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33.2 Histosols

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David Swanson

33.2.1 Introduction

While most soils of the world comprise primarily 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 carbon, depending on the particle size of the mineral component (Soil Survey Staff, 2010). Generally speaking, soils with at least 40 cm of the upper 80 cm that 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 Histosol order; they are now placed in the Histel suborder of Gelisols (Soil Survey Staff, 1998). Following the separation of Histosols and permafrost soils, Histosols occupy about 1% of the global land area while the Histel suborder of Gelisols occupies about 0.8% (Buol et al., 2003). 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 overly 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 is also used on a large scale in electricity generation. In 2009, peat burning was the source of 10% of Ireland's electricity (Public communication, 2009a), and in 2003, about 7% of Finland's electricity was a result of burning peat (Kirkinen et al., 2007). Sweden, Belarus, Latvia, Estonia, and Lithuania also utilize peat for energy; and Canada is currently investigating the possibility of using peat for power production in Ontario and Labrador (Public communication, 2009b, 2009c). In addition to being mined as an energy source, peat is mined for use as a soil amendment in agriculture and horticulture. Moreover, the agricultural value of Histosols has long been recognized. Provided that the water tables 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. 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 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 and water erosion (Lucas, 1982; Parent et al., 1982). Organic soils also have very low strength, are highly compressible (MacFarlane, 1969; MacFarlane and Williams, 1974), and gradually subside by decomposition if drained (Gesch et al., 2007). Furthermore, within the framework of current discussions of global climate change and carbon budgeting, Histosols contribute significantly to the terrestrial carbon pool. On an areal basis, C storage in Histosols is often greater than mineral soils by an order of magnitude (Rabenhorst, 1995).

33.2.2 Distribution

Histosols occur at all latitudes, but are most prevalent in the boreal forest regions of northern North America, Europe, and Asia (Figure 33.2). 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, National Wetlands Working Group, 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 (Figure 33.2). 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 widely used in agriculture in the Great Lakes region, southern Florida, and the Sacramento–San Joaquin delta region of California. In the semiarid 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 Histels in the Gelisol soil order rather than Histosols.

* The National Technical Committee on Hydric Soils (USA) has included "All Histosols except Folists" within the criteria for hydric soils (USDA-NRCS, 1996).

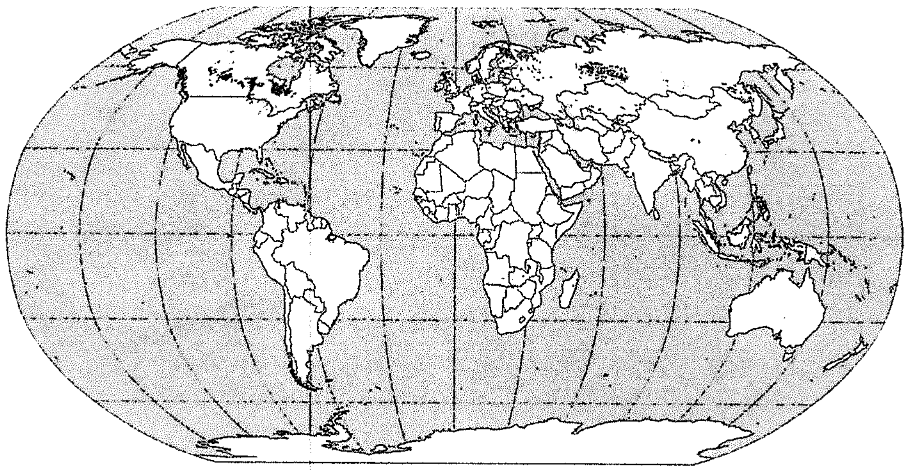

FIGURE 33.2 Worldwide distribution of Histosols. (Courtesy of USDA-NRCS, Soil Survey Division, World Soil Resources, Washington, DC, 2010.)

33.2.3 Formation of Histosols

33.2.3.1 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 that 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 in the following sections. The net accumulation of organic materials occurs when rates of additions (usually as primary plant production) exceed rates 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 relatively thin layers of organic soil materials (O horizons). In Histosols, the rates of decomposition are slowed and organic matter accumulates to the degree that the organic materials amass to a significant depth. In most cases, this is caused by saturation in soils leading to anaerobic conditions, which causes organic matter 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 matter 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 that have contributed to the organic parent material during various stages of accumulation (e.g., Wieder et al., 1994).

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 the acids are 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* (rain-fed, because all nutrients are derived from atmospheric sources). The less acid Histosols that receive base-rich groundwater or runoff are called *fens* and usually described as *minerotrophic* (fed with mineral-derived nutrients). The term *swamp* is often applied to forested systems on both fens and on mineral soil wetlands. 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).

33.2.3.2 Climate

Histosol formation is favored by wet or cold climates. Wetness and cold favor Histosol formation by hindering decomposition of organic matter. The northern Histosol-dominated regions of North America and Eurasia (Figure 33.2) have temperate and boreal climates in which average annual 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 soil organic matter. 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 water table is maintained by tidal waters, leading to coastal marshes and mangrove systems (Anderson, 1983).

Climate affects not only where Histosols occur, but also the chemistry of the resulting soils. Where average annual precipitation exceeds annual 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 may form. In less humid regions, where groundwater is required to maintain saturation of soils, only the less acid, minerotrophic Histosols can form.

Climate change scenarios suggest more frequent extreme events such as heat waves, droughts, and high precipitation events (IPCC, 2007). Increases in drought and heat will likely lead to a higher frequency of fire in the boreal zone resulting in significant losses of C from Histosols. Currently, disturbances, mainly fire, reduce the net C uptake of continental boreal peatlands by about 85% (Turetsky et al., 2002). Predicted increases in fire frequency and intensity could lead to an even greater proportion of C uptake being balanced by fire, and boreal peatlands may actually become a net C source to the atmosphere (Turetsky et al., 2002).

33.2.3.3 Topography

While climate controls the occurrence of Histosols on a regional scale, topography controls where they occur on a given landscape. The classical soil-forming factor of topography is considered broadly here to incorporate relief, geomorphic setting, and hydrologic setting. Histosols occur where the setting facilitates concentration of runoff, discharge of groundwater, or retention of precipitation (Figure 33.3a through 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 interfluves (Sjors, 1963; Walter, 1977), and in extremely humid climates, such as the British Isles and southeastern Alaska, Histosols may also occur on gentle slopes (Moore and Bellamy, 1974; Sjors, 1985; Ward et al., 2007). Histosols of floodplains receive suspended mineral matter during floods, while other Histosols obtain only dissolved material from source waters or inputs of eolian materials.

In lower latitude coastal areas, 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). Figure 33.4 shows a schematic of a cross section of a submerged upland type marsh where organic soil materials are accumulating over what were previously upland soils, forming Histosols. As sea level continues to rise, the margin of the marsh is pushed landward, and the organic materials continue to thicken such that older and deeper Histosols generally exist closer to the open water (Rabenhorst, 1997).

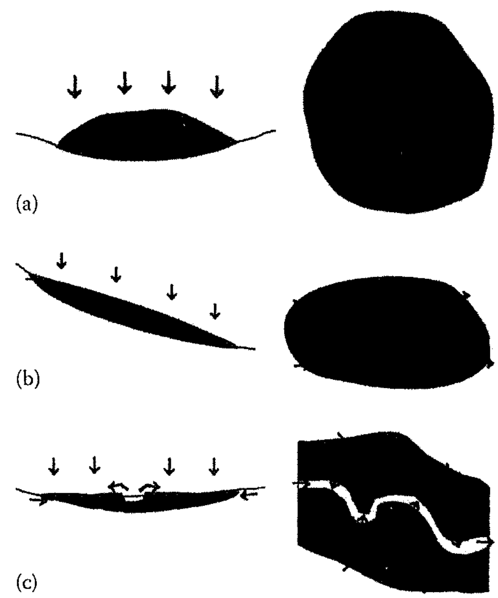

FIGURE 33.3 Settings of peatlands. Peat is shown in gray, and arrows indicate the direction of water movement (a) Bog. The domed surface of the bog precludes input of runoff or groundwater containing bases dissolved from minerals. All water is derived from the atmosphere and evaporates or runs radially 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.

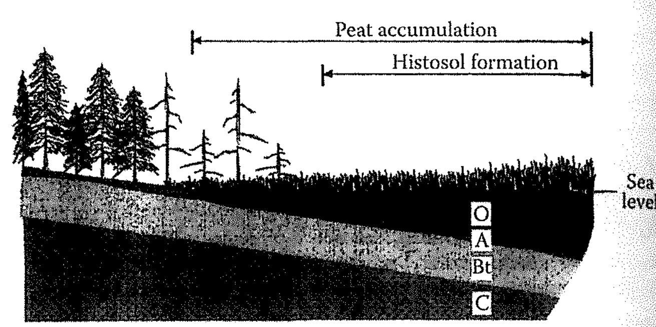

FIGURE 33.4 A schematized cross section of a submerged-upland type marsh showing development of O horizons and a Histosol (probably a Sulfihemist) over what was formerly an upland soil with an argillic (Bt) horizon.

The 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 where the topography was once level or depressional (Figure 33.5). The development of a topographic high due to peat accumulation can prevent base-rich water from moving onto the peatland and thereby facilitate formation of a highly

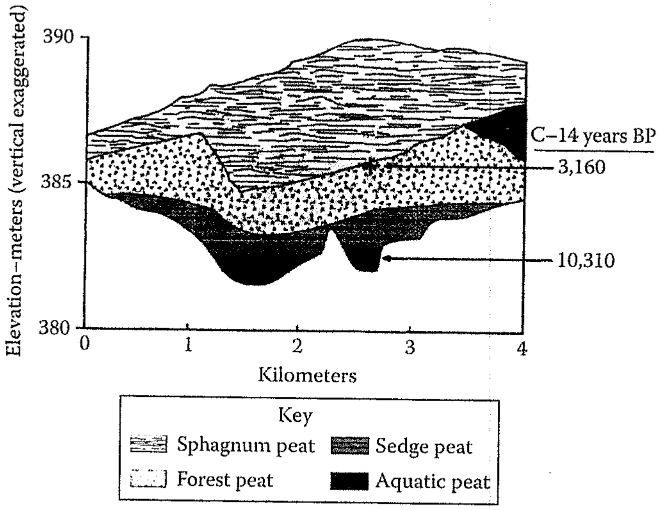

FIGURE 33.5 Stratigraphic cross section of a bog in the Myrtle Lake peatland, northern Minnesota. (Modified from Heinselman, M.L. 1970. Landscape evolution, peatland types and the environment in the Lake Agassiz Peatlands Natural Area, Minnesota. Ecol. Monogr. 40:235–261.)

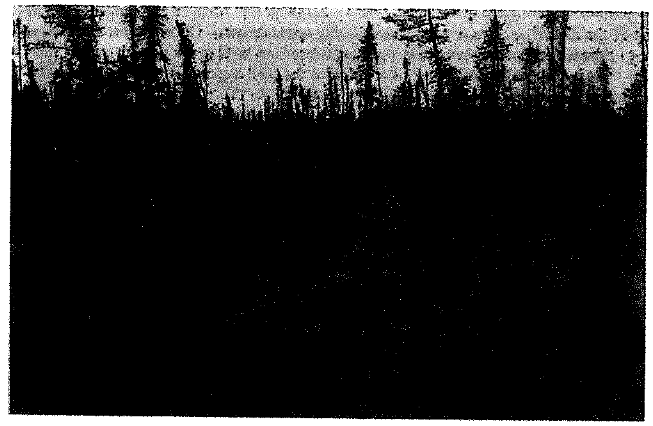

FIGURE 33.6 Typical bog vegetation on a highly acidic Histosol in boreal North America. Trees are stunted black spruce (*Picea mariana*); the largest trees visible are about 4 m tall. Understory plants include ericaceous dwarf shrubs, sedges (*Carex* sp.), and a continuous cover of *Sphagnum* sp. moss. The soil is a Typic Borohemist. Toivola peatland, northeastern Minnesota.

acidic bog. Peat accumulation also produces intriguing micro-topography on some peatlands, notably a pattern of ridges (or *strings*) and pools (Foster et al., 1983; Seppälä and 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*. Histosols may also form by *paludification*, the expansion of wetland onto what was originally drier soils. Both processes operated in the formation of the peatland depicted in Figure 33.5. Folists, unlike other Histosols, generally occur in mountainous regions (Reiger, 1983; Wakeley et al., 1996). Mountainous settings provide the high precipitation and underlying bedrock or fragmental material required for Folist formation.

