 and 6.3). Locally, they also occur in more humid regions where salts have concentrated at or near the soil surface through evaporative pumping from shallow saline ground waters. Aridisols range from base-rich to base-poor soils and exhibit a wide diversity in physical, chemical and mineralogical properties. Commonly, they represent contemporaneous xerophytic environments, but some Aridisols have well-expressed subsoil diagnostic properties such as argillic, natric, calcic, petrocalcic, etc. horizons. These likely reflect more pluvial paleoclimates; under such conditions, these diagnostic horizons would be, in part, relict. Aridisols are used extensively for rangelands, military bases, and nomadic agrarian activities. These soils are not generally considered arable land resources without irrigation. However, they are valuable land resources where an adequate quantity of high quality subsurface water is available for intensive agriculture and urban development. Irrigation and drainage must be managed for a favorable salt balance in Aridisols.

Alfisols (Chapter 6.11) and Ultisols (Chapter 6.12) are closely related soil orders that comprise, respectively, about 9.6% and 8.5% of the Earth's ice-free land area (Fig. 6.1). These soils, developed mostly under deciduous forested or savannah environments, are extensive in mid to lower latitudes. Alfisols are distributed approximately equally across tropical, temperate and boreal environments whereas the distribution of Ultisols is skewed towards tropical conditions (Table 6.2). For both Alfisols and Ultisols most of their extent is associated with ustic and udic soil moisture regimes (Table 6.3). Alfisols and Ultisols are differentiated from other orders on the basis of textural differentiation resulting from translocation and/or *in situ* neoformation of clays that form argillic and kandic horizons. Alfisols differ from Ultisols in having higher base saturation (> 35%) at a specified depth in the subsoil. Rationale for this differentiation was that Alfisols could sustain traditional slash/burn

subsistence agriculture without significant external fertilizer inputs, while Ultisols would require applied nutrients for crop production. Both orders are used extensively for forest and crop production. Major limitations are nutrient deficiencies, wind and water erosivity, and seasonal soil moisture deficits, especially for those in xeric and ustic regions. Alfisols are used extensively in developing countries of Africa, Central America, South America and India for subsistence traditional slash/burn agriculture.

Sesquioxide-rich Oxisols (Chapter 6.13) comprise about 8% of the Earth's ice-free land surface. They occur in equatorial regions intermixed with Ultisols and Inceptisols (Fig. 6.1). Nearly all the Oxisols occur in tropical environments (Table 6.2) with two-thirds under udic and one-third under ustic soil moisture regimes (Table 6.3). Oxisols have low activity clays and are nutrient poor, weakly buffered systems that have little or no horizon differentiation. This reflects (1) their high weathering intensities in low latitude, humid environments, (2) their geomorphic stability and age associated with mafic-rich, Precambrian platforms, or (3) development from polycycled, preweathered parent materials derived from highly weathered terrestrial source areas. These soils are often considered residual carcasses, but many are very productive for agriculture and forestry if properly managed for water conservation, soil erosion and nutrient inputs. With high nutrient inputs and management to conserve soil and water, Oxisols can be very productive agricultural soil systems.

6.2 Histosols

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6.2.1 Introduction

While most soils of the world are comprised primarily of mineral materials, a small but important group of soils are formed from organic materials derived from plants, or less frequently, from animals. Organic soil materials contain a minimum of 12–18% organic C, depending on the particle size of the mineral component (Soil Survey Staff, 1996). Generally speaking, soils within which at least 40 cm of the upper 80 cm are organic materials, and which do not have permafrost within 1 m of the soil surface, are Histosols. Prior to 1997, organic soils with permafrost were included in the Histosols order, but are now placed in the Histel suborder of Gelisols (Soil Survey Staff, 1998). Therefore, earlier discussions of the worldwide distribution of Histosols include organic soils of permafrost regions and might overestimate the distribution by today's concepts (Everett, 1983). Organic soil materials are commonly referred to as peat, and land covered by Histosols or Histels is known as peatland. The term mire is a synonym of peatland that is more commonly used in Europe. Histosols also include a narrowly distributed group of soils, the Folists, that consist of well-drained organic soil materials that directly overlie bedrock or coarse fragments with little or no intervening fine soil. The peat layer in Folists may be (and usually is) thinner than the 40 cm required for other Histosols.

Because of their high organic C content, many Histosols have been utilized as a combustible energy resource. Mankind has mined and burned peat since prehistoric times, and peat is still an important fuel in a number of northern countries, although it has a lower energy rating than oil or coal. In Russia, Germany, and Ireland, peat is not only utilized for domestic heating, but also on a large scale, in electricity generation. In 1991, peat burning was the source of 14% of Ireland's electricity (McNally, 1997), and as recently as the late 1980s, when fuel costs were higher, a peat-burning electric power plant was built in Deblois, ME, although the subsequent fall in fuel prices precluded the plant from

ever becoming operational. In addition to being mined as an energy source, peat is mined for use as a soil amendment in agriculture and horticulture. In addition, the agricultural value of mineral-rich Histosols has long been recognized. Provided that the watertables can be effectively managed, high yields of vegetables and other specialty crops can be produced on Histosols in such different climatic regions as Michigan and Florida (Lucas, 1982).

Because most Histosols occur in wetlands, their utilization as agricultural and energy resources has come under intense scrutiny, particularly in the United States. Histosols perform many of the beneficial functions of wetlands, and they are negatively impacted by mining, drainage or other practices associated with agriculture; thus, there are benefits derived from the preservation of Histosols in their natural state. Histosols are perhaps uniquely fragile and are highly vulnerable to degradation. When drained or dry, organic soil material is highly susceptible to wind erosion (Lucas, 1982). Organic soils also have very low strength, are highly compressible (MacFarlane, 1969; MacFarlane and Williams, 1974), and gradually disappear by decomposition, if drained. Furthermore, within the framework of current discussions of global climate change and C budgeting, Histosols contribute significantly to the terrestrial C pool. On an areal basis, C storage in Histosols is often greater than mineral soils by an order of magnitude (Rabenhorst, 1995.) A listing of great groups and subgroups in the Histosols order is presented in Table 6.4.

6.2.2 Distribution

Histosols occur at all latitudes, but are most prevalent in the boreal forest regions of northern North America, Europe, and Asia (Fig. 6.2). The world's largest expanses of Histosols occur in the West Siberian lowland (Walter, 1977) and the Hudson Bay lowland of central Canada (Sjors, 1963; Canada Committee on Ecological (Biophysical) Land Classification, 1988). At lower latitudes, Histosols occur locally on humid coastal plains, notably southeast Asia and Indonesia (Anderson, 1983).

Histosols in the United States are most widespread in lowlands of the Great Lakes region, the northeast, the Atlantic Coastal plain and Florida, the Pacific Northwest, and Alaska (Fig. 6.2). The largest expanses of Histosols in the continental United States are on the Lake Agassiz plain in north central Minnesota (Wright et al., 1992). Coastal and estuarine areas inundated by tidal water are also sites for Histosol formation, most notably along the Atlantic and Gulf coastlines. Drained Histosols are

Table 6.4 Listing of suborders, great groups and subgroups in the Histosols Order [Soil Survey Staff, 1998]

Suborder	Great Group	Subgroup
Folists	Cryofolists	Lithic, Typic
	Torrefolists	Lithic, Typic
	Ustifolists	Lithic, Typic
	Udifolists	Lithic, Typic
Fibrists	Cryofibrists	Hydric, Lithic, Terric, Fluvaquentic, Sphaginic, Typic
	Sphagnofibrists	Hydric, Lithic, Limnic, Terric, Fluvaquentic, Hemic, Typic
	Haplofibrists	Hydric, Lithic, Limnic, Terric, Fluvaquentic, Hemic, Typic
Saprists	Sulfosaprists	Typic
	Sulfisaprists	Terric, Typic
	Cryosaprists	Lithic, Terric, Fluvaquentic, Typic
	Haplosaprists	Lithic, Limnic, Halic Terric, Halic, Terric, Fluvaquentic, Hemic, Typic
Hemists	Sulfohemists	Typic
	Sulfihemists	Terric, Typic
	Luvihemists	Typic
	Cryohemists	Hydric, Lithic, Terric, Fluvaquentic, Typic
	Haplohemists	Hydric, Lithic, Limnic, Terric, Fluvaquentic, Fibric, Sapric, Typic

widely used in agriculture in the Great Lakes region, southern Florida, and the Sacramento-San Joaquin delta region of California. In the semi-arid Great Plains and mountainous west, Histosols are very rare and occur only in areas of steady groundwater discharge (Mausbach and Richardson, 1994) or humid areas at high elevations (Cooper and Andrus, 1994). Organic soils are widespread in lowlands throughout Alaska, although most of the organic soils in central and northern Alaska have permafrost, and hence, are classified as Gelisols rather than Histosols.

6.2.3 Formation

6.2.3.1 Parent Material

In contrast to the wide variety of mineral materials which may serve as parent materials for other soils, the parent materials from which Histosols are formed are organic in nature. The conditions which cause the accumulation of organic parent materials are very closely tied to the processes which form various organic soil horizons. The unique properties of Histosols result from the nature of the organic parent material.

Some of the factors which affect the nature of organic parent materials include hydroperiod, water chemistry, and vegetation type and will be discussed in more detail later. The net accumulation of organic materials occurs when rates of additions (usually as primary plant production) exceed those of decomposition. In natural soils of most ecosystems, a steady state exists between these two processes, which maintains the quantity of organic C in the surface horizons somewhere between 0.5% and 10%, although some forest soils have thin layers of organic soil materials (O horizons). In Histosols, the rates of decomposition are slowed and organic matter accumulates to significant depth. In most cases, this is caused by saturation leading to anaerobic conditions which causes soil organic

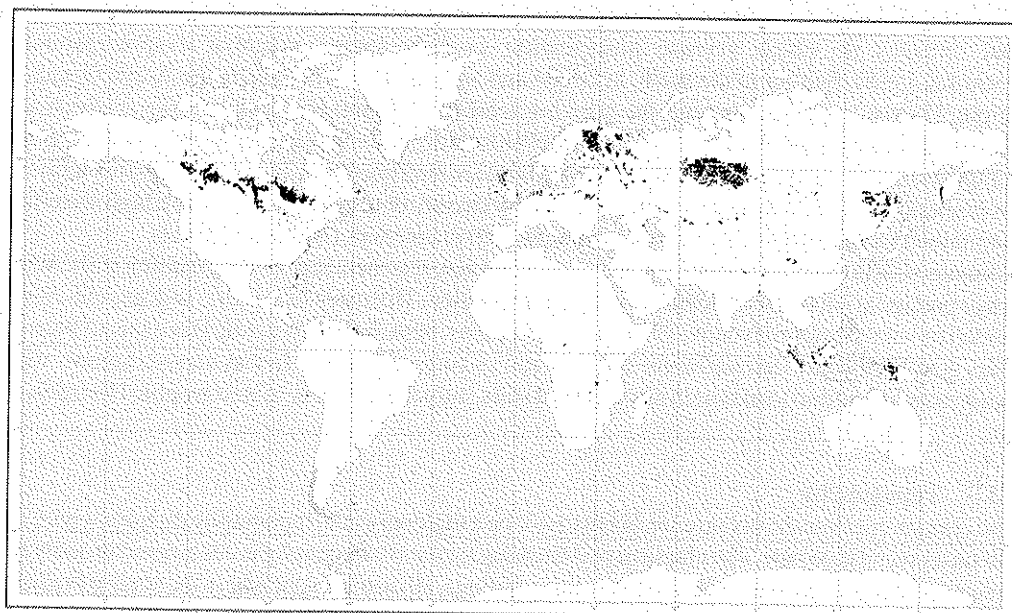


Fig. 6.2 Global distribution of Histosols. The extent of Histosols is probably significantly greater in Alaska than shown on this map due to an overestimation in the extent of permafrost (Gelisols) in Alaska used in developing this map [Courtesy of USDA-NRCS, Soil Survey Division, World Soil Resources, 1998].

matter (SOM) decomposition to be less efficient than under aerobic conditions. Occasionally, usually under cool and moist conditions, organic parent materials may accumulate without prolonged saturation, producing soils in the unique Folist suborder of Histosols. Under certain conditions and landscapes, mineral materials can be added to accumulating organic parent materials, typically by alluviation or eolian deposition. In such cases, the balance between mineral inputs and organic accumulation will determine whether a Histosol or mineral soil will form.

The accumulation of organic parent materials often occurs over long periods of time and under changing conditions. Thus, the stratigraphy of a bog may reflect many thousands of years of organic accumulation. Rarely does vegetation remain constant over such long periods. Microscopic examination of the partially decomposed peat, or evaluation of pollen or plant microfossils can provide information concerning the types of plants which have contributed to the organic parent material during various stages of accumulation.

The organic parent material of Histosols is a major source of acidity. Acids produced by the partial decomposition of organic matter cause the organic horizons of Histosols to be highly acidic unless they have been neutralized by bases that were dissolved from mineral soils or rocks and transported into the peat by groundwater. Highly acidic peatlands are called bogs and commonly described as ombrotrophic (rainfed, because all nutrients are derived from atmospheric sources). The less acid Histosols that receive base-rich groundwater or runoff are called fens (if treeless) or swamps (if forested) and usually described as minerotrophic (fed with mineral-derived nutrients). The term swamp is also applied to forested wetlands on mineral soils. Peatlands are also sometimes divided by their acidity into oligotrophic (very acid and mineral poor), mesotrophic (intermediate), and eutrophic (weakly acidic to neutral and mineral rich) classes (Moore and Bellamy, 1974; Gore, 1983).

6.2.3.2 Climate

Histosol formation is favored by wet or cold climates by hindering decomposition of organic matter. The northern Histosol-dominated regions of North America and Eurasia (Fig. 6.2) have temperate and boreal climates in which average precipitation exceeds annual potential evapotranspiration (Trewartha, 1968). Further to the north, in Greenland and the islands bordering the Arctic Ocean, for example, Histosols are rare because the growing season is so short that there is little production of organic matter. To the south of the boreal zone, Histosol formation is apparently constrained by rapid decomposition of SOM. At lower latitudes, Histosols are restricted mostly to coastal plains with very flat topography, high annual precipitation, and no dry season, or to those areas where a high watertable is maintained by tidal waters and coastal marshes with mangroves (Anderson, 1983).

Climate affects not only where Histosols occur, but also the chemistry of the resulting soils. Where average precipitation exceeds potential evapotranspiration, Histosols in suitable settings can remain saturated by rainfall alone (Ivanov, 1981). In such regions, both the highly acidic Histosols of bogs and the less acid Histosols of groundwater-fed fens and swamps may form. In less humid regions, where groundwater is required to maintain saturation of soils, only the less acid, minerotrophic Histosols can form.

6.2.3.3 Topography

While climate controls the occurrence of Histosols on a regional scale, topography dictates where they occur on a given landscape. The classical soil-forming factor of topography is considered broadly here to incorporate relief, geomorphic, and hydrologic settings. Histosols occur where the setting facilitates concentration of runoff, discharge of groundwater, or retention of precipitation. (Fig. 6.3A,B,C). These conditions are most often satisfied in topographic depressions or very flat areas. On plains with low permeability substrates in the boreal zone, Histosols may cover entire interfluvies (Sjors, 1963;

Walter, 1977), and in extremely humid climates, such as the British Isles and southeastern Alaska, Histosols may also occur on gentle slopes (Fig. 6.3B) (Moore and Bellamy, 1974; Sjors, 1985). Histosols of flood plains receive suspended mineral matter during floods (Fig. 6.3C), while other Histosols obtain only dissolved material from source waters or inputs of eolian materials.

In lower latitude coastal areas, continually rising sea level has caused brackish or saline waters to engulf drowned river valleys or to extend over formerly upland soils. This has led to the formation of coastal marsh Histosols in several geomorphic settings (Darmody and Foss, 1979). A schematized cross-section of a submerged upland type marsh where organic soil materials are accumulating over what were previously upland soils, forming Histosols is presented in Fig. 6.4. As sea level continues to rise, the margin of the marsh is pushed landward, and the organic materials continue to thicken so that older and deeper Histosols generally exist nearer to the open water (Rabenhorst, 1997).

The formation of Histosols, like other soils, is a function of topography, but Histosols are unique among soils in that their formation also modifies the topography. Accumulation of organic soil material can fill depressions and create gentle topographic highs that were once level or depressed (Fig. 6.5) which can prevent base-rich water from moving onto the peatland and thereby facilitate formation of a highly acidic bog. Peat accumulation also produces intriguing microtopography on some peatlands, notably a pattern of ridges (or strings) and pools (Foster et al., 1983; Seppälä and

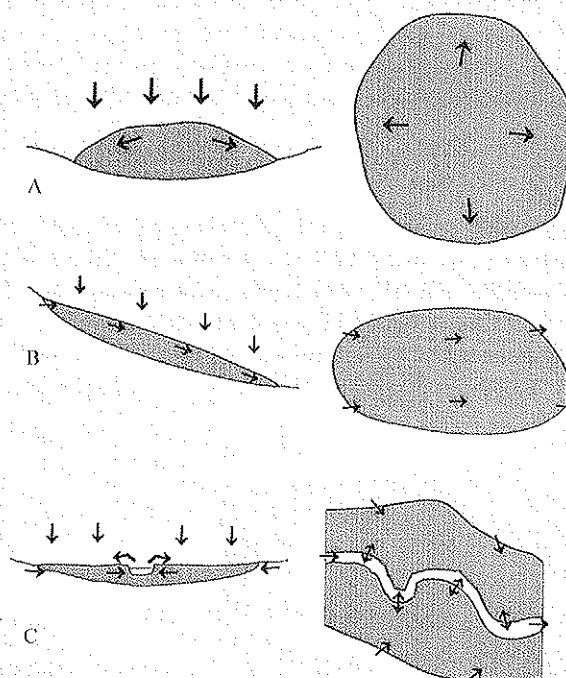


Fig. 6.3 Settings of peatlands. Peat is shown as shaded areas, and arrows indicate the direction of water movement. (A) Bog. The domed surface of the bog precludes input of runoff or groundwater containing bases dissolved from minerals. All water is derived from the atmosphere and evaporates or runs radially off the bog. (B) Fen or swamp. Runoff and groundwater from mineral soils surrounding the peatland supplement precipitation on the peatland. (C) Floodplain fen or swamp. The peatland receives precipitation and runoff or groundwater from mineral soils adjacent to the peatland. Water seeps from the peatland into the stream during periods of low water, while during floods suspended mineral matter is deposited on the peatland, increasing the content of mineral matter in the peat.

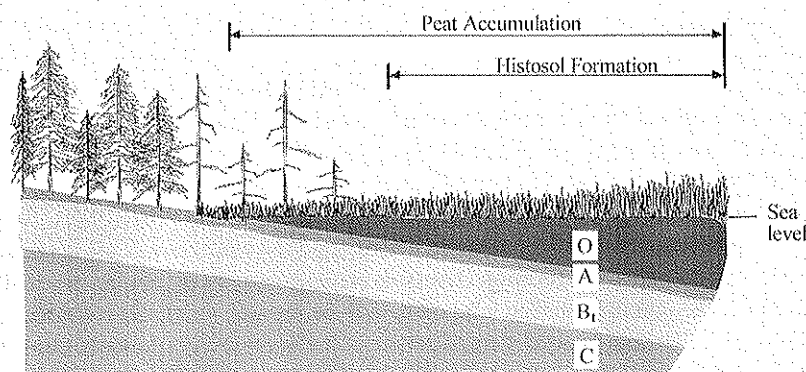


Fig. 6.4 A schematized cross-section of a submerged upland type marsh showing development of O horizons and a Histosol (probably a sulfihemist) over what was formerly an upland soil with an argillic (Bt) horizon (The vertical scale and slope are exaggerated in this diagram.)

Koutaniemi, 1985; Swanson and Grigal, 1988) (or flarks), the cause of which is still debated (Washburn, 1980).

Histosols frequently form by accumulation of organic matter in basins of lakes or ponds, a process known as lakefill or terrestrialization and by paludification, the expansion of wetland onto what was originally drier soils. Both processes operated in the formation of the peatland depicted in Fig. 6.5. Folist, unlike other Histosols, generally occur in mountainous regions (Reiger, 1983; Wakeley et al., 1996) which provide the high precipitation and underlying bedrock or fragmental material required for Folist formation.

6.2.3.4 Vegetation

The flora of Histosols vary widely as a result of the great range in climates over which they occur. Moreover, while most Histosols owe their existence to saturation of the soil by water, the depth at which saturation occurs is variable, which has a major effect on the vegetation. The vegetation of bogs in the worldwide circumboreal zone is remarkably uniform, apparently due to the limited number of plants that can tolerate the poor nutrient conditions, cold climate, and high watertable (Fig. 6.6). *Sphagnum* mosses cover the ground, along with scattered sedges (*Carex*) and cottonsedges

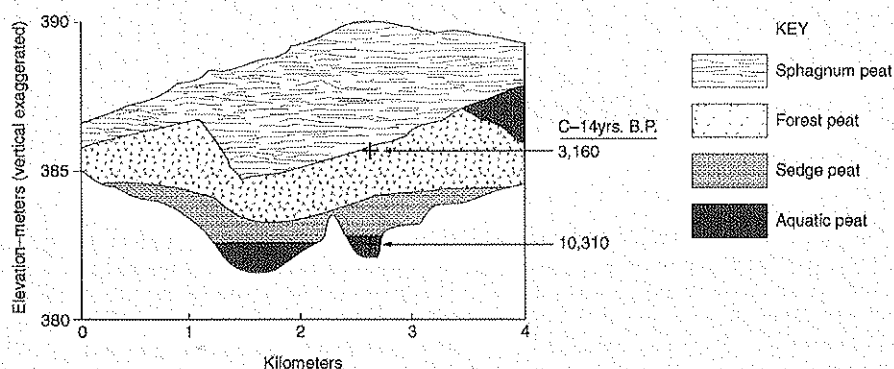


Fig. 6.5 Stratigraphic cross-section of a bog in the Myrtle Lake peatland, northern Minnesota [Modified from Heinzelman, 1970]

(*Eriophorum*). Low shrubs from the family Ericaceae are common, and trees are usually present but stunted; black spruce (*Picea mariana*) is most widespread in North America and Scot's pine (*Pinus sylvestris*) in Eurasia. The vegetation of bogs in more southerly climates includes different species, but shares with northern bogs, the *Sphagnum* moss and prevalence of nutrient-conserving evergreen plants (Anderson, 1983; Hofstetter, 1983).

The less acid conditions of groundwater-fed Histosols permit a greater diversity of plants, many of which also occur on wet mineral soils. On minerotrophic peatlands with the watertable continuously near the surface (fens), most vascular plants are aerenchymous grass-like plants (mainly sedges, family Cyperaceae) that can transport O_2 downward to their roots through hollow stems, allowing root metabolism in anaerobic soils (Crawford, 1983). Swamps occur on soils where an aerobic surface horizon is present at least during part of the growing season, allowing growth of nonaerenchymous plants, which includes most woody plants (Gill, 1970; Kozłowski, 1984). Though nutrient conditions for tree growth on these Histosols are superior to those of bogs, trees may be stunted due to short duration of aerobic conditions or thinness of the aerobic rooting zone.

Many of the plants that occur on Histosols are highly adapted to specific conditions of pH with nutrients and high watertable becoming useful indicators of these conditions (Sjors, 1963; Heinzelman, 1963, 1970; Jeglum, 1971; Vitt and Slack, 1975; Vitt and Bayley, 1984; Andrus, 1986; Janssens and Glaser, 1986; Swanson and Grigal, 1989, 1991; Glaser, 1992; Janssens, 1992). Because the high water table restricts rooting of most plants to near surface soil horizons, vegetation is a useful indicator of pH and nutrient conditions near the surface but not at depth.

Because the vegetation actually creates most of the soil material in Histosols, composition of the vegetation that formed the soil exerts a strong control over its physical and chemical properties. Peats



Fig. 6.6 Typical bog vegetation on a highly acidic Histosol in boreal North America. Trees are stunted black spruce (*Picea mariana*); the largest trees visible are about 4 m tall. Understory plants include ericaceous dwarf shrubs, sedges (*Carex* spp.), and a continuous cover of *Sphagnum* spp. moss. The soil is a Typic Borohemist. Toivola peatland, northeastern Minnesota.

are commonly divided into three broad groups on the basis of botanical composition: moss, herbaceous (sedge), and woody peat (Kivinen, 1977), which are generally deposited in bogs, fens, and swamps, respectively. The botanical origin of organic soil material may be determined by examination of plant remains (Birks and Birks, 1980; Janssens, 1983; Lévesque et al., 1988).

The changes in soil drainage and trophic conditions on peatlands that accompany peat accumulation affect the vegetation, and hence, botanical composition of the peat. As peat accumulates, the rooting zone often becomes more and more isolated from mineral nutrient sources, and portions of the peatland may become drier as the ground surfaces rise. A cross-section of a bog in Minnesota shows how a lake was filled in with aquatic peat (i.e., limnic material) (Fig. 6.4). Then, the peatland expanded onto what was originally dryland of the lakeshore as minerotrophic sedge peat accumulated, followed by minerotrophic forest peat as the surface rose and became drier. By about 3,000 years BP, the center of the peatland became isolated from minerotrophic water, allowing accumulation of highly acidic *Sphagnum* moss peat, which subsequently expanded over the entire peatland and continues to accumulate today. Such changes in vegetation over time in peatlands make it difficult to predict subsurface peat properties from surface vegetation (Swanson and Grigal, 1989).

6.2.3.5 Time

Essentially, all extant Histosols have formed since the end of the Pleistocene epoch. Most northern latitude Histosols occupy regions which were covered by glaciers during the last ice age and have formed following the glacial retreat. Reported average rates of peat accumulation in northern bogs and fens have been as high as $> 1 \text{ mm yr}^{-1}$, but more typically fall in the range of $0.2\text{--}0.7 \text{ mm yr}^{-1}$ (Table 6.5). These average rates usually are based on basal ^{14}C dates and actual rates may have been higher or lower during particular periods.

Although distant from the glacial activity, coastal Histosols at lower latitudes were also impacted by the glaciation. During the glacial maximum ($\sim 20,000 \text{ yr BP}$) when large quantities of water were tied up in the glacial ice, sea level worldwide was approximately 150 m below the present level. Melting of the ice and concurrent ocean warming caused sea level to rise at such a rapid rate ($10\text{--}20 \text{ mm yr}^{-1}$) that initially vegetation could not colonize the tidal regions. Approximately 3,000–5,000 years ago, sea-level rise slowed to a more modest pace so that marsh vegetation could become established and organic parent materials began to accumulate (Bloom and Stuvier, 1963; Redfield, 1972). As sea level has continued to rise, organic materials have accumulated in Histosols, and coastal marshes and mangroves generally have been thought to have accreted at approximately the rate of sea-level rise. In addition to the eustatic sea-level rise, sediments in transgressing coastal regions are subsiding (for example, along the Atlantic and Gulf coasts). The combination of rising sea level (presently estimated at 1 mm yr^{-1} worldwide) and coastal subsidence can be joined to yield an apparent sea-level rise, which is substantially greater. Current estimates of peat accretion in coastal areas generally range from $3\text{--}8 \text{ mm yr}^{-1}$, which are much higher than in noncoastal regions, with even higher rates reported in rapidly subsiding areas (Table 6.5). Current evidence suggests that the highest rates of sea-level rise may be too great for marsh systems to maintain, and that some of these areas are suffering marsh loss.

6.2.4 Morphological Properties

Most organic soil material is derived from terrestrial plants, and soil particles initially resemble the plants from which they were derived. As decomposition progresses, the organic matter is converted into a homogenous, dark colored mass. Some organic soil materials are derived from aquatic plants and animals that accumulate on the bottom of water bodies, producing limnic materials (Finney et al., 1974; Soil Survey Staff, 1996).

Table 6.5 Rates of peat accretion/accumulation in various organic soils

Location	Site characteristics	Peat accretion - accumulation rate mm yr ⁻¹	Source
Bergslagen, Sweden	Raised Bog	0.3 - 1.0	Derived from Almquist-Jacobson and Foster (1995)
Slave Lake, Canada	Sphagnum peat	0.3 - 0.6	Kuhry and Vitt (1996)
West Greenland	Nearshore peat	0.43	Bennike (1992)
Subarctic and Boreal Canada	Based on 138 basal ¹⁴ C dates	0.31 - 0.54	Gorham (1991)
S. Sweden and N. Germany		0.70	Tolonen (1979) (after Gorham, 1991)
S. and Central Finland		0.75	Tolonen (1979) (after Gorham, 1991)
N. Europe		0.60	Aaby (1986) (after Gorham, 1991)
Boreal USSR	Raised bogs	0.6 - 0.8	Botch and Masing (1983)
Siberian USSR	Palsa Province	0.2 - 0.4	Botch and Masing (1983)
Eurasia		0.52	Zurek (1976) (after Gorham, 1991)
Maine		0.35 - 0.75	Tolonen et al. (1988) (After Gorham, 1987)
Minnesota	Red Lake, Minnesota	0.85 - 1.15	Gorham (1987)
Louisiana	Coastal Marsh		Derived from Nyman and DeLaune (1991)
	Fresh	6.5 - 8.5	
	Brackish	5.9 - 9.5	
	Saline	7.5 - 7.6	
Chesapeake Bay, MD	Coastal Marsh, Brackish	3.3 - 7.8	Derived from Kearney and Stevenson (1991)
Chesapeake Bay, MD	Coastal Marsh, Brackish	3.5 - 7.5	Griffin and Rabenhorst (1989)
Chesapeake Bay, MD	Coastal Marsh, Brackish		Hussein (1996)
	Current (²¹⁰ Pb)	1.4-3.2	
	Long term (¹⁴ C)	0.5-1.1	
Louisiana	Barataria Basin, Coastal Marsh	7-13	Hatton et al. (1983)
Massachusetts	Barnstable Coastal Marsh	1.1 - 2.6	Redfield and Rubin (1962)

The most obvious morphological properties of organic soil horizons are related to their degree of decomposition. Master horizons Oi, Oe, and Oa are used to designate fibric, hemic and sapric horizons, respectively (Table 6.6), and are defined by the portion of the soil material which retains plant fibers after rubbing and by the color of a Na pyrophosphate-extracting solution (Soil Survey Staff, 1996). Other methodologies and rating scales have been developed and utilized for evaluating degree of decomposition; the one most broadly used outside of the soil science community and in Europe is the Von Post scale, which ranks organic materials on a scale of decomposition from 1 to 10 based on soil color, quantity of recognizable fibers, and the proportion of material remaining in one's hand when the sample is squeezed. Von Post scale numbers 1-4 correspond approximately to fibric material, 5-7 to hemic, and 8-10 sapric (Von Post and Granlund, 1926; Clymo, 1983).

There are a number of soil properties which are related to the degree of decomposition of the organic materials, including color and a variety of physical and chemical properties. Field moist colors

Table 6.6 Defining morphological criteria for organic horizons [Soil Survey Staff, 1996]

Horizon designation	Type of material	Common descriptor	Volumetric rubbed fiber (RF) content		Color of pyrophosphate extract
Oi	Fibric	Peat	RF > 3/4 or RF > 2/5	and	7/1, 7/2, 8/1, 8/2, 8/3
Oe	Hemic	Mucky Peat	2/5 > RF > 1/6	or	does not otherwise qualify for either Fibric or Sapric materials
Oa	Sapric	Muck	RF < 1/6	and	below or right of line drawn to exclude blocks 5/1, 6/2, 7/3

of sapric organic soil material are usually nearly black. Fibric peat is lighter colored and often reddish in hue, while hemic peat has an intermediate color (Table 6.7). The relation between the degree of decomposition and various physical properties is discussed later under the section on physical properties.

Undrained Histosols typically lack pedogenic structure as it is usually defined, although flattened plant remains and stratification commonly produce a plate-like structure (Lee and Manoch, 1974). This sedimentary structure often results in Histosols having different shear strength and hydraulic conductivity in horizontal and vertical directions (MacFarlane and Williams, 1974; Rycroft et al., 1975). Drained Histosols may develop pedogenic structure in the anthropogenic aerobic zone, such as granular structure due to earthworm casts and blocks or prisms due to cycles of wetting and drying (Lee and Manoch, 1974).

The mineral soil material that occurs under the peat in Histosols is typically chemically reduced as a result of saturation by water and the abundance of organic matter above it. Where paludification has occurred, soil horizons from prior mineral soils may be buried beneath the peat. In some cases, the preceding mineral soil pedogenesis may facilitate paludification by producing low permeability horizons such as placic horizons (Ugolini and Mann, 1979; Klinger, 1996). The depth to mineral soil underlying drained Histosols has been measured with some success by ground-penetrating radar (Shih and Doolittle, 1984; Collins et al., 1986). The morphology of Folists differs from that of other Histosols in that the peat is thinner and underlain by fragmental material or bedrock (Witty and Arnold, 1970; Everett, 1971; Lewis and Lavkulich, 1972).

6.2.5 Micromorphology

Micromorphological observations can provide a direct examination of the structural integrity of plant fragments and components in organic soil materials. Fox (1985) and Lévesque and Diné (1982) have

Table 6.7 Color of organic soil material in relation to degree of decomposition¹

Degree of decomposition	Median Munsell color (hue value/chroma)	N
Sapric	10YR 2/1	69
Hemic	10YR 3/2	49
Fibric	7.5YR 3/2	18

¹Data include all organic horizons of Histosols in the USDA Natural Resources Conservation Service, National Soil Survey Laboratory characterization database.

summarized the characteristics of organic soil materials at various stages of decomposition (fibric, hemic and sapric materials). Fibric materials mainly show unaltered or slightly altered plant tissues without appreciable darkening, and with little organic fine material. The plant fragments, which are only slightly decomposed, appear to be loosely arranged with a porous and open structure. Partially decomposed (hemic) materials also possess a fibrous appearance, and most fragments show incomplete degradation. The development of brown or black colors in the plant tissues is typical. Fine organic material is also present intermixed with, or adhering to, the coarser fragments of plant tissue (Fig. 6.7). Fecal pellets, which are evidence of faunal activity, can also be common. In the most highly decomposed (sapric) materials, organic fragments are sufficiently darkened and decomposed that identification of botanical origin is not possible. Fine organic material is usually the dominant component although faunal excrement is also common.

The effects of draining organic soils can sometimes be seen during microfabric examination. When an undrained sphagnum peat profile in Ireland was compared with those which had been drained for between 10 and 100 yr, the drained profiles had undergone substantial alteration and decomposition leaving little of the original tissue structures (Hammond and Collins, 1983). The fine organic material was dominant, showing some biological granulation (Pons, 1960; Lee, 1983). The change in microfabric directly corresponded to an increase in density of the material.

In another study, Lee and Manoch (1974) concluded that 50 yr of drainage and cultivation of organic soils led to significant decomposition and the development of pedogenic structure in the subsoil, whereas in the lower portion of the profile where the soil remained saturated, sedimentary structure persisted and the peat was more highly fibrous and less decomposed. The activity of soil fauna in the drained portions of the soil contributed to biological granulation and the formation of two

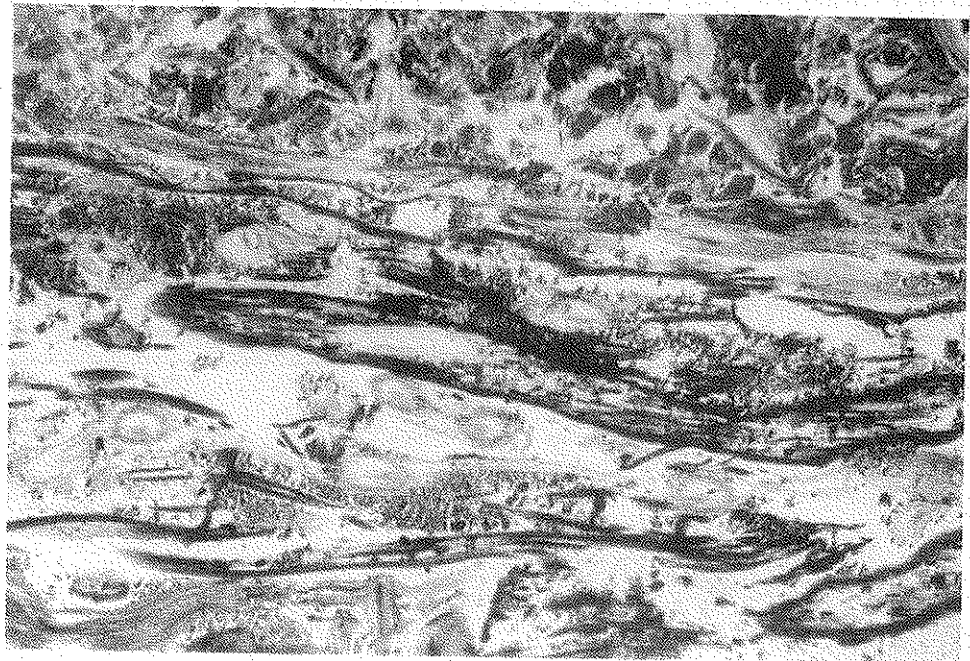


Fig. 6.7 Thin section showing organic material from hemic Oe horizon (80-88 cm) of a Typic Sulfihemist in a coastal marsh of Chesapeake Bay; the organic material reflects an intermediate degree of decomposition with some discernible plant structures and cell components intermixed with decomposed organic material; frame length = 5mm.

distinct types of surface horizons. The *moder* mostly consists of faunal excrement and usually forms in oligotrophic peats, while the *mull* is formed by an intense mixing and binding of organic with mineral particles by larger organisms such as earthworms, and usually forms in mesotrophic or eutrophic peats.

While not widely reported, following drainage and cultivation of organic soils, illuvial humus termed humilluvic material (Soil Survey Staff, 1996) may accumulate in the lower parts of acid organic soils (van Heuveln et al., 1960). Both the lower pH of the oligotrophic peat and the disturbance by cultivation apparently contribute to the dispersion of the organic fraction which can then be translocated within the soil, and accumulates within the lower horizons of the peat at the peat-mineral soil contact, or within the underlying mineral soil material.

6.2.6 Classification

The definition of organic materials for saturated soils requires a minimum of 12% organic C, if there is no clay, and a minimum of 18% organic C if the soil contains $\geq 60\%$, with a sliding scale for intermediate textures. Those soils, which are not saturated, must contain at least 20% organic C to be considered organic soil materials (Soil Survey Staff, 1998, 1996). In order for a soil to be classified as a Histosol, at least 40 cm of the upper 80 cm must be comprised of organic materials, and it must not have permafrost within 1 m of the surface. However, if the organic materials are especially low in density ($< 0.1 \text{ g cm}^{-3}$), then at least 60 cm of the upper 100 cm must be organic materials. Histosols may be buried by as much as 40 cm of overlying mineral soil materials and still be considered Histosols.

The types of differentiating characteristics used to discriminate between classes of soils at the various categorical levels are presented in Table 6.8. Basically, organic soils which are not saturated for extended periods are classified as Folists, while the saturated organic soils are differentiated according to the degree of decomposition of the organic materials in the subsurface tier (the zone approximating 40–100 cm) into Fibrists, Hemists, or Sapristis. Within the United States, some 234 soil series have been established for Histosols distributed among the classes as shown in Table 6.9. The number of series that exist within a particular class may result from many factors and should not be taken to represent the areal extent of those soils. Histosols classified at the family level are differentiated into classes based upon particle size and mineralogy (used only for terric subgroups or for those containing limnic materials), reaction ($\text{pH}_{\text{CaCl}_2}$), temperature regime, and soil depth (only used if $< 50 \text{ cm}$ deep.)

Table 6.8 Criteria utilized in the classification of Histosols

Suborder	Great group	Subgroup	Family
Degree of saturation with water	Soil temperature regime	Thickness of organic materials (Terric vs. Typic)	Particle size and mineralogy (used only for Terric subgroups or for those containing limnic materials)
Degree of decomposition of the subsurface tier	Special components (sphagnum fibers, sulfidic materials or sulfuric horizon, humiluvic materials)	Underlying materials	Reaction ($\text{pH}_{\text{CaCl}_2}$)
		Special materials contained (Limnic)	Temperature regime
		Intergrades to other great groups (Cyrlic, S phagmic)	soil depth (only used if $< 50 \text{ cm}$ deep)

Table 6.9 Number of soil series in the United States which are classified into particular taxonomic groups

Suborders	Number of Series	Great Groups		Subgroups	
		Formative Element	Number of Series	Formative Element	Number of Series
Fibrists	17	Boro [†]	60	Fluvaquentic	11
Folists	25	Cryo	22	Limnic	14
Hemists	61	Medi [‡]	98	Lithic	29
Saprists	131	Sphagno	5	Terrie	88
		Sulfi	18	Typic	76
		Tropo	31	Others	16
Total	234	Total	234	Total	234

[†] This great group is no longer used in *Soil Taxonomy* and these series have been reclassified to alternative taxa.

[‡] This great group was essentially replaced by the Haplo great group in the 1998 version of *Soil Taxonomy* (Soil Survey Staff, 1998).

6.2.7 Biological and Chemical Properties

Organic C contents are generally higher and N contents lower in less decomposed than highly decomposed peats giving high C:N ratios and ash contents for less and more highly decomposed peats, respectively (see also Table 6.12) (Lévesque et al., 1980; Lee et al., 1988).

Thus, more highly decomposed peats are generally more fertile than less decomposed peats. Some drained, sapric peats may, in fact, supply N in excess of crop requirements without fertilization. However, nutrients derived from minerals, such as P, K, and most micronutrients, are frequently deficient in Histosols (Lucas, 1982; Yefimov, 1986).

The chemical and physical properties of peats are also related to their botanical composition. Peats derived mostly from *Sphagnum* mosses tend to be more acid, less decomposed, contain less ash, have lower CEC and bulk density than woody peats; sedge peats typically have intermediate properties (Farnham and Finney, 1965; Lévesque et al., 1980).

6.2.7.1 Soil Carbon

Recent interest in the global C cycle has focused attention on the high proportion (75%) of terrestrial C stored in soils (Lal et al., 1995). As a group, wetland soils maintain a disproportionately high level of soil C, and Histosols, which are composed largely of organic matter, clearly store the highest quantities. While typical agricultural soils may contain between 2–10 kg C m⁻², reported values for Histosols typically are an order of magnitude greater, with some values as high as nearly 200 kg C m⁻² (Table 6.10). The quantity of C stored in some very deep Histosols is undoubtedly even higher.

Histosols are very dynamic and may be particularly significant in the overall terrestrial C budget. Many Histosols continue to sequester C at significant rates. This is particularly true for soils of coastal marshes, where rising sea level continues to power the engines of marsh accretion and C storage. Therefore, Histosols are generally viewed as a significant C sink. However, because Histosols contain such a high concentration of soil C, if they are drained or in some other way exposed to an aerobic environment, they may begin to oxidize and in the process yield large quantities of C to the atmosphere.

Most of the discussion of possible global warming and greenhouse gas emission has focused on rising levels of CO₂ in the atmosphere. Methane, however, is 32 times more efficient than CO₂ in trapping infrared radiation. Because many Histosols are strongly reducing (low Eh), they represent an

Table 6.10 Carbon storage values for organic soils

Site Characteristics	Location	Carbon Accumulation Rate kg m ⁻² yr ⁻¹	Quantity of stored C kg m ⁻²	Reference
Coastal Marsh	Louisiana			Derived from Nyman and DeLaune (1991)
Fresh		0.17 - 0.22		
Brackish		0.17 - 0.27		
Saline		0.21 - 0.22		
Coastal Marsh	Chesapeake Bay, Maryland	0.12 - 0.42		Derived from Kearney and Stevenson (1991)
Brackish				
Coastal Marsh	Chesapeake Bay, Maryland		59 (range 18-166)	Derived from Griffin and Rabenhorst (1989)
Brackish				
Barataria Basin, Coastal Marsh	Louisiana	0.18 - 0.30		Smith et al. (1983)
Coastal Marshes	Atlantic and Gulf Coasts		64 (range 9 - 191)	Rabenhorst (1995)
Sphagnum peat	Slave Lake, Canada	0.014 - 0.035		Kuhry and Vitt (1996)
Based on 138 basal ¹⁴ C dates	Subarctic and Boreal Canada	0.023 - 0.029		Gorham (1991)

ideal environment for the formation of CH₄. In systems where SO₄²⁻ is more abundant in the soil solution, such as in coastal or estuarine environments, SO₄²⁻ reduction is favored over methanogenesis and CH₄ production may be more limited (Bartlett et al., 1987; Widdell, 1988) (Section C, Chapter 4.3). However, in many freshwater or inland areas, Histosols may be the locus of significant CH₄ emission to the atmosphere (Table 6.11). Minerotrophic fens have higher CH₄ emissions than ombrotrophic bogs. Methane emission from Histosols seems to be directly related to the location of the watertable, with greater generation when soils are saturated to or above the surface. This is due to the fact that soil microbes in the aerobic zone utilize CH₄ as it passes through on its way to the surface. Carbon dioxide emissions, which are produced by oxidation of SOM, are generally greater when watertables drop.

6.2.7.2 Sulfides

The biogeochemical environment in which Histosols form can also be conducive to the formation of sulfides. The occurrence of Fe sulfides can, in some circumstances, lead to the generation of extreme acidity and acid sulfate soils (van Breemen, 1982) (Section B, Chapter 5). Sulfate reduction generally requires the presence of organic matter (energy source), low Eh, SO₄²⁻ (electron acceptor), and SO₄²⁻ reducing bacteria (Rickard, 1973). If sulfide is formed in the presence of a reactive Fe source, then such minerals as pyrite (FeS₂) can form. The C source and anaerobic conditions are almost always present in Histosols, but SO₄²⁻ levels may vary dramatically among environments. Many inland Histosols receive SO₄²⁻ only in small amounts as atmospheric deposition, and under these circumstances, sulfidization (Rabenhorst and James, 1992) is insignificant, and most of the S in those soils is bound in organic forms (Novak and Wieder, 1992). Some inland peats have developed acid sulfate conditions, although usually they are associated with deposits of cuprogenous earth (Lucas, 1982). On the other hand, SO₄²⁻ reduction is common in coastal Histosols which contain an abundance of SO₄²⁻ from sea water. Extensive areas of these Sulfidic soils have been identified along the Atlantic coast of the United States. The distribution of FeS₂ within coastal Histosols can be highly

Table 6.11 Reported fluxes of CO₂ and CH₄ emissions from Histosols

Location	Site Details	Notes	CO ₂ emission rate		CH ₄ emission rate	
			Mean	Range	Mean	Range
			mmol m ⁻² d ⁻¹			
Quebec Canada ¹	Gatineau Park			1.97–7.24		1.15–2.18
Wales ²	Peat Monoliths	Laboratory study	14.7	9.6–21.0	14.4	4.7–34.4
Finland ³	Natural Fen	Annual	11.3		3.49	
	Drained Fen		30.8		0.03	
Finland ⁴	Ombrotrophic	12 °C	88.2	42.5–141.3		
Alaska ^{5,6}		Summer measurements			9.2	
Boreal Canada ⁷	Swamp (n = 20)	Annual Averages			0.21	
	Fen (n = 6)				0.57	
	Bog (n = 13)				0.39	
Canada ⁸	Bog flooded	19–23 °C	0.005		0.012	
(lab study)	Bog drained		0.19		0.006	
	Fen flooded		0.009		0.58	
	Fen drained		0.14		0.025	
Alaska ⁹	Moist tundra	August			0.3	
	Waterlogged tundra				7.4	
	Wet meadows				2.5	
	Alpine fen				18	
Northern Sweden ¹⁰	Ombrotrophic bog	Summer				
	Hummocks		10.12		0.05	
	Between humocks		14.92		0.14	
	Shallow depressions		11.61		0.77	
	Deeper depressions		12.45		1.21	
	Ombro-Minerotrophic		12.49		2.73	
	Minerotrophic Fen		11.48		16.89	
Minnesota ¹¹	Bog	Sampled during August			8.2	1.2–29.2
	Fen				0.25	0.19–0.31
Georgia ^{5,6}		During midsummer			6.6	
West Virginia ^{5,6}	Mountain Bog	During midsummer			11.7	
Minnesota ^{6,12}	Forest bog	Summer measurement			3.6	0.5–5.33
	Forest fen				6.7	3.2–12
	Open bog				14	0.9–41
	Neutral fen				15	7.1–33
	Acid fen				4.8	
Minnesota ¹³	Open poor feen	Winter measurement			3.0	
	Open bog				0.7	
	Forest bog hollow				0.8	
	Hummock				0.3	
Virginia ¹⁴	Coastal Marsh	Summer				
	York River	Creek bank			0.45	0.13–0.82
	Chesapeake Bay	Short spartina			0.29	0.05–0.87
	Estuary	High marsh			0.09	0–0.36
Virginia ¹⁵	Coastal Marsh	Summer				
	Tidal creek in	Low salinity			11	5–16
	Chesapeake Bay	Moderate salinity			7.5	4–11
	Estuary	High salinity			2	0.5–2.5
West Virginia ¹⁶	Appalachian Bog		177	75–250	17.0	0–53
Maryland ¹⁶	Appalachian Bog		152	100–250	4.4	0–12
Louisiana ¹⁷	Barataria Basin, Coastal Marsh	Annual averages		41–141		
Florida ^{5,6}		During midsummer			6.0	
Malaysia ¹⁸			170			
Malaysia ¹⁹	Drained & cultivated peatland			139–727		

¹Buttler et al. (1994), ²Freeman et al. (1993), ³Nykanen et al. (1995), ⁴Silvola et al. (1996), ⁵After Gorham (1991), ⁶after Crill et al. (1988), ⁷Derived from Moore and Roulet (1995), ⁸Derived from Moore and Knowles (1989), ⁹Derived from Sebacher et al. (1986), ¹⁰Svensson and Rosswall (1984), ¹¹After Harriss et al. (1985), ¹²After Mitsch and Wu (1995), ¹³Dise (1992), ¹⁴Derived from Bartlett et al. (1985), ¹⁵Derived from Bartlett et al. (1987), ¹⁶Wieder et al. (1990), ¹⁷Smith et al. (1983), ¹⁸Wosten et al. (1997), ¹⁹Murayama and Bakar, 1996

variable and is often related to microsite differences in the availability of either reactive Fe or S^{2-} (Rabenhorst and Haering, 1989).

6.2.7.3 Acidity and Base Saturation

In Histosols of bogs, where inputs of bases are minimal because soil water is derived from rainfall that has never contacted mineral soil, base saturation is low (at least in the rooting zone) and the organic acids produced by partial decomposition of organic matter typically buffer soil water pH near 4; the soil pH_{CaCl_2} is typically 3–4 (Figs. 6.8 and 6.9) (Gorham et al., 1985; Urban, 1987). Minerotrophic (fen and swamp) peats have higher base saturation and water pH is generally above 5 ($pH_{CaCl_2} > 4.5$). The pH of minerotrophic peats is buffered by cation exchange with the soil (Bloom et al., 1983) or by carbonates if they are present. The Al^{3+} ion comprises a small proportion of the total acidity in most Histosols, because organic soil material is rich in H^+ but contains little Al which is derived from silicate minerals, and has a low solubility at $pH > 5.5$ (Table 6.12). The pH is generally lower in less than more highly decomposed peats (Table 6.12) (Lee et al., 1988).

The distinction between euic and dysic reaction classes at the family level in *Soil Taxonomy* separates the highly acid, bog peats from less acid fen and swamp peats (Farnham and Finney, 1965). Because a $pH_{CaCl_2} > 4.5$ anywhere in the control section (i.e., anywhere within 130 or 160 cm of the surface) places the soil into the euic class, a bog with highly acidic near-surface rooting zone may classify as euic rather than dysic due to the presence of higher pH horizons at depth in the soil.

Histosols which contain sulfide minerals such as FeS_2 have the potential to develop extreme acidity. Under saturated and anaerobic conditions, such sulfide-bearing soils have circumneutral pH. If they are drained, dredged, or in some other way exposed to oxidizing conditions, acid sulfate weathering takes place (van Breemen, 1982). The oxidation of pyrite can lead both to the extreme acidification of the soil ($pH < 3.5$) and also to the generation of acidity, which can be moved offsite through mobilization of acid-generating soluble salts such as $FeSO_4$.

6.2.7.4 Cation Exchange Capacity

The CEC of Histosols is quite high as a result of the high CEC of SOM (Table 6.12). The CEC is higher for more highly decomposed peats than fibric peats (Table 6.12) (Lévesque et al., 1980). The difference between the CEC of sapric and fibric peats is even more marked if the CEC is expressed on a volume rather than mass basis, as a result of the very low bulk densities of fibric materials (Table

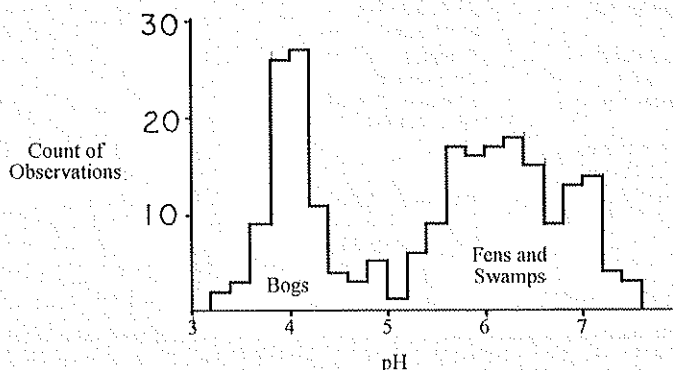


Fig. 6.8 Frequency distribution of pH in peatland surface water for 232 sites in Minnesota. Samples with surface water pH near 4 are from bogs and those with pH above 5 are from fens (unforested) and swamps (forested) [Swanson and Grigal, 1989].

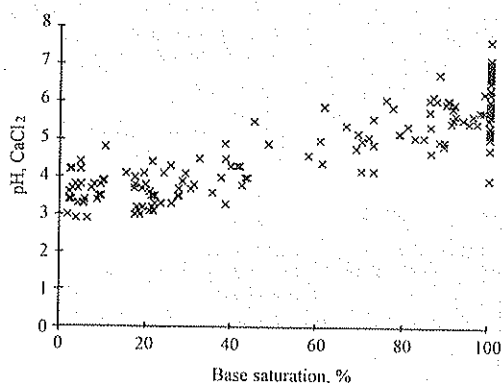


Fig. 6.9 Relationship between base saturation and pH of organic soil horizons of Histosols. Peat samples with pH in CaCl_2 of less than 4.5 (presumably deposited in bogs) generally have less than 50% base saturation. Peat samples with pH greater than 4.5 (deposited in fens and swamps) typically have high base saturation. Data are for all organic horizons of Histosols in the USDA Natural Resources Conservation Service, National Soil Survey Laboratory characterization database.

6.12). The CEC per unit soil volume is a more useful measure than CEC per unit soil mass in Histosols, because their bulk densities are very low and because roots occupy a volume rather than mass of soil. The CEC per unit volume of fibric peats ($\sim 20 \text{ cmol}_e \text{ L}^{-1}$) is comparable to that of most mineral soils ($5\text{--}20 \text{ cmol}_e \text{ L}^{-1}$) assuming mineral soil bulk density of $1.0\text{--}1.5 \text{ kg L}^{-1}$ (Holmgren et al., 1993). Even when expressed per unit volume, the average CEC of sapric and hemic peats (Table 6.12) is much higher than the $5\text{--}20 \text{ cmol}_e \text{ L}^{-1}$ of most mineral soils.

6.2.8 Physical Properties

6.2.8.1 Bulk Density

The physical properties of Histosols differ greatly from those of mineral soils. Bulk densities for organic soil materials generally are quite low, ranging from as little as 0.02 up to 0.30 g cm^{-3} . Bulk

Table 6.12 Chemical properties of organic soil material as related to degree of decomposition: mean (std. deviation, N)¹

Property	Sapric	Hemic	Fibric	AOV Probability ²
Organic Carbon, g kg^{-1}	313 (128, 129)	347 (135, 61)	372 (130, 26)	0.055
Total Nitrogen, g kg^{-1}	18 (9, 131)	16 (6, 54)	14 (5, 23)	0.058
C:N ratio	21 (10, 113)	25 (11, 48)	27.5 (10, 20)	0.007
CEC, $\text{cmol}_e \text{ g kg}^{-1}$	101 (44, 129)	88 (41, 61)	83 (33, 28)	0.046
CEC, $\text{cmol}_e \text{ L}^{-1}$	76 (42, 28)	44 (25, 25)	21 (2, 5)	0.000
Ash, g kg^{-1}	250 (nd)	178 (110, nd)	100 (50, nd)	nd
pH	5.1 (1.2, 143)	4.9 (1.2, 59)	4.5 (1.2, 27)	0.024
Al, $\text{mol Al mol}^{-1} \text{ TEA}^3$	0.074 (0.076, 49)	0.022 (0.024, 23)	0.038 (0.050, 16)	0.004

¹Data (except for ash) are for all organic horizons of Histosols in the USDA Natural Resources Conservation Service, national Soil Survey laboratory characterization database. Ash content is taken from Lee et al., 1988; data for 1300 samples of Wisconsin Histosols, nd - no data.

²F-test probability from one-way analysis of variance between fibric, hemic, and sapric peats

³KCl extractable Al divided by total NH_4 -acetate extractable acidity

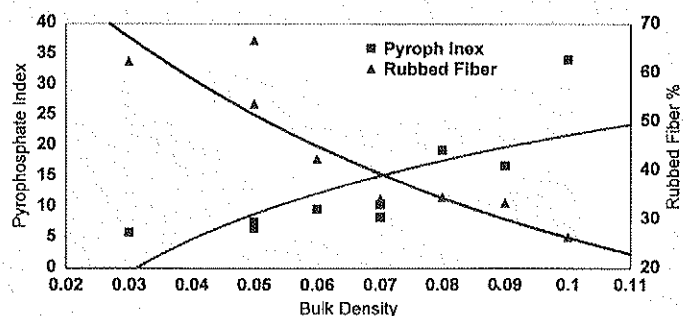


Fig. 6.10 Relationship between bulk density of organic materials and indices of decomposition such as rubbed fiber content or pyrophosphate index [Derived from Buttler et al., 1994]

density is clearly related to the degree of decomposition of the organic materials generally increasing as the materials become more highly decomposed (Fig. 6.10). Because organic materials contain varying amounts of mineral matter, this also can affect the bulk density of organic soil horizons. For coastal marsh peat, the organic matter content is generally about twice the content of organic C, with the remainder representing the mineral fraction. The relationship between the mineral content (roughly the difference remaining from twice the organic C content) and the bulk density for Oe and Oa horizons from Sulfihemists along the Atlantic Coast is illustrated in Fig. 6.11 (Griffin and Rabenhorst, 1989).

6.2.8.2 Porosity, Hydraulic Conductivity, and Water Retention

Histosols have very high porosity levels which can reach over 80% as shown by the water content at saturation (Fig. 6.12) (Boelter, 1969). The high porosity and low bulk density of organic soils would lead one to expect high rates of water transmission through Histosols. Weakly or undecomposed peat often has a fairly high hydraulic conductivity. However, as the material becomes more decomposed the hydraulic conductivity decreases (Table 6.13). The hydraulic conductivity of sapric and some hemic peats is quite low, comparable to that of clay (i.e., 10^{-5} m s^{-1}). Because the peat in the lower part of Histosol profiles is often sapric or hemic, little deep percolation occurs and water tends to evaporate or move laterally through the less decomposed surface horizons (Ivanov, 1981). Thus peat

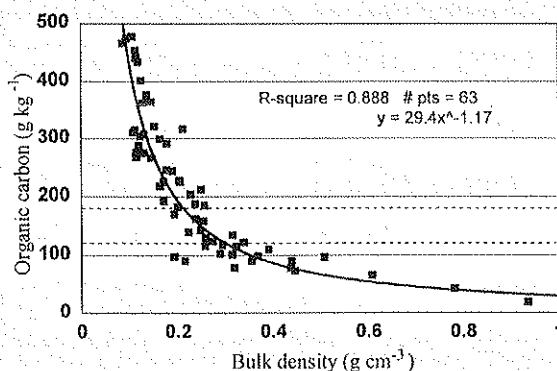


Fig. 6.11 The effect of mineral content (roughly the difference remaining from 2X the OC content) on the bulk density of Oe and Oa horizons from Sulfihemists along the Atlantic Coast. Dashed lines represent 12 and 18% OC which is necessary for soil materials to be considered organic [Derived from Griffin and Rabenhorst, 1989].

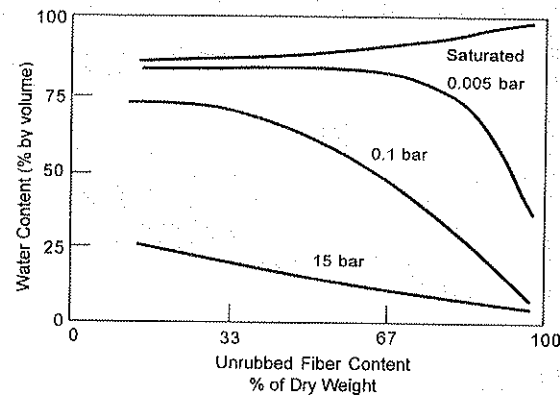


Fig. 6.12 Volumetric soil water content as a function of unrubbed fiber content at various moisture tensions [Reprinted from Boelter, 1969. *Soil Sci. Soc. Am. Proc.* 33:606-609 with permission of the Soil Science Society of America]

accumulation helps to maintain the watertable near the surface by limiting water loss to percolation. While the increased density corresponds to a decrease in porosity, this change alone is insufficient to explain the dramatic reduction in conductivity. Evidently, the pore continuity is reduced in more strongly decomposed materials.

Water retention in Histosols is also closely related to the degree of decomposition (Fig. 6.12). The large pores of slightly decomposed peats drain readily at low suction. In contrast, the fine pores of more well-decomposed peats retain considerable water at low suctions (Boelter, 1974). The water retention difference (water content at -0.01 MPa (-0.1 bar) minus that at -1.5 MPa (-15 bar); approximates plant-available soil water capacity) of most Histosols is very high; it approaches 50% in some peats and exceeds 25% in all except the least decomposed peats (Fig. 6.12).

6.2.9 Utilization and Management

6.2.9.1 Interest in Preservation of Coastal and Nontidal Wetlands

Apart from any benefits which may be achieved from managed systems, such as commercial forestry, grazing or intensive agriculture, Histosols perform a number of environmental and ecological functions. Because most Histosols occur in wetland environments, they typically provide wetland functions including wildlife habitat, flood-water control, groundwater recharge, nutrient and biogeochemical cycling, ion sorption, purification of surface and shallow groundwater, as well as functioning as an important sink for terrestrial C. The importance of peatlands as paleoenvironmental and archeological archives has also been documented (Godwin, 1981.) Throughout the years, the value which society has placed on these functions has been minimal, and the Histosols of peatlands and coastal marshes have been exploited and extensive areas destroyed. More recently people have recognized that the functions which Histosols perform in a natural setting have significant benefit for society, and legislation has been passed in the United States and elsewhere to preserve Histosols and other wetlands.

6.2.9.2 Histosols as Agricultural Resources

As was mentioned earlier, Histosols are utilized as important agricultural soils in many areas, so long as watertables can be effectively managed. For example, in Japan, over 70,000 ha of rice are grown on

Table 6.13 Physical and hydraulic properties of organic soils

Location	Site Characteristics	Bulk Density g cm ⁻³	Hydraulic Conductivity 10 ⁻⁵ cm s ⁻¹	Reference
Northern Minnesota	Undecomposed Partially decomposed Decomposed		3,810 - 15,000 13.9 - 132 0.9 - 15	$\bar{A}^\dagger = 8,650$ $\bar{A} = 63$ $\bar{A} = 5.1$ Boelter (1965)
Wyoming	mountain bog	46 cm 91 cm	0.16 - 0.22 0.0277 0.0185	Sturges (1968)
Ottawa and St. Lawrence River Valleys, Canada	Swamp and bog	0 - 60 cm 0 - 100 cm 0 - 125 cm	624 366 269	Mathur and Lévesque (1985)
Northern Minnesota	Lost River Peatland Bog Fen Margin Spring Fen	0.06 - 0.14	25 - 560 150 - 2,600 67 - 1,600	Chason and Siegel (1986)
Quebec, Canada	Gatineau park	0.03 - 0.10		Buttler et al. (1994)
Eastern New Brunswick, Canada	Sphagnum peat from raised bogs - Von Post scale	1 - 2 3 - 4 5 - 6 7 - 8	0.02 - 0.08 0.03 - 0.10 0.06 - 0.11 0.09 - 0.13	90 - 175 6.9 - 56 1.4 - 17 0.14 - 2.8 Korpilaako and Radforth (1972)
Wisconsin	Fibric Hemic Sapric	0.13 0.17 0.20, 0.24		Lee et al. (1988)
Minnesota	Bog 0 - 10 cm Von Post 1 10 - 20 cm Von Post 2 - 3 20 - 30 cm Von Post 4 - 5 30 - 40 cm Von Post 5 - 6 40 - 50 cm Von Post 5 - 7 Fen 0 - 10 cm 10 - 20 cm 20 - 30 cm		23,495 7,697 5,498 799 995 31,597 5,000 1,400	Gafni and Brooks (1990)

[†] $\bar{A}$ = average

peatlands. Peatlands in a number of tropical countries such as Cuba, Guyana, Malaysia and Indonesia have been reclaimed from mangroves and are used for the production of sugarcane (Moore and Bellamy, 1974). Extensive vegetable production on peatlands is practiced in the northern and southern United States. The agricultural use of Histosols presents some special challenges regarding nutrient management and fertility, but the most significant problem is the high watertable requiring drainage.

6.2.9.3 Impacts of Drainage

Where Histosols have been converted for higher intensity land uses, such as in agriculture, horticulture, or silvaculture, they are almost always drained to lower the watertable. Such drainage results in a number of short-term effects, such as shrinkage and consolidation due to desiccation, the loss of the buoyant force of groundwater, and compaction. Ongoing consolidation and fabric alteration are caused by the enhanced decomposition of the organic materials following the shift from an anaerobic to an oxidizing regime (Stephens and Speir, 1969). Reported rates of long-term peat subsidence range up to 10 cm yr⁻¹ but most reports are in the range of 2-5 cm yr⁻¹ (Table 6.14). This

Table 6.14 Reported rates of subsidence after drainage of organic soils

Location	Site Characteristics	Subsidence rate cm yr ⁻¹	Length of record yr	Reference
Florida	Everglades, Muck and Peat	3.2	41	Thomas (1965)
Hunts, England	Holme marsh	3.4	103	Nickolson (1951)
California	Sacramento-San Joaquin Delta, muck and peat 1 - 3 cm deep	6.4 - 9.8	26	Weir (1950)
Indiana		3.0	30+	Ellis and Morris (1945)
Northern Indiana	Muck	1.2 - 3.0 dependent on WT level	6	Jongedyk et al. (1950)
Michigan		0 - 3.6	5	Davis and Engberg (1955) [after Thomas (1965)]
Southern Ontario, Canada	Holland march, deep loose muck	3.3	19	Mirza and Irwin (1964)
Minnesota		5.1	3	Row (1940)
Florida	Everglades	3	55	Stephens and Speir (1969)
Florida	Everglades	2.3 - 1.8	20	Shih et al. (1981)
California	Sacramento-San Joaquin Delta	2.3	78	Rojstaczer and Deverel (1995)
Quebec, Canada		2.1	38	Parent et al. (1982)
Malaysia		2		Wosten et al. (1997)

Table 6.15 Peat mining by country, 1994 (in millions of tons)¹

Country	Agriculture and Horticulture Use	Fuel Use
Belarus	10	6
Canada	1	
Estonia	4.5	1
Finland	0.6	8
Germany	2.8	0.2
Ireland	0.3	6.4
Latvia	4.5	0.5
Lithuania	4.5	0.2
Russia	60	4
Sweden	0.3	1.4
Ukraine	20	1
United States	0.5	Nd ²

¹Data from Cantrell (1994); includes countries with 500,000 tons or more production²No data

consolidation is accompanied by changes in the physical properties of the peat, including higher bulk densities and lower moisture contents.

