

**A Progress Report on  
the Chemical Composition of Soils in Illinois:  
Cores 1 through 10**

by

Gary B. Dreher, Leon R. Follmer, and Yanhong Zhang

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## **PREFACE**

This progress report presents analytical chemistry data on the first ten soil cores of an anticipated total of 137 soil cores. Presently, there are not sufficient data to allow exhaustive interpretations of the analytical chemistry results. As more data become available throughout this project we will be better able to make final interpretations and to increase our confidence in the correctness of our inferences from the data. The report is primarily intended to provide our analytical chemistry information to the people from whose land the soil cores were collected. It also will be useful to others who are interested in learning about the chemical composition of soils in far-southern Illinois.

We wish to thank the landowners who allowed us to collect soil cores from their property. We are grateful for the interest they showed in our research as expressed by their questions and their desire to observe the coring and field description operations.

## INTRODUCTION

The Illinois State Geological Survey (ISGS) serves the state of Illinois as a depository of geological and chemical information about the geological materials and soils of Illinois. The ISGS also researches important issues that concern the people of the state. In general, these issues deal with the discovery and use of our natural resources, solving environmental problems, as well as serving general educational needs about earth sciences.

The ISGS has a large collection of data that were gathered over the years to serve the needs of research projects conducted for many different purposes. Thus, when questions arise about a new subject such as “the chemistry of Illinois soils,” Survey staff may be able to respond by reviewing available information and reorganizing it in useful formats, for example, data tables, figures and maps, developing a new interpretation based on the data, and presenting the new information in a variety of ways. When sufficient information is not available, however, then new research projects must be organized to address the need. This project was begun to address the general public’s increasing concerns about the chemical and mineralogical composition of soils in Illinois.

The general public is often confused about the differences between soil science and geology. The scopes of these subjects overlap and are interrelated; each presents results and discussions according to the style and terminology of their respective groups of specialists. In an agricultural sense, soil is the earth material (geologic unit) that supports agricultural activities. That is, soil is the material in which plants grow and which consequently becomes modified compared with its “parent material” or its original composition and form. In the geological sense, soil is the surficial material that has been modified over time by reactions caused by natural chemical, biological, and physical agents that cause the chemistry, mineralogy, and morphology of the original material to change through processes known as “soil formation.” To keep these concepts clear, one should think of the soil profile as being superposed over the geologic material; there are both a geologic aspect and a soil aspect in the same volume of material.

This study presents both the geologic and the soil science points of view, merging the two terminologies in a rational way. In principle, soils (soil profiles) and geologic units (often called material or stratigraphic units) are not separate physical entities. They are terms used for the same “surficial” feature from different points of view and in this sense illustrate the professional preferences or “biases” of the two disciplines.

We considered two approaches to studying the chemistry and mineralogy of Illinois soils. The first was to consider the nature of the existing data and see if they were sufficient to meet our needs. In other words, we considered whether our database would be sufficient if we supplemented it with all available data from other sources, such as the U.S. Department of Agriculture-Natural Resources Conservation Service (USDA–NRCS), engineering companies, water-well and other drillers reports, and so forth. Our preliminary assessment showed that there were many data available, but no standard framework or style existed that could be followed. So we chose to build our own comprehensive database in a universal style that tried to avoid the



professional biases of soil science and geology. Consequently, many parts of the project are still in a developmental stage. We plan to incorporate all available data eventually, but initially we chose to start building a database by selecting representative soils to be sampled and analyzed for their chemical and mineralogical composition. This will be done in a way that will enable us to handle all data we can acquire, a feat that was not possible prior to the general availability of computers. Next, we will carry out subsequent studies to fill gaps in the database following a priority plan determined by needs.

This report presents basic data acquired from the first ten (of an anticipated 137) soil cores collected and analyzed. We currently have skeletal descriptions, identifications of the geologic units and soil horizons, and results from a suite of chemical determinations from a selection of 6 samples per core. Mineralogical data are not yet available for the samples, but we have speculated in the “Results” section of this report as to which minerals contain the various elements found in the soils. These speculations were based on the available knowledge about the minerals that generally are found in Illinois soils.

There are several reasons to determine the chemical and mineralogical composition of soils: (1) to provide a coherent geochemical database by which to address the health of the environment (including the assessment of contamination of soils) and in utilizing natural resources (Darnley et al. 1995); (2) “to evaluate the contribution of soil minerals to animal and plant ecology in the State” (Jones 1986) or stated slightly differently, to correlate the toxicity of soils to plants, animals, and humans (Esser et al. 1991); (3) to relate the distributions of elemental concentrations in the state to weathering intensity and sorting of particles according to size by action of wind (Jones 1986); (4) to show the association of trace elements with soil minerals (Esser et al. 1991); (5) to supplement information required to understand the geochemical landscape of Illinois; (6) to provide correlative information for understanding the composition of lake and river sediments; and (7) to provide a well-documented reference collection of cores for other inquiries.

In a previous study conducted by the ISGS, 94 samples of soil were collected from 54 counties in Illinois (Zhang and Frost 2002). Subsamples were retained from depths of 4 to 8 inches and 28 to 32 inches below the surface. The samples were air-dried, disaggregated, riffle-split, ground, and analyzed for major, minor, and trace element composition using the techniques of x-ray fluorescence spectrometry and instrumental neutron activation analysis. Most of the samples for Zhang and Frost’s study were collected in northern Illinois.

The purpose of this research project is to determine the chemical and mineralogical characteristics of soils and underlying unconsolidated materials in Illinois, and to interpret the derived data in light of the geological processes that have acted upon the soils and their parent materials. In addition, this project will expand upon the previously collected data from other sources. A general database of chemical and mineralogical information will be created which can be used in assessing environmental conditions and to help understand the effects of soils on plant health and productivity. The data collected will contribute to our understanding of the chemical, mineralogical, and geological processes that take place during soil development.

## BACKGROUND

The present landscape of Illinois was created during the most recent part of geologic history known as the Ice Age or the Quaternary Epoch (Willman and Frye 1970, Killey 1998). All of Illinois was affected directly or indirectly by the continental glaciers that advanced from the north on about six occasions over the last million years (Follmer 1996). Large-scale glaciation started earlier in Asia, but the glaciers did not grow large enough in North America to advance into the U.S. until about 800,000 years ago. Two of the older glaciations in North America reached to the northern flanks of the Shawnee Hills in southern Illinois during the episode now known as the Illinoian (see Figure 1).

As continental glaciers advanced into the northern U.S., they crushed large amounts of bedrock into silt, sand and pebbles, dislocated boulders, and slabs of bedrock, and transported the debris to the south. Along their path they alternated between erosion of the substrate and deposition of their load. The southernmost extent of continental glaciation in North America lies in Illinois. When the glaciers stopped advancing and began melting away, they dropped their remaining load of rock debris. This resulted in most of the glaciated part of the state being covered by glacial deposits, which are collectively called drift.

The sediment carried by the glaciers tended to level the landscape. Relatively thin veneers of drift were spread across the highlands and thick deposits filled the pre-glacial bedrock valleys with drift up to 350 feet thick. Much of the drift is a homogenous mixture of fine-grained material with pebbles and a few boulders. This type of deposit was originally called boulder-clay and is now commonly called till. Interbedded with till in thick sequences of drift is outwash, which is commonly composed of stratified beds of sand and gravel, with some layers of silt and clay in places.