33.2.3.4 Vegetation

The flora of Histosols varies 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 and this 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 water table of these soils (Figure 33.6). *Sphagnum* mosses cover the ground, along with scattered sedges (*Carex*) and cotton sedges (*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 water table continuously near the surface (fens), most vascular plants are *aerenchymous* grass-like plants (mainly sedges, family Cyperaceae) that can transport oxygen downward to their roots in hollow stems, allowing root metabolism in anaerobic soils (Crawford, 1983). Fens that exhibit an aerobic surface horizon present during at least part of the growing season, allow growth of nonaerenchymous plants, including woody plants (Gill, 1970; Kozlowski, 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/nutrients and high water table; thus, plants are useful indicators of these conditions (Heinselman, 1963, 1970; Sjors, 1963; 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/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 strong control over its physical and chemical properties. Peats are commonly divided into three broad groups on the basis of botanical composition: moss, herbaceous (sedge), and woody peat (Kivinen, 1977). The botanical origin of organic soil material may be determined by examination of plant remains (Birks and Birks, 1980; Janssens, 1983; Levesque 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

TABLE 33.4 Rates of Peat Accretion/Accumulation in Various Organic Soils

Location	Site Characteristics	Peat Accretion–Accumulation Rate (mm $year^{-1}$)	References
Alberta, Canada	Fen, last 50 years	3.2–6.4	Vitt et al. (2009)
Northwest Territories, Canada	Rich fen	0.38	Robinson (2006)
	Poor fen	0.41	
	Bog	0.47	
West Siberia, Russia	Forested bog	0.40–0.70	Peregon et al. (2007)
	Shrub bog	0.25–0.60	
Western Canada	Bog	3.0	Turetsky et al. (2007)
British Columbia, Canada	Sloping, open peatland	0.94–1.7	Asada and Warner (2004)
Southern Sweden	Bog	5.2–5.9	Belyea and Malmer (2004)
West Siberia, Russia	River valley fen	0.84	Borren et al. (2004)
	Raised bog	0.67	
	Forested fen	0.42	
	Bog	0.37	
	Flow-through fen	0.38	
	Flow-through fen	0.35	
	Bog	0.42	
	Bog	1.13	
North America		0.58	Gorham et al. (2003)
North Dakota, USA			
Minnesota, USA		0.41–0.79	
Quebec, Canada		0.59–0.83	
Maine, USA		0.47–0.56	
New Brunswick, Canada		0.49–0.79	
Nova Scotia, Canada		0.40–0.48	
Newfoundland, Canada		0.39–1.05	
Alaska, USA		0.18–0.38	
West Siberia, Russia	Bog	0.57	Glebov et al. (2002)
Southeastern Norway	Bog	3–28	Ohlson and Okland (1998)
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)
Eastern United States	Last 100 years		Wieder et al. (1994)
	Minnesota bog	2.4	
	Pennsylvania fen	1.4	
	Maryland fen	1.9	
	West Virginia bog 1	3.1	
	West Virginia bog 2	2.3	
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 et al. (1983)
Siberian USSR	Palsa province	0.2–0.4	Botch and Masing (1983)
Eurasia		0.52	Zurek (1976) (after Gorham, 1991)
Maine, USA		0.35–0.75	Tolonen et al. (1988) (after Gorham, 1987)
Minnesota, USA	Red Lake, Minnesota	0.85–1.15	Gorham (1987)
Los Angeles, USA	Coastal marsh		Derived from Nyman and DeLaune (1991)
	Fresh	6.5–8.5	
	Brackish	5.9–9.5	
	Saline	7.5–7.6	

TABLE 33.4 (continued) Rates of Peat Accretion/Accumulation in Various Organic Soils

Location	Site Characteristics	Peat Accretion–Accumulation Rate (mm $year^{-1}$)	References
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	1.4–3.2	Hussein (1996)
	Current (^{210}Pb)	0.5–1.1	
	Long term (^{14}C)		
Louisiana, USA	Barataria basin, Coastal marsh	7–13	Hatton et al. (1983)
Massachusetts, USA	Barnstable coastal Marsh	1.1–2.6	Redfield and Rubin (1962)

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 rises. A cross section of a bog in Minnesota shows how a lake was filled in with aquatic peat (i.e., limnic material; Figure 33.4). Then, the peatland expanded onto what was originally dry land of the lakeshore as minerotrophic sedge peat was followed by minerotrophic forest peat as the surface rose and became drier. By about 3000 years ago, 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).

33.2.3.5 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 >3 mm $year^{-1}$, but more typically fall in the range of 0.2–0.7 mm $year^{-1}$ (Table 33.4). 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 (approximately 20,000 years ago), sea level worldwide was approximately 150 m below the present level when large quantities of water were tied up in the glacial ice. Melting of the ice and concurrent ocean warming caused sea level to rise at such a rapid rate (10–20 mm $year^{-1}$) that initially vegetation could not colonize the tidal regions. Approximately 3000–5000 years ago, sea level rise slowed to a more modest pace such 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 mangrove systems generally have been thought to have accreted at approximately the rate of sea level rise. In addition to the eustatic sea level rise, sediment in transgressing coastal regions is subsiding (e.g., along the Atlantic and Gulf coasts of the United States). The combination of rising sea level (presently estimated at 1 mm $year^{-1}$ worldwide) and coastal subsidence can be joined to yield an apparent sea level rise, which is substantially greater. Estimates of peat accretion in coastal areas generally range from 3 to 8 mm $year^{-1}$, which are much higher than in noncoastal regions, with even higher rates reported in rapidly subsiding areas (Table 33.4). 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 (Kearney et al., 2002).

33.2.4 Morphological Properties of Histosols

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, 2010).

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 33.5) and are defined by the portion of the soil material, which retains discernable plant fibers after rubbing and by the color of a Na pyrophosphate extracting solution (Soil Survey Staff, 2010). Other methodologies and rating scales have been developed and utilized for evaluating degree of decomposition; the one most broadly used 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 of sapric organic soil material are usually

TABLE 33.5 Defining Morphological Criteria for Histosol Organic Horizons

Horizon Designation	Type of Material	Common Descriptor	Volumetric Rubbed Fiber (RF) Content	Color of Pyrophosphate Extract (Value/Chroma)
Oi	Fibric	Peat	RF > 3/4 or RF > 2/5	7/1, 7/2, 8/1, 8/2, 8/3
Oe	Hemic	Mucky peat	2/5 > RF > 1/6	Does not otherwise qualify for either fibric or sapric materials
Oa	Sapric	Muck	RF < 1/6	Below or right of line drawn to exclude blocks 5/1, 6/2, 7/3

Source: Soil Survey Staff. 2010 Keys to soil taxonomy. 11th edn. USDA-NRCS, U.S. Government Printing Office, Washington, DC.

nearly black. Fibric peat is lighter colored and often reddish in hue, while hemic peat has an intermediate color (Table 33.6). 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 produces 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 man-made 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 of 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; Sheng et al., 2004; Kettridge et al., 2008).

The morphology of Folists differs from that of other Histosols in that the peat layers are thinner and underlain by fragmental material or bedrock (Witty and Arnold, 1970; Everett, 1971; Lewis and Lavkulich, 1972). Folists drain freely and as a result are less saturated than other Histosols (not saturated for >30 cumulative days per year; Soil Survey Staff, 2010).

TABLE 33.6 Color of Organic Soil Material in Relation to Degree of Decomposition[a]

Degree of Decomposition	Median Munsell Color (Hue Value/Chroma)	*N*
Sapric	10 YR (2/1)	69
Hemic	10 YR (3/2)	49
Fibric	7.5 YR (3/2)	18

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

33.2.5 Micromorphology of Histosols

Micromorphological observations can provide a direct examination of the structural integrity of plant fragments and components in organic soil materials. Levesque and Dinel (1982) and Fox (1985) have 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 (Figure 33.7). 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, so 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

FIGURE 33.7 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 discernable plant structures and cell components intermixed with decomposed organic material; frame length = 5 mm; PPL.

peat profile in Ireland was compared with those which had been drained for between 10 and 100 years, 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 materials directly corresponded to an increase in density of the material.

In another study, Lee and Manoch (1974) concluded that 50 years 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 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, 2011) 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 accumulate within the lower horizons of the peat, at the peat–mineral soil contact, or within the underlying mineral soil material.

33.2.6 Classification of Histosols

The definition of *organic materials* for saturated soils requires a minimum of 12% OC if there is no clay, and a minimum of 18% OC if the soil contains 60% or more clay, with a sliding scale for intermediate textures. Those soils which are not saturated must contain at least 20% OC to be considered organic soil materials (Soil Survey Staff, 1998, 2011). For a soil to be classified as a Histosol, at least 40 cm of the upper 80 cm must comprise organic materials, and it must not have permafrost within 1 m of the surface. However, if the organic materials are especially low in density ($<0.1\ g\ cm^{-1}$), 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 33.7. Basically, organic soils that 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 Saprists. Within the United States, some 314 soil series have been established for Histosols. Table 33.8 shows the distribution of those 314 series among the classes of the various

TABLE 33.7 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, humilluvic materials)	Underlying materials Special materials contained (Limnic) Intergrades to other great groups (Cryic and Sphagnic)	Reaction (pH in 0.01 M $CaCl_2$) Temperature regime Soil depth (only used if <50 cm deep)

TABLE 33.8 Number of Soil Series in the United States That Are Classified into Particular Taxonomic Groups

Suborders		Great Groups		Subgroups	
Formative Element	Number of Series	Formative Element	Number of Series	Formative Element	Number of Series
Fibrists	26	Cryo	58	Fluvaquentic	14
Folists	35	Haplo	198	Limnic	19
Hemists	82	Sphagno	7	Lithic	37
Saprists	171	Sulfi	17	Terric	124
		Udi	19	Typic	102
		Usti	7	Hemic	7
		Others	8	Others	11
Total	314	Total	314	Total	314

taxonomic categories. 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 (pH in 0.01 M $CaCl_2$), temperature regime, and soil depth (only used if <50 cm deep).

33.2.7 Biological and Chemical Properties

Organic carbon contents are generally higher and nitrogen contents lower in less-decomposed peats than highly decomposed peats (Table 33.9). As a result, C:N ratios are generally higher for less-decomposed peats (Table 33.9; Lee et al., 1988). The ash content (mineral component) is also higher for more highly decomposed peats (Table 33.9; 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 supply nitrogen in excess of crop requirements without fertilization. However, nutrients derived from minerals, such as phosphorus, potassium, 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 cation exchange capacity (CEC), and have lower bulk density than woody peats; sedge peats typically have intermediate properties (Farnham and Finney, 1965; Lévesque et al., 1980).

33.2.7.1 Soil Carbon

Interest in the global C cycle has focused attention on the high proportion (3/4) of terrestrial C stored in soils (Lal et al., 1995). As a group, wetland soils maintain a disproportionately high level of soil carbon, and Histosols, which are composed largely of organic matter, clearly store the largest quantities of soil carbon. Although peatlands only occupy approximately 4% of the global land surface, they store about 30% of the global soil C (Lavoie et al., 2005). While typical agricultural soils may contain between 2 and 10 kg C m^{-2}, reported values for Histosols typically are an order of magnitude greater, with some values >200 kg C m^{-2} (Table 33.10). 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 although some studies indicate that climate change either through increased frequency and intensity of fire (Turetsky et al., 2002) or through elevated decomposition as a result of rising temperatures (Billett et al., 2004) is possibly switching some peatlands to sources of C to the atmosphere. In addition, if Histosols are drained or in some other way exposed to an aerobic environment, they may begin to oxidize and yield large quantities of C to the atmosphere (e.g., Nykanen et al., 1997), although other studies indicate that drainage can increase C sequestration as a result of greater aboveground and belowground productivity (e.g., Minkkinen and Laine, 1998).

Most of the discussion of possible global warming and greenhouse gas emission has focused on rising levels of CO_2 in the atmosphere. Methane (CH_4), however, is 32 times more efficient than CO_2 in trapping infrared radiation. Because many Histosols are strongly reducing (low Eh), they represent an ideal environment for the formation of CH_4. In systems where SO_4^{2-} is more abundant in the soil solution, such as in coastal or estuarine environments, sulfate reduction is favored over methanogenesis and methane production may be more limited (Bartlett et al., 1987; Widdell, 1988; Dise and Verry, 2001). However, in many freshwater or inland areas, Histosols may be the locus of significant methane emission to the atmosphere (Table 33.11). Minerotrophic fens have higher methane

TABLE 33.9 Chemical Properties of Organic Soil Material as Related to Degree of Decomposition: Mean (Standard Deviation, N)[a]

Property	Sapric	Hemic	Fibric	AOV Probability[b]
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, cmol kg^{-1}	101 (44, 129)	88 (41, 61)	83 (33, 28)	0.046
CEC, cmol 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, mol Al mol^{-1} TEA[c]	0.074 (0.076, 49)	0.022 (0.024, 23)	0.038 (0.050, 16)	0.004

[a] Data (except for ash) is 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).

[b] F-test probability from one-way analysis of variance (AOV) between fibric, hemic, and sapric peats.

[c] KCl extractable Al divided by total NH_4 acetate extractable acidity.