6.2.9.4 Histosols as Energy Resources

Peat is mined in northern Europe and used as fuel (Table 6.15), but a large amount is used as a horticultural amendment and as an ingredient in potting soils, mixed fertilizers, components of mushroom beds, as a seed inoculant, as a material for packing of flowers, plants etc., as well as a general soil amendment to increase organic matter content in gardens, golf courses, etc (400,000 Mg in the United States) (Cantrell, 1994).

6.2.9.5 Engineering Properties

Histosols are notorious for their low strength and great compressibility, which make them poor foundation materials for roads, buildings, and other structures. Compression and settlement of peats may continue for years after loading (MacFarlane, 1969; MacFarlane and Williams, 1974; Dhowian and Edil, 1981). Special engineering techniques, such as removal of the peat or precompression before construction, are thus required. The high water content and acidity of most Histosols also make corrosion of concrete and metal structures a potential problem (MacFarlane, 1969; MacFarlane and Williams, 1974). While Histosols are poor foundation materials, their high porosity and great adsorption capacity make them very useful for treatment of wastewater. Peats have potential for treatment of municipal effluent and removal of heavy metals and hydrocarbon pollutants from wastewater (Malterer et al., 1996).

6.2.10 References

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6.3 Andisols

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6.3.1 Introduction

The Andisol order was added to *Soil Taxonomy* as a result of a proposal developed by Smith (1978) who provided the rationale for separating these soils into an order category. The name Ando soil (J. An, dark, and do, soil) connoting acid soils derived from volcanic ash which have thick, dark horizons was introduced in 1947 during reconnaissance soil surveys by United States soil scientists in Japan (Simonson and Rieger, 1967). Part of the name was retained as a formative element in the systems of nomenclature used in different global soil classification systems. The term Ando soil is still used in Japan (Wada, 1986) but evolved to Andosols in the FAO soil classification system (FAO-UNESCO, 1989) and Andepts in *Soil Taxonomy* (Soil Survey Staff, 1975). The term Andisol was introduced in *Soil Taxonomy* in 1990 (Soil Survey Staff, 1990).

The central concept of Andisols is one of deep soils with depositional stratification developing from volcanic ash, pumice, cinders or other volcanic ejecta and volcanoclastic materials, with an exchange complex dominated by amorphous materials composed of Al, Si and organic matter, or a matrix-dominated by glass, having one or more diagnostic horizons other than an ochric epipedon. Bulk density is always comparatively low in most horizons, although the actual values may vary with degree of weathering, moisture regime, and in a very few cases, with the degree of cementation by silica or other cements. The most common diagnostic horizon sequences are an umbric or mollic (rare) epipedon and a cambic horizon, or an ochric epipedon and a cambic horizon. In the driest regimes, a duripan may be found while in the wettest, a placic horizon is not uncommon. A melanic (black) epipedon was added during the testing of the proposal. While soils which are not formed on volcanic parent materials sometimes are included in the Andisol order, Andisols are classified on selected chemical, physical and mineralogical properties acquired through weathering and not on parent material alone, which is sometimes a common misconception.

Andisols possess a suite of common properties including amorphous (short-range order) mineralogy, low permanent and high variable charge, low bulk density, high C content and high P and water retention relative to other mineral soils of similar field texture. Field texture has been used

because laboratory particle size analyses of Andisols are often erroneous due to difficulties in obtaining complete soil dispersion (Warkentin and Maeda, 1980). Andisols pose unique engineering problems because of the fragility of pumice and their thixotropic properties (Section 6.3.6.4). In addition to Andisols, many other soils (Spodosols, Mollisols, Ultisols, Alfisols, Entisols, Inceptisols and Oxisols) can form in parent materials of volcanic origin.

6.3.2 Geographic Distribution

Andisols cover more than 120 million ha or nearly 1% of the Earth's surface. By far, the largest concentration is to be found along the Pacific Ring of Fire, a zone of tectonic activity and volcanoes stretching from South through Central and North America via the Aleutian Islands to the Kamchatka Peninsula of Russia through Japan, the Philippines and Indonesia to Papua New Guinea and New Zealand. Other areas include the Caribbean, central Atlantic ridge, northern Atlantic rift (Iceland) and the Mediterranean (Arnold, 1988). The largest area of Andisols in the United States occurs in Alaska (10 million ha) and in the Pacific Northwest (WA, OR, ID, CA) (Péwé, 1975; Rieger et al., 1979; Southard and Southard, 1991; Ugolini and Dahlgren, 1991; Goldin et al., 1992; Takahashi et al., 1993; McDaniel et al., 1993). Worldwide distribution of Andisols is depicted in Fig. 6.13. Volcanoes constructed during the period of intense tectonic activity in the Holocene, Pleistocene and later Tertiary periods produced about 1,000 m³ of lava and associated ejecta from which Andisols have developed (Leamy, 1984).

Andisols can occur in all soil moisture and temperature regimes (Dudal and Suparphardjo, 1961). In the Pacific Northwest, very few Andisols occur in the thermic temperature regime because the summers are too hot and dry to allow sufficient weathering to produce the required allophane and oxalate extractable Fe and Al for placement in the order. In other parts of the world, Andisols are found in thermic and even hyperthermic temperature regimes but under wetter moisture regimes (udic and

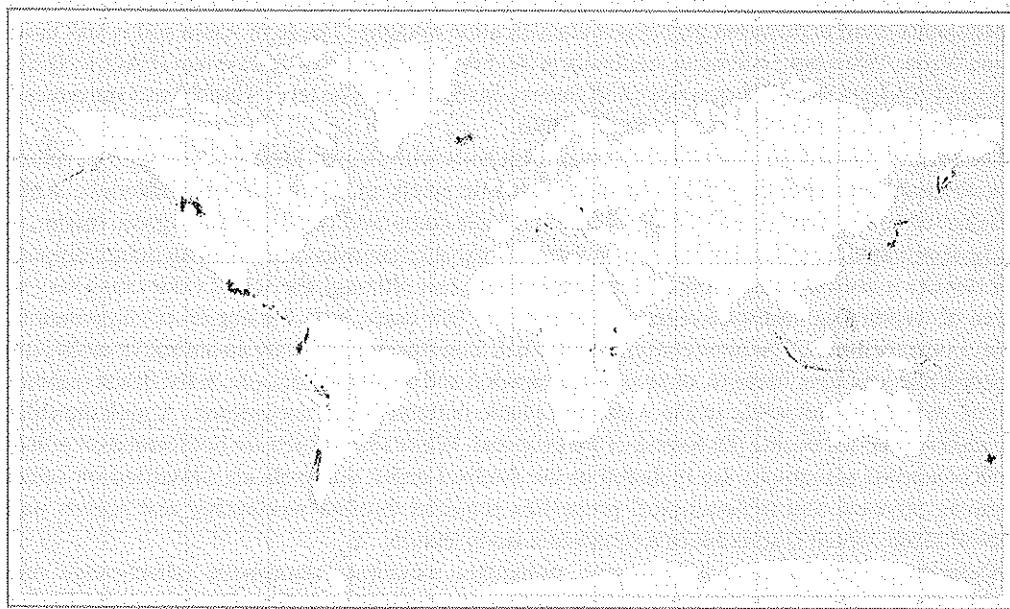


Fig. 6.13 Global distribution of Andisols [Courtesy of USDA-NRCS, Soil Survey Division, World Soil Resources, 1998]

perudic) (Derking, 1987). Because of their unique physical and chemical properties (Section 6.3.6), Andisols support a much greater population than their limited extent would suggest (Leamy, 1984). This is because they frequently occur in humid regions, which together with their excellent physical condition and high fertility made the establishment of nonshifting, early traditional agriculture possible.

Soils which classify as Andisols are also found in areas not associated with volcanic activity such as in the southern Appalachian mountains, Kyushu (Japan) and the Alps. These soils have high levels of Al associated with organic matter and similar management constraints as soils formed from volcanic ejecta, and consequently, key out as Andisols. This further highlights the importance of realizing that Andisols are not classified on parent material, but on the properties acquired during weathering.

6.3.3 Formation of Andisols

6.3.3.1 Parent Material

In most cases, the parent materials from which Andisols have formed are of volcanic origin which accounts for their close association with areas of volcanic activity. Parent materials vary from ultrafine dusts to lava flows, any of which may contain large boulders and volcanic bombs ejected by explosive eruptions. Because sequential eruptions laid down a variety of materials (ash, pumice, boulders), many Andisols are layered which allows chronological dating of the soil mass based on the known or datable age of the various layers. Andisol properties are affected by the chemical, mineralogical, textural and depositional properties of the volcanic ash which can be divided into dark and light colored varieties depending on contaminants from other rock types (Shoji, 1986). Light colored ash is acidic (andesitic and rhyolitic) containing quartz, and smaller amounts of hornblende and biotite, while dark colored ash contains mafic rich minerals, mainly olivine. Intermediate color groups are richer in plagioclases and pyroxenes. Opaque minerals are present in all three groups. Soil formation takes place faster in dark colored glassy (basaltic) than light colored glassy tephra (Shoji et al., 1993). In southcentral Alaska and the Aleutian Islands, and southeast Alaska, most Andisols are derived from light and darker colored tephra, respectively (Riehle, 1985; Ping et al., 1989), with the former being more acidic than the latter. The Vitrand suborder is the only one in which the vitric parent material plays a dominant role in the classification (Soil Survey Staff, 1998). In Iceland, Andisols form in mixtures of volcanic ash and loessial deposits with distinct layers of tephra that permit dating by radiocarbon and from historical records (Arnalds et al., 1995). These soils are highly subject to wind and water erosion when surface cover is destabilized. Accretions of volcanic ash from large cataclysmic events lead to buried soils.

6.3.3.2 Climate

Volcanic ejecta occur in all moisture and temperature regimes; consequently, the climates under which Andisols are found are very varied. Climate, which strongly controls Andisol formation, is recognized at all suborder levels except Vitrand. Combinations of temperature and moisture result in the different great groups of Andisols. Because Torrands form in arid climates, they are the least weathered and exhibit calcite accumulation in the soil profile. Ustands form in semi-arid climates and experience incomplete leaching, particularly of silica, which is translocated from the upper to lower profile and forms a duripan in some cases. Xerands, which form in Mediterranean climates with dry hot summers and cool, wet winters, have greater accumulation of organic matter than in the above less humid environments. Udands which form under humid or very humid climates are highly leached, very acidic and have high organic matter contents. Cryands which form in climates with mean annual temperatures

<8 °C with cool summers, frequently occur in Alaska, New Zealand, Iceland, and mountainous regions of the Pacific Ring of Fire countries. Aquands which occur in localized wet environments commonly have organic surface horizons and evidence of soil reduction independent of the broad climatic bias of other suborders.

To be classified as an Andisol, sufficient weathering must have taken place to produce specific quantities of acid oxalate extractable Fe and Al (Soil Survey Staff, 1998). Thus, in very dry environments, one would not expect to find Andisols. Parfitt (1990) suggested that a minimum precipitation of 1,200 mm yr⁻¹ would be required to bring about this weathering. However, in Iceland, Cryands form under as little as 700 mm yr⁻¹ (Arnalds et al., 1995). Because the occurrence of Torrands is limited, the question arises as to whether they really exist. Do they occur in seep or other areas where moisture is concentrated by other means? Because of their limited extent, this may well be the case.

The classic Andisols of Japan, New Zealand and other areas along the Pacific Ring of Fire are formed in humid environments which also tend to be fairly cool (Wada, 1986; Parfitt, 1990). They are deep organic rich soils which suggests that the preferred climate for Andisol formation is one with cool temperatures and high rainfall. Even in the tropics, the better expressed Andisols are found in isomesic temperature regimes, but that does not mean that they do not occur in warmer areas. However, more rapid weathering in the warmer areas leads to the rapid transformation to Oxisols, Ultisols, and Inceptisols.

6.3.3.3 Topography

Although Andisols are found on all types of topography at a wide range of elevations (0 to > 3,000 m), they tend to occur most frequently on mountainous to hilly terrain on the slopes of volcanoes. On the other hand, fine ash can be carried long distances and deposited on any type of landscape, while water can rework ash, both of which result in extensive flat areas of Andisols in intermountain valleys. In addition, large lava flows may result in fairly flat areas on which Andisol formation can take place. Because of the different modes of deposition of volcanic materials, no particular type of topography favors Andisol formation. However, Aquands which form in areas with restricted drainage mainly occur at low points in the landscape. These soils tend to be dark colored because of the accumulation of organic matter which forms in these anaerobic conditions.

6.3.3.4 Vegetation

Because Andisols occur in all moisture and temperature regimes on different landscapes at different elevations, vegetation is accordingly highly variable. In general, Andisols are moderately acid so vegetation associated with acid soils will predominate. Because of their relatively high water-holding capacities, Andisols, even in xerophytic areas, are likely to support more luxuriant vegetation than other nonandic soils in the same environment.

Although vegetation on Andisols is quite variable, it has a major influence on the type of Andisol formed according to Shoji and Otowa (1988). In the Pacific Northwest and Alaska, Udands generally form under forest vegetation. However, they are known to form under other vegetation types such as Japanese pampus grass (*Miscanthus sinensis*) (Melanoudands), Sitka spruce (*Picea sitchensis*) (Hydrudands) and blue joint grass (*Calamagrostis canadensis*) (Fulvudands). The organic matter is dominated by humic acids in the first cases and by fulvic acids in the last (Ping et al., 1989, 1995; Shoji, et al., 1993). Most Xerands, such as those in Washington and Oregon, form under mixed vegetation of grass and forest while Torrands form under shrubby or grass vegetation and Vitrandes under grass or forest. When the vegetation is grass as in Mollisols, a more humus-rich material develops. In Andisols, these soils are placed in great groups containing melanic (black) epipedons. In Iceland and other high latitude polar regions, Cryands form under moss, heath and grass covers (Arnalds et al.,

1995). In the tropics, the vegetation is often rain forest, particularly on the slopes of volcanoes where grassland and savanna types also occur.

6.3.3.5 Time

As with all soils, time is a major factor in the formation of Andisols. The time at which volcanic ash was deposited is regarded as time zero. Andisols generally are found on young parent materials in humid climates because, on older and more stable landscapes, weathering would have taken place for longer periods of time resulting in the transformation to other soil orders. The drier and cooler the environment, the longer it will take for an Andisol to form. Volcanic materials, either very young, in extremely dry areas, or that are recently deposited which have not weathered sufficiently to form at least the minimum requirements for andic materials, are placed in the Entisol order. According to Shoji et al. (1993), a minimum of several hundred years is required for an A horizon to form in newly deposited ash whereas for an A-B sequence, several hundred to thousands of years may be necessary. In Japan, most Andisols are 4,000–7,000 yr old whereas in SE Alaska, a Cryand was dated at 12,000 yr using ^{14}C dating. In Iceland, most of the Andisols range in age from a few hundred yr to over 10,000 yr (Arnalds et al., 1995). Episodic accumulation of volcanic and admixed eolian materials follows glacial retreat. With progressive age, A horizons formed in eolian-andic materials are transformed to a series of Bw cambic horizons, interspersed by coarse-textured tephra in marker layers (Fig. 6.14).

6.3.4 Pedogenic Processes in Andisols

Interactions of all the soil-forming factors (Jenny, 1941) are required for an Andisol to form. In general, the most common environment where Andisols form in a humid climate on young parent materials under forest vegetation. Similar processes are at work in the formation of Andisols and other soil orders, particularly Spodosols. The end products of weathering for both podzolization and andolization are very similar, namely, Fe and Al associated with organic matter with or without the presence of Si. Andisols are found on all types of landscapes and slopes.

Acid oxalate extracts Al and Fe from organic complexes and poorly crystalline oxides. Andisols contain high amounts of acid oxalate extractable Al (Al_o) and/or Fe (Fe_o). Spodosols are often high in Al_o and Fe_o as well. Pyrophosphate extractable Al (Al_p), or the amount of Al contained in organic complexes, is often proportionally higher in spodic than andic materials. Mizota and van Reeuwijk

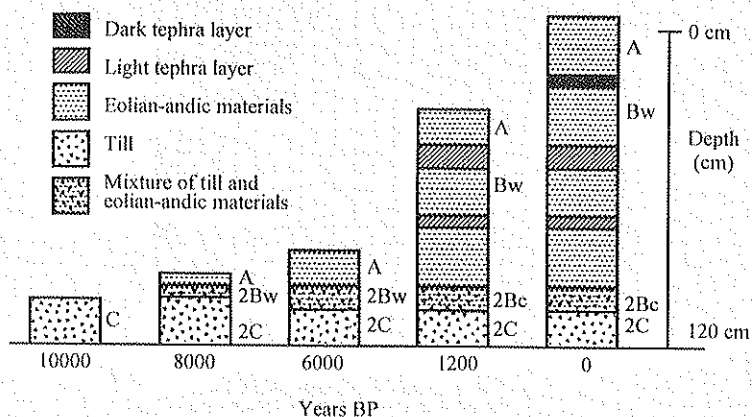


Fig. 6.14 Chronological development of Andisols in Iceland formed from tephra and eolian-andic materials [Reprinted from Arnalds et al., 1995. *Soil Sci. Soc. Am. J.* 59:161-169 with permission of the Soil Science Society of America]

(1989) have proposed a binary ratio (Al_p/Al_o) as a means of separating allophanic from nonallophanic soils. Soils with higher binary ratios are nonallophanic, are dominated by Al organic complexes, and would more likely be associated with spodic types of material. Morphologically, Spodosols often contain albic (bleached) horizons formed by the complexation and vertical translocation of metal organic complexes. Because this process typically does not occur in Andisols, they do not possess albic horizons.

The genetic concept of an Andisol is a soil formed in a humid environment on volcanic ash. A minimum amount of weathering to produce andic soil properties (Soil Survey Staff, 1996) is required. As most Andisols are formed in humid environments on easily weathered ash, weathering is quite rapid (Flach et al., 1980). Allophane which is a coprecipitate of Al, Fe and Si oxides is the first mineral to form from the weathering of volcanic glass together with organic complexes of Fe and Al (acid oxalate extractable). Under more intense weathering, halloysite and other crystalline minerals may form (Wada, 1989).

Because Andisols form under a wide range of environmental conditions, their properties vary quite widely. Common properties are low bulk densities, high levels of organic C to depth and high P fixation, all of which are used to define andic soil material. The overall processes lead to the formation of stable complexes of Al, Fe and Si with organic matter and the formation of allophane, ferrihydrite, and imogolite minerals (Southard and Southard, 1991; Ugolini and Dahlgren, 1991; Wada et al., 1992; Takahashi et al., 1993). If weathering continues, minerals become more crystalline leading to the formation of soils which would fall in other soil orders.

6.3.5 Classification of Andisols

Andisols are classified on the basis of andic soil properties which are defined by the Soil Survey Staff (1998) (Fig. 6.15):

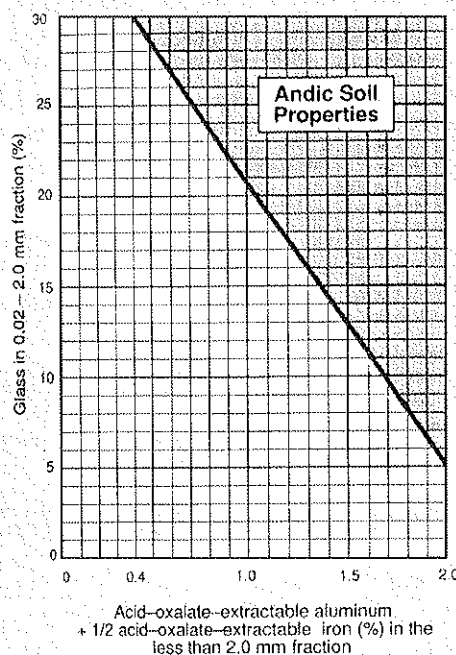


Fig. 6.15 Classification of andic versus nonandic materials as a function of glass content and ammonium oxalate extractable Al and Fe contents assuming 25% phosphate fixation [Soil Survey Staff, 1998]

Soil material must contain < 25% (by weight) of organic C and meet one or both of the following requirements in the fine earth fraction: (1) Al plus $\frac{1}{2}$ Fe (%) by ammonium oxalate $\geq 2.0\%$, a bulk density at 33 kPa $\geq 0.90 \text{ g cm}^{-3}$, and a P retention $\geq 85\%$, (2) a P retention $\geq 25\%$, $\geq 30\%$ of particles between 0.02 and 2 mm, and one of the following: (a) Al plus $\frac{1}{2}$ Fe (%) by ammonium oxalate $\geq 0.40\%$ and $\geq 30\%$ volcanic ash in the 0.02 to 2 mm fraction, or (b) Al plus $\frac{1}{2}$ Fe (%) by ammonium oxalate $\geq 2.0\%$ and $\geq 5\%$ volcanic glass in the 0.02 to 2 mm fraction, or (c) Al plus $\frac{1}{2}$ Fe (%) by ammonium oxalate between 0.40 and 2.0% and enough volcanic glass in the 0.02 to 2 mm fraction so that the glass (%) when plotted against the value obtained by adding Al plus $\frac{1}{2}$ Fe (%) falls within prescribed limits (Soil Survey Staff, 1998).

Table 6.16 Listing of suborders, great groups and subgroups in the Andisols order [Soil Survey Staff, 1998]

Suborder	Great Group	Subgroup
Aquaquands	Cryaquands	Lithic, Histic, Thaptic, Typic
	Placaquands	Lithic, Duric Histic, Duric, Histic, Thaptic, Typic
	Duraquands	Histic, Acraquoxic, Thaptic, Typic
	Vitraquands	Lithic, Duric, Histic, Thaptic, Typic,
	Melanaquands	Lithic, Acraquoxic, Hydric Pachic, Hydric, Pachic, Thaptic, Typic
	Epiaquands	Duric, Histic, Alic, Hydric, Thaptic, Typic
	Endoquands	Lithic, Duric, Histic, Alic, Hydric, Thaptic, Typic
Cryands	Duricryands	Aquic, Typic
	Hydrocryands	Lithic, Placic, Aquic, Thaptic, Typic
	Melanocryands	Lithic, Vitric, Typic
	Fulvicryands	Lithic, Pachic, Vitric, Typic
	Vitricryands	Lithic, Aquic, Oxyaquic, Thaptic, Humic Xeric, Xeric, Ultic, Alfic, Humic, Typic
	Haplocryands	Lithic, Alic, Aquic, Acrudoxic, Vitric, Thaptic, Xeric, Typic
Torrandis	Duritorrandis	Petrocalcic, Vitric, Typic
	Vitritorrandis	Lithic, Duric, Aquic, Calcic, Typic
	Haplotorrandis	Lithic, Duric, Calcic, Typic
Xerands	Vitrixerands	Lithic, Aquic, Thaptic, Alfic Humic, Ultic, Alfic, Humic, Typic
	Melanoxerands	Pachic, Typic
	Haploxerands	Lithic, Aquic, Thaptic, Calcic, Ultic, Alfic Humic, Alfic, Humic, Typic
Vitrandis	Ustivitrands	Lithic, Aquic, Thaptic, Calcic, Humic, Typic
	Udivitrands	Lithic, Aquic, Thaptic, Ultic, Alfic, Humic, Typic
Ustands	Durustands	Aquic, Thaptic, Humic, Typic
	Haplustands	Lithic, Aquic, Dystric Vitric, Vitric, Pachic, Thaptic, Calcic, Dystric, Oxic, Ultic, Alfic, Humic, Typic
Udands	Placudands	Lithic, Aquic, Acrudoxic, Hydric, Typic
	Durudands	Aquic, Acrudoxic, Hydric, Pachic, Typic
	Melanudands	Lithic, Anthraquic, Aquic, Acrudoxic Vitric, Acrudoxic Hydric, Acrudoxic, Pachic Vitric, Vitric, Hydric Pachic, Pachic, Hydric, Thaptic, Ultic, Eutric, Typic
	Hydrudands	Lithic, Aquic, Acrudoxic Thaptic, Acrudoxic, Thaptic, Eutric, Ultic, Typic
	Fulvudands	Eutric Lithic, Lithic, Aquic, Hydric, Acrudoxic, Ultic, Eutric Pachic, Eutric, Pachic, Thaptic, Typic
	Hapludands	Lithic, Anthraquic, Aquic Duric, Duric, Alic, Aquic, Acrudoxic Hydric, Acrudoxic Thaptic, Acrudoxic Ultic, Acrudoxic, Vitric, Hydric Thaptic, Hydric, Eutric Thaptic, Thaptic, Eutric, Oxic, Ultic, Alfic, Typic

Suborders are based on moisture or temperature regimes and the great groups and subgroups on other soil properties, which is evident from the names in Table 6.16. The melanic (dark humus rich soils generally formed under grass) and fulvic (brownier, generally formed under forest) great groups are the most common. Cryands, which are Andisols formed under cryic or pergelic soil temperature regimes, were recently added to the Andisol order. The classification is based on low bulk density, high P fixation and the presence of amorphous minerals (allophane and imogolite).

Under specified conditions where andic materials have influenced surface horizons of other soil orders, but where they are insufficiently thick or do not meet the andic requirements, these soils are placed in andic subgroups of other orders.

6.3.6 Andisol Properties

6.3.6.1 Morphological Features

Most Andisols have striking morphological features. They usually have multiple sequences of horizons (Fig. 6.16) as a result of intermittent volcanic fallout (Ping et al., 1988). In the field, one can easily identify dark A horizons overlying reddish brown or dark yellowish brown Bw cambic horizons with one or more buried A–Bw sequences below (Tables 6.17 and 6.19). Quite often, a thin gray or whitish layer of recent ashfall may be found at or near the surface as in the El Sueño series (Bw horizon). Because of the nature of the ashfall, the horizons are generally parallel to the surface and the transition from one horizon to the next is distinct. Such tephra or ash layers provide good markers of historical or datable volcanic ejection chronology.

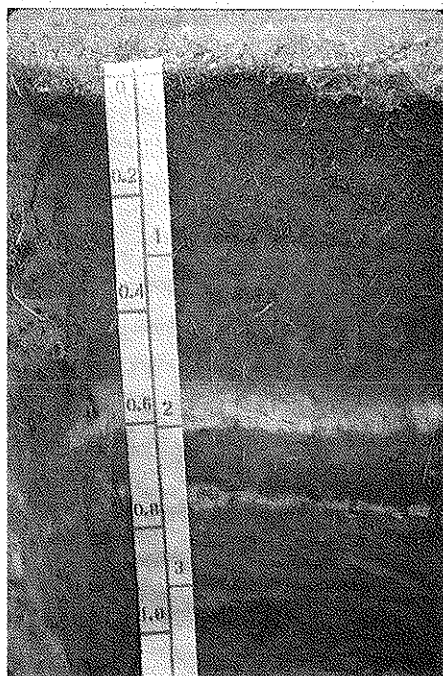


Fig. 6.16 Typical profile of a Cryand from Iceland showing a multiple stratified horizon sequence [Photo courtesy of L.P. Wilking]

Table 6.17 Selected chemical properties of some Andisols from Guatemala [Sumner et al., 1992]

Soil horizon	Depth cm	pH			Organic matter %	ECEC	CEC pH 7	AEC pH 4	Exchangeable cations					P retention %
		H ₂ O	M KCl	M NaF					Ca	Mg	Na	K	Al	
		-----cmol _c kg ⁻¹ -----												
medial, isothermic Acrudoxic Hapludand: Las Cruces series (Guatemala)														
A1	0-30	4.7	4.6	11.5	7.2	2.2	26.8	1.2	0.55	0.09	0.34	0.06	1.20	90
Bw1	30-60	4.6	4.8	10.9	2.6	1.3	29.5	1.6	0.50	0.08	0.37	0.07	0.30	95
Bw2	60-92	4.8	5.1	11.5	1.3	1.4	28.4	2.0	0.50	0.12	0.35	0.03	0.40	96
Bw3	92-185	4.3	5.0	11.4	0.9	1.3	26.9	2.2	0.50	0.03	0.26	0.03	0.50	96
ashy over medial, isothermic Thaptic Udivitrand: El Sueño series (Guatemala)														
A	0-16	5.0	4.4	10.8	4.0	3.3	13.1	0.4	1.52	0.09	0.29	0.13	1.30	88
Bw	16-75	5.7	4.9	10.7	0.2	0.8	2.2	nd	0.50	0.03	0.22	0.06	0.00	68
2Ab	75-102	5.7	5.1	11.7	6.5	4.7	39.7	nd	3.70	0.16	0.38	0.43	0.00	98
2Bwb	102-163	5.8	5.3	11.3	2.8	4.7	39.8	nd	3.67	0.21	0.37	0.41	0.00	94
2C	163-170	5.6	5.4	11.4	1.7	2.1	30.3	nd	1.33	0.09	0.52	0.11	0.00	60
medial-skeletal, isothermic Typic Haplustand: Tecuamburro series (Guatemala)														
A	0-15	6.0	5.1	10.2	5.8	17.5	39.8	0.0	15.17	1.57	0.45	0.35	0.00	92
Bw1	15-47	5.6	4.8	10.1	0.8	15.9	35.4	nd	13.60	1.80	0.38	0.07	0.00	90
Bw2	47-112	5.6	0.6	10.2	0.6	17.8	36.0	nd	15.17	1.86	0.65	0.07	0.00	95
Bw3	112-200	5.8	4.8	10.0	0.4	14.8	32.6	nd	12.15	2.17	0.47	0.03	0.00	95

In the field when excavated, Andisols give a feeling of puffiness due to the low bulk density which allows roots to penetrate to great depths. Andisols generally have granular structure in the A horizon due to perturbations caused by roots and soil fauna. Structure in the Bw horizon is generally weak subangular blocky. Some Udands formed in areas of high rainfall have higher clay contents while soils that are subjected to wet and dry cycles form columnar structure. Soils high in allophane are sticky under field moist conditions.

6.3.6.2 Mineralogical Properties

The mineralogy of Andisols is responsible for their unique physical and chemical properties. The major minerals are allophane, allophane-like minerals, imogolite and crystalline aluminosilicate clay minerals. All types of clay minerals can be found in Andisols, but those of greatest interest are allophane, imogolite and poorly crystalline Fe hydroxides (ferrhydrite) because they confer andic properties (Wada, 1986; Ping et al., 1988; Yoshinga, 1988; Shoji et al., 1993).

Volcanic ash consists of glass-containing aluminosilicates with minor amounts of ferromagnesian minerals. Upon weathering, ash releases Al and Fe into solution which then form stable complexes with organic matter derived from plant decomposition in the surface soil. These Al and Fe complexes, which are immobile, accumulate at the surface to form a dark or darkish brown A horizon. These complexes dominate the fine fraction. Normally in humid environments, dissolved Al is complexed with organic matter, but the soil solution may contain large amounts of dissolved Si which leads to the formation of secondary Si minerals (laminar opaline silica) from the saturated solution (Shoji and Masui, 1971; Ping et al., 1988). This material is circular or elliptical in shape (0.2–0.5 μm diam) and extremely thin (Fig. 6.17). Opaline silica is also found in the surface horizons of some Andisols (Aleutian Islands and Kenai Peninsula).

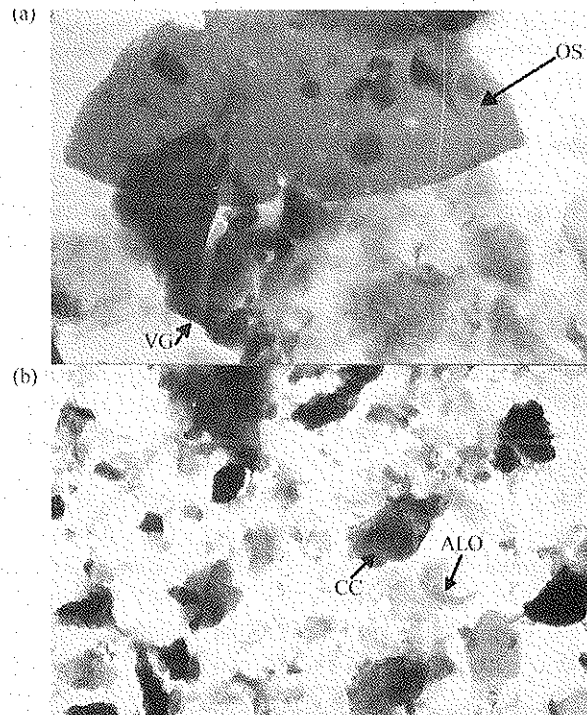


Fig. 6.17 Electron micrographs of clay minerals in Andisols from Alaska: (a) Laminar opaline silica (OS) in the A horizon of a Melanocryand, (b) allophane (ALO) and crystalline clay minerals (CC) in the Bs horizon of a Fulvucriand

As depth to Bw horizon increases, the supply of organic matter becomes more limited, and as a result, the dissolved Al and Si from volcanic glass weathering precipitate to form non- and para-crystalline minerals, mainly allophane and imogolite. Allophane has been used as a synonym for amorphous clay because it is nearly X-ray amorphous. However, under the electron microscope, it appears as hollow spheres (3.55 nm diam) which have the chemical composition $(1-2)\text{SiO}_2 \cdot \text{Al}_2\text{O}_3 \cdot (2.5-3.0)\text{H}_2\text{O}$ (Wada, 1989). Imogolite is para-crystalline and gives characteristic X-ray diffraction, IR and differential thermal analysis patterns. Under the electron microscope, it appears as smooth and curved threads or tubes with inner and outer diameters of 1.1 and 2 nm, respectively. Chemically it has the composition $(\text{OH})\text{SiO}_3 \cdot \text{Al}_2(\text{OH})_3$ (Wada, 1989). Allophane content increases with depth in the profile, usually being concentrated in the Bw and buried horizons; across the landscape, it increases with increasing age of the volcanic deposit and annual precipitation. Allophane and imogolite are soluble in acid oxalate solution and the Si dissolved is used to estimate their contents in soils (Parfitt and Henmi, 1982). Allophane in B horizons ranges from 2–12, 4–14 and 10–20% in southcentral Alaska, the Aleutian Islands and Baranof Island, respectively (Ping et al., 1989).

In contrast, the soils from Guatemala (Table 6.18) can contain halloysite, gibbsite, kaolinite, vermiculite and cristobalite in addition to amorphous allophanic materials. When the pH_{sum} is above 9.5, Fieldes and Perrott (1966) showed that it was diagnostic for the presence of allophane. In all pedons from Guatemala, this was the case which confirms the presence of this mineral. All pedons contained substantial amounts of free Fe oxides.

Ferrihydrite ($5\text{Fe}_2\text{O}_3 \cdot 9\text{H}_2\text{O}$) (Schwertmann and Taylor, 1989), also an amorphous mineral (short-range order) found in Andisols under high rainfall or in soils with restricted drainage, is a widespread

Table 6.18 Selected physical and mineralogical properties of some Andisols from Guatemala [Sumner et al., 1992]

Soil horizon	Depth cm	Clay content	Gravel content	Water content		Bulk density g cm ⁻³	Surface area m ² g ⁻¹	Free Fe ₂ O ₃ %	Mineralogy [†]
				30 kPa	1500 kPa				
				% (w/w)					
medial, isothermic Acrudoxic Hapludand: Las Cruces series (Guatemala)									
A1	0-30	7.6	0	53.0	29.0	0.56	51	3.64	AM,HA,GI,CT
Bw1	30-60	6.0	0	56.9	40.5	0.53	217	5.19	AM,HA,GI,CT
Bw2	60-92	6.4	0	60.7	41.3		244	4.44	
Bw3	92-185	6.9	0	64.6	43.6	0.62	250	3.51	AM,HA,GI,CT
ashy over medial, isothermic Thaptic Udivitrand: El Sueño series (Guatemala)									
A	0-16	5.8	2	38.7	10.8	0.86	7	0.88	AM,HA,VR,CT
Bw	16-75	2.8	4	10.0	2.9	0.92	1	0.23	AM,HA,VR,CT
2Ab	75-102	22.4	0	79.1	33.6		69	3.76	
2Bwb	102-163	23.0	0	71.5	34.6	0.53	181	3.89	
2C	163-170	26.6	0	95.2	41.4		179	2.38	AM,CT
medial-skeletal, isothermic Typic Haplustand: Tecuamburro series (Guatemala)									
A	0-15	23.8	14	44.4	27.8	0.58	44	0.82	HA,KK,AM,CT
Bw1	15-47	18.5	34	47.8	29.1		71	3.59	HA,KK,AM,CT
Bw2	47-112	18.0	32	41.1	28.8		71	4.14	
Bw2	112-200	24.5	31	49.3	32.1		80	5.34	HA,KK,AM,CT

[†] AM = amorphous, HA = halloysite, GI = gibbsite, CT = cristobalite, VR = vermiculite, KK = kaolinite

and characteristic component of young Fe-oxide accumulations precipitated from Fe-rich solutions. Ferrihydrite content can be estimated from oxalate or dithionite extractable Fe (Parfitt, 1983). In B horizons, allophane with or without ferrihydrite dominates the colloidal fraction with Al and Fe humus complexes as minor components. The lower B and C horizons can be rich in allophane and ferrihydrite but usually contain only small amounts or no Al and Fe humus complexes (Ping et al., 1989).

In the silt and sand fractions of Andisols, the dominant components are volcanic glass and other primary minerals. The glass particles (shards) which, like shattered glass, have sharp angles and edges, are very abrasive. However, in soils, the glass particles are usually coated with amorphous materials including allophane, Al and Fe oxides and their humus complexes which all contribute to aggregate formation. The oxide and organic matter coatings on glass or other mineral surfaces can be easily removed using HCl and NaOH solutions, respectively, prior to the measurement of glass content.

Mineralogical characteristics of Andisols have been reviewed by Wada (1980) and a detailed discussion of short-range order minerals is presented in Section F, Chapter 4.

6.3.6.3 Chemical Properties

Most Andisols are acid to very acid (Tables 6.17 and 6.19) and many contain elevated levels of exchangeable Al and low amounts of exchangeable basic cations. All pedons presented have high P-retention capacities due to the presence of allophane and Fe oxyhydroxides (Parfitt, 1989). Organic matter contents are high relative to nonandic soils and are concentrated in materials which have been weathered. Note the very low organic matter content in a recent ash layer (Bw) in the El Sueño series (Table 6.17).

Variable charge arising from the amorphous minerals and Fe and Al organic complexes is high in most Andisols as illustrated in Table 6.17 by the large difference between ECEC and CEC values. In

Table 6.19 Selected chemical and physical properties of some Andisols from Alaska

Soil horizon	Depth cm	pH H ₂ O	pH KCl	pH NaF	Org. C %	CEC pH 7 cmol _c kg ⁻¹	Exch. Al ¹ kg ⁻¹	Base sat. %	P reten. %	Bulk density g cm ⁻³	Allophane content %	1,500 kPa water %	Volcanic glass ³	USDA texture
Island series (Melanocryand)														
A1	0-23	5.2			9.1	37	3.1	7	90	0.65	0	30	36	sil
A2	23-40	5.6			6.1	33	1.3	10	94	0.73	2.1	27	34	vfs1
A3	40-60	6.0			3.5	13	0.1	16	82	0.64	3.6	19	34	sil
Bw	60-101	6.0			2.2	13	0.3	15	81	0.71	4.0	18	20	sil
C	101-152	5.9			0.2	5	0.6	15	20	1.00	0.5	4	2	lfs
Kachemak series (Fluvucryand)														
Bs	0-8	4.5			9.5	39	7.1	15	94	0.63	0	23	90	sil
Ab	8-20	4.8			12.4	42	7.6	10	96	0.69	0	37	86	sil
Bhs	20-30	4.9			7.3	45	7.5	5	97	0.67	1.5	36	50	sil
Bw	30-56	5.2			3.0	23	3.8	5	95	0.77	3.0	24	25	sil
2C	56-114	5.3			0.9	17	4.5	30	70	1.21	1.0	14	5	sil
Kashwitna series (Haplocryand)														
E	0-3	4.3			6.9	35	6.0	17	54	0.63	0	16	51	sil
Bs	3-20	5.5			4.7	27	1.9	6	94	0.74	4.2	20	57	sil
Eb	20-25	5.6			3.2	11	0.7	5	55	---	2.8	---	---	sil
Bsb	20-30	5.8			1.4	14	0.8	8	66	---	4.3	11	14	sil
BC	30-35	5.6			0.8	8	0.2	9	50	1.03	2.8	7	13	sil
2BC	35-46	5.9			0.2	4	---	6	34	---	0.7	3	2	gvlq
2C	46-102	5.7			---	2	---	24	16	---	0	2	2	gxs

¹Exchangeable Al extracted with M KCl²On field moist samples, wt%³In very fine sand fraction

fact, one pedon (Las Cruces series) contains substantial AEC and is poised near its point of zero charge (p.z.c.) which can be inferred from the higher pH_{KCl} than pH_{water} values. For Papua New Guinea Andisols (Hydrudands), Radcliffe and Gillman (1984) demonstrated that the p.z.c. decreased with increasing organic C and increased with increasing Fe₂O₃ content. Parfitt (1988) found that ECECs could range from 1–8 cmol_c kg⁻¹ and CECs (pH 7) from 20–40 cmol_c kg⁻¹ for allophanic A horizons. Shoji et al. (1993) demonstrated that AEC was detected in allophanic but not in nonallophanic soils.

A review of the chemical properties of variable charge soils is presented by Parfitt (1980) and the chemistry of variable charge surfaces is presented in Section B, Chapter 7.

6.3.6.4 Physical Properties

Unique physical attributes of Andisols are related to compound structural assemblages of hollow spheres and tubular threads as mineral entities into progressively larger aggregated domains. Hence, close packing of spherules and threads from 1–2 nm basic units to 0.1–1 µm microaggregated domains to aggregates of several mm in size results in low density, high porosity, high surface area, high soil water retention even at low water potentials, and the disparity in water movement from within basic units (intraaggregate pores) to interparticle porosity associated with aggregated assemblages of

different sizes. This phenomenon also accounts for the low thermal conductivity of andic materials which is 3 to 4 times less than that of the layer lattice phyllosilicates in other mineral soils. It accounts for the thixotropic character of these soils (discussed later), and the irreversible changes in bulk density (increased), water retention (decreased) and cohesive strength (increased) upon drying andic materials.

Andisols have low bulk density, usually $< 0.9 \text{ g cm}^{-3}$ due to the high organic matter and amorphous mineral contents. This makes them excellent rooting media but on the other hand, they have low bearing capacity, are highly susceptible to wind and water erosion when surface cover is removed or degraded, and present problems from an engineering standpoint. Because of the nature of ejecta, many contain appreciable amounts of gravel and stones.

Most Andisols have a smeary feeling when rubbed between the fingers when field moist. Even moist soil after prolonged rubbing can become greasy and wet. However, once the pressure is released, the excess water is reabsorbed by the soil. This property is called thixotrophy and is due to their unusually high water retention capacities (Tables 6.18 and 6.19). This high water retention is due to the allophane present which has a very high surface area ($\sim 1,000 \text{ m}^2 \text{ g}^{-1}$) due to its hollow sphere shape (Wada, 1989). Exchange of water between micropores associated with these spheres and macropores is incomplete and very slow. Field capacity moisture contents of many Andisols can exceed 100%, particularly where weathering has been intense. The Guatemalan Andisols all exhibited high surface areas in weathered horizons (Table 6.18).

Many Andisols exhibit irreversible changes on drying. On dehydration, allophane aggregates to silt or larger size particles which do not breakdown on resetting. This phenomenon can also lead to crust formation at the soil surface during hot dry periods. This also results in particle size analyses that are commonly unreliable because clay content is underestimated and sand and silt contents overestimated (Ping et al., 1988). This is well illustrated in Table 6.18 for the Las Cruces pedon where measured clay content was $\sim 6\%$ but the surface area was $50\text{--}250 \text{ m}^2 \text{ g}^{-1}$, clearly an incongruity.

Plasticity is a physical property which distinguished allophanic from all other soils containing crystalline clay minerals. Generally, field moist Andisols have high liquid (60–350%) and plastic limits (70–180%) (Warkentin and Maeda, 1980). Plasticity measurements can be used as an index of physical behavior in Andisols as a substitute for grain size distribution which is unreliable. Air-dry samples, on the other hand, show little plasticity because of the irreversible changes on drying and behave like sandy soils. The physical and mechanical characteristics of Andisols have been discussed in detail by Warkentin and Maeda (1980).

6.3.7 Management of Andisols

In many parts of the world, crop productivity on Andisols is very high, leading to population concentrations such as on Java where the soils are particularly fertile (Sutanto et al., 1988) and in Japan, the Andean portion of South America, and also Central America. Fresh deposits of ash resupply nutrients to many Andisols as they weather, thus maintaining fertility. Consequently, few inputs with the exception of P have been required for successful agricultural production on many Andisols.

However, Andisols are all acid and where acidity is particularly high, addition of lime may be required to overcome Al toxicity or alternatively acid-tolerant crops must be selected (Michaelson and Ping, 1987). Where Andisols have been acidified by N inputs, serious consequences for crop production can arise because as pH decreases, CEC declines towards the p.z.c. As a result, such soils are able to hold few basic cations for plant uptake (Sumner and Hylton, 1993). Because of their high P-fixing capacities, most Andisols require inputs of P for successful production (Michaelson and Ping, 1990). Such inputs in the form of rock phosphate are effective on very acid Andisols but in any event,

soluble P sources should be band placed or otherwise protected from intimate mixing with the soil (Sumner and Hylton, 1993). Requirements for N vary according to the ease with which organic matter is mineralized in these soils. Liming often promotes the release of substantial amounts of N from non-available reserves. Other nutrients may sometimes be deficient in Andisols depending of the chemical composition of the parent material.

Andisols, in general, are noted for their good physical properties (high water-holding capacity, good tilth, resistance to water erosion and stable aggregation). Consequently, they are excellent media for root growth and proliferation, but under heavy grazing can become compacted (Schlichting, 1988). In high latitudes, Cryands may be overgrazed, denuded, and subject to wind erosion (Arnalds, et al., 1995). Likewise, Vitrandis have high potential to be eroded by wind and have low nutrient and water holding capacities due to their sandy nature. In areas where volcanoes are or have been active, stratified ash deposits can cause textural discontinuities resulting in problems with water transmission within the profile, root growth and poor tilth. Deep plowing can often remedy these situations.

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6.4 Entisols

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6.4.1 Introduction

Entisols are mineral soils that exhibit minimal pedogenic alteration (Soil Survey Staff, 1975, 1998). The ability to support plants distinguishes Entisols from areas that are not considered soil, such as rock outcrops or some highly disturbed lands. Entisols typically have AC, ACr, or AR profiles and inherit many physical and chemical properties from their parent material. The A horizon can be classified as any taxonomic epipedon, except mollic. Subsoil horizon symbols may be used when a pedogenic pathway is inferred, but the properties of the horizon must not meet the minimum requirements of any subsurface diagnostic horizon.

Entisols form in areas where pedogenesis has been restricted or retarded. Examples include landscapes with recent deposition or erosion, resistant bedrock, inert parent material, saturation, or human activity (Buol et al., 1980). The world distribution of soil orders reveals where many of these conditions exist (Fig. 6.18). The most common areas where Entisols form are mountains, sand dunes, flood plains, coastal plains, and urban areas. Entisols comprise approximately 16% of the world distribution of soils, and are the most widespread of all orders (Soil Survey Staff, 1997).

In contrast to most other soil orders, suborders of Entisols are not defined by precipitation or temperature regime. Rather, Entisol suborders are designed to differentiate the conditions under which subsurface horizon development is impeded (Smith, 1986). The five Entisol suborders are Orthents, Psammments, Fluvents, Aquents, and Arenets (Figs. 6.19 and 6.20).

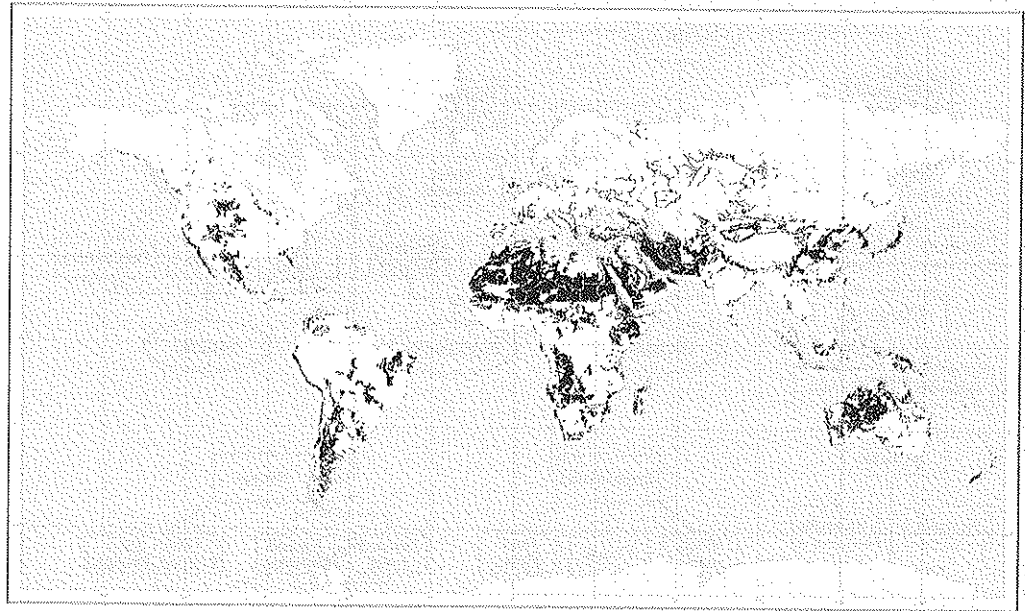


Fig. 6.18 Global distribution of Entisols [Courtesy of USDA-NRCS, Soil Survey, Division, World Soil Resources, 1998]

Parent material lithology is one factor that may inhibit pedogenesis. Resistant bedrock increases runoff and erosion, and decreases weathering. These processes promote the formation of the suborder Orthents, particularly in mountain regions (Figs. 6.19 and 6.20). Orthents may also form in sandy parent materials that have an abundance of rock fragments. When parent materials consist of quartzitic sands without coarse fragments, such as in many sand dunes, profile development is also retarded. Here, the suborder Psamments may form (Figs. 6.19 and 6.20). Both Orthents and Psamments typically have ochric epipedons.

Fluvents may develop in flood plains, fans, deltas, or mudflows when sedimentation rates exceed pedogenic rates (Section E, Chapter 1). Rapid deposition limits the effects of pedogenesis and subsurface horizon development. These soils typically have ochric epipedons (Figs. 6.19 and 6.20).

Another condition that slows pedogenesis is anaerobiosis. Persistent high watertables may prevent leaching of soil constituents, inhibit redox cycles, and preclude subsurface horizon development. These soils make up the Aquents suborder (Figs. 6.19 and 6.20). The geographical distribution of these soils may include poorly drained flood plains, hillslope seeps, and coastal marshes. Ochric or histic epipedons may form in these areas.

Lastly, Arenets and other disturbed soils are becoming more common with intensified human activity. These soils typically form in mine spoil, dredge spoil, and in other recently deposited urban or industrially related materials (Fig. 6.19). These soils may have ochric or anthropic epipedons.

6.4.1.1 Engineering and Land Use

Many engineering and land use problems arise in areas where Entisols form (Buol et al., 1980; Grossman, 1983). Erosion may occur with Orthents, rapid sedimentation with Fluvents, wind erosion or deposition with Psamments, unpredictable behavior of physical and chemical properties with Arenets, and poor drainage with Aquents. In contrast, some of the most productive agricultural soils in the world occur in flood plains. Flooding provides much needed nutrients and a water supply for crop

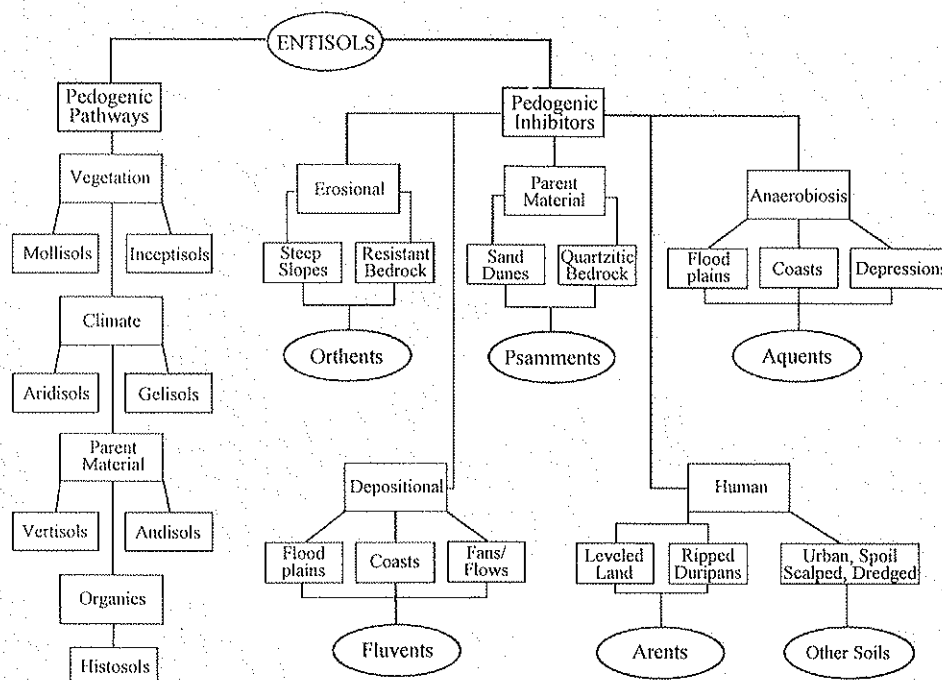


Fig. 6.19 Flow diagram illustrating (a) pedogenic pathways of soil orders from Entisols; and (b) landscape factors used to differentiate the suborders of Entisols (not included for Orthents are sandy parent materials with more than 35% rock fragments.)

production in areas where fertilization and irrigation practices are not used. Further, some of the most populated regions of the world occur in flood plains.

6.4.1.2 Genesis

The influence of the soil-forming factors (Jenny, 1941) on the development of Entisols is limited. Properties of Entisols may not reflect climate conditions because of recent deposition, erosion, or human disturbance. The primary role of biota is that of initiating the physical and chemical weathering process of freshly exposed substrate. Topography influences drainage and can contribute to the rapid formation of wet Entisols, or induce erosion in steeply sloping areas. Many of these factors also reduce the passive effects of time on soil development. Regardless of the influence of the other soil-forming factors, the properties of Entisols are largely inherited from the parent material.

Melanization (Buol et al., 1980) is the principal soil-forming process that initiates the development of Entisols. This process contributes organic matter to the parent material, promotes structural development, and may be associated with minor leaching of soluble constituents. Grossman (1983) presents data showing that bulk densities in alluvial settings initially increase in Entisols because of frequent wetting and drying cycles, which lead to compaction. Eventually, however, melanization will decrease bulk densities in surface horizons as organic matter inputs increase. In resistant bedrock, melanization rapidly decreases bulk densities as rock material is weathered and replaced by organic material or pore space. The process of melanization in the formation of Entisols can proceed as long as mollic epipedons or subsurface diagnostic horizons do not form.

6.4.1.3 Pedogenic Pathways

Entisols serve as the precursor to all other soil orders during their evolutionary history (Fig. 6.19) (Soil Survey Staff, 1998, 1975). In grassland ecosystems, Mollisols may form when steady state has been attained in the accumulation of organic matter. In forest settings, ochric epipedons and Inceptisols may form. Entisols develop into Aridisols outside of flood plains in areas with arid moisture regimes. Vertisols and Andisols are influenced more by parent material characteristics than any other soil order except Entisols. Vertisols form from Entisols in high shrink/swell parent materials where wet/dry cycles are prevalent. Andisols commonly form from Entisols in parent materials with large quantities of volcanic ash in cool and humid climates. Entisols will develop into Histosols with the accumulation of significant amounts of organic matter in association with high water tables and anaerobic conditions. Lastly, the new Gelisol order now encompasses areas formerly mapped as Entisols in cold regions. A listing of great groups and subgroups in the Entisol order is presented in Table 6.20.

6.4.2 Orthents

Orthents are loamy and clayey Entisols (better drained than Aquents), with a regular decrease in organic C with depth. They occupy approximately 10.5% of the Earth's land surface. They are more extensive than the combined area of the other Entisol suborders, which together occupy an area of about 5.7% (Soil Survey Staff, 1997). Orthents also have more taxonomic subdivisions, reflecting properties related to climate, depth to bedrock, mineralogy, wetness, fertility, shrink/swell, and bioturbation (Soil Survey Staff, 1998).

Taxonomically, Orthents commonly grade into Fluvents and Psamments. The rationale for separating Orthents from Fluvents was to maintain the old concept of flood plain soils (Fluvents), which have immense agricultural importance (Smith, 1986). The rationale for separating Orthents from Psamments was to identify soils (the Psamments) that are readily susceptible to wind erosion (Gile et al., 1981).

Although Orthents are most commonly associated with Psamments and Fluvents, they may also be related to Arenents. For example, where soils have been badly gullied, such as in some loess deposits of the southern United States, Alfisols occur on narrow ridges and Orthents in eroded areas between the ridges. When the areas are reclaimed by leveling with bulldozers, Arenents are formed from Alfisols

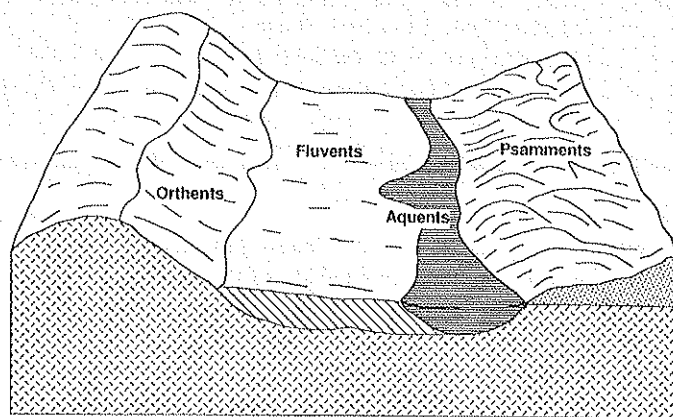


Fig. 6.20 Block diagram illustrating typical landscape positions where Entisols suborders form. Orthents on steep slopes associated with bedrock, Fluvents in floodplains, Aquents in poorly drained riparian conditions, and Psamments in sand dunes

Table 6.20 Listing of suborder, great groups and subgroups in the Entisol order [Soil Survey Staff, 1998]

Suborder	Great Group	Subgroup
Aquepts	Sulfaquepts	Haplic, Histic, Thapto-Histic, Typic
	Hydraquepts	Sulfic, Sodic, Thaptic-Histic, Typic
	Cryaquepts	Aquandic, Typic
	Psammaquepts	Lithic, Sodic, Spodic, Humaqueptic, Mollic, Typic
	Fluvaquepts	Sulfic, Vertic, Thapto-Histic, Aquandic, Aeris, Humaqueptic, Rhodic, Mollic, Typic
	Epiaquepts	Aeric, Humaqueptic, Mollic, Typic
	Endoaquepts	Sulfic, Lithic, Sodic, Aeris, Humaqueptic, Mollic, Typic
Arenis	Ustarens	Haplic
	Xerarens	Sodic, Duric, Alfic, Haplic
	Torriarens	Sodic, Duric, Haplic
	Udarens	Alfic, Ultic, Mollic, Haplic
Psamments	Cryopsamments	Lithic, Aquic, Oxyaquic, Vitrandic, Spodic, Lamellic, Typic
	Torripsamments	Lithic, Vitrandic, Haploduridic, Ustic, Xeric, Rhodic, Typic
	Quartzipsamments	Lithic, Aquodric, Aquic, Oxyaquic, Ustoxic, Udoxic, Plinthic, Lamellic Ustic, Lamellic, Ustic, Xeric, Spodic, Typic
	Ustipsamments	Lithic, Aquic, Oxyaquic, Aridic, Lamellic, Rhodic, Typic
	Xeropsamments	Lithic, Aquic Durinodic, Aquic, Oxyaquic, Vitrandic, Durinodic, Lamellic, Dystric, Typic
	Udipsamments	Lithic, Aquic, Oxyaquic, Spodic, Lamellic, Plaggeptic, Typic
Fluvents	Cryofluvents	Andic, Vitrandic, Aquic, Oxyaquic, Mollic, Typic
	Xerofluvents	Vertic, Aquandic, Andic, Vitrandic, Aquic, Oxyaquic, Durinodic, Mollic, Typic
	Ustifluvents	Aquertic, Torrtic, Vertic, Anthraquic, Aquic, Oxyaquic, Aridic, Udic, Mollic, Typic
	Torrifluvents	Ustertic, Vertic, Vitriaxerandic, Vitrandic, Aquic, Oxyaquic, Duric Xeric, Duric, Ustic, Xeric, Anthropic, Typic
	Udifluvents	Aquertic, Vertic, Andic, Vitrandic, Aquic, Oxyaquic, Mollic, Typic
Orthents	Cryorthents	Lithic, Vitrandic, Aquic, Oxyaquic, Lamellic, Typic
	Torriorthents	Lithic Ustic, Lithic Xeric, Lithic, Xerertic, Ustertic, Vertic, Vitrandic, Aquic, Oxyaquic, Duric, Ustic, Xeric, Typic
	Xerorthents	Lithic, Vitrandic, Aquic, Oxyaquic, Durinodic, Dystric, Typic
	Ustorthents	Aridic Lithic, Lithic, Torrtic, Vertic, Anthraquic, Aquic, Oxyaquic, Durinodic, Vitritorrandic, Vitrandic, Aridic, Udic, Vermic, Typic
	Udorthents	Lithic, Vitrandic, Aquic, Oxyaquic, Vermic, Typic

because fragments of argillic horizons are scattered in the smooth land left by the bulldozers (Smith, 1986). A variety of environmental conditions contribute to the absence of subsurface diagnostic horizons in Orthents. The most common factors are erosion, deposition, and resistant parent material.

6.4.2.1 Erosion and Deposition

Because steep slopes promote erosion, Orthents are common in mountainous terrains, especially where vegetation is sparse. An example of this landscape occurs in the Basin and Range area of southern New Mexico (Fig. 6.21). Mountain slopes in that area are characteristically composed of

intermixed Orthents and rock outcrops, where the Orthents are composed of A overlying R horizons (Gile and Grossman, 1979). As with steep mountain slopes, erosional scarps that separate pleistocene fan terraces (an alluvial fan deposited on a flood plain that was subsequently abandoned with channel downcutting) are also areas where Orthents form.

In some dissected arid areas, Orthents develop by episodic erosion. Here well-developed Aridisols with argillic and calcic subsurface diagnostic horizons are eroded to the extent that the parent material becomes surficially exposed. As a result, soil classification changes from Aridisols to Entisols (Argid to Calcid to Orthent).

On the scarps of Pleistocene fan terraces, another type of Orthent occurs not as the result of erosion, but as the result of deposition. Orthents in this area form in colluvial footslopes because the rate of sedimentation is greater than the rate of pedogenesis (Fig. 6.21). Consequently, no subsurface diagnostic horizons occur in this landscape position.

Orthents also commonly occupy fan terraces of Holocene age (Fig. 6.21). Although the rates of sedimentation and erosion are low, soils are too young to have yet formed subsurface diagnostic horizons. In contrast, topographically higher fan terraces of Pleistocene age have developed calcic or petrocalcic horizons, and thus classify as Aridisols (Calcids).

Many arroyo channels in southern New Mexico are also areas of Orthents (Fig. 6.21). Although these channels ephemerally carry runoff water and are in a nearly level flood plain position, they typically do not contain organic matter irregularly distributed with depth. This is because arid climates are not conducive to organic matter production or preservation. Therefore, instead of Fluvents that are common on the larger flood plains with frequent flooding, such as along the Rio Grande, the smaller arroyo flood plains have Orthents (Gile et al., 1981).

6.4.2.2 Resistant Parent Material

Exceptional resistance to weathering of some parent materials, such as quartzite-rich bedrock, prolongs the period of undistinguished horization (Buol et al., 1980). A comparison of soils formed

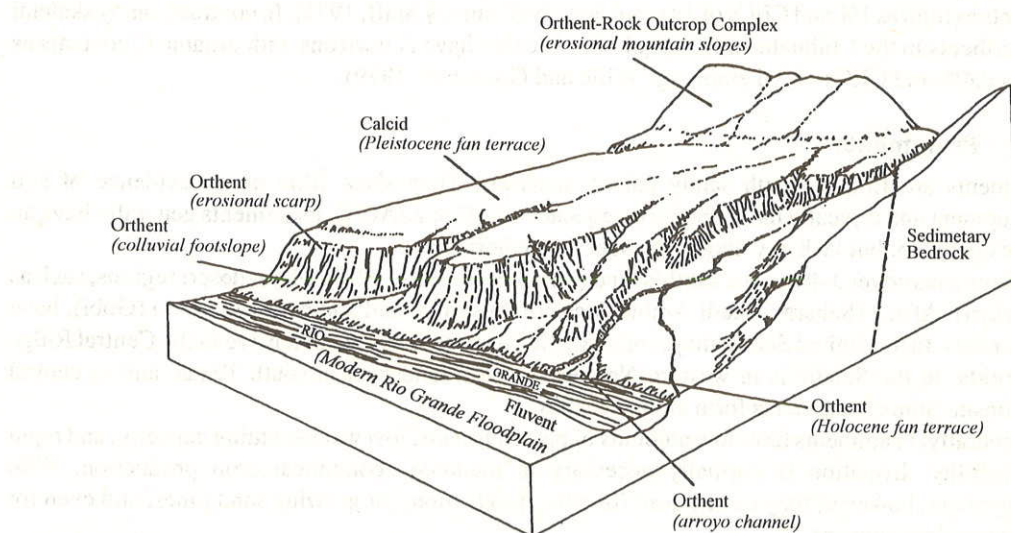


Fig. 6.21 Block diagram of a Basin and Range terrain in southern New Mexico illustrating landforms where Orthents occur [Modified from Gile and Grossman, 1979]

in monzonite versus rhyolite parent material, for instance, reveals a large difference in the nature of the soils (Gile et al., 1981). Monzonite disintegrates along grain mineral contacts and may produce a fine-textured soil, while rhyolite resists weathering because it lacks planes of weakness along which water can enter. As a result, monzonite weathers into soil (Orthents) much sooner than rhyolite.

The amount of detrital carbonate in alluvial parent materials can, like resistant bedrock, curtail the formation of diagnostic horizons and prolong the time a soil remains an Entisol. For example, soils in southern New Mexico that formed in calcareous alluvium deposited in the late Holocene often do not have diagnostic subsurface horizons. Because these soils do not have an irregular depth distribution of organic matter, they classify as Orthents (Gile et al., 1981). In contrast, neighboring soils of similar age that formed in alluvium derived from igneous rocks have faint subsurface coatings of oriented clay, which is enough development to qualify as cambic horizons and Aridisols. It appears that the formation of cambic and argillic horizons is delayed in carbonate-rich parent materials because the large amount of Ca keeps the clay flocculated and immobile.

6.4.2.3 Lower Taxonomic Levels

Orthents are subdivided at the great group level based on temperature and moisture regimes (Soil Survey Staff, 1998). For example, Torriorthents occur in desert environments and Udorthents in humid environments. At the subgroup level, Orthents are subdivided according to many features of the soils. These subgroup categories include: lithic, pergelic (frozen), vitrandic, aquic, oxyaquic, alfic (lamellae), ustic, ustertic, xerertic, vertic, haploduridic, oxyaquic, xeric, vermic, anthraquic, and dystic.

6.4.2.4 Physical and Chemical Properties

In contrast to soils that are classified according to their pedogenically produced diagnostic horizons, the physical and chemical properties of Orthents are inherited largely from their parent materials. As a result, the range of Orthent properties is as wide as the properties of the unconsolidated material in which they occur. For example, coarse-silty Udorthents in Iowa can have A horizons with organic C contents as high as 1% and CECs of $14 \text{ cmol}_c \text{ kg}^{-1}$ (Soil Survey Staff, 1975). In contrast, sandy skeletal Torriorthents in the Chihuahuan Desert of New Mexico have A horizons with organic C contents as low as 0.4% and CECs of $7.5 \text{ cmol}_c \text{ kg}^{-1}$ (Gile and Grossman, 1979).