Beyond the margins of the continental glaciers, meltwater from the glaciers filled the major valleys such as the Mississippi and Illinois Rivers and many other rivers that drain to the south. Erosion caused by the meltwater greatly widened and deepened the major river valleys and then largely filled them with stratified coarse-grained deposits. The surfaces of many of these deposits commonly are above the levels of the modern streams in terraces underlain by sand and gravel that is commonly over 100 feet thick.

The rivers that joined the master meltwater rivers south of the glacial margins were flooded by the rising level of the master rivers during the major glacial events. This caused lakes to form in the lower reaches of these rivers. An example of one of the larger flooded valleys that was formed during the last glaciation was in Gallatin and Saline Counties. The rising level of the Wabash River (caused by the flow of meltwater) caused water to back up into the Saline River lowland and cover most of the region. This lake remained for several thousand years and largely filled the lowlands with stratified silt and clay, which is over 100 feet thick above the original channel of the Saline River. These deposits differ from normal lake deposits and are called slackwater deposits.

Other types of lake deposits commonly are present in glaciated areas. Proglacial lake deposits were formed by advancing glaciers that blocked north-flowing rivers. Another type of lake deposit formed in closed basins on till plains after the glaciers stagnated. In these places variable thicknesses of drift and different rates and degrees of settlement and compaction within the drift caused basins to form.

By studying the types of stratified sediment in a basin, a geologist is able to identify the conditions that prevailed during the formation of the sediment. This information provides a valuable correlation tool for tracing the distribution of important deposits and for finding the best aquifers.

The rise and fall of the meltwater in rivers according to the seasons exposed their very wide floodplains to wind erosion during winter when melting was at a minimum. Sand and silt was blown out of these valleys onto the adjacent uplands to the east. The biggest valley through all of Quaternary time was the Mississippi, and it changed its course several times because of interruptions caused by the glaciers. Large bluffs of eolian (wind-blown) silt accumulated along the eastern margins of floodplains. This silt is very soft when first formed, but with aging it becomes hard enough to stand in vertical exposures. It was first named by German farmers who called it *loess*, meaning “loose soil” in English.

The term *loess* has become the name used by geologists and soil scientists around the world for eolian silt deposits (Follmer 1996). Loess covers all of the Midwest except where it has been eroded away (Figure 2). It is thickest along the Illinois and Mississippi Rivers and thinnest in the Chicago region. In the bluffs north of East St. Louis, the total loess thickness is as great as 80 feet in places, but it becomes thinner to the east. Within a mile east of the bluff crest the average thickness is about 25 feet, and the loess blanket gradually thins to the east until it reaches the influence of another meltwater river. The Kaskaskia and many other mid-size rivers in Illinois were minor sources of loess, as indicated by wind-blown dune sand in and along their valleys, but their contributions were relatively small and had little effect on the regional eastward thinning pattern of the loess.

The next master meltwater river to the east of the Mississippi drainage was the Wabash. Within 10 miles of the Wabash River the loess begins to thicken and grows to over 10 feet thick in the west bluff of the Wabash. As might be predicted, the loess is even thicker on the east side of the Wabash, again indicating that the Wabash was a major source of loess, and confirming that the prevailing winds blew toward the east. Another important loess source in northwest Illinois was from barren till plains that lacked a vegetative cover. A large amount of loess came from deflation (wind erosion) off the glacial deposits in central Iowa (Putman et al. 1988). All glaciated landscapes were barren for some time during the main interval of glacial activity and were subjected to wind erosion. Eventually, these landscapes stabilized and became vegetated; that is, landscape disturbance (erosion and sedimentation) slowed enough for vegetation to be re-established.

In only a few places are Wind-blown dune sand deposits are significant in only a few places. The

largest of these are in Mason County; in the Green River Lowland of Whiteside, Henry, and Lee Counties; and the eastern Kankakee County area. During the peak of the last glaciation favorable conditions allowed “sand seas” to form. When the climate changed back to a warmer condition about 10,000 years ago, the dune-sand deposits were stabilized by vegetation. In geologic terms, this change in conditions marked the beginning of the last geologic interval called the Holocene or “Recent” in common terms.

The dune sand and loess have physical properties and chemical compositions that make them especially good parent materials for agricultural soils. Both materials contain a blend of minerals that, during the early stages of weathering, produce chemicals and by-products that make fertile soils. Weathering processes, particularly the chemical reactions called hydrolysis and oxidation, cause the dark-colored minerals and the feldspars of rock particles in the sand or loess to be slowly altered, releasing ions (sodium, calcium, iron, magnesium, potassium, phosphorus, sulfur, and many others) and producing by-products (clay minerals) that cause the cation exchange capacity (CEC) of the developing soil to increase over time. The silt-size particles in loess generally are more altered or weathered than the coarser grains in the dune sands and because loess deposits can hold significantly more water than dune sand, the water is available to plants long after a rainfall.

The chemical composition of the Midwestern glacial deposits is near optimum for plant growth. In general, the composition of most glacial deposits can be characterized as containing variable proportions of five types of pulverized rocks: igneous, metamorphic rocks of the Canadian shield, and the sedimentary rocks sandstone, limestone, and shale. The only significant drawback to the glacial deposits as parent material for soils is the presence of so much limestone in the mixture; that the soils developed in the glacial deposits are alkaline. After aging for a few thousand years, however, the excess limestone is leached from the upper horizons of the soils and they can become more acidic.

In areas where the bedrock or the glacial sediment overridden by the glacier is dominated by one rock type or a limited combination of types, the resulting soil developed in this material may be unusually coarse or fine. For example, in areas where the glacier rode over extensive areas of shale or fine-grained lake deposits, the resulting soil is likely to be unusually fine-grained and rich in the clay minerals that are most likely to be present.

Loess deposits cover all the uplands of Illinois except where erosion has removed them. Such erosion was common on slopes along valleys, but the loess is missing in some places for reasons we do not fully understand. Exceptionally large floods that occurred near the end of the glacial conditions scoured away both the loess cover and some of the bedrock in some parts of Illinois. The best example of this in Illinois is along the Kankakee River west of Kankakee. When a large moraine in northern Indiana that had impounded a large pro-glacial lake was overtopped and failed, the catastrophic flood that coursed down the river overflowed the normal floodplain, transgressed onto the lower parts of the upland, and removed all the loose sediment down to bedrock over a large area. The height and scope of the erosion caused by the Kankakee Flood (Willman and Frye 1970) may seem incredible, but are real.

All major loess deposits were formed in direct response to the glacial environment (Follmer 1996). The link to the aggradation, or building up, of outwash in large valleys is clear. As outwash accumulated in the master valleys it was subjected to annual wind storms. Sand dunes on or adjacent to these surfaces provide direct evidence for wind erosion. The loess deposits are thickest along the main river valleys (Figure 2), which provide further evidence that outwash-valleys were the main sources for the loess. In Illinois, four distinct deposits of loess have been verified in field studies. At a few locations there are indications that there might be six separate loess units in succession. In many places silt units are commonly found interbedded with other forms of the drift, but it is not yet possible to correlate among them with any confidence. In other words, we have pieces of the puzzle but do not yet know how they fit together.