TABLE 33.10 Carbon Storage Values for Organic Soils

Site Characteristics	Location	Carbon Accumulation Rate (kg m^{-2} year^{-1})	Quantity of Stored C (kg m^{-2})	References
10 year averages	West Siberia			Golovatskaya and Dyukarev (2009)
Pine peatland		0.021		
Stunted pine peatland		0.11		
Sedge fen		0.10		
Fen, last 50 years	Alberta, Canada	0.14–0.25		Vitt et al. (2009)
Various peatland types	Alberta, Canada		53–165, mean = 129	Beilman et al. (2008)
Bog, last 100 years	Western Canada	0.09		Turetsky et al. (2007)
Fen 1	Saskatchewan, Canada		29–210	Robinson (2006)
Fen 2			20–120	
Rich fen	Northwest territories, Canada	0.014		Robinson (2006)
Poor fen		0.018		
Bog		0.019		
Rich fen	Western Canada	0.025		Yu (2006)
Tropical peatlands	Micronesia	0.3		Chimner and Ewel (2005)
Sloping, open peatland	British Columbia, Canada	0.007–0.039		Asada and Warner (2004)
Bog	Southern Sweden	0.060–0.072		Belyea and Malmer (2004)
River valley fen	West Siberia	0.069		Borren et al. (2004)
Raised bog		0.033		
Forested fen		0.034		
Bog		0.027		
Flow-through fen		0.021		
Flow-through fen		0.020		
Bog		0.019		
Bog		0.040		
Variety peatland types	West Siberia		<30 to >300	Sheng et al. (2004)
Various peatland types	North Dakota, USA	0.038		Gorham et al. (2003)
	Minnesota, USA	0.021–0.038		
	Quebec, Canada	0.022–0.028		
	Maine, USA	0.022–0.030		
	New Brunswick, Canada	0.017–0.019		
	Nova Scotia, Canada	0.019–0.041		
	Newfoundland, Canada	0.008–0.016		
	Alaska, USA			
Raised bog	Southern Sweden			Malmer and Wallen (1999)
Hummock		0.05–0.18		
Hollow		0.03–0.15		
Last 100 years	Eastern USA			Wieder et al. (1994)
Minnesota bog		0.16		
Pennsylvania fen		0.14		
Maryland fen		0.15		
West Virginia bog 1		0.15		
West Virginia bog 2		0.18		
Coastal marsh	Los Angeles, USA			Derived from Nyman and DeLaune (1991)
Fresh		0.17–0.22		
Brackish		0.17–0.27		
Saline		0.21–0.22		
Coastal marsh Brackish	Chesapeake Bay, MD	0.12–0.42		Derived from Kearney and Stevenson (1991)
Coastal marsh Brackish	Chesapeake Bay, MD		59 (range 18–166)	Derived from Griffin and Rabenhorst (1989)
Barataria Basin, coastal marsh	Louisiana, USA	0.18–0.30		Smith et al. (1983)
Coastal marshes	Atlantic and Gulf Coasts, USA		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)

TABLE 33.11 Reported Fluxes of CO_2 and CH_4 Emissions from Histosols

			CO_2 Emission Rate ($mmol\ m^{-2}\ day^{-1}$)		CH_4 Emission Rate ($mmol\ m^{-2}\ day^{-1}$)		
Location	Site Details	Notes	Mean	Range	Mean	Range	References
West Siberia	10 year averages	Growing season					Golovatskaya and Dyukarev (2009)
	Pine peatland		45.0		0.030		
	Stunted pine peatland		28.5		0.045		
	Sedge fen		29.9		3.30		
Quebec, Canada	Poor fen—control	Growing season		42–250		0.95–3.55	Strack and Waddington (2007)
	Poor fen—w/water table drawdown			46–242		0.63–6.40	
Western Canada	Bog	Transplant experiment	55		0.4		Turetsky et al. (2007)
Northern England	Acidic blanket peat	Growing season				0.15–4.2 (range for all treatments)	Ward et al. (2007)
	Control			5.4–327			
	Burned			10.9–491			
	Grazed			5.4–436			
Micronesia	Forested peatland	Annual	198				Chimner and Ewel (2004)
	Cultivated for taro		110				
New York, USA	Conifer/maple peatland	Annual	103	8.6–216	0.05	−0.17–0.69	Coles and Yavitt (2004)
Ontario, Canada	Mesocosms—controlled temperature, water table, and humidity	Net production reported		6.1–602		−0.25–1.1	Blodau and Moore (2003)
Quebec, Canada	Poor fen—control	Growing season			8.9		Strack et al. (2004)
	Poor fen—w/water table drawdown				4.0		
Minnesota, USA	Poor fen—control	Growing season			15.1	6.2–31.9	Dise and Verry (2001)
	Ammonium nitrate added				16.0	7.5–30.0	
	Ammonium sulfate added				10.2	5.0–15.6	
Minnesota, USA	Bog	Mesocosms		275–2500		1.3–41.3	Updegraff et al. (2001)
	Fen	Growing season					
Quebec, Canada	Gatineau Park			1.97–7.24		1.15–2.18	Buttler et al. (1994)
Wales	Peat monoliths	Laboratory study	14.7	9.6–21.0	14.4	4.7–34.4	Freeman et al. (1993)
Finland	Natural fen	Annual	11.3		3.49		Nykanen et al. (1995)
	Drained fen		30.8		0.03		
Finland	Ombrotrophic	12 C	88.2	42.5–141.3			Silvola et al. (1996)
Alaska, USA		Summer measurements			9.2		After Gorham (1991), after Crill et al. (1988)
Boreal Canada	Swamp (n = 20)	Annual averages			0.21		Derived from Moore and Roulet (1995)
	Fen (n = 6)				0.57		
	Bog (n = 13)				0.39		
Canada (lab study)	Bog flooded	19–23 C	0.005		0.012		Derived from Moore and Knowles (1989)
	Bog drained		0.19		0.006		
	Fen flooded		0.009		0.58		
	Fen drained		0.14		0.025		
Alaska	Moist tundra	August			0.3		Derived from Sebacher et al. (1986)
	Waterlogged tundra				7.4		
	Wet meadows				2.5		
	Alpine fen				18		
Northern Sweden	Ombrotrophic bog	Summer					Svensson and Rosswall (1984)
	Hummocks		10.12		0.05		
	Between hummocks		14.92		0.14		
	Shallow depressions		11.61		0.77		
	Deeper depressions		12.45		1.21		
	Ombrominerotrophic		12.49		2.73		
	Minerotrophic fen		11.48		16.89		

TABLE 33.11 (continued) Reported Fluxes of CO_2 and CH_4 Emissions from Histosols

			CO_2 Emission Rate (mmol m^{-2} day^{-1})		CH_4 Emission Rate (mmol m^{-2} day^{-1})		
Location	Site Details	Notes	Mean	Range	Mean	Range	References
Minnesota	Bog	Sampled during August			8.2	1.2–29.2	After Harriss et al. (1985)
	Fen				0.25	0.19–0.31	
Georgia, USA		During midsummer			6.6		After Gorham (1991), after Crill et al. (1988)
West Virginia, USA	Mountain bog	During midsummer			11.7		After Gorham (1991), after Crill et al. (1988)
Minnesota, USA	Forest bog	Summer measurements			3.6	0.5–33	After Crill et al. (1988), after Mitsch and Wu (1995)
	Forest fen				6.7	3.2–12	
	Open bog				14	0.9–41	
	Neutral fen				15	7.1–33	
	Acid fen				4.8		
Minnesota, USA	Open poor fen	Winter measurements			3.0		Dise (1992)
	Open bog				0.7		
	Forest bog hollow				0.8		
	Hummock				0.3		
Virginia	Coastal marsh	Summer					Derived from Bartlett et al. (1985)
	York river	Creek Bank			0.45	0.13–0.82	
	Chesapeake Bay estuary	Short Spartina			0.29	0.05–0.87	
		High marsh			0.09	0–0.36	
Virginia	Coastal marsh	Summer					Derived from Bartlett et al. (1987)
	Tidal Creek in Chesapeake Bay estuary	Low salinity			11	5–16	
		Moderate salinity			7.5	4–11	
		High salinity			2	0.5–2.5	
West Virginia	Appalachian bog		127	75–250	17.0	0–53	Wieder et al. (1990)
Maryland	Appalachian bog		152	100–250	4.4	0–12	Wieder et al. (1990)
Louisiana, USA	Barataria basin and Coastal marsh	Annual averages		41–141			Smith et al. (1983)
Florida, USA		During midsummer			6.0		After Gorham (1991), after Crill et al. (1988)
Malaysia	Ombrotrophic bog		170				Wosten et al. (1997)
Malaysia	Drained and cultivated peatland			139–727			Murayama and Bakar (1996)

emissions than ombrotrophic bogs (Dise, 1993). Methane emission from Histosols seems to be directly related to the location of the water table, with greater generation when soils are saturated to or above the surface (Updegraff et al., 2001; Strack et al., 2004). Greater methane emissions occur when soils are saturated. If aerobic zone exists in the profile, soil microbes will utilize the methane as it passes through on its way to the soil surface. Carbon dioxide emissions, which are produced by oxidation of soil organic matter, have been found to be both greater (Nykanen et al., 1997) and unchanged (Updegraff et al., 2001; Strack and Waddington, 2007) following the lowering of water tables.

33.2.7.2 Sulfides

The biogeochemical environment in which Histosols form can also be conducive to the formation of sulfides. The occurrence of iron sulfides can, in some circumstances, lead to the generation of extreme acidity and acid sulfate soils (Van Breemen, 1982). Sulfate reduction generally requires the presence of organic matter, which serves as an energy source, low redox potentials, sulfate that functions as an electron acceptor, and sulfate reducing bacteria (Rickard, 1973). If sulfide is formed in the presence of a reactive iron source, then such minerals as pyrite (FeS_2) can form. The C source and anaerobic conditions are almost always present in Histosols, but sulfate (SO_4^{2-}) levels may vary dramatically among environments. Many inland Histosols receive SO_4^{2-} 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 S forms (Novak and Wieder, 1992). Some inland peats have developed acid sulfate conditions, although usually they are associated with deposits of coprogeneous earth (Lucas, 1982). Sulfate reduction is common in coastal Histosols, which contain an abundance of SO_4^{2-} from sea water. Extensive areas of these Sulfihemist soils have been identified along the Atlantic coast of the United States. The distribution of pyrite within coastal Histosols can be highly variable and is often related to microsite differences in the availability of either reactive iron or sulfide (Rabenhorst and Haering, 1989).

33.2.7.3 Acidity and Base Saturation

In bog Histosols, 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 in $CaCl_2$ is typically 3–4 (Figures 33.8 and 33.9; Gorham et al., 1985; Urban, 1987). Minerotrophic (fen) peats have higher base saturation, and pH in water is generally above 5 ($CaCl_2$ pH above 4.5). The pH of minerotrophic peats is buffered by cation exchange within the soil (Bloom et al., 1983) or by carbonates if they are present. The aluminum ion, which is derived from silicate minerals, comprises a small proportion of the total acidity in most Histosols because of its low solubility at pH > 5.5 (Table 33.9). Less-decomposed peats generally have lower pH than more highly decomposed peats (Table 33.9; Lee et al., 1988).

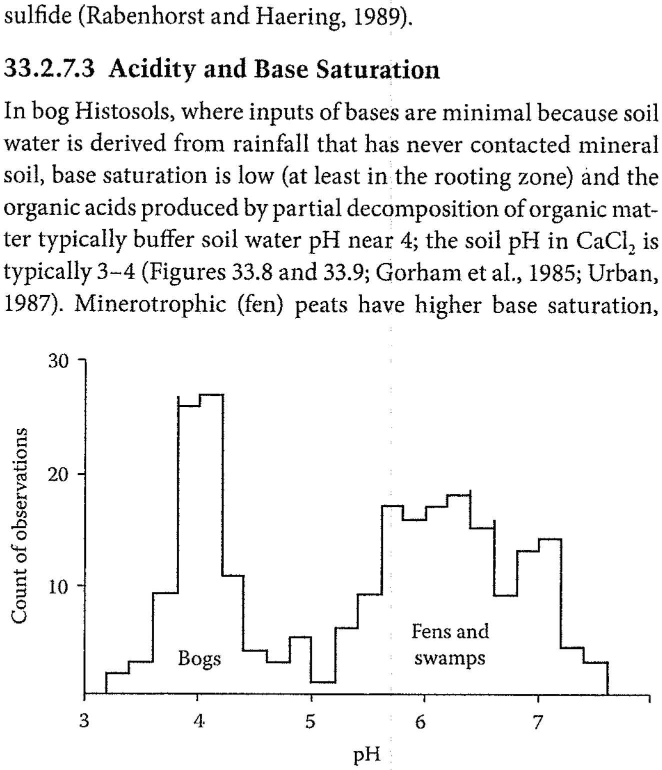

FIGURE 33.8 Frequency distribution of pH in peatland surface water for 232 sites in Minnesota. Samples with surface water pH near 4 are from bogs and those with pH above 5 are from fens (unforested) and swamps (forested). (From Swanson, D.K., and D.F. Grigal. 1989. Vegetation indicators of organic soil properties in Minnesota. Soil Sci. Soc. Am. J. 53:491–495.)

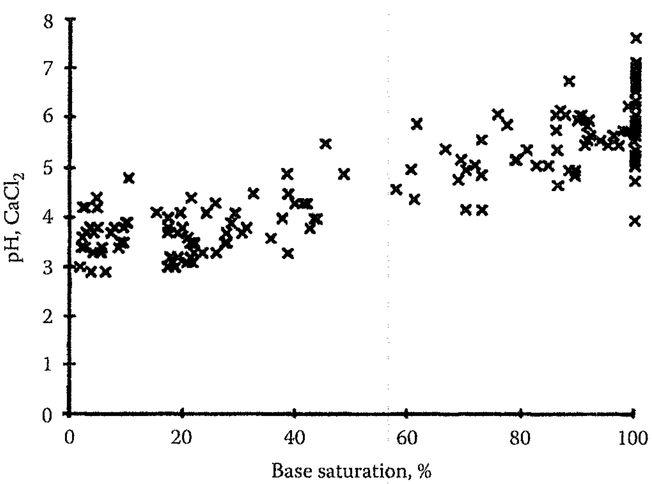

FIGURE 33.9 Relationship between base saturation and pH of organic soil horizons of Histosols. Peat samples with pH in $CaCl_2$ of less than 4.5 (presumably deposited in bogs) generally have less than 50% base saturation. Peat samples with pH greater than 4.5 (deposited in fens and swamps) typically 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.

