PROPERTIES OF THE CLAY FRACTION OF ALFISOLS AND ULTISOLS IN THAILAND

INTRODUCTION

The clay fraction of many soils in Thailand is dominated by kaolin. Variable amounts of illite, smectite and inhibited vermiculite are also constituents of these soils along with quartz, anatase and iron oxide minerals. These soil kaolins give broad basal reflections and have poorly ordered structures but little is known of their properties (Yoothong *et al.*, 1997). Kaolin is one of the most widespread clay minerals in soils, especially those of warm, moist climates (Dixon, 1989). The deeply weathered saprolite underlying many highly weathered soils is dominated by kaolin and soils developed from such materials are also often dominated by kaolin (Gilkes *et al.*, 1973). Kaolin is the dominant clay mineral in the pallid zone subsoil of lateritic profiles and mostly formed by *in situ* chemical weathering of parent rocks (Anand *et al.*, 1985).

The characteristics of kaolin may strongly influence the physical and chemical properties of soils. Many soils in Thailand are inherently very low in chemical and physical fertility so that many management problems occur. These have in some instances been related to the dominance of kaolin in the clay fraction (Suddhiprakarn *et al.*, 1985; Yoothong *et al.,* 1997).

Ultisols are widespread in Thailand covering about 45 percent of the total country area. They occur in every region, especially in the Northeast Plateau, Southeast Coast and Peninsular Thailand. Alfisols (9% of the total area) occur in drier areas of the country, mainly in the Central Highlands, Central Plain, North and West Continental Highlands and some parts of the Northeast Plateau. Ultisols have developed over deep saprolite of diverse rock types and many of them developed on colluvium and alluvium (Kheoruenromne, 1991; Suwanasing, 1992).

By definition based on soil taxonomy (Soil Survey Staff, 1999; Ahrens and Arnold, 2000), Alfisols have base saturation of greater than 35 percent, whereas

Ultisols contain lower base saturation (<35%). However, there are similarities between these two soil orders in Thailand. They generally have low cation exchange capacity (CEC) and common clay minerals. Alfisols are often composed of smectite and inhibited vermiculite is rather common in the Ultisols. In most cases, both soil orders are found to be dominant by kaolin in their clay fraction (Soil Survey Staff, 1999; Ahrens and Arnold, 2000)

With kaolin as the major clay mineral species in the soils, the fertility status of these soils is low. Nevertheless, with proper management particularly in the Tropics, these soils can be used extensively for economic crop production.

Hypothesis

Reactions at mineral surfaces largely control the fertility and chemical reactivity of Thai Alfisols and Ultisols, so, the better knowledge of mineral components will assist in the management of fertility and also of contamination of these soils.

1. Kaolin group and iron oxide are the most abundant in Thai soils.

2. Trace element concentrations in Thai soils are largely unknown and lead, cadmium, arsenic, copper and zinc may be the pollutants of Thai soils.

Objectives

1. To identify and quantitatively determine the clay minerals in topsoil and subsoil samples of Alfisols and Ultisols in Thailand

2. To determine the intrinsic properties of kaolin mineral in selected soils using a number of analytical techniques (e.g. XRD, XRF, chemical analysis and transmission electron microscopy)

3. To elucidate the geochemistry of these Alfisols and Ultisols in Thailand relating to some factors and conditions that influence different concentrations of their major and trace elements.

LITERATURE REVIEW

1. Mineralogy of the clay fraction

The clay fraction is most important component of surface deposits and soils, whether in term of origin, agriculture or engineering use, or as a factor in the environmental, physical, chemical and biological processes (Velde, 1992). Clay is cohesive, it holds soil aggregates together and controls soil structure. Clay retains moisture, and thereby influences the movement of water and air in soil. Clay retains plant nutrients against leaching and sometimes against plant uptake (Brady and Ray, 2002; Singh and Gilkes, 1992a).

The clay fraction of soils is composed of primary and secondary minerals of variation sizes and compositions. Despite the variation in composition, the inorganic fractions are predominantly silicates and oxides in most soils. They are usually distinguished into secondary minerals inherited from various primary minerals. Primary minerals present in soils are chemically unchanged, whereas secondary minerals are minerals that have been formed by weathering and alteration of primary minerals. Clay minerals which are mostly hydrous aluminosilicates are common secondary minerals. The most stable form of the clay mineral is probably kaolin which dominates highly weathered soils. Clay minerals are a major product of weathering processes and also a major component of many sedimentary rocks (Allen and Hajek, 1989). Clay minerals are of special importance in soil chemistry, since they have a different surface chemistry from that of larger mineral grains. Clay materials exhibit bulk physical properties different from gravel, sand or silt. Many of the minerals in soil clay fractions are crystalline (ordered) in structure, whereas others may be structurally disordered and some minerals in the clay fractions may be amorphous (e.g. silica, alumina and iron oxide gels) (Tan, 1993).

1.1 Classification of clay minerals

Geologists have used the word “clay” in two senses; as a size term, to refer to material of any composition whose average grain size is less than approximately 0.002 mm, and as a mineralogical term, to refer to a group of minerals with a specific range of composition and a particular crystallographic structure. The two meanings often overlap, because the fine-grained part of a soil or sediment most commonly consists largely of clay minerals.

Clay minerals have been classified in various ways, but a usual classification, based on their crystal chemistry, is shown in Table 1. The clay minerals are divided into five major groups by their structural arrangement (Velde, 1992). This study has emphasized the kaolin group minerals because of the dominance of these minerals in the clay fraction of Thai soils (Suddhiprakarn *et al.*, 1985; Hart *et al.*, 2003; Trakoonyingcharoen, 2005; Thanachit, 2006).

1.1.1 The kaolin minerals

Structure of kaolin: Kaolin group minerals are hydrated aluminosilicates, with a general chemical composition Al_2O_3: SiO_2: H_2O =1:2:2 or $2Al_2O_3$•$4SiO_2$ •$4H_2O$ per formula (Tan, 1993). The formula for kaolin is $Si_4Al_4O_{10}(OH)_8$, with very little isomorphous substitution The layer consists of a single sheet of hexagonal Si-O network and layer of Al-O. Layers are joined together in kaolin crystals by weak hydrogen or electrostatic bonds (Grim, 1968; Brown and Brindley, 1980).

Monoclinic structure

The unit layer in its ideal geometrical form has a reflection plane of symmetry, and a monoclinic structure can be developed by stacking the layers so that it is mirror plane, the b axis normal to this plane and the c axis will be inclined to the ab plane at an angle depending on the displacement of successive layer parallel to c. The basal 001 reflections are easily recognized by x-ray diffraction. This reflections show that the structure is formed of layers having a thickness of about 0.715 nm.

Single crystal diffraction diagrams of kaolinite can be obtained by electron diffraction showing hk0 net pattern, and basal and general hkl reflections can be obtained by tilting the crystal with repeat to the electron beam (Gilkes, 1994).