The youngest loess is named after Peoria, Illinois, and can be traced across the Midwest from Ohio to Colorado (Follmer 1996). Most of the accumulation of the Peoria Loess [formally called Peoria Silt (Hansel and Johnson 1996)] occurred during the last glaciation from about 10,000 to 25,000 years ago (commonly called the Late Wisconsinan Age in the Midwest). The Peoria Loess formed while a glacier was advancing into northeastern Illinois. Part of this loess was deposited in front of the glacier and was overridden, part was deposited on the glacier and subsequently washed off, and the youngest part was deposited after the glacier melted away. Beyond the limit of the Late Wisconsinan glacier, these intervals of Peoria Loess merge and appear as a single uninterrupted geologic unit.

In many parts of Illinois, the Peoria Loess overlies another loess unit, the Roxana (Silt) Loess. Named after Roxana, Illinois, where it is up to 40 feet thick, this loess is nearly as widespread as the Peoria and has a similar thinning pattern to the east of the major rivers in the Midwest. It formed during the middle portion of the Wisconsinan Age. The reason this loess was deposited is a mystery, because we have not been able to relate it to any glacier to the north. Presumably the evidence of the southern limit of this glacier's advance is buried beneath the Late Wisconsinan deposits on the north side of the Great Lakes. Numerous radiocarbon dates from the Roxana Loess in Illinois show that it formed between 25,000 and 55,000 years ago.

In a few places in southern Illinois, the Peoria and Roxana lie directly over a third loess we correlate with the Loveland Loess of Iowa. The Loveland is well known up the Missouri River valley and can be traced down the Mississippi River Valley to Louisiana. It formed during the next-to-the last glaciation. On the basis of correlations with ice cores from Greenland and ocean sediment records around the world (Imbrie and Imbrie 1979), we believe that the age of the Illinoian can be now constrained to the period between 180,000 to 125,000 years ago. Previous estimates place the older boundary back to 300,000 years ago, but we believe this age should be rejected because no glacial sediments that old have been found in the Midwest (Follmer 1996).

In southern Illinois near the Mississippi River a fourth loess has been found and was correlated with the Crowley's Ridge Loess of Arkansas. Probable correlations are known in Illinois and up the Missouri River Valley to Nebraska. Up to two older loesses have been observed in Illinois and in exposures along the Missouri River near St. Charles, Missouri. All across the glaciated part of Illinois, silt units that may be loess have been observed in the older parts of the glacial

sequence. These silt units have mostly been observed in areas of thick drift where preglacial valleys have been filled with glacial deposits. At this time, we do not know much about these deposits. The oldest glacial deposits in Illinois are approximately 800,000 years old (Follmer 1996).

The loess units are distinguished from each other by physical and mineralogical properties. The most important characteristics are those caused by soil formation. In geologic terms, the soil at the surface that has developed mostly in the Peoria Loess is called the modern soil. Soil profiles, developed in the tops of buried loess units are called paleosols, which means ancient, or fossil soils. The fossil soils' characteristics indicate that the unit was once at the ground surface and exposed to active soil formation. Some other buried glacial deposits also have a paleosol, developed in their upper part; that is, there are soil features in the upper parts of some important geologic units. These buried soils are important markers for mapping the distribution of important units. The classification of the major Quaternary geologic units is based on these key markers (Follmer 1982)

The general surficial (Quaternary) deposits map of Illinois (Figure 3) shows the distribution of the major Quaternary units (Lineback 1981) as they would appear if they were not buried by loess deposits. The deposits of the last glaciation are named after Wisconsin. In Illinois the deposits of this most recent glaciation are mostly limited to the northeast quarter of the state. The next older glaciation is named after Illinois because its deposits cover most of the state. Deposits from some older glaciations have been found in scattered places in Illinois, especially in the lowermost deposits in buried bedrock valleys, but the evidence of their stratigraphic succession and age is limited. Until definitive work is done on these older units, they are meanwhile grouped into an indefinite time period called pre-Illinoian (Lineback et al. 1979).

Where it has not been eroded, the Peoria Loess is the parent material of the modern soil across most of the nearly flat uplands of Illinois. In valleys, the upper part of the alluvium is mostly derived from the loess that has been eroded from the surrounding hills in the watershed. Where the Peoria Loess is missing, particularly on sloping land where it has been stripped off by erosion, the older loesses are almost always missing also. At such sites the soils are formed in other glacial deposits or bedrock.

The important variations in the physical characteristics of modern soil from place to place are given soil names by the USDA–NRCS Soil Survey. We use these names in combination with the geologic names of the parent materials assigned by the ISGS in describing the soil cores we collect. In places where the thickness of the Peoria Loess is less than the solum thickness, i.e., where the soil horizons have developed into the underlying geologic unit, the soil profile is described as having formed in two materials and, in some places, three. In southern Illinois, the Peoria Loess across much of the flat upland on the Illinoian till plain is less than 5 feet thick and underlain by the Roxana Loess. In these situations the modern soil features (roots, and so forth) in the Peoria commonly extend into the Roxana, thus blurring the boundary between them. Where such conditions are found we have grouped the two geologic units together and call it Wisconsinan loess after the time interval during which it was deposited.

The important factors that govern the development of a soil are the nature of the parent (geologic) material (loess, in most of Illinois), the geographic relief (the slope of the land surface), climate (temperature and rainfall patterns), organisms (plants, animals, and microorganisms), and time (Jenny 1941; see also Luxmoore 1994). On a glaciated landscape, soil quality increases with time until the peak or optimum chemical conditions are attained. Eventually, the quality will decline as continued weathering depletes the minerals that supply nutrients. To a certain extent, fertilizers can maintain peak conditions if sufficient inputs are made. The amount and composition of fertilizer to apply is difficult to determine because the balance of minerals that provides the best soil is not well known. Some plants have special requirements, particularly in the balance of trace elements.

The mineralogical and chemical composition of the parent material governs the chemical composition of the soil (see Figure 4). Certain minerals in the parent material, such as mica, feldspar, and hornblende, are weathered over time to form clay minerals, such as kaolinite, illite, and smectite (Jackson and Sherman 1953). Weathering of the primary minerals results from chemical and biochemical reactions of the minerals with water, organic acids, carbon dioxide, and oxygen (Hassett and Banwart 1992). In general, the chemical elements that make up the primary minerals are simply rearranged by weathering processes into new minerals. Most elements that were present in trace concentrations in the parent material remain in the resulting soil (Kabata-Pendias 2001). If the minerals in the soil parent material are easily weathered, then the soil develops rapidly. If, however, the minerals are resistant to weathering, such as quartz sand, then development of the soil will be delayed (Hassett and Banwart 1992).