The distinction between euic and dysic reaction classes at the family level in *Soil Taxonomy* separates the highly acid, bog peats (dysic) from less acid fen peats (euic) (Farnham and Finney, 1965). Because a pH (in $CaCl_2$) greater than 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 a 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 pyrite have the potential to develop extreme acidity. Under saturated and anaerobic conditions, such sulfide bearing soils have circumneutral pH. If it is drained, dredged, or in some other way exposed to oxidizing conditions, the soil will undergo acid sulfate weathering, a process that is the result of sulfide mineral oxidation and the concomitant production of sulfuric acid (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 $FeSO_4$.

33.2.7.4 Cation Exchange Capacity

The CEC of Histosols is quite high as a result of the high cation exchange of organic matter (Table 33.9). The CEC is higher for more highly decomposed peats than fibric peats (Table 33.9; Lévesque et al., 1980). The difference between the CEC of sapric and fibric peats is even more dramatic 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 33.9). 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, near 20 $cmol_c\ L^{-1}$, is comparable to that of most mineral soils (5–20 $cmol_c\ L^{-1}$) assuming mineral soil bulk density of 1.0–1.5 kg L^{-1} (Holmgren et al., 1993). Even when expressed per unit volume, the average CEC of sapric and hemic peats (Table 33.9) is much higher than the 5–20 $cmol_c\ L^{-1}$ of most mineral soils.

33.2.8 Physical Properties

33.2.8.1 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 0.8 g cm^{-3} (Table 33.12). Bulk density is related to the degree of decomposition

TABLE 33.12 Physical and Hydraulic Properties of Organic Soils

Location	Site Characteristics	Bulk Density (g cc^{-1})	Hydraulic Conductivity (10^{-5} cm s^{-1})	References
Alberta Canada	Rich fens	0.04–0.12		Vitt et al. (2009)
Western Canada	Bog	0.056		Turetsky et al. (2007)
Manitoba, Saskatchewan and Alberta, Canada	Various peatland types	0.07–0.26		Bauer et al. (2006)
Northwest Territories, Canada	Rich fen	0.03–0.14		Robinson (2006)
	Bog	0.08–0.20		
Western Canada	Rich fen	0.18		Yu (2006)
Micronesia	Tropical peatlands	0.11–0.13		Chimner and Ewel (2005)
West Siberia	Various peatland types	0.05–0.41		Sheng et al. (2004)
England	Blanket peats (fens)	0.15–0.27	0.01–1.04, mean = 0.24	Holden and Burt (2003)
New Zealand	Bog	0.06		Schipper and McLeod (2002)
Poland	Various peatland types	0.07–0.60		Bogacz (2000)
Finland	Drained and harvested fens		0.4–60	Klove (2000)
Southern Sweden	Bog			Malmer and Wallen (1999)
	Lichen hummock	0.18		
	Sphagnum hummock	0.27		
	Sphagnum lawn	0.27		
Finland	Various peatland types	0.12–0.16		Minkkinen and Laine (1998)
Southeastern Norway	Bog	0.2–0.8		Ohlson and Okland (1998)
Northern Minnesota	Undecomposed		3,810–15,000, mean = 8,650	Boelter (1965)
	Partially decomposed		13.9–132, mean = 63	
	Decomposed		0.9–15 mean = 5.1	
Wyoming	Mountain bog 46 cm	0.16–0.22	0.0277	Sturges (1968)
	91 cm		0.0185	
Ottawa and St. Lawrence River Valleys, Canada	Swamp and bog			Mathur and Levesque (1985)
	0–60 cm		624	
	0–100		366	
	0–125		269	
Northern Minnesota	Lost river peatland	0.06–0.14		Chason and Siegel (1986)
	Bog		25–560	
	Fen margin		150–2,600	
	Spring fen		67–1,600	
Quebec, Canada	Gatineau Park	0.03–0.10		Buttler et al. (1994)
Eastern New Brunswick, Canada	Sphagnum peat from raised bogs			Korpijaako and Radforth (1972)
	Van post scale 1–2	0.02–0.08	90–175	
	3–4	0.03–0.10	6.9–56	
	5–6	0.06–0.11	1.4–17	
	7–8	0.09–0.13	0.14–2.8	
Wisconsin, USA	Fibric	0.13		Lee et al. (1988)
	Femic	0.17		
	Sapric	0.20, 0.24		
Minnesota	Bog			Gafni and Brooks (1990)
	0–10 cm Von Post 1		23,495	
	10–20 cm Von Post 2–3		7,697	
	20–30 cm Von Post 4–5		5,498	
	30–40 cm Von Post 5–6		799	
	40–50 cm Von Post 5–7		995	
	Fen			
	0–10 cm		31,597	
	10–20 cm		5,000	
	20–30 cm		1,400	

of the organic materials with bulk density generally increasing as the materials become more highly decomposed (Figure 33.10). 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. Figure 33.11 illustrates the relationship between the mineral content (roughly the difference remaining from twice the OC content) and the bulk density for Oe and Oa horizons from Sulfihemists along the Atlantic Coast (Griffin and Rabenhorst, 1989).

33.2.8.2 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 (Figure 33.12; Boelter, 1969). The high porosity and low bulk density of organic

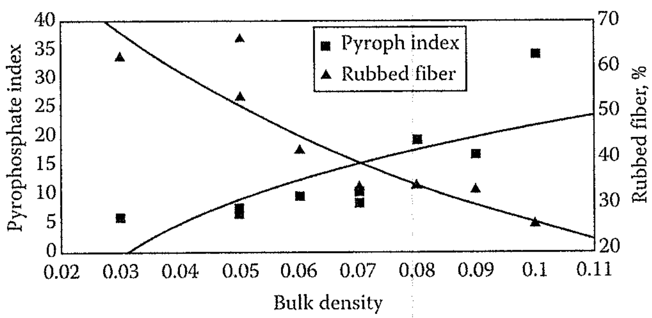

FIGURE 33.10 Relationship between bulk density of organic materials and indices of decomposition such as rubbed fiber content or pyrophosphate index. (Derived from Buttler, A., H. Dinel, and P.E.M. Levesque. 1994. Effects of physical, chemical and botanical characteristics of peat on carbon gas fluxes. Soil Sci. 158:365–374.)

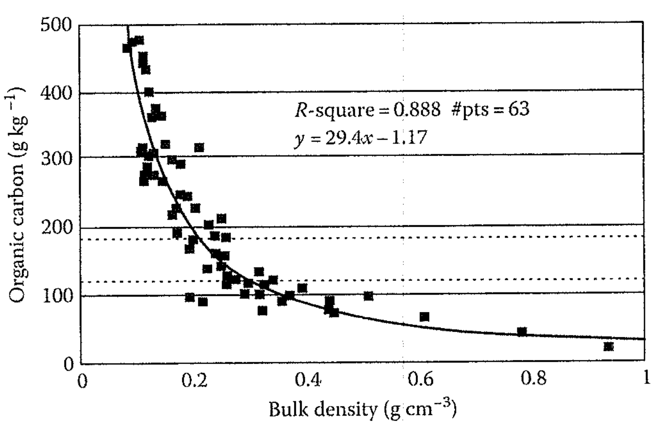

FIGURE 33.11 The effect of mineral content (roughly the difference remaining from 2x the OC content) on the bulk density of Oe and Oa horizons from Sulfihemists along the Atlantic Coast. Dashed lines represent 12% and 18% OC, which is necessary for soil materials to be considered organic. (Derived from Griffin, T.M., and M.C. Rabenhorst. 1989. Processes and rates of pedogenesis in some Maryland tidal marsh soils. Soil Sci. Soc. Am. J. 53:862–870.)

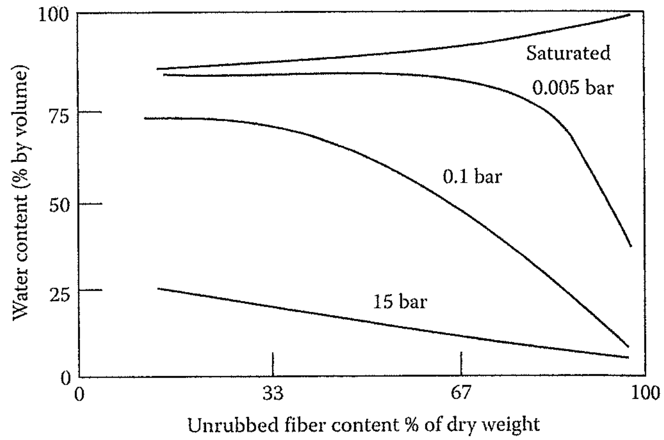

FIGURE 33.12 Volumetric soil water content as a function of unrubbed fiber content at various moisture tensions. (After Boelter, D.H. 1969. Physical properties of peats as related to degree of decomposition. Soil Sci. Soc. Am. J. 33:606–609.)

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 33.12). The hydraulic conductivity of sapric and some hemic peats is quite low, comparable to that of clay (i.e., 10^{-5} m s^{-1} or less). 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).

Water retention in Histosols is also closely related to the degree of decomposition (Figure 33.12). 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 minus that at 1.5 MPa suction; approximates plant-available soil water capacity) of most Histosols is very high, approaching 50% in some peats and exceeding 25% in all others except the least-decomposed peats (Figure 33.12).

33.2.9 Utilization and Management of Histosols

33.2.9.1 Interest in Preservation as 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, floodwater 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 archaeological archives has also been documented (Godwin, 1981). Throughout the years, the value society has placed on these functions has been minimal, and the Histosols of peatlands and coastal marshes have been exploited and extensive areas have been destroyed. More recently, people have recognized that the functions which Histosols perform in a natural setting have significant benefit for society. Thus, legislation has been passed in the United States and elsewhere to preserve Histosols and other wetlands.

33.2.9.2 Histosols as Agricultural Resources

As was mentioned earlier, Histosols are utilized as important agricultural soils in many areas, so long as water tables can be effectively managed. For example, in Japan, there are over 70,000 ha of rice that is grown on peatlands. Peatlands in a number of tropical countries such as Cuba, Guyana, Malaya, and Indonesia have been reclaimed from mangroves and are used for the production of sugarcane (Moore and Bellamy, 1974). Within the United States, there has also been extensive growth of vegetables on peatlands in the Northern United States (such as Michigan), and there has also been extensive agricultural development in the Everglades region of Florida. The agricultural use of Histosols presents some special challenges regarding nutrient management and fertility, but the most significant problem is the high water tables requiring drainage.

33.2.9.3 Impacts of Drainage

Where Histosols have been converted for higher intensity land uses, such as in agriculture, horticulture, or silviculture, they are almost always drained to lower the water table. 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. There is also ongoing consolidation and soil alteration caused by the enhanced decomposition of the organic materials following the shift from an anaerobic to an oxidizing regime (Stephens and Speir, 1969; Minkkinen and Laine, 1998). Reported rates of long-term peat subsidence range up to 10 cm $year^{-1}$ but most reports are in the range of 2–5 cm $year^{-1}$ (Table 33.13). This consolidation is accompanied by changes in the physical properties of the peat, including higher bulk densities and lower moisture contents.

33.2.9.4 Histosols as Energy Resources

Peat is mined mainly in northern Europe and used as fuel (Table 33.14). In addition, there is extensive mining and export of peat as a horticultural amendment. Important examples of the latter uses include ingredients for potting soils, and mixed fertilizers, components of mushroom beds, as a seed inoculant, as a material for packing of flowers and other plants as well as a general soil amendment to increase organic matter content in gardens, golf courses, etc. Approximately 635,000 metric tons of peat are utilized annually for these types of uses in the United States (Public communication, 2009c).

33.2.9.5 Engineering Properties

Histosols are notorious for their low strength and great compressibility, which make them poor foundation materials for

TABLE 33.13 Reported Rates of Subsidence after Drainage of Organic Soils

Location	Site Characteristics	Subsidence Rate (cm $year^{-1}$)	Length of Record (year)	References
New Zealand	Bog	3.4	40	Shipper and McLeod (2002)
Finland	Various peatland types	0.4	60	Minkkinen and Laine (1998)
Florida, USA	Everglades, muck, and peat	3.2	41	Thomas (1965)
Hunts, England	Holme marsh	3.4	103	Nickolson (1951)
California, USA	Sacramento–San Joaquin delta, muck, and peat 1–3 m deep	6.4–9.8	26	Weir (1950)
Indiana, USA		3.0	30+	Ellis and Morris (1945)
Northern Indiana, USA	Muck	1.1–3.0 dependent on WT level	6	Jongedyk et al. (1950)
Michigan, USA		0–3.6	5	Davis and Engberg (1955) (after Thomas, 1965)
Southern Ontario, Canada	Holland marsh and deep loose muck	3.3	19	Mirza and Irwin (1964)
Minnesota, USA		5.1	3	Row (1940)
Florida, USA	Everglades	3	55	Stephens and Speir (1969)
Florida, USA	Everglades	2.3–1.8	20	Shih et al. (1981)
California USA	Sacramento–San Joaquin delta	2.3	78	Rojstaczer and Deverel (1995)
Quebec, Canada		2.1	38	Parent et al. (1982)
Malaysia	Ombrotrophic bog	2	21	Wosten et al. (1997)

TABLE 33.14 Peat Mining by Country, 2007 (in Thousands of Metric Tons)[a]

Country	Horticultural Use	Fuel Use	Total
Argentina	15	0	15
Australia	nd	nd	7
Belarus	100	2400	2500
Burundi	0	10	10
Canada	1250	0	1250
Denmark	300	0	300
Estonia	1300	600	1900
Finland	900	8200	9100
France	200	0	200
Germany	120	0	120
Ireland	500	3800	4300
Latvia	nd[b]	nd	1000
Lithuania	nd	nd	307
Moldova	0	475	475
New Zealand	27	0	27
Norway	30	0	30
Poland	500	0	500
Russia	nd	nd	1300
Spain	nd	nd	60
Sweden	380	900	1280
Ukraine	nd	nd	395
United States	635	0	635

[a] Data from Public communication (2009c).
[b] No data.