Table 1 Classification of principle clay minerals

Sheet	Population of octahedral sheets	Expansion	Species	Ideal formula
1:1	Dioctahedral	Non-swelling	Kaolinite	$Al_4Si_4O_{10}(OH)_8$
		Swelling	Halloysite	$Al_4Si_4O_{10}(OH)_8.4H_2O$
2:1	Trioctahedral	Non-swelling	Septechlorite	$Mg_6Si_4O_{10}(OH)_8$
	Dioctahedral	Swelling	Montmorillonite	$(Al_3Mg)Si_8O_{20}(OH)_4$
		Non-swelling	Illite	$K_{0-2}A_{l4}(Al,Si)_8O_{20}(OH)_4$
	Trioctahedral	Swelling	Vermiculite	$Mg(Al,Fe,Mg_4)(Al_2Si_6)O_{20}(OH)_4$ $.nH_2O$
2:1:1	Trioctahedral	Non-swelling	Chlorite	$Mg_6(OH)_{12}(Al,Mg_5)(Al_2Si_6)O_{20}(OH)_4$

Sources: Velde (1992); Richard *et al.* (1997); Olsen *et al.* (2000)

Triclinic structure

The structural analysis of kaolinite on the basis of the triclinic unit cell is considerably easier than for the monoclinic unit cell. This cell has

$$d(001) = c(1-\cos^2 \alpha - \cos^2 \beta)^{1/2} = 0.715 \text{ nm}$$

It contains only a single structure layer, and the analysis resolves itself into finding the correct orientation of the layer with respect to the a and b axes. There are six ways of arranging the structural layer to conform with the cell dimensions, and only one of these gives calculated intensities in satisfactory agreement with those observed. Oxygen and hydroxyl ions are grouped in pairs, and this suggests that the layers are bound together by hydrogen or hydroxyl type bonds. It is reasonable to expect that the layers will be stacked with respect to one another so that these O-OH distances are equal or at least approximately equal.

High-defecte kaolin structure

Considerable variation exists in the degree of structural regularity of kaolin group minerals. Disorder in the lattice produces blurring and weakening of reflections, and usually of particular groups of reflections, depending on the disordering process. In the idealized layer structure, the OH ions in the whole hydroxyl layer are located in lines parallel to *b* axis and at intervals of *b*/3. Poorly ordered kaolins are disordered along the *b* axis with the unit layer randomly displaced by multiple $b_o/3$. The structural layers can be displaced parallel to *b* by n*b*/3, without altering the OH-O bonds between the adjacent layers. These poorly ordered polymorphs also have a slightly higher first-order spacing than the highly ordered polymorphs.

The triclinic angle $\alpha = 91.6^o$ of well-ordered kaolinite becomes effectively 90^o in these disordered forms, which are therefore at least pseudo-monoclinic. The positions taken up by the Al atom determine the deformations of the octahedral part of the layer structure and thus creates an α angle of 91.6^o. These broken bonds give kaolinite a small amount of permanent negative layer charge and finite exchange capacity (Plancon and Zacharie, 1990).

This kaolin group has several members (including kaolinite, halloysite, dickite and nacrite) and a general formula of $Al_2Si_2O_5(OH)_4 \bullet nH_2O$. The different minerals are polymorphs (Bailey, 1980; Velde, 1992). The general structure of the kaolin group is composed of silicate sheets (Si_2O_5) bonded to aluminum oxide/hydroxide layers ($Al_2(OH)_4$) resembling gibbsite layers. The silicate and gibbsite layers are tightly bonded together with only weak bonding existing between the silicate-gibbsite paired layers (Allen and Hajek, 1989). The term kaolins will be used in this thesis to refer to these minerals unless a particular polymorph is discussed.

Properties of kaolin minerals: Kaolin is formed from the decomposition of silicates under moderately to strongly acid, leaching conditions which remove the more soluble metal ions, such as Ca, Mg, Na and Si. The residual aluminum and silicon crystallize to form kaolin (Brady and Ray, 2002) and some trace elements e.g. Cr may also be retained by kaolin (Singh and Gilkes, 1992a). The extent to which other metals are retained in the kaolin structure and is a focus of this thesis. Many soils of the tropics have a clay fraction dominated by minerals of the kaolin group with various amounts of sesquioxides (Juo, 1980; McCrea *et al.*, 1990).

Many properties of these soils are related, directly or indirectly, to the dominant kaolin mineralogy of the soils. The chemical properties include low pH and weak pH buffering, Al toxicity, low available P and high P fixation capacity, deficiencies of Na, Ca, Mg, K and micronutrients, and low cation exchange capacity (CEC) (Schwertmann and Herbillon, 1992). Soil physical processes, particularly those relating to aggregation and dispersion, may depend on the crystal properties of kaolin and some kaolinitic soils exhibit poor structure whereas other kaolinitic soils are well structured (e.g. Oxisols) (Chittleborough and Walker, 1988; Dixon, 1989). However, despite the dominance of kaolin in soils of the Tropics, little is known of the extent to which the properties of kaolin vary in tropical soils and how this variation affects soil properties.

The structure of kaolin is simple but it has complex crystal and surface chemistries. The CEC of kaolin is partly pH dependent (Ferris and Jepson, 1975), and is due to both ionic substitution and pH dependent surface charge on several distinct surfaces (Bolland *et al*., 1976). The increase in CEC with increasing structural defects in kaolin can be associated with a higher surface charge density at the surface of kaolin crystals (Tari *et al.,* 1999). Kaolin in highly weathered soils often has a low degree of structural order (henceforth referred to as "crystallinity") and small crystal size, with a high specific surface area (Hughes and Brown, 1979; Montes *et al.,* 2002 Varajao *et al.,* 2002). Consequently kaolin may provide an appreciable proportion and probably most of the cation exchange capacity in the solum of kaolin-rich tropical soils. The crystallinity, chemical and morphological properties of kaolin are considered to be mostly consequences of pedo-environmental factors of soil formation, although distinct parent mineral influences can be recognized in some instances *e.g.* epitactic and topotactic replacement of mica by kaolin (Gilkes and Suddhiprakarn, 1979; Singh and Gilkes, 1992c). Several workers have suggested that the degree of defect structure in kaolin may be due at least partly to the extent of replacement of Al^{3+} by Fe^{3+} ions (Mestdagh *et al.,* 1980; Brindley *et al.,* 1986). Cases *et al.* (1986) concluded that decreasing "crystallinity" is usually accompanied by an increase in specific surface area. Parent rock type may also play an important role, as more low-defect kaolin often occurs in subsoils of well drained soils over granitic or siliceous sedimentary rocks (Jungerius and Levelt, 1964; Webster, 1965). Pedoclimate may affect the nature of soil kaolins; for example structural order decreases as the length of the dry season decreases (Hughes and Brown, 1979).

Very little is known of the crystal chemical properties of kaolin in Thai soils despite kaolin often being the dominant clay mineral (Trakoonyongcharoen, 2005; Trakoonyongcharoen *et al.,* 2006). Hart *et al.* (2003) investigated kaolin in purified samples from 10 Thai soils including Alfisols, Ultisols and Oxisols on parent materials as diverse as granite, basalt, sandstone, shale, limestone and mixed colluvium. They observed that very small, high-defect and iron substituted kaolin was the dominant form of this mineral but no systematic association of kaolin properties and soil type/parent rock could be identified due to the small number and great diversity of the soil samples.