The term “climate” encompasses both temperature and the long-term availability of water. The higher the temperature of the developing soil, the faster the soil will develop through chemical and biological processes. The availability of water is important in determining the types of plants and organisms that thrive in the developing soil, which chemical and biological reactions will occur, and in the movement of soil particles and dissolved chemical species from one location to another on the landscape or in the soil profile. As water seeps through the soil profile it causes chemical elements to be leached from the surface horizon (eluviation) and to be deposited in lower horizons (illuviation) (Hassett and Banwart 1992).

Topography influences the ability of precipitation to infiltrate the soil profile. In steep terrains water is more likely to run off and to erode the soil surface than to infiltrate. In flatter terrains the opposite is true. In enclosed basins, water may stand on the soil surface for long periods if the soil pores have very small diameters or have been plugged by fine particles, as in a pond (Hassett and Banwart 1992).

Organisms in the soil present major factors in soil formation. Microscopic organisms are the major promoters of the degradation of organic matter. When plants die, it is the soil microorganisms that cause the rapid breakdown of plant tissues. The degraded organic matter is very important, along with minerals, in maintaining soil fertility (Hassett and Banwart 1992). Microorganisms also catalyze chemical reactions. That is, a particular chemical reaction would occur even if microorganisms were absent, but because certain types of microorganisms depend

on particular chemical elements in the soil, they act as promoters of the reaction—the microorganisms cause the reaction to proceed at a faster rate than it would in the absence of the microorganisms (Hassett and Banwart 1992).

Exposure of the soil profile to weathering factors for long time periods allows the other soil formation factors to more fully act on the soil (Hassett and Banwart 1992).

Since the Industrial Revolution, metals have been introduced to the soil surface through atmospheric outfall of particulate matter generated by various industries, by capture of small particles (aerosols) in raindrops, and by the application of fertilizers and other agricultural chemicals to the soils (Kabata-Pendias 2001). For example, the concentration of molybdenum in soils near a molybdenum processing plant in western Pennsylvania was found to form a plume of contamination in the surface soil in the direction of the prevailing winds. The molybdenum concentration decreased with downwind distance from the plant (Hornick et al. 1976). At about 1 mile from the processing plant the molybdenum concentration was about 30 mg/kg, but at 5 miles it had decreased to about 6 mg/kg. Prior to “Clean Air” regulations emissions from coal burning plants could cause widespread dispersion of metals at large distances from their source (Mattigod and Page 1983).

Other factors that affect the distribution of metals in soil are churning, or bioturbation, of the soil by earthworms, ants, termites, other invertebrates, and burrowing mammals, such as moles, chipmunks, and gophers (Paton et al. 1995). Plants also accumulate metals during their growth cycles. When the plants die, they are decomposed by microorganisms, which releases the metals back into the soils (Kabata-Pendias 2001). If the plants are not recycled to the soil, as in many farming operations, then periodic fertilization in greater amounts is required. The leaching of metals and transport of colloidal-sized particles (0.001 to 1  $\mu\text{m}$  diameter) causes metals generally to move downward through the soil column, but capillary action can cause metals dissolved in the soil water to move upward (Simonson 1978).

Various chemical reactions also operate on metals and the soil parent materials. These include sorption and desorption, dissolution and precipitation, occlusion and coprecipitation, oxidation and reduction, chelation and release by organic chemicals, and fixation and release by biological organisms. The reactions are affected by the pH of the soil, the availability of oxygen, the presence of various types of clay minerals, the rates of various chemical reactions, the presence of and nature of various kinds of animals and microorganisms, and the reaction of organic chemicals with metals and clay minerals. All these variables affect how the metals are held in the soil. For example, these factors determine whether a particular metal is bound to the surface of a clay mineral or on an oxide/hydroxide, or whether it is present as a discrete oxide, hydroxide, or other compound (Hassett and Banwart 1992).

## **DEFINITION OF TERMS**

Sorption and desorption in a soil refer to an interaction between small particles, such as colloidal-sized clay minerals or organic materials, and solutes dissolved in the soil water. The



dissolved solutes are attracted to the surfaces of the particles because of differences in electrical charge, and tend to become sorbed on the particles to the point of equilibrium. Equilibrium refers to the condition in which the concentration of an adsorbed species and the concentration of that species in the soil solution have reached a balance; that is, the concentrations remain unchanged. If the concentration of the solute in the soil solution is greater than is necessary to achieve equilibrium, then the solute will be adsorbed on soil particles until equilibrium is attained. If the concentration of the solute in the soil solution is less than required to achieve equilibrium, then the solute will be desorbed, that is, it will be released from the solid particles into the solution until equilibrium is again attained (Hassett and Banwart 1992).

Dissolution and precipitation refer to the process of solid materials entering into or separating from a solvent. For example, when a small amount of sugar is stirred into water, the sugar enters the solution; that is, it dissolves. If the water is then allowed to evaporate, the sugar eventually separates from the solvent (water) as crystals; that is, it precipitates.

Occlusion refers to the physical enclosure or capture of small amounts of fluid, such as soil solution, in a rapidly growing crystal. Coprecipitation refers to the precipitation of a substance that would normally be soluble with another substance that would normally precipitate (Fisher 1961).

Oxidation and reduction refer to the transfer of one or more electrons between solute ions. The ion, or electrically charged atom, from which the electron is transferred is oxidized, and the ion which receives the electron is reduced. In soils, iron and manganese commonly undergo oxidation-reduction reactions. Under oxidizing conditions, the soil is commonly well-drained and aerated, and both iron and manganese precipitate, usually as oxides and/or hydroxides (called oxyhydroxides). If the soil becomes saturated with water and oxygen is excluded (producing reducing conditions), iron and manganese oxides and hydroxides dissolve. In the overall process the iron and manganese gain electrons during reduction and lose electrons during oxidation (Birkeland 1999).

As plant residues decompose, many organic compounds are formed, some of which will dissolve in the soil solution. Metals, such as cobalt, nickel, iron, and manganese, are readily captured by the organic molecules to form what are known as chelate compounds (Fritz and Schenk 1966). Microorganisms sometimes accumulate metal ions, effectively removing the ions from the soil solution. However, when the microorganism dies and decomposes, the accumulated ions will be released again (Weatherley et al. 1980).

The reactions described above can all occur at the same time in a soil. The factors that govern the reactions are temperature, pH, the depth to which oxygen can penetrate the soil column and the rate at which it can be replaced as it is used in chemical and biological reactions, the degree of saturation of the soil by water, and the number and types of animals and microorganisms in the soil. The various chemical and biological reactions determine how metals are held in the soil.

## **FIELD METHODS**

The sampling plan adopted was to collect cores of soils at the points of intersection, or nodes, of a rectangular grid, 20 miles on a side, laid on top of a state map. The grid was established in the Universal Transverse Mercator projection by drawing the first node 10 miles west of Lake Michigan and 10 miles south of the Illinois–Wisconsin border, in Lake County, northeastern Illinois. The remainder of the grid was drawn from this starting point using Geographic Information System (GIS) computer software. The grid comprises 137 nodes, or sampling points (see Figure 5). At a rate of approximately 25 sampling sites per year, six years will be required to complete the sample collection.