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, 1980). 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).

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33.3 Andisols

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33.3.1 Introduction

Andisols are soils that typically form in loose volcanic ejecta (tephra) such as volcanic ash, cinders, or pumice. They are characterized by andic properties that include physical, chemical, and mineralogical properties that are fundamentally different from those of soils of other orders. These differences resulted in a proposal to recognize these soils at the highest level in the USDA soil classification system (Smith, 1978). In 1990, Andisols were added to *Soil Taxonomy* as the 11th soil order (Soil Survey Staff, 1990; Parfitt and Clayden, 1991). A very similar taxonomic grouping, Andosols, is 1 of the 32 soil reference groups recognized in the World Reference Base for Soil Resources (IUSS Working Group WRB, 2006). Andisols (and Andosols) are classified on the basis of selected chemical, physical, and mineralogical properties acquired through weathering and not on parent material alone. Both soil names relate to two Japanese words, *anshokudo* meaning "dark colored soil" (*an*, dark; *shoku*, color or tint; *do*, soil) and *ando* meaning "dark soil." *Ando* was adopted into western soil science literature in 1947 (Simonson, 1979).

The central concept of Andisols is one of deep soils commonly with depositional stratification developing mainly from ash, pumice, cinders (scoria), or other explosively erupted, fragmental volcanic material (referred to collectively as tephra) and volcaniclastic or reworked materials. Andisols occur much less commonly on lavas. Unlike many other soils, Andisol profiles commonly undergo "upbuilding pedogenesis" as younger tephra materials are deposited on top of older ones. The resulting profile character is determined by the interplay between the rate at which tephras are added to the land surface and classical "top-down" processes that form soil horizons. Understanding Andisol genesis in many instances thus requires a stratigraphic approach combined with an appreciation of buried soil horizons and polygenesis.

The coarser fractions of Andisols are often dominated by volcanic glass. This glass weathers relatively quickly to yield a fine colloidal or nanoscale fraction (1–100 nm) dominated by short-range order materials composed of "active" Al, Si, Fe, and organic matter, especially humus. Previously described erroneously as "amorphous," short-range order materials comprise extremely tiny but structured nanominerals, the main ones being allophane and ferrihydrite (Hochella, 2008). A useful collective descriptor for them is "nanocrystalline" (Michel et al., 2007). Another colloidal constituent, imogolite, comprises long filamental tubes and therefore has both short- and long-range order (Churchman, 2000). The nanominerals, chiefly allophane, ferrihydrite, and metal–humus complexes, are responsible for many of the unique properties exhibited by Andisols.

Despite covering less of the global ice-free land area than any other soil order (~1%), Andisols generally support high population densities, about 10% of the world's population (Ping, 2000). This is because they typically have exceptional physical properties for plant growth and, in many localities, high native fertility because relatively frequent additions of tephra can renew potential nutrient sources (Ugolini and Dahlgren, 2002; Dahlgren et al., 2004). The majority of Andisols occur in humid regions where there is adequate rainfall. Andisols often have high organic carbon contents. These and other factors make Andisols generally well suited for agriculture production and historically allowed establishment of nonshifting agricultural practices. Despite their generally favorable properties for plant growth, Andisols do pose some engineering and fertility challenges. These soils have low bulk densities, resulting in low weight-bearing capacity. Andisols also exhibit thixotropy and sensitivity, properties that cause them to behave in a fluid-like manner when loading pressures are applied (Neall, 2006; Arnalds, 2008). Andisols may exhibit substantial fertility limitations, including P fixation, low contents of exchangeable bases (especially K) and other nutrients, and strong acidity and Al toxicity (Shoji and Takahashi, 2002; Dahlgren et al., 2004; Lowe and Palmer, 2005).

33.3.2 Geographic Distribution

Andisols cover approximately 124 million ha or about 0.84% of the Earth's ice-free surface (Soil Survey Staff, 1999). They are closely associated with areas of active and recently active volcanism, and their global distribution is depicted in Figure 33.13. The greatest concentration of Andisols is 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, Taiwan, the Philippines, and Indonesia to Papua New Guinea and New Zealand. Other areas include the Caribbean, central Atlantic ridge, northern Atlantic rift, the Mediterranean, parts of China, Cameroon, the Rift Valley of east Africa, and southern Australia (Soil Survey Staff, 1999). There are numerous volcanic islands where Andisols are common, including Iceland, the Canary Islands, Azores, Galapagos Islands, Hawaiian Islands, the West Indies, and various small islands in the Pacific.

The global distribution of Andisols encompasses a wide variety of climatic conditions—cold-to-hot and wet-to-dry. This suggests that climate is less important to the formation of Andisols than is proximity to volcanic or pyroclastic parent materials. Nevertheless, the majority of Andisols are found in higher-rainfall regions of the world. Almost two-thirds of Andisols occur in humid regions (udic soil moisture regimes) while fewer than 5% occur in aridic moisture regimes (Mizota and van Reeuwijk, 1989; Wilding, 2000). Approximately half of the world's Andisols occur in the tropics, with the remaining half being split between boreal and temperate regions (Wilding, 2000; IUSS Working Group WRB, 2006).

There are almost 15.6 million ha of Andisols in the United States (Soil Survey Staff, 1999). The largest areas occur in Alaska

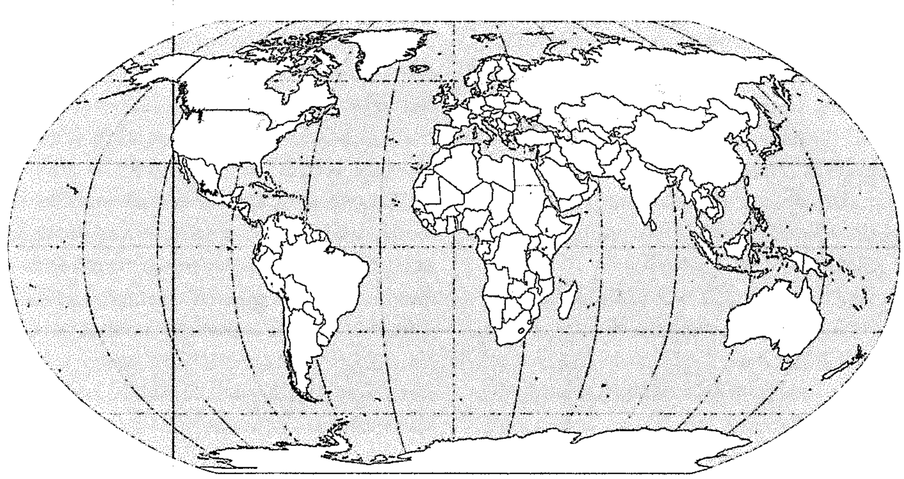

FIGURE 33.13 Global distribution of Andisols. (Courtesy of USDA-NRCS, Soil Survey Division, World Soil Resources, Washington, DC, 2010.)

(~10 million ha) and in Washington, Oregon, Idaho, northern California, and western Montana (Pewe, 1975; Rieger et al., 1979; Ping et al., 1989; Southard and Southard, 1991; Ugolini and Dahlgren, 1991; Goldin et al., 1992; Takahashi et al., 1993; McDaniel and Hipple, 2010). In the Pacific Northwest region of Washington, Idaho, and Oregon, most Andisols are forested and occur at mid-to-high elevations in cooler temperature regimes (McDaniel et al., 2005). Few Andisols are found in warmer temperature regimes because the summers are normally too hot and dry to allow sufficient weathering or leaching to produce the required andic properties.

Iceland contains ~7 million ha of Andisols. These represent the largest area of Andisols in Europe (Arnalds, 2004). Andisols also occur in France, Germany, Spain, Italy, and Romania (Buol et al., 2003; Kleber et al., 2004; Quantin, 2004; IUSS Working Group WRB, 2006; Arnalds et al., 2007). In New Zealand, ~3.2 million ha of Andisols occur on the North Island, the majority now supporting agriculture or forestry (Parfitt, 1990; Lowe and Palmer, 2005). Japan has ~6.9 million ha of Andisols (Wada, 1986; Takahashi and Shoji, 2002).

Some soils classified as Andisols are also found in humid areas not associated with volcanic activity such as in the southern Appalachian Mountains, parts of Kyushu (Japan), Scotland, Spain, and the Alps. These soils have large quantities of Al or Fe associated with humus (see Section 33.3.3.2) and similar management constraints as those of soils formed from volcanic ejecta and also key out as Andisols. These attributes further highlight the importance of realizing that Andisols are not classified on parent material but on the properties acquired during weathering and leaching. By the same token, soils other than Andisols, such as Entisols, Inceptisols, Spodosols, Mollisols, Oxisols, Vertisols, Alfisols, or Ultisols, may form in association with volcanic or pyroclastic materials (e.g., Shoji et al., 2006).

33.3.3 Andisol Properties

33.3.3.1 Morphological Features

Most Andisols have distinct morphological features. They usually have multiple sequences of horizons (Figure 33.14) resulting from the intermittent deposition of tephras and ongoing top-down soil formation referred to as upbuilding pedogenesis (see Section 33.3.5.1). A horizons are typically dark, often overlying reddish brown or dark yellowish brown Bw cambic horizons. Buried A–Bw sequences are common (Figure 33.14). Layers representing distinct tephra-fall events are common, often manifested as separate Bw horizons or as BC or C horizons if the tephra shows limited weathering or is relatively thick. Horizon boundaries are typically distinct or abrupt where these thicker layers occur.

Andisols are usually light and easily excavated because of their low bulk density and weakly cohesive clay minerals. The high porosity allows roots to penetrate to great depths. Andisols

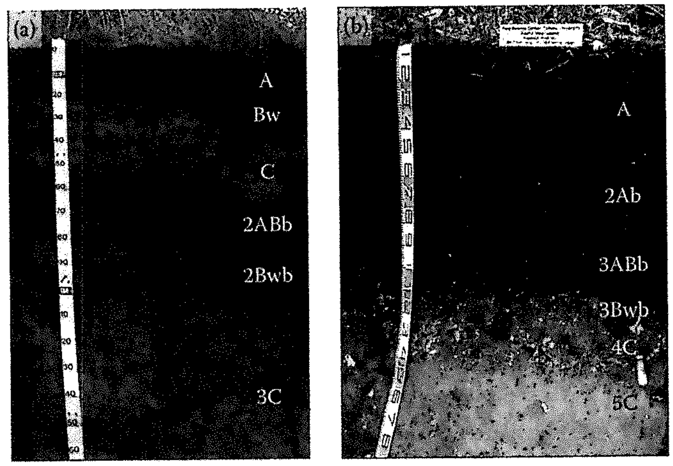

FIGURE 33.14 Andisol profiles (a) allophanic Taupo soil (Udivitrand) from New Zealand (scale divisions = 10 cm) and (b) nonallophanic Tohoku Farm soil (Melanudand) from Japan with thick (~1 m), dark, strongly humified horizons (melanic epipedon).

generally have granular structures in A horizons, but the structure in Bw horizons is generally weak subangular blocky, often crushing readily to crumb structure. Some Andisols (Udands) formed in areas of high rainfall have higher clay contents while soils that are subjected to wet and dry cycles form prismatic structures. At higher water contents, soils containing as little as 2% allophane have a characteristic greasy feel (Parfitt, 2009), an indication of sensitivity.

33.3.3.2 Mineralogical Properties

Tephra parent materials weather rapidly to form nanominerals that are responsible for many of the unique physical and chemical properties associated with Andisols. Although a wide range of clay minerals can be found in Andisols (such as gibbsite, kaolinite, vermiculite, smectite, crystalline Fe oxides such as hematite and goethite, and cristobalite), those of greatest interest are allophane, imogolite, ferrihydrite, and the Al- and Fe-humus complexes because they confer the characteristic andic properties (Dahlgren et al., 2004; Parfitt, 2009).

Allophane is nearly X-ray amorphous, but under an electron microscope it is structured over short distances, appearing as nanoparticles of hollow spheres 3.5–5 nm in diameter that have the chemical composition $(1\text{–}2)SiO_2 \cdot Al_2O_3 \cdot (2\text{–}3)H_2O$ (Figure 33.15a) (Wada, 1989; Churchman, 2000; Brigatti et al., 2006; Theng and Yuan, 2008). The most common type of allophane is the so-called Al-rich allophane with an Al:Si molar ratio of ~2 (it is sometimes called proto-imogolite allophane). There is also Si-rich allophane with an Al:Si ratio ~1 (also referred to as halloysite-like allophane).