Genesis of kaolin: Many soils of the humid Tropics have clay fractions dominated by minerals of the kaolin group and sesquioxides (iron and aluminum oxides) (Schulze *et al.,* 1999). Among the factors advanced to explain variations in kaolin crystallinity have been profile drainage, parent material and profile depth. Well crystalline kaolin occurs at depth in well drained soils formed over granitic and sedimentary rocks. All members of the kaolin group form primarily during hydrothermal alteration or weathering of feldspars under acid conditions; but kaolinite and halloysite are probably the only members formed in soils (Swindale, 1975; Deer *et al*., 1978; Hughes and Brown, 1979; Manning, 1995).

Kaolinite is a product of a chemical change, which affects an aluminous precursor phase, such as feldspar or muscovite (Manning, 1995).

The process of kaolinization of alkali feldspar is:

$$\underset{\text{Feldspar}}{2(\mathrm{Na,K})\mathrm{AlSi_3O_8}} + \mathrm{H_2O} \longrightarrow \underset{\text{Kaolinite}}{\mathrm{Al_2Si_2O_5(OH)_4}} + \underset{\text{Quartz}}{4\mathrm{SiO_2}} + 2(\mathrm{Na,K})^+$$

Weathering occurs at low temperature (up to 40°C) and close to the Earth's surface. The depth of weathering profiles depend on the age, topography, climate and via various intermediate minerals and configuration of the unsaturated zone.

Processes of the weathering from various precursors are represented by the examples,

1. Feldspar ⟶ Smectite (or illite) ⟶ Halloysite (Kaolinite)
2. Biotite ⟶ Mixed-layer minerals/Vermiculite ⟶ Kaolinite

Hydrothermal alteration of aluminosilicate-bearing rock is a widely accepted mechanism for the formation of kaolinite. Velde (1992) concluded that the kaolin group of minerals are stable at relatively low pH. Kaolin is unstable above approximately 300°C. The hydrothermal formation of kaolin from alkali feldspar in an acid solution via muscovite (or illite) via an intermediate product in aqueous solution can be regenerated by the following equations (Dixon, 1989; Velde, 1992).

$$\underset{\text{Orthoclase}}{3KAlSi_3O_6} + 2H^+ \longrightarrow \underset{\text{Muscovite}}{KAl_3Si_3O_{10}(OH)_2} + \underset{\text{Quartz}}{6SiO_2} + 2K^+$$

$$\underset{\text{Muscovite}}{2KAl_3Si_3O_{10}(OH)_2} + 2H^+ + 3H_2O \longrightarrow \underset{\text{Kaolin}}{3Al_2SiO_5(OH)_4} + 2K^+$$

Halloysite

Halloysite has a general composition of $Al_2O_3{\bullet}2SiO_2{\bullet}4H_2O$, and it is similar in structure to kaolinite. Halloysite is often tubular in form. This is contrast to kaolinite, which commonly consists of hexagonal plates (Tan, 1993). The formation and stability of halloysite may follow the weathering sequence (Dixon, 1989):

Igneous rock ⟶ Montmorillonite ⟶ Halloysite ⟶ Metahalloysite ⟶ Kaolinite

Halloysite has a single layer of water molecules between individual aluminosilicate layers, which increases the basal spacing to approximately 0.1 nm (Manning, 1995).

1.1.2 Smectite

Smectite minerals play a major role in establishing the chemical and physical properties in many soils throughout many climatic zones of the world. Their most important chemical properties include high CEC, ion selectivity and molecular sorption (Table 2). Their most important physical properties include expansion and collapse, retention of large quantities of water, small particle size and extremely large specific surface area. The chemical properties aid in the maintenance of soil fertility, although their physical properties sometimes make cultivation and management of soil difficult. Most smectites are stable in neutral, poorly drained environments and in soils where leaching of Si and bases is restricted (Allen and Hajek, 1989; Borchardt, 1989; Velde, 1992).

1.1.3 Vermiculite

Vermiculite occurs in soils in all climate zones and is commonly an alteration product of muscovite, biotite or chlorite. Vermiculite occurs in all soil orders, although it is more abundant in soils of temperate and subtropical zones than in soils of tropical zone. Vermiculite can form in many different soil environments where either mica or chlorite is present in the parent material (Allen and Hajek, 1989; Douglas, 1989; Velde, 1992).

1.1.4 Inhibited Vermiculite

Inhibited vermiculite is abundant in more acidic highly weathered soils. The association of inhibited vermiculite with abundance of kaolin can be related to the occurrence of mica and feldspar in the deeper horizons of the soils. In the more acidic surface soil horizons the interlayer K has been removed from mica by exchange and replaced by aluminum hydroxide polymers (Norrish and Pickering, 1983). Rich (1968) suggested that moderately acidic conditions, low organic matter content, oxidizing conditions and frequent wetting and drying cycles are optimal environmental conditions for pedogenic inhibited vermiculite. Harris *et al.* (1980) found that the amounts of inhibited vermiculite increased with age relative to kaolin in surface soils sampled along a chronosequential transect. Usually the interlayer region consists of both hydrated cations and hydroxide polymers. Frequently, inhibited vermiculite is relatively abundant in surface horizons and decreasing in abundances with depth (Barnhisel and Bertsch, 1989).

Table 2 Cation exchange capacity (CEC), anion exchange capacity (AEC) and specific surface area (SSA) of clay minerals

Mineral	CEC (cmol kg^{-1})	AEC (cmol kg^{-1})	SSA (m^2g^{-1})
Kaolinite	2-15	7-20	5-30
Halloysite	5-50	n.a.	10-45
Smectite	80-150	20-30	800
Vermiculite	100-200	4	400
Illite	10-40	4-17	20
Chlorite	10-40	5-20	10

n.a.= not available
Sources: Bain and Smith (1987); Orlov (1992); Richard *et al.* (1997)

1.1.5 Illite

Illite is essentially a group name for non-expanding, clay-sized, micaceous minerals with soil illite being aluminosilicate dioctahedral. It is structurally similar to muscovite in that its basic unit is a layer composed of two inward-pointing silica tetragonal sheets with a central octahedral sheet (Sroden and Eberl, 1984).

However, illite has on average slightly more Si, Mg, Fe, and water and slightly less tetrahedral Al and interlayer K than does muscovite (Bailey, 1980). The weaker interlayer forces caused by the lower layer charge and fewer interlayer cations in illite allows for more variability in stacking (Grim, 1968). Illites, which are the dominant clay minerals in argillaceous rocks, formed by the weathering of silicates (primarily muscovite and feldspar) and through the alteration of other clay minerals (Deer *et al.,* 1978). Formation of illite is generally favored by alkaline conditions and by high concentrations of Al and K in solution. Glauconite is an iron rich illite which forms authigenically in marine environments and occurs primarily in pelletal form (Allen and Hajek, 1989; Fanning *et al.*, 1989; Velde, 1992). Members of the illite group are characterized by intense 0.1 nm 001 and a 0.33 nm 003 peaks that remain unaltered by ethylene glycol or glycerol solvation, potassium saturation, and heating to 550 degrees C (Fanning *et al.*, 1989)

1.1.6 Other minerals

The clay fraction may also contain other minerals including palygorskite-sepiolite, which are chain-structured clay minerals, quartz, sesquioxides, titanium oxides, pyrophyllite, talc, sulfates, carbonates and phosphates (Allen and Hajek, 1989). Residual quartz, zircon, ilmenite, and other resistant primary minerals present in soils are mostly coarse in size (> 5µm). They have low specific surface area and do not exhibit colloidal properties (Allen and Hajek, 1989; Drees *et al.*, 1989; Milnes and Fitzpatrick, 1989).