The geographic locations of the grid nodes calculated by the GIS were used to locate the target sampling points on appropriate topographic maps, plat maps, soil maps, and highway maps. County assessment supervisors or county clerks were contacted to confirm the current ownership of the various properties upon which the grid nodes were located. Landowners were then contacted by letter to inform them of the research project and to inform them that ISGS personnel would like to visit with them to discuss the project and their willingness to participate by granting permission for a core to be taken from their property. Most landowners that we contacted in 1998 were willing to participate. For the few who did not wish to participate, an alternate site was selected from a nearby location which had the same soil type as that at the original target sampling location.

The initial visits with landowners were made during September 1998. Cores were collected at 26 locations during November and December 1998, after harvest was completed. The gray circular symbols on the map in Figure 5 indicate the approximate locations of the sampling sites. The core number, the county in which the core was collected, and the length of the core are recorded in Table 1.

A Giddings® hydraulically operated coring device mounted on a two-ton pickup truck was used to collect all cores. A combination of unsplit and split core barrels was used, depending on conditions encountered in the soil. The core barrel was pushed into the soil/sediment with no applied rotation. The cores were briefly described in the field as they were collected. Each core segment, approximately 2 feet long, was wrapped in plastic food wrap, then overwrapped with heavy-duty aluminum foil. The cores were labeled and placed in core boxes for transport and storage.

## **LABORATORY METHODS**

At the ISGS, the cores were unwrapped, trimmed to remove any smeared and/or oxidized material from the outer surface, and described in more detail than done at the time of collection. Samples were selected from the cores on the basis of lithologic changes for chemical analysis. Samples from cores 1 through 8 were dried at 60°C for 24 hours. Samples from cores 9 and 10 were dried at 50°C. It was decided that the lower temperature would be more likely to prevent loss of mercury, if present, from the samples. The dried samples were then disaggregated to approximately <4 mm size by passing them through a miniature jaw crusher. The samples were further disaggregated to pass a sieve with 2-mm openings by placing the material between two

sheets of clean white paper and crushing the particles with a wooden rolling pin. The samples were then split by the “cone and quarter” method to a reduced sample mass of about 30 grams. This subsample was then ground in a Spex® 8500 Shatterbox® and passed through a 208-µm (no. 65) sieve in preparation for chemical analyses. All analyses were conducted in ISGS laboratories.

### **Wavelength-Dispersive X-ray Fluorescence Spectrometry**

Wavelength-dispersive x-ray fluorescence spectrometry (XRF) was used to determine the concentrations of the following major and minor elements [silicon (Si), aluminum (Al), iron (Fe), potassium (K), calcium (Ca), magnesium (Mg), sodium (Na), titanium (Ti), phosphorus (P), manganese (Mn), and sulfur (S)] and trace elements [barium (Ba), cadmium (Cd), chromium (Cr), copper (Cu), molybdenum (Mo), nickel (Ni), lead (Pb), rubidium (Rb), tin (Sn), strontium (Sr), vanadium (V), zinc (Zn), and zirconium (Zr)] in the samples.

To prepare the samples for major and minor element determinations, the samples were ground in a Spex® 8500 Shatterbox® with ceramic grinding surfaces to pass a 208-µm sieve. The ground samples were dried at 110°C overnight. Then the samples were ignited in platinum-rhodium crucibles at 1000°C for one hour to determine the loss on ignition and to ensure complete oxidation of the samples. Loss-on-ignition data provide a measure of the amounts of structural water and organic- and carbonate-carbon in the samples.

Six-tenths of a gram of the ignited sample was mixed with 5.4 g of flux mixture (50% lithium tetraborate plus 50% lithium metaborate) and fused in a 95% platinum-5% gold crucible in the propane flame of a Claisse-bis® Fluxer. The melt is automatically poured into the crucible cover, which also serves as a flat circular mold, for cooling. The resulting glass disk was stored in a desiccator prior to analysis.

Pressed pellets of samples were prepared for the determination of volatile elements, such as Cl and As, which would be partially or completely lost during fusion, and trace elements, which generally become too dilute in the fused disk samples.

In this preparation, 10 g of the as-received sample was ground in a tungsten carbide grinding container and then ground in a Spex® 8500 Shatterbox® for three minutes. A portion of the ground sample (6.3 g) was mixed with 0.7 g of Chemplex® X-Ray Mix Powder as a binder. This mixture was placed in a tungsten carbide grinding capsule that contained a tungsten carbide puck. The capsule was then placed in a Spex Shatterbox® pulverizer to ensure thorough mixing of the sample and binder. The sample mixture was placed in a steel die, a layer of 2 g of cellulose powder was added on top of the sample, and the mass was pressed into a 35-mm diameter disk under a pressure of 20 tons/in<sup>2</sup> for two minutes. The cellulose provides a reinforcing backing for the disk. The prepared pellets were stored in a desiccator prior to analysis.

### **Mercury**

Total mercury in the soil sample was determined by cold vapor atomic absorption spectrometry (CVAAS). In CVAAS mercury is reduced to elemental mercury in the vapor state immediately prior to passing the vapor through a detection cell. Metal ions, including mercury, are dissolved from the soil sample by mixing the sample with aqua regia, a mixture of concentrated hydrochloric and nitric acids. Mercury combined in both inorganic and organic compounds are oxidized in the aqua regia to the mercurous ion by potassium permanganate. Excess potassium permanganate is reduced by addition of a solution of hydroxylamine hydrochloride prior to analysis. The solution in the sample tube is drawn off by use of a peristaltic pump and combined with an acidic carrier solution. The sample is directed to a reaction chamber where it is combined with a solution of stannous chloride, whereby mercury ions are reduced to elemental mercury. A stream of argon is allowed to pass through the reaction solution. The elemental mercury is carried by the argon stream to the detection cell. The amount of mercury in the argon stream is converted to concentration in the original soil sample. The method detection limit is approximately 3 µg of mercury per kilogram of soil.

### **Total Carbon and Inorganic Carbon**

Total and inorganic carbon were determined in the less-than 208-µm samples using a Coulometrics Inc.<sup>®</sup> carbon analyzer. For determinations of total carbon a weighed amount, 10 to 30 mg, of the sample was heated for 10 minutes in a tube furnace at 950°C through which a stream of oxygen was allowed to flow. Carbon in the samples reacted with the oxygen to form carbon dioxide (CO<sub>2</sub>) gas. The generated CO<sub>2</sub> was absorbed in a solution of ethanoldiamine, with which it reacts to form acid. The acid thus released was titrated by an electrical current until a neutral pH was attained. The amount of current required to reach the end-point of the titration is an indirect measure of the amount of carbon in the original sample.

The method for the determination of inorganic carbon was similar, except that rather than the sample being heated in a tube furnace as in the determination of total carbon, the sample was submerged in a dilute solution of hydrochloric acid which reacted with the carbonate to generate CO<sub>2</sub>.

### **Soil pH**

Five grams of the oven-dried (50°C) <2 mm sample was weighed into a 50-mL disposable plastic beaker. Five mL of deionized water was added to the beaker, which was swirled and allowed to stand for 5 to 10 minutes. The pH value of the slurry was determined by a solid-state pH electrode immersed in the slurry. The pH was determined with a Corning<sup>®</sup> Model 314i ISFET pH meter.