6.4.3 Psamments

Psamments are Entisols with sandy parent materials. They show little or no evidence of soil development and typically have a horizon sequence of AC or A/AC-C. Psamments generally have an ochric epipedon, but lack any diagnostic subsurface horizons.

Psamments cover 3.4% of the Earth's land (Soil Survey Staff, 1997). Major desert regions, such as in northern Africa (Sahara), Saudi Arabia, western Australia, and southern Mongolia (Gobi), have Psamments. In the United States, these soils exist in all states. They are extensive in the Central Ridge of Florida, in the Sandhills in western Nebraska, in the sandsheet in south Texas, and in central Wisconsin. Some Psamments form along beaches.

Generally, Psamments have low amounts of plant nutrients, low water-holding capacity, and rapid permeability. Irrigation is normally necessary to maintain economical crop production. With management, however, they can be used for citrus production, for growing sand pines, and even for real estate development.

Natural vegetation associated with Psamments in Florida consists of bluejack (*Quercus incana*), blackjack oak (*Quercus marilandica*), turkey oak (*Quercus laevis*), longleaf threeawn (*Aristida*

stricta), and bluestem (*Andropogon* sp.). In the major cattle-growing states of Texas and Nebraska, Psamments are used for spring and summer grazing of native grasses. In central Wisconsin, truck crops are an important use.

6.4.3.1 Lower Taxonomic Levels

Psamments are restricted to soils that have a texture of loamy fine sand or coarser in the control section (family particle size class) and < 35% by volume of particles > 2 mm diameter (Soil Survey Staff, 1998). The latter distinguishes Psamments from sandy Orthents.

Psamments occur in a wide range of climates. Six of the seven great groups reflect a climatic/geographic bias. Only Quartzipsamments do not indicate a geographic location, but rather the dominant mineralogy, quartz. Psamments exist from cryic or pergelic (Cryopsamments) soil temperature regimes to areas in the world where temperature at a depth of 50 cm from the soil surface does not change more than 6 °C between mean summer and winter soil temperatures (Tropopsamments). Also, they are located in torric (Torripsamments), ustic (Ustipsamments), xeric (Xeropsamments), and udic (Udipsamments) soil moisture regimes.

Subsurface features present in some Psamments include a lithic contact (Lithic), aquic or oxyaquic conditions, lamellae (Argic), plinthite, or durinodes. Thus, subsoil features vary greatly and the variation is noted in the subgroup taxonomic name. Examples include Lithic Cryopsamments, Plinthic Quartzipsamments, and Aquic Durinodic Xeropsamments.

At the subgroup taxonomic level, climatic/geographic conditions are also described. As an example, Aridic Ustipsamments are located in several soil temperature regimes, but must have a period of time in which the soil is dry in all parts. The length of time that the soil must be dry is defined by the soil temperature regime. Xeric Torripsamments are defined by both soil temperature (Torri) and moisture (Xeric) regimes.

6.4.3.2 Genesis

Genetically, Psamments develop in areas high in sand content (> 85%). The parent material can be geologically young, as in eolian or alluvial materials or old, as in sandy residual bedrock. Sand dunes of either marine (Florida) or eolian (Saudi Arabia or Nebraska) origin are common parent materials. Some Psamments form in materials weathered from sandstone bedrock (e.g., South Dakota or Montana). Depending on the mineralogy of the sand, Psamments (Quartzipsamments) can be very resistant to weathering.

6.4.3.3 Representative Psamments

Psamments from Florida (Astatula series: hyperthermic, uncoated, Typic Quartzipsamments), Nebraska (Valentine series: mixed, mesic Typic Ustipsamments), and Saudi Arabia (Torripsamments) were selected to represent different geographic areas and, therefore, different genetic pathways. But morphologically, these soils are very similar. The soils are also excessively drained.

Astatula, Valentine, and the Torripsamments from Saudi Arabia have thick C horizons with thin A horizons. If an A horizon is present, it is light colored and < 15 cm thick. The A horizon is subject to continual mixing by wind action. Astatula soils form in eolian and marine sands and are extensive in peninsular Florida on upland slopes that range from 0 to 30%. These soils occur in high rainfall areas (about 1,270 mm yr⁻¹), are highly leached, and are typically strongly acid (pH < 5.5). Valentine soils form in sand dunes in the sandhills of Nebraska on slopes that range from 0 to 60%. The sand dunes generally are oriented in a northwest to southeast direction (prevailing wind direction). In addition to Nebraska, Valentine soils occur extensively in Kansas, Montana, New Mexico, South Dakota, Texas,

and Wyoming (> 2 million ha). The Torripsamments in Saudi Arabia form in sand dunes and also in alluvial sandy deposits on plains and stream terraces. The dunes range in height from < 2 to > 10 m. The Torripsamments in these desert areas (< 100 mm yr⁻¹ precipitation) typically have an accumulation of bases, and are alkaline in reaction (pH > 8.5).

6.4.3.4 Physical and Chemical Properties of Representative Psamments

Physical properties and chemical properties are presented for two selected Psamments (Table 6.21). Organic C contents typically are very low with pH ranging from strongly acid in the Psamments of Florida that occur in high rainfall areas (Astatula series), to strongly alkaline soils in Saudi Arabia (data not shown) that occur in the very dry desert regions. The sand size fraction of the Astatula and Valentine series typically is dominated by fine sand. Clay content of these Psamments is typically < 5%.

6.4.4 Fluvents

Fluvents are Entisols that typically form in alluvium of rapidly aggrading flood plains, fans, deltas, and in some cases, mudflows. In the United States, the greatest abundance of Fluvents occur within the numerous and large floodplains of the Midwest (Lindbo, 1997). World coverage of Fluvents is 2.2% (Soil Survey Staff, 1997).

When rates of deposition exceed rates of pedogenesis in alluvial settings, A-C soil profiles develop (Table 6.22). Fluvents typically have ochric epipedons, but no subsurface diagnostic horizons (Soil Survey Staff, 1975, 1998). Organic C content in Fluvents must decrease irregularly with depth, or be above 0.2% at a depth of 125 cm (Table 6.22). The rationale of the Fluvent suborder is that deposition inhibits diagnostic horizon formation and is accompanied by fluctuations in organic C levels associated with stratification.

Fluvents are differentiated from other Entisol suborders by: (1) having textures that are finer than loamy very fine sand and with less than 35% rock fragments (Psamments), (2) occurring on slopes less than 25% and not having bedrock within 25 cm (Orthents), (3) not having aquic conditions (Aquepts), (4) not having fragments of diagnostic horizons (Arenets), and (5) having mean annual soil temperatures greater than 0 °C (Soil Survey Staff, 1998, 1975).

6.4.4.1 Lower Taxonomic Levels

Great groups of the Fluvents are differentiated by either soil temperature or soil moisture regime (Soil Survey Staff, 1996). Fluvents occur in udic (Udifuvents), ustic (Ustifuvents), xeric (Xerofluvents), and torric (Torrifluvents) soil moisture regimes. They also occur in cryic (Cryofluvents), isomesic, isothermic, and isohyperthermic (Tropofluvents) soil temperature regimes. At the subgroup taxonomic level, Entisols intergrade to the Andisol (andic), Mollisol (mollic), and Vertisol (vertic) soil orders, and to numerous other soil moisture and temperature regimes not used at the suborder categorical level (Soil Survey Staff, 1998).

6.4.4.2 Climate, Vegetation, and Topography

Fluvents are soils that are not in equilibrium with the effects of climate, vegetation, and in some cases, topography. Thus, Fluvents can occur in most climates associated with most vegetation communities, and on a variety of slopes. Climate can, however, indirectly influence the chemical and physical properties of Fluvents. Alluvium associated with flood plains, fans, deltas, or mudflows may be inherited from upland soils where properties are more likely to reflect equilibrium with regional climate, vegetation, and topography.

Table 6.21 Soil characterization data for the Astula fine sand [Sodek et al., 1990] and Valentine fine sand [Soil Survey Staff, 1966]

Horizon	Depth	Very Coarse Sand			Medium Sand			Very Fine Sand			pH ^a	Organic Carbon	Extractable				Cation Exchange Capacity	Base Saturation
		Sand	Coarse Sand	Sand	Sand	Medium Sand	Fine Sand	Fine Sand	Silt	Clay			Ca	Mg	Na	K		
Astauala fine sand (Typic Quartzipsamments)																		
A	0-13	---	0.1	18.8	76.7	2.1	0.9	1.4	5.5	0.5	0.5	0.1	0.0	0.0	0.0	2.7	23	
C1	13-64	---	0.1	18.4	76.5	2.4	1.6	1.5	5.2	0.2	0.2	0.0	0.0	0.0	---	1.5	5	
C2	64-175	---	0.1	19.7	75.8	2.2	0.6	1.6	5.0	0.1	0.1	0.0	0.0	0.0	---	0.7	5	
C3	175-203	---	0.1	19.4	76.2	2.0	0.6	1.7	5.0	0.0	0.0	0.0	0.0	0.0	---	0.9	6	
Valentine fine sand (Typic Ustipsamments)																		
A	0-10	0.2	4.2	12.7	50.2	24.7	4.1	3.9	6.0	0.1	0.1	3.6	0.6	0.1	0.3	5.9	78	
AC	10-25	0.1	4.6	13.3	55.1	21.6	2.2	3.1	6.2	0.5	0.5	2.1	0.4	0.1	0.3	3.7	78	
C1	25-45	---	3.6	11.2	56.2	24.0	1.8	3.2	6.5	0.2	0.2	2.0	0.6	0.1	0.1	0.1	88	
C2	45-81	---	4.4	18.6	58.9	14.7	1.0	2.4	6.7	0.1	0.1	1.5	0.2	0.1	0.2	2.4	79	
C3	81-137	0.1	3.8	10.9	59.1	22.1	1.3	2.9	6.8	0.0	0.0	1.8	0.8	0.1	0.1	2.8	100	

^aDetermined in a 1:1 H₂O solution for the Astula fine sand and in a 1:5 CaCl₂ solution for the Valentine fine sand

Table 6.22 Soil characterization data for Fluvents from Iowa (Udifuvents) and New Mexico (Torrifluvents)

Horizon	Depth cm	Sand	Silt	Clay	Organic Carbon
-----%					
Udifuvents ^a					
Ap	0-14	5.1	78.4	16.5	1.28
C1	14-31	6.9	78.6	14.5	0.57
C2	31-42	4.1	81.0	14.9	0.60
C3	42-56	16.5	73.0	10.5	0.45
C4	56-66	8.2	75.3	16.5	0.57
C5	66-98	34.1	55.3	10.6	0.42
C6	98-125	13.0	71.6	15.4	0.52
Torrifluvents ^b					
A1	0-5	74.8	16.2	9.0	0.66
A2	5-13	66.9	21.1	12.0	0.71
2A	13-18	73.1	16.1	10.8	0.49
3A	18-38	60.4	25.2	14.4	0.54
4C	38-43	69.5	19.2	11.3	0.33
5C	43-64	62.6	25.1	12.3	0.41
6C	64-81	77.8	13.5	8.7	0.19
7C	81-119	81.7	11.5	6.8	0.24

^a Missouri River flood plain, Monona County, Iowa. Modified from Soil Survey Staff, 1975, p. 614^b Gardner Springs Arroyo, Dona Ana County, New Mexico. Modified from Gile and Grossman, 1979, p. 874

6.4.4.3 Hydrology

Flood frequency, flood magnitude, and sediment load control the amount of deposition that occurs on flood plains. Prior to 1,000 yr BP, many North American rivers were forming cumulative Mollisols and Inceptisols in response to moderate frequency and low magnitude floods (Hall, 1990; Waters, 1991; Blum and Valastro, 1994; Arbogast and Johnson, 1994; Waters and Nordt, 1995). Between 1,500 and 500 yr BP there was widespread regional river channel trenching. This event transformed many broad flood plains into infrequently flooded surfaces, herein called flood terraces. Flood terraces flood on occasion, but the frequency is low enough that pedogenesis is beginning to transform the associated soils into orders beyond Entisols. In Australia, Walker and Coventry (1976) reported that flood plains that flooded once each year maintained soil development in the Entisol stage, whereas when flood frequencies were on the order of 1 to 10 yr, Mollisols began forming. Reservoirs built along major rivers during the last 100 yr have also reduced the flood frequency of many flood plains.

Countering late Holocene flood reduction by channel trenching or reservoirs, historic flood magnitudes are increasing in some areas because of intensified land use, increased runoff and channel peak discharge. As a consequence, it appears that some flood plains that were flooded infrequently during the last 500 to 1,000 yr, may be transforming back to more frequently flooded flood plains because of historic land use practices. This has resulted in deposition of post-settlement alluvium across numerous flood plains (Knox, 1987; Grossman, 1983; Scully and Arnold, 1981; Daniels and Jordan, 1966).

Deposition of post-settlement alluvium has buried many late Holocene soils in flood terrace positions. In most cases, if a recent flood deposit is < 50 cm thick, it is treated as a mapping phase and the buried soil is classified (Soil Survey Staff, 1975, 1998). If a recent flood plain deposit is over 50 cm thick, the recent material is classified, which may be a Fluvent (Fig. 6.22).

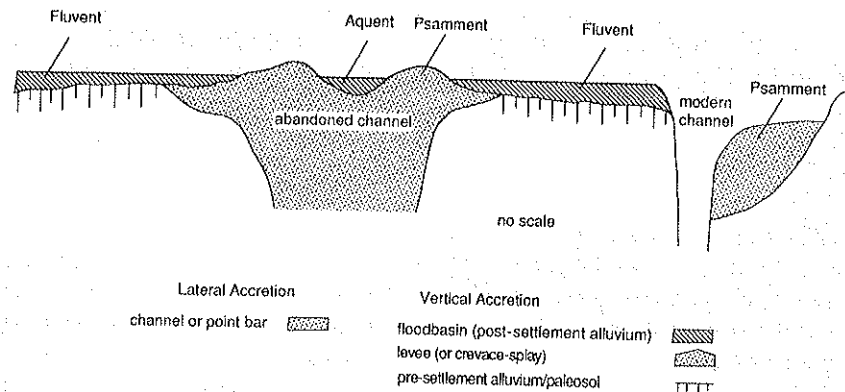


Fig. 6.22 Soil stratigraphic cross-section of an idealized alluvial floodplain

6.4.4.4 Time

Under conditions of landscape stability, the evolution of Entisols into other soil orders typically occurs in association with one or more properties approaching steady state. Surface horizons, and consequently Fluvents, can develop almost immediately after deposition because of the inheritance of detrital organic matter (Nordt and Hallmark, 1990). In general, organic matter inputs attain steady state in < 1,000 yr as Entisols evolve into other soil orders (Dickson and Crocker 1953a, 1953b; Parsons et al., 1970; Walker and Coventry, 1976; Hallberg et al., 1978; Schafer et al., 1980; Nordt and Hallmark, 1990). In some humid regions, mollic epipedons and Mollisols can develop in as little as 100 yr (Ruhe et al., 1975). In other humid regions, Entisols can develop into Inceptisols within 500 to 600 yr (Bilzi and Ciolkosz, 1977; Scully and Arnold, 1981). Fluvents in arid regions can persist for a thousand years in low carbonate parent material and up to several thousand years in carbonatic parent material (Gile, 1975). After this time, subsurface diagnostic horizons and Aridisols form.

6.4.4.5 Parent Material

Flood plains, fans, and deltas are constructed by the alluvial processes of vertical and lateral accretion (Reeding, 1982). Lateral accretion results from lateral channel migration and deposition of bar and channel facies (Fig. 6.22). Vertical accretion results from overbank deposition, which entails levee, floodbasin, and sometimes crevasse splay facies. The distribution of facies in a flood plain system is dependent on stream type (Chorely et al., 1985). Braided channels in either floodplains or alluvial fans are dominated by laterally accreted channel and bar deposits. Straight to meandering streams tend to deposit thick vertical accretion facies. Anastomosing streams have multiple bifurcating and reconnecting channels with a high proportion of fine grained lateral accretion deposition. Deltas are dominated by a distributary channel network with a complex array of lateral and vertical accretion deposits. Fluvents are most likely to form in lateral accretion deposits and in rapidly aggrading vertical accretion facies.

Sedimentological packages of meandering streams in humid regions typically display a fining upward sequence consisting of lower channel, intermediate point bar, and upper flood basin or levee facies. In some cases, Fluvents may form entirely within only one of these facies and not exhibit a noticeable textural change within the solum. If a Fluvent develops in both a point bar and flood basin facies, for example, a fining upward sequence may exist in the profile. In other situations, reverse grading may be encountered. For example, a levee may be deposited over a floodbasin facies, thus

producing a Fluvent with a coarsening upward textural distribution. In addition, other suborders of Entisols other than Fluvents may form in flood plains. Psammments may form in sandy channel and point bar deposits and Aquents in poorly drained topographic positions (Fig. 6.22). In sum, recognizing the alluvial stratigraphy of flood plains can enhance the understanding of the genesis of Fluvents (Section E, chapter 1).

6.4.4.6 Physical and Chemical Properties

The texture, chemistry, mineralogy, and color of Fluvents are determined largely by the character of the soils and parent materials (bedrock, till, loess, etc.) that are being eroded from the surrounding drainage basin. For example, permeability, water-holding capacity, and nutrient retention vary greatly depending on the texture and mineralogy of the parent material. In essence, Fluvents can have a wide variety of physical and chemical properties that are for the most part inherited from their parent materials. Most alluvial soils have high fertility levels and support some of the most important agricultural areas of the world (Edelman and van der Voorde, 1963; Radwanski, 1968; Sidhu et al., 1977). In the tropics, however, upland soils tend to be depleted in nutrients such that their alluvial counterparts produce low fertility flood plain soils (Edelman and van der Voorde, 1963).

6.4.5 Aquents

Aquents are wet Entisols. They exist widely in the United States, except in Arizona and New Mexico with the largest concentrations in south Florida (Soil Survey Staff, 1990). Worldwide, Aquents are not extensive, covering only approximately 0.1% of the Earth's land (Soil Survey Staff, 1997). Areas where Aquents occur include salt marshes of coastal tidal areas, freshwater marshes and swamps, alluvial flats and backswamps of flood plains, and outwash plains that receive new deposits of sediment at frequent intervals. Depending on where Aquents are located, some can be used for agricultural production when drained. Many undrained areas of Aquents are used for pasture or wildlife habitat.

Aquents are defined as having aquic conditions and sulfidic materials within 50 cm of the mineral soil surface, or permanent saturation and a reduced matrix below a depth of 25 cm from the mineral soil surface, or texture and color requirements if a dense, lithic, or paralithic contact exists or aquic conditions are present for some time in some years in a layer between 40 and 50 cm from the mineral soil surface (Soil Survey Staff, 1998). The soil properties of Aquents are highly variable. One common feature of all Aquents is their saturation with water at or near the surface for much of the year. Aquents can range in texture from sandy to clayey.

Wetness in Aquents is normally inferred by the color distribution generated by the redox state of Fe. In most Aquents, this process produces redoximorphic features (Fe concentrations or depletions) within the upper 50 cm of the surface. Redoximorphic features are most common in nonsandy Aquents. The sandy Aquents (Psammaquents) often lack redoximorphic features because of the very low Fe content of the quartz parent materials. Many Aquents have hydric soil indicators (thin layer of muck on the soil surface, strong sulfidic odor, or a gray and reduced soil matrix in the upper 15 cm of the soil). In Florida according to F.S. 17-340 (Florida Statutes, 1995), all Sulfaquents and Hydraquents are considered to be hydric soils. Many Psammaquents are also hydric according to the county listings of hydric soils.

6.4.5.1 Lower Taxonomic Levels

Aquents are associated with landscapes that accumulate soil water. They exist in cryic to isohyperthermic soil temperature regimes. Sulfaquents and Hydraquents are located along coastal

areas in tidal marshes. Sulfaquents, as the name implies, have sulfidic materials. Hydraquents are thixotropic soils, meaning that they cannot support appreciable overburden pressures (n value > 0.7). Psammaquents have a texture of loamy fine sand or coarser in all layers within the control section for the family particle size class. Fluvaquents occur on flood plains and are defined by the amount of organic C and its distribution in the soil. Cryaquents form in the colder climatic areas of Alaska and other higher latitudes that receive significant snowfalls, and soil development is slow.

6.4.5.2 Genesis

Aquents form where pedogenetic development does not occur or is very slow under wet conditions. The soil must be saturated for long periods of time, and in some areas, organic matter can accumulate to create a histic epipedon. The saturation in Aquents can be episaturated (perched watertable) or endosaturated (apparent watertable). The time needed for Aquents to develop can be short. On flood plains and other areas in which water flows across the area, Aquents commonly have stratified layers of sand, silt, clay and organic matter. Aquents of flood plains may also have buried histic epipedons, and Aquents with a high clay content may have cracks and slickensides.

Aquents from Florida (Pellicer series: fine, smectitic, nonacid, hyperthermic Typic Sulfaquents), Egypt (Edko series: fine-loamy, mixed, nonacid, hyperthermic Mollic Fluvaquents), and Alaska (Beluga series: coarse-loamy, mixed, nonacid, Typic Cryaquents) were selected to represent Aquents from different geographic areas and to reflect different genetic pathways. The Pellicer soils form in clayey marine sediments in tidal marshes along the Atlantic Ocean coastal areas of Florida (Carlisle et al., 1981). These soils are very poorly drained and are subject to daily tidal flooding. The representative Aquent from Egypt (Edko area) has anthric saturation and oximorphic features in the form of brownish Fe accumulations. The soil was sampled in a depressional area south of Lake Burullus near Resheed City in northern Egypt (Rasheed et al., 1992). It developed in recent fluvio-lacustrine deposits in an arid climate. The Beluga soils are extensive in flood plains and alluvial fans in southern Alaska and form in stratified alluvium and colluvium derived from soft clayey shale and sandstone. These soils are poorly drained and are saturated with water at the surface for portions of the year.

6.4.5.3 Physical and Chemical Properties

Physical properties such as particle size distribution, and chemical properties such as organic C content, pH, extractable bases, and CEC are presented for the Pellicer and Edko soils (Table 6.23). The sand, silt, and clay content of the Pellicer soil varies considerably within the profile, reflecting the different conditions under which the materials were deposited. Organic C content of the Pellicer soils is high in the surface horizon, the result of the accumulation of organic materials under wet conditions. The Pellicer soils are high in S content (1 to 4%), and this is reflected in the extremely low pH values (< 3.5). The sand, silt, and clay content of the Edko soil is fairly evenly distributed in each horizon. The pH of this soil is very high (> 8.5) due to the high Na content.

6.4.6 Arenets

Humans affect soils in many ways. Some Entisols have been created where humans have deposited or exposed soil parent material by earth-moving operations. Some of these operations include highway construction, surface mining, landfill construction, or dredging. In other cases, as in Arenets described below, Entisols have been created by humans because of severe physical disturbance of previously existing soils.

Table 6.23 Soil characterization data for the Pellicer clay [Carlisle et al., 1981] and Edko soil [Rasheed et al., 1992]

Horizon	Depth	Very					Very			pH ^a	Organic				Extractable			Cation	Base						
		Coarse Sand	Coarse Sand	Medium Sand	Fine Sand	Fine Sand	Silt	Clay	Carbon		Ca	Mg	Na	K	Exchange Capacity	Saturation									
cm																			%		cmol. kg ⁻¹			%	
Pellicer clay (Typic Sulfaquents)																									
A	0-51	0.1	0.5	2.1	3.5	9.3	23.6	60.9	2.7	14.3	11.7	24.3	27.7	1.0	129.6	50									
C1	51-89	0.1	1.1	15.9	21.6	11.3	16.6	33.4	3.2	7.1	6.8	15.7	20.7	1.1	78.6	56									
C2	89-140	0.1	2.6	38.7	34.1	10.9	7.5	6.1	3.1	0.8	1.3	3.3	5.0	0.2	15.4	63									
Edko soil (Mollic Fluvaquents)																									
A1	0-6		4.6		37.4		26.4	31.6	8.6	3.3	12.0	16.0	64.4	1.1											
A2	6-23		4.0		29.1		36.0	30.9	8.4	1.6	14.0	24.0	67.2	0.9											
C	23-56		0.4		29.9		34.8	29.9	8.6	0.9	10.0	14.0	65.2	4.5											
2C	56-76		0.4		29.9		34.8	29.9	8.6	0.9	10.0	14.0	65.2	4.5											

^a Determined in a 1:1 solution of distilled water

Arents are a suborder of Entisols that are physically disturbed by the actions of humans (Soil Survey Staff, 1998). Arents are restricted to those soils which have, in one or more layers between 25 and 100 cm from the mineral soil surface, 3% or more by volume of fragments of diagnostic horizons that are not arranged in any discernible order (Soil Survey Staff, 1998). This suborder was placed in *Soil Taxonomy* for situations where previously existing soils had been significantly physically modified. One situation where this is known to have happened is in California where duripans have been ripped by heavy equipment to improve the soils for the growth of almond trees, based on observations in the 1960s (Simonson, 1969). The ripping was possible only with moderately developed duripans because thick, strongly developed duripans were impossible to rip. The value of the land was sufficiently improved to permit successful commercial almond production. As a result of the duripan ripping, the great group classification of the soils changed from Durixeralfs to Xerarents.

In formerly gullied lands in loess bluffs in western Tennessee, land leveling was used to improve the soils for agricultural purposes. Fragments of argillic horizons from the original Alfisols permitted the new soils to be identified as Arents after the land disturbance. Thus, Hapludalfs (e.g., soils of the Memphis soil series) were reclassified to Udarents following the leveling. In Texas, reclamation of lignite surface mine lands with mixed spoil materials gives rise to Arents, especially near the margins of the mine lands. Where the overburden is thin relative to thick Paleudalfs in premine condition.

6.4.6.1 Soils in Anthromorphic Parent Materials

In many soils with human-deposited or human-exposed parent materials, fragments of diagnostic horizons are not identifiable; thus a classification of Arents would not be appropriate. Such soils commonly have been classified in the suborder of Orthents and subdivided into great groups according to the soil moisture regime (e.g., into Udorthents in the eastern United States) (Soil Survey Staff, 1998). This classification has seemed insufficient to many soil scientists because it does not reflect the role of humans in the genesis of the soils (Short et al., 1986a; Strain and Evans, 1994). Although these soils are commonly thought to be highly variable, studies have shown a degree of variability similar to that of some soils in naturally deposited parent materials (Short et al., 1986b).

Many soils in materials deposited by bulldozers or other earth-moving equipment have an irregular distribution of organic C with depth. In theory, these soils would classify as Fluvents at the suborder taxonomic level (Soil Survey Staff, 1998). In contrast to Fluvents, however, human-disturbed soils are not stratified by natural processes, are not flooded in most years, and are more dense. Because these soils behave more like Orthents than Fluvents, they have been grouped, during soil surveys such as that of Washington, DC (Smith, 1976), into Orthents by ignoring their irregular organic-C depth distribution.

Because of the inadequacies of the present classification system, new suborders and/or subgroups of Entisols have been proposed for soils in human-deposited or human-exposed parent materials (Fanning and Fanning, 1989). At present, there is considerable interest in an international committee for anthropogenic soils (ICOMANTH) for the addition of new taxonomic classes (Section E, Chapter 6.1). However, the taxonomic level at which the recognition should take place has yet to be firmly decided. The suborder of Arents, discussed above, established a precedence for recognition at the suborder level. Fanning and Fanning (1989) have supported this viewpoint. However, some others (Strain and Evans, 1994; Kosse, 1988) have suggested that the recognition be at the order level called Anthrosols.

For soils in earth spoil, such as those deposited during surface mining operations, a suborder of Spolents (for spoil) was proposed by West Virginia workers (Sencindiver, 1977; Smith and Sobek, 1978). During the soil survey of Washington, DC (Smith, 1976), a suborder of Urbents was considered for soils in miscellaneous urban fill that had object artifacts such as brick or concrete in the particle size

control section (generally between a depth of 25 and 100 cm). For soils with garbage of an organic nature within 2 m of the soil surface, which would readily subside and generate methane under anaerobic conditions, a suborder of Garbents was proposed (Fanning and Fanning, 1989; Fanning, 1991). Proposals for special diagnostic characteristics (urbic, garbic, spolic, and dredged material, and the scalped land surface) for *Soil Taxonomy* were proposed by Fanning and Fanning (1989). These characteristics were employed to propose special subgroup classes, such as Urbic Garbic Udorthents for sanitary landfills where the cover material qualified as urbic material (Fanning and Fanning, 1989). It was proposed that soils on scalped land surfaces (an engineering perspective) could be placed in Scalpic subgroups.

Soils created by dredging operations commonly qualify as Sulfaquents immediately after sulfidic-dredged materials have been deposited (Fanning and Fanning, 1989). Under aerobic conditions at dredged material sites, sulfuric horizons commonly form from sulfidic materials in a few weeks or months. This causes the soils to become Sulfaquepts, a great group of Inceptisols. Thus, the period that such soils remain as Entisols, before becoming a soil of a different order, may be very brief.

With the present strong interest in improving knowledge about the classification of human-influenced soils, much progress may be expected in the near future. This is needed as soil surveys of lands with human-influenced soils are becoming more widespread. A soil survey of the City of New York is presently underway, and many of the problems associated with the soils discussed in this section will be encountered. For example, sulfide-bearing clays have been used as landfill cover materials, which can lead to the formation of acid sulfate soils (Kargbo et al., 1993; Fanning and Burch, 1997). A more resolving classification system can enhance future interpretations and mapping techniques of soil surveys in these areas. Further, it has been suggested (Fanning, 1990) that a discipline of pedotechnology is needed to improve the design of soils brought into existence through the action of humans.

6.4.7 References

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6.5 Inceptisols

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6.5.1 Introduction

Inceptisols serve two important functions in *Soil Taxonomy*. First, they group soils with incipient soil development (L. *Inceptum*, beginning); and second, they ensure that the classification system is fully bifurcated—that is, all soils fall within some taxa of the classification system. Because of the latter, Smith (1986) referred to Inceptisols as the wastebasket order. They are the repository for all soils that do not meet class differentiae of other orders. While the central concept of Inceptisols is that of soils in cool to very warm humid and subhumid regions that have a weakly developed subsoil horizon (cambic) and a light colored surface horizon (orchric), the order contains a wide variety of soils (Soil Survey Staff, 1998a,b). In some areas they contain soils with minimal development, while in other areas they represent soils with well-expressed diagnostic horizons that fail the criteria of other orders.

Inceptisols, because of their broad and inclusive nature, served as the precursor for new soil orders (Andisols and Gelisols) and new taxa in other orders. For example, shrink/swell soils with aquic conditions and those with cryic soil temperature regimes previously placed within Inceptisols are now Vertisols (Aquerts and Cryerts). Inceptisols with isomesic or warmer *iso* temperature regimes, previously called Tropepts, are now distributed among other great groups and families of Inceptisols with *iso* temperature regimes; Inceptisols as markers of past cultural habitats and anthropogenic activities (Plaggepts) are now termed Anthrepts; and Ochrepts, a previous suborder of Inceptisols, has now been incorporated into Inceptisol suborders reflecting soil moisture bias. Table 6.24 illustrates the dynamics of these changes and how suborders of Inceptisols have evolved with increased knowledge over the past 40 years.

The central concept of Inceptisols includes soils that have undergone modifications of the parent material by structural development and alteration sufficient to differentiate them from Entisols. Pedogenesis may include eluviation (losses of constituents); the translocation of clay, iron, silica, alumi-

Table 6.24 Inceptisol suborders reflecting the evolution of *Soil Taxonomy*

7th Approximation (1960) ^a	<i>Soil Taxonomy</i> (1975) ^b	<i>Keys to Soil Taxonomy</i> (1990) ^c	<i>Soil Taxonomy</i> (1998) ^d
Andepts	Andepts	Aquepts	Anthrepts
Aquepts	Aquepts	Ochrepts	Aquepts
Ochrepts	Ochrepts	Plaggepts	Cryepts
Umbrepts	Plaggepts	Tropets	Udepts
	Tropets	Umbrepts	Ustepts
	Umbrepts		Xerepts

^a Soil Survey Staff (1960)^b Soil Survey Staff (1975)^c Soil Survey Staff (1990)^d Soil Survey Staff (1998a,b)

num, carbonates, bases, and organic matter; and the formation of redoximorphic features by aquic (hydromorphic) conditions. Inceptisols occur in all known climates, except under aridic and pergelic conditions, and have many kinds of diagnostic horizons and epipedons. Hence, while the central concept expresses soils with incipient development, Inceptisols are an order with inordinate diversity. No attempt will be made herein to consider their full diversity. Rather, examples of the great range in physical, chemical, biological and mineralogical attributes of this order pertinent to land use and behavior will be illustrated. For a more extensive coverage of these attributes the reader is referred to online resources (<http://www.statlab.iastate.edu/soils/soildiv/>).

6.5.2 Distribution

The worldwide distribution of Inceptisols is illustrated in Fig. 6.23. Because these soils lack a sharply focused central concept, they occur indiscriminately globally. Where Inceptisols reflect youthfulness, they occur in high gradient mountainous regions, along major river systems as deltaic/fluvial plains, and as soils developed from carbonate-rich bedrocks or sediments in positions of geomorphic instability. Where Inceptisols serve to bifurcate *Soil Taxonomy* there is no particular pattern to their occurrence, but they are commonly dispersed among more strongly developed Alfisols, Ultisols and Oxisols in tropical and subtropical regions (Fig. 6.23).

6.5.3 Formation of Inceptisols

Because soils placed in Inceptisols are varied and represent many different pedogenic processes and combinations of soil-forming factors, examples will be given that typify central concepts or exhibit unique characteristics.

6.5.3.1 Parent Material

Foss et al. (1983) state that Inceptisols develop on geologically young sediments or landscapes and/or under environmental conditions that inhibit soil development (e.g., coarse-textured siliceous deposits that are resistant to weathering or parent materials within cool climates that inhibit pedogenesis). Given the above constraints, parent materials include almost all types of igneous, metamorphic, and sedimentary materials (residuum, alluvium, colluvium, loess/eolian, glacial drift, etc.). Inceptisols are excluded from soils having a sandy texture throughout because it is believed that pedological features indicative of eluviation, weathering, translocation, and transformation (e.g., development of color Bw horizons) take a very short time to form in subsoils with such low specific surface area

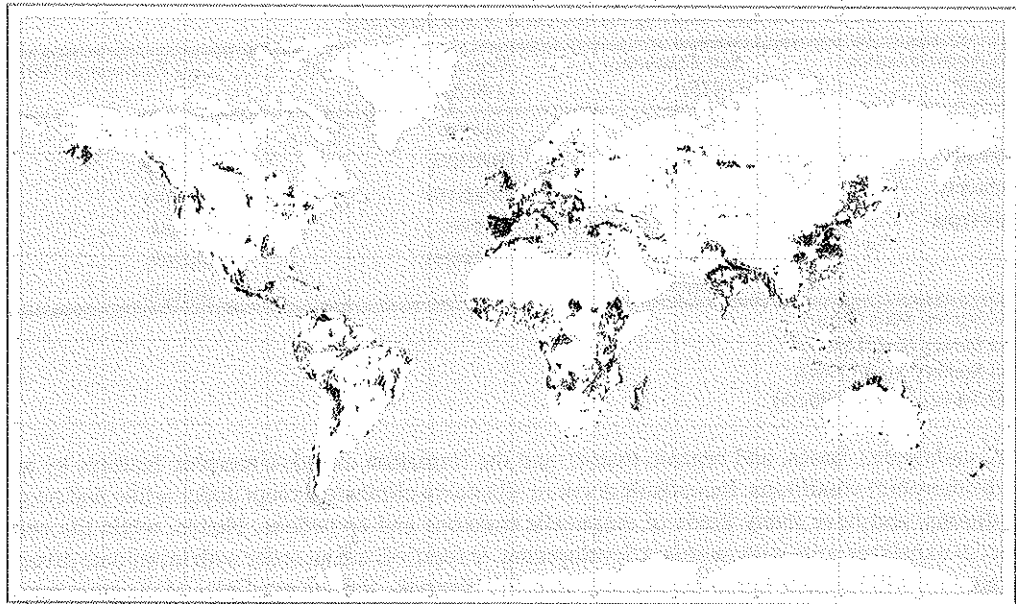


Fig. 6.23 Global distribution of Inceptisols [Courtesy of USDA-NRCS, Soil Survey Division, World Soil Resources, 1998]

coarse-textured materials. The kinds and arrangement of horizons, and their chemical, physical and mineralogical properties, are controlled to a major degree by the kinds of parent material from which Inceptisols have developed. For example, fine-textured parent materials give rise to clayey Inceptisols, acidic parent materials result in base-poor (*Dystr*) Inceptisols, and calcareous parent materials form neutral or alkaline, base-rich (*Eutr*) Inceptisols. Depth, color, organic carbon content and drainage frequently reflect the combination of geomorphic position, landscape instability, and nature of the parent materials.

Table 6.25 illustrates the diversity in selected physical, chemical and mineralogical properties of soils developed from six different parent materials [e.g., Mississippi River alluvium (*Fluvaquentic Epiaquepts*); noncalcareous glacial outwash (*Aeric Fragiaquepts*); Permian red bed shales (*Vertic Calciustepts*); basic igneous rock (*Oxic Haplustepts*); colluvium weathered from sandstone, siltstone and shale (*Typic Dystrudepts*); and Coastal Plains alluvium (*Typic Sulfaquepts*)]. Abridged attributes of these soils follow with detailed morphological, physical, chemical and mineralogical data found online at <http://www.statlab.iastate.edu/soils/soildiv/>:

Fluvaquentic Epiaquepts (Conciene series): These soils developed from loamy alluvial sediments that have weathered into soils with moderately expressed cambic horizons. Their nearly level to depressional landforms favor restricted drainage, especially under humid climates. Under slightly better drained conditions, soils developed in this alluvium may form argillic horizons in a relatively short time (hundreds to several thousand years). This is because the eluviation-illuviation process proceeds relatively rapidly in such permeable, base-rich, medium textured parent sediments.

Vertic Calciustepts (Vernon series): The calcareous clays and shales of the Permian red beds restrict water percolation and profile development. Well-drained soils that develop in these sedimentary materials have a calcic or salic horizon whereby lithogenic carbonates and soluble salts have been translocated to subsoils and precipitated as pedogenic products at the terminus of the wetting

Table 6.25 Physical and chemical properties of selected pedons in the Inceptisol order. See <http://www.statlab.iastat.e.edu/soils/soildiv/> for complete morphological and characterization data sets

Horizon	Depth, cm	Clay	Sand	OC ^a	Extr. Fe	CEC/ clay	WRD ^c	CEC	BS [*]	CaCO ₃	pH _{H₂O}	COLE [†]
				-----%			cm cm ⁻¹	cmol _c kg ⁻¹		-----%		cm cm ⁻¹
Conciene series (S88-LA-047-001) (fine-silty, mixed, superactive, hyperthermic, Fluvaquentic Epiaquepts) Louisiana, USA												
Ap1	0-14	13.8	20.4	1.94	0.6	1.25	0.25	17.3	100	---	6.8	0.024
Bg2	51-78	15.9	17.6	0.30	0.4	0.91	0.22	14.5	100	1	7.7	0.009
Bgssb	180-215	50.4	4.3	0.71	1.0	0.69	0.19	34.9	100	1	7.5	0.116
Vernon soil series (S80-TX-253-001) (fine, mixed, semiactive, thermic, Vertic Calcicustepts) Texas, USA												
A1	0-6	31.5	23.7	1.46	0.9	0.65	---	20.6	100	7	8.3	---
Bca2	45-65	40.5	13.8	0.21	1.7	0.34	0.08	14.8	100	23	8.5	0.042
Cx2	160-190	52.9	5.2	0.04	1.9	0.48	---	25.4	100	11	8.7	---
Kahana series (S89-HI-009-004) (very fine, kaolinitic, semiactive, isohypothermic Oxic Haplustepts) Hawaii, USA												
Ap	0-38	71.9	3.3	1.20	5.7	0.21	---	14.8	26	---	4.7	---
Ba2	56-87	59.8	7.1	0.45	5.9	0.23	---	14.0	71	---	6.2	---
Ba4	120-155	48.3	12.1	0.33	5.9	0.28	---	13.7	80	---	6.3	---
Volusia series (S86-NY-025-001) (fine-loamy, mixed, mesic, Aeric Fragaquepts) New York, USA												
Ap	0-20	---	---	4.50	2.0	---	---	22.5	44	---	5.4	---
Bw	20-29	---	---	0.95	1.8	---	---	11.3	33	---	5.5	---
Eg	29-42	---	---	0.27	1.2	---	---	6.3	30	---	5.2	---
2Bx3	90-145	---	---	0.16	1.5	---	---	10.9	68	---	6.2	---
Feds Creek series (S83-KY-195-018) (coarse-loamy, mixed, semiactive, mesic, Typic Dystrudepts) Kentucky, USA												
A	0-9	13.5	38.0	2.17	1.2	0.75	0.28	10.1	37	---	5.0	1.3
Bw2	42-76	14.7	43.6	0.20	1.2	0.34	0.17	5.0	20	---	4.8	0.4
BC	122-152	17.4	36.8	0.43	1.5	0.41	0.13	7.2	31	---	4.9	0.5
Gimhae series (S85-FN-515-012) (fine-silty, mixed, active, mesic, Typic Sulfaquepts) Korea												
Ap1	0-9	33.2	4.6	1.97	1.5	0.34	0.19	11.3	80	---	5.1	0.2
Bg1	15-32	34.4	5.2	1.59	1.9	0.38	---	13.2	42	---	4.0	---
Bg3	66-125	19.4	29.2	1.64	0.9	0.58	0.28	11.3	100	---	3.6	1.4
Vidhas soil (S94-FN-120-009) (fine-silty, mixed, superactive, thermic, Fluventic Haploxerepts) Albania												
Ap	0-25	21.4	12.4	0.98	0.9	0.75	0.17	16.1	100	20	8.1	0.6
Bw1	49-83	31.1	0.9	0.59	0.9	0.66	0.18	20.5	100	20	8.3	1.6
Bw2	83-116	59.8	1.9	0.69	1.3	0.52	0.09	31.2	100	15	8.2	1.4
Moran series (S85-CO-051-005) (loamy-skeletal, mixed, superactive, Humic Dystricrepts) Colorado, USA												
Oi	8-0	---	---	16.6	---	---	---	---	---	---	---	---
A1	0-19	17.5	46.1	5.07	2.3	1.33	---	23.3	33	---	4.8	---
A2	19-64	13.1	54.0	3.36	2.3	1.40	---	18.4	24	---	5.2	---
Bw	64-98	14.5	50.7	3.05	2.3	1.32	---	19.2	22	---	5.2	---

^a OC = organic C, ^b WRD = water retention difference, ^c BS = base saturation, [†] COLE = coefficient of linear expandability

front. The depth of carbonate and salt translocation, degree of calcic or salic horizon expression, and thickness of the cambic horizons are conditioned by the landform position, texture and permeability of the parent materials.

Oxic Haplustepts (Kahana series): These well-drained, clayey soils have developed in basic igneous rocks in Hawaii. Under a warm, tropical environment, the weathering of nonresistant mafic-rich minerals and fine-grained to amorphous materials has been rapid. An oxic-like (sesquioxide rich, subactive clay) cambic subsoil has developed. On a stable landform position, under an ustic moisture regime, subsequent weathering over time will result in an Oxisol with low activity clay.

Aeric Fragiaquepts (Volusia series): These soils have developed from weakly or noncalcareous glacial outwash of stratified sand, silt, and gravel. Upon weathering, these poorly sorted materials form loamy pans that have a brittle or fragile rupture upon deformation (fragipans). Weathering discontinuities in these materials enhanced by textural stratification are believed to result in weak cementation between closely packed skeleton grains (Smeck et al., 1989). The bonding agents between the grains are believed to be hydrous oxides of Al and Si and/or translocated clay coatings or bridges. Restricted drainage would complement transport discontinuities in these soils to form weathering discontinuities.

Typic Dystrudepts (Feds creek series): These well-drained soils form under udic moisture regimes in loamy colluvium weathered from sandstone, siltstone and shale. The parent materials consist of acid channery-sized skeletal materials. Because the chemical and physical buffering capacity of the colluvium is low, and the parent materials are siliceous clasts, very acid conditions develop upon weathering. Profile development within the solum, however, is minimal because the preweathered clasts are resistant to further decomposition and disintegration. Hence, release of clay and other products of weathering for subsequent translocation and transformation is limited. Again, the nature of the parent material has been pivotal in governing the intensity, modes, and mechanisms of pedogenesis in these coarse-textured, highly permeable, acid upland soils.

Typic Sulfaquepts (Gimhae series): These soils are common along coastlines with tidal influence. Here the parent materials are medium textured alluvial deposits, usually associated with deltas or depositional regions of rivers that empty into the ocean. These wet environments are favorable for the accumulation of metal sulfides (e.g., pyrite) which upon drainage (natural or artificial) oxidize to form sulfuric acid and other acid sulfate weathering products (van Breeman, 1982). Sulfaquepts are important soils because of their land use, distribution, and behavior are conditioned by their acidic character.

6.5.3.2 Climate

Inceptisols occur in subhumid to humid climates from equatorial to tundra regions where, if they have permafrost within 2 m, they are replaced by Gelisols (Soil Survey Staff, 1998a,b). Inceptisols cannot have an aridic (torric) moisture regime unless they have an Anthropic epipedon upon which they become Anthrepts.

Soil temperature and moisture are used at the suborder level in *Soil Taxonomy* to reflect major controls on current soil-forming processes. Such controls not only influence rates, modes and mechanisms of physical and chemical weathering, but biological and geomorphic processes that influence relief, landscape stability, topography and soil drainage. Because Inceptisols form over a wide range of climatic conditions, excluding aridic and pergelic, the climatic impact on Inceptisols is similar to that of most other orders.

Soil moisture regime is used as a differentiation for four Inceptisol suborders (Aquepts, Udepts, Ustepts and Xerepts), soil temperature regime for one suborder (Cryepts), and past human cultural impacts for one suborder (Anthrepts). For Cryepts it was judged that temperature constraints on

Table 6.26 Listing of suborders, great groups and subgroups for Inceptisols Order (Soil Survey Staff, 1998a)

Suborder	Great Group	Subgroups
Aquepts	Sulfaquepts	Salidic, Hydraquentic, Typic
	Petraquepts	Histic Placic, Placic, Plinthic, Typic
	Halaquepts	Vertic, Aquandic, Duric, Aeric, Typic
	Fragiaquepts	Aeric, Humic, Typic
	Cryaquepts	Sulfic, Histic Lithic, Lithic, Vertic, Histic, Aquabdic, Fluvaquentic, Aeric, Humic, Aeric, Humic, Typic
	Vermaquepts	Sodic, Typic
	Humaquepts	Hydraquentic, Histic, Aquandic, Cumulic, Fluvaquentic, Aeric, Typic
	Epiaquepts	Vertic, Aquandic, Fluvaquentic, Fragic, Aeric, Humic, Mollic, Typic
	Endoaquepts	Sulfic, Lithic, Vertic, Aquandic, Fluvaquentic, Fragic, Aeric, Humic, Mollic, Typic
Anthrepts	Plagganthrepts	Typic
	Haplanthrepts	Typic
Cryepts	Eutrocryepts	Humic Lithic, Lithic, Andic, Vitrandic, Aquic, Oxyaquic, Lamellic, Xeric, Ustic, Humic, Typic
	Dystrocryepts	Humic Lithic, Lithic, Andic, Vitrandic, Aquic, Oxyaquic, Lamellic, Spodic, Xeric, Ustic, Humic, Typic
Ustepts	Durustepts	Typic
	Calcustepts	Lithic Petrocalcic, Lithic, Torrtic, Vertic, Petrocalcic, Gypsic, Aquic, Aridic, Udic, Typic
	Dystrustepts	Lithic, Andic, Vitrandic, Aquic, Fluventic, Oxidic, Humic, Typic
	Haplustepts	Aridic Lithic, Lithic, Udertic, Torrtic, Vertic, Andic, Vitrandic, Anthraquic, Aquic, Oxyaquic, Oxidic, Lamellic, Torrtic, Udifluventic, Fluventic, Gypsic, Haplocalcic, Calcic Udic, Calcic, Aridic, Dystric, Udic, Typic
Xerepts	Durixerepts	Aquandic, Andic, Vitrandic, Aquic, Entic, Typic
	Calcixerepts	Lithic, Vertic, Petrocalcic, Sodic, Vitrandic, Aquic, Typic
	Fragixerepts	Andic, Vitrandic, Aquic, Humic, Typic
	Dystroxerepts	Humic Lithic, Lithic, Aquandic, Andic, Vitrandic, Fragiaquic, Fluvaquentic, Aquic, Oxyaquic, Fragic, Fluventic Humic, Fluventic, Humic, Typic
	Haploxerepts	Humic Lithic, Lithic, Vertic, Aquandic, Andic, Vitrandic, Gypsic, Aquic, Lamellic, Fragic, Fluventic, Calcic, Humic, Typic
Udepts	Sulfudepts	Typic
	Durudepts	Aquandic, Andic, Vitrandic, Aquic, Typic
	Fragiudepts	Andic, Vitrandic, Aquic, Humic, Typic
	Eutruudepts	Humic Lithic, Lithic, Aquertic, Vertic, Andic, Vitrandic, Anthraquic, Fragiaquic, Fluvaquentic, Aquic Dystric, Aquic, Oxyaquic, Fragic, Lamellic, Dystric Fluventic, Fluventic, Arenic, Dystric, Rendollic, Humic, Ruptic-Alfic, Typic
	Dystrudepts	Humic Lithic, Lithic, Vertic, Aquandic, Andic, Vitrandic, Fragiaquic, Fluvaquentic, Aquic Humic, Aquic, Oxyaquic, Fragic, Lamellic, Humic, Psammentic, Fluventic Humic, Fluventic, Spodic, Oxidic, Humic Pachic, Humic, Ruptic-Alfic, Ruptic-Ultic, Typic

pedogenesis, use and management were more important than moisture constraints, even though both are important; the soil moisture bias was introduced at the great group level (Table 6.26). Soil temperature regime was used as a differentiation at the family level for all orders except Gelisols and those soils that have a cryic soil temperature regime. Anthrepts have no particular climatic bias and are of limited geographical extent.

Aquepts form under aquic or excess wetness conditions (saturation with water, periodic reduction, and formation of contemporaneous redoximorphic features, redox concentrations and/or depletions). Udepts form under better drained, strongly leaching environments, where precipitation exceeds evapotranspiration. These soils generally lack salts and carbonates in the solum and have undergone moderate to strong weathering. Generally, summer periods of soil moisture deficit are insufficient to negatively impact mesophytic agricultural crops. Ustepts are formed under drier climates with significant periods of summer soil moisture deficit. Precipitation is less than evapotranspiration, at least during significant periods of the summer. They commonly exhibit incomplete leaching, weak to moderate weathering intensities and pedogenic salts and/or carbonates in the lower solum. Moisture deficit is sufficiently severe to limit crop growth during summer months. Xerepts are similar to Ustepts but are formed in climates where winter precipitation is dominant. These soils are common in Mediterranean regions of Europe and along the Pacific coastal zones of western United States. The Cryepts serve as an interface between the cold soils of high latitudes with pergelic conditions (permafrost within 2m) and cold soils of temperate regions or orographic mountainous zones without permafrost within 2m.

In Table 6.25 the climatic effects on soil classification and corresponding properties are illustrated for a few pedons. Vertic Calcustepts (*Vernon series*) and Oxidic Haplustepts (*Kahana series*) are examples of soils with an ustic moisture regime and contrasting cambic horizon composition, development and chemical characteristics. The Calcustepts are incompletely leached, have carbonates throughout the pedon and have a clay-mineral composition that favors shrink/swell activity with changing moisture content. Comparisons between Fluventic Haploxerepts (*Vidhas series*) and Humic Dystrocryepts (*Moran series*) reflect contrasting temperature and precipitation regimes due to elevational differences. Although parent materials are different, climate is the major determinant governing increased organic matter contents in the Humic Dystrocryepts, which formed under udic moisture and cryic temperature regimes about 1050m elevation. This is compared to the Fluventic Haploxerepts formed under xeric moisture and thermic temperature regimes at an elevation of about 90 m. The Dystrocryepts have accumulated much greater organic carbon contents in the histic layer (Oi) overlying an umbric epipedon and a low base subsoil. In contrast, the Haploxerepts have accumulated much less organic matter in the ochric epipedon and the presence of carbonates distributed throughout the profile implies weak leaching potentials. Climate, in this example, is considered an intensity factor whereby the rate of soil development increases as temperature and rainfall increase, but decreases as either decreases. Limited soil moisture during periods of optimal soil weathering limits profile development in the Haploxerepts. In contrast, the Dystrocryepts have optimal moisture for pedogenesis, but the soil temperature is limiting.

6.5.3.3 Topography

The topography associated with many Inceptisols is a sloping to strongly sloping landscape subject to erosion rates that are equal to or greater than soil formation rates. Many Inceptisols occur on hillsides of major mountains throughout the world. Profile development here is constrained by unstable landforms and cool climates. Youthful Inceptisols are common also on nearly level fluvial, lacustrine and deltaic plains (or associated depressions) where profile development is limited by age of the sedi-

ments and/or shallow ground water tables. Coastal estuaries are also probable locations for Inceptisols, especially those where metallic sulfides and sulfates are likely to accumulate.

6.5.3.4 Vegetation

It is difficult to isolate vegetation as an independent soil-forming factor governing the synthesis of Inceptisols because vegetation is dependent on climate and topography. Interdependency of these factors is easily demonstrated by comparing the Humic Dystricrypts (*Moran series*) and Vertic Calciustepts (*Vernon series*) of Table 6.25. Dystricrypts support mountain meadow and alpine vegetative communities such as Kobres (*Kobresia sp.*), Asters (*Aster sp.*), and Yarrow (*Achillea sp.*) under cold, moist climates. Under hotter, drier conditions of the Calciustepts, native vegetation is mixed, midgrass prairie consisting of buffalograss (*Buchloe dactyloides*), grama grasses [blue (*Bouteloua gracilis*), hairy (*B. hirsuta*) and sideoats (*B. curtipendula*)] and tobosa grass (*Hilaria mutica*). Primary productivity, accumulation of soil organic matter, and leaching potentials are governed by effective soil moisture and temperature regimes. Effects on soil properties are remarkably evident in Table 6.25 for these two soils. Primary productivity for the Dystricrypts is limited by soil temperature regime while a seasonal soil moisture deficit limits primary productivity for Calciustepts. Higher organic matter accumulation, low turnover rates, and moderate to strong leaching potentials (acid subsoils) are consequences in Dystricrypts. In contrast, lower organic matter contents, high turnover rates and less effective leaching potentials (calcareous solum) occur in Calciustepts.

6.5.3.5 Time

Many examples are available in the literature to demonstrate that time is a soil-forming factor (Jenny, 1941, 1980). Some might argue that factors other than time have more influence on soil formation. While this may be true in some cases, time in combination with the intensity factor (climate) has been responsible for many attributes of Inceptisols. Inceptisols are usually considered immature soils, but some have advanced stages of development and are just short of completing their evolutionary cycle from Entisols to Oxisols. For example, Oxic Haplustepts are similar to Oxisols but either have oxic conditions too deep in the soil (greater than 150 cm) or have CEC charge characteristics just a little too high for oxic horizon requirements—a prerequisite for all Oxisols. Upon erosion and truncation, some Inceptisols have soil properties at depths associated with old stable landforms prior to dissection. Further, several taxonomic classes of Inceptisols have morphological features that could be interpreted either as youthful or an advanced stage of pedogenesis. Arenic Eutrudepts is one such example in which the soils may have developed from recent sandy sediments or may be residuum from weakly consolidated sandstone on old stable landforms.

6.5.4 Morphology

Inceptisols have many kinds of diagnostic horizons and epipedons but the most common sequence of horizons is an ochric or umbric epipedon overlying a cambic or fragipan horizon. However, a cambic horizon is not required if an umbric, histic, or plaggen epipedon is present or if there is a fragipan, duripan, placic, calcic, petrocalcic, gypsic, petrogypsic, salic or sulfuric horizon (Soil Survey Staff, 1998a,b). They can also have anthropic and mollic epipedons but not argillic, kandic, or natric horizons unless buried below recent sediments. Oxic horizons are permitted only if the upper boundary is deeper than 150 cm. Spodic horizons are permitted only if they are less than 10 cm thick or if the upper boundary is deeper than 50 cm.

It is clear from the above discussion that the number and kinds of processes active in Inceptisols and the definition of Inceptisols are unavoidably complex because the order serves dual functions in

grouping soils of minimal pedogenic development and bifurcating *Soil Taxonomy*. In summary, they are youthful soils in rejuvenated landscapes but pedogenetically advanced on more stable landforms. Their attributes are highly variable and seldom uniquely definitive.

6.5.5 Pedogenic Processes

Pedogenic processes common to Inceptisols have been briefly considered under Section 6.5.3. Such processes are covered in greater detail in Section E, Chapters 2 and 3. The processes germane to Inceptisols are combinations and complexes of subprocesses and reactions found in most other soil orders. The nature of these processes, including their terminology, are reviewed by Marbut (1921), Kellogg (1936), Byers et al. (1938), Simonson (1959), Arnold (1983), Buol et al. (1980), Fitzpatrick (1986), and Fanning and Fanning (1989).

Biological processes including establishment of macro and micro fauna and flora, microbial and fungal mineralization, decomposition (humification) and accumulation of organic matter, biotic transformations (e.g., ammonification, nitrification, denitrification and nitrogen fixation), and pedoturbation are important in epipedon and cambic horizon genesis. Chemical processes including hydration, hydrolysis, solution, mineral synthesis, and oxidation/reduction reactions are responsible for both epipedon and cambic horizon synthesis. Physical processes contribute to the formation of pedogenic structure (obliteration of at least 50% of the "rock" structure) and transport of materials within the solum. They include aggregation, expansion/contraction (shrink/swell), freeze-thaw, and mass transport by solution or suspension (translocation).

Formation of a cambic horizon requires conversion of rock structure to soil structure (peds) and evidence of at least one of the following: oxidation/reduction (aquic conditions); neoformation of clay; translocation of salts, carbonates, clay and/or organic metal complexes; or liberation of free iron oxides. The cambic horizon is a prerequisite for most Inceptisols, but exceptions occur for certain epipedons or subsurface diagnostic features (Section 6.5.3).

Illuviation and/or neosynthesis of clay and organic metal complexes cannot be of such magnitude that argillic, natric, kandic, or spodic horizons are formed, unless buried. Processes responsible for formation of oxic horizons include desilication, neoformation of low activity clays and/or residual concentration of secondary oxides and oxyhydroxides of iron and aluminum (Kellogg, 1936; Sivarajasingham et al., 1962; Cline, 1975 and Jenny, 1980). Acid sulfate soils are formed by processes termed sulfidization and sulfuricization. These processes are responsible for the accumulation of metal sulfides in soils and sediments, oxidation of these sulfides, formation of sulfuric acid, ferrolisis, soil acidification and mineral dissolution (Kittrick et al., 1982; van Breeman, 1982; Fanning and Fanning, 1989).

6.5.6 Classification

In the 1938 classification scheme (Baldwin et al., 1938) and in most subsequent revisions in the U.S. prior to *Soil Taxonomy*, Inceptisols were included in a number of great soil groups including Ando, Hydrol Humic Latosols, Humic Gley, Tundra, Half Bog, Sols Bruns Acides, Brown Forest, Regosols, Lithosols, Aluvial, and numerous other minor groups (Foss et al., 1983). Even at that time this was a very extensive and diverse taxa of soils. In the Canadian soil classification system, the wet Inceptisols (Aquepts) were classified as Gleysols but most of the better drained members were Brunisols. In the legend of the FAO/UNESCO soil map of the world (Section E, Chapter 5), most well-drained Inceptisols would be classed as Cambisols or Phaeozems where the base saturation was low. The wet Inceptisols would be classed mostly as Gleysols except those with acid sulfate conditions, which would be grouped with Fluvisols.

Inceptisol suborders, great groups and subgroups are presented in Table 6.26. Rationale for these taxa have been considered previously in this chapter and in Section E, Chapter 4. The following discussion of suborders has been taken directly from *Soil Taxonomy* (1998b) to illustrate the general nature and characteristics of Inceptisols suborders.

6.5.6.1 Anthrepts

These are the Inceptisols that are the more or less freely drained that have either an anthropic or a plaggen epipedon. Most of these soils have been used as cropland or places of human occupation for many years. They can have almost any temperature regime, and almost any vegetation. Most of them have a cambic horizon. Only two subgroups are recognized at this time, those with plaggen epipedons indicative of ancient cultural sites (*Plaggenanthrepts*), and those with anthropic epipedons reflecting other long-term human impacts on soils such as irrigated farmlands or housing areas. Most of these soils are in Eurasia or Northern Africa.

6.5.6.2 Aquepts

These are the wet Inceptisols. Their natural drainage is poor or very poor and, if they have not been artificially drained, groundwater is at or near the soil surface at some time during normal years but typically not during at all seasons. They mostly have a gray to black surface horizon and a gray subsurface horizon with redox concentrations that begin at a depth of less than 50 cm. A few have a brownish surface horizon that is less than 50 cm thick.

Most Aquepts have developed in late Pleistocene or younger deposits in depressions, nearly level plains, or flood plains. They occur from the Equator to latitudes with discontinuous permafrost. The common features of most of these soils are the gray and rusty colors of redoximorphic features at a depth of 50 cm or less and the shallow groundwater or artificial drainage. They may have almost any particle size class except fragmental, and any reaction class, any temperature regime except pergelic, and almost any vegetation. Most of them have a cambic horizon, and some have a fragipan. It is possible that some have a plaggen epipedon.

Table 6.26 lists the 9 Aquepts. These suborders are based upon limiting features or horizons that impact use, management and behavior. Specifically, these include soils with acid sulfuric horizons within 50 cm of the surface associated with oxidation of metal sulfides (*Sulfaquepts*); soils with a restrictive cemented subsurface horizon that forms a continuous phase within 100 cm of the soil surface (*Petraquepts*); salty (salic) or alkali (natric) subsoil horizons (*Halaquepts*); soils with a fragipan (*Fragiaquepts*); soils with a cryic temperature regime (*Cryaquepts*); soils with strongly bioturbated layers by macrofauna such as crayfish, worms, and mammals (*Vermaquepts*); soils with darkened, organic-enriched surface horizons (*Humaquepts*); soils with a perched water table (*Epiaquepts*); soils that are saturated from a groundwater source (*Endoaquepts*).

6.5.6.3 Cryepts

Cryepts are the cold Inceptisols of high mountains or high latitudes. They cannot have permafrost. The vegetation is mostly conifers or mixed conifers and hardwood trees. Few of them are cultivated. These soils may be formed in loess, drift, alluvium, or solifluction deposits, mostly late Pleistocene or Holocene in age. They commonly have a thin, dark brownish ochric epipedon and a brownish cambic horizon. Some have bedrock within 100 cm of the surface. In the United States these soils are moderately extensive in the high mountains of the West as well as other mountainous areas of southern Alaska and the world.

Two suborders of Cryepts are recognized (Table 6.26). Those that are calcareous or have high base saturation are *Eutocryepts*, and Cryepts with low base saturation are *Dystrocryepts*.

6.5.6.4 Udepts

Udepts are mainly the more or less freely drained Inceptisols that have a udic or perudic moisture regime. They formed on nearly level to steep surfaces mostly of late Pleistocene or Holocene age. Some, where the soil moisture regime is perudic, formed in older deposits. Most of them had or now have a forest vegetation, but some have shrub or grass. A few have been formed from Mollisols by truncation of the mollic epipedon (*Eutrudepts*), mostly under cultivation. Most of them have an ochric or umbric epipedon and a cambic horizon with low base saturation (*Dystrudepts*). These were the Sols Bruns Acides of earlier classification schemes (Foss et al., 1983). Some also have a sulfuric horizon (*Sulfudepts*), a fragipan (*Frugiudepts*), or a duripan (*Durudepts*). In the United States, Udepts are most extensive on the Appalachian Mountains, the Allegheny Plateau, and the West Coast; they also occur extensively in Eurasia,

6.5.6.5 Ustepts

Ustepts are the more or less freely drained Inceptisols that have a ustic moisture regime. They have dominantly summer precipitation or an isomesic, hyperthermic or warmer temperature regime. They formed mostly in Pleistocene or Holocene deposits or on steep slopes. Many of these soils are calcareous at a shallow depth and have a Bk or a calcic horizon (*Calciustepts*). A few have formed from Mollisols by truncation of the mollic epipedon, mostly under cultivation. Most of them have an ochric or umbric epipedon and a cambic horizon (*Haplustepts*). Some have a duripan (*Durustepts*), especially in areas where a labile silica source is associated with pyroclastic deposits or volcanic ash. Ustepts with low base saturation (*Dystrustepts*) occur in areas with polycycled preweathered acidic sediments or outcrops of acid bedrocks common in West Africa and isolated areas of the United States. The native vegetation commonly was grass, but some supported trees and savannas. Ustepts are of moderate extent in the United States. They are most common on the Great Plains mostly in Montana, Texas, and Oklahoma. They occur extensively in ustic sectors of the Americas, Eurasia, West Africa, Australia, and several island countries.

6.5.6.6 Xerepts

Xerepts are the more or less freely drained Inceptisols that have a xeric moisture regime, dominantly winter precipitation. They have a frigid, mesic, or thermic temperature regime. They formed mostly in Pleistocene or Holocene deposits or on steep slopes. Many of these soils are calcareous at a shallow depth and have a Bk, a calcic or a petrocalcic horizon (*Calcixerepts*). Others have low base saturation (*Dystrixerepts*). Most of them have an ochric or umbric epipedon and a cambic horizon (*Haploxerepts*). Some have a duripan (*Durxerepts*) and a few have a fragipan (*Fragixerepts*). The native vegetation was commonly coniferous forest on those with a thermic temperature regime. These soils are of moderate extent in the United States. They are most common near the West Coast in the states of California, Oregon, Washington, Idaho, and Utah. They are major soils of the Mediterranean regions of Eurasia and Northern Africa.

6.5.7 Physical, Chemical and Mineralogical Properties

Because of the diverse nature of Inceptisols, few generalized statements can be made about their physical, chemical, biological, and mineralogical properties. Their properties are nearly as inclusive as all the other soil orders collectively. Inceptisols span the global regions from intensively weathered to minimally developed soils. For example, they are acidic to alkaline in reaction, weak to strongly physico-chemically buffered, have low to high organic matter contents, low to high hydraulic transport functions, low to high water retention values, fertile to infertile in nutrient status, etc. The Soil

Survey manual (Soil Survey Staff, 1993) identify class ranges for the above classes and for many other soil attributes described and measured. Table 6.25 illustrates the kind and magnitude of diversity vested in many of these Inceptisol attributes. The database on-line for Inceptisols (<http://www.statlab.iastate.edu/soils/soildiv/>) further documents this aspect.

The reader is referred to discussions of other orders in Sections 6.2-6.13 for physical, chemical, biological, and mineralogical attributes likely to be associated with Inceptisols. For example, those Inceptisols with andic surface materials are similar to Andisols; those with vertic shrink/swell features to Vertisols; those with calcic, gypsic, salic, petrocalcic and petrogypsic horizons to Aridisols; those with translocation and/or neosynthesis of clay and organic metal colloids to Alfisols, Ultisols and Spodosols; those with high sesquioxide contents and/or low activity clay mineral suites to Oxisols; and those with cryic temperature regimes to Gelisols.