Iron oxide minerals: The commonest sesquioxide minerals in most soils are the iron minerals hematite (α-Fe_2O_3) and goethite (α-FeO.OH) (Schwertmann, 1988; Schwertmann and Taylor, 1989; Kämpf *et al.,* 2000; Cornell and Schwertmann, 2003).

Goethite (α-FeOOH) occurs in almost every soil type and climatic region (Singer *et al.,* 1998). It is responsible for the yellowish brown (7.5YR-10YR) colour of many soils and weathered materials. Low temperature, humid climate, pH near 4, high water activity, and high organic matter are among the conditions that favour goethite formation (Schwertmann and Taylor, 1989; Schwertmann, 1993; Brennan and Lindsay, 1998).

Hematite (α-Fe_2O_3) is usually bright red (5R-2.5YR) and is responsible for the red colour of many soils even when considerable goethite is present (Bigham *et al.*, 1978; Torrent *et al.*, 1983; Torrent and Schwertmann, 1987). Hematite forms in relatively dry, warm soils more slowly than in moisture environments (Ibanga *et al.*, 1983).

Other oxide groups: Manganese oxides and hydroxides are important constituents in soils (D'Amore *et al.*, 2004). Birnessite and vernadite (MnO_2) are common Mn minerals in soils. Titanium oxides occur as both inherited and pedologically formed species with ilmenite, rutile and anatase being the most common (Milnes and Fitzpatrick, 1989). Rutile is ubiquitous in soil derived from sediments of mixed provenance. It is sometimes used as an index mineral in soil genesis studies because of its pronounced stability in soil environments (Taylor *et al.,* 1983). Anatase is the most common TiO_2 mineral in soils. It is a major constituent in some highly weathered soils on mafic rocks such as in Hawaii (Walker *et al.*, 1969), and is ubiquitous (1-4%) in Western Australia laterites (Anand and Gilkes, 1984). The presence of quartz is consistent with the quite highly weathered condition of most of these soils (Dixon, 1989).

2. Genesis and properties of Alfisols

Alfisols occupy about 13 percent of the land area of the planet (Wilding, 2000; FAO, 2001). Because of their good natural fertility, location in humid and subhumid regions, and responsiveness to good management they are widely used for agriculture. Most Alfisols are in temperate regions, but these soils are also extensive in tropical and subtropical zones (Hallmark and Franzmeier, 2000; Buol *et al.*, 2003).

2.1 Definitions

According to Soil Taxonomy (Soil Survey Staff, 1999; Ahrens and Arnold, 2000), Alfisols form in semiarid to humid areas and have a clay enriched and nutrient enriched subsoil (argillic horizon). They commonly have a mixed vegetative cover but were originally dominantly under forest. Alfisols have the potential to be very productive if conserved, but can also degrade rapidly if eroded. The diagnostic features are the ochric epipedon and the argillic horizon. A typical horizon sequence would be A, E, Bt and C (Soil Survey Staff, 1999; Hallmark and Franzmeier, 2000).

Alfisols are generally moist and contain an argillic horizon, that is, a subsoil horizon in which clay has accumulated. This horizon of clay accumulation forms because clay particles are washed from the upper horizons (eluviation) and deposited in the lower ones (illuviation) (Arnold and Riecken, 1964; Allen and Hole, 1968; Nielsen and Hole, 1964).

Alfisols are defined as soils that have an argillic horizon and with base saturation of more than 35 percent (Soil Survey Staff, 2003).

2.2 Environmental conditions

2.2.1 Climate

Alfisols may develop from calcareous parent material in a humid climate or in strongly weathered parent material in a subhumid climate where the rain does not penetrate deep enough into the soil to leach out the exchangeable cations. In some subhumid areas, the effectiveness of the rain in leaching the soil is reduced by salt spray (Guerrero, 1963; Kheoruenromne, 1991; Buol *et al.*, 2003).

The climatic conditions under which Alfisols form are thermic or warmer and mesic or cooler. Therefore, most Alfisols are in temperate regions, but these soils are also extensive in tropical and subtropical zones. Alfisols can occur generally in zones with a temperature ranges from below 0^{o}C to above 22^{o}C. Important for the development of Alfisols are periods of high moisture content and sufficiently high soil temperature, to break down the primary mineral components and to leach the weathered products, and a period of relatively low moisture content which permits the precipitation and accumulation of secondary minerals. Most Alfisols have a udic, ustic, or xerix moisture regime, and some experience aquic conditions, but they are not known to occur under a perudic moisture regime (Kheoruenromne, 1991; Soil Survey Staff, 2003). The suborder Aqualfs requires higher soil moisture conditions compared to other suborders of the Alfisols (Soil Survey Staff, 2003).

2.2.2 Vegetation

Most Alfisols form under broadleaf deciduous forest, but they also occur under grassland and prairie vegetation. In forested ecosystems, the trees deliver the bulk of their annual production of organic matter above ground, which is different from grassland soils. In those ecosystems the organic matter is enriched by the huge root system of the grass or prairie cover. While present vegetation may be deciduous forest, earlier vegetation may have been grass or conifers (Brady and Ray, 2002; Buol *et al.*, 2003).

2.2.3 Relief

In most Alfisols, drainage is not restricted with the water table occuring below the solum during major portions of the nonfrozen period. For instance, the suborder Aqualfs is often functionally related to landscape position. Alfisols develop under several drainage conditions ranging from excessive on hill crest and steep slopes (e.g. Lithic Hapludalfs) to poorly drained footslopes and level plains (e.g. Albaqualfs). Alfisols do not develop on very steep slopes, alluvial floodplains, or in very poorly drained depressions. High elevations combined with limited rainfall favor Alfisol formation in the Tropics (Kheoruenromne, 1991; Buol *et al.*, 2003).

2.2.4 Parent Material

The parent material has a major impact on the formation of clay minerals within Alfisols. The resistance to weathering and the composition of primary minerals in combination with the other soil forming factors determine which clay minerals are formed. Generally, a wide variety of clay minerals including kaolin, illite, smectite and vermiculite can occur in Alfisols (Yoothong *et al.*, 1997; Kaket *et al.*, 2006). It should be stressed that several clay minerals do have a potential to adsorb exchangeable bases (high CEC), which is a criterion that should be met to qualify as an Alfisol.

2.2.5 Time

Most Alfisols are present on relatively old landscapes (Kheoruenromne, 1991; Buol *et al.*, 2003). Alfisols need a long period of time for development. Several studies postulate that the time to develop Alfisols is at least 200 y, at which time development of an argillic horizon has commenced, to 1000 y for development of a clear expression of an Alfisol profile, and even longer periods, depending on the other soil forming factors (Buol *et al.*, 2003).