### **Soil Texture**

Soil texture was determined by the method of Indorante et al. (1990). Up to nine samples and a blank were processed at a time. In this method, 10 g of oven-dried (50°C) <2 mm sample was placed in a 500 mL plastic, wide-mouth, screw-cap bottle. Ten mL of a 10% solution of sodium

hexametaphosphate ( $\text{NaPO}_3$ )<sub>6</sub> was added to the bottle followed by 140 mL of deionized water. The blank contained ( $\text{NaPO}_3$ )<sub>6</sub> and deionized water only. All bottles were sealed and placed on an oscillating shaker and shaken at 120 strokes per minute overnight. After shaking, 250 mL of deionized water was added to each bottle. The bottles were then shaken end-to-end by hand for 25 seconds to dislodge settled sand and silt, then end-over-end for 15 seconds. At the end of the 15-second period, the bottles were immediately placed in a covered, static water bath at 25°C for 3 hours and 36 minutes to allow particles larger than 2  $\mu\text{m}$  to settle from the top 5 cm of the suspension. The samples were placed in the water bath at two-minute intervals to provide ample time between samples to withdraw aliquots for further processing. At the end of the settling period, the bottles were removed from the water bath at two-minute intervals in the same sequence in which they were placed into it. The tip of an Oxford® Macro-Set adjustable pipet was carefully inserted into the suspension to a depth of 5 cm, and exactly 10 mL of the suspension was withdrawn from the suspension during a 15- to 20-second period. The pipet had been previously calibrated by weighing and averaging 10 replicate volumes of deionized water at the desired volume. The suspension was delivered to a numbered, weighed aluminum weighing pan. When aliquots of all the samples and the blank had been delivered to their respective weighing pans, the pans were placed in an oven to dry overnight at 110°C.

After the aliquots of suspended clay fraction had been withdrawn, the contents of each bottle were poured through a 3.5-inch diameter, 62  $\mu\text{m}$  (No. 230) stainless steel sieve to separate the sand-size particles from the silt- and clay-size particles. Each bottle was thoroughly rinsed with deionized water. Successive rinsates were poured through the sieve until no particulate matter could be observed in the bottle. The sand was rinsed several times to remove all silt-size and smaller particles.

The sand was then backwashed from the sieve with deionized water into a beaker, and the sand was quantitatively filtered through a numbered and weighed circle of Whatman No. 41 filter paper. The filter paper was rinsed three times with deionized water, folded, and dried overnight in an oven at 110°C.

After drying, both the aluminum weighing pans and the filter papers were weighed. The weight of clay in the weighing pans was corrected for the weight of ( $\text{NaPO}_3$ )<sub>6</sub> in the blank. The clay and sand contents were calculated for each sample. Silt content was calculated by subtracting the percentages of clay and sand from 100%.

Organic matter in the sample has been found to distort the determination of clay-size content, therefore, the samples that contained more than 1% organic carbon were treated with 30% hydrogen peroxide ( $\text{H}_2\text{O}_2$ ). We also determined it necessary to treat the upper two samples from each core with  $\text{H}_2\text{O}_2$ . Approximately 12 grams of <2 mm sample was weighed into a 500-mL beaker. Five mL of deionized water, 5 mL of 30%  $\text{H}_2\text{O}_2$  and one drop of glacial acetic acid were added to the beaker in a fume hood and the beakers were covered with watch glasses. When the reaction subsided, the beakers were transferred to a covered water bath at 60°C. Additional 5-mL aliquots of 30%  $\text{H}_2\text{O}_2$  were added to the beakers at 15- to 20-minute intervals until a total of 30 mL of 30%  $\text{H}_2\text{O}_2$  had been added to each beaker. The covered beakers were allowed to stand in

the covered water bath overnight. The beakers were then dried in an oven at 50°C overnight. The treated samples were removed from the beakers with a plastic spatula. A 10 g portion of each H<sub>2</sub>O<sub>2</sub>-treated sample was weighed into 500-mL plastic bottles for texture determinations as described above.

## **RESULTS AND DISCUSSION**

### **Sampling Plan**

The square-grid sampling plan (Fig. 5) was chosen as the pattern for coring locations because we wanted to collect a set of cores that was representative of the distribution of soil associations throughout the state. If a complex distribution of soil associations is sampled according to a regular grid sampling plan, given a sufficient number of samples, the statistical distribution of the soils that are cored will be representative of the distribution of soil associations in the state.

Table 2 provides the percentage of the state covered by each soil association according to Fehrenbacher et al. (1984) and the number of cores to be collected through this project from each soil association. A comparison of the two distributions (percent coverage of each soil association and the percentage of cores to be collected in each soil association) is shown in Figure 6. As shown in the figure, the two distributions are reasonably similar. The correlation coefficient of the two distributions is 0.82 at the 95% confidence level.

Table 2 also shows the difference between the percentage of the state's area covered by each soil association and the percentage of all the cores to be collected from each soil association. The distribution of differences is nearly normal with a mean of 0.01 and a standard deviation of  $\pm 1.04$  (Figure 7). Therefore, the sampling procedure under-represents about as many soil associations as it over-represents, and the sampling procedure is shown to be unbiased.

### **Soil Texture**

The sand, silt, and clay contents of each sample from each core are listed in Tables 3 through 12, and depicted graphically in Figure 8. The proportion of sand, silt, and clay in a sample determines the textural class of the sample. We use the textural classes defined by the USDA Soil Survey Division Staff (1993). The USDA terminology is useful for several reasons: (1) it is popular and used in all U.S. soil survey reports, (2) it is quantitative and easy to determine, and (3) it is based on empirical factors—there is a natural tendency for sediments to occur in these classes and the textural classification has about the maximum practical number of classes to use. Of the ten cores discussed in this report, the texture of the uppermost or surficial horizon (the plow layer in most cases) of eight of the cores was silt loam and of the other two (cores 1 and 4) silty clay loam. The parent materials of the soils at sites 1 and 4 are alluvium, and were deposited as floodplain sediments. The soil at site 6 was collected from the bed of ancient Lake Saline, and is an example of a soil developed in the deposits left by a lake caused by glacial flooding. Although the uppermost sample of core 6 was silt loam, samples from below the uppermost sample were silty clay in texture.

The clay content of all the subsamples of all the cores ranged from about 15% to 75%. The silt content ranged from about 24% to 86%. The soil texture classification, soil type, and developmental environment of the uppermost soil samples are listed in Table 13. In general, the clay content of the samples increased with depth, as shown in Figure 8.

In general, the silt size fraction of a soil comprises principally quartz ( $\text{SiO}_2$ ), with feldspar and carbonate minerals (where present), plus small amounts of heavy minerals, such as zircon ( $\text{ZrSiO}_4$ ). Quartz and zircon are two of the most resistant minerals in soils along with rutile ( $\text{TiO}_2$ ) and ilmenite ( $\text{FeTiO}_3$ ) and other iron-bearing minerals. In addition, quartz and other siliceous minerals are resistant to physical abrasion, which means that much energy is required to grind these minerals to the silt-size range. The results of our analyses showed a moderately strong correlation (Table 14) of the silt-size fraction with silicon. Quartz grains probably reached a size limit, also known as “terminal grade,” below which they could not be ground by glacial action. For quartz that is ground by glacial action, terminal grade is between about 31 and 62  $\mu\text{m}$  (Dreimanis and Vagners 1969 1971). With artificial grinding, the terminal grade for quartz is about 16 to 32  $\mu\text{m}$  (Gaudin 1926), slightly finer than that observed by Dreimanis and Vagners, but still in the range of 16 to 62  $\mu\text{m}$ .