The sand and silt mineralogy commonly consists of quartz, mica and feldspars with minor components of weatherable heavy minerals (opaques). The opaque minerals in soils derived from basic bedrocks are feldspars, hematite, magnetite, ilmenite and rutile. Inceptisols rich in sesquioxides (Oxic subgroups) are comprised of oxyhydroxides of Fe, Al and Ti including gibbsite, hematite, goethite, boehmite, rutile and anatase. Those soils associated with drier climates (Ustepts and Xerepts) contain soluble salts (e.g., NaCO_3 , Na_2SO_4), gypsum and carbonates (calcite and dolomite). Inceptisols with andic and vitrandic materials (pyroclastics) contain amorphous or short-range order minerals such as allophane, ferrihydrite and glasses. Phyllosilicate clay minerals range from smectite, mica, kaolinite, chlorite, vermiculate and mixed layer assemblages to rather unique suites dominated by serpentine and glauconite. In acid sulfate soils (e.g., Sulfaquepts and Sulfudepts), jarosite is a common constituent that marks very acid conditions associated with metal sulfide oxidation. Pedogenic gypsum is also common in these soils.

6.5.8 Management

For management purposes, Inceptisols can be subdivided into three land uses, namely forestry, pasture production and agronomic cropping.

6.5.8.1 Forestry

Most Inceptisols under forested land use occur in mountainous regions on slopes ranging from 3 to 90%. On steep terrains, management systems other than natural regrowth are environmentally unacceptable and practically impossible. Most forested Inceptisols have carbon contents sufficiently high as long as the surface is not eroded, that indigenous nutrient recycling is sufficient to sustain growth without external fertilizer amendments. On less sloping terrain, many large, commercial timber companies find fertilizer amendments, especially phosphorous applications, to be economically beneficial. Management techniques to control competition from unwanted species and disease suppression commonly involve controlled burns or physical removal of dead or unwanted species. Harvesting methods depend on the slope gradient; they range from surface skidding to aerial removal by helicopter or cable lifts. Harvest schemes may involve clear cutting or selective harvesting. Damages from erosion and compaction of harvest operations on less sloping Inceptisols are commonly ameliorated with surface tillage operations to break up surface crusts, compacted zones and to establish soil and water conservation buffers.

6.5.8.2 Pasture Production

Most of the Inceptisols under pasture management are planted to improved pastures that respond to N, P and K fertilization, especially if the pasture is intensely managed for animal or forage produc-

tion. Areas, which occur under native range, or under shifting traditional agriculture as in developing countries, should be managed using best management practices for stocking rate, pasture rotation and fallow period. These vary from one region to another, not only in terms of soil conditions but the ability of the operator to provide investment inputs. Commonly under traditional agriculture, the inputs are minimum and the soil dictates the best management and utilization practices to be followed. Inceptisols with an ustic or xeric moisture regime are limited by insufficient rainfall. Irrigation may be used to supplement natural precipitation on some Inceptisols, but the cost of this investment for improved pastures may make it economically unattractive. However, many examples of irrigated improved pastures occur in Europe and the United States where the economics are favorable and adequate high quality aquifers are available. Surface or subsurface drainage may be required to remove surface water or lower the watertable for optimal forage production, especially for Aquepts.

6.5.8.3 Agronomic Cropping

Inceptisols are the major soils on which agronomic crops are produced in some parts of the world. Aquepts, Cryepts, Udepts, Ustepts, and Xerepts are very productive and valuable agricultural land resources, if properly managed. Crops produced depend on climate (length of growing season, dependability of precipitation, seasonal periods of soil moisture excess or deficit, photoperiodism, degree days, etc.), but because of Inceptisol diversity, most major food, feed and fiber crops are included. Irrigation is used on vegetable, citrus, and other important cash or speciality crops when summer periods of moisture deficit become extreme. The highly weathered Inceptisols within the tropics are used for sugarcane, pineapple, cotton and some upland rice production with many of these crops being produced under irrigation. Most of these soils respond to N, P, and K fertilizers and some soils require lime in order to neutralize acidity and sustain production.

Cryepts are used mostly for cereal crops because of the short growing season, photoperiodism and few number of degree days. Sometimes these soils are used for cool season vegetables and forage production. Where these soils interface with Gelisols, they are constrained by cold soil temperatures, freeze-thaw heaving, and erosion induced by wind, water and mass movement (solifluction). They may be under alpine meadow.

Aquepts used for agronomic production usually require surface or internal drainage but depending on soil temperature, a wide range of crops are grown including cotton, sugarcane, rice and corn in the subtropical and tropical climates and corn, soybeans, sorghum and some cereal crops in temperate climates. Sulfaquepts (and Sulfudepts) require specialized management because when they are drained mainly for lowland rice production, the sulfuric horizon produces large amounts of sulfuric acid. Because the reactions involved were not well understood when most of these soils were drained, vast wastelands were created. Management schemes, which reduce acid production, have been developed by Ponnampetuma et al. (1973), van Breeman (1982) and Coly (1996). Reclamation and management practices for acid sulfate soils have been proposed by van Breeman (1982), Rimwanich and Suebsuri (1983), Cisse et al. (1993) and Coly (1996). They are discussed in detail in Section G, Chapter 6. Briefly the formation of a sulfuric horizon should be prevented or minimized by controlling the depth of the water table upon drainage (keeping the soils saturated to prevent pyrite oxidation). Sulfaquepts can be reclaimed by inducing reducing conditions within the sulfuric horizon and/or flushing the soil to remove the sulfuric acid. The flushing can be accomplished using a combination of saline and fresh water to remove the soluble salts. The addition of CaCO_3 is required to neutralize any acidity produced and CaSO_4 is required when saline water has been used for reclamation to remove excess sodium (Cisse et al., 1993; Coly, 1996).

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6.6 Gelisols

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6.6.1 Introduction

Gelisols, which are the permafrost-affected soils, constitute the twelfth and newest soil order. They comprise 18 million km² or about 13% of the Earth's land surface and occur in the Arctic, Antarctic, Subarctic, Boreal, and some alpine regions under cold continental, subhumid or semi-arid, and arid conditions (Bockheim et al., 1994). They support unvegetated to continuously vegetated tundra, subarctic and boreal forest, and some alpine tundra. Gelisols are of global concern because they contain many protected areas, support numerous indigenous populations who depend on the land and surrounding oceans for sustenance, and may be subject to considerable impacts from human development (oil, coal and gas exploration and mining) and global warming.

Gelisols are defined as soils having permafrost within 100 cm of the soil surface, or gelic materials within 100 cm of the soil surface and permafrost within 200 cm of the soil surface (Soil Survey Staff, 1998). Gelic materials, in turn, are seasonally or perennially frozen mineral or organic soil materials that have evidence of cryoturbation (frost churning), ice segregation, or cracking from cryodesiccation. Gelic materials contrast with other kinds of materials, such as andic or spodic materials, in being defined entirely on the basis of physical and thermal characteristics, rather than chemical properties.

A representative Gelisol is shown in Fig. 6.24. Gelic materials occur in both the active layer and the upper part of the permafrost as evidenced by cryoturbation, denoted in soil descriptions by the subscript *jj* (*y* in the Canadian system).

6.6.2 Permafrost and the Occurrence of Gelisols

Gelisols only occur in areas containing permafrost within 100 cm of the soil surface if the soil is not cryoturbated or 200 cm of the surface if the soil is cryoturbated. Permafrost is defined here as a thermal

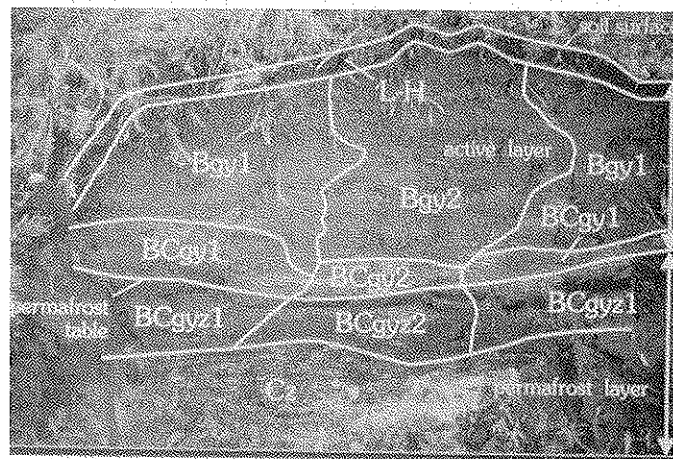


Fig. 6.24 An Aquaturbel developed on an earth hummock in northern Canada. The Canadian soil horizon nomenclature is used, where L-H = O, *y* = *jj*, and *z* = *f* in *Soil Taxonomy*.

condition in which a material (including soil materials) remains below 0°C for two or more years in succession. Permafrost may be ice-cemented (designated in soil descriptions with the subscript *f*) or in the case of insufficient interstitial water, dry (designated as *ff*). In the frozen layer, a variety of ice lenses, vein ice, segregated ice crystals, and ice wedges are evident. An important consideration is that the permafrost is in dynamic equilibrium with the environment.

Permafrost, which comprises about 26% of the Earth's surface, including the Antarctic and Greenland ice sheets (Black, 1954), is differentiated into two broad zones on land in the circumpolar regions: the continuous and the discontinuous (Péwé, 1991). In the continuous zone, permafrost is present nearly everywhere except under large lakes and rivers that do not freeze to the bottom (Fig. 6.25). The discontinuous zone is divided into the widespread discontinuous permafrost subzone and the sporadic discontinuous subzone in the northern and southern parts, respectively.

The distribution and thickness of permafrost are influenced by natural surface features such as snow and vegetation cover, topography, and bodies of water, but as Fig. 6.25 shows, are most affected by regional climate.

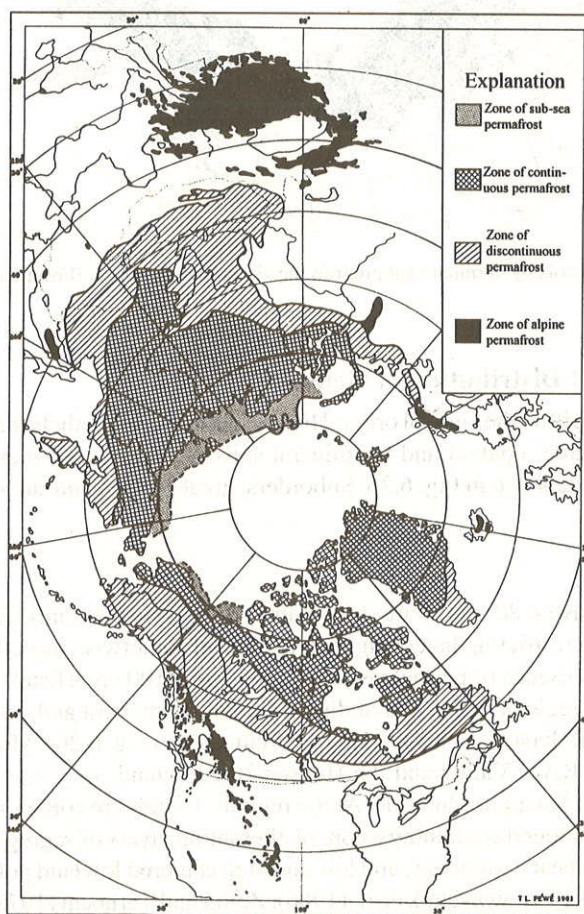


Fig. 6.25 Distribution of permafrost in the Northern Hemisphere [From Péwé, 1983. *Arctic Alpine Res.* 15:146 with permission from the Regents of the University of Colorado]

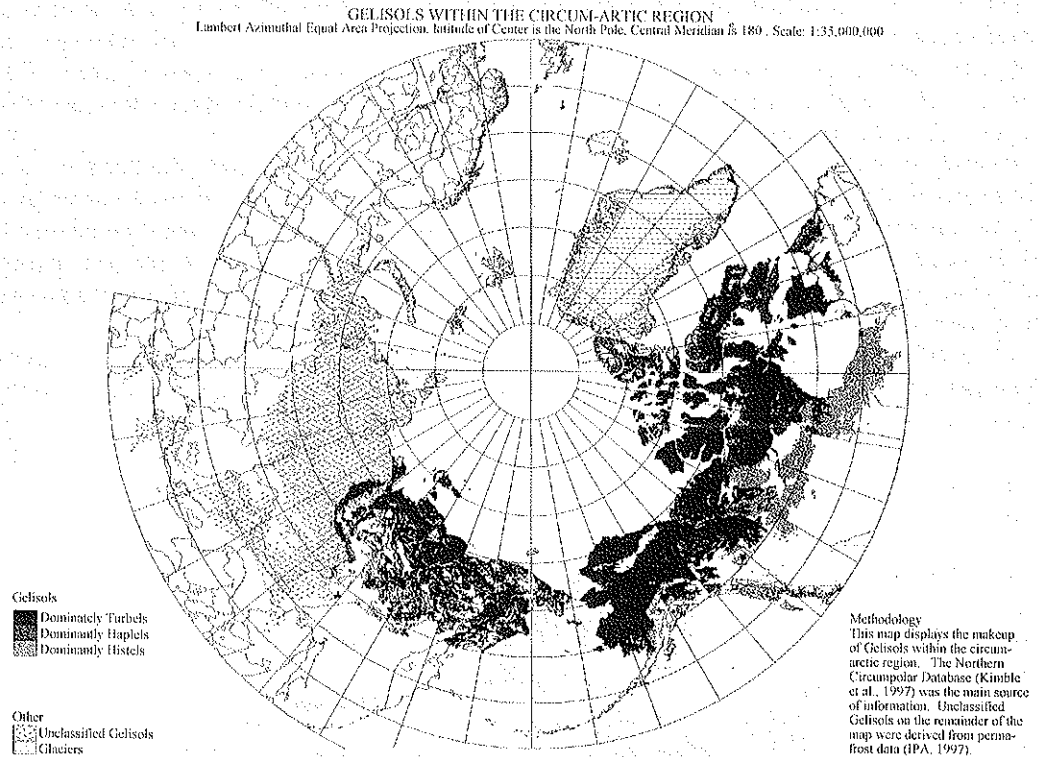


Fig. 6.26 Distribution of Gelisols by dominant suborder in the circum-arctic region [From Kimble et al., 1997 with permission]

6.6.3 Description and Distribution of Gelisols

There are three suborders within the Gelisol order: Histels, Turbels, and Orthels that are differentiated on the basis of organic matter content and for mineral soils whether or not there is cryoturbation. Distribution of Gelisols is depicted in Fig. 6.26. Suborders, great groups, and subgroups are shown in Table 6.27.

6.6.3.1 Histels

Histels are Gelisols that have $\geq 80\%$ or more (by volume) of organic materials from the soil surface to a depth of 50 cm, or to a restricting layer (Fig. 6.27A). Histels otherwise meet the requirements of a Histosol except for the presence of permafrost within the upper 100 cm. Histels occur primarily in Subarctic and Low Arctic regions in the area of discontinuous permafrost and comprise 5.3% of the Gelisols in areas where soils have been mapped in the circumarctic (Fig. 6.26). Most of these soils are located in the Mackenzie River Valley and the Hudson Bay lowlands of Canada. In general, they mostly occur in the Boreal, Subarctic, and Low Arctic regions. Histels are commonly associated with palsas (a peaty permafrost mound containing a core of alternating layers of segregated ice and peat or mineral soil), peat plateau, peat hummocks, and low and high centered lowland polygons. These soils have been discussed fully by Holowaychuk et al. (1966), Zoltai and Tarnocai (1971), Tarnocai (1972, 1973), Zoltai and Tarnocai (1974, 1975), Everett (1979), and the National Wetlands Working Group (1988).

Table 6.27 Listing of suborders, great groups and subgroups of Gelisols order [Soil Survey Staff, 1998]

Suborder	Great Group	Subgroups
Histels	Folistels	Lithic, Glacic, Typic
	Glacistels	Hemic, Sapric, Typic
	Fibristels	Lithic, Terric, Fluvaquentic, Sphagnic, Typic
	Hemistels	Lithic, Terric, Fluvaquentic, Typic
	Sapristels	Lithic, Terric, Fluvaquentic, Typic
Turbels	Histoturbels	Lithic, Glacic, Ruptic, Typic
	Aquiturbels	Lithic, Glacic, Sulfuric, Ruptic-Histic, Psammentic, Typic
	Anhyturbels	Lithic, Glacic, Petrogypsic, Gypsic, Nitric, Salic, Calcic, Typic
	Molliturbels	Lithic, Glacic, Vertic, Andic, Vitrandic, Cumulic, Aquic, Typic
	Umbriturbels	Lithic, Glacic, Vertic, Andic, Vitrandic, Cumulic, Aquic, Typic
	Psammoturbels	Lithic, Glacic, Spodic, Typic
	Haploturbels	Lithic, Glacic, Aquic, Typic
	Orthels	Lithic, Glacic, Ruptic, Typic
Orthels	Aquorthels	Lithic, Glacic, Sulfuric, Ruptic-Histic, Andic, Virandic, Salic, Psammentic, Typic
	Anhyorthels	Lithic, Glacic, Petrogypsic, Gypsic, Nitric, Salic, Calcic, Typic
	Mollorthels	Lithic, Glacic, Vertic, Andic, Vitrandic, Cumulic, Aquic, Typic
	Umbrorthels	Lithic, Glacic, Vertic, Andic, Vitrandic, Cumulic, Aquic, Typic
	Argorthels	Lithic, Glacic, Natric, Typic
	Psammorthels	Lithic, Glacic, Spodic, Typic
	Haplorthels	Lithic, Glacic, Aquic, Typic

Histels are divided into five great groups, including the Folistels, Glacistels, Fibristels, Hemistels, and Sapristels. With the exception of the Glacistels, these great groups follow the suborders in the Histosol order (Section E, Chapter 6.2). The Glacistels contain a glacic layer within 50 cm of the surface that is 30 cm or more thick and contains 75% or more ice (by volume).

6.6.3.2 Turbels

Turbels are mineral soils that occur in areas with patterned ground. Patterned ground is a general term for any ground surface with a discernibly ordered, more or less symmetrical, morphological pattern of ground and, where present, vegetation (Washburn, 1980). They show marked influence of cryoturbation (Fig. 6.27B). These soils occur throughout the circumpolar regions and are the dominant suborder, accounting for more than 57% of the Gelisols mapped in the circumarctic (Fig. 6.26). Field and laboratory data for these soils are contained in publications by Douglas and Tedrow (1960), Day and Rice (1964), Holowaychuk et al. (1966), Everett (1968), Tedrow et al. (1968), Tedrow (1970), Zoltai and Tarnocai (1974), Pettapiece (1975), and Tarnocai and Smith (1992).

Most horizons and layers within the active layer of Turbels are strongly affected by cryoturbation. Turbels commonly contain irregular or broken horizons, involutions, organic matter that usually accumulates on the surface of the permafrost, oriented rock fragments, and silt caps and silt-enriched subsoil horizons. Permafrost occurs within 200 cm of the soil surface. These soils are differentiated into seven great groups that link them with other orders not containing permafrost, including the Histoturbels (40 to 80% organic materials by volume in the upper 50 cm), Aquiturbels (aquic conditions), Anhyturbels (anhydrous conditions) [Anhydrous conditions refer to soils of cold deserts

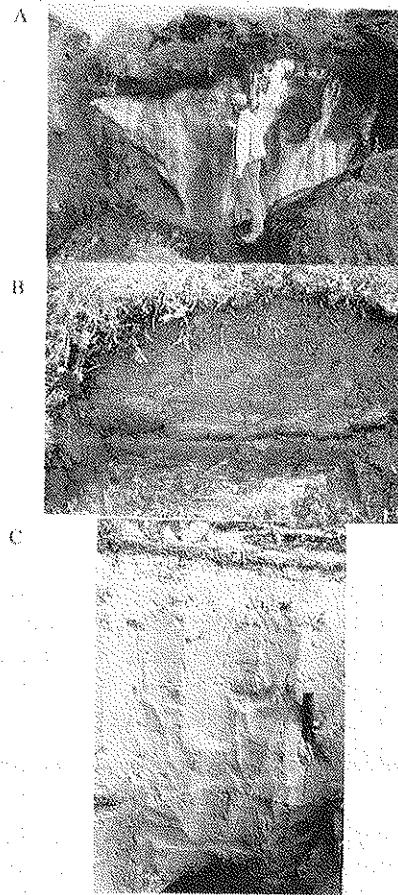


Fig. 6.27 Examples of the three suborders of Gelisols: (A) Histel, (B) Turbel, and (C) Orthel.

and other areas with permafrost (often dry permafrost) and low precipitation (usually $< 50 \text{ mm yr}^{-1}$ water equivalent). Anhydrous soil conditions are similar to the aridic and torric soil moisture regimes except that the soil temperature is $< 0^{\circ}\text{C}$ (Soil Survey Staff, 1998).], Molliturbels (mollic epipedon), Umbrturbels (umbric epipedon), Psammoturbels (sandy texture), and Haploturbels (other Turbels).

6.6.4 Orthels

Orthels are mineral soils containing permafrost within the upper 100 cm, but that lack cryoturbation (Fig. 6.27C). These soils comprise $< 3.6\%$ of the Gelisols mapped in the circumarctic and occur primarily in areas with dry permafrost such as floodplains and the dry valleys of Antarctica. They are divided into great groups that parallel the Turbels, i.e., Historthels, Aquorthels, etc. Subgroups within each great group of Gelisols are listed in Table 6.27.

6.6.5 Cryopedogenic Processes

Cryopedogenic processes that lead to gelic materials are driven by the physical volume change of water to ice, moisture migration along (1) thermal, (2) hydrostatic pressure, (3) solute concentration,

and (4) electrical potential gradients in the frozen (or unfrozen) system (Marion, 1995), or thermal contraction of the frozen material by continued rapid cooling. These processes include freezing and thawing, cryoturbation, frost heaving, cryogenic sorting, thermal cracking, and ice segregation (Tedrow, 1977; Washburn, 1980) (Fig. 6.28).

It should be emphasized that cryopedogenic processes are soil-forming processes characteristic of soils with permafrost and should not be viewed as operating against the other soil-forming processes in lower latitude soils; rather, they are distinctive processes producing horizons and properties that are uncommon to other soil orders. Processes common to the other soil orders operate in Gelisols but at a lesser magnitude because of the dominance of cryopedogenic processes.

6.6.6 The Pedon as a Basic Soil Unit

The pedon is the basic soil unit for sampling in *Soil Taxonomy* (1975); it is an especially important concept for describing, classifying, and sampling Gelisols. The pedon is defined so as to encompass the full cycle of patterned ground with a 1 or 2 m linear interval or a half cycle with a 2 to 7 m cycle (Fig. 6.29A). This interval is suitable for most patterned ground features such as earth hummocks, circles, nets, and nonsorted polygons. In the case of large scale (> 7 m) ice-rich polygons, such as those that occur along the Alaskan Coastal Plain, two pedons are delineated: one within the center of the polygon and the other within the ice wedge (Fig. 6.29B). If no patterned ground exists, the pedon is arbitrarily selected but approximates about 1 m² in area.

Scaled sketches of a pedon showing soil horizons, including patches of cryoturbated material, should be drawn on graph paper in the field, or Polaroid photographs should be taken and the horizons annotated directly on the print (Fig. 6.24). Samples should be collected from each horizon across the pedon and composited for subsequent laboratory characterization. In the case of highly cryoturbated soils, the areal percentage of each horizon is reported in soil descriptions rather than depth intervals (Kimble et al., 1993).

6.6.7 Historical Approaches to Classification of Permafrost-Affected Soils

Tedrow (1977) provided a comprehensive review of early approaches to classification of soils in the cold regions, emphasizing the zonal systems. Of the soil taxonomic systems used today, only the Canadian (Agriculture Canada Expert Committee on Soil Survey, 1987) and United States (Soil Survey Staff, 1998) approaches recognize soils with permafrost in a separate order. The World Reference Base (WRB) for Soil Resources (Section E, chapter 5.4) is considering permafrost-affected soils in a separate soil unit (Tarnocai, 1997, personal communication). The proposed WRB system has received the preliminary approval of the international soil community, and is remarkably parallel to the Canadian and United States taxonomic systems with regard to permafrost-affected soils (Table 6.28).

Whereas permafrost-affected soils are identified as Gelisols (from the Greek word, *gelid*, meaning very cold) in *Soil Taxonomy*, they are called Cryosols (from the Greek word, *kraios*, meaning cold or ice) in the other two systems. Each system divides permafrost-affected soils into three categories: organic soils, cryoturbated mineral soils, and other mineral soils. The soils are further delineated on the basis of key properties that link them with soils of lower latitudes.

6.6.8 Properties of Gelisols

Gelisols/Cryosols encompass a vast array of soils in terms of chemical and mineralogical properties. However, with the exception of the Histels, Gelisols are uniform in terms of physical and thermal properties.

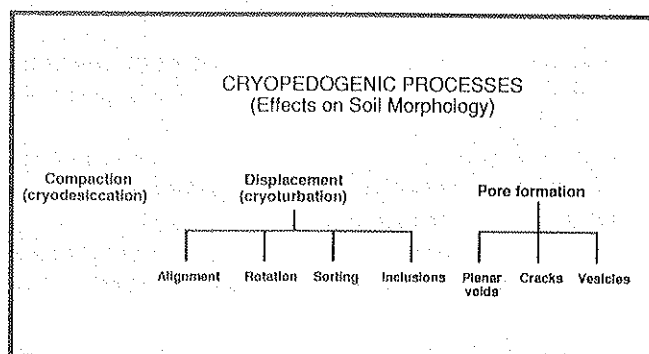


Fig. 6.28. Conceptual framework showing the interrelationships of the effects of cryopedogenic processes on development of specific fabric types in Gelisols [From Fox, 1994 with permission]

6.6.8.1 Macromorphology

The most common macroscopic soil features are due to cryoturbation and include irregular or broken horizons and incorporation of organic matter in lower horizons, especially along the top of the permafrost. Oriented stones and displacement of soil materials are common in Gelisols. Freezing and thawing produce granular and platy structures in surface horizons and blocky, prismatic or massive structures in subsurface horizons. The massive structure is due to cryostatic pressure and desiccation that develop when the two freezing fronts, one from the surface and the other from the permafrost, merge during freeze back in the autumn. The perennially frozen layer commonly contains ground ice in the form of segregated ice crystals, vein ice, ice lenses and wedges, and thick ground ice.

The granular, platy, or blocky structures of the surface mineral horizons are also the result of cryopedogenic processes such as the freeze-thaw process and vein-ice formation (ice segregation process). The subsurface horizons often have massive structures and are associated with higher bulk densities, especially in fine-textured soils.

Almost all Gelisols contain ice in the form of crystals, lenses, layers (vein ice), wedges, or massive ground ice, often to a thickness of several m. Soil texture is one of the factors controlling ice content in mineral soils with fine-textured Gelisols generally having higher ice contents than coarse-textured soils. Although coarse-textured soils have a relatively low ice content, they are often associated with ice wedges in the form of polygons. Histels have an ice content of 60 to 90% on a volume basis (Zoltai and Tarnocai, 1975).

The active layer that is subject to annual thawing and refreezing lies above the permafrost and not only supports biological life, but also protects the underlying permafrost. The thickness of the active layer is controlled by soil texture and moisture, thickness of the surface organic layer, vegetation cover, aspect and latitude.

Dilatancy, often confused with thixotropy, is common in soils with high silt content, greatly affects the trafficability of the soil, and is frequently present during the thaw period. Dilatancy is a property of granular masses expanding due to the increase of space between rigid particles upon displacement of the particles. When dilatant soils dry out, a characteristic vesicular porosity develops.

Salt crusts, patches, and pans are common in polar desert soils of the high arctic and cold desert soils of Antarctica (Tedrow, 1977; Bockheim, 1997).

6.6.8.2 Micromorphology

When viewed in thin sections, Gelisols contain a variety of fabrics resulting from desiccation and displacement due to alignment, rotation, sorting, and inclusions (Fig. 6.28). These features are

accompanied by planar voids, cracks and vesicles during pore formation. More specifically, the fabric of Gelisols varies from granular (granitic and granoidic) in the surface horizons to mainly porphyroskelic with fragmic and fragmoidic components in subsurface mineral horizons (Pawluk and Brewer, 1975; Fox, 1985; Smith et al., 1991). The micromorphology of Gelisols can show evidence of matrix displacement and movement, with resultant reorganization of skeleton grains into circular or elliptical patterns, producing an orbicular fabric (Fox and Protz, 1981). Ice lensing and vein-ice development lead to the formation of lenticular fabrics, while cryoturbation and cryodesiccation can lead to granic or granoidic fabrics (Smith et al., 1991). In addition, suscitic and conglomeric fabrics also are common in Gelisols (Fox and Protz, 1981).

6.6.8.3 Thermal Characteristics

Soil temperatures for a Gelisol located north of the arctic treeline in Canada are shown in Fig. 6.30. The unique thermal signature that separates Gelisols from all other soils is the presence of a perennially frozen layer, usually below 50 cm. Because of this frozen layer, Gelisols have a steep

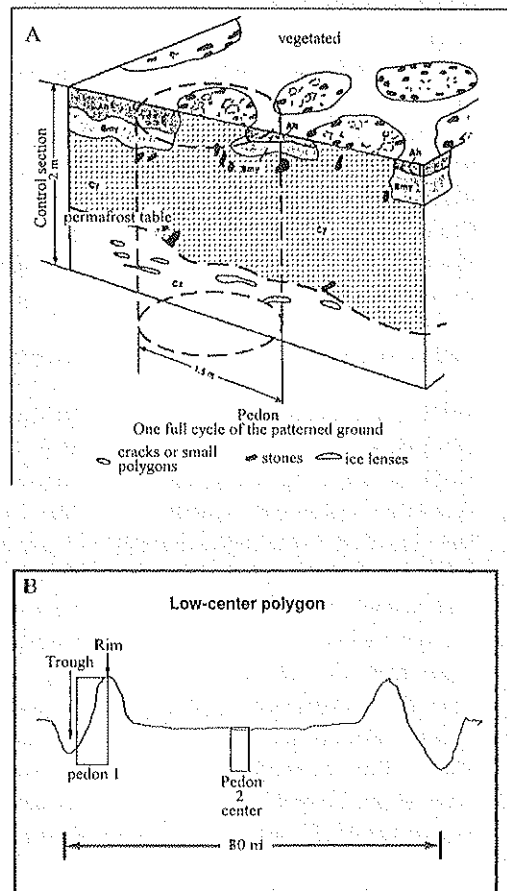


Fig. 6.29 The pedon concept as it applies to Gelisols and form of patterned ground, including (A) a nonsorted circle [Tarnocai and Smith, 1992] and (B) a large-scale, low-centered nonsorted polygon. The Canadian soil horizon nomenclature is used, where Ah = A, Bmy = Bwjj, Cy = Cjj, and Cz = Cf.

Table 6.28 Comparison of soil taxa among *Soil Taxonomy*, the Canadian system, and the proposed World Reference Base (FAO)

<i>Soil Taxonomy</i> (Soil Survey Staff, 1998)	World Reference Base	Canadian (Agriculture Canada Expert Committee on Soil Survey, 1987)
Gelisol (order)	Cryosol (soil unit)	Cryosol (order)
Histels (suborder)	Histic Cryosols (soil subunit)	Organic Cryosols (great group)
Folistels	(no equivalent)	(no equivalent)
Glacistels	Glacic Histic Cryosols	Glacic Organic Cryosols
Fibristels	Fibric Histic Cryosols	Fibric Organic Cryosols
Hemistels	Mesic Histic Cryosols	Mesic Organic Cryosols
Sapristels	Humic Histic Cryosols	Humic Organic Cryosols
[Terric Fibristels]	(no equivalent)	Terric Fibric Organic Cryosols
[Terric Hemistels]	(no equivalent)	Terric Mesic Organic Cryosols
[Terric Sapristels]	(no equivalent)	Terric Humic Organic Cryosols
Turbels	Turbic Cryosols	Turbic Cryosols
Histoturbels	Stagnic & Gleyic Turbic Cryosols	Gleysolic Turbic Cryosols
Aquiturbels	Stagnic & Gleyic Turbic Cryosols	Gleysolic Turbic Cryosols
Anhyturbels	Salic, Gypsic & Calcic Turbic Cryosols	(no equivalent)
Molliturbels	Mollic Turbic Cryosols	Brunisolic Turbic Cryosols
Umbristurbels	Umbric Turbic Cryosols	Brunisolic Turbic Cryosols
Psammoturbels	Regic Turbic Cryosols	Regosolic Turbic Cryosols
Haploturbels	Regic Turbic Cryosols	Orthic Turbic Cryosols
[Glacic subgroups]	Glacic Turbic Cryosols	(no equivalent)
[Sulfuric Aquaturbels]	Thionic Turbic Cryosols	(no equivalent)
[Andic Molli- & Umbristurbels]	Andic Turbic Cryosols	(no equivalent)
[Vitrandic Molli- & Umbristurbels]	Tephric Turbic Cryosols	(no equivalent)
[Spodic Psammoturbels]	Spodic Turbic Cryosols	(no equivalent)
Orthels	Haplic Cryosols	Static Cryosols
Historthels	Stagnic & Gleyic Haplic Cryosols	Gleysolic Static Cryosols
Aquorthels	Stagnic & Gleyic Haplic Cryosols	Gleysolic Turbic Cryosols
Anhyorthels	Salic, Gypsic & Calcic Haplic Cryosols	(no equivalent)
Mollorthels	Mollic Haplic Cryosols	Brunisolic Static Cryosols
Umbrorthels	Umbric Haplic Cryosols	Brunisolic Static Cryosols
Argiorthels	(no equivalent)	(no equivalent)
Psammorthels	Regic Haplic Cryosols	Regosolic Static Cryosols
Haploorthels	Regic Haplic Cryosols	Regosolic Static Cryosols
[Glacic subgroups]	Glacic Haplic Cryosols	(no equivalent)
[Sulfuric Aquaorthels]	Thionic Haplic Cryosols	(no equivalent)
[Andic Molli- & Umbrorthels]	Andic Haplic Cryosols	(no equivalent)
[Vitrandic Molli- & Umbrorthels]	Tephric Haplic Cryosols	(no equivalent)
[Spodic Psammorthels]	Spodic Haplic Cryosols	(no equivalent)

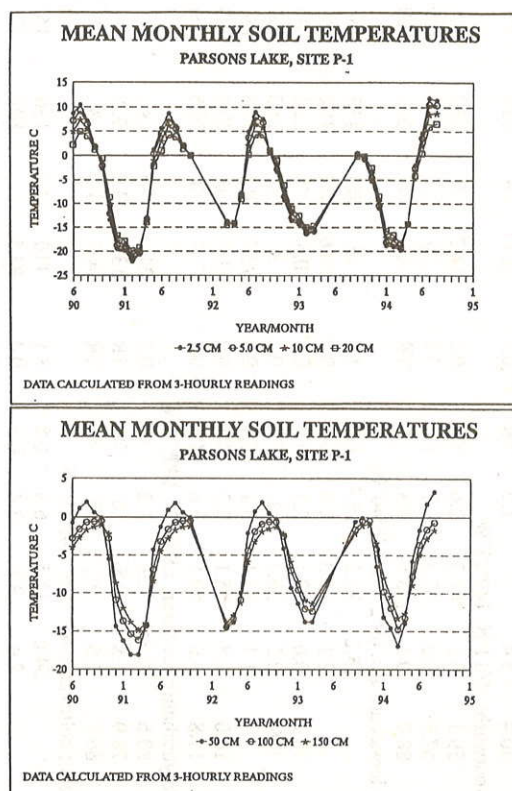


Fig. 6.30 Soil temperature variation in a Turbel from Parson Lake, Canada (68°58'09'N, 133°32'54'W)

vertical temperature gradient. If these soils are associated with certain types of patterned ground, the horizontal temperature gradient can also be large. For example, in the case of Gelisols associated with earth hummocks, the soil temperature at the center of the hummock can decrease from 12 °C at the surface to 0 °C at 50 cm during the summer months. Soil temperatures at comparable depths under the interhummock depression, < 1 m away and at equivalent depths, can be 5 to 7 °C lower (Tarnocai and Zoltai, 1978).

6.6.8.4 Physical and Chemical Characteristics

Histels contain weakly to well-decomposed moss, sedge, woody and amorphous peats. The pH of these soils ranges from 2.5 to 7.0 (Table 6.29). Histels derived from moss peats (Fibristels) tend to have lower pH values than other organic soils with permafrost. Turbels and Orthels developed from calcareous parent materials have a $\text{pH}_{\text{CaCl}_2}$ of 7 and a high base saturation. On the other hand, where these soils are developed from acidic parent materials, $\text{pH}_{\text{CaCl}_2} < 5.5$.

Anhyturbels and Anhyorthels in polar and cold deserts often have high salt contents and electrical conductivities. Although these soils have many features common to Aridisols (Section E, Chapter 6.10), they have patterned ground, dry permafrost in the upper meter over ice-cemented permafrost, and show evidence of thermal cracking (Bockheim, 1997).

Gelisols, especially the Histels and Turbels, often contain large amounts of organic C in the upper m, with values commonly ranging from 3 kg C m⁻³ in Haplorthels to over 100 kg C m⁻³ in Sapristsels

Table 6.29 Analytical characteristics of selected Gelisols

Horizon	Depth cm	pH	BD Mg m ⁻³	C	N	Clay %	Silt	Sand	Extr. P	mg L ⁻¹				cmol _c kg ⁻¹				Base	
										Ca	Mg	Na	K	Ex. Acid.	CEC	Ex. Al	NH ₄ OAc	Sat.	%
Typic Hapluturbel; Dry Acidic Tundra (Empetrum-Betula-Dryas-Arctostaphylos); till/congelifractate; 68°37'N, 149°18'W																			
Ajj	2-46	4.91	0.56	8.68	0.53	9.8	53.2	37.0	0.7	2.4	1.2	TR [†]	0.1	39.4	11.5	36.1		10	
Bwj	4-40	4.83	0.87	2.3	0.13	18.4	46.2	35.4	1.2	0.7	0.8	TR	TR	25.2	9.1	20.5		8	
BCjj	0-34	5.14	0.68	9.31	0.26	11.0	55.3	33.7	3.5	0.7	0.4	0.1	0.1	34.5	9.7	26.6		5	
C	8-90	6.94	nd	0.58	0.05	5.2	44.4	50.4	5.4	7	1.2	TR	0.1	3.1	0	10.8		77	
Typic Molliturbel; Dry Nonacidic Tundra (Carex-Dryas-Tomentypnum); loess/outwash; 70°11'N, 149°17'W																			
A	1-21	6.93	nd	5.66	0.37	14.0	26.8	59.1	0.1	61.3	2.6	TR	0.1	11.2	0	69.3		92	
Oa/Ckjj	21-42	7.55	nd	15	0.96	16.0	16.1	67.9	0.5	47.8	1.2	0.1	0.1	3.6	0	30.3		100	
2Ck	42-61	7.88	nd	2.37	0.04	5.7	6.1	88.2	0.1	35.4	0.8	0	TR	0	0	3.3		100	
Ruptic-Histic Aquiturbel; Moist Nonacidic Tundra (Dryas-Salix-Eriophorum); loess/outwash; 69°27'N, 148°40'W																			
Oi	0-8	8.37	0.12	35.7	1.25			15.2	84.1	15.4	0.2		1.5	16	0.1	113		89	
Oeji1	0-40	7.81	0.20	24.2	1.25			2.1	84.3	6.8	0.3		0.2	11.6	0.1	102		89	
Bw	0-12	8.01	1.07	3.55	0.24	19.8	70.0	10.2	0.0	47.1	3.1	TR	0.1	0.8	0	22.1		100	
Bg	0-88	7.92	1.19	3.8	0.29	23.5	66.9	9.6	0.5	36.3	2	TR	0.2	3	0	24.3		100	
Oeji2	0-50	7.55	0.43	20.12	1.29			1.1	81.1	5.1	0.1		0.3	13.7	0.1	97.8		89	
2Ajj	0-85	7.73	1.09	7.27	0.39	16.7	69.3	14.0	1.8	55.1	2.3	TR	0.2	3.7	0	29.5		100	
2Cg/Oeji2m	65-85	7.40	nd	10.93	0.74	18.1	70.1	11.8	3.0	69.5	3.8	TR	0.2	7.4	0	50.8		100	
Typic Umbrothel; Moist Acidic Prostrate Shrub-Grass Tundra (Salix-Carex-Polytrichum); residuum; 68°46'N, 149°35'W																			
A	0-8	4.21	0.57	7.13	0.48	3.8	15.6	80.6	10.3	1.9	1.2	0	0.3	21.4	3.3	18.1		19	
Bw	8-22	4.05	1.30	2.05	0.12	5.2	15.9	78.9	1.8	0.5	0	TR	TR	13.9	3.9	9.2		7	
BC	22-31	4.64	1.38	0.44	0.05	1.2	11.9	86.9	7.5	0.2	0	0	TR	4.6	1.5	3		7	
Typic Histoturbel; Moist Acidic Tussock Tundra (Eriophorum-Sphagnum-Betula); colluvium; 68°37'N, 149°19'W																			
Oi	0-9	4.46	0.04	44.34	0.85			34.0	11.5	6.3	0.8		3.1	110	2.5	136		16	
Oe1	9-17	5.06	0.10	39.8	1.40			2.4	9.3	2.4	0.3		0.4	93.5		104		12	
Oe2	17-23	5.37	0.37	24.3	1.34			1.2	3.9	1.6	0.1		0.3	63.2		59.9		10	
Bg	23-40	5.17	1.46	3.04	0.11	20.0	41.9	38.1	0.4	0.5	0.4	TR	TR	14.7	4.2	15.8		6	
BCg	40-48	5.70	1.09	3.46	0.14	20.5	39.8	39.7	0.6	0.5	0.4		TR	15	3.9	14.7		6	
Cg/Oe2m	48-80	6.39	nd	4.24	0.19	19.0	38.9	42.1	1.5	0.7	0.4	TR	TR	16	3.3	14		9	
Typic Fibristel; Wet Acidic Tundra (Eriophorum-Sphagnum); organic basin deposits; 68°37'N, 149°19'W																			
Oi1	0-13	4.48	0.03	43.56	0.65			10.4	1.9	0.8	0.1		0.3	57.2	4.1	109		3	
Oi2	13-33	4.52	0.19	44.82	2.58			4.0		2.6	0.1		0.2	107	4.8	77.1		14	
Oe	33-44	4.66	0.35	25.2	1.17			0.6	3.4	0.7	TR		TR	75.2	6.9	51.3		8	
Oa	44-62	4.99	0.52	18.5	1.02			0.5	3.6	0.7	TR		TR	58.2	9	46.6		9	

† Tr = trace

and averaging about 50 kg C m^{-3} (Michaelson et al., 1996; Tarnocai, 1997). This is similar to Histosols of other regions (Section E, Chapter 6.2). Less than half of the organic C in Turbels is in the active layer, with the remainder in the upper 30 to 50 cm of permafrost (Michaelson et al., 1996).

The particle size distribution of Gelisols varies from clayey to coarse gravelly sand. The composition of the fine earth fraction is commonly dependent on the composition and age of the parent materials.

6.6.9 Special Problems in Managing Gelisols

Gelisols present special problems in terms of management, not only because of frost churning, heaving, sorting and cracking, but also because melting of segregated ice following a disturbance to the thermal regime leads to subsidence, or thermokarst. To preserve the integrity of structures (buildings, roads and pipelines) in permafrost soil, it is important to maintain the negative thermal balance of the soil. This is achieved by using special construction methods. For agricultural development it is important to determine the ice content of the soil; otherwise, after clearing of the land or within a few years after cultivation begins, severe subsidence and thermokarst can develop (Péwé, 1982).

6.6.9.1 Global Warming and Trace Gas Emissions

General circulation models (GCM) predict that with a projected twofold increase in atmospheric CO_2 by the year 2050, the mean air temperature of the Earth's surface could increase by 1.5 to 4.5°C (Maxwell and Barrie, 1989); however, warming at the high northern latitudes could be on the order of 4 to 5°C . Indeed, sea ice variations over the past several decades are compatible with a distinct warming of air temperatures in the arctic, especially during the winter and spring (Chapman and Walsh, 1993). In addition, permafrost temperatures in northernmost Alaska have increased by 2 to 4°C during the last few decades (Lachenbruch and Marshall, 1986).

As mentioned previously, Gelisols are large C sinks. The concern is that warming in the circumpolar regions could increase the thickness of the active layer and enhance heterotrophic respiration, releasing additional CO_2 to the atmosphere (Oechel and Billings, 1992; Shaver et al., 1992; Waelbroeck et al., 1997) and Gelisols would become a C source.

6.6.9.2 Human-Caused Disturbances

Arctic regions contain vast energy reserves, including fossil fuels (coal, oil and gas), biomass, and hydropower. The extraction of fossil fuels and minerals and deforestation may have dramatic long term effects on arctic ecosystems as they have a low resistance and resilience to disturbance (Reynolds and Tenhunen, 1996). As the world's population continues to expand, there will be increased pressure for development at the high latitudes. For example, Siberia already contains several cities that were developed in permafrost and have in excess of 500,000 inhabitants.

The circumarctic contains numerous indigenous peoples who are dependent on the terrestrial and marine ecosystems for food, fuel, and shelter. Special efforts must be made to ensure that these ecosystems remain sustainable.

The polar regions are less diverse biologically than other life zones. However, ancient (~ 2.5 million years old) permafrost contains viable microorganisms that may give clues to the evolution of microbes (Gilichinsky, 1993).

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6.7 Vertisols

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6.7.1 Introduction

Vertisols are clay soils which exhibit impressive volume change due to shrink/swell processes when soil moisture conditions change. The intrinsic shrink/swell behavior is considered the dominant process involved in Vertisols and results in significantly greater spatial and temporal variability of soil properties than in any other soil order. Thus, it is no surprise that Vertisols remain the most difficult and challenging type of land to successfully manage worldwide for natural resource and engineering applications.

Despite a voluminous literature, some aspects of the Vertisol order are still under debate or not fully understood. Nonetheless, considerable changes in the knowledge base of Vertisols have occurred,

particularly in the last two decades. This subsection provides a condensed overview of Vertisols characteristics. For a more detailed discussion about any specific aspect of Vertisols, the reader is referred to Ahmad (1982), Probert et al. (1987), Wilding and Puentes (1988), Kimble (1991), Ahmad and Mermut (1996), and Coulombe et al. (1996b).

6.7.2 Distribution and Formation

Geographically, Vertisols have been reported in more than 100 countries. They are estimated to occupy 308.5 Mha (10^6 ha) which corresponds to 2.36% of the ice-free land area (Table 6.1) (USDA-SCS, 1994). World extent by suborder is discussed in Section 6.7.5 with Australia and India having the greatest extent with a respective estimate of 80 Mha each (Dudal and Eswaran, 1988; Isbell, 1991). Sudan follows with approximately 50 Mha and then the United States, China, and Ethiopia with 12–15 Mha each. The global distribution of Vertisols is illustrated in Fig. 6.31.

In the United States, Vertisols are reported in 25 states and territories (Coulombe et al., 1996b). More than half (~6.5 Mha) occurs in Texas with South Dakota, California, and Montana having 1.5 Mha, 1 Mha, and 0.6 Mha, respectively. The remaining states have less than 0.25 Mha of Vertisols and vertic intergrades (Fig. 6.32).

Vertisols can be derived from a wide variety of parent materials of varying age and under differing climatic conditions (Coulombe et al., 1996a,b). Most Vertisols have developed in recent times (Quaternary). Vertisols begin to develop at the time when the geomorphic surface is exposed and undergo surficial weathering processes. While most Vertisols typically occur at low elevation and on footslope to depressional positions in the landscape, some can occur at higher elevations and on higher slope gradients and positions on the geomorphic landscape. Natural vegetation of Vertisols is typically grassland and savanna, but mixed/deciduous forested and scrub/shrub can occur, depending on the ecological succession of the vegetation and human impact.

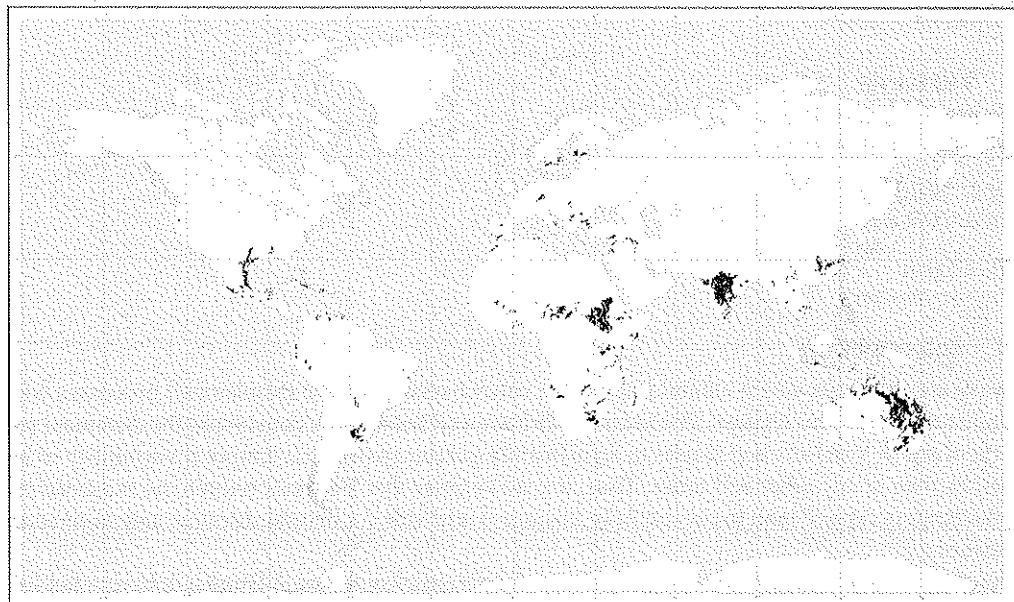


Fig. 6.31 Global distribution of Vertisols [Courtesy of USDA-NRCS, Soil Survey World Soil Resources, 1998]

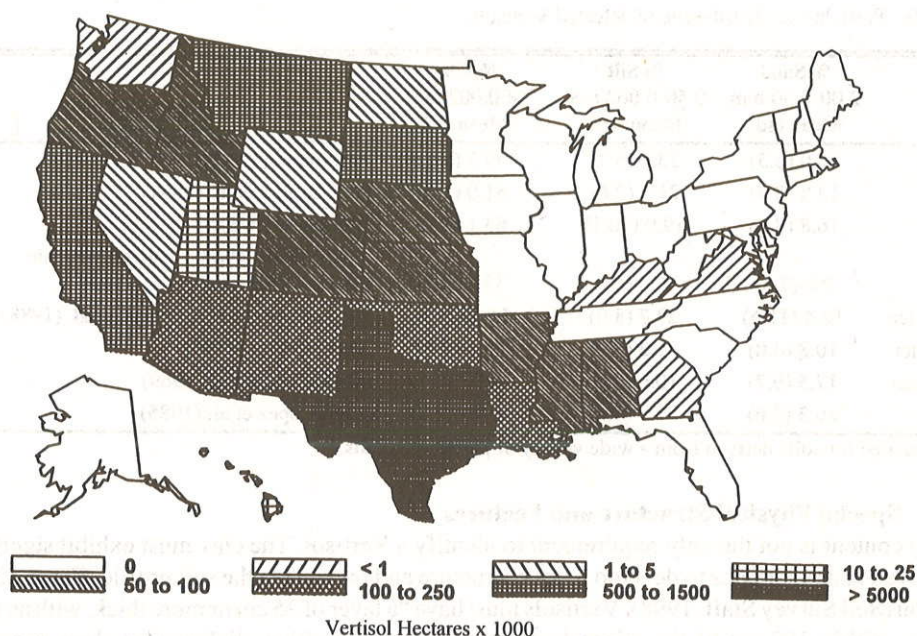


Fig. 6.32 Distribution of Vertisols in the United States [Reprinted from Coulombe et al., 1996. Overview of Vertisols. *Adv. Agron.* 57:289-375. Copyright Academic Press with permission]

6.7.3 Morphology

6.7.3.1 Color

In general, Vertisols are dark in color and commonly lack distinct horizonation. Vertisols with gray, brown, or red pigmentation have also been reported in various regions of the world. In the original version of *Soil Taxonomy* (Soil Survey Staff, 1975), color was considered at the great group category as *Pellic* and *Chromic* for low (< 1.5) and high (≥ 1.5) chroma color, respectively. In the most recent *Keys to Soil Taxonomy* (Soil Survey Staff, 1998), the color criterion is at the subgroup category by considering the chromic intragrade only, dark Vertisols being the norm. Chromic Vertisols require a color value of 4 or more (moist) or 6 or more (dry) or a chroma of 3 or more. The chroma requirement does not apply to the subgroup of Aquerts.

6.7.3.2 Texture

Vertisols are fine-textured soils. According to *Soil Taxonomy* (Soil Survey Staff, 1975; 1998), Vertisols must have: "a weighted average of 30% or more clay in the fine earth fraction either between the mineral soil surface and a depth of 18 cm or in an Ap horizon, whichever is thicker, and 30% or more clay in the fine earth fraction of all horizons between a depth of 18 cm and either a depth of 50 cm, or a densic, lithic, or paralithic contact, duripan, or petrocalcic horizon if shallower." Clay content of Vertisols can be as high as 90%, particularly for those derived from pyroclastic deposits. If the clay content requirement is not met, but the soil type exhibits considerable shrink/swell activities as measured by the coefficient of linear extensibility (COLE), these soils would be considered as vertic intergrades to other soil orders. Particle size distribution for selected Vertisols is listed in Table 6.30.

Table 6.30 Particle size distribution of selected Vertisols

Regions	% Sand 2.00-0.50 mm Mean (sd)	% Silt 0.50-0.002 mm Mean (sd)	% Clay < 0.002 mm Mean (sd)	References
Australia	16.0 (2.5)	33.6 (3.7)	50.3 (5.5)	Boettinger (1992)
India	13.8 (3.7)	21.9 (2.6)	61.0 (5.2)	Hirekurubar et al. (1991)
Sudan	16.8 (8.4)	19.9 (10.1)	63.1 (9.7)	Blokhuis (1993)
Texas	9.6 (7.3)	36.3 (6.9)	53.5 (3.7)	Kunze et al. (1963); Yule and Ritchie (1980a); Hallmark et al. (1986)
West Africa	18.4 (11.6)	27.7 (8.0)	54.0 (15.4)	Beavington (1978); Yerima et al. (1988)
El Salvador	10.8 (4.0)	31.2 (7.3)	58.1 (8.5)	Yerima et al. (1985)
West Indies	17.5 (9.7)	10.6 (3.0)	74.9 (11.4)	Ahmad and Jones (1969)
Uruguay	26.3 (5.6)	27.0 (8.1)	46.7 (13.3)	Lugo-Lopez et al. (1985)

Data calculated for soils derived from a wide variety of parent materials.

6.7.3.3 Special Physical Structure and Features

The clay content is not the only requirement to identify a Vertisol. The clay must exhibit significant shrink/swell characteristics to develop unique structure and features in the soil profile (Fig. 6.33). By definition (Soil Survey Staff, 1998), Vertisols must have "a layer of 25 cm or more thick, with an upper boundary within 100 cm of the mineral soil surface that has either slickensides close enough to intersect" or "wedge shaped aggregates which have their long axes tilted 10 to 60 degrees from the horizontal." Slickensides are features which exhibit shiny and smooth surfaces at the interface of the peds (Fig. 6.34). The presence of slickensides in a soil horizon is designated in the field as Bss and is the diagnostic characteristic common to all Vertisols.

Wedge-shaped aggregates are structural units that generally result from the intersection of slickensides. The presence of cracks (Fig. 6.35) that open and close in response to drying and wetting is also a requirement for Vertisols (Soil Survey Staff, 1975; 1998). Since cracking occurs in soils other than Vertisols, criteria for depth and duration when cracks remain open are the basis to discriminate and infer soil moisture regimes of Vertisols.

Gilgai and diapir are other features which may be present in Vertisols. Gilgai is an Australian aboriginal term to describe a microtopography that consists of mounds, shelves, and/or depressions, either in circular, linear, or complex patterns (Fig. 6.36). Since gilgai does not occur in all Vertisols, it was abandoned as a facultative criterion in *Soil Taxonomy* (Soil Survey Staff, 1998).

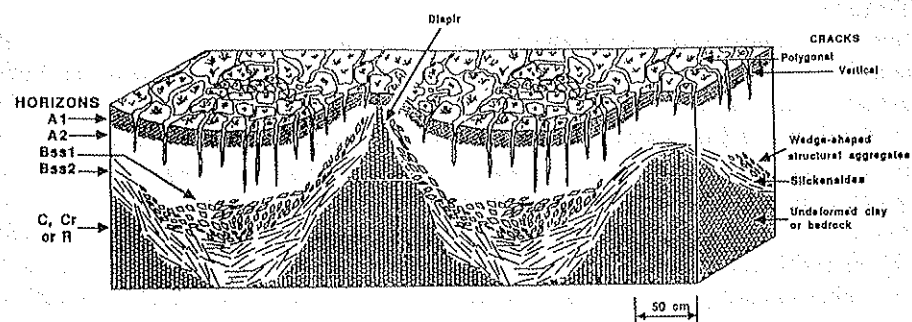


Fig. 6.33 Schematic of morphological structure and features of vertisols [Reprinted from Coulombe et al., 1996. Overview of Vertisols. Adv. Agron. 57:289-375. Copyright Academic Press with permission]

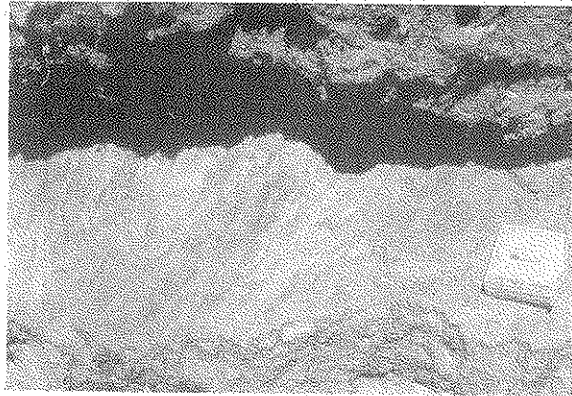


Fig. 6.34 Slickenside in a Vertisol of Texas [Reprinted from Coulombe et al., 1996. Overview of Vertisols. *Adv. Agron.* 57:289-375. Copyright Academic Press with permission].

Diapir, also called mukarra by the Australians, is another feature which may occur in Vertisols. This feature is a protrusion of soil material which penetrates from the subsoil through the upper layers and can be identified by a contrast in color and/or texture (Fig. 6.37). Diapir may occur independent of gilgai; however, when a gilgai is present, the tip of the diapir appears under the mound.

6.7.4 Genesis

Genesis of Vertisols which leads to the formation of characteristic features and structure is attributed to shrink/swell processes induced by changes in soil moisture conditions and can be seen at macroscopic to submicroscopic scales of resolution (Fig. 6.38). For decades, shrink/swell phenomena were attributed to the paradigm of the interlayer expansion/collapse of crystallographic structure of



Fig. 6.35 Typical polygonal cracking pattern of a Vertisol of Mississippi [Courtesy of Dr. Dennis Mengel, CH2M HILL]



Fig. 6.36 Circular gilgai under ponded conditions in a Vertisol of Texas [Reprinted from Coulombe et al. 1996. Overview of Vertisols. *Adv. Agron.* 57:289-375. Copyright Academic Press with permission]

clay minerals such as smectites. However, changes in the interlayer spacing of smectites contribute only 10–30% of the change in volume when saturated with Ca (Greene-Kelly, 1974; Tessier, 1984). Tessier (1994) reported that the type of microstructure, e.g., crystallite, domains, or quasicrystals (Fig. 6.38) and the surface area of the clay mineral phases govern the shrink/swell potential. The microstructure and the porosity will accommodate the changes in water content and potential of the clay water systems. Other determinant factors involved in the shrink/swell phenomenon include, but are not limited to, predominance of cations present on the exchange sites, the electrolyte concentration of the soil solution, the amount and location of the charge in the crystallographic structure of clay minerals which may impact the structural organization of the layers and the interlayer distance which may differ from smectite to smectite, particle rigidity, and stress history, etc. Models and factors involved in shrink/swell processes have been extensively reviewed by Tessier (1984), Quirk (1994), and Coulombe et al. (1996a).

Despite advances in the knowledge of clay-water behavior, controversy remains regarding the theoretical models that can account for the genesis of various structures and features (Coulombe et al., 1996b). Coulombe et al. (1996b) also reviewed some of the models proposed for the genesis of

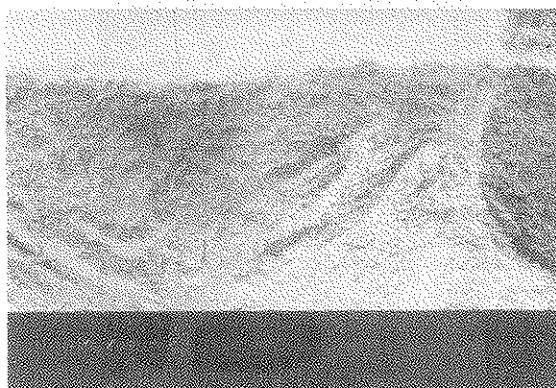


Fig. 6.37 Cross-sectional view of a Vertisol of Texas showing the diapiric structure [Reprinted from Coulombe et al. 1996. Overview of Vertisols. *Adv. Agron.* 57:289-375. Copyright Academic Press with permission]

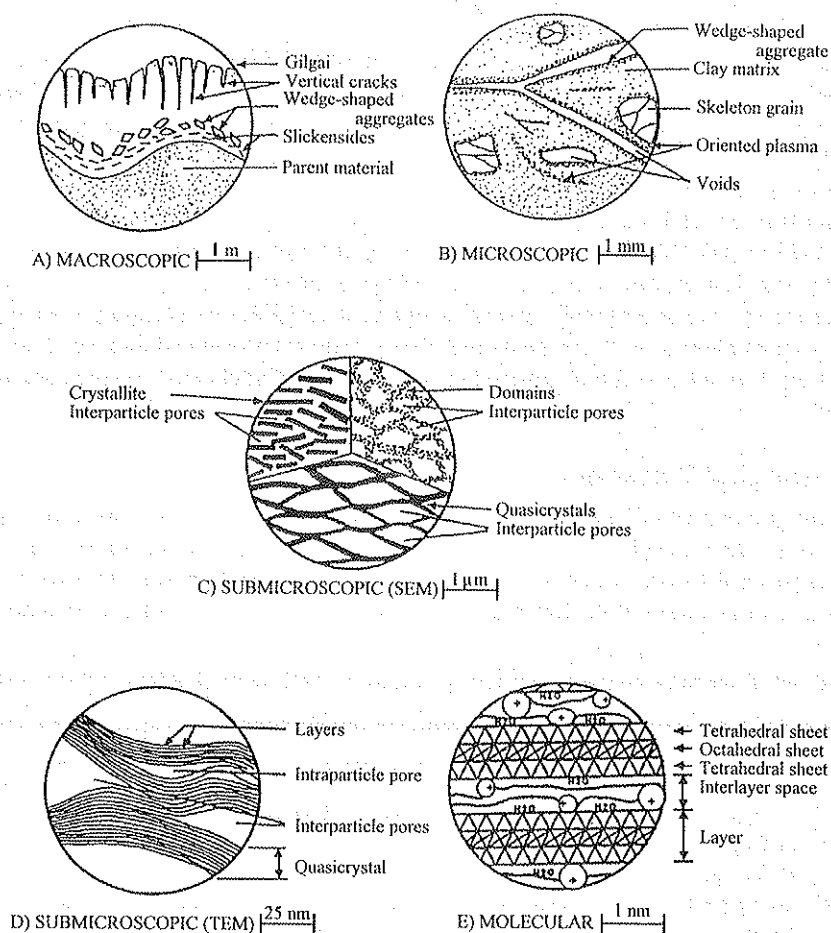


Fig. 6.38 Shrink/swell features which occurred at different scales of resolution [Reprinted from Coulombe et al., 1996. Overview of Vertisols. Adv. Agron. 56:289-375. Copyright Academic Press with permission]

features and structure in Vertisols. At the present time, the application of the Coulomb-Mohr theory of shear failure constitutes the best model to account for the formation and occurrence of features and structure in Vertisols. Shear failure is more likely to occur under moist or wet than dry conditions, since a lower normal stress is required to promote shear failure. Even though shrink/swell phenomena are significant in Vertisols, other pedogenic processes occur such as accumulation of salts (e.g., salic, natric, calcic, and gypsic great groups), silica (duric great group) and depth of saturation by water (epi and endo great groups).

6.7.5 Classification

Vertisol is a globally accepted term to designate fine-textured soils with extensive shrink/swell behavior. The term originates from *vertere* (L.) meaning to churn or turn over and *sol* for soils (Dudal and Eswaran, 1988). It was first introduced as a soil order in the 7th Approximation (Soil Survey Staff, 1960) which was superseded by *Soil Taxonomy* as an international soil classification system.

Coulombe et al. (1996b) discussed the history and evolution of the Vertisol order in the United States soil classification systems. Vertisols are also recognized in other international soil classification systems such as FAO-UNESCO (1988) and the French systems (CPCS, 1987; Baize and Girard, 1996) (Section E, Chapter 5). However, some variations in the criteria used to describe them may occur from one classification to another.

The Vertisol order is subdivided into 6 suborders, 23 great groups and 155 subgroups (Soil Survey Staff, 1998) (Table 6.31). The suborders are identified according to their moisture and temperature regimes (aquic, cryic, torric, udic, ustic, and xeric). The great group categories reflect morphogenetic markers and processes, while subgroup categories represent intragrade or intergrade concepts toward other soil orders or categories. At the present time, Vertisols do not possess a diagnostic horizon. They are discriminated from other soil orders based on the presence of slickensides which are diagnostic characteristics.

6.7.6 Mineralogical Properties

For decades, the popular belief has been that Vertisols must be dominated by a smectitic mineralogy based on extreme shrink/swell behavior. Kaolinite was generally considered as a nonswelling accessory clay mineral (phyllosilicate) of secondary importance in Vertisols. However, Vertisols containing abundant kaolinite, illite, halloysite, hydroxy-interlayered smectite, or a combination of

Table 6.31 Listing of suborders, great groups and subgroups for Vertisols order [Soil Survey Staff, 1998]

Suborders	Great groups	Subgroups
Aquepts	Salaquepts	Aridic, Ustic, Leptic, Entic, Chromic, Typic
	Duraquepts	Aridic, Xeric, Ustic, Aeris, Chromic, Typic
	Natraquepts	Typic
	Calciquepts	Aeric, Typic
	Dysaquepts	Sulfaqueptic, Alic, Aridic, Ustic, Aeris, Leptic, Entic, Chromic, Typic
	Epiaquepts	Halic, Sodis, Aridic, Xeric, Ustic, Aeris, Leptic, Entic, Chromic, Typic
	Endoaquepts	Halic, Sodis, Aridic, Xeric, Ustic, Aeris, Leptic, Entic, Chromic, Typic
Cryepts	Humicryepts	Sodic, Typic
	Haplocryepts	Sodic, Chromic, Typic
Xerepts	Durixererts	Halic, Sodis, Aquic, Aridic, Udic, Haplic, Chromic, Typic
	Calcixererts	Lithic, Aridic, Petrocalcic, Leptic, Entic, Chromic, Typic
	Haploxerepts	Lithic, Halic, Sodis, Aridic, Aquic, Udic, Leptic, Entic, Chromic, Typic
Torrepts	Salitorrepts	Aquic, Leptic, Entic, Chromic, Typic
	Gypsiteorrepts	Chromic, Typic
	Calcitorrepts	Petrocalcic, Leptic, Entic, Chromic, Typic
	Haplotorrepts	Halic, Sodis, Leptic, Entic, Chromic, Typic
Udepts	Dystrudepts	Lithic, Salic, Aridic, Udic, Aquic, Leptic, Entic, Chromic, Typic
	Hapludepts	Lithic, Aquic, Oxyaquic, Leptic, Entic, Chromic, Typic
Ustepts	Dystrustepts	Lithic, Aquic, Aridic, Udic, Leptic, Entic, Chromic, Typic
	Salustepts	Lithic, Sodis, Aquic, Aridic, Leptic, Entic, Chromic, Typic
	Gypsiteustepts	Lithic, Halic, Sodis, Aridic, Udic, Leptic, Entic, Chromic, Typic
	Calcistepts	Lithic, Halic, Sodis, Petrocalcic, Aridic, Udic, Leptic, Entic, Chromic, Typic
	Haplustepts	Lithic, Halic, Sodis, Petrocalcic, Aridic, Leptic-Udic, Entic-Udic, Chromic-Udic, Udic, Leptic, Entic, Chromic, Typic

interstratified clay minerals have been also found in various regions of the world (Coulombe et al., 1996a). Recognition of Vertisols with these characteristics has caused geoscientists to reconsider the Vertisol paradigm and mechanisms involved in the shrink/swell processes of soils not dominated by smectite. As previously mentioned, comprehension of shrink/swell behavior is beyond the crystallographic level, involving changes in microstructure and porosity associated with changes in water content and potential. Only under very specific conditions will the expansion/collapse of the interlayer space of clay minerals play a role in shrink/swell phenomena (Tessier, 1984; Wilding and Tessier, 1988; Coulombe et al., 1996a).