2.3 General properties

Alfisols in tropical region vary considerably in their chemical and physical characteristics (Brady and Ray, 2002; Buol *et al.*, 2003). Permeability and bulk density are relatively low in most Alfisols. Alfisols naturally have a high base saturation so they are generally not deficient in Ca, Mg and K. However, the high base saturation of Alfisols does not necessary mean that they are rich in bases (Singh, 1991). Many tropical Alfisols have a low cation exchange capacity and exchangeable cations as they also contain low activity clay such as kaolinite (Moncharoen and Vijarnsorn, 1978; Soil Survey Staff, 1999; Ahrens and Arnold, 2000; Buol *et al.*, 2003). Some Alfisols may have high activity clay minerals such as smectite, vermiculite, illite or chlorite as the dominant clay species (Kheoruenromne and Suddhiprakarn, 1984; Suddhiprakarn *et al.*, 1985; Yoothong *et al.*, 1997). The CEC of a soil indicates its ability to exchange cation and retain nutrients and it is a measure of the soil quality (Bear, 1965). The type of clay minerals and their relative amount in the clay fraction should reflect the CEC values of the soils (White, 1987). For example, the CEC of the kaolin is relatively low (about 10 cmol kg^{-1}), whereas the CEC is high for smectite and vermiculite. So, the presence of smectite and vermiculite could increase the CEC of the soils (Yong *et al.*, 1992).

3. <u>Genesis and properties of Ultisols</u>

3.1 Definitions

Ultisols are soils that have an argillic or kandic horizon as a subsurface horizon enriched with illuvial clay, with low base saturation (Soil Survey Staff, 1999; West and Beinroth, 2000). They have lower pH and organic matter than Alfisols and often are redder in colour. They have lower pH and organic matter than Alfisols and often are redder in colour. With soil amendments they are productive (Brady and Ray, 2002; Buol *et al.*, 2003). The surface soil usually is coarser-textured, making it easy to till, while the clay-enriched subsoil stores water for plant roots. A typical horizon sequence would be A, E, Bt, BC, C. These soils form either in a wetter climate or in a more strongly weathered parent material than do Alfisols. Leaching has removed most of the exchangeable cations from the soil profile. As a result, the soils are strongly to extremely acid throughout. Ultisols have less than 35 percent base saturation in the lower part of the argillic horizon (Kheoruenromne, 1991; Brady and Ray, 2002; Soil Survey Staff, 1999, 2003).

3.2 Environmental conditions

3.2.1 Climate

Ultisols are formed in climatic regions where precipitation exceeds potential evapotranspiration during some periods each year. Also, the precipitation amount has to exceed the water storage capacity of the soil for some time of the year to allow water to percolate through the soil. This is essential to maintaining the low base status. Ultisols are found in tropical areas, where they tend to have somewhat finer textured E horizons, containing more organic matter and iron, than do the majority of Ultisols formed in temperate climate. Ultisols also may form in frigid soil temperature regimes. Xeric, perudic, udic, ustic, and aquic soil moisture regimes are present in various Ultisols (Miller, 1983; Kheoruenromne, 1991; Brady and Ray, 2002; Buol *et al.*, 2003).

3.2.2 Vegetation

Many Ultisols formed under forest vegetation (e.g. North Australia is under mixed hardwood, pine, oak, hickory forest) although savanna or even swamp vegetation is possible (Kheoruenromne, 1991; Brady and Ray, 2002). Because of their low base status most Ultisols are used for timber production but they are also used in agriculture, where liming and fertilization is required to decrease acidity and increase soil fertility. Where adequate agricultural management is applied Ultisols are quite productive (Miller, 1983; Soil Survey Staff, 1999; Brady and Ray, 2002).

3.2.3 Relief

There are no relief limitations to where Ultisols might form. They may occupy hillslopes or level upland areas. The position they occupy is controlled by the relationship between geomorphology and other factors of formation and the resulting rates and degree of expression of pedogenic processes (Allen and Fanning, 1983; Miller, 1983; Buol *et al.*, 2003).

3.2.4 Parent Material

Almost all Ultisols form in acidic parent materials containing few weatherable minerals. Extensive leaching over a long period is characteristic of Ultisols formed in more basic parent materials (Buol *et al.*, 2003).

3.2.5 Time

The time required to develop Ultisols depends on the rate of pedogenic processes. A Pleistocene or older age is assigned to most parent materials or land surfaces where Ultisols occur. The geologic age of parent materials, however, serves only to fix an absolute maximum on possible periods of time involved in soil formation. The actual time periods involved may be, and generally are, much less (Miller, 1983; Kheoruenromne, 1991; Buol *et al.*, 2003).

3.3 General properties

Most Ultisols are formed in weathered parent rock thus the subsurface horizons are underlain by a saprolite zone. A major characteristic of Ultisols is low base saturation throughout the soil profile with slightly higher base contents in the upper soil horizon due to biocycling (Brady and Ray, 2002; Buol *et al.*, 2003). The low base saturation status is mainly due to formation in parent material high in silica but low in bases. In some soils, the low base status results from intense leaching of parent material initially high in content of weatherable minerals, while in others, a low base status and small quantities of weatherable minerals were initial parent material characteristics. In many Ultisols there are continuous losses of bases through leaching and erosion, therefore, the base saturation remains low. Associated with low base (low nutrient) content is a high soil acidity. Surface horizons rarely have pH values less than 5.0 or greater than 5.8 (Buol *et al.*, 2003; Grunwald, 2005). In general, the pH values decrease with depth to a minimum of 4.0 to 5.5 in the argillic horizon. In highly weathered and leached Ultisols a decreasing pH is evident throughout the solum (Soil Survey staff, 1999; Grunwald, 2005).

Organic matter content and the thickness of the surface horizon increase in most Ultisols with decreasing internal soil drainage and aeration. The organic matter content of many Ultisols is low compared to other soil orders such as Mollisols or Alfisols.

4. Geochemistry of Alfisols and Ultisols

Element concentrations in soils are mainly related to the chemical composition of parent materials and the change due to weathering which differ for soil under various climatic environments (Aubert and Pinta, 1977; Topp *et al.*, 1984; Alloway, 1995; Kabata-Pendias and Pendias, 2001). The importance of pedogenic processes on trace element concentration specific to each soil order is secondary to the influence of parent material. Mineral content of the parent material is one of the most important factors governing the amount of trace elements in a soil, irrespective of classification or the extent of weathering (Lee *et al.*, 1997). Soils in most orders can have a diversity of parent materials affecting their mineralogy and texture, and also their total trace element concentration (Burt *et al.,* 2003). The spatial distribution of elements during weathering is particularly complicated because elements are affected by various pedogenic processes including dissolution of primary minerals, formation of secondary minerals, redox processes, transport of materials and ion exchanges (Middleburg *et al.,* 1988).