Imogolite has the composition $(OH)SiO_3 \cdot Al_2(OH)_3$ and has both long- and short-range order. Under an electron microscope, it appears as long smooth and curved hollow threads or tubules with inner and outer diameters of ~0.7 and 2 nm, respectively

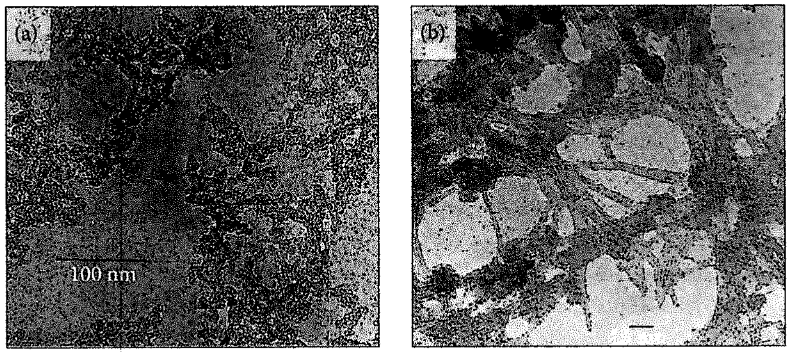

FIGURE 33.15 Micrographs of (a) allophane and (b) imogolite (external diameter of nanotubes is ~2 nm). (Reproduced with the kind permission of the Mineralogical Society of Great Britain and Ireland from a paper by Parfitt, R.L. 2009. Allophane and imogolite: Role in soil biogeochemical processes. Clay Miner. 44:135–155. With permission.)

(Figure 33.15b). These nanotubes typically appear as bundles of two or more threads 10–30 nm thick and several micrometers long (Theng and Yuan, 2008). Imogolite in Japan can be seen with the naked eye as a whitish gel film infilling pores in coarse pumice particles (Wada, 1989).

Allophane and imogolite both have high surface areas, ranging from 700 to 1500 m^2 g^{-1} (Parfitt, 2009), and this feature, coupled with their variable surface charge characteristics and exposure of $(OH)Al(OH_2)$ groups at wall perforations (defects), explains their strong affinity for water, metal cations, organic molecules, and other soil minerals (Harsh et al., 2002; Theng and Yuan, 2008). Even small amounts contribute huge reactive surface areas in soils (Lowe, 1995). Allophane and imogolite are soluble in ammonium (acid) oxalate solution, and the Si dissolved is used to estimate their contents in soils (Parfitt and Henmi, 1982; Parfitt, 2009). *Soil Taxonomy* uses oxalate-extractable Al (and Fe) to help define andic soil properties (see Figure 33.16 and Section 33.3.4). Allophane content of B horizons is quite

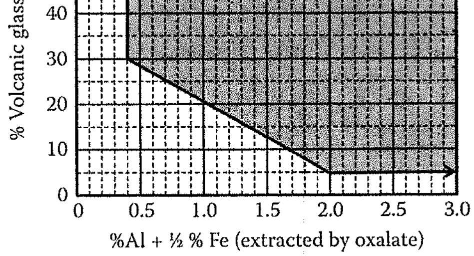

FIGURE 33.16 Andic soil properties as defined by *Soil Taxonomy*. (From Soil Survey Staff. 2010. Keys to soil taxonomy. 11th edn. USDA-NRCS, Washington, DC.)

variable, ranging from about 2% in slightly weathered or metal–humus-dominated systems to >40% in well-developed Andisols. It typically increases with depth in upper subsoils, usually being highest in the Bw and buried horizons. But in many Andisol profiles in New Zealand, allophane decreases and halloysite concomitantly increases with depth in lower subsoils either because of the downward migration of Si into lower profiles or because of changes in climate during upbuilding, or both. Imogolite is more commonly found in B horizons under carbonic acid weathering regimes than in A horizons where organic acid weathering dominates (Dahlgren et al., 2004). Allophane may occur dispersed as groundmass, as coatings, bridges, or infillings (in vesicles or in root channels), or it may be disseminated through pseudomorphs of glass or feldspar grains (Jongmans et al., 1994, 1995; Bakker et al., 1996; Gérard et al., 2007).

Ferrihydrite is common in many Andisols, especially those associated with more basic parent materials, has a composition of $Fe_5HO_8 \cdot 4H_2O$ and imparts a reddish brown color (hues of 5YR–7.5YR; Bigham et al., 2002). Made up of spherical nanoparticles 2–5 nm in diameter (Schwertmann, 2008), ferrihydrite has large, reactive surface areas ranging from ~200 to 500 m^2 g^{-1} (Childs, 1992; Jambor and Dutrizac, 1998). Its abundance is commonly estimated from the amount of Fe extracted by ammonium oxalate solution multiplied by 1.7 (Parfitt and Childs, 1988). It is a widespread and characteristic component of young Fe-oxide accumulations precipitated from Fe-rich solutions in the presence of organic matter, such as in Iceland (Arnalds, 2004), and elsewhere, including New Zealand, Japan, and Australia where its precipitation may be inorganic or bacteria-driven (Childs et al., 1991; Lowe and Palmer, 2005). Ferrihydrite can transform to hematite via solid-state transformation or goethite through dissolution and reprecipitation (Schwertmann, 2008).

Metal–humus complexes are significant components of some Andisol colloidal fractions. These Al- and Fe-organic complexes are immobile and accumulate in dark or black surface horizons where organic materials are abundant, and dark (melanic) horizons may extend to depths as much as 2 m (see Figure 33.14b). Metal–humus complexes represent the active forms of Al and Fe in nonallophanic Andisols as described below (Dahlgren et al., 2004).

Halloysite is a relatively fast-forming 1:1 layer silicate that often exhibits tubular or spheroidal morphology (White and Dixon, 2002; Joussein et al., 2005). Its formation is favored in seasonally dry environments where higher Si concentrations are maintained (Shoji et al., 1993). These include areas of lower rainfall, restricted drainage, and Si-rich parent materials (Lowe, 1986; Churchman, 2000; Churchman and Lowe, 2011, Section 20.1). Halloysite surfaces are characterized by some permanent negative charge, allowing retention of cations across a wide range of pH values.

The soil solution in Andisols in a range of locations may contain large amounts of dissolved Si, which leads to the formation by nucleation of secondary silica minerals from the saturated solution (Ping et al., 1988; Shoji et al., 1993; Ping, 2000; Nanzyo, 2002, 2007; Waychunas and Zhang, 2008). Termed laminar opaline silica, this material is circular or elliptical in shape (0.2–0.5 μm diameter) and extremely thin. Precipitation of the silica may be aided by evaporation or freezing of soil water, or via plant-related processes related to Si uptake and recycling (Lowe, 1986; Drees et al., 1989; Churchman, 2000; Henriet et al., 2008). Such silica polymorphs can be distinguished from biogenic forms of silica (phytoliths) because the latter have more complex shapes inherited from biological cells (Kondo et al., 1994; Nanzyo, 2007).

Andisols dominated by allophane with subordinate imogolite and ferrihydrite in upper horizons are referred to as *allophanic Andisols*. These contrast with a second, strongly acid group known as *nonallophanic Andisols* in which metal–humus complexes dominate the colloidal mineralogy. Nonallophanic Andisols are common in Japan especially where they account for about 30% of soils formed on tephras (Takahashi and Shoji, 2002) and are known in around 20 other countries (Saigusa and Matsuyama, 2004). Examples of soils from each group are shown in Figure 33.14. In Table 33.15, the Thingvallasveit and Tirau soils are examples of allophanic Andisols; the Tohuku Farm soil is an example of a nonallophanic Andisol. The mineralogical differences between these two groups of Andisols lead to several important different physical and chemical properties (especially the strong acidity of nonallophanic Andisols) and significant management implications (Dahlgren et al., 2004).

In the silt and sand fractions of Andisols, the dominant components are volcanic glass (a mineraloid) and various primary minerals. The glass particles (shards) which, like shattered glass, have sharp angles and edges, are very abrasive. However, these glass particles are usually coated with colloidal materials including allophane, ferrihydrite, and other Fe oxides and their humus complexes, which all contribute to aggregate formation. It is noteworthy that volcanic glass is often quite vesicular and porous in nature (as is pumice), and thus can retain water and has more chemical activity than other common sandy materials (Ping, 2000; Lowe and Palmer, 2005).

33.3.3.3 Chemical Properties

One of the common characteristics of Andisols is accumulation of relatively large quantities of organic matter, both in the allophanic (moderate pH) and nonallophanic Andisols (low pH; Table 33.15). Allophanic Andisols typically contain up to ~8%–12% C, whereas nonallophanic soils may contain up to ~25%–30% C (Mizota and van Reeuwijk, 1989). The residence time of C in Andisols, as measured by ^{14}C, is much greater than that of other soil orders (Parfitt, 2009). In addition, upbuilding pedogenesis leads to the storage of C in lower parts of profiles, and especially in buried A horizons that are sealed off and isolated from most surface processes.

Andisols are almost always acid, with most pH (H_2O) values ranging from 4.8 to 6.0 (Shoji et al., 1993; Dahlgren et al., 2004). Uncultivated, nonallophanic Andisols with high organic matter contents typically have a pH (H_2O) < 4.5 (IUSS Working Group WRB, 2006).

One of the key factors affecting the chemistry of Andisols is the variable surface charge associated with the colloidal fraction (Qafoku et al., 2004). Electrical charges on colloid or nanoparticle surfaces can be either positive or negative, and change as a function of pH. Surfaces have positive charge at lower pH and retain anions, while cations are retained by the negative charge at higher pH. This variable charge greatly affects the behavior of ions that are retained in soils. Cation retention capacity of Andisols makes the soils susceptible to metal and radioactive fallout ^{137}Cs pollution (Adamo et al., 2003; Sigurgeirsson et al., 2005). Andisols also exhibit anion exchange properties, which can be important for nutrient retention (e.g., NO_3^- and SO_4^{2-}) (Shoji et al., 1993).

Measured cation exchange capacity of Andisols is relatively high, often 20 to >50 $cmol_c\ kg^{-1}$. However, because the dominant colloids have variable charge, much of this CEC is pH dependent. This is especially true in allophanic Andisols (Dahlgren et al., 2004) and means that CEC decreases with decreasing pH. And because most Andisols are acid, CEC measurements made at pH 7 or 8.2 will be artificially high. In andic soils of the Pacific Northwest of United States, average CEC values determined using unbuffered extractants (effective CEC or ECEC) are approximately one-fourth of those determined at pH 8.2, 6.5 $cmol_c\ kg^{-1}$ vs. 26.4 $cmol_c\ kg^{-1}$ (McDaniel et al., 2005). This phenomenon needs to be considered when measuring CEC and base saturation or interpreting these data. The relatively low ECEC of Andisols can limit their ability to retain and exchange Ca, Mg, and K. Some representative cation exchange characteristics of Andisols are presented in Table 33.15.

The active Al and Fe compounds in Andisols (allophane/imogolite, metal–humus complexes, and ferrihydrite) also have the ability to sorb and strongly bind anions such as phosphate and fluoride (Shoji et al., 1993; Dahlgren et al., 2004; Parfitt, 2009). Much of this sorption is not reversible, leading to large quantities of phosphate being rendered unavailable for plant uptake. As described in Section 33.3.4, the amount of P retention in soils is used to define andic soil properties (Soil Survey Staff, 2010). Similarly, quantities of active Al and Fe compounds can also be estimated by reacting soil with NaF solution. Sorption of F^- releases OH^- into solution, thereby raising the pH. A resultant pH greater than ~9.5 indicates the presence of allophane/imogolite

TABLE 33.15 Selected Chemical and Mineralogical Properties of Representative Andisols

Horizon	Depth (cm)	pH H$_2$O	pH NaF	Clay (%)	Organic C %	ECEC[a]	CEC pH 7 (cmol$_c$ kg^{-1})	Sum of Exch. Bases	[b]Exch. Al^{3+}	P Retention	Al in Al-Humus Complexes[c] (%)	Allophane[d]	Ferrihydrite[e]
Thingvallasveit Series[f] (Haplocryand)—Iceland													
A1	0–12	5.4	11.0	—	7.9	—	31.9	9.9	—	99	—	13	8.5
A2	12–28	5.7	10.7	1	7.6	—	44.6	13.1	—	99	0.1	16	11.1
Bw1	28–61	5.6	10.8	—	7.4	—	43.8	10.1	—	99	—	15	12.7
Bw2	61–68	5.7	11.1	3	4.2	—	32.3	6.5	—	99	—	16	6.0
2Bw3	68–87	5.6	10.5	—	2.1	—	24.8	4.4	—	98	—	17	4.4
2C	87–142	5.5	10.3	1	0.8	—	11.8	2.8	—	99	—	13	3.6
Tirau Series[g] (Hapludand)—New Zealand													
Ap	0–18	5.6	9.6	19	7.9	14.6	29.7	14.4	0.2	88	0.7	9	1.0
Bw1	18–32	6.2	9.9	11	2.0	5.6	11.3	5.6	0	98	0.2	14	1.2
Bw2	32–48	6.2	9.8	13	1.0	5.2	10.5	5.2	0	98	0.2	16	1.2
Bw3	48–65	6.4	9.5	20	0.5	6.1	12.5	6.1	0	91	0.1	10	0.7
2BCb	65–90	6.4	9.4	18	0.5	7.5	13.2	7.5	0	86	0.1	8	0.8
Tohoku Farm[h] (Melanudand)—Japan													
A1	0–14	4.2	10.5	18	11.5	13.3	52.2	2.3	11.0	83	1.5	0.3	—
2A2	14–30	4.6	11.6	5	8.0	8.4	49.1	0.4	8.0	93	2.5	1.1	—
3A3	30–57	4.8	11.6	12	13.2	8.4	56.1	0.4	8.0	93	2.7	1.4	—
4Bw1	57–80	5.2	11.3	8	2.8	2.0	18.4	0.1	1.9	93	0.9	6.7	—
4Bw2	80–126	5.2	11.0	16	0.9	0.9	13.7	0.2	0.7	94	0.6	7.4	—
4Bw3	126–160	5.3	10.9	20	0.6	0.8	10.9	0.2	0.6	89	0.5	7.0	—
4C	160–200	5.3	10.9	22	0.5	1.8	11.5	0.4	1.4	84	0.5	4.9	—

[a] Effective cation exchange capacity.
[b] Exchangeable Al^{3+} extracted with KCl.
[c] Estimated using pyrophosphate-extractable Al.
[d] Estimated using oxalate-extractable Si (Parfitt, 1990).
[e] Estimated using oxalate-extractable Fe (Parfitt and Childs, 1988).
[f] Data are from Arnalds et al. (1995).
[g] Data are from Bakker et al. (1996).
[h] Data (Pedon #86P0091) are from Soil Survey Staff 2011.

and/or Al–humus complexes, and because of this, NaF field test kits can be used for field identification of Andisols (Fieldes and Perrott, 1966; IUSS Working Group WRB, 2006).