Nonetheless, smectites remain a universal and significant component of most Vertisols. Characteristics of smectites in Vertisols, which may also apply to other clay minerals and other soil orders, include (1) a greater thermodynamic stability than reference minerals; (2) a higher layer charge (> 0.45 per half unit cell) which may be as high as vermiculite or illite and in return impact the availability of some ions such as K and NH_4 ; and (3) a higher Fe content in their crystallographic structure, particularly when derived from basaltic parent materials containing abundant ferromagnesian minerals (Coulombe et al., 1996a). Despite the importance of clay minerals, a broad variety of minerals have been reported in Vertisol profiles and parent materials (Coulombe et al., 1996a).

6.7.7 Chemical Properties

6.7.7.1 pH

Vertisols are commonly known as being neutral to alkaline in reaction, since the majority are derived from calcareous or base-rich parent materials. However, acid Vertisols (Ahmad, 1985) exist and have been recognized as the Dystric great group (Soil Survey Staff, 1998). Dystric Vertisols must exhibit pH values of 4.5 or less in 0.01M CaCl_2 (5 or less in 1:1 water) and an electrical conductivity of the saturation extract of less than 4.0 dS m^{-1} at 25°C with a total thickness of 25 cm or more within 50 cm of the mineral soil surface. Processes leading to acidification include removal of carbonates and bases by leaching or due to chemical processes such as (1) hydrolysis, (2) acid sulfate weathering, which is the result of the oxidation of sulfidic minerals such as pyrite to sulfuric acid and sulfate minerals, and (3) ferrolysis, which involves destabilization of crystallographic structure of clay minerals by proton attack. Ranges in pH of selected Vertisols are listed in Table 6.32.

6.7.7.2 Cation Exchange Capacity and Exchangeable Cations

Cation exchange capacity (CEC) of Vertisols generally ranges from 20 to $45 \text{ cmol}_c \text{ kg}^{-1}$ (soil) or higher depending on the amount and type of mineral phases and soil organic matter (SOM) (Table 6.32). Calcium and Mg, and to a lesser extent, K and Na are the main exchangeable cations. When exchangeable Na percentage (ESP) > 15 , structural deterioration occurs. Soil dispersion may occur at low ESP values when electrolyte concentration is low while at higher electrolyte concentrations, and even ESP, soil particles may remain flocculated (Section G, Chapter 2). However, in the case of high electrolyte concentration, the osmotic potential gradient between soil and plant roots increases and induces wilting and eventually death of the plants that are not tolerant to these saline conditions. Presence of some minerals such as gypsum, even in trace concentrations, can be sufficient to overcome the dispersing power of Na and flocculate soil particles.

Other cations such as Al, Mg, and H may increase and replace exchangeable Ca, particularly under acid conditions due to dissolution of minerals (Ahmad, 1985). An increase in the concentration of these cations and a decrease in CEC indicates that chemical degradation has occurred.

Table 6.32 Chemical characteristics of selected Vertisols

Regions	pH (1:1 water)	CEC (cmol _c kg ⁻¹ soil)	Extractable Bases (cmol _c kg ⁻¹ soil)				References
			Ca	Mg	K	Na	
	Mean (sd)	Mean (sd)	Mean (sd)	Mean (sd)	Mean (sd)	Mean (sd)	
Australia	7.0 (0.6)	33.7 (2.5)	22.4 (3.4)	15.7 (3.4)	0.1 (0.1)	0.9 (0.4)	Boettinger (1992)
India	8.2 (0.2)	55.8 (4.7)	53.4 (5.6)	*	0.9 (0.1)	1.4 (0.5)	Hirekurubar et al. (1991)
Sudan	7.3 (0.4)	40.6 (11.0)	n/a	n/a	n/a	2.95 (3.9)	Blokhuis (1993)
Texas	7.4 (0.9)	41.0 (5.4)	52.9 (23.0)	6.8 (3.5)	0.7 (0.3)	2.7 (3.4)	Kunze et al. (1963); Hallmark et al. (1986)
West Africa	7.4 (0.8)	30.3 (8.0)	17.9 (7.9)	7.5 (3.1)	0.7 (0.4)	3.2 (3.3)	Beavington (1978); Yerima (1986)
El Salvador	6.7 (0.6)	45.5 (3.4)	33.9 (10.3)	11.6 (2.5)	0.5 (0.2)	1.9 (1.5)	Yerima et al. (1985)
West Indies	7.3 (0.6)	48.0 (11.7)	51.8 (23.6)	8.1 (5.8)	0.4 (0.3)	2.4 (2.1)	Ahmad and Jones (1969)
Uruguay	6.5 (0.5)	33.9 (7.9)	23.1 (7.5)	8.0 (2.3)	0.3 (0.1)	0.3 (0.1)	Lugo-Lopez et al. (1985)

n/a: not available; * Extractable Mg was compiled with data for extractable calcium.

Data were calculated for soils derived from a wide variety of parent materials.

6.7.8 Biological Properties

Biological properties involve the quantity and type of organic matter as they relate to pedofauna and land use (Table 6.33). Generally, organic C content ranges from 3 to 60 g kg⁻¹ but most commonly from 3 to 35 g kg⁻¹ (Coulombe et al., 1996a). The SOM varies with (1) climatic conditions (higher under humid and cooler regimes), (2) microtopography (higher under depressional positions than under mound positions of a gilgai) (Kunze and Templin, 1956; Wilding and Tessier, 1988), (3) depth (decreases with depth in the soil profile, which strongly refutes the argument that complete self-mixing occurs in Vertisols) (Wilding and Tessier, 1988; Southard and Graham, 1992; Coulombe, 1997), and (4) tillage (higher in virgin than cultivated areas) (Kunze and Templin, 1956; Skjemstad and Dalal, 1987; Coulombe, 1997).

Inferences of SOM content based on soil color can be misleading. Black Vertisols may contain as low as 3 g kg⁻¹ of organic C. The dark color is the result of strong clay organic complexes in which smectites are coated with humic materials in the presence of Ca (Singh, 1954, 1956). The resistance to degradation is due to this complex rather than the recalcitrance of the humic materials to degradation (Skjemstad et al., 1986; Skjemstad and Dalal, 1987).

Pedofauna in Vertisols are generally diversified (Stephan et al., 1983). Humus type is a mull with a C/N ratio ranging from 10–25 (Table 6.33), unless the soil conditions are poorly drained which may lead to the formation of a hydromull or hydromoder. Microbial biomass may represent up to 5% of total organic C (Jocteur Monrozier et al., 1991). The loci of the organic matter such as polysaccharides and fungal hyphae are at interparticle positions in the domains and quasicrystals of clay minerals (Chenu, 1989; Chenu and Jaunet, 1990).

Loss of organic C can occur shortly after a change in land use systems from native/virgin conditions to cultivation. After a few decades or more of continuous cultivation, loss of organic C can be significant (Skjemstad et al., 1986; Skjemstad and Dalal, 1987; Puentes, 1990; Puentes and Wilding, 1990; Coulombe, 1997).

Table 6.33 Biochemical and biological characteristics of surface horizons of selected Vertisols

Regions	Soil Series (and/or Parent Material)	Land Use	OC (g kg ⁻¹)	Total N (g kg ⁻¹)	C/N ratio	Total P (mg kg ⁻¹)	References
Australia	Langlands-Logie clays	Virgin	23.0	2.41	9.5	n/a	Skjemstad et al. (1986)
	(Alluvial clayey sediments)	20 yr cultivation	11.4	1.12	10.2	n/a	Skjemstad et al. (1986)
		35 yr cultivation	8.9	0.89	10.0	n/a	Skjemstad et al. (1986)
		45 yr cultivation	7.8	0.77	10.1	n/a	Skjemstad et al. (1986)
	Black Earth	10 yr pasture	14.2	0.9-2.6	5.8-12.2	n/a	Jocteur Monrozier et al. (1991)
Texas	Houston Black	Virgin	20.9-48.2	1.37-3.53	13.69-21.10	0.35-0.55	Coulombe (1997)
		20 yr restoration	13.7-24.0	0.87-1.45	18.37-19.81	0.23-0.29	Coulombe (1997)
		Pasture	15.5-33.3	0.97-2.02	18.18-25.83	0.29-0.44	Coulombe (1997)
		Cereal rotation	12.2-15.6	0.83-1.03	14.74-20.28	0.34-0.43	Coulombe (1997)
		Row crop rotation	15.6-19.3	0.96-1.45	13.36-19.39	0.29-0.48	Coulombe (1997)
Nigeria	Lacustrine Plain	Cereal	0.26-0.58	0.06-0.08	5.2-7.2	5-10	Beavington (1978)
Trinidad	Princes Town Clay	> 100 yr Sugar cane	31.0-38.0	2.7-3.5	11	29-58	Ahmad and Jones (1969)
Jamaica	Carron Hall Clay	Unimproved pasture	42.0-60.0	4.5-5.7	9-11	13-15	Ahmad and Jones (1969)
Barbados	Black Soil	> 300 yr Sugar cane	12.0-15.0	0.8-1.0	15	250-500	Ahmad and Jones (1969)
Antigua	Fitches Clay	Mixed scrubs/grasses	32.0-38.0	2.2-2.3	15-17	13-17	Ahmad and Jones (1969)
Dominica	Black Soil	Scrub	11.0-28.0	1.0-2.1	11-13	2-4	Ahmad and Jones (1969)

n/a: not available

6.7.9 Physical Properties

6.7.9.1 Moisture Retention and Shrink/Swell Behavior

Soil moisture conditions are extremely critical in Vertisols because they govern all physical properties and behavior. Due to their clay content, Vertisols exhibit high moisture retention characteristics at all water potentials. Plant-available water in Vertisols is determined as the difference between water held at -0.033 and -1.5 MPa. Ranges in moisture retention values for selected Vertisols are presented in Table 6.34. Moisture content can vary considerably depending on the amount and kind of clay, climatic conditions, and land use.

Shrink/swell behavior severely limits the utilization of Vertisols. Under normal field conditions, most of the volume change occurs at water potentials > -1.5 MPa (Yule and Ritchie, 1980a; Wilding and Tessier, 1988), but it is not clear whether volume change is isotropic (equidimensional) or anisotropic. Current knowledge suggests that individual undisturbed peds shrink and swell isotropically (Yule and Ritchie, 1980a; Cabidoche and Voltz, 1995). Yule and Ritchie (1980b) reported that shrinkage in large (73 cm diam, 140 cm long) and small (10 cm diam) cores of a Texas Vertisol was similar. However, under field conditions, Cabidoche and Voltz (1995) reported anisotropic volume change in a Vertisol from Guadeloupe (French West Indies) with vertical slightly

larger than the horizontal movements (anisotropy ratio ≈ 0.8). They attributed the anisotropy to movement of the peds along oblique slickensides, which highlights the importance of spatial arrangement or geometry of the peds in a soil profile on shrink/swell behavior.

6.7.9.2 Bulk Density and Coefficient of Linear Extensibility

Bulk density values for selected Vertisols are presented in Table 6.34. Factors that influence bulk density include (1) moisture content, (2) clay content, but more specifically the particle density of the mineral fraction, and (3) methodology used to determine bulk density (core vs. Saran® coated clods). Bulk density is routinely used to estimate total porosity and assess structural conditions of soils, but in several cases, bulk density has not been a sensitive indicator of structural changes in Vertisols. Bulk density should be coupled with morphological characteristics and other properties for a thorough assessment of soil structural conditions.

COLE, which is an index of soil shrinkage, is correlated with total and fine clay contents, surface area, water retention at field capacity, and ESP (Anderson et al., 1973; Yule and Ritchie, 1980a; Yerima et al., 1989). COLE values for selected Vertisols are presented in Table 6.34.

6.7.9.3 Consistence and Atterberg Constants

Vertisols exhibit very narrow ranges of soil moisture and consistence favorable for workability and trafficability. Consistence is estimated from the pressure required to deform a soil clod placed between the index finger and thumb. Consistence of Vertisols varies from friable to firm when moist, and from firm to extremely firm when dry. When wet, Vertisols become very plastic and sticky. Thus, determination of optimal moisture conditions is extremely critical for proper timing of cultivation and the energy/power required.

The Atterberg constants (liquid limit, plastic limit, and plasticity index) (Section A, Chapter 2) are empirical tests used by geoscientists and engineers to determine the mechanical characteristics of a

Table 6.34 Physical and mechanical characteristics of selected Vertisols

Regions	% Water 0.033 MPa Mean (sd)	% Water 1.5 Mpa Mean (sd)	Bulk Density 0.033 MPa (Mg m ⁻³) Mean (sd)	Bulk Density Oven Dry (Mg m ⁻³) Mean (sd)	COLE (cm cm ⁻¹) Mean (sd)
India	38.3 (2.9)	17.8 (0.9)	n/a	1.3 (0.1)	n/a
Texas	42.2 (3.2)	23.4 (2.3)	n/a	1.8 (0.2)	n/a
West Africa	31.4 (6.5)	n/a	1.4 (0.1)	1.4 (0.1)	0.10 (0.03)
El Salvador	n/a	n/a	1.1 (0.1)	1.1 (0.1)	0.16 (0.03)

	Liquid Limit (%) Mean (sd)	Plastic Limit (%) Mean (sd)	Plasticity Index (%) Mean (sd)	References
India	61.0 (2.4)	34.0 (1.2)	26.9 (1.3)	Hirekurubar et al. (1991)
Texas	67.1 (3.8)	22.7 (4.3)	44.3 (3.6)	Kunze et al. (1963); Yule and Ritchie (1980a)
West Africa	n/a	n/a	n/a	Yerima (1986)
El Salvador	n/a	n/a	n/a	Yerima et al. (1985)

n/a: not available

Data calculated for soils derived from a wide variety of parent materials.

soil. Values for Vertisols are presented in Table 6.34. In general, values of Atterberg constants (1) increase as the clay content increases, and (2) decrease in the order: smectite > illite > kaolinite.

6.7.9.4 Structure/Porosity and Related Properties

Structure and porosity are extremely critical components of Vertisols. Soil processes such as gas exchange, heat transfer, water flow, and movement of solutes and organics are dependent on structure and porosity. One of the typical characteristics of Vertisols is the propensity to form large pores (macro- and mesopores) through wetting/drying cycles and/or biotic activities. The presence of macro/mesopores can be fairly significant for various processes that occur in Vertisols. For instance, they allow greater nitrification rates than in the rest of the soil mass (Kissel et al., 1974) and promote movement of water and solutes through preferential or bypass flow (Kissel et al., 1973, 1974; Bouma, 1983; Bouma and Loveday, 1988). Lin et al. (1996) reported, for a Texas Vertisol, that macropores (> 0.5 mm effective diam.) and mesopores (0.06 mm – < 0.5 mm effective diam.) contributed 89% and 10% of the water flow, respectively. Numerous factors influence infiltration. Infiltration rates reported for selected Vertisols are presented in Table 6.35. Because in shrink/swell clay soils, hydraulic conductivity (K) does not meet the assumption of uniform flow in a homogenous soil (Darcy's law), infiltration rates are preferred over K (Lin, 1995; Lin and McInnes, 1995). Nonetheless, infiltration rates can be highly variable in space and time and remain dependent on a variety of factors including land utilization and infiltration methods (Table 6.35).

6.7.10 Land Use and Management Considerations

6.7.10.1 Natural Resources

Agriculture

For many regions of the world, Vertisols are highly productive and sustainable resources for various crop production systems. However, despite their resilience, Vertisols in developed as well as developing countries are undergoing degradation after only decades of cultivation due to intensive and/or inappropriate management practices. Soil properties may partially recover from the effects of cultivation but it is unlikely that native conditions can be rejuvenated even after a restoration period of 20 years or more (Puentes, 1990; Puentes and Wilding, 1990; Coulombe, 1997). One of the difficulties in successfully managing Vertisols involves technology transfer. Vertisols require site-specific management which is discussed in detail by Ahmad and Mermut (1996).

Wetlands

Vertisols subjected to aquic and udic soil moisture conditions can potentially be used for the creation or restoration of wetland habitats (Section G, Chapter 4). Besides the presence of hydric conditions, a requirement for successful establishment of desired wetland functions includes development of wetland hydrology and the presence of hydrophytic vegetation. Hydrological conditions can be supported by a permanent or seasonally perched watertable, or surface water. This causes the clay to swell and, consequently, drastically reduce permeability. In the process, an organic mat will develop on the floor of the wetland over time, further contributing to the decrease in permeability. Alternatively, the soil surface could be mechanically puddled such as for rice production, which alters the soil structure and breaks pore continuity and connectivity. Wetland hydrology is critical for development of hydric soil conditions and establishment of hydrophytic vegetation.

Table 6.35 Infiltration values under saturated conditions for surface horizons of selected Vertisols with reference to land use and methodology

Regions	Soil Series (and/or Parent Material)	Land Use	Methodology (Reference)	Infiltration ($\mu\text{m s}^{-1}$)	References
India	(Chlorite schist)	n/a	Double ring infiltrometer (Marshall and Stirk, 1950)	3.3	Hirekurubar et al. (1991)
	(Shale)	n/a	Double ring infiltrometer (Marshall and Stirk, 1950)	5.6	Hirekurubar et al. (1991)
	(Granite gneiss)	n/a	Double ring infiltrometer (Marshall and Stirk, 1950)	0.8	Hirekurubar et al. (1991)
	(Deccan trap)	n/a	Double ring infiltrometer (Marshall and Stirk, 1950)	1.2	Hirekurubar et al. (1991)
	(Limestone)	n/a	Double ring infiltrometer (Marshall and Stirk, 1950)	11.1	Hirekurubar et al. (1991)
Texas	Heiden	15-25 yr Pasture	Double-ring infiltrometer (Bouwer, 1986)	0.70-0.94	Puentes (1990)
	Heiden	15-25-yr cultivation	Double-ring infiltrometer (Bouwer, 1986)	0.14	Puentes (1990)
	Houston Black	Virgin	Column method (Bouma, 1982)	1.86	Coulombe (1997)
			Ponded infiltrometer (Prieksat et al., 1992)	1.77	Potter (pers. comm.)
			Disc permeameter (Perroux and White, 1988)	2.13	Lin (pers. comm.)
			Column method (Bouma, 1982)	1.04	Coulombe (1997)
		20-yr restoration	Ponded infiltrometer (Prieksat et al., 1992)	1.77	Potter (pers. comm.)
			Disc permeameter (Perroux and White, 1988)	2.29	Lin (pers. comm.)
			Column method (Bouma, 1982)	0.61	Coulombe (1997)
			Ponded infiltrometer (Prieksat et al., 1992)	1.26	Potter (pers. comm.)
		Pasture	Disc permeameter (Perroux and White, 1988)	1.86	Lin (pers. comm.)
			Column method (Bouma, 1982)	0.39	Coulombe (1997)
			Ponded infiltrometer (Prieksat et al., 1992)	1.75	Potter (pers. comm.)
			Disc permeameter (Perroux and White, 1988)	1.40	Lin (pers. comm.)
		Cereal rotation	Column method (Bouma, 1982)	0.88	Coulombe (1997)
			Ponded infiltrometer (Prieksat et al., 1992)	1.65	Potter (pers. comm.)
			Disc permeameter (Perroux and White, 1988)	1.37	Lin (pers. comm.)
			Column method (Bouma, 1982)		
		Row crop rotation	Ponded infiltrometer (Prieksat et al., 1992)		
			Disc permeameter (Perroux and White, 1988)		
			Column method (Bouma, 1982)		
			Ponded infiltrometer (Prieksat et al., 1992)		

n/a: not available

6.7.10.2 Engineering***Environmental Engineering***

Due to their extensive shrink/swell behavior, Vertisols are generally not suitable for waste and wastewater disposal. Water would generally percolate too fast when dry and too slow when wet. In

either case, this would contribute to potential contamination of surface and ground water and potential environmental health hazards. For similar reasons, Vertisols that exhibit a permanent (endoaquic) or seasonally perched (epiaquic) watertable would not be a suitable filter for wastewater disposal.

On the other hand, Vertisols could be used as constructed treatment wetlands for final disposal of tertiary effluent provided that wetland hydrology is maintained at saturation to enhance low soil permeability and to reduce groundwater recharge. Treatment wetlands are effective technologies for water quality improvement of various wastewaters while being beneficial for wildlife habitats (Kadlec and Knight, 1995).

Civil Engineering

The shrink/swell nature of Vertisols severely limits civil engineering applications. Vertisols must be stabilized for construction of buildings, roads, pipelines and utility corridors, to name a few. For example, several major cities in Texas (Dallas, Houston, San Antonio, and Corpus Christi) have been built on Vertisols. The swell/shrink properties of Vertisols make effective stabilization very difficult. The most economic choice may be to select stable soil for construction sites. If construction on a Vertisol is necessary, on-site investigation is required. Morphological properties of these soils often indicate movement at depths between 2–4 m due to natural moisture variability (M. E. Bloodworth, personal communication). These depths of movement indicate that prevention is the best way to minimize damage to structures on Vertisols and other expansive soils. Homeowners should maintain moist soils around the foundations of their dwellings that are built on steel reinforced concrete slabs to minimize movement. Application of gypsum to reduce ESP minimizes excessive swelling of smectite clay and maintains a favorable pH for plant growth in a climate with a seasonal water deficit.

Prior to construction of roads, streets and many structures that are to be built on Vertisols, lime [$\text{Ca}(\text{OH})_2$] stabilization of the soil is required to a depth of about 0.3 m or less (Eades and Grim, 1960; National Lime Association, 1991). Although this is beneficial, it obviously is only a partial treatment and complete stabilization would be extremely expensive. Larger multistory buildings sited on vertic soils and deep smectitic clays typically are supported by concrete piers with large bell-shaped feet at about 4 m depth to avoid contact with expansive soil.

Many other commercial methods to stabilize Vertisols are being promoted, some of which may be as successful as lime stabilization. Persons unfamiliar with these soils should proceed with caution because of the magnitude of the task and the unproven methods that have been marketed. One such treatment with a sulfonated naphthalene was tested under laboratory conditions on three Vertisols with some success (Marquart, 1995), but the results were inconsistent, depending on soil conditions and level of treatment. The method is not recommended without proof of its effectiveness on a given soil.

It is imperative for construction workers to seriously consider the use of shoring materials and other safety precautions when a Vertisol is excavated. In Texas, from 1980 to 1985, 50 construction workers were killed due to destabilization of Vertisols that occurred during excavation (USDA-SCS, 1986).

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6.8 Mollisols

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6.8.1 Introduction

Mollisols are generally characterized as soils with thick dark surface horizons (mollic epipedons) resulting from organic C incorporation. The terms mollic and Mollisol are derived from the Latin *molli*, soft. While the soil taxon, Mollisols, is defined by an exact set of soil property criteria, these criteria arose from conceptual ideas and empirical observations of soil landscapes with thick, dark, and often fertile surface horizons. Mollisols can form under multiple environmental conditions that facilitate accumulation of organic C in the upper soil profile. Although Mollisols correspond to the Chernozem soils of the Russian and older U.S. classification systems, the Mollisol soil order includes soils that are beyond the central concept of Chernozems (Fanning and Fanning, 1989).

Mollisols are among the most important soils for food and fiber production due to relatively high levels of native fertility coupled with climatic conditions conducive to plant growth. For this reason, the characteristics of Mollisols have been studied extensively, with early efforts focusing on the Chernozems of the Russian Steppes. During the late 19th century, droughts and crop failures afflicted the Russian Chernozem region resulting in widespread famine. In an effort to determine if the droughts could be prevented or moderated, Dokuchaev (1883, 1886) initiated an extensive study of the climate, soils, vegetation, and topography of the Chernozem Steppes. Based on the physical evidence that he and his colleagues collected, Dokuchaev theorized that the Chernozems of the Russian Steppes formed under specific conditions of climate, vegetation, and topography. These ideas contradicted the in vogue theories that Chernozems originated under either aquatic or marsh conditions. Dokuchaev's (1883, 1886) classic studies not only provided insight into the potential processes responsible for the formation of Chernozems, and subsequently Mollisols, but also articulated a conceptual framework for soil genesis that is still one of the foundations of modern pedology.

6.8.2 Geography and General Characteristics

6.8.2.1 Geographic Distribution

Extensive areas of Mollisols are distributed throughout the midlatitudes of the world predominantly on subhumid steppes and prairies. (Fig. 6.39). Mollisols occur under xeric, ustic, udic, and aquic soil

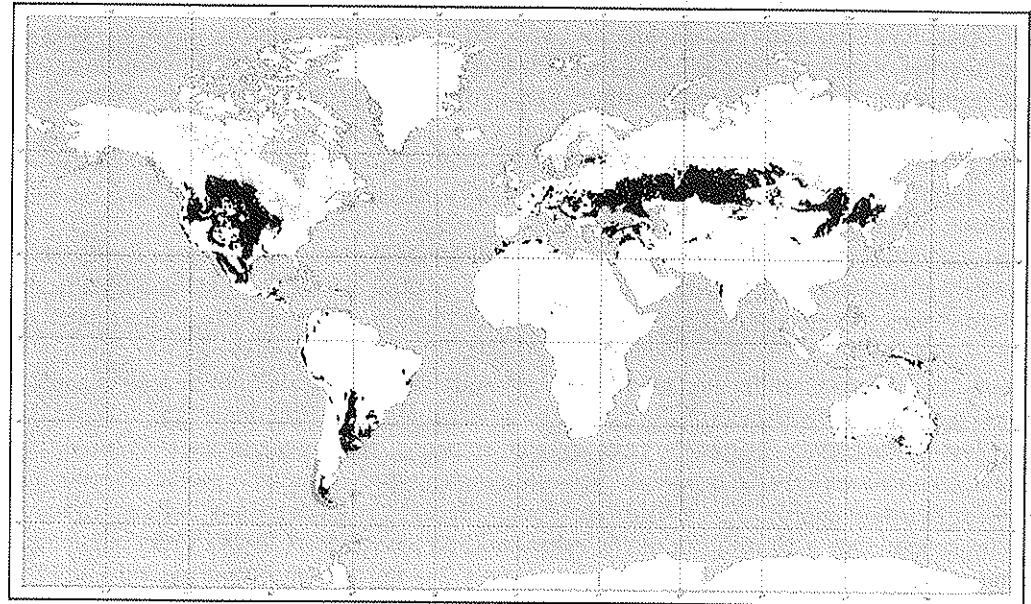


Fig. 6.39 Global distribution of Mollisols [Courtesy of USDA-NRCS, Soil Survey Division, World Soil References, 1998]

moisture regimes and cryic, frigid, mesic, thermic, and hypothermic soil temperature regimes. Spatially, Mollisols are the 8th (of 12) most common soil order and are estimated to cover 9,005,433 km² or approximately 7% of the ice-free land area on earth (Table 6.1). The distribution of Mollisol suborders can be distinguished by differences in climate with the exception of wet Mollisols (Aquolls and Albolls), which typically occur in localized areas subject to saturated and reduced soil conditions (Table 6.36).

6.8.2.2 General Characteristics of Mollisols

The general concept of Mollisols is that of dark colored soils of semi-arid to subhumid grassland ecosystems. The dark color reflects soil organic matter (SOM) enrichment in the upper portion of the

Table 6.36 Estimated occurrence of Mollisol suborder by general climatic condition (based on 1997 estimates by the USDA-NRCS)

Climatic conditions	Major geographic locations	Mollisol suborder	Approximate land area (km ²)
Semi-Arid	Great Plains of United States and Canada, Southern Russia, Pampas Region of Argentina, Yucatan Peninsula of Mexico	Ustolls	5,244,636
Humid	Black Sea Region of Russia, Coastal Regions of United States, North-Central United States, Western Europe	Udolls, Rendolls	1,526,878
Cold	Mountainous regions of the Western US, Canadian Plains, higher latitudes across Kasikstan, Ukraine, and Russia	Borolls	1,163,797
Mediterranean	Turkey, Argentina, Palouse Region of United States	Xerolls	924,394
Seasonally Saturated	Red River Valley of North, riparian zones, Gulf coastal plain, glaciated regions of the US	Aquolls, Albolls	145,728

profile (Fig. 6.40). The formation of dark surface horizons is termed melanization, which is actually a combination of several processes involving the addition of organic matter to the soil in the form of plant residues and its subsequent transformation into humus (Buol et al., 1989). Distinguishing features of Mollisols include the presence of a mollic epipedon and high base status (Soil Survey Staff, 1998). A mollic surface layer by itself is not diagnostic for Mollisols, as it can be a feature of soils belonging to other orders. Mollisols are differentiated in these cases by the presence of high base status ($> 50\%$ base saturation) horizons that underlie the mollic epipedon, or by the lack of features associated with high shrink/swell clays (some Vertisols). A wide variety of subsurface horizons can occur beneath the mollic epipedon for Mollisols, including albic, cambic, argillic, calcic, petrocalcic, natric, duripans, and gleyed horizons. The nature of any diagnostic horizon is usually expressed at the great group level in classification with the exception of Albolls (albic) and Aquolls (gleyed).

A mollic epipedon is defined in *Soil Taxonomy* (Soil Survey Staff, 1975) by several quantitative criteria including horizon thickness, organic C content, color, consistence, structure, and base status (Soil Survey Staff, 1975; 1996). Although a mollic epipedon must contain at least 1% SOM (0.6% organic C) on a weight basis, many Mollisols have higher SOM contents. A mollic epipedon must generally be a minimum of 18 cm thick for shallow and 25 cm thick for deep soils. Soils which are very shallow require only a 10 cm thick mollic epipedon (Soil Survey Staff, 1998). Color requirements include a value (a measure of lightness/darkness) that is 3 or less for a moist sample and 5 or less for a dry sample. In both cases, the chroma (intensity or purity of color) is 3 or less. Thus, the colors of a mollic epipedon range from black or very dark brown when moist and are somewhat lighter (higher value) when dry. The color criteria were established to separate high organic matter soils formed under grasslands from their lower organic matter counterparts under forest. Mollic epipedons often have well-developed granular structure that is quite friable. By definition, a mollic epipedon cannot be both hard and massive when dry. A minimum of 50% base saturation is required for a mollic epipedon indicating a dominance of basic cations, primarily Ca and Mg, occupying the CEC.

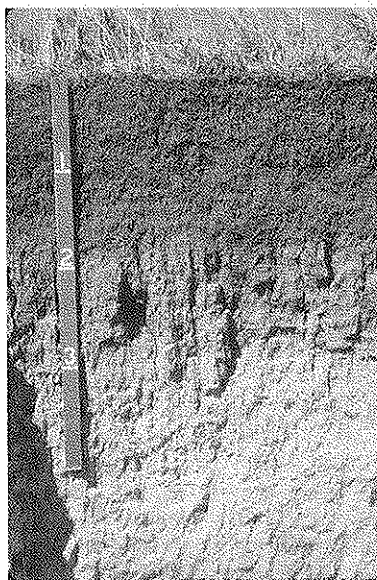


Fig. 6.40 Soil profile of a Mollisol showing the thick, dark mollic epipedon. The mollic epipedon extends to depth of almost 60 cm (2 feet); scale is in feet.

The high base status of Mollisols generally translates into a high level of native fertility. Calcium and Mg are the dominant exchangeable cations, and both of these are required in fairly large quantities for plant growth. The pH values for Mollisols may be quite variable ranging from strongly acid (5.1–5.5) to strongly alkaline (8.5–9.0). However, many Mollisols have pH values somewhere in the middle of this range and are generally considered to be favorable for plant growth without widespread use of liming agents. Their high native fertility is one of the main reasons for their extensive use in agriculture throughout the world.

Mollisols generally have not undergone intensive weathering and their mineralogy is, therefore, often dominated by minerals inherited from parent material. Many Mollisols have formed in recently deposited parent materials such as glacial till and loess, which has not allowed sufficient time for significant mineral weathering to occur. Mineral weathering is also limited by the cooler temperatures and/or lack of moisture that are associated with many Mollisols. Clay mineralogy of Mollisols is typically dominated by 2:1 layer silicates, including clay mica (also referred to as illite), smectite, and vermiculite (Allen and Fanning, 1983). Generally, kaolinite, a 1:1 layer silicate, is commonly found in Mollisols although in small quantities (Allen and Hajek, 1989) but exceptions such as the Waialua soil from Hawaii (dominated by kaolinite) are sometimes found (USDA-NRCS, 1997).

Many of the physical characteristics of Mollisols are influenced by SOM and the associated biological processes. The relatively high SOM content of Mollisols is conducive to the formation of water stable aggregates which typically have crumb or granular structure, and are formed and maintained by the interaction of plant roots, microbial structures, and biological molecules (Tate, 1992). This type of stable structure is important for good water infiltration and reduced susceptibility of soils to erosion. Organic matter plays a considerable role in enhancing the water-holding capacity of Mollisols, as it can hold up to 20 times its weight of water (Stevenson, 1994). The dark colors associated with Mollisols are attributable to the humic fractions of soil SOM which are black and mask the lighter colors associated with fulvic fractions (Schultze et al., 1993).

6.8.3 Classification

Soils classified as Mollisols in *Soil Taxonomy* can be divided into seven suborders. While all of these soils have a mollic epipedon, they are separated into the various suborders largely on the basis of their climatic regimes. Mollisols include soils that were formerly classified as Chernozems, Brunizems, Chestnut and Rendzina soils and some soils formerly classified as Brown, Brown Forest, Humic Gley, and Planosols (Fenton, 1983). Table 6.37 lists the suborders, great groups, and subgroups in the Mollisol order.

6.8.3.1 Albolls

Albolls have a well-developed, light colored horizon (albic) and a fluctuating watertable. The albic horizon forms as pigmenting agents such as SOM, clays, and Fe oxides are removed from mineral grains (primarily quartz) exposing their light color. In Albolls, the albic horizon typically occurs just below the mollic surface layer and directly above a less permeable horizon of clay accumulation. Albolls often have a watertable at or near the surface for several months during the winter and spring. In the United States, Albolls are most extensive in loess deposits of the midwest (Soil Survey Staff, 1975).

6.8.3.2 Aquolls

Aquolls are the wettest Mollisols and are characterized by the presence of redoximorphic features below a very dark brown or black epipedon. Depending on the duration of wet conditions and Fe

Table 6.37 Listing of suborders, great groups, and subgroups in the Mollisols order [Soil Survey Staff, 1998]

Suborder	Great Group	Subgroups
Albolls	Natralbolls	Leptic, Typic
	Argialbolls	Xerertic, Vertic, Argiaquic Xeric, Argiaquic, Xeric, Aquandic, Typic
Aquolls	Cryaquolls	Vertic, Histic, Thapto-Histic, Aquandic, Argic, Calcic, Cumulic, Typic
	Duraquolls	Natric, Vertic, Argic, Typic
	Natraquolls	Vertic, Typic
	Calciquolls	Petrocalcic, Aeris, Typic
	Argiaquolls	Arenic, Grossarenic, Vertic, Abruptic, Typic
	Epiaquolls	Cumulic Vertic, Fluvaquentic Vertic, Vertic, Histic, Thapto-Histic, Aquandic, Duric, Cumulic, Fluvaquentic, Typic
	Endoaquolls	Lithic, Cumulic Vertic, Fluvaquentic Vertic, Vertic, Histic, Thapto-Histic, Aquandic, Duric, Cumulic, Fluvaquentic, Typic
Rendolls	Cryendolls	Lithic, Typic
	Haprendolls	Lithic, Vertic, Inceptic, Entic, Typic
Cryolls	Duricryolls	Argic, Typic
	Natricryolls	Typic
	Palecryolls	Aquic, Oxyaquic, Abruptic, Pachic, Ustic, Xeric, Typic
	Argicryolls	Lithic, Vertic, Andic, Vitrandic, Abruptic, Aquic, Oxyaquic, Pachic, Alfic, Ustic, Xeric, Typic
	Calcicryolls	Lithic, Petrocalcic, Pachic, Ustic, Xeric, Typic
	Haplocryolls	Lithic, Vertic, Andic, Vitrandic, Aquic Cumulic, Cumulic, Fluvaquentic, Aquic, Oxyaquic, Calcic Pachic, Pachic, Fluventic, Calcic, Ustic, Xeric, Typic
Xerolls	Durixerolls	Vertic, Vitritorrandic, Vitrandic, Aquic, Paleargidic, Abruptic Argiduridic, Cambidic, Haploduridic, Argidic, Argiduridic, Haplic Palexerollic, Palexerollic, Haplic Haploxerollic, Haploxerollic, Haplic, Typic
	Natrixerolls	Vertic, Aquic Duric, Aquic, Aridic, Duric, Typic
	Palexerolls	Vertic, Vitrandic, Aquic, Pachic, Petrocalcic, Duric, Aridic, Petrocalcic, Ultic, Haplic, Typic
	Calcixerolls	Lithic, Vertic, Aquic, Oxyaquic, Pachic, Vitrandic, Aridic, Vermic, Typic
	Argixerolls	Lithic Ultic, Lithic, Torrtic, Vertic, Andic, Vitritorrandic, Vitrandic, Aquultic, Aquic, Oxyaquic, Alfic, Calcic Pachic, Pachic Ultic, Pachic, Argiduridic, Duric, Calcicargidic, Aridic, Calcic, Ultic, Typic
	Haploxerolls	Lithic Ultic, Lithic, Torrtic, Vertic, Vitritorrandic, Vitrandic, Aquic Cumulic, Cumulic Ultic, Cumulic, Fluvaquentic, Aquic Duric, Aquultic, Aquic, Oxyaquic, Calcic Pachic, Pachic Ultic, Pachic, Torrtic, Duridic, Calcic, Torripsammentic, Torriorthetic, Aridic, Duric, Psammentic, Fluventic, Vermic, Calcic, Entic Ultic, Ultic, Entic, Typic
	Durustolls	Natric, Haploduridic, Argiduridic, Entic, Haplic, Typic
	Natrustolls	Leptic Torrtic, Torrtic, Leptic vertic, Glossic Vertic, Vertic, Aridic leptic, Leptic, Aquic, Aridic, Duric, Glossic, Typic
	Calcustolls	Salidic, Lithic Petrocalcic, Lithic, Torrtic, Udertic, Vertic, Petrocalcic, Gypsic, Pachic, Aquic, Oxyaquic, Aridic, Udic, Typic
Ustolls	Paleustolls	Torrtic, Udertic, Vertic, Pachic, Petrocalcic, Calcic, Aridic, Udic, Calcic, Entic, Typic
	Argiustolls	Aridic Lithic, Alfic Lithic, Lithic, Torrtic, Udertic, Vertic, Andic, Vitritorrandic, Vitrandic, Pachic, Aquic, Oxyaquic, Alfic, Calcic, Aridic, Udic, Duric, Typic
	Vermustolls	Lithic, Aquic, Pacic, Entic, Typic
	Haplustolls	Salidic, Rutic-Lithic, Lithic, Torrtic, Pachic Udertic, Udertic, Vertic, Torroxie, Oxie, Andic, Vitritorrandic, Vitrandic, Aquic Cumulic, Cumulic, Anthraquic, Fluvaquentic, Pachic, Aquic, Oxyaquic, Torrtic, Torriorthetic, Aridic, Fluventic, Duric, Udorthentic, Udic, Entic, Typic
	Natrudolls	Petrocalcic, Leptic Vertic, Glossic vertic, Vertic, Leptic, Glossic, Calcic, Typic
	Calciudolls	Lithic, Vertic, Aquic, Fluventic, Typic
Udolls	Paleudolls	Vertic, petrocalcic, Aquic, Pachic, Oxyaquic, Calcic, Typic
	Argiudolls	Lithic, Aquertic, Oxyaquic Vertic, Pachic Vertic, Alfic Vertic, Vertic, Andic, Vitrandic, Aquic, Pachic, Oxyaquic, Lamellic, Psammentic, Arenic, Abruptic, Alfic, Oxie, Calcic, Typic
	Vermudolls	Lithic, Haplic, Typic
	Hapludolls	Lithic, Agertic, Vertic, Andic, Vitrandic, Aquic Cumulic, Cumulic, Fluvaquentic, Aquic, Pachic, Oxyaquic, Fluventic, Vermic, Calcic, Entic, Typic

reduction, expression of Fe-depleted zones can range from entire horizons with predominantly dull gray colors to horizons that have a strongly contrasting mottled gray color pattern. They typically occupy the lower lying areas of the landscape where water accumulates and a high watertable exists. The most extensive areas of Aquolls in the United States occur in the Red River Valley of the North, along flood plains and terraces of major rivers in the central part, and along the Gulf coastal plain (Soil Survey Staff, 1975). Most uses of Aquolls, whether urban or agricultural, are limited by wetness. Many Aquolls are suitable for agriculture if artificially drained. Aquolls typically have black surface horizons underlain by gray colors indicating reduction and/or depletion of Fe (Table 6.38). Organic C accumulates in the surface horizon due to reduced rates of organic matter decomposition resulting from seasonal anaerobic conditions. The presence or absence of carbonates in Aquoll profiles is quite variable and can be used as an indicator of wetland hydrology. Carbonates tend to be removed from recharge wetlands by leaching, whereas discharge wetlands are enriched often to the soil surface. Depressional basins in subhumid climates often have recharge hydrology resulting in leaching of carbonates and formation of argillic horizons (Mausbach and Richardson, 1994).

6.8.3.3 Rendolls

Rendolls form in highly calcareous parent materials of humid or cold regions, often under forest vegetation. These soils typically consist of a mollic surface layer overlying mineral material that contains > 40% CaCO_3 by weight (Soil Survey Staff, 1998). Rendolls are associated with parent materials such as chalk, limestone, highly calcareous glacial till, and shell deposits. These soils are not extensive in the United States where only a few soils with this classification have been recognized. Rendolls are extensive in western Europe (Smith, 1986).

6.8.3.4 Xerolls

Xerolls form in areas that receive the majority of precipitation during the winter and spring months when temperatures are cool (Mediterranean climates). Little precipitation is received during the growing season and, as a result, Xerolls frequently become dry at some point during the summer. This type of moisture regime is very effective for leaching CaCO_3 and other soil constituents in higher precipitation zones. Conversely, Xerolls of the drier areas are subject to long periods of soil moisture deficits during the growing season. In the United States, Xerolls comprise < 5 % of the land area occupied by Mollisols and are restricted to the western states. An example of a Xeroll profile from the Palouse region of northern Idaho is shown in Fig. 6.41, with selected data in Table 6.39. Important

Table 6.38 Selected morphological, physical, and chemical properties of an Aquoll from southern Minnesota; Location: Waseca County, Minnesota; Mean annual air temperature: 6.2 °C; Mean annual precipitation: 836 mm; Vegetation: Reed Canary Grass (*Phalaris* Parent material: *arundinacea* L.), Cattails (*Typha* sp.), Wisconsinan-age glacial till; Landscape position: depression, 0-1% slope; Classification: fine-loamy, mixed, mesic, Typic Endoaquoll

Horizon	Depth	Texture	Moist Color	Clay	pH	Organic C	Cation Exchange Capacity	CaCO_3
	(cm)			(%)		(%)	(cmol _c kg ⁻¹)	(%)
Ap	0-28	clay loam	Black	37.5	7.5	4.5	45.5	1
A	28-41	clay loam	Black	35.9	7.2	1.3	34.9	trace
AB	41-55	clay loam	Very dark gray	32.8	7.4	0.5	28.9	trace
Bg	55-71	clay loam	Olive gray	28.6	7.5	0.4	23.0	7
Cg1	71-105	loam	Olive gray	26.3	7.5	0.2	20.5	11
Cg2	105-125	loam	Gray	25.3	7.6	0.2	18.8	12

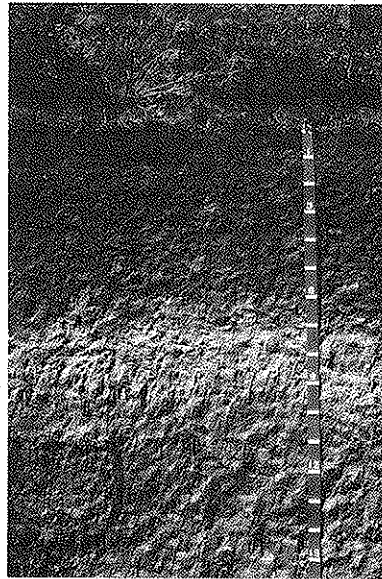


Fig. 6.41 Soil profile of a Xeroll from northern Idaho. Accompanying data are presented in Table 6.39; scale is in decimeters.

features of this soil include the lack of CaCO_3 in the profile, acidic conditions, and dense subsoil horizons (94-152+ cm) in the lower part of the profile. Effective leaching by winter precipitation is responsible for the lack of CaCO_3 and the low pH. The dense subsoil horizons restrict downward water movement and result in the formation of perched watertables during the winter and spring months. Xerolls are used primarily for production of wheat, hay, and some timber.

6.8.3.5 Cryoborolls

Cryoborolls occur in cold climates having a mean annual soil temperature $< 8^\circ\text{C}$ with cool summers. Cryoborolls are generally associated with a short growing season that limits their use for agricultural

Table 6.39 Selected morphological, physical, and chemical properties of a Xeroll (Southwick series, 93-ID-29151) from northern Idaho; Location: $460^\circ44'\text{N}$ Lat., $116^\circ 50' \text{W}$ long., Latah County, Idaho; 13 km east of Moscow; Elevation: 825 m; Mean annual air temperature: 7°C ; Mean annual precipitation: 635 mm; Vegetation: ponderosa pine, Idaho fescue, snowberry; Parent material: Late Pleistocene and Holocene loess; Landscape position: near summit of gently sloping (4%) spur ridge; Classification: fine-silty, mixed, mesic Boralfic Argixerolls

Horizon	Depth (cm)	Moist Color	Sand -----	Silt (%)	Clay -----	Bulk Density (g cm^{-3})	pH	Organic C (%)
A	0-18	very dark brown	6.9	72.2	21.0	--†	6.5	3.09
AB	18-38	very dark grayish brown	7.9	72.7	19.3	1.25	5.9	1.69
Bw	36-71	dark brown	7.7	70.0	22.3	1.30	5.8	1.20
BE	71-58	brown	7.3	74.0	18.7	1.37	5.6	0.46
E	58-94	grayish brown	8.7	78.3	13.0	1.48	5.5	0.46
Btb1	94-114	brown	5.5	63.2	31.3	1.76	5.2	0.42
Btb2	114-152+	brown	6.6	63.8	29.6	1.75	5.6	0.17

† not sampled

production. They can, however, be productive for small grains and hay production. Data from a representative Cryoboroll profile from Montana are presented in Table 6.40. Large quantities of organic C are associated with such soils giving a strong black color (~ 10% C in the 0–6 cm layer) (Table 6.40). Limited leaching has resulted in the removal of CaCO_3 from the upper 61 cm of this profile and the development of acid pH values. Base saturation increases with depth and is generally ~100% in lower horizons where CaCO_3 has accumulated. These soils are mainly used for winter wheat and hay production.

6.8.3.6 Ustolls

Ustolls occur in subhumid or semi-arid climates receiving significant amounts of spring and summer rain (350–900 mm annual precipitation), which is characteristic of the ustic soil moisture regime (Soil Survey Staff, 1975). As a result, the potential productivity of these soils for agriculture varies considerably. In the higher rainfall areas, drought sensitive crops, such as corn and soybeans, can be produced, while in the drier regions, summer fallow which allows recharge of soil moisture in a fallowed field (typically every other year), is often necessary for crop production (Soil Survey Staff, 1975). Ustolls are extensive throughout the central Great Plains and are the dominant suborder in the United States.

6.8.3.7 Udolls

Udolls occur in humid regions having typically formed in late Pleistocene or Holocene deposits under tallgrass prairie. Udolls are characterized by having moderate precipitation, well-distributed throughout the year (Soil Survey Staff, 1975). Because extended soil moisture deficits are not experienced, most Udolls have been placed into agricultural production. Udolls are extensive in the Midwest and comprise one of the most productive grain-producing regions of the world (Soil Survey Staff, 1975). Corn and soybeans, which are among the principal crops produced on Udolls, are generally not produced or produced with higher risks on other Mollisols because of climatic limitations. Data for a Udoll from Iowa are presented in Table 6.41. Although the parent material is calcareous loess, the relatively high annual precipitation has completely removed CaCO_3 from the upper 175 cm. Calcium is still the dominant soil cation and, with Mg, occupies most of the CEC.

Table 6.40 Selected morphological, physical, and chemical properties of a Boroll (S71 MT 31-1) from Montana (data courtesy of Dr. G.A. Nielsen, Montana State University); Location: Gallatin County, Montana; Elevation: 1585 m; Mean annual air temperature: 5 °C; Mean annual precipitation: 510 mm; Vegetation: bluebunch wheatgrass, Idaho fescue, prairie junegrass, sagebrush; Parent material: mixed alluvium/colluvium; Landscape position: alluvial/colluvial fan, 4% slope

Horizon	Depth	Moist Color	Clay	pH	Organic C	Cation Exchange Capacity	CaCO_3	Base Saturation
	(cm)		(%)		(%)	(cmol kg^{-1})	(%)	(%)
A1	0-6	black	28.1	5.5	9.88	43.7	nd†	70
A2	6-18	black	28.6	5.8	5.27	36.8	nd	80
Bt1	18-29	dark brown	32.8	6.1	1.88	32.1	nd	84
Bt2	29-42	brown	28.9	6.4	1.11	32.4	nd	100
Bt3	42-61	brown	27.1	7.0	0.87	34.0	nd	100
Bk1	61-86	brown	25.0	7.7	0.55	27.2	11	100
Bk2	86-130	brown	26.2	8.0	0.28	26.8	8	100
2C	130-175	brown	26.3	8.2	0.21	25.6	2	100

† not detected

Table 6.41 Selected morphological, physical, and chemical properties of a Udoll (S63 Iowa-83-2) from Iowa (data from Appendix IV, Soil Survey Staff, 1975); Location: Shelby County, Iowa; 135 km west of Des Moines; Mean annual air temperature: 9 °C; Mean annual precipitation: 720 mm; Vegetation: cropland; Parent material: Late Pleistocene loess; Landscape position: axis of short interfluvial, 3% slope; Classification: fine-silty, mixed, mesic Typic Hapludoll

Horizon	Depth	Texture	Moist Color	Clay	pH	Organic C	Cation Exchange Capacity	Exch. Ca
	(cm)			(%)		(%)	(cmol _c kg ⁻¹)	-----
Ap	0-18	silty clay loam	very dark brown	30.4	5.6	2.20	22.0	13.9
A	18-33	silty clay loam	very dark brown	33.5	5.7	1.87	22.9	14.7
			very dark	32.8	5.8	1.11	21.6	14.8
AB	33-46	silty clay loam	grayish brown					
Bw1	46-69	silty clay loam	dark brown/ brown	30.4	5.8	0.58	20.0	14.8
Bw2	69-86	silty clay loam	dark brown/ brown	28.2	5.9	0.33	20.7	14.7
BC1	86-110	silt loam	yellowish brown	26.9	5.9	0.21	20.4	14.6
BC2	110-125	silt loam/ silty clayloam	yellowish brown/ olive gray	28.0	6.0	0.17	20.7	15.2
C1	125-145	silt loam	brown/ olive gray	26.9	6.0	0.11	20.8	14.6
C2	145-175	silt loam	yellowish brown/ olive gray	25.7	6.2	0.10	19.7	13.8

6.8.4 Pedogenic Processes

Soils classified as Mollisols develop under a variety of environmental conditions through several distinct genetic pathways. Soil genesis typically occurs over time frames where environmental conditions, and hence, the theoretical equilibrium state are variable. The morphological imprints left by grassland vegetation are ephemeral in pedogenic time frames, and hence, are probably indicative of the soil climate over the past few hundreds to thousands of years. The genesis of Mollisols will be discussed from two different perspectives (factors and processes of soil formation) in an attempt to elucidate the development of these unique soils.

6.8.4.1 Factors of Soil Formation

The factor approach applied by Jenny (1941, 1980) to the soil-forming factor model is predicated on the hypothesis that soil properties result from the interaction of at least five site and flux factors. This model proposes that soil bodies result from the action of climate, organisms, and relief acting upon parent material over time and provides a useful conceptual framework by which soil differences can be attributed to state or site factors.

Organisms

To a large degree, Mollisols are distinguished from other soils on the basis of organisms, or more specifically, the overriding effects of grassland vegetation. The reason why grassland vegetation imparts distinctive soil morphologies is discussed later. While some Mollisols have developed under consistent vegetative conditions, abundant evidence suggests that many were forested during glacial and post glacial periods (Walker, 1966; Ruhe, 1970; Wright, 1970). Similar evidence also suggests that the climate has changed sufficiently during the Holocene to allow fluctuations between forest and

prairie vegetation, especially along humid to subhumid ecotones (Walker, 1966; Geis and Boggess, 1968). Most soils classified as Mollisols (with the possible exception of Rendolls, Aquolls, and Albolls) have probably had grassland vegetation at some time during their development. Because soil characteristics reflect the integration of the soil forming factors over long time periods (centuries to millennia), the contemporary environment may or may not accurately reflect the conditions under which a soil has developed. Mollisols are common in transitional areas between humid and subhumid environments such as the north central United States. Changes in climate during the Holocene have probably resulted in the formation and subsequent degradation of Mollisols as forest encroached on areas that were previously prairies (Fenton, 1983).

The mixing and incorporation of organic matter within the mollic epipedon have also been attributed to soil faunal activity. Of particular importance are burrowing organisms, such as gophers, prairie dogs, worms, and ants (Thorp, 1948; Munn, 1993). The net effect of burrowing activities is to move organic matter lower in the soil profile and to mix soil materials, resulting in a rather homogeneous epipedon.

Climate

As previously suggested, climate is also a key factor in the formation of Mollisols, primarily through its effect on vegetation. Grassland ecosystems are associated with broad ranges in mean annual precipitation and temperature. Consequently, Mollisols occur under a wide range of soil temperature and moisture regimes. In the United States, these soils are found from as far south as Texas to the Canadian border. Mollisols are most common in the zone between the arid deserts of the western and the forests of the midwestern states, a range of approximately 300 mm to more than 900 mm in annual precipitation.

The effects of climate on Mollisol formation can be demonstrated by comparing soil morphology along climatic gradients. Munn et al. (1978) examined a warm/dry to cool/moist sequence of grassland soils in Montana. They observed an increase in annual aboveground biomass production with increasing precipitation and decreasing temperature (Fig. 6.42). Similarly, SOM content, which increased as well, is expressed by increasing thickness of the mollic epipedon. This suggests that the dark, organically enriched mollic epipedon provides a record of the balance between annual site production and microbial decomposition. As production increases with increasing precipitation, decreasing temperature favors the accumulation of SOM by decreasing microbial decomposition rates. Leaching of CaCO_3 in the soils was greater with increasing annual precipitation, indicating that depth to carbonate serves as an indicator of rainfall within a geographic region.

Relief

The local topography or hillslope position affects hydrologic processes, especially erosion and sedimentation and the timing, duration, and depth of seasonal soil saturation. Many glacial till landscapes of the eastern portion of the central United States contain both Mollisols and Alfisols. Aquolls and Albolls occur in the wetter, lower lying portions of the landscape (darker soil surface) (Fig. 6.43). Conversely, Alfisols occupy the adjacent higher positions having better drainage (Allen and Fanning, 1983).

Richardson et al. (1992) proposed a framework to view wetland hydrology based on differences in how water flows through depressional basins (recharge, flowthrough, or discharge). Differences in basin hydrology result in distinctive differences in soil morphology along hillslopes (Thompson and Bell, 1996; Bell and Richardson, 1997). Examination of a toposequence of Mollisols from south central Minnesota illustrates typical catena relationships (Fig. 6.44). Along this hillslope continuum, the following changes occur from upper to lower landscape positions: (1) increased darkness (organic

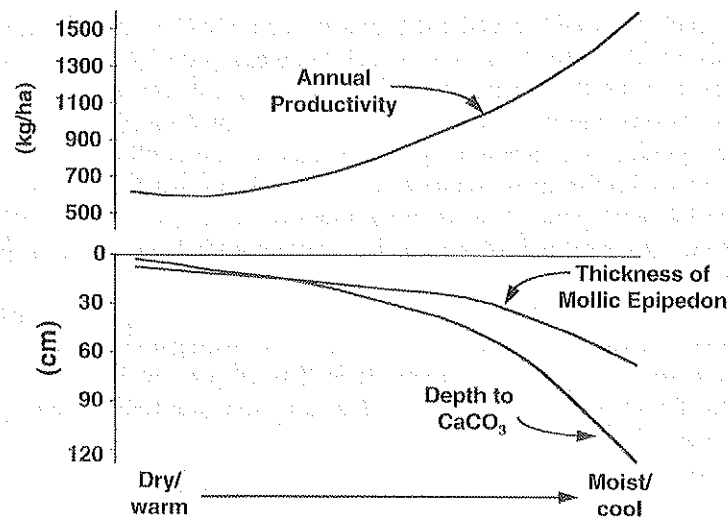


Fig. 6.42 Generalized relationship between annual aboveground productivity of rangelands in western Montana, depth to CaCO_3 , and thickness of mollic epipedon. [Adapted from Munn et al., 1978. Soil Sci. Soc. Am. J. with permission of the Soil Science Society of America]

C) and thickness of the surface horizon, (2) decreased depth to redoximorphic features, and (3) decreased chroma of the subsurface horizons. These morphological differences can often be traced back to differences in soil processes, such as erosion, deposition, and reduction, that are affected by hydrology and the shape and position on the hillslope. Numerous studies have documented increased duration of saturation in lower landscape positions for similar landscapes (Khan and Fenton, 1994; Bell et al., 1995; Thompson and Bell, 1996). For the example from Waseca, Minnesota (Fig. 6.44); higher carbonate concentrations between 45 to 90 cm at the toeslope (Fig. 6.45) are probably indicative of discharge hydrology and subsequent enrichment of the soil profile from carbonate laden groundwaters. The distribution of organic C with depth along the hillslope clearly indicates a trend of



Fig. 6.43 Glacial till landscape in central Indiana showing patterns of Alfisols and Mollisols. Wet Mollisols (Aquolls) occupy wetter, lower lying portions of the landscape and appear as the dark areas. Alfisols are found in the higher, better drained positions and appear as the lighter areas in the photograph [Photo used with permission of the Soil Science Society of America].

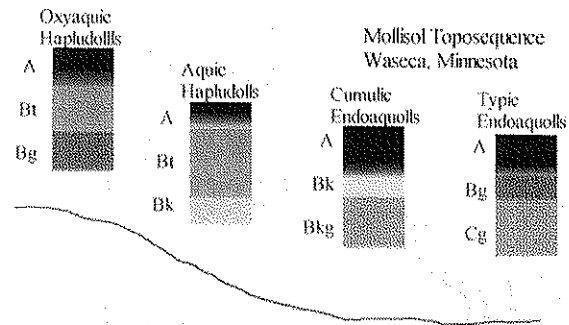


Fig. 6.44 Toposequence of soil profiles along a hillslope in glacial tills near Waseca, Minnesota.

increasing organic C in the surface horizons from upper to lower hillslope positions (Fig. 6.46). Organic C concentrations tend to converge at a depth of approximately 75 cm for all landscape positions where concentrations fall below 0.5%. These same spatial relationships cannot be assumed for all hillslopes as differences in soil stratigraphy, regional hydrology, and climate can cause different spatial patterns to develop.

Parent Material

Mollisols occur on a variety of substrate materials, but are most commonly associated with unconsolidated sediments from coastal, riverine, or glacial depositional environments, including loess (Soil Survey Staff, 1975). Loess is a common parent material of Mollisols in southern and central Russia, and Duchaufour (1982) has suggested that the permeability and CaCO_3 content of loess is especially conducive to the formation of Mollisols. Many Mollisols have formed in calcareous parent materials resulting in appreciable quantities of CaCO_3 within the soil profile, particularly those Mollisols in lower precipitation areas. Rendolls form exclusively in highly calcareous parent materials such as chalk, limestone, and shell deposits.

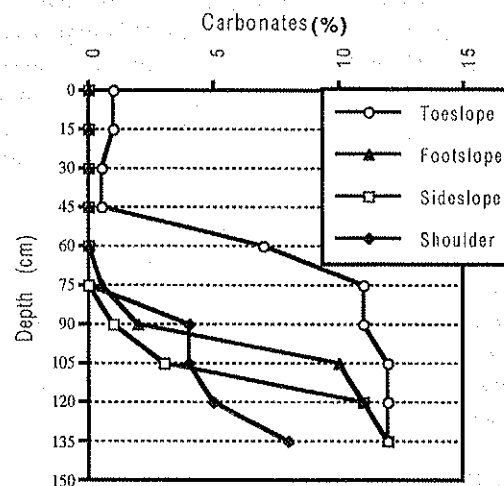


Fig. 6.45 Depth distribution of soil carbonate concentrations at four hillslope positions in glacial till near Waseca, Minnesota.

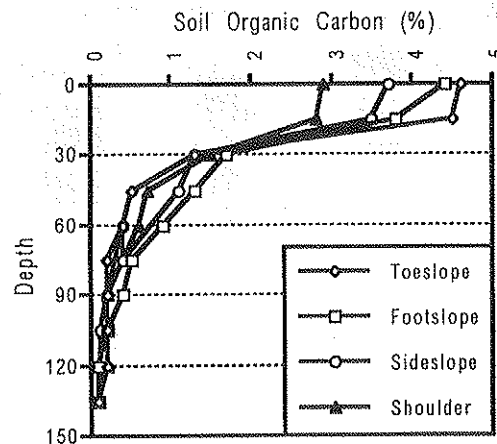


Fig. 6.46 Depth distribution of soil organic carbon at four hillslope positions in glacial till near Waseca, Minnesota.

Time

Most Mollisols in the United States and Russia appear to be relatively young and have formed on late Pleistocene and Holocene deposits and surfaces (Soil Survey Staff, 1975). However, the recognition of Mollisols in intertropical areas may warrant re-examination of this concept. It has been shown that accumulation of SOM occurs relatively fast. Schafer et al. (1980) who compared organic C levels in newly created mine soils with those of adjacent undisturbed grassland soils of eastern Montana, found that only 30 years was required for C to build up in the top 10 cm to the level in undisturbed soils. In Iowa, the rate of formation of a Udoll surface horizon was estimated at 0.08 cm yr^{-1} (Buol et al., 1989). Foss et al. (1985) suggested that mollic epipedons developed in $< 900 \text{ yr}$ in the Red River Valley of the North in Minnesota and North Dakota. From these and other studies, it is clear that the formation of a mollic surface horizon can occur in relatively short time spans.

6.8.4.2 Genetic Pathways for Mollisol Formation (Soil Processes)

The accumulation of organic C in the upper soil profile is the salient morphological feature distinguishing Mollisols from other soil orders. The C content is high when the long-term rate of addition and retention exceeds that of decomposition. Rates of annual root production and decomposition in grasslands result in high rates of organic C turnover in the soil to depths often approaching 50 to 100 cm (Dahlman and Kucera, 1965; Dormaar and Sauerbeck, 1983). Thorp (1948) estimated that 113 to 409 kg ha^{-1} of raw OM (dry weight) were added annually for short grass and 136–500 kg ha^{-1} for tall grass prairies. Alternatively, soil organic C accumulates when rates of microbial decomposition are low due to either anaerobic (wet) conditions or cool soil temperatures (Jenny, 1930). Three primary pathways for Mollisol genesis based on the mode of organic C accumulation can be identified: (1) high rates of accumulation in grasslands, (2) low rates of decomposition under anaerobic (wet) conditions, and (3) low rates of decomposition in cold climates (Fig. 6.47). These pathways are not necessarily mutually exclusive. For example, many wet soils may have also developed under grassland, but wet grassland soils frequently have higher organic C contents than drier soils in the same landscape (Figs. 6.44 and 6.46).

Grassland Mollisols (Udolls, Ustolls, and Xerolls)

A comparison of the chemical composition and growth forms of forests and grasslands explains why soils under these covers have distinctive morphologies. In forests, organic C tends to accumulate on

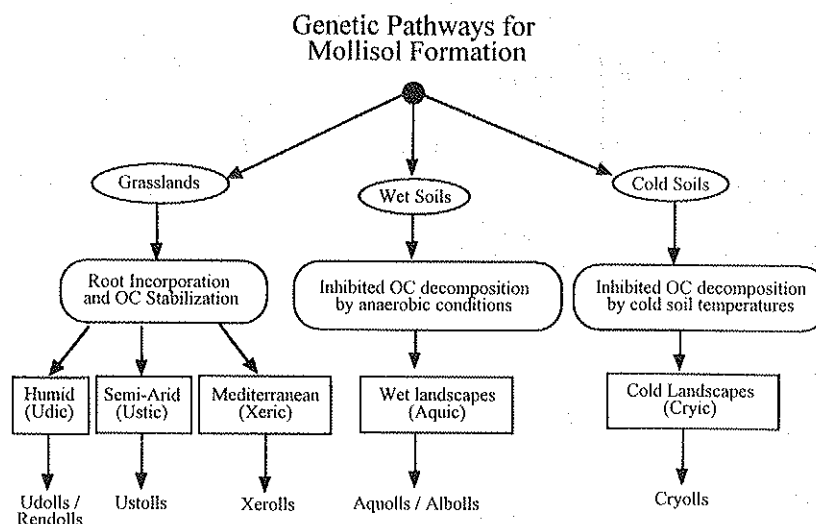


Fig. 6.47 Theoretical pathways for the genesis of Mollisols

the soil surface from annual decomposition cycles of forest floor litter. Despite the fact that the forested soils (Alfisols) receive more annual precipitation, they contain less organic C throughout the mineral soil and lack the dark soil colors associated with Mollisols. Much of the organic matter associated with the forest soils is contained in litter layers lying on top of the mineral soil. Leachate from the decomposition of forest floor litter is often composed of carbonic, flavic, and/or tannic acids facilitating translocation (eluviation) of soil materials deeper within the soil profile resulting in morphological features usually associated with Alfisols and Spodosols. These conditions and processes inhibit the accumulation and stabilization of organic C in the upper soil layers.

By contrast, photosynthesis and other metabolic processes in grassland vegetation quickly transport organic C to dense and fibrous root systems. Estimates indicate that grasslands have the highest annual additions of C to soil of any of the terrestrial ecosystems, including tropical forests (Bolin et al., 1979). Related research has demonstrated that the greatest biomass production in grassland ecosystems is below ground (Caldwell, 1975; Lauenroth and Whitman, 1977; Jenny, 1980). As such, Mollisols have distinctive organic matter profiles with depth that differ markedly from Alfisols, where most of the biomass is produced aboveground (Fig. 6.48). The annual proliferation of roots and their subsequent decomposition is responsible for the deep accumulation of SOM leading to the formation of mollic epipedons whose thickness is largely determined by the depth and amount of grass roots (Hole and Nielsen, 1968; Cannon and Nielsen, 1984). Incorporation of organic C is also facilitated by the mixing of near surface soil horizons by ants (Formicidae), earthworms (*Lumbricus*), and other soil fauna (Curtis, 1959; Baxter and Hole, 1967). Schlesinger (1977, 1991) has estimated that temperate grassland soils have a mean organic matter content of 19.2 kg C m^{-2} , ranking them behind only soils of wetlands and tundra/alpine ecosystems on a global scale.