Many studies have examined relationships among elements (major and trace) and between elemental concentrations and other soil properties in non-contaminated soils (Shuman, 1985; Pardue *et al.,* 1992; Holmgren *et al.,* 1993; Ma *et al.*, 1997; Chen *et al.,* 1999). In Alfisols and Ultisols, trace element concentrations reported were as follows: Mn > (Zn, Cr, Ni, Cu)> (Pb, Co) > (Cd, Hg). For Alfisols, Cr, Cu, Mn, Ni, Zn and Hg concentrations were higher in subsoil than in topsoil, whereas the reverse applies for Co and Pb, suggesting that metals partition within a pedon depending on their affinities for organic matter and clay size inorganic constituents (Liu *et al.,* 2002; Burt *et al.,* 2003). Zn and Ni were positively related to clay content in US Alfisols but organic carbon and free iron (Fe_d) concentrations were not related to heavy metal concentrations (Lee *et al.,* 1997). In Bangkok, Thailand, the average heavy metal (Cd, Cr, Cu, Fe, Mn, Ni, Pb and Zn) concentrations in soils were lower than in soils in Manila, London or Hamburg but were comparable to those in Hong

Kong soils (Wilcke *et al.*, 1998). Relatively little is known of the trace element concentrations in Thai Alfisols and Ultisols.

Ultisols are widespread in Thailand covering about 45 percent of the total country area. They occur in every region, especially in the Northeast Plateua, Southeast Coast and Peninsular Thailand. Alfisols (9% of the total area) occur in drier areas of the country, mainly in the Central Highlands, Central Plain, North and West Continental Highlands and some parts of the Northeast Plateau. Ultisols have developed over deep saprolite of diverse rock types and many of them developed on colluvium and alluvium (Kheoruenromne, 1991).

By definition based on soil taxonomy (Soil Survey Staff, 1999), Alfisols have base saturation of greater than 35 percent, whereas Ultisols contain lower base saturation (<35%). However, there are similarities between these two soil orders in Thailand. They generally have low CEC and common clay minerals (Changprai, 1987; Kheoruenromne, 1991; Yoothong *et al.,* 1997). Alfisols are often composed of smectite and inhibited vermiculite is rather common in the Ultisols. In most cases, both soil orders are found to be dominanted by kaolin in their clay fraction (Kheoruenromne, 1991). Nonetheless, there have been no systematic studies on the concentration of major and trace elements in these two soil orders to be a baseline for further studies on these soils.

5. Minerals in clay fraction of Alfisols and Ultisols

5.1 Alfisols

In the Tropics, many and perhaps most Alfisols have a low CEC-kandic horizons and exchangeable cations as they also contain low activity clay such as kaolin (Moncharoen and Vijarnsorn, 1978; Kheoruenromne, 1991; Bhattacharyya *et al.,* 1992; Soil Survey Staff, 1999). The abundance of relatively unweathered clay minerals provides many temperate zone Alfisols with a high CEC (Buol *et al.,* 2003). Some Alfisols may have high activity clay minerals such as smectite, vermiculite, illite or chlorite as the dominant clay species (Kheoruenromne and Suddhiprakarn, 1984; Kheoruenromne, 1991; Yoothong *et al.*, 1997). Potassium and ammonium fixation is an important process to be considered in Alfisols that have notable contents of vermiculite and slightly weathered mica in the clay fraction (Buol *et al.,* 2003).

5.2 Ultisols

Clays in Ultisols usually consist mostly of the 1:1 type (kaolin) or gibbsite. There are less clay minerals of the 2:1 type. Therefore, the cation exchange capacity, exchangeable cations and water holding capacity are relatively low in most Ultisols (Miller, 1983; Kheoruenromne, 1991). These limitations can be overcome by the application of lime to decrease acidity and the application of fertilizers to add bases to the soils, but Ultisols are commonly not as productive as Mollisols or Alfisols. Clay content increases from A, E or upper B horizons to a maximum in the upper part of the argillic horizon, then decreases regularly with depth into the C horizon (Kheoruenromne, 1991; Brady and Ray, 2002; Buol *et al.*, 2003).

Iron oxides, released from other minerals through weathering or inherited as such from the parent material, are important pedogenic and taxonomic indicators in Ultisols. Goethite is the dominant crystalline form of iron oxides in most Ultisols and commonly associated with lesser quantities of hematite, maghemite, and magnetite (Kheoruenromne and Suddhiprakarn, 1984; Suddhiprakarn *et al.,* 1985; Schwertmann and Taylor, 1989). The amounts of hematite are generally greater in Ultisols developed from basic rocks and are more abundant in tropical than in temperate regions. This accounts for the red colour in well-drained tropical Ultisols as compared to that in other regions. The red or yellow colours of the argillic horizon and underlying materials in many Ultisols are due to iron oxides (Schwertmann and Kämpf, 1985; Barrón and Torrent, 1986; Schwertmann and Taylor, 1989; Singh and Gilkes 1992d; Schwertmann, 1993). In most Ultisols, various proportions of the soil are comprised of reddish and grayish or light-coloured mottles (Grunwald, 2005). This condition is normally associated with segregation of Fe-oxides due to alternating oxidation and reduction. Reduction forms of relatively soluble Fe^{2+} which may migrate to more oxidizing locations before reoxidization or they may reoxidize and precipitate *in situ* on existing Fe^{3+} compounds. Repetitions of the cycle result in development of zones with high and low free iron contents corresponding to the reddish and grayish colours (Schwertmann and Taylor, 1989; Schwertmann, 1993 Brennan and Lindsay, 1998). The behavior of Mn in oxidizing and reducing environments is analogous to that of Fe. Continued segregation and concentration of oxides by alternating oxidizing and reducing conditions forms plinthite or fragipan (Grunwald, 2005).

6. Relationships of minerals in clay fractions with geochemistry of Alfisols and Ultisols

Clay minerals may contain negligible amounts of trace elements as structural components, but their sorption capacities for trace elements play the most important role (Millot, 1971). The affinity of trace ions for the clay surface has been examined by many investigators (Barrer, 1978; Kabata-Pendias and Pendias, 2001) and while many aspects have been clarified, still much remains to be learned about the adsorption processes (Andersson, 1977a, 1977b; Bowden *et al,* 1980; Abd-Elfattah and Wada, 1981; Kabata-Pendias and Pendias, 2001). Although clay minerals vary in chemical composition and in their nature, some general surface properties can be given for each mineral group (Table 3) that are responsible for values of the specific surface area of soil materials (Ahlrichs, 1972; Dobrzanski *et al.,* 1977). The capacity values (CEC) vary with the type of clay in the following sequence: smectite, vermiculite> illite, chlorite> kaolinite> haloysite (Bain and Smith, 1987; Orlov, 1992; Richard *et al.*, 1997).

The ability of the clays to bind the metal ions is correlated with their CEC (Czarnowska, 1974; Alloway, 1995), and usually the greater the CEC, the greater the amounts of cation adsorbed (Kabata-Pendias and Pendias, 2001).

Table 3 Microcations sorption properties of some clay minerals

Mineral	Total sorption of microcations (µM g^{-1})	Sorption of cations (mg kg^{-1})			
		Cd^{2+}	Mn^{2+}	Zn^{2+}	Hg^{2+}
Kaolinite	30-70	3.1	3.5	3.4	0.46
Smectite	390-460	60-86	72-116	88-108	0.4-2.2
Vermiculite	-	98	92	98	9.7
Illite	65-95	-	-	-	-

Source: Kabata-Pendias and Pendias (2001)

Clay containing exchangeable transition metals cations (mainly Cu, Fe and Co) are known to act as electron or proton acceptors, thus they can be activators in transformations, decompositions and polymerization of the adsorbed organic species. The mechanisms of coprecipitation and mobilization of certain trace cations with alumina of silica gels may play a significant role in their behavior in the particular soil, especially in the tropical climate zone (Korte *et al*., 1976; Jenkins and Jones, 1980). There are suggestions that metal ions (mainly Cu^{2+}) can be substituted for aluminum in the mineral structures, while soluble acid promotes adsorption of Co, Ni and Zn by clay (Tiller, 1968; McBride, 1978).