33.3.3.4 Physical Properties

Unique physical attributes of Andisols are related to structural assemblages of hollow spheres and tubular threads as mineral entities into resilient, progressively larger (silt-sized) aggregated domains. This aggregation results in low density, high porosity, high surface area, and high soil water retention even at low water potentials. The structural arrangement accounts for the low thermal conductivity of andic materials, which is three to four times less than that of the phyllosilicates in other mineral soils. It also accounts for the thixotropic and sensitivity character of these soils and several irreversible changes in physical properties that occur upon drying (Ping, 2000; Neall, 2006).

Andisols have low bulk density, usually <0.9 g cm^{-3}, because of the high organic matter and nanomineral contents (Table 33.16) and well-developed aggregation. This results in good tilth and makes them excellent rooting media. On the other hand, low bulk densities result in low weight-bearing capacity and make Andisols highly susceptible to wind and water erosion when surface cover is removed or degraded (Ping, 2000; Nanzyo, 2002; Dahlgren et al., 2004; Neall, 2006). Because of the nature of volcanic ejecta and its distribution, many Andisols in proximal locations contain appreciable amounts of gravel and stones, and coarse-grained tephra layers may have adverse effects on hydrological properties by interrupting capillary movement of water. Some additional adverse physical properties include a high glass content that can reduce the quantity and biodiversity of soil organisms such as earthworms, and the presence occasionally of impenetrable horizons such as thin Fe pans (placic horizons) in higher rainfall areas (Neall, 2006).

Andisols typically exhibit high water retention because of the presence of allophane, ferrihydrite, and metal–humus complexes, which have high surface areas as noted previously. As a result, moisture contents of many Andisols can exceed 100% on a weight basis, even at soil moisture tensions of 1500 kPa—this feature is illustrated by the data for the Hilo soil (Table 33.16). Water retention is greatest in Andisols that have undergone significant weathering and hence have high clay contents.

Most field-moist Andisols have a greasy feel when rubbed between the fingers and exhibit smeariness—these can be both indicators of the properties known generally as sensitivity and thixotropy (Soil Survey Staff, 1975; Torrance, 1992). Thixotropy is a reversible gel–sol transformation that occurs when shear forces are applied to a moist soil. The applied shear force causes the soil to abruptly lose strength, sometimes to the point of behaving as a fluid. When the shear force is removed, the soil will recover some or all of its original strength. Sensitivity (the term used more commonly by engineers) is the ratio of undisturbed to disturbed (remolded) shear strength, that is, the maximum strength of

TABLE 33.16 Selected Physical Properties of Representative Andisols

		Texture					Bulk Density (g cm^{-3})		1500 kPa H_2O (%)	
Depth (cm)	Horizon	Field	Lab	Sand (%)	Silt (%)	Clay (%)	Oven Dry	Moist	Air-Dry	Moist
Bonner Series[a] (Vitrixerand)—Idaho										
0–4	A	Sil	Sil	30.6	63.8	5.6	0.75	0.68	20.3	—
4–20	Bw1	Sil	Sil	38.1	58.0	3.9	—	—	10.0	—
20–48	Bw2	Sil	Sil	39.3	59.0	1.7	0.96	0.94	9.9	—
48–69	Bw3	—	Cosl	52.6	44.1	3.3	—	—	6.8	—
69–89	Bw4	—	Cosl	52.7	44.8	2.5	0.80	0.80	6.8	—
89–152	2C	—	Cos	88.6	9.8	1.6	—	—	2.2	—
Tirau Series[b] (Hapludand)—New Zealand										
0–18	Ap	Sil	L	35	46	19	0.75	—	23.3	31.1
18–32	Bw1	Sl	Sil	33	56	11	0.71	—	15.7	33.7
32–48	Bw2	Sl	Sil	33	54	13	0.69	—	16.5	33.6
48–65	Bw3	Sl	L/Sil	30	50	20	0.79	—	22.0	34.7
65–90	2BCb	Sil	Sil	20	62	18	0.87	—	20.4	33.9
Hilo Series[c] (Hydrudand)—Hawaii										
0–18	Ap1	Sicl	Sl	66.2	29.7	4.1	—	—	28.4	54.1
18–36	Ap2	Sicl	Cos	86.8	12.1	1.1	0.82	0.51	27.7	107.2
36–60	Bw1	Sicl	Cos	91.8	7.9	0.3	1.66	0.41	25.0	112.6
60–92	Bw2	Sicl	Cos	93.1	6.9	—	1.61	0.25	25.1	132.5
92–108	Bw3	Sicl	Cos	94.9	4.7	0.4	1.41	0.30	26.0	122.4

Sil, Silt loam; Sicl, Silty clay loam; Cosl, Coarse Sandy loam; Cos, Coarse sand; Sl, Sandy loam; L, Loam.

[a] Data (Pedon #78P0553) are from Soil Survey Staff 2011.

[b] Data are from Bakker et al. (1996).

[c] Data (Pedon #89P0658) are from Soil Survey Staff 2011.

an undisturbed specimen compared with the residual strength remaining after force or strain is applied (Mitchell and Soga, 2005; Neall, 2006). In sensitive materials, the original strength is generally not recovered—unlike thixotropic materials—after removal of the force. Both sensitivity and thixotropic behavior are best expressed in Andisols having high water retention and lack of layer silicates to provide cohesion and can pose significant engineering problems.

Many Andisols exhibit irreversible changes upon drying. Allophane nanospheres collapse upon dehydration and form larger aggregates that do not break down upon rewetting. This phenomenon can cause crust formation at the soil surface during hot dry periods. It also results in well-known unreliable particle size analyses of air-dried Andisols. Normally, clay content is underestimated and sand and silt contents are overestimated (Ping et al., 1989; Dahlgren et al., 2004). However, reliable sand-, silt-, and crystalline clay-size fraction data were obtainable for andesitic Andisols in New Zealand by Alloway et al. (1992) who analyzed grain-size distributions of residual material following selective dissolution of nanominerals and organic constituents via ammonium oxalate. Other irreversible changes that occur upon desiccation include increases in bulk density, decreases in water retention, and increases in cohesive strength. Note the difference in water retention and bulk density values between dried and moist samples in Table 33.16.

Plasticity in Andisols is different from that of soils containing layer silicate clay minerals. Generally, field-moist Andisols have high liquid (60%–350%) and plastic limits (70%–180%) (Warkentin and Maeda, 1980; Neall, 2006). The low plasticity index (0–10) also clearly separates Andisols from other soils. Plasticity measurements can be used as an index of physical behavior in Andisols and as a substitute for particle size analysis, which usually is not reliable. Air-dry samples, on the other hand, often show low plasticity because of the irreversible changes on drying and behave like sandy soils.

33.3.4 Classification of Andisols

Andisols are classified on the basis of having andic soil properties, which are quantitatively defined in *Soil Taxonomy* (Soil Survey Staff, 2010). In general, andic soil properties consist of combinations of properties that develop as tephras and other volcanic materials weather. These include relatively low bulk density values ($\leq$0.90 g cm^{-3}), relatively high phosphate retention (>25%–85%), the presence of volcanic glass, and the presence of nanoscale weathering products containing Al, Fe, and Si (Soil Survey Staff, 2010). The criterion of percentage Al extracted by ammonium oxalate (Al_o) plus half the percentage Fe extracted by ammonium oxalate (Fe_o) from short-range order nanominerals is used to quantitatively define andic properties. The quantity 0.5 Fe_o is used to normalize the criterion because Fe has an atomic weight (56), approximately twice that of Al (27). The majority of soils with a glass content and percentage of Al_o + $0.5Fe_o$ that falls within the shaded area of Figure 33.16 have andic soil properties. Soils possessing at least 36 cm of material with andic soil properties are classified as Andisols (Soil Survey Staff, 2010). It is emphasized that freshly deposited volcanic ash does not have andic soil properties and would not be classified as an Andisol. It is not until some weathering has occurred—sufficient to generate Al_o + $0.5Fe_o$ totaling at least 0.4%—that andic properties are developed.

Andisols are separated into suborders primarily on the basis of soil moisture and/or temperature regimes (Table 33.17). Aquands are poorly drained and have a water table at or near the soil surface for much of the year. These soils are usually restricted to low-lying landscape positions and have dark surface horizons. Because of excessive wetness, Aquands typically require drainage in order to be used for crop or pasture production.

Gelands are very cold Andisols that have a mean annual soil temperature (MAST) $\leq$0°C (Soil Survey Staff, 2010). Relatively little is known about the distribution of these soils, but they are found at higher latitudes in areas of either current or recent volcanic activity.

Cryands are cold Andisols having a cryic temperature regime (0°C < MAST $\leq$ 8°C and summers are cool; Soil Survey Staff, 2010). They are found at higher latitudes and higher elevations. In the United States, Cryands are found mainly in Alaska and mountainous regions of the Pacific Northwest (Ping, 2000; McDaniel and Hipple, 2010). They are extensive in Iceland (Arnalds and Kimble, 2001) and also on the Kamchatka Peninsula and occur in mountainous regions elsewhere including in eastern Africa, the Andes, and (uncommonly) New Zealand (Ping, 2000; Lowe and Palmer, 2005). Globally, there are ~26 million ha of Cryands, and they are the third most common suborder, representing ~28% of Andisols (Soil Survey Staff, 1999; Wilding, 2000).

Torrands are Andisols of very dry environments where moisture for plant growth is very limited. This lack of soil moisture slows down weathering processes and leaching, thereby inhibiting development of andic soil properties. Torrands are found in Oregon and Hawaii in the United States. They are the least extensive of any of the Andisol suborders, with only ~100,000 ha occupying the global ice-free land area (<1% of Andisols).

Xerands occur in temperate regions with xeric soil moisture regimes, which are characterized by cool, moist winters and very warm, dry summers (Mediterranean climates). In the United States, they occur primarily in northern California, Oregon, Washington, and Idaho where they have formed under coniferous forest. Elsewhere, Xerands occur in scattered localities including Italy, Canary Islands, Argentina, and South Australia (Broquen et al., 2005; Lowe and Palmer, 2005; Inoue et al., 2010). Xerands are uncommon globally (~4% of Andisols).

Vitrands are the only suborder that is not defined by a climatic regime. These Andisols are relatively young and only slightly weathered. They tend to be coarse-textured and have a high content of volcanic glass that may be strongly vesicular or pumiceous. In the United States, Vitrands are found in Washington, Oregon, and Idaho. They also occur in Argentina (Broquen et al., 2005) and are common in the North Island of New Zealand (Lowe and Palmer, 2005). Globally, there are ~28 million ha of Vitrands (~31% of Andisols) making them and

TABLE 33.17 Listing of Andisol Suborders and Great Groups

Suborders	Great Groups
Aquands—poorly drained Andisols with a water table at or near the surface for much of the year	
	Gelaquands—Aquands of very cold climates (mean annual soil temperature <0°C)
	Cryaquands—Aquands of cold climates (0°C < mean annual soil temperature ≤8°C)
	Placaquands—Aquands with a thin pan cemented by Fe, Mn, and organic matter
	Duraquands—Aquands with a cemented horizon
	Vitraquands—Aquands with coarse textures dominated by glassy materials
	Melanaquands—Aquands with a thick, dark, organic matter-rich surface layer
	Epiaquands—Aquands with a perched water table
	Endoaquands—Aquands with a groundwater table
Gelands—Andisols of very cold climates (mean annual soil temperature ≤0°C)	
	Vitrigelands—Gelands with coarse textures dominated by glassy materials
Cryands—Andisols of cold climates (0°C < mean annual soil temperature ≤8°C)	
	Duricryands—Cryands with a cemented horizon
	Hydrocryands—Cryands with very high water-holding capacity
	Melanocryands—Cryands with a thick, very dark, organic matter-rich surface layer
	Fulvicryands—Cryands with a thick, organic matter-rich surface layer
	Vitricryands—Cryands with coarse textures dominated by glassy materials
	Haplocryands—other Cryands
Torrands—Andisols of very dry climates	
	Duritorrands—Torrands with a cemented horizon
	Vitritorrands—Torrands with coarse textures dominated by glassy materials
	Haplotorrands—other Torrands
Xerands—Andisols of temperature regions with warm, dry summers and cool, moist winters	
	Vitrixerands—Xerands with coarse textures dominated by glassy materials
	Melanoxerands—Xerands with a thick, very dark, organic matter-rich surface layer
	Haploxerands—other Xerands
Vitrands—slightly weathered Andisols that are coarse textured and dominated by glassy material	
	Ustivitrands—Vitrands of semiarid and subhumid climates
	Udivitrands—Vitrands of humid climates
Ustands—Andisols of semiarid and subhumid climates	
	Durustands—Ustands with a cemented horizon
	Haplustands—other Ustands
Udands—Andisols of humid climates	
	Placudands—Udands with a thin pan cemented by Fe, Mn, and organic matter
	Durudands—Udands with a cemented horizon
	Melanudands—Udands with a thick, very dark, organic matter-rich surface layer
	Hydrudands—Udands with very high water-holding capacity
	Fulvudands—Udands with a thick, organic matter-rich surface layer
	Hapludands—other Udands

Source: Soil Survey Staff. 2010. Keys to soil taxonomy. 11th edn. USDA-NRCS, Washington, DC.