Once added to the soil, organic materials undergo further decomposition by complex, microbially mediated processes that results in the formation of a relatively stable organic fraction. The large annual additions of C to Mollisols and subsequent cycling results in the formation of an active (bioavailable) as well as a very stable (labile) organic fraction. The humus contained in Mollisols appears to be more stable than that found in other soils (Stevenson, 1994), which is possibly related to the chemical

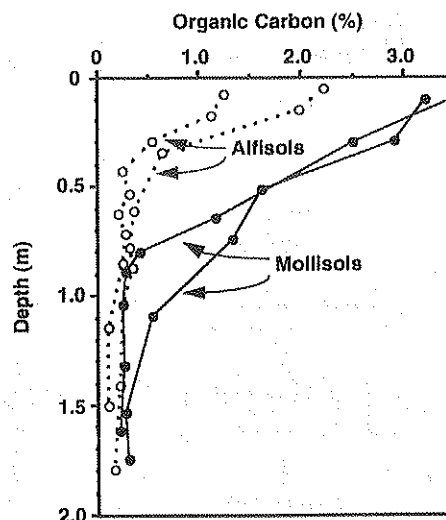


Fig. 6.48 Depth distribution of organic C in Mollisols and Alfisols of the Palouse region of northern Idaho. The Mollisols have formed under Idaho fescue grassland with approximately 580 mm of annual precipitation; the Alfisols have formed under grand fir forest with approximately 830 mm of annual precipitation [Data courtesy of Soil Characterization Laboratory, University of Idaho].

composition of grasses and soil parent material. Grasses have a higher ratio of humic to fluvic acids (Glazovskaya, 1985). Novak and Smeck (1991) found higher concentrations of humic substances in the surface horizons of Mollisols compared to Alfisols in southwestern Ohio where these soils are adjacent in the same landscape. Minimal differences were found in fluvic acid content. The combination of humic substance in the presence of calcareous soil parent material results in the formation of Ca humates that are thought to bind silicate grains to organic C which becomes stabilized (Evans and Russell, 1959; Stevenson, 1994). Formation of these complexes increases the resistance of SOM to physical disintegration, chemical extraction, and further biological change through microbial decomposition (Oades, 1989; Stevenson, 1994). Estimates as to the extent of organic matter-mineral associations in grassland soils vary. As much as 80% of the organic C in a mollic epipedon may be so closely associated with the mineral fraction that it cannot be separated by physical means (Bremner, 1965; Greenland, 1965; McKeague et al., 1986). Researchers have been able to determine mean residence time (MRT) of organic fractions of Mollisols in several different environments. The MRT represents the average length of time that an organic fraction has been present in the soil. Oldest MRT values measured in soils are associated with Mollisols and Histosols (Oades, 1989), for example, between 1255 and 2973 years for Mollisols in Canada and the United States (Anderson and Paul, 1984; Hsieh, 1992).

The stabilization of organic C in Mollic epipedons is also facilitated by soil clays. Organic C associated with fine clays is protected from further rapid decomposition lengthening the turnover time of otherwise labile humic substances from days to months, years, or even decades (Anderson and Paul, 1994). Anderson (1979) found that as much as 50% of the total humus in some grasslands is associated with clay. From these and other studies, it appears that texture is an important factor in determining the stable level of organic C in Mollisols as well as other soils. Texture influences the water-holding capacity of a soil and the quantity of clays available to form complexes with organic matter. Nichols (1984) observed a good correlation ($r = 0.86$) between soil organic C and clay content in 65 Mollisols

and associated soils of Texas, Oklahoma, and New Mexico where mean annual temperatures exceed $\sim 15^{\circ}\text{C}$ (Fig. 6.49). These data indicate that water-holding capacity and the protection afforded to humus through formation of mineral organic complexes may control the equilibrium organic C contents in Mollisols of warmer regions. In contrast, clay content does not appear to exert such strong control of organic C contents in Mollisols of cooler regions. McDaniel and Munn (1985) found little correlation between SOM and clay in 137 Mollisols and associated soils in Montana and Wyoming having mean annual soil temperatures $< 8^{\circ}\text{C}$ (Fig. 6.50). Comparison of these data with those of Nichols (1984) suggests that it is temperature rather than moisture that becomes the determining factor in establishing an equilibrium SOM content. Furthermore, organic matter may be able to persist in cold soils without the stabilizing influence afforded by complexation with clays (McDaniel and Munn, 1985).

Wet Mollisols (Albolls, Aquolls)

An alternate genetic pathway for the development of Mollisols involves organic C accumulation under anaerobic soil conditions due to reduced rates of organic matter decomposition (Ponnamperuma, 1972; Gambrell and Patrick, 1978; Tate 1980). These Mollisols have an aquic soil moisture regime and are classified as Aquolls or Albolls. The decomposition of organic matter is less efficient under anaerobic conditions and a net accumulation of SOM will occur if rates of biomass production are sufficiently high. If this rate of organic matter accumulation is high enough, organic soils (Histosols) will develop. In many landscapes, Mollisols are found at the transition between organic and mineral soils.

In general, Aquolls have black (N 2/0) surface horizons with predominant gleyed or depleted horizons directly beneath the surface horizon (Table 6.38). Increases in the duration of soil saturation and anaerobic conditions are usually associated with higher soil organic C contents (Khan and Fenton, 1994; Bell et al., 1995). Topographically, organic C increases downslope evidenced by the darker soil surfaces in lower landscape positions (Fig. 6.44) associated with higher surface concentrations of organic C in depressional areas (Fig. 6.46). Thompson and Bell (1996) found good agreement between a profile darkness index based on the thickness and darkness of the surface horizons and the duration of saturation for soil hydrosequences at several locations in the United States. They also

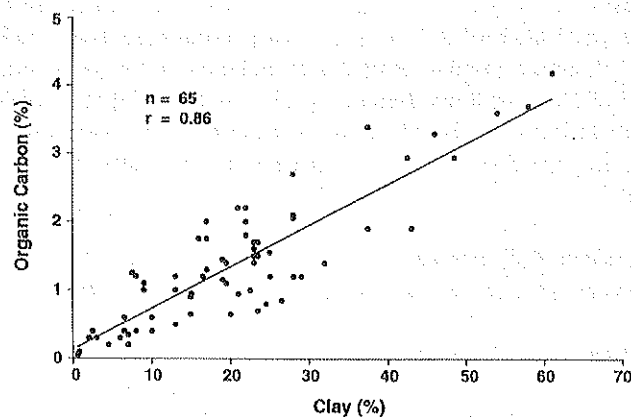


Fig. 6.49 Relationship between clay and organic matter content in warm grassland soils of the southern United States [Adapted from Nichols, 1984. *Soil Sci. Soc. Am. J.* 48:1382-1384 with permission of the Soil Science Society of America]

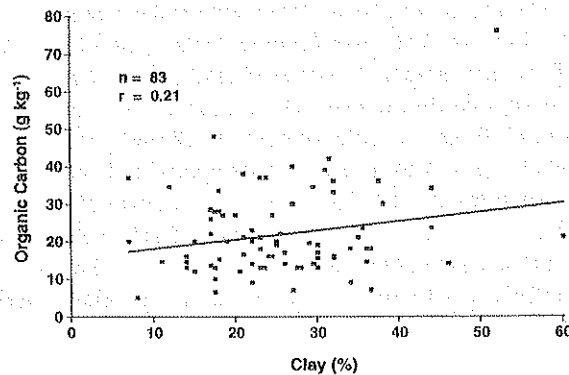


Fig. 6.50 Relationship between clay and organic matter content in cool grassland soils of the northern United States [Adapted from McDaniel and Munn, 1985. *Soil Sci. Soc. Am. J.* 49:1486-1489 with permission of the Soil Science Society of America]

found that the profile darkness index was highly correlated with soil organic C for the soils that were derived from glacial deposits.

Soil saturation, the development of anaerobic conditions and subsequent reduction of Fe(III) to Fe(II), causes the formation of distinctive color patterns (redoximorphic features) in the horizons beneath the mollic epipedon. Biochemical reduction results in the translocation of Fe compounds, which are the primary determinants of soil color in low organic matter horizons. The processes involved in the development of redoximorphic features are discussed in Section B, Chapter 5. If the soil water is stagnant, Fe(II) which has a distinctive bluish or greenish color creating a reduced (or gleyed) soil matrix is often present. If the soil water is moving, Fe(III) coatings on mineral grains will be removed as Fe^{2+} ions, leaving the mineral grains the dull, gray color of the minerals, known as depletions. Under conditions of fluctuating water tables, Fe(II) may be reoxidized creating discrete concentrations of orange or reddish Fe(III) (Vepraskas and Sprecher, 1997). Bell and Richardson (1997) discuss specific soil process and morphologies associated with Aquolls and Albolls.

Cold Mollisols (Cryolls)

The formation of Mollisols under cold conditions (mean annual soil temperature $< 8^\circ\text{C}$) is somewhat analogous to the formation of wet Mollisols. Low soil temperatures (similar to lack of soil O_2) reduce microbial activity and facilitate organic matter accumulation if the long term rate of biomass production exceeds that of decomposition. Again, organic soils (Borofolists) will develop where rates of organic matter accumulation are high and Cryolls in areas where rates are lower. Cryolls, Aquolls, and Albolls usually develop under grassland, but can develop under other types of vegetation where conditions are favorable.

6.8.4.3 Associated Pedogenic Processes

The following processes are also important in the formation of some Mollisols.

Carbonate Translocation

The dissolution of CaCO_3 and its subsequent precipitation as secondary carbonates lower in the soil profile is common in many calcareous parent materials where sufficient leaching takes place. Carbonates in layers of maximum accumulation are more finely divided than those associated with the

parent material supporting secondary carbonate formation (Redmond and McClelland, 1959). Total precipitation is important in determining the depth at which CaCO_3 is deposited (Jenny, 1941). In many Xerolls where precipitation is low but exceeds evapotranspiration especially in winter, CaCO_3 is readily mobilized and reprecipitated at depth. In the Palouse region of eastern Washington and northern Idaho, CaCO_3 has been removed from the upper 1.5 m of the profile in Xerolls receiving more than 530 mm of rainfall (Barker, 1981). In contrast, Ustolls of the Great Plains receiving comparable amounts of total rainfall typically have well-developed zones of CaCO_3 accumulation within 1 m of the soil surface. The Ustolls receive much of their precipitation during the growing season when evapotranspirational demand is high, resulting in less moisture being available for mobilization of CaCO_3 . As previously discussed, spatial patterns of carbonate removal and accumulation along hillslopes can often be used to interpret soil hydrology.

Clay Translocation

The movement and accumulation of clay are common soil-forming processes that occur in soils occupying relatively stable landscape positions and depression-focused recharge depressional basins. Clay translocation causes differences in soil texture that can affect soil water movement and the subsequent genesis of certain soil morphological features. This process is thought to be important in the formation of Albolls. Appreciable clay movement does not occur until carbonates have been removed from the upper portions of the soil profile (Duchaufour, 1982; Fanning and Fanning, 1989). Carbonate removal releases clay previously cemented which may be subject to further chemical alteration prior to movement (Fenton, 1983). In calcareous parent materials, subsoil clay accumulation commonly occurs immediately above the zone of CaCO_3 accumulation. In Mollisols, subsurface clay accumulation is highly variable and is not always observed.

In a study of Mollisols and associated soils in the forest steppe-dry steppe transition in Eastern Europe and the forest-prairie transition in the United States Great Plains, Bronger (1991) was unable to find micromorphological evidence of clay illuviation in Mollisols commonly described as having argillic horizons. While Nettleton et al. (1969) proposed that the lack of illuvial clay skins in some fine-textured soils was due to disruption by shrink/swell clays, Bronger (1991) suggested that increased clay contents in Mollisol subsoils are either lithologic discontinuities or are polygenetic in origin and, as such, are relict features that formed under older and moister climates. Sobecki and Wilding (1983) investigated Texas Coast Prairie Mollisols and found that argillic horizons were confined to microtopographic lows, suggesting depression-focused recharge. Weakly developed Bt horizons on microtopographic highs were determined to be relict features based on micromorphological analysis. They suggested that processes leading to both carbonate accumulation and clay illuviation in the same horizon were incompatible under the current humid climate.

Erosion and Sedimentation

Erosional processes result in the redistribution of surface soil material from upper to lower hillslope positions. This process often results in thickening of the mollic epipedons in concave slope positions. Soils with these overthickened surface horizons are often classified as cumulic at the subgroup level. Anthropogenically accelerated erosion causes significant reduction in the thickness of A upland horizons which no longer meet the surface horizon thickness criteria for Mollisols. These eroded soils will typically be classified as Inceptisols or Alfisols (if an argillic horizon is present). Erosional losses as a genetic pathway for Alfisols or Inceptisols which is contrary to the conceptual models for the genesis of these soils, has been a long-standing taxonomic dilemma in the classification of Mollisols (Fenton, 1983).

6.8.5 Land Use

Human impact on the use of Mollisols has increased dramatically over the past century. Early cultures had little influence on grassland soils other than the use of fire as a management tool. In the plains of North America, burning was used to stimulate the subsequent year's production and attract buffalo (Warkentin, 1969; Anderson, 1987). With the advent of the moldboard plow, Mollisols became widely used as arable soils. During the early 1900s, artificial drainage (tile drains and surface ditching) and flood protection measures lowered watertables sufficiently for the conversion of extensive areas of Aquolls and other marginally wet Mollisols to productive cropland. In the United States, where Mollisols make up ~25% of the land area, much of the agriculturally based, westward settlement was a direct result of the ease with which productive Mollisols could be cleared by fire. The high level of native fertility and favorable physical properties meant that most areas of Mollisols were converted to row cropping. Virgin areas of Mollisols and associated prairies are now quite rare in the United States.

Mollisols are among the most agriculturally productive soils in the world. Although Mollisols make up only ~7% of the Earth's ice free land area, a large part of the world's wheat and other small grains are grown on Ustolls and Xerolls (Troeh and Thompson, 1993). Because Mollisols are commonly found in subhumid climates, lack of sufficient soil moisture can limit the production of traditional agricultural crops. Soil moisture conservation strategies, such as fallowing and residue management, must be implemented for sustainable dryland farming. Unfortunately, the use of unsustainable farming practices not designed to protect the soil from erosion or to conserve soil organic C, has led to a decline in soil quality and subsequent productivity in some regions of the world. Long-term cultivation has reduced the organic C content of Mollisols; reductions of ~35 % in C contents in 60 to 70 yr of cultivation have been documented (Tiessen et al., 1982). These reductions are often associated with degradation of soil structure and increased susceptibility to erosion. Additionally, the conversion of prairie to cropland further increases the erosion potential and has resulted in devastating wind erosion during cyclic droughts in portions of North America and Asia. These processes of soil degradation result in loss of organic matter and organically bound nutrients and subsequent declines in soil productivity. The use of appropriate soil conservation practices can greatly reduce soil degradation and maintain soil productivity.

As population pressure has increased, conversion of Mollisols from arable to urban uses increased. Many of the same characteristics that make Mollisols preferable for cropland also favor many non-agricultural uses. The conversion of these prime agricultural soils and the associated decline in the resource base for food and fiber production has generated much debate in the United States and stimulated farmland preservation efforts.

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6.9 Spodosols

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6.9.1 Introduction

The formative element of the word, Spodosols, was derived from the Greek, *spodos*, meaning wood ashes. The name is connotative of the bleached eluvial (E) horizon that often overlies the distinctive reddish brown spodic (Bs or Bhs) horizon, which is characterized by an accumulation of active amorphous materials composed of organic matter and Al with or without Fe (Soil Survey Staff, 1998). Most Spodosols are sandy. The strongest grade of structure is in B horizons as a result of the accumulation of C, Al, and Fe. The humus and sesquioxides provide the cohesion in the sandy materials. Organic matter also produces structure in A horizons. Sandy E and C horizons which usually are single grained, have a high proportion of macropores, and are rapidly permeable. The sandy nature of Spodosols gives them a relatively low water-holding capacity. The available water-holding capacity increases as the amount of amorphous materials in B horizons increases (Shetron, 1974).

Typical Spodosol profiles (Fig. 6.51) have four major horizons. A or O horizons have a black or dark brown color (Table 6.42), as a result of accumulation of organic materials. E horizons have an ashy gray color, frequently albic (Soil Survey Staff, 1998), as a result of the weathering of nonresistant minerals and subsequent eluviation of amorphous materials. The Bs or Bhs horizons have dark reddish brown colors as a result of the illuvial accumulation of organometallic complexes; usually the reddest hue, lowest value and lowest chroma occur in the upper spodic subhorizon. The distinctive reddish brown color of Spodosol subsoils is related to amorphous material (Mokma, 1993). With depth, hues become more yellow, and value and chroma increase. C horizons usually have a yellower hue and higher value than the spodic horizon (Table 6.42; Fig. 6.51). The POD index is a numerical index of Spodosol development based on differences between eluvial and illuvial horizon colors (Schaetzl and Mokma, 1988). Redder Munsell hues and thicker illuvial horizons result in higher POD indices. The POD index of the Typic Haplorthod in Table 6.42 is 8.

The E horizons generally have uncoated sand grains and edge weathering on ferromagnesian minerals (Fig. 6.52). As organometallic complexes are immobilized in B horizons, they form coatings on most sand grains (Fig. 6.53). With time these coatings become thicker. As vegetation removes water from the horizon, desiccation occurs allowing van der Waals bonds and protonic bridges (De Coninck, 1980) to form, causing the horizon to become at least partially hydrophobic. Desiccation is also responsible for the development of cracked coatings (Fig. 6.54), the presence of which is a criterion for spodic materials (Soil Survey Staff, 1998). Many Spodosols also have thin Al and/or Fe

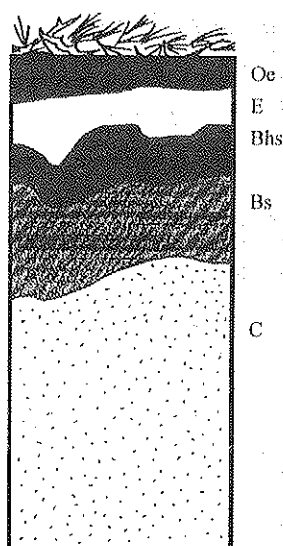


Fig. 6.51 Typical Spodosol profile

oxide coatings on sand grains in C horizons. For a more comprehensive treatment of Spodosols, the reader is referred to McKeague et al. (1983).

6.9.2 Distribution

Spodosols are found on all continents, but are most extensive in cool, humid climates (Fig. 6.56). They occupy about 2.5% of the ice-free land surface (Table 6.1). Spodosols occupy about 3,354,000 km². Some Spodosols have also formed in hot, humid areas in quartz-rich sands that are nearly devoid of weatherable minerals. Spodosols are absent in arid regions. Most Spodosols in the United States are in New England, New York, Great Lakes states, Alaska, along the Atlantic coast in the South, and in high mountains in the West.

6.9.3 Formation of Spodosols

6.9.3.1 Parent Materials

Sand fractions are composed primarily of quartz with lesser amounts of orthoclase, plagioclase, mica, pyroxene, and amphibole (Haile-mariam and Mokma, 1995). Accessory minerals are often highly

Table 6.42 Selected morphological properties of a Typic Haplorthod

Horizon	Depth (cm)	Texture	Moist Color
O	0-4	s	N 2/0
A	4-8	s	10YR 2/1
E	8-26	s	10YR 5/2
Bhs	26-38	s	7.5YR 3/2
Bs	38-57	s	7.5YR 3/4
BC	57-96	s	10YR 4/4
C	96-157	s	10YR 5/4

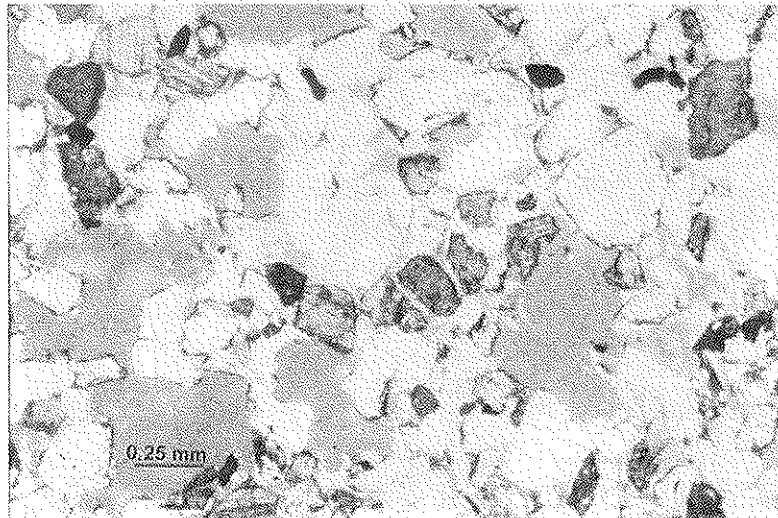


Fig. 6.52 Microphoto of E horizon in polarized light. Most grains are quartz. A few feldspar grains are conspicuous due to twinned diffraction.

weathered, opaque grains that are not easily identifiable. Less resistant minerals have weathered from E and B horizons. Parent materials contain little or no carbonate minerals and are either naturally acidic or intensively leached. Silt fractions are composed dominantly of quartz with lesser amounts of orthoclase and plagioclase. Most Spodosols have little silicate clay. Phyllosilicate minerals present in clay fractions of parent materials include mica, chlorite, and small amounts of kaolinite and vermiculite. Clay mineral distributions in Spodosols suggest that mica transforms to vermiculite and hydroxy interlayered vermiculite in B horizons, and to smectite through an intermediate vermiculite

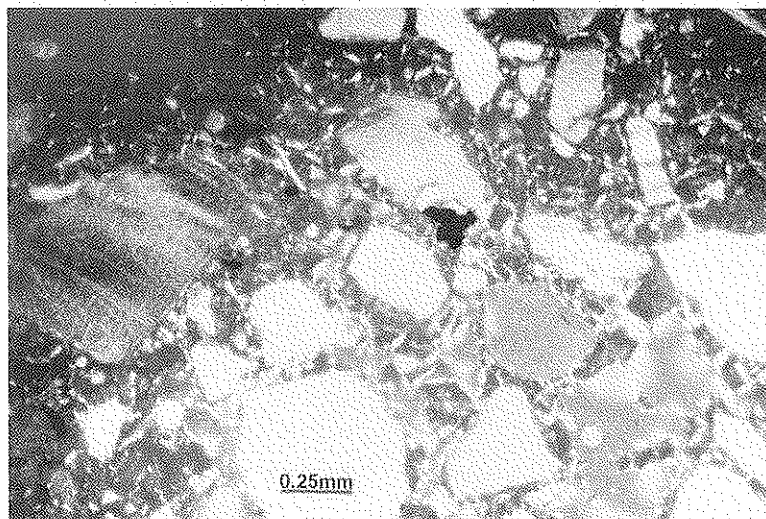


Fig. 6.53 Microphoto of Bs horizon showing distribution of sesquioxide coatings on sand grains in plain transmitted and reflected light

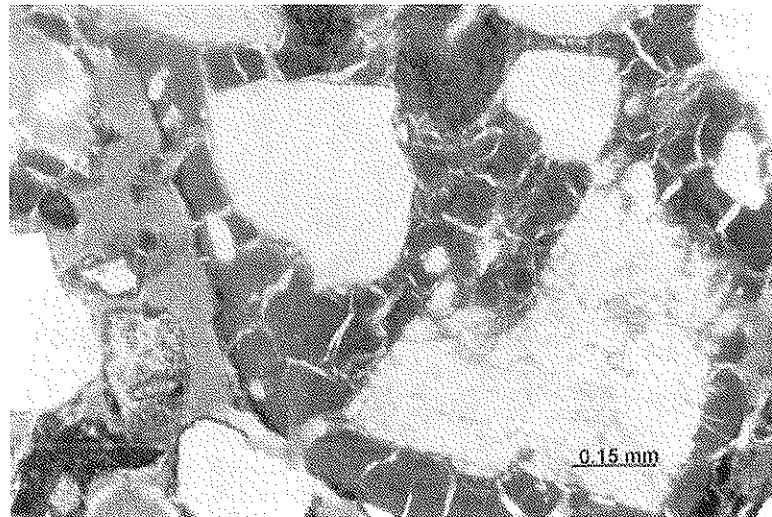


Fig. 6.54 Microphoto of Bsm horizon (ortstein) showing development of sesquioxide matrix with desiccation cracking in plain transmitted and reflected light

state, in E horizons. Nonphyllosilicate minerals present in the < 2 mm fraction are quartz and small amounts of feldspars.

As the silicate clay content increases, the translocation of humus and sesquioxides decreases; the thickness of E and B horizons decreases, and B horizon color is less red and less dark (Gardner and Whiteside, 1952). Silicate clay tends to inhibit translocation of humus and sesquioxides by adsorbing

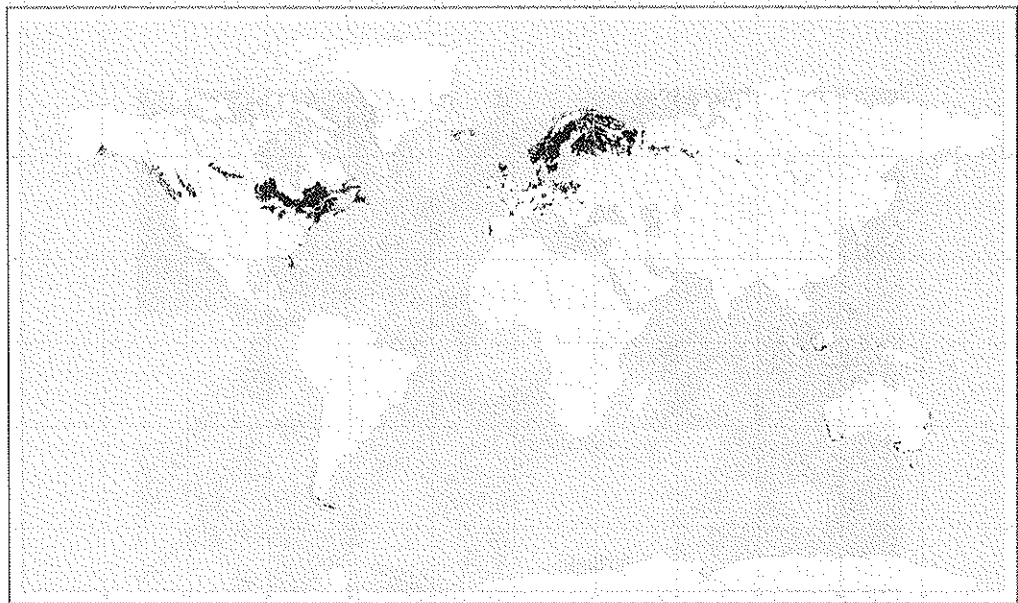


Fig. 6.55 Global distribution of Spodosols [Courtesy USDA-NRCS, Soil Survey Division, World Soil Resources, 1998]

these components. Also, weathering of silicate clays in acid, upper horizons may release many cations, immobilizing chelate complexes and inhibiting translocation.

6.9.3.2 Topography

Slope Position and Aspect

In Michigan Spodosols on north to northeast facing slopes are more strongly developed than those on south to southwest facing slopes (Table 6.43) (Hunckler, 1996).

Soil Drainage

Spodosols form on surfaces that have very deep to shallow watertables. Development of spodic horizons is influenced by depth and duration of water saturation (Mokma, 1991; Mokma and Sprecher, 1994). Spodic horizons do not appear to form in soils that are permanently saturated near the soil surface. Ratios of resistant minerals in the sand fraction of subsoils to those in parent materials tend to decrease from well (1.08) to somewhat poorly (1.06) to poorly drained (1.04) pedons. Ratios of quartz contents in the silt fractions of the subsoils to those in parent materials tend to be greater for horizons that were not water saturated (1.04) than for those that were saturated (0.93). This suggests that water saturation hinders weathering of sand and silt sized minerals, as well as translocation processes. Poorly drained soils in Spodosol landscapes tend to have less C, Al, and Fe in their B horizons than do better drained Spodosols. Some of these have particularly low Fe levels because it was chemically reduced and leached from the soil. In some low lying landscapes, hydrologic discharge may result in reversal of translocation, such that the spodic horizon may accumulate materials eluviated from the C horizon, as the water table moves upward (Evans and Mokma, 1996). These soils were formerly referred to as Groundwater Podzols (Baldwin et al., 1938).

Maximum accumulations of C, Al, and Fe occur usually in somewhat poorly drained Spodosols (Table 6.44) in which the saturated zone acts as a barrier to water movement, and organometallic complexes are carried from A and E horizons to B horizons, but rarely move through the B horizons in significant amounts. In these soils, intensity of oxalate extractable Fe illuviation into spodic horizons is strongly related ($R^2 = 0.7$) to the frequency of watertable fluxes (Evans and Mokma, 1996). Spodic horizon chemistry is unresponsive to duration of saturation and watertable fluctuation, but C

Table 6.43 Mean optical density of oxalate extract and extractable Al and Fe in paired pedons on opposing aspects [Adapted from Hunckler, 1996]

Aspect	Horizon	ODOE [#]	Al _o [†]	Al _p [‡]	Fe _o [†]	Fe _p [‡]
g kg ⁻¹						
N-NE	E	0.0*	0.2*	0.1*	0.2*	0.1*
	Bs1	0.3*	4.7	2.5*	5.1*	2.5*
	Bs2	0.1	3.5*	1.7*	2.3	1.1
S-SW	E	0.1	0.3	0.2	0.4	0.2
	Bs1	0.1	3.5	1.4	3.5	1.1
	Bs2	0.1	2.1	0.9	1.8	0.6

[#] ODOE = optical density of oxalate extract

[†] Al_o and Fe_o = oxalate extractable Al and Fe

[‡] Al_p and Fe_p = pyrophosphate extractable Al and Fe

* Indicates or promotes significantly stronger podzolization when comparing equivalent horizons of paired pedons on opposing aspects.

Table 6.44 Weighted index amounts of pyrophosphate extractable Al and Fe in soils of a Michigan hydrosequence

Drainage Class	Horizon	Al	Fe
-----g kg ⁻¹ x cm-----			
Well	E	1	2
	Bs1	42	47
	Bs2	29	16
Moderately Well	E	1	1
	Bs1	48	19
	Bs2	24	18
Somewhat Poorly	E	2	2
	Bs1	75	112
	Bs2	76	36
Poorly	Bg	4	0

horizon chemistry is more responsive and that of E horizons is most responsive to watertable fluctuation patterns (Evans and Mokma, 1996).

In well-drained Spodosols, organometallic complexes can be carried by percolating waters into, and even through, B horizons. The brownish colored lakes and streams in Spodosol areas support this; however, spodic horizons are a significant sink for Fe, Al, and dissolved organic C (DOC) that effectively decreases stream content of these elements (McDowell and Wood, 1984).

Vegetation

Spodosols form under coniferous or deciduous forest but not under prairie vegetation. More Al and Fe occurs in B horizons formed under conifers than under deciduous trees (Messenger et al., 1972; De Kimpe and Martel, 1976). In mixed forests, Spodosols are more strongly expressed under some species, such as hemlock (*Tsuga canadensis*) and Kauri pine (*Agathis australis*), than under other species. Since organic C is involved in the translocation of C, Al, and Fe, the more leaf/needle litter produced, the more organic compounds to form organometallic complexes.

In some areas of sandy Spodosols, fire frequency affects the formation of B horizons by influencing forest successional events (Mokma and Vance, 1989). When the interval between severe fires is 5 to 50 yr, jack pine (*Pinus banksiana* L.) survives and regenerates. Formation of the Bs horizon is slow because much of the organic matter in the forest floor is destroyed by the frequent fires. Relatively small amounts of organic compounds are generated for complexing Al and Fe (Table 6.45). The E horizons are thin or nonexistent and B horizons have small amounts of C, Al, and Fe. The soils in this environment are weakly developed. When the fire interval is 50 to 350 yr, red (*Pinus resinosa* Ait.)

Table 6.45 Weighted index amounts of pyrophosphate extractable Al, Fe and C in B horizons of a developmental sequence of soils [Adapted from Vance et al., 1986]

Subgroup	Fire frequency	Al	Fe	C	Al + Fe	Al + Fe + C
				-----g kg ⁻¹ x cm-----		
Typic Udipsamment	5 - 50	62	33	196	95	291
Entic Haplorthod	50-350	65	44	275	109	384
Typic Haplorthod	>350	68	59	388	127	515

and/or white (*Pinus strobus* L.) pine succeed the jack pine. More needle litter accumulates as a result of fewer fires; thus, more organic acids and chelating substances are available to translocate Al and Fe from E horizons to B horizons. The soils in this environment are primarily Entic Haplorthods. When fire intervals exceed 350 yr, succession will likely produce a northern hardwood forest [sugar maple (*Acer saccharum* Marsh) and beech (*Fagus grandifolia*)] which is more shade tolerant and more fire resistant. The accumulation of C, Al, and Fe in B horizons continues, but is likely to be slower under hardwoods because less Al and Fe and more Ca are being recycled (Messenger, 1975). The soils in this environment are primarily Typic Haplorthods.

Treethrow is another form of vegetative influence in Spodosol formation. Tree uprooting disrupts soil horizons and arrests and regresses soil development. Portions of the Spodosol profile are removed, producing a pit. The soil material is deposited on top of adjacent soil producing a mound. The resulting microtopography in an area of Haplorthods had Entic Haplorthods or Spodic Udipsamments in pits and Typic Udipsamments on mounds (Schaetzl, 1990).

Climate

Accumulations of C, Al and Fe in spodic horizons tend to be greater in cooler areas of the Appalachian Mountains than in warmer areas (Stanley and Ciolkosz, 1981). In New England, Spodosols are the dominant soil order in frigid and cryic temperature regimes, but occur less frequently in mesic areas, except when aquic conditions are present (Evans, 1996). Similarly, many Aquods also occur in sandy parent materials in Florida. In Michigan, strong spodic development is associated with belts of deep snowpack that were in place before the soil froze. Meltwater from the snow could then flush freely through the soil in the spring (Schaetzl and Isard, 1991).

Time

Spodic horizons can form in a few hundred years if conditions are optimum, but they usually form more slowly. E horizons have been reported in soils that were 300 yr old or less (Gjems, 1960; Ugolini, 1968). Weak Spodosol profiles have been reported after about 2000 yr of soil development (Burgess and Drover, 1953; Franzmeier and Whiteside, 1963). Even though weak Bs horizons form in a few thousand yr, it takes 8000 to 10,000 yr of soil development for the majority of soils on a surface to qualify as Spodosols (Table 6.46) (Franzmeier and Whiteside, 1963; Barrett and Schaetzl, 1992).

6.9.4 Pedogenic Processes in Spodosols

6.9.4.1 Eluviation of Humus and Sesquioxides

Formation of the Spodosol profile is thought to involve the formation, translocation and immobilization of organometallic complexes. Organic acids are produced in forest canopies and fine roots and by incomplete decomposition of the forest floor. These acids form soluble organometallic complexes with the Al and Fe released during mineral weathering in A and E horizons. Translocation of clay, resulting in argillic horizon formation, may also occur in Spodosols. The clay moves downward to form E and Bt horizons, and when the E horizon becomes sufficiently coarse, a shallow E-Bhs or Bs horizon sequence forms in it. Before the clay eluviates, it adsorbs the Al and Fe released during mineral weathering and they are not available to form organometallic complexes.

Acidification and Mineral Weathering

Organic acids move downward through O and A horizons with percolating waters. They acidify the upper mineral horizons initiating intense weathering of minerals resulting in the release of large amounts of sesquioxides, particularly Al, which often shows a strong depth correlation with pH

Table 6.46 Color, optical density of oxalate extract and extractable Al and Fe of pedons from different aged surfaces [Adapted from Barrett and Schaetzl, 1992]

Surface Age	Horizon	Color	ODOE [#]	Al _o [†]	Al _p [†]	Fe _o [‡]	Fe _p [‡]
yr BP				g kg ⁻¹			
3,000	E	10YR 4/2	0.02	0.1	0.1	0.3	0.1
	Bs	10YR 4/4	0.01	0.2	0.1	0.1	0.1
	C	10YR 6/4	0.00	0.1	0.1	0.1	0.0
4,000	E1	10YR 4/1	0.01	0.2	0.1	0.1	0.0
	E2	10YR 5/3	0.03	0.2	0.1	0.3	0.1
	Bs1	10YR 5/6	0.02	0.7	0.3	0.7	0.3
	Bs2	7.5YR 5/8	0.02	2.3	0.5	0.9	0.2
	Bs3	10YR 5/4	0.02	0.9	0.3	0.4	0.2
	BC	10YR 5/6	0.02	1.3	0.3	0.5	0.1
	C	10YR 6/4	0.01	0.4	0.1	0.2	0.1
10,000	E1	10YR 4/2	0.01	0.1	0.0	0.1	0.0
	E2	7.5YR 4/2	0.05	0.5	0.2	0.7	0.4
	Bs1	7.5YR 4/6	0.09	3.6	1.2	2.3	1.2
	Bs2	7.5YR 3/4	0.22	5.6	1.7	1.4	0.6
	Bs3	10YR 5.6	0.04	3.1	0.8	0.6	0.2
	BC	10YR 6/6	0.02	2.0	0.6	0.4	0.1
	C	10YR 6/4	0.01	1.3	0.4	0.3	0.1
11,000	E	10YR 5/2	0.01	0.1	0.1	0.3	0.2
	Bhs	7.5YR 3/2	0.28	2.8	0.9	3.2	1.3
	Bs1	7.5YR 4/6	0.10	3.3	1.0	1.7	0.6
	Bs2	10YR 4/6	0.04	1.9	0.6	0.7	0.2
	BC	10YR 5/4	0.02	0.9	0.3	0.2	0.1
	C	10YR 6/4	0.02	0.8	0.3	0.3	0.1

[#] ODOE = optical density of oxalate extract[†] Al_o and Fe_o = oxalate extractable Al and Fe[‡] Al_p and Fe_p = pyrophosphate extractable Al and Fe

* Indicates or promotes significantly stronger podzolization when comparing equivalent horizons of paired pedons on opposing aspects.

(McDowell and Wood, 1984; Mokma and Vance, 1989; Freeland and Evans, 1993). The Si, Al, Ca, and Mg in the leachate of granitic parent materials increased with increasing DOC concentration, which coincided with decreasing pH (McCracken et al., 1996). Iron release was also greatest in materials leached with the highest DOC concentration, but at low and medium concentrations, Fe was actually retained by the granitic materials, presumably because the organometallic complex precipitated out of solution (Burman, 1985).

Chelation and Eluviation

Organic compounds form chelate complexes with the Al and Fe released during weathering of minerals. Organic compounds, capable of complexing metal ions such as Al and Fe, are produced in forest canopies, canopy leachate and SOM. The kinds of compounds differ depending on tree species. The complexes are translocated downward by percolating waters.

6.9.4.2 Illuviation of Humus and Sesquioxides

C, Al and Fe

Immobilization of low molecular weight phenolic compounds, such as protocatechuic acids, may occur as a result of increases in pH. Immobilization of larger phenolic compounds is thought to occur

when the amount of Al and Fe taken up by the organic molecules exceeds a critical limit (McKeague et al., 1983). Morphological expression of podzolization has been related to the content of phenolic compounds. Seven phenolic compounds have been extracted from spodic horizons (Vance et al., 1986). Vanillic and *p*-hydroxybenzoic acids dominate in surface horizons, whereas protocatechuic acid is found in greater concentrations in B horizons (Table 6.47). The distribution of protocatechuic acid parallels that of Al and Fe in the Spodosol profiles suggesting that it plays a role in the podzolization process (Table 6.48). Of the seven phenolic compounds, only protocatechuic acid has *ortho*-dihydroxy functional groups which have the potential to form chelate complexes with Al and Fe. Content of low molecular weight phenolic compounds increases in A horizons as the proportion of coniferous to deciduous tree species in the forest overstory increases (Evans, 1980).

Ortstein

Ortstein is a spodic horizon in which sands are cemented, primarily by organic matter and Al with lesser amounts of Fe and little or no Si (McKeague et al., 1967; McKeague and Wang, 1980; Lee et al., 1988). Most roots do not penetrate ortstein. Ortstein may occur in a horizontal orientation which tends to be root restrictive, in vertical orientation which is less root restrictive, or as nodules. Ortstein in somewhat poorly drained Spodosols is more continuous and more strongly cemented with horizontal orientation, whereas that in better drained Spodosols is less continuous and less strongly cemented with vertical orientation (Mokma et al., 1994). Somewhat poorly drained soils tend to form ortstein more readily than well-drained soils, presumably due to increased accumulation of amorphous Fe and Al (Langley-Turnbaugh, 1995). Saturated zones provide a longer residence time in B horizons for organometallic complexes, thus the complexes have more time to become immobilized by attaching themselves to sand grains. Continued accumulation of the complexes may fill pores and the sand grains become cemented together. Roots are unable to penetrate ortstein, but water is able to slowly percolate through these horizons. Root and faunal activity occurs above the cemented B horizon. Ortstein reduces the permeability to moderate or moderately rapid rates (10 cm hr^{-1}), and Spodosols with ortstein tend to have lower potential forest productivity than those without ortstein. The site index for red pine on a Typic Haplorthod without ortstein is 72, whereas that with 60 to 100% ortstein is 59.

Table 6.47 Amounts of low molecular weight phenolic compounds in a Typic Haplorthod [Adapted from Vance et al., 1986]

Horizon	PA*	PHBA	VA	PHB	V	CA	FA
	mg g ⁻¹ x cm						
A	2.1	4.2	0.6	0.1	0.4	0.3	0.4
E	0.6	0.1	0.8	0.1	0.1	0	0
Bhs	5.6	3.8	7.0	0.5	0.5	0.1	0.1
Bs	13.4	4.6	8.0	1.3	1.7	0	0
BC	0.4	0.2	0.1	0.1	0.1	0	0
C	0.6	1.3	0.6	0.3	0.6	0	0

* PA = protocatechuic acid
 PHBA = *p*-hydroxybenzoic acid
 VA = vanillic acid
 PHB = *p*-hydroxybenzaldehyde

V = vanillin
 CA = *trans* - *p* - coumaric acid
 FA = ferulic acid.

Table 6.48 Weighted index contents of protocatechuic acid (PA) and pyrophosphate extractable Al_p and Fe_p in pedons of a sandy developmental sequence of Spodosols [Adapted from Vance et al., 1986]

Subgroup	Horizon	PA	Al_p	Fe_p
		mg kg ⁻¹ x cm	g kg ⁻¹ x cm	
Typic Udipsamment	Oa			
	A/E	0.5	3	5
	Bw1	2.0	36	24
	Bw2	0.8	21	10
Entic Haplorthod	Oe			
	A	0.3	1	1
	E	1.2	3	6
	Bs	9.9	22	48
	BC	2.4	35	14
	C	1.3	26	13
	Oe			
Typic Haplorthod	A	2.1	3	4
	E	0.6	3	6
	Bhs	5.6	32	31
	Bs	13.4	46	17
	BC	0.4	4	2
	C	0.6	25	12

Silicon

Silicon has recently been proposed as a component of podzolization (Farmer et al., 1980; Childs et al., 1983; Parfitt and Saigusa, 1985; Wang et al., 1986; Jakobsen, 1989; Ugolini et al., 1991; Freeland and Evans, 1993) and Si data have been interpreted as an accumulation of imogolite or allophane. Imogolite and/or allophane are identified by various methods, namely, petrography, infrared spectroscopy, energy dispersive X-ray analysis, electron microscopy, or inference from chemical data as oxalate extractable Si is nearly exclusively derived from allophane or imogolite. In some Spodosols accumulations of apparent imogolite coincide with cemented horizons with lighter colors than those typical of the C-rich ortstein (McKeague and Sprout, 1975; McKeague and Kodama, 1981; Childs et al., 1983; Freeland and Evans, 1993). Imogolite-cemented horizons usually occur as a lower Bs horizon, whereas ortstein usually occurs as upper Bhs or Bs horizons. There are no specific provisions in *Soil Taxonomy* for imogolite- or allophane-cemented horizons; thus, these horizons are considered duripans to acknowledge the role of Si. Such duripans are very different, however, from those most commonly identified in arid to semi-arid climates in base-rich soils (Section E, Chapter 6.10). Furthermore, duripans are defined as cemented horizons due to an illuvial accumulation of silica (Soil Survey Staff, 1996), and there has been controversy as to whether Si in spodic horizons has been translocated or accumulates from *in situ* weathering (Ugolini et al., 1991). Other studies (Ugolini and Dahlgren, 1987; Gustafsson et al., 1995) indicated imogolite/allophane formed in B horizons after Al had been translocated as organometallic complexes. Strong positive correlations between oxalate extractable Al and Si, but not between Al and C or between Fe and Al, occur in Haplorthods (Freeland and Evans, 1993) and Aquods (Evans and Mokma, 1996). It is possible that, following translocation, organic complexes are broken down by soil microbes, thus oxidizing and decreasing soil C levels. In weathering studies of granitic materials, however, close correlations have been found between Al and Si release rates but not between Fe and Al release rates in response to leaching with DOC (McCracken et al., 1996).

6.9.5 Classification: From Marbut to Soil Taxonomy

Soils with translocation of organic matter Al and Fe were classified as Podzols by Marbut (1936). Podzols were Pedalfers that had developed under forest cover, dominated by conifers. Soils with Podzol-like profiles and shallow water tables were recognized as Groundwater Podzols but not normal Podzols. No great soil group was established for these wet soils. Baldwin et al. (1938) added the groundwater Podzol great soil group for soils with Podzol-like profiles and somewhat poor drainage. Also, the great soil group, Brown Podzolic, was added for soils that have morphological properties of an immature Podzol. No modifications were made in the classification of Podzols by Thorp and Smith (1949) before the new approach to soil classification was begun in 1951. The Spodosol order was created for soils that had a spodic horizon (Soil Survey Staff, 1960, 1975). Emphasis was placed on the presence of the spodic horizon rather than on the albic or E horizon. Spodosols include primarily the soils that had been classified as Podzols, Brown Podzols and Groundwater Podzols. Not all Podzols classified as Spodosols. Four suborders of Spodosols were recognized: Aquods, Ferrodos, Humods and Orthods (Soil Survey Staff, 1960). The Suborder Ferrodos was dropped and the suborder Cryods was added in 1992 (Soil Survey Staff, 1992). Aquods have aquic conditions, Cryods have a cryic or pergelic soil temperature regime, Humods have a layer with at least 6% organic C within the spodic horizon, and Orthods have a spodic horizon that contains C, Al and Fe with no one element dominating. The criteria for the spodic horizon also changed in 1992. Oxalate extractable Al and Fe and the optical density of the oxalate extract (Soil Survey Staff, 1992) are used rather than

Table 6.49 Listing of suborders, great groups, and subgroups in the Spodosols order [Soil Survey Staff, 1998]

Suborders	Great groups	Subgroups
Aquods	Cryaquods	Lithic, Placic, Duric, Andic, Entic, Typic
	Alaquods	Lithic, Duric, Histic, Alfic Arenic, Arenic Ultic, Arenic Umbric, Arenic, Grossarenic, Alfic, Ultic, Aeric, Typic
	Fragiaquods	Histic, Plaggeptic, Argic, Typic
	Placaquods	Andic, Typic
	Duraquods	Histic, Andic, Typic
	Epiaquods	Lithic, Histic, Andic, Alfic, Ultic, Umbric, Typic
	Endoaquods	Lithic, Histic, Andic, Argic, Umbric, Typic
Cryods	Placocryods	Andic, Humic, Typic
	Duricryods	Aquandic, Andic, Aquic, Oxyaquic, Humic, Typic
	Humicryods	Lithic, Aquandic, Andic, Aquic, Oxyaquic, Typic
	Haplocryods	Lithic, Aquandic, Andic, Oxyaquic, Aquic, Entic, Typic
Humods	Placohumods	Andic, Typic
	Durihumods	Andic, Typic
	Fragihumods	Typic
	Haplohumods	Lithic, Andic, Plaggeptic, Typic
Orthods	Placorthods	Typic
	Durorthods	Andic, Typic
	Fragiorthods	Aquic, Alfic Oxyaquic, Oxyaquic, Plaggeptic, Alfic, Ultic, Entic, Typic
	Alorthods	Oxyaquic, Arenic Ultic, Arenic, Entic Grossarenic, Entic, Grossarenic, Plaggeptic, Alfic, Ultic, Typic
	Haplorthods	Entic Lithic, Lithic, Fragiaquic, Aqualfic, Aquentic, Aquic, Alfic Oxyaquic, Oxyaquic Ultic, Fragic, Lamellic, Oxyaquic, Andic, Alfic, Ultic, Entic, Typic

pyrophosphate and dithionite-citrate extractable Al and Fe, extractable C, clay content, and CEC (Soil Survey Staff, 1975). The Spodosol order in the 8th Edition of the *Keys to Soil Taxonomy* (Soil Survey Staff, 1998) has 4 suborders, 20 great groups, and 108 subgroups (Table 6.49).

6.9.5.1 Relationship to Andisols

Both Spodosols and Andisols are characterized by significant amounts of amorphous, hydrated oxyhydroxide compounds of Fe, Al, and Si that are chemically active. In Spodosols, these materials are present as illuvial accumulations in Bs horizons, but in Andisols such compounds are typically distributed throughout the soil profile due to weathering of the volcanic parent material (Table 6.50) (Parfitt and Saigusa, 1985) (Section E, Chapter 6.3). Carbon contents are often greater in Spodosols than in Andisols, especially in subsurface horizons in which illuvial humus accumulates. In some cases, Spodosols also form in volcanic ash parent materials. These Spodosols may contain greater than usual amounts of clay but remain distinguished from Andisols on the basis of profile distributions of active amorphous materials (Takahashi et al., 1989).

6.9.6 Chemical Properties

Organic C is highest in O and A horizons followed by the upper B horizon (Table 6.49). E horizons have an intermediate amount of organic C and usually less Al and Fe than C horizons as a result of their eluviation. Organic C contents decrease with depth from the upper B horizon to the C horizon. The distributions of Al and Fe often parallel each other, presumably because Fe and Al are comobilized by organic acids (Mokma and Vance, 1989; Mokma, 1991, 1993). In these soils, the amounts of Al and Fe are greatest in upper B horizons and decrease with depth. In other studies, however, Al is more closely related to pH (McDowell and Wood, 1984; Freeland and Evans, 1993), while Fe and C distributions are parallel. In those instances, maximum accumulation of Al tends to occur below horizons in which C and Fe reach a maximum.

The levels of Al in B horizons of Spodosols may be toxic to plants. Root growth, stem growth and leaf development may be reduced by the Al toxicity (Hoyle, 1971). Reductions in root growth associated with high Al levels are often complicated by low levels of other nutrients which are common in Spodosols. Root tolerance differs widely among plants.

Table 6.50 Extractable Fe, Al, Si and organic C in horizons of a Spodosol and an Andisol [Adapted from Parfitt and Saigusa, 1985]

Sample	Horizon	Dithionite #		Fe _o	Acid Oxalate		Pyrophosphate		C _p
		Fe _d	Al _d		Al _o	Si _o	Fe _p	Al _p	
----- % -----									
Tihoi Spodosol	HO	0.14	0.16	0.07	0.15	0	0.01	0.13	4.02
	E	0.15	0.05	0.03	0.04	0	0.03	0.04	0.77
	Bhs	0.70	0.60	0.50	0.56	0	0.44	0.49	4.21
	Bhs	0.37	1.01	0.26	0.96	0.01	0.23	0.88	5.79
	C	0.82	1.56	0.63	1.60	0.36	0.58	1.30	3.92
Taupo Andisol	A1	0.33	0.56	0.21	0.47	0	0.19	0.48	2.90
	A3	0.37	0.47	0.28	0.83	0.30	0.15	0.27	1.03
	B	0.32	0.38	0.19	0.80	0.34	0.07	0.17	0.72
	C	0.22	0.24	0.13	0.74	0.37	0.02	0.11	0.31

Dithionite-citrate-bicarbonate extract.

Table 6.51 Selected chemical properties of a sandy Typic Haplorthod

Horizon	Org C	ODOE	pH	Al _d	Al _o	Al _p	Fe _d	Fe _o	Fe _p	Si _o	P Retention
	%						g kg ⁻¹				%
A	3.4	0.06	3.6	0.4	0.3	0.2	0.3	0.4	0.3	tr	4.5
E	0.2	0.03	4.0	0.2	0.1	0.1	0.2	0.2	0.2	tr	5.6
Bhs	1.1	0.23	4.2	2.0	2.8	1.5	2.1	2.8	2.1	0.4	24.1
Bs	0.6	0.13	4.5	1.9	2.7	1.7	1.1	1.1	1.1	1.1	21.5
BC	0.2	0.04	4.8	0.8	1.2	0.7	0.3	0.4	0.3	0.5	11.2
C	-	0.03	4.9	0.4	0.5	0.4	0.2	0.2	0.2	0.3	7.2

Spodosols are acidic soils having low native fertility. A and E horizons are very strongly to extremely acid with gradual increases in pH with depth (Table 6.51).

Spodic horizons have large amounts of variable charge as a result of the accumulation of amorphous C, Al, and Fe which have the ability to retain large amounts of P (Table 6.51) not readily available to plants.

6.9.7 Management of Spodosols

Many Spodosols are used for forestry and the major products are lumber, pulpwood, and Christmas trees. Forestry has been the dominant economic activity on these soils, but agriculture is also important. Major constraints to productivity are low available P levels, low CEC, acidity and drought susceptibility.

6.9.7.1 Low Input Cultural Systems

After loggers removed most of the timber, extensive areas of Spodosols were settled by farmers and converted to crop production. These soils produced a good first crop of fall-planted small grains, such as wheat. After two or three years, yields decreased and land was abandoned. The sandy nature of these soils makes them droughty which together with acidity and low CEC make them infertile. Some Spodosols with argillic (Bt) horizons which are not as infertile as the more sandy Spodosols are used to support dairy enterprises. The application of animal manure returns nutrients and the organic matter contributes to increased water- and nutrient-holding capacities.

6.9.7.2 High Input Cultural Systems

Because of the droughty and infertile nature of Spodosols, their utilization requires high inputs, especially P. Therefore, high value crops such as potatoes and blueberries are grown. Potatoes are grown on these acid soils to avoid potato scab disease. Blueberries, both high and low bush, are acid-loving plants that do well on Spodosols. Cranberries, another acid-loving crop, are produced on wet Spodosols.

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6.10 Aridisols

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6.10.1 Concept of Aridisols

Aridisols are soils that do not have water available to mesophytic plants for long periods. Limited supplies of water may be caused by a large excess of evapotranspiration over precipitation, low

water-holding capacity due to a shallow soil depth or restricted infiltration, or to low osmotic potential due to salinity. Generally speaking, water for plant growth is available for fewer than 90 consecutive days, and there are long periods of water deficit. The lack of water also inhibits many other soil processes, including mineral weathering and removal of weathering products by leaching. As a result, Aridisols are characterized by the accumulation of relatively soluble weathering products, which are more extensively leached, or removed entirely, from soils of more humid regions.

As the order name implies, Aridisols are associated with arid climates, but not all soils in regions with arid climates are Aridisols. Approximately one-third of the Earth has an arid climate (Shantz, 1956), but Aridisols occupy less than half of the arid land area. Soils of arid regions are also classified in the Gelisol, Oxisol, Vertisol, Andisol, Mollisol and Entisol orders.

6.10.2 Geographical Distribution

6.10.2.1 Global

Aridisols occupy approximately 12% (about 16.0 million km²) of the land surface of the Earth (Fig. 6.56, Table 6.52), and rank second behind Entisols in areal extent (Table 6.1). Large areas of these soils occur in southwestern and central Asia, throughout most of Australia, western North and South America, and on the continental margins of Africa. The distribution of Aridisols coincides quite closely with arid climates associated with the descending branches of the Hadley cell of global atmospheric circulation (Ahrens, 1991) at about 30° N and S latitudes (subtropical high pressure). Their distribution is further influenced by orographic influences (rain shadows), particularly on the western margins of North and South America and in central Asia, and by long distances from oceanic sources of moisture, as in the case of the continental interiors of North America and Asia.

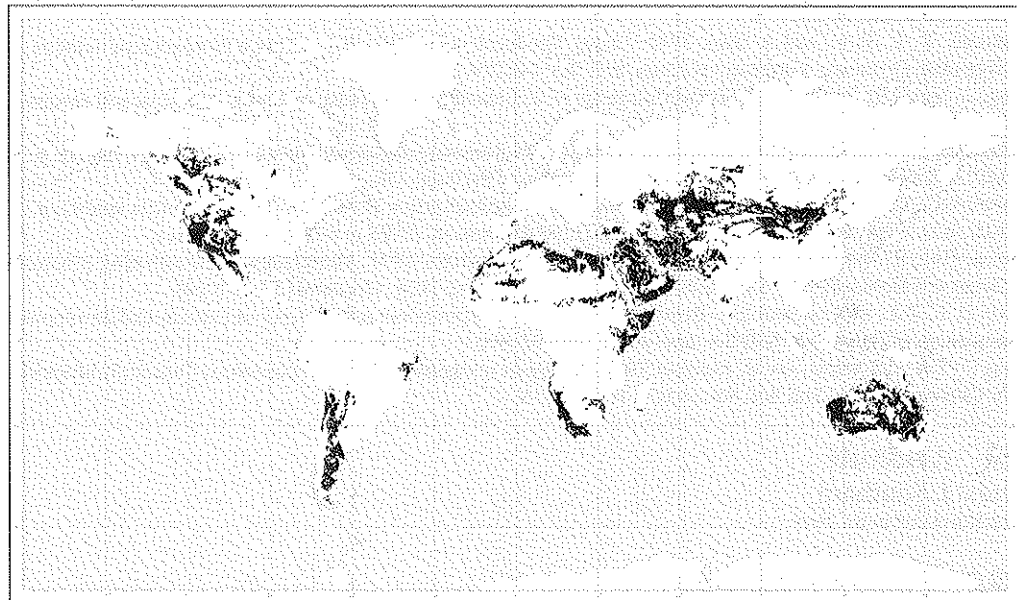


Fig. 6.56 Global distribution of Aridisols [Courtesy of USDA-NRCS, Soil Survey Division, World Soil Resources, 1998]

Table 6.52 Estimated extent (km²) of Aridisol suborders globally (USDA-NRCS, 1997a) and in the United States [Compiled from Fosberg et al., 1991; Hipple et al., 1990; Johnson, 1990; and Witty, 1990]

State	Cryids	Salids	Durids ^a	Gypsids	Argids	Calcids	Cambids	Total
WORLD	948,365	890,118	349,580	682,963	5,407,965	4,872,554	2,897,853	16,049,398
AZ	0	0	2,366	112	22,613	10,040	6,376	41,507
CA	0	45	914	0	4,616	346	350	6,271
CO	0	0	3,531	53	24,231	3,261	2,206	33,382
HA	0	5	0	0	0	0	82	87
ID	1,428	0	8,001	0	4,595	12,993	4,367	31,384
KS	0	0	1,319	0	0	0	29	1,348
MT	0	94	3,700	39	3,684	3,090	6,395	17,002
NE	0	0	74	0	26	0	137	237
NV	0	246	13,976	196	23,651	4,837	7,910	50,816
NM	0	115	1,557	3,425	34,132	36,189	7,986	83,404
ND	0	0	241	0	0	0	35	276
OK	0	0	481	0	0	808	0	1,289
OR	0	0	1,282	0	61	0	2,209	3,552
SD	0	12,052	463	17	4,396	0	1,839	18,767
TX	0	669	8,134	1,189	1,136	25,457	4,735	41,320
UT	0	696	2,311	264	1,987	9,611	1,876	16,745
WA	0	0	894	0	14	0	6,474	7,382
WY	0	0	336	0	2,403	317	350	3,406
USA TOTAL	1,428	13,922	49,580	5,295	127,545	106,949	53,356	358,075

^a Includes Durids in the USA and red and brown hardpan soils in Australia (Hubble et al., 1983; Isbell, 1990), although some of these soils may be in other orders.

6.10.2.2 United States

Aridisols occupy approximately 358,000 km² in 18 of the western states, including Hawaii, generally west of the 100th meridian (Table 6.52). They are most extensive in New Mexico, Nevada, Arizona, Texas, Colorado, and Idaho, occupying more than 30,000 km² in each of those states.

The most southerly Aridisols in the United States occur in the subtropical high pressure belt and receive most of their precipitation in the summer from subtropical convective thunderstorms. The Aridisols of the intermountain west have more continental climates, but precipitation is still dominated by summer thunderstorms driven by orographic lifting in the Rocky Mountains. Farther west, the Aridisols of the Great Basin region receive most of their precipitation in the winter from Pacific frontal storms.

6.10.3 Soil-Forming Environment

The most important aspect of the Aridisol soil-forming environment is the limited water available for plant growth, mineral weathering, and leaching of soluble weathering products. A typical water budget (Fig. 6.57) is dominated by a long period of deficit during the summer when potential evapotranspiration greatly exceeds precipitation. The deficit period is followed by a partial recharge of water-holding capacity, then rapid utilization of that stored water during the brief growing season. The lack of a surplus of precipitation over evapotranspiration and water-holding capacity prevent wholesale leaching of weathering products from Aridisols. Nonetheless, many Aridisols have thin, but morphologically well-developed, soil profiles due to the concentration of the limited water in a relatively small volume of soil. Further, the typically erratic distribution of rainfall in arid regions (Archibold, 1995) causes occasional periods of relatively high precipitation in the winter months, when deeper

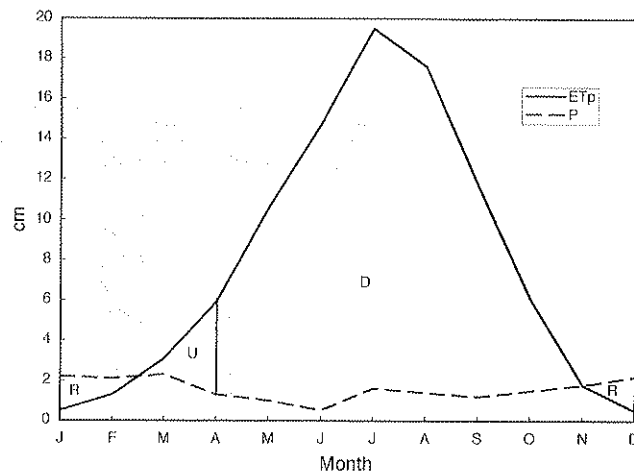


Fig. 6.57 Water budget for St. George, Utah, USA, calculated by the Thornthwaite (1948) method, assuming a soil water-holding capacity of 200 mm. R = recharge of soil water-holding capacity, U = utilization of stored water, D = deficit of precipitation (P) with respect to potential evapotranspiration (ETp)

leaching could occur. In this case, the depth of leaching and horizon development may reflect the rainfall of the extreme years, rather than that of average years.

Soil temperature regimes of Aridisols range from cryic to isohyperthermic (Soil Survey Staff, 1998).

Vegetation on Aridisols is dominated by drought-escaping annual grasses and forbs and drought-enduring evergreen shrubs (Archibold, 1995). Aridisols transitional to xeric or ustic soil moisture regimes, or in frigid or cryic soil temperature regimes, often have a significant component of perennial grasses (Passey et al., 1982). The spatial distribution of plants on the landscape is generally very

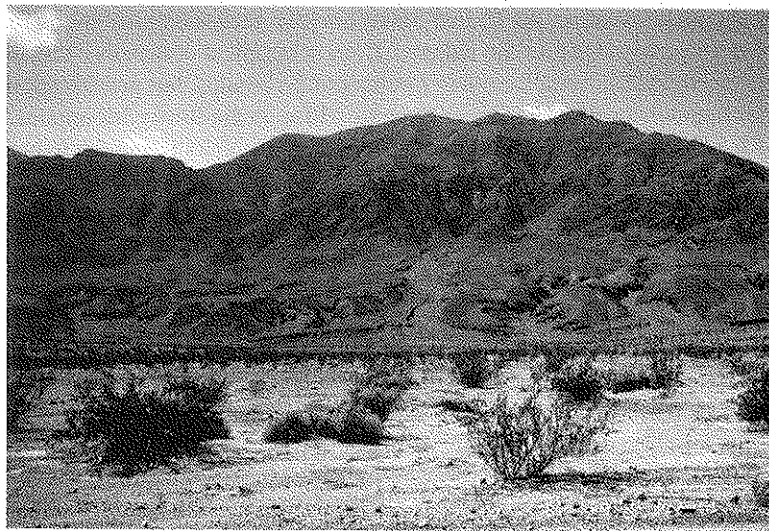


Fig. 6.58 An arid region landscape with desert shrub vegetation (foreground) and active, high-gradient alluvial fans (background), in Owens Valley, California. Coppice dunes around the base of creosote bush (*Larrea tridentata*) often support microbiotic crusts.

heterogeneous (Eckert et al., 1978; Schlesinger and Jones, 1984), giving rise to islands of fertility (Schlesinger et al., 1990) occupied by the shrubs and understory grasses, forbs, and microbiotic crusts (Johnston, 1997). These islands tend to be relatively stable, resistant to erosion (Otterman and Gornitz, 1983), and often trap eolian very fine and fine sand to form a coppice dune (Gile, 1966; Reid et al., 1993) at the base of the shrub (Fig. 6.58). These coppice dunes range in height from a few cm, around the base of shrubs, to 1 m or more, in which case they dominate the landscape microtopography.

Most Aridisols form from sediments transported by wind or water on relatively low gradient landscapes. Lithology of the sediments varies widely. Truncation of steep landforms and slow weathering rates retard the formation of subsurface soil horizons and result in Entisols on most erosional landforms. Some Aridisols have formed in residual rock material on erosional landforms, for example, on old and highly weathered granitic pediments in the Mojave Desert (Boettinger and Southard, 1991; 1995) and on stable, steep hills where episodic erosion has been minimal (Nettleton and Peterson, 1983).

Landscapes with Aridisols range in age from late Holocene to early Pleistocene or older (Nettleton and Peterson, 1983; Machette, 1985). The youngest Holocene landforms generally have not experienced sufficient weathering to produce Aridisols and are dominated by Entisols. Most Aridisols on Pleistocene landscapes have probably experienced fluctuating climatic conditions (cooler, moister Pleistocene pluvials) that significantly affected the rate and intensity of weathering.

6.10.4 Soil Morphology and Physical Properties

6.10.4.1 Surface Horizons

Pedogenic processes in Aridisols have produced numerous surface soil features. Of special interest are (1) surface crusts, (2) vesicular horizons, and (3) desert pavement. A soil crust, widespread, but not present everywhere, is a surficial layer of uncemented, coherent fine earth that can be broken free from underlying soil. The crust is generally less than 10 to 20 mm thick and is often platy in the upper part and massive below. Water infiltration of the crust is slow, in contrast to the rapid infiltration in uncrusted coppice sand dunes. Crusts form by raindrop impact, localized sheet erosion, and by repeated wetting and drying of loamy soil material (Nettleton and Peterson, 1983).

Beneath the crust, especially in silty eolian material, a horizon with many fine, vesicular pores (Fig. 6.59) often forms (sometimes referred to as an Av horizon). The vesicles result from repeated saturation and desiccation and progressive enlargement of air bubbles trapped beneath the thin, saturated surface layer (Springer, 1958; Miller, 1971).

Desert pavement, a surficial layer of rock fragments, usually one or two fragments thick, occurs on many Aridisols (Fig. 6.60). The origin of the pavement has been attributed to loss of fine particles by wind deflation or water erosion, surficial weathering and splitting of some stones (Cooke, 1970), vertical sorting of coarse fragments by repeated wetting and drying (Springer, 1958), and by rafting of surface fragments by infiltrating eolian material (McFadden et al., 1987). The pavement itself serves as a dust trap (Yaalon and Ganor, 1973; Gile, 1975a; Peterson, 1977), whose effectiveness decreases as the surface fragments become more interlocked, until the land surface is nearly completely covered by the pavement. At this point, the surface probably neither retains much new eolian material, nor loses much material to erosion, unless disturbed.