The strong adsorption of divalent trace cations (Cu, Pb, Zn, Ni, Co, Cd, and Sr) by freshly precipitated alumina gels is suggested by Kinniburgh *et al.* (1976) to play a role determining the availability to plants and the movement of some of the cations through the soil.

The mineral composition of the parent materials commonly governs the amounts of trace elements in a soil, irrespective of soil classification or the extent of weathering (Lee *et al*., 1997). Soils in most orders can have a diversity of parent materials which affects their mineralogy and texture, and also their total trace element concentrations (Burt *et al.,* 2003). The spatial distributions of elements that develop in soil profiles and catena during weathering are particularly complicated because elements are affected by various pedogenic processes including dissolution of primary minerals, formation of secondary minerals, redox processes, transport of materials and ion exchange (Marshall, 1977; Beinroth, 1981; Schwertmann, 1985; Middleburg *et al.,* 1988).

The minerals of the smectite group can expand and contract in response to charge and size of the absorbed cation between the clay platelets (Harris *et al.,* 1980; Barnhisel and Bertsch, 1989). Thus, their sortion capacity will differ when saturated with different cations (Barnhisel and Bertsch, 1989). The microcations sorbed by smectite are also easily released into the liquid phase and, therefore, can be an important pool of micronutrient supply to plants growing in particular soil conditions (Borchardt, 1989). These data are consistent with soil kaolin containing appreciable proportions of the Ni, Cu and Zn and some of the Co and Pb present in these highly weathered soils. This research has also identified a possible role for kaolin as a host for minor elements as structural ions which has significant implications for soil fertility and geochemical exploration.

7. Geology and physiography of Thailand

7.1 Location

Thailand is divided into six physiographic regions according to Moormann and Rojanasoonthon (1972): the Central Plain, Northeast Plateau, North and West Continental Highlands, Central Highlands, Southeast Coast and Peninsular Thailand. The Central Plain occupies the central part of the country and is characterized by the broad alluvial plain, known as the Bangkok Plain. Surrounding the plain is mainly terraces with peneplain and monadnocks (Rua and Nutalaya, 1980; Takaya and Thiramongkol, 1982). The major physiographic features of the Southeast Coast consist of hills and broad terraces. The region is bordered on the North and East by hills and on the South and West by the Gulf of Thailand. The Northern Plateau is bordered on the West and South by hills and on the North and East by Mekong River. The main landscape features are terraces, of which three important levels have been recognized: low, medium and high. The Central Highlands are situated between the Central Plain on the West and the Northeast Plateau on the East. The main landscape features are hills, high plateaus, peneplains and terraces. The North and West Continental Highlands consist of two major landscape units, hills and valleys which are composed of a series of north-south parallel hills. The Peninsular Thailand is composed of a series of north-south oriented parallel mountain ranges. The major landscape features are hills, marine terraces and with peneplains as a minor element of the landscape (Department of Mineral Resources, 1987; Suwanasing, 1992).

7.2 Climate of Thailand

Thailand has three main seasons: a rainy, a cool and dry and a hot and dry season. The eastern part of the Southeast Coast and most parts of the Peninsular Thailand are characterized by the tropical monsoon. The northern mountainous region is characterized by a humid subtropical climate. Most of continental areas of the country is characterized by a tropical savanna climate (Köppen, 1931). Temperature varies between 24-26 ^{o}C in the North, between 28-30 ^{o}C in the central region and between 26-28 ^{o}C in the rest of the country. The average annual rainfall ranges between 1,000 and 4,000 mm (Meteorological Department, 2000). In the southern and southeast regions, however, rainfall is generally the highest compared to that in the rest of the country. As a consequence, the southern and southeast regions do not have a pronounced dry season (Yoothong *et al.*, 1997; Meteorological Department, 2000).

7.3 Seasons in Thailand

As a tropical country, only the wet and dry seasons are succinctly differentiated (Meteorological Department, 2000). This is especially true in the Peninsular South and the lower mainland sections. For the upper mainland sections, however, a greater distinction between temperature differences during the dry season justifies a broad separation of summer and winter. Such a seasonal change is the results of monsoon wind directions, interacting with the location and terrain.

The southwest monsoon carries moisture from the Indian Ocean, normally from May to mid-October; but for the upper mainland sections, it is effective from mid-May to

September. Its initial strength correlates with that of the high pressure center over the Indian Ocean in the Southern hemisphere (Meteorological Department, 2000).

The northeast monsoon which carries moisture from the Indian Ocean is normally active from mid-October to mid-February, sometimes extending into March. It causes a cool, dry season for the upper mainland section. The eastern shore of the lower South, however, receives this northeast monsoon after it has traveled across the South China Sea, picking up moisture and bringing more rain. Therefore, its wet season is effectively extended, with even higher precipitation, until January or later.

The switch-over or inter-monsoon period, from mid-February to mid-May, coincides with the sun's position moving northward across the equator. Increased heat absorption by the land mass and much reduced rain produces a very hot climate, especially for the upper mainland sections. The wind can be very variable, with occasional storms. The lower section of the land often receives southerly winds coming up from the Gulf of Thailand, sometimes referred to as a southeast monsoon, making it generally cooler than the upper mainland (Panichapong and Hemsrichart, 1982; Meterological Department, 1994).

7.4 Soil moisture regimes of Thailand

Moncharoen (1983) distinguished five soil moisture regimes: aquic, peraquic, udic, ustic and perudic. The different regimes occur in the following regions:

1. aquic, in low-lying paddy lands throughout the country;

2. peraquic, in tidal flats and swamp areas, especially along the Gulf of Thailand;

3. udic, in the peninsular region, with an annual rainfall higher than 2,000 mm, and with more than 160 days of rain, as well as in the northern region where hills, between 1,000 and 1,600 m in altitude, are covered with an evergreen vegetation;

4. perudic, in areas above 1,600 m MSL and

5. ustic, in remaining areas of the country.

8. Alfisols in Thailand

The total area of Alfisols of Thailand is about 49.5 square kilometers. Alfisols (9% of the total area) occur in drier areas of the country, mainly in the Central, North and Northeast. Most Alfisols have relatively high pH and have low values of cation exchange capacity, plant available nutrients and medium to high base saturation. The dominant minerals of the clay fraction may be kaolin, vermiculite, illite with lesser amounts of an aluminum interlayered 14 Å clay mineral (HIV) together with sesquioxides (Changprai, 1987). Alfisols in Thailand can be divided into three suborders: Aqualfs, Udalfs and Ustalfs as shown in Table 4.

Aqualfs are seasonally sataurated, or if artificially drained, display evidence of former wetness such as mottles, low chroma, and Fe-Mn concretions. They occur in every region of Thailand. Most Aqualfs have some artificial drainage or other water control and are used as cropland for rice (Kheoruenromne, 1991).