## Results of Chemical Analyses

A list of the correlation coefficients of various chemical constituents in the soil samples is given in Table 14. The results of the chemical analyses of the subsamples from each of the 10 cores are presented in Tables 15 through 24 and Figures 9 through 29. Major and minor element concentrations as oxides (silicon through sulfur in the tables) are listed first, followed by trace element concentrations (barium through zirconium). Major elements are those whose concentrations are greater than 1%, minor elements are those whose concentrations are between 0.1 and 1%, and trace elements are present at less than 0.1%, or 1000 mg/kg. The concentrations of all major and minor elements, as oxides, are listed as weight percent; trace element concentrations are listed as mg/kg.

**Correlation Coefficients** A correlation coefficient is a numerical description of the relationship of one constituent with another. If two constituents possess a positive correlation coefficient it means that as the concentration of one constituent increases from one sample to another, the concentration of the second constituent increases also. If the correlation coefficient is negative, then as the concentration of the one constituent increases, the concentration of the other decreases.

For example, the correlation coefficient between alumina ( $\text{Al}_2\text{O}_3$ ) and the clay-size fraction is 0.91 (Table 14). That is, in about 91% of the samples, as the content of the clay-size fraction increased the alumina content also increased. Because two constituents are positively correlated does not necessarily mean that they are always present in the same ratio. Therefore, correlation coefficients do not prove conclusively that any certain mineral is present in a sample; they are merely suggestive.

We do not have sufficient data with only ten cores to draw conclusions about similarities or differences between cores, and certainly not to make any suggestions about regional trends. It is, however, our goal, when we do have sufficient data, to study similarities and differences in constituent trends relative to soil horizons and to study regional or state-wide patterns. We did calculate correlation coefficients to highlight the relationships between constituents based on all samples analyzed from all ten cores. From this statistical analysis we are able to observe general relationships between constituents.

The calculation of correlation coefficients helped confirm soil chemical properties and relationships that are known from previous studies. For example, it is known that silt-sized particles are composed predominantly of silica ( $\text{SiO}_2$ , Brady and Weil 1999) and that zirconium (Zr), which is found principally as zircon ( $\text{ZrSiO}_4$ ) in soils, also occurs predominantly in the silt-size particles. The correlations between silica, zirconium, and the silt-size and sand+silt-size fractions are strong and confirm these previously known observations.

Clay minerals are important components of all soils. They occur principally in the clay-size fraction and contain Si as one of the major constituents. Because  $\text{SiO}_2$  is the major constituent of the silt- and sand-size fractions, we observed that as the amounts of these two fractions increased, the amount of  $\text{SiO}_2$  in the samples also increased. That is, the correlation between Si and the two fractions was strong. As the silt and sand fractions increased, the clay-size fraction decreased and there was a negative correlation between Si and the clay-size fraction. Aluminum, another major constituent of clay minerals and other minerals in the clay-size fraction, demonstrated a strong positive correlation with the clay-size fraction.

The clay-size fraction was also correlated with iron (0.74), potassium (0.58), magnesium (0.70), chromium (0.57), nickel (0.60), rubidium (0.82), vanadium (0.87), and zinc (0.58). Therefore, these elements correlated with each other. Iron, potassium, and magnesium are commonly structural members, that is, they are necessary parts, of certain clay minerals, although these elements are also found as exchangeable ions on clay minerals. Potassium is found in illite, as are iron and magnesium. In addition, iron commonly is found in coatings of iron oxyhydroxides on other minerals (Wilding et al. 1977) and as concretions in many soils.

The other elements noted above, chromium, copper, nickel, rubidium, vanadium, and zinc, are known as soil trace elements because they occur at trace concentrations, or less than 1000 mg/kg. These metals are readily adsorbed by, or attached to, clay minerals, or in some cases, trapped (occluded) within the clay mineral structure. The iron oxyhydroxide minerals are strong adsorbers of many trace elements, as well.

The correlation of 0.78 between  $\text{CaO}$  and  $\text{SO}_3$  indicates an association between these two elements. Whether gypsum ( $\text{CaSO}_4 \cdot 2\text{H}_2\text{O}$ ) or anhydrite ( $\text{CaSO}_4$ ) or some other chemical combination involving calcium and sulfur was present in the samples is difficult to say. During preparation of the samples for chemical analyses, they were dried in an oven at  $50^\circ\text{C}$ . As the water evaporated from a sample, any dissolved solids would have precipitated. The observed concentration of sulfur as  $\text{SO}_3$ , was only about 0.1 to 0.15% in most samples, so the amount of

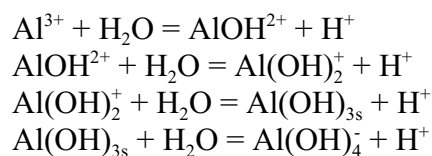


any Ca associated with the SO<sub>3</sub> would be small.

The strong correlation of 0.98 between CaO and inorganic (or carbonate) carbon suggests the presence of calcite (CaCO<sub>3</sub>) in the samples as analyzed. There was also a correlation of 0.60 between Ca and organic carbon. Plants require Ca as part of their nutrition. When plants die and the organic matter decays, the Ca is released to the soil (Severson and Shacklette 1988). In fact, Ca, among other elements is found in all living organisms including plants, animals, and microbes (Nebel 1987). When these organisms die in contact with the soil the Ca is released to the soil. Sulfur trioxide (SO<sub>3</sub>) was correlated with both inorganic carbon (0.69) and organic carbon (0.81). The correlation between SO<sub>3</sub> and inorganic carbon is probably a result of their mutual correlations with CaO. Sulfur is also found in all living organisms (Nebel 1987), thus the correlation between SO<sub>3</sub> and organic carbon. The fact that two parameters show large positive or negative correlation coefficients does not necessarily demonstrate a direct cause-and-effect relationship between them.

The correlation between lead (Pb) and manganese (Mn) was 0.62, and that between Pb and organic carbon was also 0.62. Lead may be incorporated into both Mn oxides and organic matter in soils (Kabata-Pendias 2001).

**Soil pH** Soil pH is a measure of the acidity or hydrogen ion (H<sup>+</sup>) content of the soil solution. Various substances are important in maintaining soil pH, depending on the pH range. At pH values between 4.5 and 5.5, aluminum in the soil solution, which may originate from clay minerals, buffers the pH of the soil solution according to the following chemical reactions (Hassett 1989):

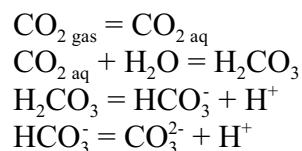


If small amounts of gibbsite, Al(OH)<sub>3</sub>, or clay minerals in the soil dissolve to contribute Al<sup>3+</sup> to the soil solution, the Al<sup>3+</sup> becomes hydrolyzed to AlOH<sup>2+</sup>, which results in the addition of H<sup>+</sup> to the solution, which results in greater acidity. If additional H<sup>+</sup> enters the soil solution, the reaction between AlOH<sup>2+</sup> and Al<sup>3+</sup> is driven to the left and H<sup>+</sup> is consumed in the formation of H<sub>2</sub>O and Al<sup>3+</sup>. Aluminum, therefore, serves as an acidity buffer for pH between pH values of about 4.5 and 5.5 (Sparks 1995). Above pH 5.5, the solubility of Al<sup>3+</sup> is low enough that it is not effective in buffering soil pH.