Desert varnish, typically yellowish to reddish brown on the undersides of desert pavement rock fragments, and nearly black on upper surfaces, includes 2:1 silicate clay minerals and oxyhydroxides of Fe and Mn. These constituents are probably of eolian origin and appear to be microbially precipitated (Dorn and Oberlander, 1981). The varnish occurs most frequently on stable, mafic rock frag-

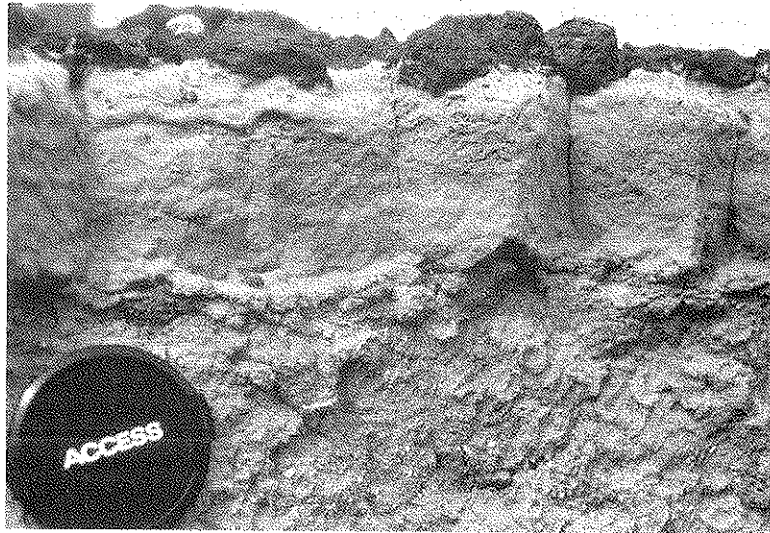


Fig. 6.59 A vesicular surface horizon (an Av horizon) rafts desert-varnished basalt pebbles of a desert pavement in the Cima volcanic field, in the Mojave Desert, California. The lens cap is about 60 mm in diameter.

ments and generally is thickest (as much as 100 μm thick) and darkest on Pleistocene pavements (Nettleton and Peterson, 1983).

The low organic C content and relatively light colors of the surface soil horizon are characteristic of the ochric epipedons of virtually all Aridisols (dark anthropic epipedons occur, but are very rare and localized).

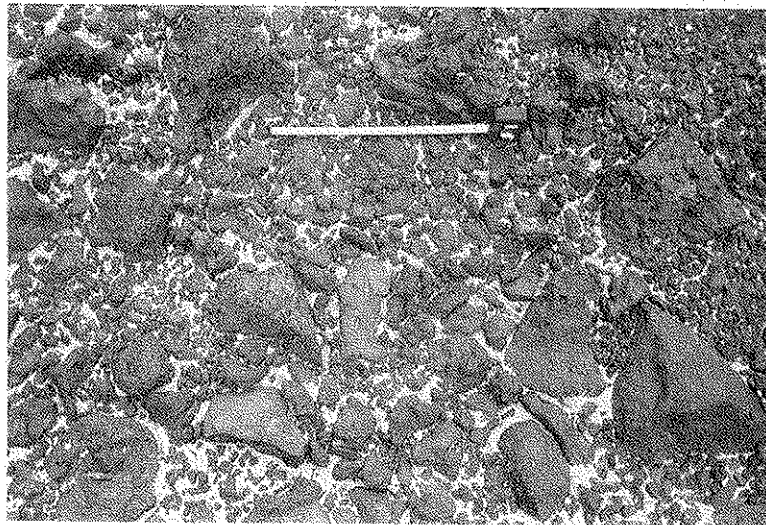


Fig. 6.60 Desert varnished surface clasts interlock to form a nearly continuous desert pavement on Pleistocene basalt flows in the Mojave Desert. The clasts act as a dust trap to catch eolian materials derived from nearby playas and alluvial fans. The length of the tape measure scale is about 30 cm.

Table 6.53 Summary of properties of subsurface horizons in Aridisols [Soil Survey Staff, 1998]

Horizon	Properties
Argillic	Clay accumulation relative to overlying horizon, evidence of clay translocation, minimum thickness of 7.5 cm if loamy or clayey, 15 cm if sandy
Calcic	CaCO ₃ accumulation, generally ≥ 15 cm thick, $\geq 15\%$ CaCO ₃ equivalent, $\geq 5\%$ more CaCO ₃ than underlying horizon
Cambic	Evidence of alteration, reddening, development of soil structure, loss of carbonates
Duripan	Silica-cemented, does not slake in water or acid, evidence of opal accumulation
Gypsic	Gypsum accumulation, ≥ 15 cm thick, $\geq 5\%$ gypsum, $\geq 1\%$ visible secondary gypsum, thickness in cm multiplied by % gypsum has product ≥ 150
Petrocalcic	CaCO ₃ -cemented, ≥ 10 cm thick, or ≥ 1 cm thick if a laminar cap is underlain directly by bedrock
Petrogypsic	Gypsum-cemented, ≥ 10 cm thick, $\geq 5\%$ gypsum and thickness in cm multiplied by % gypsum has product ≥ 150
Natric	Properties of argillic horizon, plus columnar or prismatic structure, or degraded blocky structure, sodium adsorption ratio (SAR) ≥ 13 or exchangeable sodium percentage (ESP) ≥ 15
Salic	Accumulation of salts more soluble than gypsum, ≥ 15 cm thick, electrical conductivity (EC) ≥ 30 dS m ⁻¹ , thickness in cm multiplied by EC has product ≥ 900

6.10.4.2 Subsurface Horizons

Subsurface horizons include cambic, argillic, natric, calcic, petrocalcic, gypsic, petrogypsic and salic horizons, and duripans. None of these horizons is unique to Aridisols, although the occurrence of salic, gypsic, and petrogypsic horizons in other soil orders is relatively rare. A summary of properties of these horizons is given in Table 6.53.

6.10.5 Chemical Properties**6.10.5.1 Organic Carbon and Nitrogen**

Organic C contents are among the lowest of all soils, often $< 1\%$ on a mass basis, due to low biomass production and rapid decomposition. Accumulation of organic C is confined to relatively shallow soil depths in the thin ochric epipedon. Organic C content generally is inversely related to mean annual soil temperature. In cool and cold arid regions (frigid and cryic soil temperature regimes), organic C contents may rival those of semi-arid to subhumid climates (Table 6.54).

Nitrogen occurs mostly in organic matter, and is usually quickly nitrified after the organic matter is decomposed. The NO₃⁻ produced is susceptible to some leaching, but more importantly, to runoff during overland flow and accumulation in a playa environment, where denitrification can occur (Schlesinger et al., 1990).

6.10.5.2 Soluble Salts

Limited leaching and concentration of soluble ions by evaporation near the soil surface causes accumulation of soluble salts (gypsum and more soluble evaporites) (Section 6.10.6). The distribution of salts in Aridisol profiles reflects the relative solubilities of the compounds and the relative importance of downward leaching from precipitation versus the upward movement of groundwater due to

Table 6.54 Selected properties^a of representative Aridisol pedons [Data from Soil Survey Staff, 1975 and USDA-NRCS, 1997b]

Depth cm	Horizo n	Sand %	Silt %	Clay %	OC ^a %	N ^a %	CCE ^a %	BD ^a g cm ⁻³	pH ^b	EC ^a dS m ⁻¹	SAR ^a
Xeric Calcicryid											
0 - 10	A	32	56	12	2.98	0.275	15	0.95	7.7	1.7	tr
10 - 18	Bk1	30	55	15	1.83	0.189	80		8.2	1.0	1
18 - 33	Bk2	34	44	22	0.78	0.078	41	1.17	7.8	12.9	18
33 - 58	Bk3	35	41	24	0.37	0.037	73	1.27	8.0	19.0	21
58 - 84	Bkq1	31	34	35	0.38		54	1.11	7.9	19.6	19
84 - 107	Bkq2	31	30	39	0.46		88		7.7	30.2	19
107 - 152	Bkq3	38	32	30	0.23		62		7.8	24.3	18
Typic Haplosalid											
0 - 11	A	1	50	49	0.69	0.064	4		8.4 ^b	22.5	67
11 - 15	Cyz	1	44	55	0.38	0.039	2		8.7	116.7	281
15 - 20	Czy	2	55	43	0.60	0.063	3		8.8	122.5	553
20 - 36	Cz	10	53	37	0.44	0.046	4		8.4	83.8	201
36 - 56	C1	17	54	29	0.27	0.030	6		8.4	66.7	110
56 - 76	C2	2	58	40	0.26	0.033	13		8.6	63.0	110
76 - 105	C3	4	78	18	0.18	0.019	11		8.6	37.7	74
105 - 150	C4	4	53	43	0.21	0.028	13		8.6	38.3	34
Typic Haplodurid											
0 - 3	A1	86	12	2	0.11	0.013	3		8.2 ^b	0.4	1
3 - 10	A2	81	14	5	0.08	0.015	5		8.3	0.4	1
10 - 41	A3	75	14	11	0.15	0.021	7		8.3	0.3	1
41 - 58	Bqkm1	87	7	6	0.20	0.019	14		8.1	0.4	4
58 - 76	Bqkm2	81	13	6	0.03		9		8.4	0.4	10
76 - 94	C	87	9	4	0.01		3		8.5	0.5	15
94 - 140	Cqkm	84	12	4	0.01		14		8.3	1.0	7
140 - 175	C'	90	8	2			4		8.4	0.9	6
Typic Calcigypsid											
0 - 0.5	A1	31	43	26	1.27	0.112	12		7.9	2.5	
0.5 - 5	A2	28	42	30	0.96	0.094	12		7.9	3.8	3
5 - 13	A3	29	42	29	0.93	0.088	12	1.37	7.9	4.0	3
13 - 33	Bk1	25	40	35	0.63	0.066	19		7.8	9.2	9
33 - 56	Bk2	22	40	38	0.63	0.068	26	1.22	7.4	29.1	8
56 - 84	Cy1	20	30	50	0.26	0.025	10	1.30	7.6	21.3	
84 - 115	Cy2	10	50	39	0.14		8		7.7	17.4	2
115 - 140	Cy3	15	41	44	0.16		14		7.8	16.6	6
140 - 180	Cy4	14	43	43	0.18		11		7.9	21.3	7
Typic Haplargid											
0 - 3	A1	58	28	14	0.98	0.095	-	1.49	6.9 ^b	0.5	
3 - 12	A2	60	23	17	0.26	0.038	-	1.50	6.6	0.5	
12 - 23	BA1	57	22	21			-	1.63	6.9	0.3	
23 - 46	Bt	46	24	30			-	1.47	7.5	0.2	
46 - 66	Btk1	40	31	29			3	1.39	7.9	0.2	
66 - 84	Btk2	33	31	36			8	1.41	8.0	0.2	
84 - 110	Btk3	31	39	29			15	1.34	8.1	0.3	
110 - 130	BCtk1	34	40	26			16	1.45	8.1	0.2	
130 - 160	BCtk2	33	44	23			18	1.60	8.3	1.0	7
160 - 180	BCtk3	27	49	24			12	1.42	8.1	2.7	11
180 - 230	C	32	49	19			9	1.37	7.9	6.8	10

^a OC = organic carbon, N = nitrogen, CCE = calcium carbonate equivalent, BD = bulk density at 33 kPa water potential, EC = electrical conductivity, and SAR = sodium adsorption ratio.

^b pH of 1:1 soil:water mixture, unless noted by ^b, then pH of saturated paste.

Table 6.54 cont.

Depth cm	Horizon	Sand %	Silt %	Clay %	OC ^a %	N ^a %	CCE ^a %	BD ^a g cm ⁻³	pH ^b	EC ^a dS m ⁻¹	SAR ^a
Argic Petrocalcic											
0 - 5	A	85	5	10	0.25		-				
5 - 18	BAt1	87	4	9	0.13		-	1.86 ^c			
18 - 25	BAt2	82	5	13	0.17		-	1.78			
25 - 36	Bt	80	5	15	0.14		1	1.70			
36 - 48	Btk	77	6	17	0.24		16	1.53			
48 - 74	Bkm1	68	9	23	0.15		75				
74 - 100	Bkm2	71	11	18	0.05		65	1.68			
100 - 150	Bkm3	75	6	19	0.05		51	1.66			
Typic Haplocambid											
0 - 6	A	47	43	10	0.52		tr	1.41	8.0		
6 - 15	Bw1	44	45	12	0.31		tr	1.26	8.3		
15 - 33	Bw2	46	43	11	0.27		tr	1.22	8.5		
33 - 48	Bk1	44	44	12	0.35		4	1.22	8.7	0.7	5
48 - 60	Bk2	65	29	6	0.28		tr	1.24	8.7	1.7	10
60 - 87	Bk3	49	44	7	0.22		tr	1.17	8.2	7.8	15
87 - 135	C1	39	52	9	0.22		tr	1.24	7.9	13.3	14
135 - 150	C2	66	25	9	0.12		tr	1.30	7.8	11.8	12
150 - 175	C3	62	29	9	0.12		tr	1.30	7.7	11.1	11
175 - 205	C4	64	22	14	0.10		tr	1.30	7.9	10.6	9

^a OC = organic carbon, N = nitrogen, CCE = calcium carbonate equivalent, BD = bulk density at 33 kPa water potential, EC = electrical conductivity, and SAR = sodium adsorption ratio.

^b pH of 1:1 soil:water mixture, unless noted by ^b, then pH of saturated paste.

^c Bulk density at 33 kPa water potential, unless noted by ^c, then air-dry.

evaporation. Typically, well-drained Aridisols have accumulations of the most soluble salts (chlorides and sulfates) at greater depths than those that are less soluble (gypsum and calcium carbonate), whereas more poorly drained Aridisols have accumulations of soluble salts closest to the surface. The distribution of these salts within the soil profile is reflected in the electrical conductivity (EC) of the soil solution (Table 6.54).

6.10.5.3 Calcium Carbonate

Calcium carbonate is present in almost all Aridisols, except for some Aridisols on stable platforms in West Africa and Australia. Inputs of Ca from mineral weathering or from the atmosphere, either dissolved in rain water or as particles of Ca-bearing minerals, coupled with relatively low solubility, cause CaCO₃ to accumulate. Once in soil solution, Ca can react with HCO₃⁻ from rainwater or that produced by the reaction of soil CO₂ (from respiration) with water. Loss of water by evaporation, or reduction of the partial pressure of CO₂ (Pco₂), drives the precipitation of CaCO₃ from solution.

Aridisols on Holocene or latest Pleistocene landscapes are often calcareous in all horizons, because the relatively dry conditions of the Holocene prevent leaching of carbonate from surface horizons (e.g., the Typic Haplodurid and Typic Calcigypsid in Table 6.54). In contrast, Aridisols on older Pleistocene landscapes are often carbonate-free in the upper horizons, presumably due to more effective leaching during the cooler, moister Pleistocene (e.g., the Typic Haplargid and Argic Petrocalcic in Table 6.54). Aridisols of West Africa are acidic and lack carbonates because they form on old pre-weathered platform surfaces or from more recent polycyclic sediments derived therefrom (L.P. Wilding, 1999, personal communication).

6.10.5.4 Exchangeable Sodium and Alkalinity

The accumulation of Na on the cation exchange complex is a common phenomenon in Aridisols and is related to conditions in soil solution [Na adsorption ratio (SAR)] (Table 6.54). Hydrolysis of exchangeable Na, particularly when soil solution electrolyte concentrations are low (i.e., little or no soluble salts) can produce alkaline soil reactions near pH 10. Swelling and dispersion of soil clays can result, in some cases causing plugging of soil pores and reducing permeability (Section G, Chapter 2). Clay dispersion also enhances the mobility of clay particles in the soil solution and aids the accumulation of clay in subsoil horizons, particularly natric horizons.

6.10.6 Mineralogy

Most minerals in Aridisols are inherited from the parent material (Nettleton and Peterson, 1983; Allen, 1990). The arid climate provides little surplus water for mineral hydrolysis or for leaching of weathering products. As a result, primary silicates experience little chemical weathering and Si-rich smectites are expected to dominate the clay fraction. Some Aridisols are kaolinitic and reflect paleoenvironmental conditions in regions of West Africa and Australia (L.P. Wilding, 1999, personal communication).

In certain soil-forming environments, there is clear evidence that mineral weathering is an important process and that some minerals are pedogenic. For example, on the deeply weathered granitic pediments of the Mojave Desert, alteration of biotite to vermiculite is extensive. This alteration and neosynthesis of Al-rich smectites from feldspar weathering are the main sources of phyllosilicates in the clay fraction (Boettinger and Southard, 1995). It is likely that much of this weathering is a result of more humid conditions during the Pleistocene.

The pedogenic formation of the fibrous clays, palygorskite [ideal structure: $\text{Si}_8\text{Mg}_5\text{O}_{20}(\text{OH})_2(\text{OH}_2)_4 \cdot 4\text{H}_2\text{O}$] and sepiolite [ideal structure: $\text{Si}_{12}\text{Mg}_8\text{O}_{30}(\text{OH})_4(\text{OH}_2)_4 \cdot 8\text{H}_2\text{O}$] (Singer, 1989) has been documented in the alkaline, Si-preserving, environment of petrocalcic horizons.

In most other Aridisols, the sand and silt fractions reflect relatively little mineral weathering, and soil mineral composition is dominated by the mineralogy of the parent material. Generally, the silicate mineral suite is dominated by quartz, feldspar, amphiboles, and micas. A few Aridisols contain abundant inherited zeolites (Allen, 1990).

The least intensively leached Aridisols may have significant contents of relatively soluble, nonsilicate minerals, including calcite (CaCO_3), gypsum ($\text{CaSO}_4 \cdot 2\text{H}_2\text{O}$), halite (NaCl), thenardite (Na_2SO_4),

Table 6.55 Mineralogical classes of Aridisol soil series in the United States [USDA-NRCS, 1997a]

Mineralogy class	Sandy and loamy, including skeletal families	Clayey and clayey-skeletal families	Substitute class families	Contrasting particle- size class families
Mixed	1651	126		82
Smectitic		316		9
Carbonatic	136	4		
Gypsic	31			
Siliceous	11			
Glassy			7	2
Illitic		3		
Kaolinitic		2		

mirabilite ($\text{Na}_2\text{SO}_4 \cdot 10\text{H}_2\text{O}$), epsomite ($\text{MgSO}_4 \cdot 7\text{H}_2\text{O}$), nahcolite (NaHCO_3), and trona ($\text{Na}_2\text{CO}_3 \cdot 10\text{H}_2\text{O}$) (Doner and Lynn, 1989).

The mixture of minerals in the sand and silt fractions and the dominance of smectites (usually montmorillonite) in the clay fraction are reflected by the mineralogical classes of *Soil Taxonomy* (Table 6.55).

6.10.7 Soil-Forming Processes

6.10.7.1 Eolian Accretion

The significant role of eolian additions to many Aridisols is well documented (Yaalon and Ganor, 1973; Gile and Grossman, 1979; Reheis, 1987). The role of eolian dust is particularly important in many Holocene Aridisols, where nearly all of the carbonate, gypsum, and opaline silica are derived from an external source (Harden et al., 1991). In some cases, eolian material constitutes nearly the entire fine-earth fraction and accounts for the nearly stone-free vesicular horizons below many desert pavements (McFadden et al., 1987). It is clear that an external source of carbonates is required for the development of thick petrocalcic horizons in soils formed from noncalcareous parent materials (Gile et al., 1966). The large volume of primary rocks that would need to be weathered to produce the carbonates present in many Aridisols is unrealistic, given the limited availability of water for silicate mineral weathering (Birkeland, 1984). Landscapes derived from rock types that tend to weather to soils with few surface clasts are less likely to have significant eolian accretion because an effective dust trap is absent (Boettinger and Southard, 1991). Those Aridisols derived from limestone bedrocks or sediments derived therefrom may have some accumulation of eolian dusts, but dust impact in terms of carbonate source may be of lesser importance in these soils (Rabenhorst and Wilding, 1986).

6.10.7.2 Mineral Weathering

Mineral weathering in Holocene Aridisols is minimal, due to the aridity of the Holocene climate. Aridisols on Pleistocene landscapes show clear signs of weathering of primary silicate minerals to produce Ca- and Mg-rich clays (Harden et al., 1991), interstratified mica-vermiculite (Eghbal and Southard, 1993a), vermiculite and Al-rich smectite (Boettinger and Southard, 1995), and Ca and silica in calcareous duripans (Boettinger and Southard, 1991; Eghbal and Southard, 1993c).

6.10.7.3 Clay Accumulation

The processes by which clay accumulates in Aridisols are similar to those that occur in other soil orders (Section E, Chapters 6.11, 6.12). Argillic and natric horizons, ranging in texture from sand to clay, and in thickness from 7.5 to 75 cm, generally begin at shallow depths (4 to 25 cm) below the soil surface, and, thus, are shallower than argillic and natric horizons in other soils.

Major processes of clay accumulation in subsurface horizons include disaggregation and dispersion of clays in eolian dust accompanied by burial by additional increments of dust (eolian accretion), *in situ* weathering of primary minerals, illuviation of dispersed clay from surface horizons, selective loss of fine particles from surface horizons by deflation, sheet flow, or dissolution, and neosynthesis of clay minerals from soil solution. Mineral weathering to produce clays susceptible to dispersion and translocation is primarily a relict of wetter Pleistocene climatic conditions (Nettleton and Peterson, 1983). Further, wetter conditions of the Pleistocene pluvials were probably required to dissolve and leach carbonates from surface horizons, the removal of Ca being a prerequisite for dispersion and translocation of the clays.

Most Aridisols with argillic horizons occur on the earliest Holocene and older landscapes, and the minimum time required to form an argillic horizon is largely a function of the carbonate content of the parent material and the rate at which Ca is added by eolian dust (Gile, 1975b). Exchangeable Na increases the rate of subsurface clay accumulation. Enhanced dispersion of clays, even in calcareous soils, causes natric horizons to form in less than 6,600 years (Alexander and Nettleton, 1977; Peterson, 1980).

6.10.7.4 Calcium Carbonate and Soluble Salt Redistribution

The depth and magnitude of CaCO_3 accumulation in many Aridisols illustrate the importance of varying eolian input and Pco_2 (McFadden et al., 1991). Formation of pedogenic Bk horizons is by dissolution of carbonates from upper horizons and precipitation from soil solution in lower horizons. Biogenic CO_2 raises Pco_2 to levels as high as 1.2%, and the relative rates of CO_2 production and loss to the atmosphere by diffusion generally cause Pco_2 to increase with soil depth (McFadden et al., 1991). Plant roots take up water preferentially over dissolved HCO_3^- , causing CaCO_3 to precipitate due to desiccation.

The progressive accumulation of carbonate in many Aridisols (Table 6.56, Fig. 6.61) reflects variable, but more or less continuous, input of Ca in eolian material, or from indigenous calcareous parent materials, and a large expansion of the volume of the calcic and petrocalcic horizons as carbonate precipitates and forces matrix grains apart. Cementation of horizons occurs when carbonate content reaches 25 to 60% (Machette, 1985).

Gile et al. (1965, 1966) proposed that a master K horizon be used to designate the more advanced stages of carbonate accumulation. The proposal was not formally adopted (the formal designation is Bkm), but K horizon is still used (Machette, 1985), as are the terms caliche and calcrete, although these latter terms are not well defined. The preferred terms for horizons of CaCO_3 accumulation are calcic horizon (often designated Bk) if not cemented, and petrocalcic if cemented (Soil Survey Staff, 1996). Many petrocalcic horizons may be relicts of older land forms and not related to present soil profiles, although upper, laminar layers may be the result of present-day processes of localized dissolution and reprecipitation (Harden et al., 1991; Eghbal and Southard, 1993b).

Gypsic and salic horizons are typical of basins with playas (Driessen and Schoorl, 1973) where leaching is limited by a high watertable and where eolian salts are recycled from the playa to surrounding terrain. The distribution of these salts in soil profiles is often seasonally dynamic due to their high solubility and the ease with which they are moved downward by leaching during the wet season and upward by evapotranspiration during the dry season (Eghbal et al., 1989).

Table 6.56 Stages of calcium carbonate accumulation in Aridisols [Gile et al., 1966; Machette, 1985]

Stage	General character	Carbonate morphology-----	
		Gravelly soils	Nongravelly soils
I	Weakest expression of macroscopic carbonate	Thin, discontinuous pebble coating, usually on undersides.	Few filaments or faint coatings
II	Carbonates segregated	Continuous pebble coatings, some interpebble fillings.	Few to common nodules
III	Carbonate continuous, plugged horizon forms in last part	Many interpebble fillings.	Many nodules and internodular fillings
Any soil			
IV	Thin laminar layer	Laminar layer < 1 cm thick over plugged horizon.	
V	Thick laminar layer	Laminar layer > 1 cm thick over plugged horizon. Pisolites. Vertical faces laminated.	
VI	Multiple laminae, recemented, relaminated	Multiple generations of laminae, breccia, and pisolites. Many case-hardened vertical faces.	

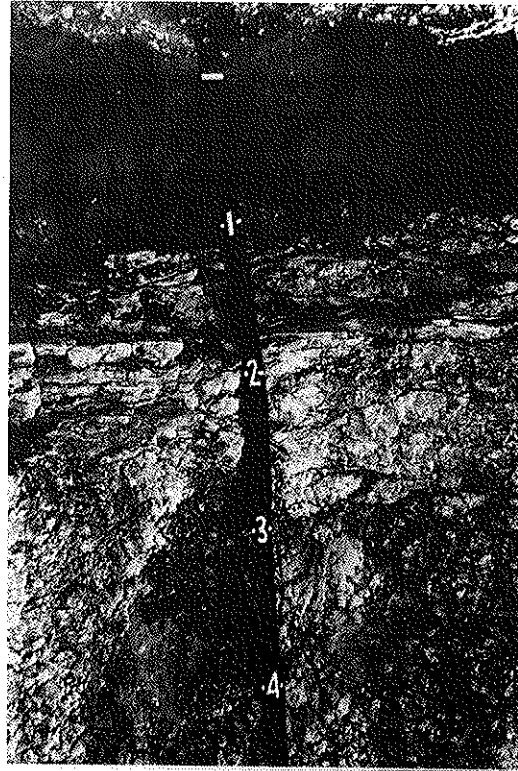


Figure 6.61 Profile of an Argic Petrocalcic in New Mexico, with a platy, Stage V petrocalcic horizon about two feet (60 cm) thick. The scale on the tape is in feet.

6.10.7.5 Silica Accumulation

Silica accumulates in the form of opal ($\text{SiO}_2 \cdot n\text{H}_2\text{O}$) (Flach et al., 1973; Chadwick et al., 1987a), and possibly as chalcedony and quartz (Flach et al., 1973; Smale, 1973). Incipient cementation produces horizons that are very hard when dry and brittle when moist. More complete cementation of the soil by opal produces a duripan. Unlike the cementation process in petrocalcic horizons, wherein silicate grains in the matrix are forced apart, cementation by opal tends to be less disruptive of the silicate grains in the soil fabric, and relatively small amounts of silica (about 10% by weight) cause signifi-

Table 6.57 Ratios of 1,500 kPa water content and CEC to clay in Aridisols with indurated (typic) duripans and less strongly cemented (haplic) duripans from Nevada and Arizona [After Southard et al., 1990]

1,500 kPa water/clay		CEC/clay	
Bw or Bt horizon	Bqm or Bqkm horizon	Bw or Bt horizon	Bqm or Bqkm horizon
Typic (n = 9)			
0.56 ± 0.15^a	1.15 ± 0.22	1.02 ± 0.39	2.03 ± 1.23
Haplic (n = 15)			
0.52 ± 0.14	1.47 ± 0.71	1.09 ± 0.46	3.24 ± 1.92

^a Mean $\pm$ standard deviation.

cant cementation of clay particles (Nettleton and Peterson, 1983), significantly changing water retention characteristics and CEC (Table 6.57).

Duripans have been called silcretes and duricrust (Jackson, 1957), although some are largely of geologic, not pedologic, origin. Most duripans in Aridisols have a considerable content of CaCO_3 , as much as 70% by weight (Southard et al., 1990), but at least half the volume of a duripan does not slake in acid, as petrocalcic horizons do. Many of these duripans have indurated, laminar upper layers as much as 2 cm thick that overlie massive and less strongly cemented, highly calcareous material. These duripans can be distinguished from a petrocalcic horizon only by the acid treatment. In many cases only the upper laminar layer survives the acid treatment intact. The silica cements are derived by weathering of siliceous rocks (Boettinger and Southard, 1991), volcanic ash (Chadwick et al., 1987b), and eolian dust (Harden et al., 1991).

6.10.7.6 Polygenesis

Evidence of leaching below the calculated average depth of water storage (Arkley, 1963) is often observed in Aridisols and ascribed to more humid Pleistocene paleoclimates (Nettleton and Peterson, 1983), and to variation in the accretion of eolian material and rates of weathering attributable to Quaternary climate change (Wells et al., 1987). Evidence for climatically driven shifts in the dominant soil-forming processes includes engulfment of carbonate-free argillic horizons, presumably leached of carbonates during the Pleistocene, by eolian-derived carbonates during the more arid Holocene (Nettleton and Peterson, 1983) and micromorphologic evidence of episodic deposition of carbonates, opal, and clay in argillic horizons and duripans (Eghbal and Southard, 1993c).

6.10.8 Classification

At the order level, Aridisols are differentiated from other soils by an aridic soil moisture regime, an ochric or anthropic epipedon, and one or more of the following subsurface horizons within 100 cm of the soil surface: argillic, cambic, natric, salic, gypsic, petrogypsic, calcic, petrocalcic, or duripan (Table 6.58). These subsurface horizons, and the cryic soil temperature regime, are used to differentiate seven suborders. Prior to 1994, only two suborders existed: Argids, with an argillic or a natric horizon, and Orthids. Great groups of the Orthids were differentiated on the basis of the other diagnostic horizons now used to identify the suborders (Witty, 1990), thus placing more emphasis on the presence of a horizon of clay accumulation.

The Cryids have a cryic soil temperature regime and occur at high latitudes and high elevations. These soils, in which soil processes are inhibited not only by lack of water, but also by low temperatures, are common in the intermountain western region of the United States (Hipple et al., 1990). The Argids, Durids, and Petrocalcids (petrocalcic horizon) occur on the oldest geomorphic surfaces, usually late Pleistocene and older, as on the least eroded parts of dissected alluvial fans or on hillslopes protected from significant erosion by vegetation. Cambids and other Calcids are found on geologically younger side slopes and surfaces of intermediate age, usually latest Pleistocene and younger. Gypsid and Salids often occur near playa margins, where salts are concentrated at or near the soil surface by upward flux from a watertable driven by evaporation. Less commonly, they occur in association with Cambids and Calcids where parent materials are saline or where gypsum is added with eolian material (Reheis, 1987; Eswaran and Gong, 1991).

The great groups are differentiated on the basis of a number of other properties, including aquic conditions, cementation of horizons by gypsum or carbonates, and presence of diagnostic subsurface horizons not identified at the suborder level, generally at depths between 100 and 150 cm. Subgroups

Table 6.58 Listing of suborders, great groups and subgroups of Aridisols order [Soil Survey Staff, 1998]

Suborders	Great groups	Subgroups
Cryids	Salicryids	Aquic, Typic
	Petrocryids	Xereptic, Duric Xeric, Duric, Petrogypsic, Xeric, Ustic, Typic
	Gypsicryids	Calcic, Vitrixerandic, Vitrandic, Typic
	Argicryids	Lithic, Vertic, Natric, Vitrixerandic, Vitrandic, Xeric, Ustic, Typic
	Calcicryids	Lithic, Vitrixerandic, Vitrandic, Xeric, Ustic, Typic
Salids	Haplocryids	Lithic, Vertic, Vitrixerandic, Vitrandic, Xeric, Ustic, Typic
	Aquisalids	Gypsic, Calcic, Typic
	Haplosalids	Duric, Petrogypsic, Gypsic, Calcic, Typic
Durids	Natridurids	Vertic, Aquic Natrargidic, Aquic, Natrixeralfic, Natrargidic, Vitrixerandic, Vitrandic, Xeric, Typic
	Argidurids	Vertic, Aquic, Abruptic Xeric, Abruptic, Haploxeralfic, Argidic, Vitrixerandic, Vitrandic, Xeric, Ustic, Typic
	Haplodurids	Aquicambidic, Aquic, Xereptic, Cambidic, Vitrixerandic, Vitrandic, Xeric, Ustic, Typic
Gypsid	Petrogypsid	Petrocalcic, Calcic, Vitrixerandic, Vitrandic, Xeric, Ustic, Typic
	Natrigypsid	Lithic, Vertic, Petronodic, Vitrixerandic, Vitrandic, Xeric, Ustic, Typic
	Argigypsid	Lithic, Vertic, Calcic, Petronodic, Vitrixerandic, Vitrandic, Xeric, Ustic, Typic
	Calcigypsid	Lithic, Petronodic, Vitrixerandic, Vitrandic, Xeric, Ustic, Typic
	Haplogypsid	Lithic, Leptic, Sodc, Petronodic, Vitrixerandic, Vitrandic, Xeric, Ustic, Typic
Argids	Petroargids	Petrogypsic Ustic, Petrogypsic, Duric Xeric, Duric, Natric, Xeric, Ustic, Typic
	Natrargids	Lithic Xeric, Lithic Ustic, Lithic, Vertic, Aquic, Durinodic Xeric, Durinodic, Petronodic, Glossic Ustic, Haplic Ustic, Haploxeralfic, Haplic, Vitrixerandic, Vitrandic, Xeric, Ustic, Glossic, Typic
	Paleargids	Vertic, Aquic, Arenic Ustic, Arenic, Calcic, Durinodic Xeric, Durinodic, Petronodic Ustic, Petronodic, Vitrixerandic, Vitrandic, Xeric, Ustic, Typic
	Gypsiargids	Aquic, Durinodic, Vitrixerandic, Vitrandic, Xeric, Ustic, Typic
	Calciargids	Lithic, Xerertic, Ustertic, Vertic, Aquic, Arenic Ustic, Arenic, Durinodic Xeric, Durinodic, Petronodic Xeric, Petronodic Ustic, Petronodic, Vitrixerandic, Vitrandic, Xeric, Ustic, Typic
	Haplargids	Lithic Ruptic-Entic, Lithic Xeric, Lithic Ustic, Lithic, Xerertic, Ustertic, Vertic, Aquic, Arenic Ustic, Arenic, Durinodic Xeric, Durinodic, Petronodic Ustic, Petronodic, Vitrixerandic, Vitrandic, Xeric, Ustic, Typic
	Petrocalcids	Aquic, Natric, Xeralfic, Ustalfic, Argic, Calcic lithic, Calcic, Xeric, Ustic, Typic
	Haplocalcids	Lithic Xeric, Lithic Ustic, Lithic, Vertic, Aquic Durinodic, Aquic, Duric Xeric, Duric, Durinodic Xeric, Durinodic, Petronodic Xeric, Petronodic Ustic, Petronodic, Sodc Xeric, Sodc Ustic, Sodc, Vitrixerandic, Vitrandic, Xeric, Ustic, Typic
	Aquicambids	Sodic, Durinodic Xeric, Durinodic, Petronodic, Vitrixerandic, Vitrandic, Fluventic, Xeric, Ustic, Typic
	Petrocambids	Sodic, Vitrixerandic, Vitrandic, Xeric, Ustic, Typic
Cambids	Anthracambids	Typic
	Haplocambids	Lithic Xeric, Lithic Ustic, Lithic, Xerertic, Ustertic, Vertic, Durinodic Xeric, Durinodic, Petronodic Xeric, Petronodic Ustic, Petronodic, Sodc Xeric, Sodc Ustic, Sodc, Vitrixerandic, Vitrandic, Xerifluventic, Ustifluventic, Fluventic, Xeric, Ustic, Typic

are differentiated by a large number of properties. Some common criteria are shallow lithic contacts, soil properties that grade toward Andisols, and soil moisture regimes that border on ustic and xeric.

6.10.9 Management

Agricultural use of Aridisols is limited chiefly by the lack of water. Where water is not available, most Aridisols are used for grazing by livestock and wildlife. Where irrigation systems have been

developed, most Aridisols present some problem with irrigation and generally are less well suited for irrigation than are the Entisols (e.g., Torriorthents and Torrifluvents) that are commonly associated with the Aridisols. Land leveling to allow flood, furrow, or sprinkler irrigation often exposes calcic, petrocalcic, natric, or argillic horizons, or duripans.

Under irrigation, control of salinity in the rooting zone is critical to the long-term success of any crop management system. Flushing of salts by an amount of water in excess of the crop evapotranspiration demand (a leaching fraction) helps reduce salinity in the root zone. The excess water and dissolved salts must be removed by subsurface drainage. The presence of shallow subsurface horizons with slow permeability (duripans and natric and petrocalcic horizons) compounds the irrigation and drainage management problems for crop production, as well as for septic tank leach fields and irrigation of lawns in urbanized areas. Soil subsidence, due to solution of gypsum, and corrosion of concrete are common problems where soils containing gypsum are irrigated (Nettleton et al., 1982). Concrete corrosion occurs when sulfate reacts with Na to form mirabilite or thenardite (section 6.10.6), or with Ca and Al to form ettringite $[\text{Ca}_6\text{Al}_2(\text{SO}_4)_3(\text{OH})_{12}\cdot 26\text{H}_2\text{O}]$. Crystallization of these sulfate minerals increases the volume of solids, leading to the disintegration of the concrete.

Other than N, most major plant nutrients are abundant in Aridisols, although micronutrient availability is often limited by low solubility at alkaline pH. Careful management of P is also often needed due to sorption of P by CaCO_3 .

6.10.10 References

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6.11 Alfisols

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6.11.1 Introduction

The central concept of Alfisols embraces soils of the semi-arid to humid climates with light colored surfaces and subsoils moderately rich in basic cations, formed, at least in part, by movement of clay from overlying horizons into the subsoil. Most Alfisols developed under deciduous forest vegetation in humid climates and grass or savannah vegetation in drier regimes. In loamy and silty Alfisols of the

humid regions, fragipans are not uncommon; in drier climates, incomplete leaching of salts and carbonates can result in concentrations of secondary carbonates, gypsum, amorphous silica or enrichment of exchangeable Na. As the processes of formation of Alfisols require both time and energy, most are found on geomorphic surfaces that are of Pleistocene age or older. They form on a wide variety of parent materials.

In previous classification systems in the United States, most Alfisols were in the Gray-Brown Podzolic Great Group, but others were in the Low-Humic Gley, Noncalcic Brown, Reddish-Brown Lateritic, Reddish Brown, Reddish Chestnut, Gray Wooded and Planosol Great Groups. Further, natric great groups or subgroups of Alfisols also include salt-affected soils that were previously classified as Solonetz and Soloth soils (Soil Survey Staff, 1975).

6.11.2 Distribution

Alfisols occur on all continents except Antarctica (Fig. 6.62). They occupy about 9.6% of the ice-free Earth's surface (Table 6.1). They are found between the latitudes of 65° N and 45° S and concentrated in the Northern Hemisphere, specifically North America, Europe and east Central Asia between latitudes of 30° N and 65° N; also, significant occurrences are found in subsaharan Africa, India, South America, and Australia. Cryalfs are found within the colder reaches such as Northern Europe and into Western Russia. Ustalfs are common in subsaharan Africa, Eastern Brazil, the eastern portion of India, and Southeastern Australia. Udalfs are prominent in Central Europe and Eastern Australia, while Xeralfs are found in countries bordering the Mediterranean Sea.

Within the United States, Cryalfs are found in the north central states and at higher elevations in the western states. Udalfs are prominent in the eastern portion of the Midwest, along the Mississippi and Ohio River valleys while Ustalfs occur to the south and west of the Great Plains as well as the southern portion of the High Plains. Most Xeralfs are restricted to the west coast states.

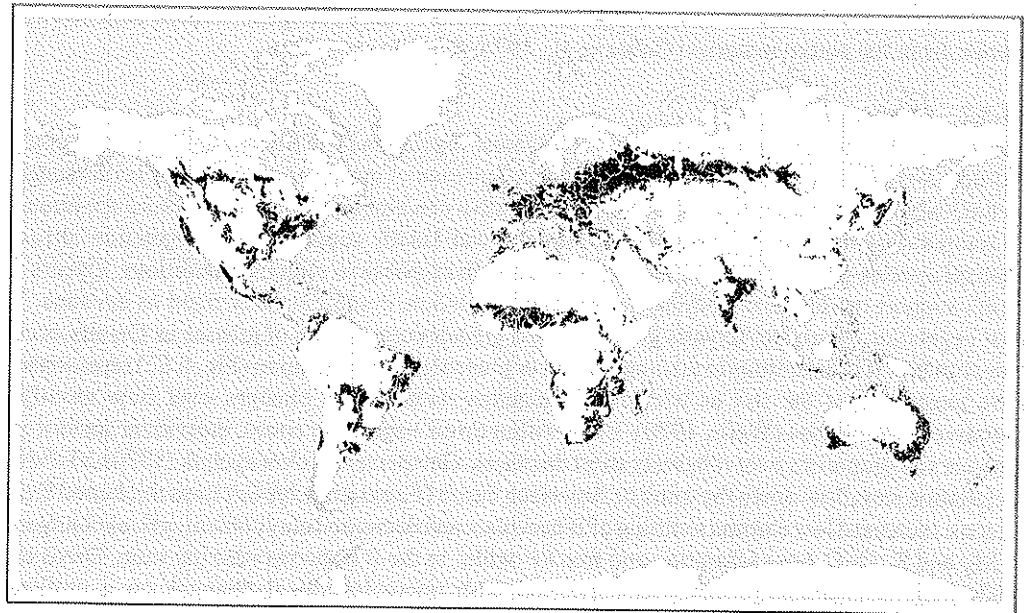


Fig. 6.62 Global distribution of Alfisols [Courtesy of USDA-NRCS, Soil Survey Division, World Soil Resources, 1998]

6.11.3 Factors of Soil Formation

6.11.3.1 Parent Material

Alfisols form on a wide variety of parent materials which includes glacial till (Borchardt et al., 1968; Smeck et al., 1968; Smith and Wilding, 1972; Ranney et al., 1975; Franzmeier et al., 1985), loess (Bouma et al., 1968; Grossman and Fehrenbacher, 1971), eolian sands (Davis, 1970; Allen et al., 1972; Miles and Franzmeier, 1981), residuum from limestone (Reynders, 1972), sandstone (Anderson et al., 1975; Stahnke et al., 1983; Chittleborough, 1989; Delgado et al., 1994), siltstones and shales (Gallez et al., 1975; Delgado et al., 1994), acid igneous and metamorphic rock (Goss and Allen, 1968; Ranney and Beatty, 1969; Yerima et al., 1989; Kooistra et al., 1990); mafic igneous and metamorphic rock (Gallez et al., 1975; Bhattacharyya et al., 1993; Juo and Wilding, 1996), lacustrine sediments (Borchardt et al., 1968; Ranney and Beatty, 1969), colluvium derived from acid shales and siltstones (Bailey and Avers, 1971), basalt (Weitkamp et al., 1996), coastal plain sediments (Vepraskas and Wilding, 1983; Juo and Wilding, 1996) and alluvium (Parsons et al., 1968; Ranney et al., 1975; Chittleborough et al., 1984; Klich et al., 1990; Delgado et al., 1994). The literature is replete with examples of Alfisols developing from two or more parent materials, for example, loess over till (Foss and Rust, 1968; Allan and Hole, 1968; Ransom et al., 1987), loess over basalt (Blank and Fosberg, 1991), loess over shale (Ranney et al., 1975), and colluvium over residuum (Weitkamp et al., 1996). Uniformity of parent material also continues to be a question in evaluating the influence of lithogenic inherited versus pedogenic and translocated clay in the argillic horizon (Ruhe, 1984a). In eolian sands, the argillic (zone of clay accumulation) horizon usually consists of a sequence of thin bands called lamellae that contain somewhat more clay than the soil material between the bands.

Lotspeich and Coover (1962) stated that "texture of the parent material controls the texture of the soil because texture is a nearly permanent characteristic of soil," particularly in arid and semi-arid regions. While this may be true in Alfisols for which the parent materials have undergone previous cycles of weathering and are rich in quartz and resistant minerals (Anderson et al., 1975), many examples are available where weathering gives rise to textures and a pedogenic mineral suite markedly different from the parent material (Gallez et al., 1975; Paeth et al., 1971).

6.11.3.2 Organisms

Most Alfisols formed primarily under hardwood forest vegetation (Soil Survey Staff, 1975), but some suborders and great groups support a significant component of grass vegetation. This is illustrated in Fig. 6.63 where soil classes with abbreviated vegetative descriptions from the literature are arranged in a temperature-moisture regime matrix. Aqualfs and Udalfs support forest vegetation although dominant species and composition of forests change in response to climatic conditions. Ustalfs and Xeralfs support moderate to dominant grassland vegetation, commonly described as savanna. The forest vegetation of moister Cryalfs gives way to a greater composition of grasses as Cryalfs occupy dryer climates. It is hazardous to assume present or presettlement vegetation (Fig. 6.63) represents the native vegetation of the soil as both evolve together with climate changes over time modifying vegetative communities (Wells, 1970). Even within broad vegetative zones, individual species may influence soil properties over relatively short distances. Gersper and Holowaychuk (1970a,b) showed precipitation stemflow down the trunk of American beech (*Fagus grandifolia*) trees thickened E (A2) horizons, increased low chroma mottling of B horizons, and decreased clay content, pH, exchangeable Ca, Mg and K, CEC, base saturation and free Fe oxides in the B horizons near the stem. Davis et al. (1995) believe available P that was correlated with microelevational changes on a cultivated

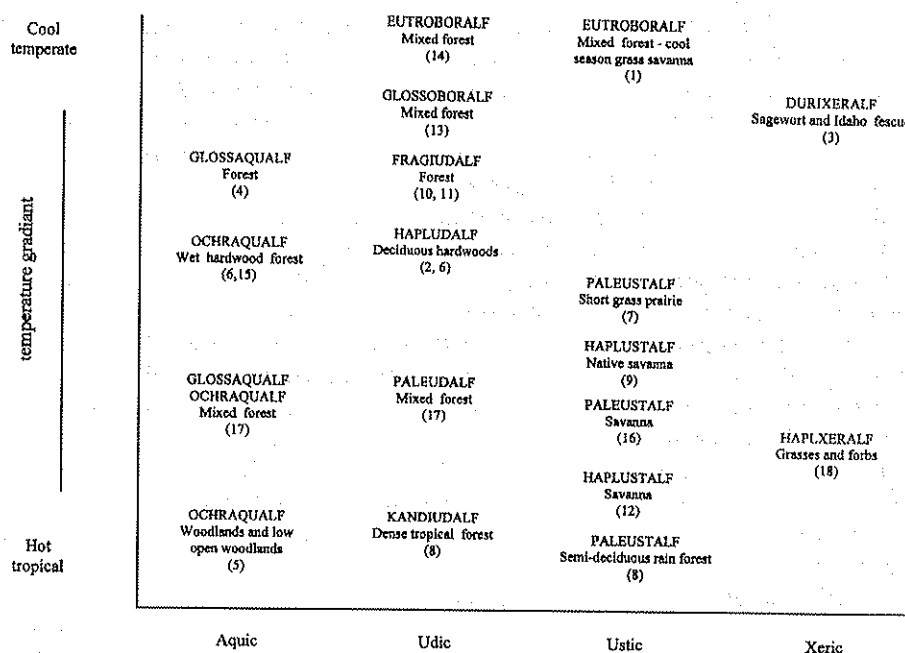


Fig. 6.63 Generalized native vegetation for selected Alfisols as a function of temperature and moisture regimes. [(1) Anderson et al., 1975; (2) Bailey and Avers, 1971; (3) Blank and Fosberg, 1991; (4) Bouma et al., 1969; (5) Coventry and Williams, 1984; (6) Cremeens and Mokma, 1986; (7) Davis, 1970; (8) Gallez et al., 1975; (9) Goss and Allen, 1968; (10) Lozet and Herbillon, 1971; (11) Miller et al., 1971; (12) Moberg and Esu, 1991; (13) Ranney and Beatty, 1969; (14) Schaetzl, 1996; (15) Smith and Wilding, 1972; (16) Stahnke et al., 1983; (17) Vepraskas and Wilding, 1983; (18) Weitkamp et al., 1996] Those soils listed as Ochraqualfs would be placed as Epigualfs or Endoqualfs, and those as Boralfs as Cryalfs in the current 1998 *Keys to Soil Taxonomy* [Soil Survey Staff, 1998]

Psammentic Paleustalf in Niger is related to location of bushes and trees in the savanna or to cultivation.

Fauna dwelling in Alfisols have also been credited with affecting soil formation. Nielson and Hole (1964) found earthworm population masses of 2218 kg ha⁻¹ under a deciduous forest and earthworm middens numbering 274,000 ha⁻¹. Essentially all annual leaf fall was moved to the middens by earthworms. Wiecek and Messenger (1972) described calcite spheroids in acid A horizons of Hapludalfs under forest vegetation that are produced in earthworm calciferous glands and were responsible for close-range variability in soil pH of A horizons. Hugie and Passey (1963) describe burrowing activity of western cicadas, crediting them with development of cylindrical structure as they burrowed through loamy soils. Their numbers were greatly reduced on coarse- and fine-textured materials. Ant, termite, and crustacean activity have also been described (Thorp, 1949). Crayfish activity in Ochraqualfs of the Texas coast has been so intense that up to 75% of the subsoil volume are comprised of their krotovinas prompting the call for a new great group (Vermaqualfs) to accommodate these soils highly modified by fauna (Vepraskas and Wilding, 1983). Arkley and Brown (1954) observed sufficient rodent activity to suggest that mima mounds in western states are the result of pocket gophers.

6.11.3.3 Climate

In the process of clay eluviation, percolating water has mobilized silicate clays in upper horizons, carried them downward in the pedon with final deposition in subsoil horizons. This occurs most

readily in climates in which precipitation exceeds evapotranspiration for at least several months each year. Consequently, Alfisols occur in semi-arid to humid climates. In temperate and cool climates, Alfisols tend to form a belt between the Mollisols of the grasslands and the Spodosols and Inceptisols of very humid climates (Soil Survey Staff, 1975). In warmer climates, they often lie between the Aridisols of arid regions and the Inceptisols, Ultisols, and Oxisols of warm, humid climates.

Within the Alfisol areas, several soil properties vary with climate. Early works of Jenny (1935) and Jenny and Leonard (1934) serve to broadly illustrate soil changes over long transects where gradients of temperature and precipitation exist. When crystalline rocks of similar composition and geomorphic age weather across a temperature gradient, soil clay content in the upper solum increases with mean annual temperature. Along the 11 °C annual isotherm from eastern Colorado to western Missouri, they found that with increased annual precipitation, depth to free carbonates, N and clay contents increased, and pH decreased. Partitioning climatic influence over long distances is difficult at best. Climate implies both temporal and spatial fluxes in a number of parameters including temperature, precipitation, radiation, humidity, and wind speed and direction. Locally, relief, slope, and aspect can provide a microclimate somewhat dissimilar to the broader regional climate. Cooper (1960) illustrated the role of slope aspect on microclimate in Michigan Udalfs where he found south-facing slopes had higher light intensities, maximum air temperature, evaporation rate, and soil temperature and lower soil moisture than north-facing slopes. The B horizons of soils on the southern aspect were redder and finer in texture, and solums were thinner than those of the northern exposure. Torrent and Nettleton (1979) proposed a textural index (based upon fine silt:total silt ratios) to assess chemical weathering. Using soils, primarily Udalfs and Aqualfs developed from loess along the Mississippi Valley from Minnesota to Mississippi, they related the textural index describing weathering of silts to clays with increasing mean annual temperature. Subsequently, Ruhe (1984a), utilizing a smaller data set of Hapludalfs along a similar Minnesota to Mississippi transect, interpreted the trends as sedimentation dependent. In a companion study on the same Hapludalfs, Ruhe (1984b) related the more intensive and deeper leaching of bases in the southern sector to both historic and paleo climate differences in annual and effective precipitation.

6.11.3.4 Relief

Alfisols occur mostly on hillslopes that are of moderate gradient and are flat or convex in cross-section. On steeper slopes, clay translocation is limited because the soil is very shallow or stony or because material moves downslope, resulting in Inceptisols or Entisols. In depressional areas, downward water movement might be limited, or soil organic matter (SOM) content of upper horizons is high enough for the soils to be Mollisols.

Relief is responsible for the distribution of Aqualfs in many settings where water moves to lower topographic positions either by overland or throughflow. In these lower positions, epiaquic (perched water table) or endoaquic conditions may occur. Cremeens and Mokma (1986) described soil hydrosequences in Michigan where well- and moderately well-drained Hapludalfs yielded to somewhat poorly drained Ochraqualfs and poorly drained Haplaquepts and Argiaquolls. Argillans (clay films) and parameters related to movement of clay from eluvial to illuvial zones were greatest in the better drained soils. Working with seasonally saturated Alfisols in a toposequence on the Texas Coastal Plain, Vepraskas and Wilding (1983) found soil pH generally increased as did base saturation and exchangeable Na levels from upper to lower topographic positions. Also, smectite was the predominant clay mineral in the upper sola of the soils in the lower positions while kaolinite was dominant in higher positions. Cremeens and Mokma (1986) also found smectite in the wettest sites of their study with its absence in the better drained soils. Goss and Allen (1968) found appreciable

accumulation of exchangeable Na downslope in a Natrustalf and evidence of accumulation of smectite due to relief.

6.11.3.5 Time

Alfisols in areas with moderate and cool temperatures are mostly on late Pleistocene surfaces as most of these surfaces were affected by glaciation. In warmer areas, they may be on older surfaces. Usually, there has not been enough clay translocation for the soils to be classified as Alfisols on surfaces younger than late Pleistocene. On old surfaces, leaching of bases commonly has progressed to the point where the soils are classified as Ultisols instead of Alfisols.

The length of time for development of soil and specific soil properties has long been of scientific interest (Jenny, 1941). In recent years, interest in soil age has been increasing as evidenced by the growing volume of literature (Bilzi and Ciolkosz, 1977; Bockheim, 1980; Little and Ward, 1981; Meixner and Singer, 1981; Harden, 1982; Muhs, 1982; Chittleborough et al., 1984; Catt 1991; Markewich and Pavich, 1991; Delgado et al., 1994). Part of this renewed interest is due to application of soil interpretation to subjects of current interest such as climate change, C sequestration rates, landscape stability, and archaeology. Improved and new methodologies to directly determine soil age such as radiocarbon assay, thermoluminescence, amino acid analysis, palaeomagnetic measurement and U-Th disequilibrium series have augmented dendrochronology and recorded historic events (mudflows, volcanic eruptions, etc.) and indirect methods of correlation of stratigraphic units. As soil (and landscape) age increases, so does the concern for significant changes in other factors of state, particularly climate and its effect on vegetation. The time required for the development of Alfisols is incumbent upon the formation of argillic and kandic horizons. Bilzi and Ciolkosz (1977) believe that 2,000 to 12,000 years are required to develop fragipan-like features and argillic horizons in alluvium in Pennsylvania under udic soil moisture and mesic soil temperature regimes. In northeast Iowa on earthen mounds of varying age built by Indians, Parsons et al. (1962) found that soil genesis had taken place with A horizon properties remaining relatively constant with increasing age beyond 1000 yr, E horizon platy structure developing within 1,000 yr (maximum expression required > 2500 yr) and the blocky structure plus clay translocation sufficient to meet argillic horizon definitions in the B horizon occurring between 1,000 and 2,500 yr BP. In comparison, the nearby Fayette soil (Typic Hapludalf) had developed over a 14,000 yr period. In preconditioned material used by Indians to construct mounds and ridges in Louisiana, argillic horizons in Ultisols and Alfisols had developed in 5,000 to 5,400 yr (Saunders et al., 1997).

In the Midwest, most Alfisols are found on surfaces affected by late Wisconsinian glaciation (12,000–25,000 yr BP). In the Upper Midwest, many Alfisols formed in late Wisconsinian till deposited by glacier ice, and in outwash and lacustrine deposits from the glacier when it melted. Also, much of the Midwest and the lower Mississippi Valley was blanketed with late Wisconsinian age loess that was calcareous when it was deposited. Soils formed in it are relatively high in basic cations. This loess also covered surfaces on which soils had formed previously and had been eroded. These buried soils (paleosols) were enriched with basic cations (especially Ca and Mg) that were leached from the loess and translocated downward.

Alfisols of ustic and xeric soil moisture regimes and developed from alluvium have been included in a number of chronosequence studies (Table 6.59). The annual precipitation ranged from 170 to 420 mm. The youngest Alfisol (Ultic Haplustalf) was 3,740 yr BP and the oldest (Paleustalf) was 763,000 yr BP, while the youngest Paleustalf was 9,220 yr BP. Much older Haploxeralfs were noted in the California studies, but in the xeric moisture regime sufficient development of the argillic horizon to permit classification as a Palaxeralf had not occurred.

Table 6.59 Ages of Ustalfs and Xeralfs included in chronosequence studies. All soils developed from alluvium. Dating methods in the studies include amino acid, U-series, uplift rates, and stratigraphic correlations

Soil Class	Age, yrs BP	MAP (mm)	Location	Source
Typic Haploxeralf	250,000	310	California	Meixner and Singer (1981)
Typic Haploxeralf	40,000	410	California	Harden (1982)
Typic Haploxeralf	130,000			
Typic Haploxeralf	600,000			
Typic Natrixeralf	60-107,000	170	California	Muhs (1982)
Typic Natrixeralf	120-134,000			
Typic Natrixeralf	375-460,000			
Ultic Haplustalf	3,740	420	New South Wales, Australia	Chittleborough et al. (1984)
Rhodic Paleustalf	9,220			
Ultic Paleustalf	26,700			
Ultic Paleustalf	27,040			
Ultic Paleustalf	29,000			
Paleustalf	42,500	--	Victoria, Australia	Little and Ward (1981)
Paleustalf	210,000			
Paleustalf	227,000			
Paleustalf	763,000			

6.11.4 Morphology

The morphology of Alfisols ranges greatly across the broad spectrum of conditions under which they occur. In all Alfisols, however, there is an increase in clay content from the A and E horizons near the surface to the B horizons below. In most Alfisols, this clay increase meets the concept of an argillic horizon where the illuvial Bt (t from *ton* Gr, clay) horizon shows evidence of oriented clays. In older

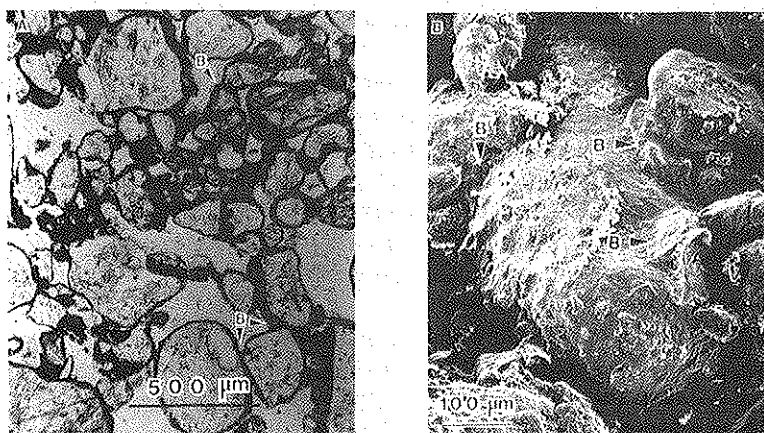


Fig. 6.64 The Bt2 horizon (51-72 cm depth) of a sandy clay, siliceous, subactive, isohyperthermic Psammentic Paleustalf from Niger. A. Plane polarized light micrograph of ferriargillans around quartz skeleton grains and clay bridging (B) of sand grains; B. Scanning electron micrograph of clay bridging (B) between sand grains [West et al. (1984)]

soils dominated by low activity clays (low CEC of the clay fraction), evidence of oriented clay may be absent but clay increase is sufficient for a kandic horizon.

Where some of the clay in the Bt horizon is oriented, that is, stacked like a pile of papers or playing cards, the oriented clay may be on ped surfaces where it forms clay films (thin layers of oriented clay) that often cover and may be a different color from sand grains in the ped (see Section E, Chapter 6.12 for electron micrograph of a clay film). In sandy soils, the oriented clay forms bridges between adjacent sand grains (Fig. 6.64) and may line root channels. Often, this clay mentioned above actually contains a fair amount of silt. In many Alfisols, the transitional EB or BE horizons have silt coatings instead of clay on the outside of peds. In these, and some other Alfisols, especially those formed in calcareous parent materials, the best developed clay films are in the lower Bt horizon. Apparently, clay is stripped from the ped coatings in the upper B or BE horizon and is deposited on ped surfaces deeper in the profile, illustrating that clay translocation is a dynamic process, and that the argillic horizon is moving down in the profile.

Before a soil can be classified in *Soil Taxonomy*, germaine diagnostic horizons and soil properties (defined in Section E, Chapter 4.5) must be identified. Several are essential or common in Alfisols and their occurrence in the various suborders of Alfisols is shown in Table 6.61.

6.11.5 Suborders: Their Morphology, Distribution, and Use

The Alfisol Order is subdivided into five suborders, one based on temperature and four based on soil moisture regime. Cryalfs are the Alfisols in cold climates. Of the four suborders based on soil moisture regime, three (Udalfs, Ustalfs, Xeralfs) follow broad geographical climatic belts. The distribution of Aqualfs depends on local landscape and parent material factors. A complete listing of the Great Groups and Subgroups in each suborder appears in Table 6.60.

6.11.5.1 Aqualfs

These are the wet Alfisols in which the watertable is at or near the surface (50 cm) for at least several weeks each year, but may be very deep during the dry season. The watertable may be continuous to the aquifer, or it may be held up by a layer with low permeability in or just beneath the soil surface. Their location in the landscape is controlled mainly by local conditions of parent material and relief. They occur as small areas in larger areas of Cryalfs, Udalfs, Ustalfs, Xeralfs or other soil suborders, but they are more common in zones of Udalfs than Ustalfs. They may have several kinds of diagnostic horizons (Table 6.61). Nearly all Aqualfs had forest vegetation at some time in the past (Soil Survey Staff, 1975). Most are cultivated after they were drained by underground tiles or ditches. Rice is a common crop in warmer regions.

Aqualfs display quantifiable redoximorphic features (gray colors) immediately below a depth of 25 cm from the mineral soil surface. In the absence of quantifiable redoximorphic features, a positive reaction to active Fe^{2+} (α , α -dipyridyl) can be used. In this process (gleying), soil horizons are saturated with water, microbes deplete available O_2 , causing Fe(III) in brownish and reddish Fe oxide minerals to be reduced. The resulting Fe^{2+} is released into the soil solution, much of which is leached from the soil. The result is that the horizon, or much of it, takes on the gray color of the silicate minerals. Gleyed horizons are designated with a g as Btg.

Gleying in Aqualfs may be noted in the upper 25 cm of the soil. Gleyed soils are generally found on lower topographic positions receiving additional water, or on broad, level interfluvies where surface drainage may be very slow. The A horizon, commonly overlain by O horizon material in varying stages of decomposition, is usually either too thin, or too low in organic C to meet requirements of a mollic epipedon. A few may qualify as an umbric or mollic epipeon, but have insufficient bases in the subsoil

Table 6.60 Listing of suborders, great groups and subgroups in the Alfisols order [Soil Survey Staff, 1998]

Suborder	Great Group	Subgroup
Aqualfs	Cryaqualfs	Typic
	Plinthaqualfs	Typic
	Duraqualfs	Typic
	Natraqualfs	Vertic, Vermic, Albic Glossic, Albic, Glossic, Mollic, Typic
	Fragiaqualfs	Vermic, Aeric, Plinthic, Humic, Typic
	Kandiaqualfs	Arenic, Grossarenic, Plinthic, Aeric Umbric, Aeric, Umbric, Typic
	Vermaqualfs	Natric, Typic
	Albaqualfs	Arenic, Aeric Vertic, Chromic Vertic, Vertic, Udollic, Aeric, Aquandic, Mollic, Umbric, Typic
	Glossaqualfs	Histic, Arenic, Aeric Fragic, Fragic, Aeric, Mollic, Typic
	Epiaqualfs	Aeric Chromic Vertic, Aeric Vertic, Chromic Vertic, Vertic, Aquandic, Aeric Fragic, Fragic, Arenic, Grossarenic, Aeric Umbric, Udollic, Aeric, Mollic, Umbric, Typic
	Endoaqualfs	Aquandic, Chromic Vertic, Vertic, Aeric Fragic, Fragic, Arenic, Grossarenic, Udollic, Aeric Umbric, Aeric, Mollic, Umbric, Typic
Cryalfs	Palecryalfs	Andic, Vitrandic, Aquic, Oxyaquic, Xeric, Ustic, Mollic, Umbric, Typic
	Glossocryalfs	Lithic, Vertic, Andic, Vitrandic, Aquic, Oxyaquic, Fragic, Xerollic, Umbric Xeric, Ustollic, Xeric, Ustic, Mollic, Umbric, Eutric, Typic
	Haplocryalfs	Lithic, Vertic, Andic, Vitrandic, Aquic, Oxyaquic, Lamellic, Psammentic, Inceptic, Xerollic, Umbric Xeric, Ustollic, Xeric, Ustic, Mollic, Umbric, Eutric, Typic
Ustalfs	Durustalfs	Typic
	Plinthustalfs	Typic
	Natrustalfs	Salidic, Leptic Torrtic, Torrtic, Aquertic, Aridic Leptic, Vertic, Aquic Arenic, Aquic, Arenic, Petrocalcic, Leptic, Haplargidic, Aridic, Mollic, Typic
	Kandiustalfs	Grossarenic, Aquic Arenic, Plinthic, Aquic, Arenic Aridic, Arenic, Aridic, Udic, Rhodic, Typic
	Kanhaplustalfs	Lithic, Aquic, Aridic, Udic, Rhodic, Typic
	Paleustalfs	Aquertic, Oxyaquic Vertic, Udertic, Vertic, Aquic Arenic, Aquic, Oxyaquic, Lamellic, Psammentic, Arenic Aridic, Grossarenic, Arenic, Plinthic, Petrocalcic, Calcic, Aridic, Kandic, Rhodic, Ultic, Udic, Typic
	Rhodustalfs	Lithic, Kanhaplic, Udic, Typic
	Haplustalfs	Lithic, Aquertic, Oxyaquic Vertic, Torrtic, Udertic, Vertic, Aquic Arenic, Aquatic, Aquic, Oxyaquic, Vitrandic, Lamellic, Psammentic, Arenic Aridic, Arenic, Calcic, Aridic, Kanhaplic, Inceptic, Calcic Udic, Ultic, Calcic, Udic, Typic
Xeralfs	Durixeralfs	Natric, Vertic, Aquic, Abruptic Haplic, Abruptic, Haplic, Typic
	Natrixeralfs	Vertic, Aquic, Typic
	Fragixeralfs	Andic, Vitrandic, Mollic, Aquic, Inceptic, Typic
	Plinthoxeralfs	Typic
	Rhodoxeralfs	Lithic, Vertic, Petrocalcic, Calcic, Inceptic, Typic
	Palexeralfs	Vertic, Aquandic, Andic, Vitrandic, Fragiaquic, Aquic, Petrocalcic, Lamellic, Psammentic, Arenic, Natric, Fragic, Calcic, Plinthic, Ultic, Haplic, Mollic, Typic
	Haploxeralfs	Lithic Mollic, Lithic Ruptic-Inceptic, Lithic, Vertic, Aquandic, Andic, Vitrandic, Fragiaquic, Aquatic, Aquic, Natric, Fragic, Lamellic, Psammentic, Plinthic, Calcic, Inceptic, Ultic, Mollic, Typic
Udalfs	Natrudalfs	Vertic, Glossaquic, Aquic, Typic
	Ferrudalfs	Aquic, Typic
	Fraglossudalfs	Andic, Vitrandic, Aquic, Oxyaquic, Typic
	Fragiudalfs	Andic, Vitrandic, Aquic, Oxyaquic, Typic
	Kandiudalfs	Plinthaquic, Aquic, Oxyaquic, Arenic Plinthic, Grossarenic Plinthic, Arenic, Grossarenic, Plinthic, Rhodic, Mollic, Typic
	Kanhapudalfs	Lithic, Aquic, Oxyaquic, Rhodic, Typic
	Paleudalfs	Vertic, Andic, Vitrandic, Fragiaquic, Plinthaquic, Glossaquic, Albaquic, Aquic, Anthraquic, Oxyaquic, Fragic, Arenic Plinthic, Grossarenic Plinthic, Lamellic, Psammentic, Arenic, Grossarenic, Plinthic, Glossic, Rhodic, Mollic, Typic
	Rhodudalfs	Typic
	Glossudalfs	Aquertic, Oxyaquic Vertic, Vertic, Aquandic, Andic, Vitrandic, Fragiaquic, Aquic, Oxyaquic, Fragic, Arenic, Haplic, Typic
	Hapludalfs	Lithic, Aquertic Chromic, Aquertic, Oxyaquic Vertic, Chromic Vertic, Vertic, Andic, Vitrandic, Fragiaquic, Fragic Oxyaquic, Aquic Arenic, Albaquic, Glossaquic, Aquatic, Aquollic, Aquic, Anthraquic, Oxyaquic, Fragic, Lamellic, Psammentic, Arenic, Glossic, Inceptic, Ultic, Mollic, Typic

Table 6.61 Occurrence of soil horizons or properties in soils of the suborders of Alfisols

Suborder	Soil horizon or property									
	Argillic horizon	Natric horizon	Kandic horizon	Glossic horizon	Fragipan	Duripan	Plinthite	Very red argillic	Very deep argillic	Other
Aqualfs	X	X	X	X	X	X	X			Worm activity, very cold *
Cryalfs	X			X					X	
Udalfs	X	X	X	X	X			X	X	Fe nodules
Ustalfs	X	X	X			X	X	X	X	
Xeralfs	X	X			X	X	X	X	X	

* Also: abrupt clay increase at upper boundary of the argillic horizon, and perched water table

to be classified as Mollisols. Most Aqualfs have well-developed E horizons although the underlying Bt and Btg horizons often show only minimally expressed clay films. Boundaries between E and Bt and Btg horizons often show evidence of mixing by fauna (i.e., crayfish) or tongues and stringers of E horizon material between structural units of Bt and Btg material (glossic material).