Udalfs are the more or less freely drained than Alfisols that have aquic and ustic moisture regimes. In Thailand, Udalfs have moderate fertility and have brown, yellow or red colour (Kheoruenromne, 1991). Some of the Udalfs on older surfaces are underlain by limestone or other calcareous sediments. Many Udalfs are on stable surfaces and retain most of their eluvial horizon above the argillic or kandic illuvial horizon.

Ustalfs are the Alfisols of subhumid to semiarid regions. They have an ustic moisture regime. They do not have, near surface, redoximorphic features with low chroma or aquic conditions for some time in a normal year. Moisture moves through, these soils to deeper layers only in occasional wet years. The clay in many of these soils is mostly kaolin (Suddhiprakarnm and Kheoruenromne, 1987; Soil Survey Staff, 1999; Hart *et al.,* 2003).

Table 4 Classification of Alfisols in Thailand

Suborder	Great group	Subgroup	Series	Mineralogy class
Aqualfs	Endoaqualfs	Typic	La-ngu Hang Dong	Mixed Mixed
		Typic Plinthic	Phan	Kaolinitic
		Typic Aeric	Lampang	Mixed
		Aeric	San Sai	Siliceous
		Aeric Plinthic	Tha Tum	Mixed
			Deum Bang	Kaolinitic
	Natraqualfs	Typic	Kula Ronghai	Mixed
Udalfs	Paleudalfs	Plinthaquic	Phak Kat	Mixed
	Hapludalfs	Typic	O Lum Chiak	Mixed
Usalfs	Natrustalfs	Aquic	Nong Kae	Mixed
	Kandiustalfs	Grossarenic	Ban Pai	Siliceous
	Paleustalfs	Aquertic	Nam Len	Smectite
		Oxyaquic	Nam Duk	Mixed
		Plinthic	Phayao	Mixed
		Typic (Plinthic)	Sa Kaeo	Kaolinitic
	Rhodustalfs	Typic	Surin	Kaolinitic
			Sikhio	Siliceous
	Haplustalfs	Typic	Kamphaeng Saen	Mixed
		Ultic	Li	Mixed
		Arenic	Nam Phong	Siliceous
		Oxyaquic (Ultic)	Yang Talat	Siliceous
		Oxyaquic (Arenic)	Maha Sarakham	Siliceous
		Oxyaquic	Phetchaburi	Mixed
		Aquic	Wichian Buri	Mixed
		Aquic Arenic	Ubon	Mixed

Source: Kheoruenromne (1991); Soil Survey and Classification Division (1999a, b, c and d)

9. Ultisols in Thailand

The wide variety of soils in Thailand is a result of interactions between the diverse climates, parent rocks and physiography of the country (Kheoruenromne, 1991). Soils exhibit diverse physical, chemical and mineralogical properties. Many Thai soils are Ultisols (45% of the total area) which occur in every region, especially in the Northeast, Southeast Coast and South (Kheoruenromne, 1991; Theerawongse, 1990). Ultisols have developed over deeply weathered regolith or colluvium and old alluvium (Kheoruenromne, 1991; Suwanasing, 1992). Most of them have low cation exchange capacity, base saturation, available phosphorus and potassium. The dominant clay minerals are kaolin and lesser aluminum interlayered vermiculite (Changprai, 1987; Yoshinaga *et al.,* 1989; Hart *et al.,* 2003; Trakoonyingcharoen, 2005). Ultisols in Thailand can be divided into four suborders: Aquults, Humults, Udults and Ustults. Some Ultisols in Thailand and their taxonomy classes are shown in Table 5.Aquults are the Ultisols of wet areas where ground water is very close to the surface during part of each year. These soils are mostly grayish or olivine in subsoil and mainly in alluvium and marine deposits that are of Pleistocene age or older (Suddhiprakarn and Kheoruenromne, 1987; Kheoruenromne, 1991; Soil Survey Staff, 1999; Buol *et al.,* 2003). They are in every region of Thailand and are normally used for rice production.

Humults are the more or less freely drained, humus-rich Ultisols. They generally occur in the North of Thailand. They commonly receive high rainfall but also have a moisture deficit during some seasons (Soil Survey Staff, 1999; Buol *et al.,* 2003).

Udults are the more or less freely drained, relatively humus poor Ultisols that have a udic moisture regime. They are in Southeast Coast and Peninsular regions of the country. Many have been used as cropland (Kheoruenromne, 1991; Soil Survey Staff, 1999; Buol *et al.,* 2003).

Ustults are the more or less freely drained Ultisols that have an ustic soil moisture regime and have a relatively low content of organic carbon (Buol *et al.,* 2003). They occur in the Northeast region. They are commonly used as cropland or pasture (Kheoruenromne, 1991; Soil Survey Staff, 1999; Buol *et al.,* 2003).

Table 5 Classification of Ultisols in Thailand

Suborder	Great group	Subgroup	Series	Mineralogy class
Aquults	Plinthaqualfs	Kandic	Satun	Kaolinitic
		Typic (Aeric)	Yan Ta Khao	Mixed
		Typic	Klaeng	Kaolinitic
	Kandiaquults	Typic	Khok Khain	Mixed
			Roi Et	Mixed
	Paleaquults	Aeric Plinthic	Phen	Kaolinitic
		Aeric (Plinthic)	Renu	Mixed
		Plinthic	Chiang Rai	Kaolinitic
		Aeric Uumbric	Sungai Padi	Mixed
		Typic	Bang Nara	Kaolinitic
	Epiaquults	Aeric (Plinthic)	On	Kaolinitic
	Endoaquults	Typic	Sungai Kolok	Kaolinitic
Humults	Kandihumults	Typic	Khlong Chak	Kaolinitic
	Palehumults	Typic (Kandic)	Doi Pui	Kaolinitic
		Typic	Pak Chan	Kaolinitic
	Haplohumults	Typic	Khlong Teng	Mixed
Udults	Plinthudults	Typic (Kandic)	Khao Khat	Kaolinitic
		Typic (Aquic)	Wang Tong	Kaolinitic
	Kandiudults	Aquic	Sai Buri	Mixed
		Rhodic	Fang Deang	Mixed
		Typic	Nong Khla	Kaolinitic
			Krabi	Kaolinitic
			Pathiu	kaolinitic
	Paleudults	Aquic	Songkhla	Mixed
		Typic	Sawi	Mixed
			Chumphon	Kaolinitic
	Hapludults	Typic	Phato	Mixed
Ustults	Kandiustults	(Oxyaquic)	Korat	Siliceous
		Rhodic	Mea Taeng	Kaolinitic
		Typic	Don Rai	Mixed
			Nong Mot	Kaolinitic
	Paleustults	Plinthaquic	Khemarat	Kaolinitic
		(Oxyaquic)	Chakkarat	Mixed
		Typic (Aquic)	Lam Kao	Mixed
		Typic (Oxyaquic	Phon Pisai	Mixed
		Typic (Kandic)	Ban Chong	Kaolinitic
		Typic	Yasothon	Siliceous
	Haplustults	Plinthaquic	Borabu	Mixed
		Petroferric	Sakon	Mixed
		Kanhaplic	Phu Sana	Mixed
		Typic	Phon Ngam	Mixed

Source: Kheoruenromne (1991); Soil Survey and Classification Division (1999a, b, c and d)