Udands (similar in extent) the two most extensive suborders (Soil Survey Staff, 1999; Wilding, 2000).

Ustands occur in tropical or temperature regions that have ustic soil moisture regimes. The ustic soil moisture regime is characterized by an extended dry period, but moisture is normally present at a time when conditions are suitable for plant growth (Soil Survey Staff, 2010). Ustands are of fairly limited extent in the United States and are found mainly in Hawaii. They also occur in Mexico, on Pacific islands, and in eastern Africa (Dubroeucq et al., 1998; Takahashi and Shoji, 2002). About 6.3 million ha of Ustands are known globally (~7% of Andisols).

Udands are Andisols with a udic moisture regime, which is common to humid climates. Udands are thus characterized by well-distributed precipitation throughout the year and limited periods of soil moisture stress. In the United States, Udands are found in western parts of Washington and Oregon and in Hawaii. Elsewhere, they occur commonly on other parts of the Pacific Rim including in Patagonia (Argentina), Mexico, Japan, the Philippines, Indonesia, and New Zealand (Takahashi and Shoji, 2002; Van Ranst et al., 2002; Broquen et al., 2005; Prado et al., 2007). In total, Udands occupy nearly ~28 million ha of ice-free land globally (~30% of Andisols; Soil Survey Staff, 1999; Wilding, 2000).

Each of the suborders is further separated into great groups. These are listed and described in Table 33.17. A variety of characteristics are used to define great groups, including the presence or absence of certain types of soil horizons, moisture and temperature regimes, glass content and texture, and water retention.

There are also soils of other orders that have been influenced to a lesser degree by andic materials (Parfitt, 2009). Andic soil properties have developed, but their distribution within the soil is not of sufficient thickness for the soils to be classified as Andisols. These soils are therefore classified as andic subgroups of other orders. In cases where andic materials have been extensively mixed with other parent materials, or andic properties are only weakly expressed, soils are classified as vitrandic subgroups of other soil orders (Soil Survey Staff, 2010). In the United States, such soils are extensive in the Pacific Northwest where they are transitional to higher-elevation, forested Andisols (McDaniel and Hipple, 2010).

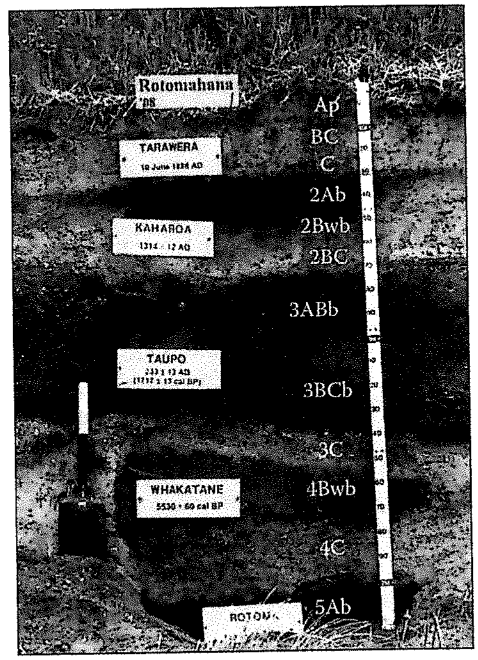

FIGURE 33.17 Udivitrand (Rotomahana soil) from New Zealand composed of multiple tephra deposits and associated buried soils and horizons; scale divisions = 10 cm; the oldest tephra layer (Rotoma) at bottom of exposure is 9505 ± 25 calibrated years BP.

33.3.5 Formation of Andisols

33.3.5.1 Parent Material and Stratigraphy

In most cases, the parent materials from which Andisols have formed are of explosive volcanic origin (rather than effusive) and thus fragmental and unconsolidated. Such materials range in size from ash (<2 mm) and lapilli (2–64 mm) through to large angular blocks or part-rounded bombs (>64 mm)—that is, from fine dust to boulders. Collectively, these materials are termed pyroclastic deposits or tephra (Gk *ashes*) (Alloway et al., 2007). Tephra deposits typically are loose and very coarse and thick close to source vents but become markedly finer and thinner with increasing distance away from source so that at more than ~100 km from vent most comprise mainly ash-sized material (equivalent to sand or finer particles). The accumulation at a particular site of numerous tephra deposits from sequential eruptions from one or more volcanoes leads usually to the formation of Andisols with distinctive layered profiles and buried soil horizons, forming multisequal profiles (Figure 33.17). Such layered profiles, together with their andic soil properties, are special features of Andisols. Study of the layers and attaining ages for them (tephrostratigraphy) is an important aspect of understanding Andisol formation (Lowe and Palmer, 2005; Lowe and Tonkin, 2010).

During periods of quiescence between major eruptions, soil formation takes place, transforming the characteristics of the unmodified tephra via normal top-down pedogenesis whereby the materials are altered in a downward-moving front to form subsoil horizons. However, when new tephras are added to the land surface, upbuilding pedogenesis takes place. The frequency and thickness of tephra accumulation (and other factors) determine how much impact the top-down processes have on the ensuing profile character, and if "developmental" or "retardant" upbuilding, or both, will take place. Two contrasting scenarios can be considered.

In the first scenario, successive thin tephra deposits (ranging from millimeters to centimeters in thickness) accumulate incrementally and relatively infrequently, so that developmental upbuilding ensues. Such a situation occurs typically at distal sites. The thin materials deposited from each eruption become incorporated into the existing profile over time. Top-down pedogenesis continues as the tephras accumulate but its effects are lessened because any one position in the sequence is not exposed to pedogenesis for long before it becomes buried too deeply for these processes to be effective as the land surface gently rises (Figure 33.18). This history thus leaves the tephra materials with a soil fabric inherited from when the tephra was part of the surface A horizon or subsurface Bw horizon (Lowe and Palmer, 2005; Lowe and Tonkin, 2010). Each part of the profile has been an A horizon at one point, as illustrated in Figure 33.18.

In the second scenario, tephra accumulation is more rapid, as occurs in locations close to volcanoes, or when a much thicker layer (more than a few tens of centimeters) is deposited from a powerful eruption. In the latter case, the antecedent soil is suddenly buried and isolated beyond the range of most soil-forming processes (i.e., it becomes a buried soil horizon or paleosol). A new soil will thus begin forming at the land surface in the freshly deposited material. This scenario typifies retardant upbuilding, which recognizes that the development of the now-buried soil has been retarded or stopped, and the pedogenic "clock" reset to time zero for weathering and soil formation to start afresh. An example of a multisequal Andisol profile formed via retardant upbuilding pedogenesis since ~9500 calendar

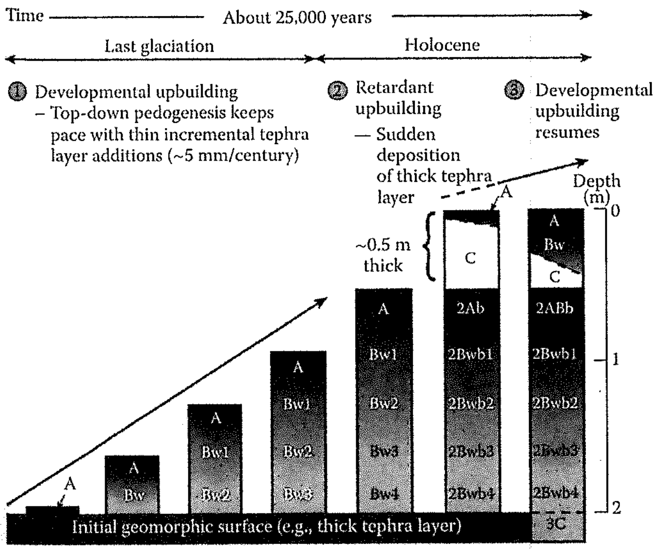

FIGURE 33.18 Model of upbuilding pedogenesis profiles over ~25,000 years in the Waikato region, New Zealand. (From Lowe, D.J., and P.J. Tonkin. 2010. Unravelling upbuilding pedogenesis in tephra and loess sequences in New Zealand using tephrochronology, p. 34–37. *In* Proc. IUSS 19th World Soil Congress, Symposium 1.3.2 Geochronological techniques and soil formation, Brisbane, Australia August 1–6, 2010. Published on DVD and online at: http://www.iuss.org. With permission.)

(cal) years ago is shown in Figure 33.17. Each of five successive tephra deposits shows the imprint of top–down pedogenesis, as depicted by their horizonation. But the sudden arrival of a new deposit every few thousand years or so on average has buried and effectively isolated each of the weakly developed "mini" soil profiles as the land surface rises.

In addition to stratigraphic factors, Andisols are markedly affected by the mineralogical and physicochemical compositions of the parent tephras, or associated deposits derived from remobilization of volcanic and other material (collectively termed volcaniclastic deposits). For example, the marked influence of windblown dust (mainly basaltic glass) on soil properties in Iceland was described by Arnalds (2010). Tephras differ widely according to the chemical makeup of magmas of the volcanoes that generated them. The chemistry of magmas, especially Si content, governs the way a volcano erupts. Three main magma types and resulting eruptives can be identified according to their chemical composition—rhyolitic (≥~70% SiO_2), andesitic/dacitic (~50%–70% SiO_2), and basaltic (≤~50% SiO_2). All magmas generate volcanic glass, a noncrystalline, easily weatherable mineraloid, and various other primary silicate minerals (Shoji et al., 1993; Nanzyo, 2002; Smith et al., 2006; Alloway et al., 2007; De Paepe and Stoops, 2007). Glass especially provides much of the Si and Al required to dissolve and re-form as allophane or other aluminosilicate clay minerals, and the amounts differ according to the magma composition as shown in Figure 33.19. Feldspars also release Si and Al via weathering for clay formation.

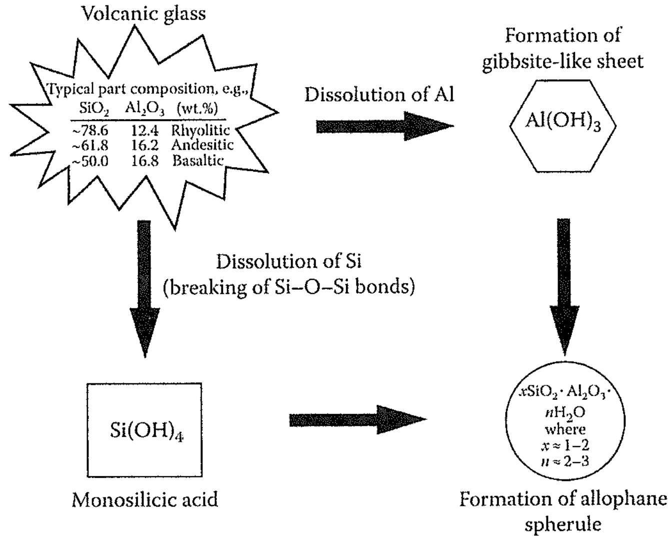

FIGURE 33.19 Volcanic glass compositions and dissolution of Al and Si to form allophane. (After Hiradate, S., and S.-I. Wada. 2005. Weathering processes of volcanic glass to allophane determined by ^{27}Al and ^{29}Si solid-state NMR. Clays Clay Miner. 53:401–408.)

Generally, the basaltic and intermediate tephras tend to weather more readily than rhyolitic tephras, and in all cases glasses weather very quickly (Neall, 1977; Kirkman and McHardy, 1980; Colman and Dethier, 1986; Hodder et al., 1996; Shikazono et al., 2005). Compared with hard rock, the fragmental tephra components, especially vesicular glass and pumice fragments, have a much greater surface area and higher porosity and permeability, and so break down to constituent compounds very readily. As this weathering occurs, various elements including Si and Al are released into chemical solution either for subsequent leaching, complexing with humic materials, plant uptake, or synthesis (neoformation) into relatively stable nanominerals and other clay minerals (Lowe, 1986; Vacca et al., 2003; Dahlgren et al., 2004). Andisols developed on (base-rich) basaltic eruptives are inherently more fertile than those on more siliceous eruptives (Wolff-Boenisch et al., 2004). Weakly weathered tephras give rise to the glass-rich Andisols that belong to the Vitrand suborder.

33.3.5.2 Climate

Climate plays an important role in Andisol formation, and soil moisture and temperature regimes are used to differentiate all the suborders except Vitrands (Ping, 2000). The majority of Andisols, as reported previously, are found under udic soil moisture regimes, and around half are found in the tropics, with the rest occurring in temperate or boreal regions. Climatic conditions help govern the combinations of processes, collectively referred to as andisolization, that occur in soils developing on tephras. Andisolization is the in situ formation of andic soil materials comprising nanominerals composed of "active" Al, Si, Fe, and humus (Dahlgren et al., 2004). The process is discussed in more detail in Section 33.3.6.