6.11.5.2 Cryalfs

Cryalfs are restricted to frigid and cryic temperature regimes, and generally occur in more mountainous and sloping landscapes that are generally well drained. They formed in North America, eastern Europe, and Asia above 49° N latitude and in some high mountains of lower latitude where they occur at lower altitudes than Spodosols or Inceptisols. Most have been under coniferous forest because of the cool, short growing season. Commonly, parent materials are residuum or colluvium from local bedrock. Most are not cultivated, and thus, have thin to moderately thick (2–10 cm) O horizons in varying stages of organic matter decomposition. Many lack A horizons, but when present, they are thin (5 cm) with granular or subangular blocky structure. Most have E horizons of platy or subangular blocky structure that are relatively thick (15 to 30 cm). The Bt horizons range from thin to thick (15 to 135 cm), display a wide range of clay film features, and commonly have angular or subangular blocky structure. Many have lithic or paralithic contacts, and most contain significant quantities of coarse fragments throughout the solum.

6.11.5.3 Udalfs

These are the freely drained Alfisols of humid regions, extensive in the United States and Europe. They formed under forest vegetation at some time during their development. The forests were mainly deciduous, but in colder regions, may have been mixed coniferous and deciduous. Many Udalfs have been cleared and intensively farmed; while in some that have been severely eroded, the Bt horizon is immediately below or may be incorporated into the Ap horizon (Soil Survey Staff, 1975). Many have fragipans, and some have natric and other diagnostic horizons (Table 6.61). Corn and soybeans are common crops on Udalfs.

In well-drained, temperate humid environs under forest, uncultivated Udalfs have thin (2–5 cm) O horizons with organic matter in various stages of decomposition. Most also have a thin (10–15 cm) A horizon commonly of granular structure darkened by organic matter and underlain by a thin (10–15 cm), light colored E horizon with platy structure, a zone of maximum eluviation of silicate clays.

Because many Alfisols are cultivated, the A and E horizons have been mixed by plowing producing a light color. The intense eluviation in E horizons results in a concentration of sand and silt, and in extreme cases, all or most coloring compounds have been removed leaving skeletal grains whose color dominates the horizon. Often there is a transitional EB or BE horizon between the E and Bt horizons. As compared to the thinner A and E horizons, the Bt horizon is thick, possesses subangular or angular blocky structure, and contains significantly more silicate clay. At least a portion of the silicate clay occurs as clay films on structural or pores surfaces. In the Midwest where most of the parent material is calcareous and associated with the last glaciation, complete leaching of carbonates from the solum has occurred while C horizons remain calcareous.

6.11.5.4 Ustalfs

These are the Alfisols of subhumid to semi-arid regions. Moisture moves through these soils to deeper layers only in occasional years (Soil Survey Staff, 1975). If there are carbonates in the parent material or in dust, the soil may have a carbonate enriched horizon (calcic) below the argillic. Original vegetation may have been xerophytic trees, savanna, or grassland. Dryland crops for Ustalfs include drought tolerant selections such as sorghum, wheat, cotton, and millet. Ustalfs which are common in the southern Great Plains, do not have fragipans, but may have several other kinds of diagnostic horizons (Table 6.61).

In the United States, Ustalfs tend to be on older landforms than Udalfs, and have thicker sola; many are in the Paleustalf great group because of either an abrupt textural change between E and Bt horizons or a thick Bt horizon enriched in clay throughout. There is less tendency for E horizons to display platy structure; while with less rainfall, E horizons are often absent. The Bt horizons often have prismatic parting to angular or subangular blocky structure; at the drier end of Ustalfs, sufficient secondary carbonates commonly accumulate to form calcic horizons in the lower portion of the argillic horizon or just below it.

6.11.5.5 Xeralfs

These occur in regions that have a Mediterranean climate, where precipitation occurs during the cool season and the summers are hot and dry. In some winters water moves through the entire soil profile to deeper layers. Most Xeralfs border the Mediterranean Sea or lie to the east of an ocean, such as those in the western United States and Australia. Where there is no irrigation, grains are common crops but with irrigation, a variety of crops can be grown. Grapes and olives are grown in the warmer areas (Soil Survey Staff, 1975). Occurrence of E horizons is uncommon, and A horizons with weak subangular blocky structure to massive condition tend to range from 15 to 25 cm thick. Most argillic horizons are relatively thin (15 to 75 cm) with significant quantities of coarse fragments throughout the pedon.

6.11.6 Soil Formation Processes

6.11.6.1 Clay Translocation

Alfisols and Ultisols form mainly by eluviation of silicate clay from A and E to Bt horizons. In Alfisols, however, the base saturation of deep subsoil layers is greater than it is in Ultisols. Alfisols are similar to the Gray Brown Podzolic soils of previous soil classification systems. The name Alfisol was derived from the term Pedalfers, a coined word used in previous classifications derived from *ped*, as in pedology (the science of natural soils), *Al*, aluminum, and *Fe*, iron. The implication is that in Pedalfers, Al and Fe accumulate in the subsoil, in contrast to Pedocals, in which Ca accumulates.

The concept of Gray Brown Podzolic soils originated at a time when soil colloids (clays) were thought to be mainly amorphous. Then, soil scientists determined total chemical composition of the

colloidal fraction, and interpreted the results by examining $\text{SiO}_2/\text{Al}_2\text{O}_3$ and $\text{SiO}_2/(\text{Fe}_2\text{O}_3 + \text{Al}_2\text{O}_3)$ ratios of the various horizons. These ratios showed that both Podzols (Spodosols) and Gray Brown Podzolic soils had B horizons enriched in Fe and Al, but that Podzol B horizons were also enriched in organic C. This led Marbut (1935) to conclude that the differences in the processes leading to the two kinds of soil were of degree, not kind, and that Gray Brown Podzolic soils were less developed than Podzols. The word Podzolic is a descriptor previously used for many Alfisols.

Hendricks and Fry (1930) reported two key discoveries that laid the groundwork for the current understanding of the processes by which Alfisols form. First, they showed that soil colloids are mainly crystalline silicate minerals, and that when clay suspensions are allowed to dry, the plate-shaped minerals lay flat to form masses that have a regular crystallographic, or at least optical, arrangement and are large enough to study with a polarizing microscope. This discovery led to an understanding that in Alfisols the material translocated is mainly silicate clay minerals (with Fe oxides on their surfaces), but in Spodosols, it is mainly amorphous complexes of organic matter with Al and usually Fe. Second, they provided the basis for interpreting what are now called clay films.

These ideas were gradually applied to studies of soil genesis. Jenny and Smith (1935) produced early stages of argillic horizon formation in columns of pure quartz sand by first coating the sand grains with electropositive iron hydroxide and then passing dispersed clay through the column. Brown and Thorp (1942) noted that the B horizons in the Miami soil (Hapludalf) from Indiana and similar soils contained much more clay than did the A and E horizons. They concluded, however, that the clay increase in the B horizons was only partly due to illuviation. Although the types of clay were identified by X-ray diffraction and thermal methods, soil formation processes were still discussed mainly in terms of $\text{SiO}_2/(\text{Fe}_2\text{O}_3 + \text{Al}_2\text{O}_3)$ ratios. Frei and Cline (1949) concluded that the apparent loss of clay from A and E horizons and the apparent gain in clay in B horizons of a New York soil could be accounted for in at least three possible ways: (1) clay migrates downward as a sol in percolating water; (2) clay is synthesized in the B horizon from soluble weathering products that were released there or moved down in the profile; or (3) the apparent increase in clay was really due to loss of other materials, such as CaCO_3 . They showed, too, that there was a strong concentration of clays with a high degree of optical continuity on peds in the B horizon of a Gray Brown Podzolic soil, but in the upper B horizon, these coatings were highly degraded. According to their interpretation, clay that was mobilized in A and E horizons and stripped from ped surfaces in the upper B horizon, moved down the profile in suspension, and was deposited on ped surfaces as films of oriented clay particles. Thorp et al. (1959), in another study of the Miami soil of Indiana, showed that the small particle size of clays, their association with organic compounds, and shrink/swell cycles in upper horizons all enhanced their mobility.

In summary, the movement of clay from A and E horizons to Bt horizons can be viewed as the net result of three subprocesses: dispersion, transport, and deposition (Jenny, 1980). Dispersion, the release of individual clay particles from aggregates, is favored by the replacement of exchangeable Ca^{2+} with Na^+ , by adsorption of humus molecules on clay particles, and by mechanical disruption of aggregates which may be caused by wetting and drying, freezing and thawing, or by mechanical disruption by faunal activity. Small particles move more readily than large particles, so that illuvial clay, as in clay films, contains more fine than immobile clay, such as that in ped interiors. Little clay is dispersed in calcareous soil materials because Ca^{2+} ions on exchange sites favor clay flocculation. Clay that is mobilized in A and E horizons moves as a percolating sol or as slurry creep along ped surfaces. Deposition is favored when Na^+ is replaced by Ca^{2+} , resulting in clay flocculation by an increase in pH, or by reaction of the silicate clay with oxides such as Fe oxide. When the water of a slurry enters into a ped, the clay it contains is deposited on the ped surface, like passing the slurry

through filter paper. In some cases, the slurry moves down to the wetting front in the subsoil, and clay is deposited as water evaporates or is taken up by plants. Texture discontinuities may also arrest the percolation of suspensions.

Because many Alfisols formed in calcareous materials, CaCO_3 must be dissolved and the weathering products leached from the soil before much clay can move. Usually H^+ replaces a portion of the Ca^{2+} in these Alfisols, causing leached horizons to become acidic. The shallowest depth at which carbonate minerals are found, as detected by dilute HCl , serves as an index of the intensity and length of soil formation.

It should be stressed that, although illuviation of clays is diagnostic for the argillic horizon, much or even most clay in the Alfisol Bt horizons originates from other sources to include inheritance from parent material, clay neoformation, *in situ* weathering of lithogenic minerals, and dissolution of silt and sand sized carbonates (Smeck and Wilding, 1980).

6.11.6.2 Natric Horizons

Natric horizons are argillic horizons high in exchangeable Na^+ . A natric horizon in an Alfisol seems to be an oxymoron; if rainfall has been high enough to move sufficient clay for the horizon to qualify as an argillic horizon, Na^+ should have long since been leached from the soil. However, natric horizons do exist in Alfisols indicating that there are circumstances in which they form. Some might be caused by a change in climate where the argillic horizon was formed during a previous wetter period, but rainfall is now less, and the leaching of Na^+ is reduced. Other soils with natric horizons occur in small areas where Na^+ has accumulated. Although the original source of much of the Na is feldspar minerals, Na^+ moves about in the landscape once it is released into solution. Often, soils with natric horizons are low in the landscape because solutions containing Na^+ move downslope either on the soil surface or through subsurface horizons, thus concentrating Na^+ in the low areas. In some soils, Na^+ is retained in the soil because the hydraulic conductivity of B horizons is very low, and leaching of Na^+ is retarded. This seems to be a self-perpetuating process; Na^+ accumulates because permeability is low, while Na^+ promotes dispersion, further slowing permeability. Furthermore, some soils with natric horizons were formerly saline soils. When soluble salts are leached from a soil, Na^+ may remain behind on the exchange sites.

6.11.6.3 Fragipan Formation

Fragipans occur in many Alfisols of the humid temperate regions, especially those of silty and loamy textures. These subsurface horizons restrict the entry of water and roots into the soil matrix (prisms) and often underlie argillic horizons. Commonly, the fragipan has a relatively low content of organic matter and high bulk density relative to horizons above. Most fragipans consist of prisms more than 10 cm across with light colored coatings on the prism faces. These light colored coatings form a polygonal pattern as shown in Fig 6.65. Material within prisms has a brittle fracture, a tendency for a clod to rupture suddenly rather than undergo slow deformation when pressure is applied. Fragments of the fragipan slake (fall apart) when submersed in water.

There are many theories on how fragipans form, summarized in two reviews (Smalley and Davin, 1982; Smeck and Ciolksz, 1989). One reason for the diversity in ideas is that there may be several kinds of horizons included under the name fragipan, each with its own set of processes of formation. Most soil scientists believe that they have formed, or are forming, under present day conditions. The main evidence that fragipans are genetic horizons related to the current land surface is that they are roughly parallel to the soil surface, and the upper boundary has a relatively narrow range of about 50 to 100 cm depth. Some scientists, however, believe that they formed much earlier. The following are some of the various theories of fragipan formation.

Paleosols

According to this theory, fragipans formed in earlier times and were buried by younger deposits in which fragipans are not forming. In western Tennessee, Buntley et al. (1977) observed that fragipans are mainly in older loess deposits that were buried by younger Peorian loess in which no fragipans form.

Rapid Initial Development

Some workers believe that fragipans formed quickly in their life history. Fragipan formation has been related to periglacial conditions (Payton, 1992) and to permafrost (FitzPatrick, 1956; van Vliet and Langohr, 1981). The polygonal pattern formed by coatings on the coarse prismatic structure is similar to that found in periglacial regions. Others present evidence that the structure of loess collapses when it first dries after deposition (Bryant, 1989), and that this denser material becomes the fragipan.

Much glacial till is very dense because it was compacted by the mass of glacier ice. Subsoil horizons that form in dense till may inherit some of their strength from the parent material (Lindbo and Veneman, 1989), but they might also develop fragipan properties through soil-forming processes. In

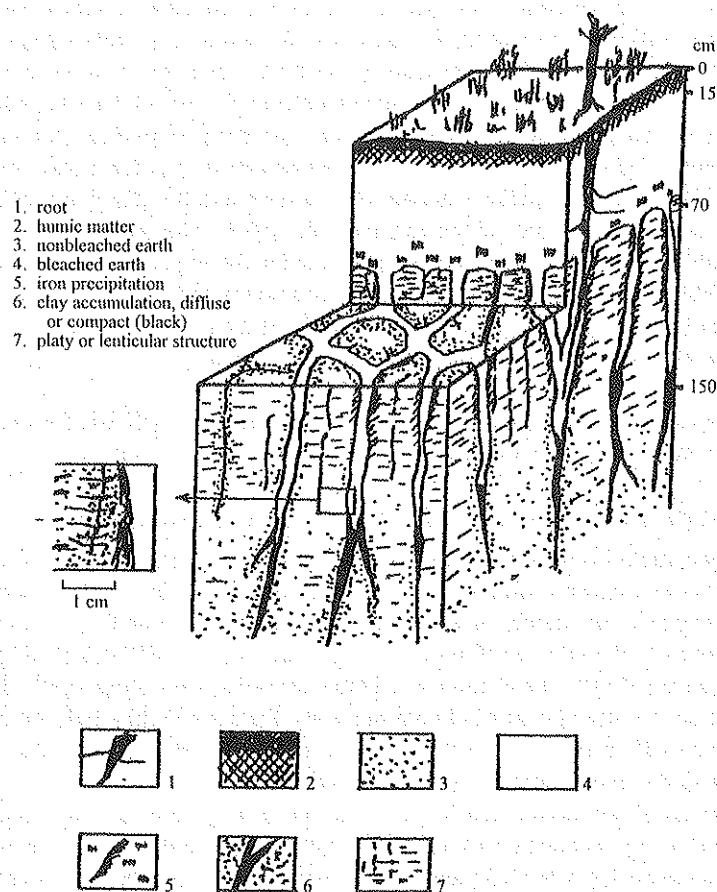


Fig. 6.65 Schematic of a fragipan in an imperfectly drained silty soil [Reprinted from van Vliet and Langohr, 1981. Correlation between fragipans and permafrost with special reference to silty deposits in Belgium and northern France. *Catena* 8:137-154. Copyright with permission of Elsevier Science]

these soils, which are common from eastern Ohio to New England, it is difficult to decide whether the dense material should be considered a fragipan or a slightly modified parent material. The great strength of these horizons is due in part to their high density, but it may also be due to cementation by Si or carbonates (McBurnett and Franzmeier, 1997).

Chemical Cementation

Fragipans could also be cemented by Fe-, Al-, or Si-rich materials. Iron-rich materials cause the cementation of ortstein layers, and silica bonds particles together to form duripans. Although duripans form mainly in aridic and xeric climates, in more humid climates, they grade into fragipans (Soil Survey Staff, 1998). The idea that fragipans are cemented by silica was first proposed in the 1930s and 1940s but was discredited and ignored for about 40 years before revitalized in the late 1970s. The proposal requires certain interactions of parent materials and climate (Franzmeier et al., 1989). In addition to undergoing periodic leaching, soils must also become seasonally dry, which commonly occurs in areas with a udic moisture regime under hardwood forest. Also, downward percolation of soil solutions must be arrested, rather than passing quickly through subsoil layers. This could be caused by the depth of penetration of the wetting front, by discontinuities in parent materials (Smeck et al., 1989), or by slowly permeable, deeper subsoil layers. Silicate minerals weather in upper horizons, and the silica released moves down, usually in the winter and early spring, to the subsoil where it remains for a time. When the trees leaf out, water taken up by roots concentrates silica in solution eventually causing it to precipitate. Compared with grasses, hardwood trees take up relatively little Si, and thus have the potential to concentrate Si in the soil. The precipitated silica bridges between clay particles, or more likely between Fe oxide minerals adsorbed on silicate clay surfaces, causing weak cementation. The greater Si uptake in grasses may be the reason that fragipans do not form under prairie vegetation in humid climates (Franzmeier et al., 1989). The precipitation of Si must be somewhat irreversible, or else any cementation that formed in the summer would be destroyed in the winter, and no net fragipan formation would take place. Fragipans do not form readily on steep slopes, in deep loess, or in loess over more permeable materials, such as outwash. In these soils, the soil solution moves laterally downslope or through the profile, preventing the accumulation of Si-rich soil solutions.

Most of these theories of fragipan formation individually have deficiencies. Perhaps various combinations of them can explain how a particular soil forms.

6.11.7 Selected Physical, Chemical and Mineralogical Properties

An increase in clay content between the E and/or A horizon(s) and underlying Bt horizons occurs in all Alfisols; however, this textural contrast ranges from little to great. This is illustrated (Fig. 6.66) by the change in clay content with depth for a Psammentic Haploxeralf from California (Torrent et al., 1980), a Typic Hapludalf from Illinois (Grossman and Fehrenbacher, 1971) and a Udertic Paleustalf from Texas (Klich et al., 1990). The Psammentic Haploxeralf has a minimal argillic horizon, with a clay increase of about 4% above that of eluvial horizons. By contrast, the Paleustalf shows a clay increase of more than 40% between the surface horizon and the finest textured portion of the argillic horizon. The Hapludalf is intermediate in textural contrast.

Alfisols with Bt horizons dominated by fine clay (< 0.0002 mm, usually smectitic) display significant volume change between wet and dry states. This volume change is reported as the coefficient of linear extensibility (COLE) and can be determined from bulk density values associated with moist and dry, natural clods (Grossman et al., 1968). The Paleustalf in Fig. 6.66 has COLE values > 0.07 (7% volume change between moist and dry states) for horizons containing 30% or more clay.

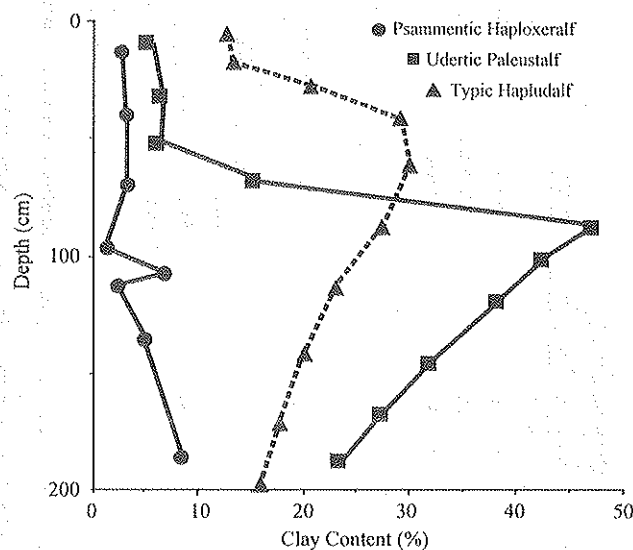


Fig. 6.66 Clay distribution in three Alfisols of contrasting degrees of argillic horizon development [Sources are Torrent et al., 1980; Grossman and Fehrenbacher, 1971; and Klich et al., 1990 for the Psammentic Haploxeralf, Typic Hapludalf and Udertic Paleustalf, respectively]

Because such soils are intergrades to Vertisols and possess similar features such as high subsoil clay content and shrink/swell potential, and slickensides, they present similar problems for use as construction materials (i.e., unstable subgrades for roads and houses).

Texture and clay content affect water-holding and water transmission properties of soils since water-related properties are dependent on void size and continuity. When the size of voids change greatly over a short distance, water movement is restricted. Alfisols with lamella type argillic horizons may exhibit slower drainage and better soil-plant relations due to greater available water retention than similar textured soils without lamella. Under more extreme conditions, strongly contrasting textures of eluvial and argillic horizons may result in water perched above the argillic horizons for several weeks as in Natraqualfs and Albaqualfs. In Alfisols with compact, clayey argillic horizons, root growth may be limited primarily to structural planes due to the high bulk density of ped interiors. Jones (1983) showed that root exclusion due to soil bulk density is a function of clay content.

Although organic C is greatest in the surface horizons, Alfisols tend to have greater CEC in the argillic horizon where clay and smectite contents are greatest. By definition, Alfisols must have a base saturation of 35% or more at the shallower of the following depths: (1) 1.25 cm below the top of the argillic horizon, (2) 1.8 m below the soil surface, (3) 75 cm below the top of a fragipan, or (4) immediately above a densic, petroferic, lithic or paralithic contact (just above continuous hard or soft bedrock). Most Alfisols have base saturation percentages well above 35%; those that approach low base status are recognized as intergrades to Ultisols by classification in Ultic subgroups. Generally, base saturation (and pH) decrease initially with depth in Alfisols, often reaching a minimum in the E horizons or upper portions of the argillic horizon, then increase again with depth. Alfisols under deciduous forest vegetation have relatively high base saturations in the surface horizon due to cycling of basic cations from depth to the surface in leaf fall. The relationship between soil pH and base saturation (Beery and Wilding, 1971; Blosser and Jenny, 1971; Ranney et al., 1974) is often used in

Table 6.62 Mineralogy of the argillic horizon of selected Alfisols

Classification	Location	Parent material	Minerals Reported																Ze	Source
			Clay fraction %(< 0.002 mm)												Sand and/or silt-fraction **					
			Ch	Ve	In	Sm	Mi	Ka	Qz	GH	Ps	Kf	Pl	Mi	Qz	Vg	AP			
																			%	
Glossoboralfs	Wisconsin	Lacustrine sediments	1	2		2	2	1				12	18	8	55			Raney and Beatty (1969)		
Eutroboralf	New Mexico	Sandstone				2	2	2				-18-			76	6		Anderson et al. (1975)		
Hapludalf	Wisconsin	Thin loess over till	1	2		2	2	2	1			13	14	8	61			Borchardt et al. (1968)		
Hapludalf	Oregon	Greenish tuff & breccia	1			3					22	52	3	1	t	6	11	Paeth et al. (1971)		
Haploxeralf	Spain	Calcareous alluvium			1	2	3	1				-a-		t	t			Delgado et al. (1994)		
Rhodoxeralf		Red sandstone			1	1	3	1				-a-		c				Goss and Allen (1968)		
Haplustalf	Texas	Granite				2	2	1												
Natraustalf						2	2	1												
Paleudalf	Texas	Coastal plains sediments				1	1	3+		1								Vepraskas and Wilding (1983)		
Haplustalf?	India	Shist/quartzite/phyllite				t	2	3	1	1		31			65			Sahu et al. (1990)		
Kandiudalf	Nigeria	Olivine basalt						3	t	1								Gallez et al. (1975)		
(reclassified)																				
Plinthustalf		Basement complex			t		1	3		1										
Paleustalf	Niger	Eolian sands						3		1								West et al. (1984)		

*Relative amounts: 3 = >50%; 2 = 10 to 50%; 1 = <10%; t = trace

Ch = chlorite; Vr = vermiculite; In = interstratified 2:1 minerals; Sm = smectite; Mi = hydrous mica or illite; Ka = kaolinite; Qz = quartz; GH = goethite and/or hematite.

** Values are for a major fraction of sand or silt.

Ps = Pseudomorphs of clay minerals; Kf = K feldspar; Pl = plagioclase feldspar; Mi = mica; Qz = quartz; Vg = volcanic glass; AP = amphibole and/or pyroxene; Ze = zeolite; a = abundant; c = common; t = trace. +; Chloritized vermiculite also noted

field operations to separate Alfisols from Ultisols, but care must be exercised as this relationship changes with mineralogy.

The mineralogy from a range of Alfisols illustrating variations in parent materials, climates, and degree of development is presented in Table 6.60. Most Alfisols contain substantial quantities of 2:1 clay minerals such as smectite and hydrous mica, and perhaps some vermiculite, chlorite and kaolinite. However, for more intensely weathered Alfisols (Kandiudalf, Paleudalf, Paleustalf, and Plinthustalf), kaolinite is the dominant clay mineral in the argillic horizon. Similarly, the most common mineral dominating the sand and silt fraction is quartz (Table 6.62) which comprised > 90% of the sand fraction in the Paleustalf and Paleudalf. The Udalf from Oregon which formed from tuff and breccia, was dominated by plagioclase feldspars. Most Alfisols contain significant quantities of weatherable minerals including K feldspars, plagioclase and mica in the sand and silt fractions.

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6.12 Ultisols

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6.12.1 Introduction

Ultisols are a group of soils with an argillic or kandic horizon and low base saturation in lower subsoil horizons. These soils may be found on a variety of parent materials and under a range of climatic conditions, though most have developed under forest vegetation in humid climates. Ultisols are generally considered to be less productive than soils in many of the other orders, but in many regions of the world, they are the most productive soils available. With proper management of organic residues, fallow periods, and/or chemical inputs, the productivity of this resource can be enhanced and maintained. This section presents a brief overview of Ultisol genesis and properties. For a more comprehensive review, the reader is referred to Miller (1983) and West et al. (1997).

6.12.2 Classification: Historical and Current

As soil classification has evolved in the United States soils currently considered to be Ultisols have been included with the Red soils, Yellow soils, or Lateritic soils (Marbut, 1928), the Yellow Podzolic or Red Podzolic great soil groups (Baldwin et al., 1938), and the Red-Yellow Podzolic great soil group (Kellogg, 1949). The order Ultisols was introduced in the 7th approximation (Soil Survey Staff, 1960) to include soils with an argillic horizon and base saturation $< 35\%$ in lower subsoil horizons. Since publication of *Soil Taxonomy* (Soil Survey Staff, 1975), the classification of Ultisols has been revised to include kandic horizons as criteria for placement in the order and to include soils in any temperature regime. For a comprehensive definition of Ultisols, the reader is referred to the *Keys to Soil Taxonomy* (Soil Survey Staff, 1998). Table 6.63 lists current orders, suborders and great groups in Ultisols (Soil Survey Staff, 1998).

6.12.3 Geographic Distribution

Worldwide, Ultisols are currently estimated to cover about 11,054,000 km² which is 8.4% of the ice-free global landmass (Table 6.1; Fig. 6.67). About 80% of Ultisols are in tropical regions, and about 18% of the tropics is covered by Ultisols (Eswaran, 1993). About 50% of Ultisols are Udults, 35% Ustults, and 12% Aquults with Humults and Xerults comprising the remaining 3% (Table 6.1). In the United States, Ultisols cover about 860,000 km² (Table 6.64), and occur in 30 states and Puerto Rico (Fig. 6.68). The largest concentration of Ultisols in the United States is in the east central, southeast, and south central parts of the country. Ultisols are also extensive on the older islands of Hawaii, along the west coast, and in unglaciated regions of the northeast and north central parts of the country.

6.12.4 Genesis of Ultisols

Soil as an independent natural body whose properties are a function of five soil-forming factors: local climate, parent materials, organisms, relief, and age (Dokuchaev, 1883; Jenny, 1941), is a unifying philosophy in pedology. As such, Ultisol genesis will be briefly discussed in terms of these factors.

6.12.4.1 Climate

The definition of Ultisols implies two rainfall/evapotranspiration conditions. First, there must be some time during the year when evapotranspiration exceeds precipitation, as this appears to be a prerequisite for the formation of an argillic horizon (van Wambeke, 1991). Second, precipitation must exceed the capacity of the soil to retain water during some time of the year so that water percolates through the solum to remove basic cations and maintain low base saturation ($< 35\%$ by sum of cations [pH 8.2] which corresponds roughly to 50% base saturation measured at pH 7.0) in the subsoil (Miller, 1983). Most Ultisols occur in regions with mean annual air temperatures above 6 °C. Thus, the climate conducive to the formation of Ultisols is typically humid tropical or humid warm temperate. As many Ultisols have evolved over long time periods, effects of paleoclimate on Ultisol development cannot be ignored.

6.12.4.2 Parent Material

The definition of Ultisols requires the presence of either an argillic or a kandic horizon, both of which contain silicate clays. Therefore, the parent material of Ultisols must be one that contains either phyllosilicates or primary minerals that can weather to produce them. As the vast majority of rocks meet one or both of these criteria, a wide variety of geologic formations are capable of producing parent materials suitable for Ultisol formation. Residual parent materials that have weathered *in situ*

Table 6.63 Listing of suborders, great groups and subgroups in the Ultisols order [Soil Survey Staff, 1998]

Suborder	Great Group	Subgroup
Aquults	Plinthaquults	Kandic, Typic
	Fragiaquults	Aeric Plinthic, Umbric, Typic
	Albaquults	Vertic, Kandic, Aeric, Typic
	Kandiaquults	Acraquoxic, Arenic Plinthic, Arenic Umbric, Arenic, Grossarenic, Plinthic, Aeric, Umbric, Typic
	Kanhaplaquults	Aquandic, Plinthic, Aeric Umbric, Aeric, Umbric, Typic
	Paleaquults	Vertic, Arenic Plinthic, Arenic Umbric, Arenic, Grossarenic, Plinthic, Aeric, Umbric, Typic
	Umbraquults	Plinthic, Typic
	Epiaquults	Vertic, Aeric Fragic, Fragic, Arenic, Grossarenic, Aeric, Typic
	Endoaquults	Arenic, Grossarenic, Aeric, Typic
Humults	Sombrihumults	Typic
	Plinthohumults	Typic
	Kandihumults	Andic Ombroaquic, Ustandic, Andic, Aquic, Ombroaquic, Plinthic, Ustic, Xeric, Anthropic, Typic
	Kanhaplohumults	Lithic, Ustandic, Andic, Aquic, Ombroaquic, Ustic, Xeric, Anthropic, Typic
	Palehumults	Aquandic, Andic, Aquic, Plinthic, Oxyaquic, Ustic, Xeric, Typic
	Haplohumults	Lithic, Aquandic, Aquic, Andic, Plinthic, Oxyaquic, Ustic, Xeric, Typic
Udults	Plinthudults	Typic
	Fragiudults	Arenic, Plinthaquic, Glossaquic, Aquic, Plinthic, Glossic, Humic, Typic
	Kandiudults	Arenic Plinthaquic, Aquic Arenic, Arenic Plinthic, Arenic Rhodic, Arenic, Grossarenic Plinthic, Grossarenic, Acrudoxic Plinthic, Acrudoxic, Plinthaquic, Aquandic, Andic, Aquic, Plinthic, Ombroaquic, Oxyaquic, Sombic, Rhodic, Typic
	Kanhapludults	Lithic, Fragiaquic, Arenic Plinthic, Arenic, Acrudoxic, Plinthaquic, Andic, Aquic, Ombroaquic, Oxyaquic, Fragic, Plinthic, Rhodic, Typic
	Paleudults	Vertic, Spodic, Arenic Plinthaquic, Aquic Arenic, Plinthaquic, Fragiaquic, Aquic, Anthraquic, Oxyaquic, Lamellic, Arenic, Plinthic, Psammentic, Grossarenic Plinthic, Plinthic, Arenic Rhodic, Arenic, Grossarenic, Fragic, Rhodic, Typic
	Rhodudults	Lithic, Psammentic, Typic
	Hapludults	Lithic Ruptic-Entic, Lithic, Vertic, Fragiaquic, Aquic Arenic, Aquic, Fragic, Oxyaquic, Lamellic, Psammentic, Arenic, Grossarenic, Inceptic, Humic, Typic
	Plinthustults	Haplic, Typic
	Kandiustults	Acrustoxic, Aquic, Arenic Plinthic, Arenic, Udandic, Andic, Plinthic, Aridic, Udic, Rhodic, Typic
Ustults	Kanhaplustults	Lithic, Acrustoxic, Aquic, Arenic, Udandic, Andic, Plinthic, Ombroaquic, Aridic, Udic, Rhodic, Typic
	Paleustults	Typic
	Rhodustults	Lithic, Psammentic, Typic
	Haplustults	Lithic, Petroferric, Aquic, Arenic, Ombroaquic, Plinthic, Kanhaplic, Typic
	Palexerults	Aquandic, Aquic, Andic, Typic
	Haploxerults	Lithic Ruptic-Inceptic, Lithic, Aquic, Andic, Lamellic, Psammentic, Arenic, Grossarenic, Typic

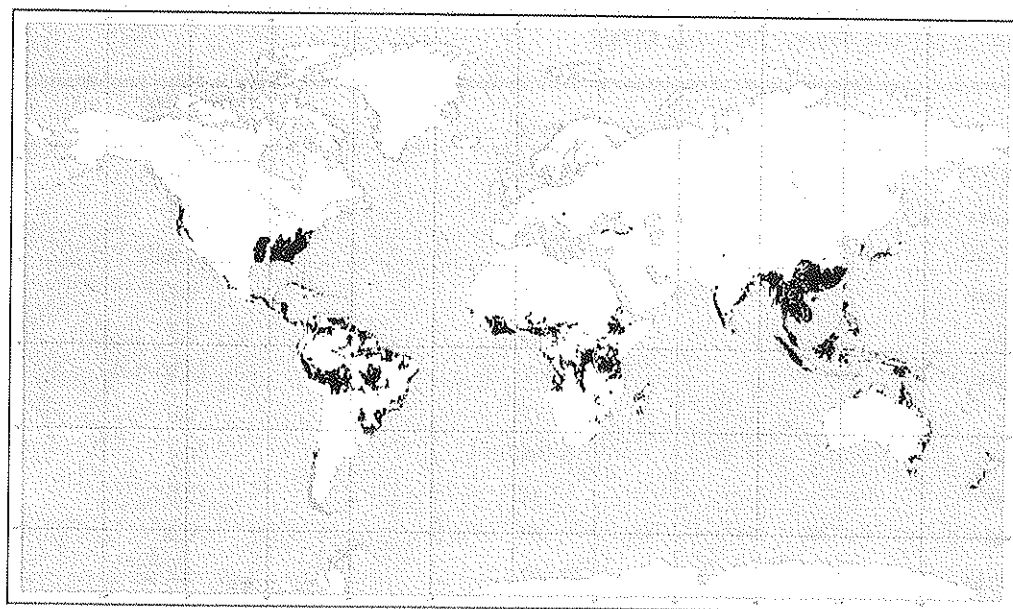


Fig. 6.67 Global distribution of Ultisols [courtesy of UDSA-NRCS, Soil Survey Division, World Soil Resources, 1998]

are usually saprolite. In contrast, transported sediments that form Ultisol parent materials have frequently been preweathered and may have gone through more than one weathering cycle (Fiskell and Perkins, 1970; Allen and Fanning, 1983).

6.12.4.3 Biota

Although Ultisols occur under many tropical and subtropical ecosystems, forests are the dominant vegetation on most natural Ultisol landscapes. The dynamics in forest soils appear to accelerate the formation of argillic horizons through the desiccating effect of tree roots that absorb water, but not colloids in suspension. In Africa and South America, vast areas of Ultisols are under savanna vegetation. It cannot be ascertained, however, whether the soils formed under the present vegetation or under forests in one or more humid paleoclimates.

6.12.4.4 Relief

Ultisols occur on almost all landforms. The one characteristic that Ultisol landforms have in common, however, is geomorphic stability over long periods of time. Geomorphic studies of tropical landscapes invariably show that Ultisols occupy geomorphic positions that are younger and less stable than the surfaces where Oxisols occur, but older and more stable than those of soils in other orders with which they are geographically associated (Beinroth et al., 1974; Lepsch and Buol, 1974; Beinroth, 1981). In temperate regions where Oxisols are absent, Ultisols have developed on the oldest, most stable and highly weathered landscapes such as the central Missouri Ozarks and the coastal plain of the southeastern United States (Scrivner et al., 1966; Daniels et al., 1971).

Table 6.64 Distribution of Ultisols in the conterminous United States and Puerto Rico [Courtesy of USDA-NRCS, Soil Survey Division, National Soil Survey Center]

State	Area (km ²)	Percentage of Total
Alabama	97,965	11.4
Arkansas	68,644	8.0
California	8,418	1.0
Delaware	3,421	0.4
Florida	28,106	3.3
Georgia	114,802	13.3
Idaho	17	0.0
Indiana	3,158	0.4
Illinois	44	0.0
Kansas	79	0.0
Kentucky	27,402	3.2
Louisiana	23,395	2.7
Massachusetts	22	0.0
Maryland	16,004	1.9
Missouri	28,534	3.3
Mississippi	49,826	5.8
North Carolina	85,057	9.9
New Jersey	7,617	0.9
New York	579	0.1
Oregon	11,195	1.3
Ohio	7,137	0.8
Oklahoma	16,784	1.9
Pennsylvania	39,375	4.6
Puerto Rico	1,784	0.2
South Carolina	54,720	6.3
Tennessee	44,317	5.1
Texas	24,967	2.9
Virginia	69,910	8.1
Washington	3,574	0.4
West Virginia	25,435	2.9
Total	862,288	100.0

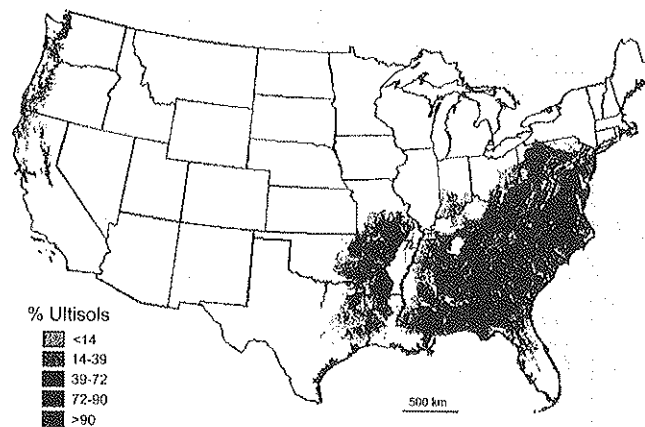


Fig. 6.68 Distribution of Ultisols in the United States [courtesy of USDA-NRCS, Soil Survey Division, National Soil Survey Center]

6.12.4.5 Time

The time required for the formation of Ultisols depends on the other pedogenic factors, notably weatherability of the parent rock, composition of transported parent materials, and climate and its fluctuations over time. There is much room for variability in these conditions, and consequently, for the age of Ultisols. The Soil Survey Staff (1975) concluded "that formation of the argillic horizon ordinarily requires a few thousands of years." This relatively short time may suffice to produce a minimal expression of an argillic horizon. The development of the thick argillic horizons that typify the Pale great groups of Ultisols, however, would certainly take longer. As such, age of Ultisols has been reported to range from < 12,000 yr for soils on preweathered sediments to > 15 million yr on marine sediments in the Coastal Plain of the southeast United States. (Daniels et al. 1971; Gamble et al., 1970; Miller, 1983).

6.12.5 Morphological Properties

6.12.5.1 Argillic and Kandic Horizons

Other than low base saturation, the one property that distinguishes Ultisols from all but one other soil order (Alfisols, Section E, Chapter 6.11) is the requirement of a clay increase between surface and subsoil horizons in the form of either an argillic or kandic horizon. This clay increase essentially determines the morphology of Ultisols and influences many chemical and physical properties. The depth, thickness, and clay content of the argillic or kandic horizon, however, can vary widely, and this variation has considerable impact on interpretation of these soils for management. Fig. 6.69 illustrates the range in depth and clay content of argillic or kandic horizons that can occur in Ultisols.

In addition to a clay increase, the presence of oriented illuvial clay as clay coatings on pore walls (Fig. 6.70) or bridges between grains is required for an argillic horizon and clay coatings are commonly present in kandic horizons. Illuvial clay is pedogenically significant in terms of land surface age and stability, water movement through the soil, and periods of moisture deficit. In addition, clay coatings may retard water movement between pores and the matrix, slow diffusion of ions into and out of the bulk soil, and have different exchange properties than the bulk soil.

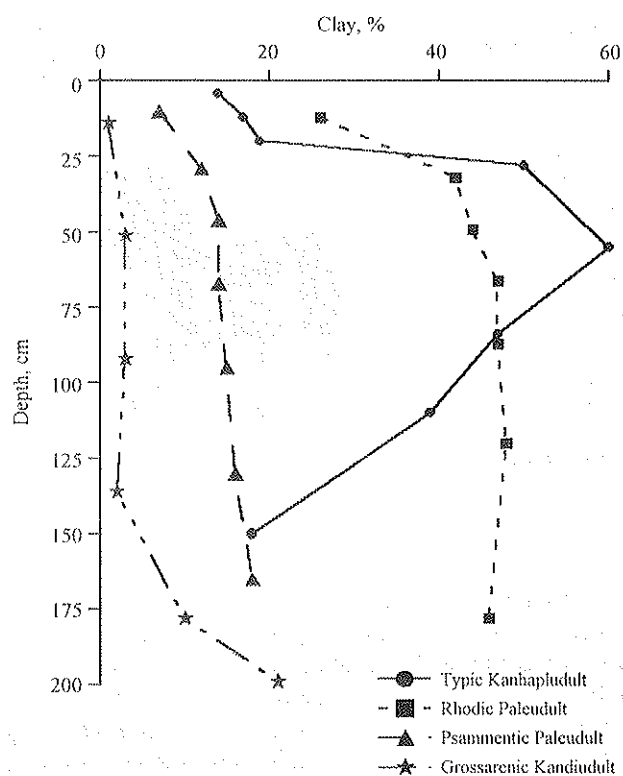


Fig. 6.69 Clay distribution with depth for four subgroups of Ultisols

6.12.5.2 Soil Color

Color of surface horizons in Ultisols is commonly brown to yellowish brown to dark brown as would be expected for soils with relatively low amounts of organic C. However, in cool climates or poorly drained conditions with slow organic matter decomposition, thick, dark, acidic surface horizons

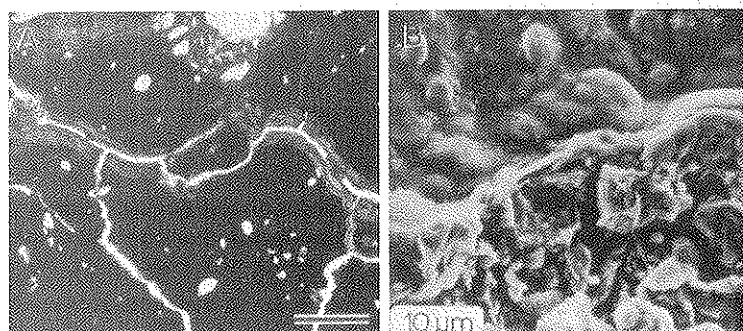


Fig. 6.70 (A) Thin section micrograph of clay coatings in Btv2 horizon of a Plinthic Kandiudult. Plane-polarized light. Bar length = 0.5 mm; (B) Scanning electron micrograph of clay coating in 2Btx1 horizon of a Typic Fragiudult [Original micrograph from Michael Thompson, Iowa State University]

(umbric epipedons) are common. Many well-drained Ultisols have reddish Bt horizons because of the presence of hematite as a pigmenting agent. Yellowish-brown subsoil horizons, also common in Ultisols, lack hematite because either the parent material lacked mineral precursors for hematite, environmental conditions precluded hematite formation (Schwertmann and Taylor, 1989), or hematite was preferentially reduced and lost under short periods of saturation leaving behind more stable goethite (Macedo and Bryant, 1989; Dobos et al., 1990). Under extended periods of saturation and reduction, Ultisols will develop redox concentrations and depletions or gray matrices as Fe is reduced and mobilized.

6.12.5.3 Plinthite

Plinthite is defined in *Soil Taxonomy* as an Fe-rich, humus-poor mixture of clay with quartz and other diluents (Soil Survey Staff, 1998). This material most often occurs in lower subsoil horizons, is weakly indurated, and is normally low in basic cations and primary silicate minerals other than quartz. Kaolinite is commonly the only clay mineral present with abundance. The reader is referred to van Wambeke (1991) for a complete discussion of the genesis and kinds of plinthite and petroplinthite.

From a pedological perspective, plinthite is an interesting soil feature, but it also has significance in terms of movement of water through soils. Daniels et al. (1978) suggested that 10% platy plinthite would perch water. A similar amount of nodular plinthite was not considered to be water restrictive, but horizons subjacent to the plinthic horizon did restrict water movement. Movement of water and solutes through horizons containing plinthite is primarily restricted to areas of redox depletions (Carlan et al., 1985; Blume et al., 1987; Shaw et al., 1997). Because of water perching by plinthic or subjacent horizons, plinthite in soils suggests that water and solutes are moving from high to low landscape positions as shallow subsurface flow (Hubbard and Sheridan, 1983; Shirmohammadi et al., 1984).

6.12.6 Mineralogical Properties

6.12.6.1 Clay Minerals

The clay fraction of subsoil horizons of most Ultisols is dominated by kaolinite because of the stability of kaolinite as compared to other phyllosilicates (Jackson et al., 1948; Allen and Fanning, 1983). In addition, gibbsite, K mica, K feldspar, and amorphous silica may transform directly to kaolinite without passing through other phyllosilicate intermediates (Garrels and Christ, 1965). As such, kaolinite in soils has been shown to be an early weathering product of biotite and K feldspar in acid gneiss and schist parent materials (Robertus and Buol, 1985; Robertus et al., 1986; Norfleet and Smith, 1989).

In addition to kaolinite, hydroxy Al-interlayered vermiculite (HIV) is often abundant in A and E horizons of Ultisols and may be a major component of upper Bt horizons of these soils (Bryant and Dixon, 1964; Fiskell and Perkins, 1970; Carlisle and Zelazny, 1973; Harris et al., 1980; Karathanasis et al., 1983). Appreciable amounts of mica, smectite, and other 2:1 minerals have also been reported to occur in Ultisols (Nash, 1979; Harris et al., 1984; Karathanasis et al., 1986; Nash et al., 1988). The 2:1 minerals in these Ultisols were inherited from the parent materials instead of forming *in situ* from primary minerals or other phyllosilicates.

A common but minor mineral component of many Ultisols is gibbsite (Carlisle and Zelazny, 1973; Hajek and Zelazny, 1985). Because this mineral is an end product of silicate mineral weathering (Jackson et al., 1948), minor amounts should be expected to be found in Ultisols. Clay fractions of Ultisols developed from acid gneiss and schist in the southeastern United States, however, have been

reported to contain as much as 70% gibbsite formed as a product of K feldspar weathering at the rock-saprolite interface (Losche et al., 1970; McCracken et al., 1971; Norfleet and Smith, 1989).

Iron oxides and oxyhydroxides are also common, but minor components in the clay fraction of Ultisols. Goethite is the most common Fe oxide mineral in most Ultisols, but many well drained Ultisols have appreciable amounts of hematite. Hematite formation is favored by warm temperatures and distinct dry seasons (Schwertmann and Taylor, 1989), which are environmental conditions commonly associated with Ultisols.

6.12.6.2 Sand and Silt Minerals

The sand and silt fraction of Ultisols commonly is dominated by quartz, and many Ultisols have > 90% quartz and other resistant minerals in their sand and silt separates. However, weakly to moderately developed Ultisols may have appreciable quantities of mica, feldspars, and other weatherable minerals in their coarse fraction (Norfleet and Smith, 1989; Robertus et al., 1986; Bockheim et al., 1996).

6.12.7 Physical Properties

6.12.7.1 Bulk Density, Coefficient of Linear Extensibility (COLE), Water Retention, and Hydraulic Conductivity (K)

Bulk density, COLE, water retention, and hydraulic conductivity of Ultisols vary considerably depending on soil and horizon properties including particle size distribution, organic matter content, mineralogy, macroporosity, parent material, presence of plinthite, and, for surface horizons, management. A horizons of uncultivated Ultisols often have lower bulk density than subsoil horizons,

Table 6.65 Ranges in selected physical properties of Ultisols

Property	Value	References
Bulk Density (Mg m^{-3})		
A horizon	0.54-1.85	Bruce et al. (1983)
Bt horizon	1.05-1.70	Hubbard et al. (1985)
		Quisenberry et al. (1987)
		Southard and Buol (1988a)
		Norfleet and Smith (1989)
COLE (Bt horizons)	0.006-0.157	Karathanasis and Hajek (1985)
		Griffin and Buol (1988)
		Southard and Buol (1988a)
Water retention ($\text{cm}^3 \text{cm}^{-3}$)		
A horizons		Bruce et al. (1983)
-10 kPa	0.13-0.21	Hubbard et al. (1985)
-30 kPa	0.09-0.18	Quisenberry et al. (1987)
-1,500 kPa	0.05-0.09	
Bt horizons		
-10 kPa	0.20-0.42	
-30 kPa	0.17-0.40	
-1,500 kPa	0.09-0.33	
Saturated Hydraulic Conductivity ($\mu\text{m s}^{-1}$)		
Bt horizons	0.01-13.1	Hubbard et al. (1985)
Btv horizons	0.3-1.3	Southard and Buol (1988b)

but with tillage and the accompanying loss of organic matter, compaction, and aggregate breakdown, the A horizon may become as dense or more dense than underlying Bt horizons (Table 6.65) (Hubbard et al., 1985). Shrink/swell potential for kaolinitic Ultisols is low (Table 6.63) (Hubbard et al., 1985; Southard and Buol, 1988a), but COLE values up to 0.157 have been reported for Bt horizons from Ultisols with smectitic mineralogy (Karathanasis and Hajek, 1985; Griffin and Buol, 1988). However, smectitic Ultisols with high shrink/swell potential often lack wide cracks common to Vertisols and Vertic subgroups of other orders (Karathanasis and Hajek, 1985).

Ultisols that are placed in Typic subgroups and fine-loamy or fine particle size families would generally be considered capable of retaining and supplying sufficient moisture for plant growth in humid climates (Table 6.65). However, water often becomes a limiting factor for crop growth and yield for Ultisols in sandy or coarse-loamy particle size families or that have thick sandy epipedons (Arenic and Grossarenic subgroups).

Most studies of hydraulic properties of Ultisols have found macroporosity to be the best predictor of saturated hydraulic conductivity (K_s) (O'Brien and Buol, 1984; Southard and Buol, 1988b). Preferential flow paths in Ultisols in the Piedmont and Coastal Plain of Georgia have textures similar to the matrix of the horizon, but the preferential flow zones have a significantly greater number and area of pores > 0.05 mm diameter (Table 6.66) (Franklin, 1994; Shaw et al., 1997). Formation of coarse pores in Ultisols is most often attributed to microfabric alteration and channel formation by roots and burrowing organisms (O'Brien and Buol, 1984; Southard and Buol, 1988b; Franklin, 1994; Shaw et al., 1997).

6.12.7.2 Infiltration and Surface Crusting

In many Ultisols, one factor affecting infiltration of water is formation of a surface crust or seal. For 25 Ultisols from the southeastern United States tested under 50–90 mm h^{-1} simulated rainfall, infiltration rate decreased rapidly during the first 10 to 20 mm of the rainfall event, and the infiltration rate at the end of rainfall was less than 10 mm h^{-1} for all of the soils (Miller and Radcliffe, 1992). In general, Ultisols in this region with sandy loam texture and high amounts of water dispersible clay are most prone to crusting. Soils higher in clay tend to take longer to form a surface crust, and sands and loamy sands form only weakly expressed crusts (Miller and Bahrudin, 1986; Radcliffe et al., 1991; Miller and Radcliffe, 1992; Chiang et al., 1993).

Table 6.66 Mean pore area for dyed and undyed areas after break through of methylene-blue for Ultisols from the Piedmont and Coastal Plain of Georgia. [Franklin (1994); Shaw et al. (1997)]

Horizon	Mean Pore Area [†]	
	Dyed	Undyed
	-----%-----	
Typic Kanhapludult (Piedmont)		
A	19.0	3.0
BA	17.7	5.0
Bt	15.8	2.4
Bt	14.4	2.8
Plinthic Kandiudult (Coastal Plain)		
Bt	5.0	1.1
Btv	5.9	0.8
BC	NA [‡]	0.9

[†] Area of pores > 0.05 mm equivalent circular diameter determined from impregnated polished blocks with image analysis

[‡] The BC horizon had insufficient dyed area for evaluation of pore area.

Table 6.67 Comparison of organic C, aggregate stability, infiltration rate, and percentage moist days under conventional and no-till cropping systems [West et al., 1991; Bruce et al., 1992]

Cropping System ^H	Organic C	Aggregate Stability	Infiltration Rate ^I		Moist Days [†] % of growing season
			residue	residue removed	
	-----g kg ⁻¹ -----		-----mm h ⁻¹ -----		
CTG	10.4	580	36	22	29
NTG	23.3	890	50	46	49

^H CTG—Conventional-tilled grain sorghum (*Sorghum bicolor* (L.) Moench); NTG—No-till grain sorghum with clover winter cover.

^I Infiltration rate at the end of a 1 h of rainfall simulation at approximately 64 mm h⁻¹.

[†] Moist days are defined as the days with soil moisture tension > 0.1 Mpa.

Because energy inputs from rainfall are needed to form surface crusts, crusting is minimal if the soils have residue or vegetative cover and organic matter is maintained in the upper part of the soil. For Ultisols in the Piedmont of Georgia, infiltration rates after 1 h of simulated rainfall were about 40% higher for soils that had been under no-tillage for 5 yr than for soils that had been conventionally tilled (Table 6.67) (Bruce et al., 1992).

6.12.8 Chemical Properties

The requirement that Ultisols have base saturation < 35% in the lower part of the subsoil carries with it certain associated properties including low pH, potentially high Al saturation, appreciable weathering and associated kaolinitic mineralogy, and in many cases, relatively high contents of Fe and Al oxides and oxyhydroxides. This suite of properties has significant impact on the chemical properties of Ultisols and how these properties affect use and management of this soil order.

6.12.8.1 Cation and Anion Exchange Capacity

The CEC of Ultisol horizons depends on the amount and type of clay, Fe, Al, and Mn oxide content, organic matter content, soil solution electrolyte concentration, and pH of the horizon. Reported CEC values for Ultisol horizons range from < 3 to > 20 cmol_c kg⁻¹ (Sanchez, 1976; Carlisle et al., 1985; Karathanasis et al., 1986). Generally, Ultisols with kaolinitic mineralogy have mostly pH-dependent charge and low CEC, while those with mixed or smectitic mineralogy have a greater proportion of fixed charge and higher CEC values.

Even when the net charge is negative, many Ultisols have an appreciable positive charge and anion exchange capacity (AEC). Gillman and Sumner (1987), in a study of four Ultisols from the Piedmont of Georgia, reported AEC values at the pH of the soil that ranged from 0.1 to 1.2 cmol_c kg⁻¹, which was attributed to Fe and Al oxides in these soils. For Ultisols from Georgia and South Africa, Grove et al. (1982) reported AEC values ranging from 0.03 to 1.91 cmol_c kg⁻¹. Measurable AEC has also been reported for Ultisols from tropical Australia and Peru (Gillman and Sumpter, 1986; Gillman and Sinclair, 1987).

6.12.8.2 Acidity and Exchangeable Aluminum

Below a pH of about 5.5, Al released from weathering of primary and secondary minerals is present as part of the exchange complex. An Al saturation of greater than 60% (measured as a percentage of the unbuffered CEC or ECEC) has been reported as the threshold that results in soil solution Al concentrations > 1 mg L⁻¹ which may reduce yield in Al sensitive crops (Nye et al., 1961; Kamprath,

1970; Sanchez, 1976). Farina and Channon (1991), however, observed yield reductions in maize for Al saturations > 25% of the ECEC.

6.12.8.3 Phosphorus

Phosphate is specifically adsorbed by Fe, Al, and Mn oxides and amorphous or poorly crystalline aluminosilicates. Because of eluviation and weathering, surface horizons of most Ultisols have low contents of these components. Thus, P sorption is generally lower in Ultisols than in Oxisols and Andisols (Sanchez, 1976; van Wambeke, 1991). However, differences in crystallinity of Fe oxides among soils may confound this relationship within and among orders (Pratt et al., 1969; Fox et al., 1971).

6.12.9 Biological Properties

6.12.9.1 Organic Matter

A common misconception is that Ultisols have low contents of organic matter compared to other soil orders. While Ultisols generally have lower amounts of organic C than Mollisols, Oxisols, and Andisols, levels of organic C in Ultisols are similar to those in Alfisols both within and across geographic areas (Sanchez, 1976). There is also little difference in organic C content between Ultisols in temperate and tropical climates (Buol, 1973; Sanchez, 1976).

Organic matter contents in Ultisols decrease rapidly when these soils are converted from forest or pasture to intensive cultivation. For Ultisols in Georgia, Giddens (1957) found that organic matter contents decreased from 20.5 to 16.6 g kg⁻¹ in 24 months after conversion of forest to intensive cultivation. Use of no-tillage, cover crops, and residue management, however, can help to restore and maintain organic C levels in Ultisols at precultivation levels. Bruce et al. (1992) reported that use of no-tillage and winter cover crops increased organic C levels in the upper 15 mm of degraded Ultisols from 10.4 to 23.3 g kg⁻¹ over a 5 yr period. Associated with the increase in organic C in these soils was a significant increase in water stable aggregates and significant decreases in runoff and interrill soil loss (Table 6.68) (West et al., 1991; Bruce et al., 1992; West et al., 1992).

6.12.9.2 Biologic Populations and Processes

Populations of microorganisms in Ultisols are not appreciably different from soils in other orders. Surface horizons of most Ultisols have acid pH values. Thus, fungi may comprise a greater proportion of the microbial biomass in these soils because of the greater ability of fungi to survive under acid conditions than other microorganisms (Alexander, 1977). Low pH has been reported to reduce rates of nitrification and denitrification in pure cultures (Alexander, 1977). Natural adaptation by

Table 6.68 Interrill soil loss and runoff from three Ultisols with varying tillage and surface cover under 50 mm h⁻¹ rainfall intensity for 1 hour on 5–9% slopes [West et al., 1991]

Tillage	Surface condition	-----Interrill soil loss-----			-----Total runoff-----		
		Soil 1	Soil 2	Soil 3	Soil 1	Soil 2	Soil 3
		-----g m ⁻² -----			-----mm-----		
Conventional	Crusted	0.46	0.56	0.33	16	23	17
	Tilled	0.72	0.79	0.48	18	23	16
No-till	Bare	0.14	0.19	0.07	3	3	6
	Residue	0.07	0.03	0.02	1	1	1

microorganisms to local field conditions, however, probably alters the effects of pH and other environmental conditions observed in the laboratory.

6.12.10 Conclusions

Worldwide, Ultisols are a widespread soil resource, which because of their extensive weathering, low base saturation, and associated properties, are often considered to be unproductive. It is true that Ultisols, in general, have lower native fertility than Mollisols and Alfisols, and in areas where these orders are abundant, Ultisols are the less desirable soils. In many regions of the world, however, Ultisols are dominant or are the most productive soils available. Only with a thorough understanding of the genesis, properties, and response to management of these soils can the productivity of this valuable resource be maintained and enhanced.

6.12.11 References

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6.13 Oxisols

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6.13.1 Introduction

The deep, red and highly weathered soils of the tropics have long fascinated pedologists, particularly those from the temperate region. The uniqueness of these soils lies not only in their properties, but also in their geographic distribution being confined almost exclusively to the tropics. As this account is intentionally concise, ample references are provided for those seeking additional or more detailed information.

6.13.2 Historical Background

In the early literature, the highly weathered soils of the tropics were identified as red soils, red loams or red earths. In an historical report on travels in South India, Buchanan (1807) described soil material used for construction and called it laterite (Buchanan, 1807). The term was adopted by the pedologic community and is still used today with a wide variety of meanings (Alexander and Cady, 1962; Maignen, 1966). In 1949, a group of scientists proposed the term Latosols (Cline, 1975) that soon became popular and, by the midfifties, the concept of Latosols as highly weathered soils with a low negative charge was firmly established. Several soil classification systems developed during the fifties and sixties recognized such soils in a separate class with the nomenclature shown in Table 6.69.

The term Oxisol originated around 1954 (Smith, 1963; 1965) during the development of *Soil Taxonomy* (Soil Survey Staff, 1975). In 1978, an International Committee on the Classification of Oxisols (ICOMOX) was formed (Buol and Eswaran, 1988) to initiate discussions that would lead to an improved classification. Through circular letters and four international soil classification workshops held in Brazil (1976), Malaysia and Thailand (1978), Rwanda (1981), and Brazil (1986), the current state of knowledge of these soils was assembled and documented (Camargo and Beinroth, 1978; Beinroth and Paramanathan, 1978; Beinroth and Panichapong, 1978; Beinroth et al., 1983; Beinroth et al., 1986).

Table 6.69 Synonyms for Oxisols in other classification systems

Synonym	Classification System	Source
Sols Ferrallitiques	French	Aubert, 1958
Latosols	Brazilian	Bennema et al., 1959
Solos Ferraliticos	Portuguese	Botelho da Costa, 1954, 1959
Oxysols	Ghana	Charter, 1958
Ferralsols		FAO/UNESCO, 1971-1976
Rotlehm	German	Harrassowitz, 1930
Laterites		Maignen, 1966
Red Earths	British	Robinson, 1951
Kaolisols	Belgian	Tavernier and Sys, 1965

6.13.3 Concept and Classification of Oxisols

6.13.3.1 Perceptions

The rationale of the taxon of Oxisols is based largely on the concept of Latosols which was developed to designate "all zonal soils having their dominant characteristics associated with low silica:sesquioxide ratios, low base exchange capacity, low activity of clays, and low content of weatherable minerals" (Kellogg, 1949; Cline, 1975). These characteristics were selected because they were thought to manifest the effects of advanced pedogenesis under tropical conditions.

6.13.3.2 Definition

Simplified from the *Keys to Soil Taxonomy* (Soil Survey Staff, 1998), Oxisols may be defined as those soils that fail to meet the criteria definitive for Gelisols, Histosols, Spodosols, and Andisols but have either (1) an oxic horizon within 150 cm of the mineral soil surface and no kandic horizon within this depth, or (2) 40% or more clay in the surface horizon and a kandic horizon within 100 cm of the mineral soil surface that meets the weatherable mineral properties of the oxic horizon.

The definition of Oxisols has undergone significant changes since it was first published in *Soil Taxonomy* (Soil Survey Staff, 1975). Consequently, more soils now qualify for Oxisols; this should be kept in mind when consulting the earlier literature. It should also be mentioned that not all intensively weathered soils are Oxisols. Highly weathered soils with a marked clay increase below a light textured surface horizon are by definition excluded from the Oxisols and are classified as kandic great groups of Ultisols and Alfisols (van Wambeke, 1992; Soil Survey Staff, 1998).

6.13.3.3 Diagnostic Criteria

Key to the identification of Oxisols is the presence of either an oxic horizon or a kandic horizon underlying a surface horizon with $\geq 40\%$ clay. The key properties of the oxic horizon are its charge characteristics and negligible amounts of weatherable minerals. The charge characteristics are defined by the magnitude of the variable charge ($\text{CEC by NH}_4\text{OAc} < 16 \text{ cmol}_c \text{ kg}^{-1} \text{ clay}$), and the permanent charge estimated by the Effective Cation Exchange Capacity ($\text{ECEC} < 12 \text{ cmol}_c \text{ kg}^{-1} \text{ clay}$). The low weatherable mineral requirement ensures that there are few weatherable minerals that could release plant nutrients. The particle size class of the oxic horizon is sandy loam or finer and its upper boundary is diffuse. To be diagnostic, the oxic horizon must have a minimum thickness of 30 cm and occur within 150 cm of the mineral soil surface.

The kandic horizon (Moormann, 1985) shares some of the properties of the oxic and argillic horizons. Like the argillic horizon, clay increases with depth but the charge characteristics are those of the oxic horizon. To be diagnostic for Oxisols, the kandic horizon must meet the weatherable mineral requirements of the oxic horizon, occur beneath a surface horizon that has $\geq 40\%$ clay, and have its upper boundary within 100 cm of the soil surface.

Suborders of Oxisols are differentiated on the basis of soil moisture regimes, aquic conditions, the presence of a histic epipedon, and soil color. Defining criteria for great groups are the kandic and sombric horizons, plinthite, acric properties, and base saturation. The 14 differentiae used to establish subgroups are: andic-like properties, $\text{Al}+\text{Fe}$ percentage, aquic conditions, CEC, pH ($\text{pH}_{\text{KCl}} - \text{pH}_{\text{H}_2\text{O}}$), histic epipedon, kandic horizon, lithic contact, organic C content, petroferic contact, plinthite, soil color, soil depth, and the sombric horizon.

At the family level, criteria include particle size class, mineralogy class, and soil temperature regimes. Soil mineralogy is an important factor and eight mineralogical classes are recognized (Table 6.70). When the soil has $> 18\% \text{ Fe}_2\text{O}_3$ or gibbsite in the fine earth fraction ($< 2 \text{ mm}$), the mineralogical