In the pH range of 5.5 to 6.8, there are three mechanisms that act to maintain soil solution pH: (1) H<sup>+</sup> and basic cations of the soil exchange-complex buffer the pH, (2) atmospheric carbon dioxide (CO<sub>2</sub>) dissolves in and reacts with water, and (3) weak acidic groups of soil mineral matter such as silicate groups of clay minerals (for example, smectite) and organic matter exchange H<sup>+</sup> with the soil solution. Soil organic matter is more important than clay minerals in controlling pH and adsorption of various cations in soils (Helling et al. 1964; Yuan et al. 1967)

If some external source of acid makes the soil solution more acidic, some of the  $H^+$  ions in the soil solution are adsorbed by the soil in exchange for basic cations, which go into solution. Conversely, if the soil solution becomes more alkaline, some of the basic cations become attached to the soil solids in exchange for  $H^+$  ions, which enter the soil solution and make the solution more acidic. In this way, the soil solids act to resist change in the pH of the soil solution.

Atmospheric carbon dioxide ( $CO_2$ ) contributes to the control of soil solution pH by reactions with water to produce carbonic acid and intermediate carbonate and bicarbonate anions, as in the following reactions (Hassett 1989):



In the pH range 5.5 to 7.2, weak acidic groups such as  $-AlOH$  and  $-SiOH$  on the edges of clay minerals and  $-COOH$  groups on carboxylic acids in organic matter serve to buffer, or protect, the soil solution against changes in pH by consuming or releasing hydrogen ion ( $H^+$ ).

Above pH 7.2 the pH is controlled by the precipitation or dissolution of carbonate minerals such as calcite and dolomite. As calcite dissolves in the soil solution it releases carbonate ion ( $CO_3^{2-}$ ), which reacts with water to produce bicarbonate ions, which, as illustrated in the equations given above, react to produce or consume  $H^+$  ions. The pH value at which calcite or dolomite precipitate depends on the concentration of  $CO_2$  in the gas phase. The greater the amount of carbon dioxide in the gas phase, the lower the solubility of calcite.

The pH values of the samples from the 10 cores ranged from 3.74 (very acidic) to 8.06 (moderately alkaline), with a median value of 5.9. Of the 65 samples, 51 had a pH between 4 and 7. The approximate pH range of most soils found globally is 3.0 to 8.5 (Baas Beeking et al. 1960). As shown in Figure 9 and Tables 15 through 24, the pH value in seven of the 10 cores became more acidic with increasing depth to about 4.0 feet, then became more alkaline with increasing depth.

**Means and Ranges of Element Concentrations** The mean and range of the concentrations of all elements determined in the uppermost samples from the 10 cores in this project are compared in Table 25 with the results obtained by Shacklette and Boerngen (1984) for loess and silty soils or loamy and clay soils, and Severson and Shacklette (1988), according to the availability of their data. The data from Shacklette and Boerngen for trace elements were for surface horizons from throughout the U.S., whereas the data for most major and minor elements were for Illinois soils without regard to soil texture. Our ranges of values were within the ranges observed by Shacklette and Boerngen for similar soil textures for about 50% of the elements determined.

The excursions of concentrations outside the ranges observed by Shacklette and Boerngen did

not necessarily mean that the soil sample was contaminated, nor that plants grown in that soil will absorb any of those elements in toxic amounts. Much depends on how tightly bound the elements are by the clay minerals, iron and manganese oxides and hydroxides, and soil organic matter. The pH of the soil solution is also very important in determining the solubility of various elements.

**Silicon** According to Kabata-Pendias (2001), quartz, or  $\text{SiO}_2$ , is the most resistant common mineral in soils. It seems likely to us that a large portion of the quartz in the source rocks for the glacial deposits in Illinois was originally sand-sized material. Glacial transport reduced the size of the quartz grains by grinding them to a size that approached the terminal grade in the silt range. The grinding does not appear to have reached completion in the natural setting, but had there been increased transport distance an increase in the amount of coarse silt (31 to 62  $\mu\text{m}$ ) would have been favored (Dreimanis and Vagners 1971). Experimental tests have indicated that the terminal grade for quartz is about 25  $\mu\text{m}$  (Dreimanis and Vagners 1971). In the glacial deposits of Illinois, the fraction with the maximum amount of quartz ranges in size from medium silt (0.006 to 0.02 mm) to fine sand (0.125 to 0.250 mm).

Cores 1, 4, and 6 contained less Si and more clay-size particles than the other seven cores. Cores 1 and 4 were collected from floodplain (alluvial) deposits and core 6 from lacustrine deposits. Both floodplain and lacustrine sediments generally contain more clay-sized particles than loessial deposits because of sorting during transport in flowing water. When water flow is fast, even sand-size and larger particles are suspended in the water. As the flow rate slows the larger particles settle out of suspension until only the smallest particles are left in suspension. Farther from the channel during flood stage the slower the rate of flow of the water. Although core 8 was taken from alluvial deposits, it was located very close to the source stream. Core 8, therefore, contained more sand than cores 1 and 4, which were collected 2 to 3 miles from the source rivers. The Si contents of cores 1, 4, and 6 ranged from about 63 to 69% (except the uppermost and lowermost samples from core 6), whereas the range of Si contents of the other seven cores was about 71 to 81 %. There were only a few Si values from the seven cores that were outside this range. In general, the Si content decreased from the surface horizon into the B horizon.

**Aluminum** The aluminum content of soils was inherited from the parent materials (Kabata-Pendias 2001). The concentration of Al in the cores generally increased with depth, although in some cores the concentration passed through a maximum at some depth. These trends reflected the trends in clay content of the subsamples. The Al contents of the cores collected in alluvial and lacustrine soils, cores 1, 4, and 6, were greater than in the other cores, but the trends with depth in these three cores were similar to the other cores.

**Iron** Iron in soils occurs principally as oxyhydroxides, most commonly as goethite ( $\text{FeOOH}$ ) in soils of temperate, humid regions. Iron minerals commonly occur as coatings on clay minerals, silt, and sand particles, and cements in concretions. The iron oxides and hydroxides typically are very fine-grained, possess large surface areas, and are active adsorbers of other cations, particularly metals such as copper, nickel, vanadium, and zinc (Kabata-Pendias 2001).

Iron has an affinity for organic molecules and forms complex compounds with them. These organo-iron complexes may be largely responsible for the migration of iron through the soil profile (Kabata-Pendias 2001). The Fe concentrations in these cores tended to pass through a maximum with depth, an indication of downward migration.

**Potassium** Typically, potassium occurs in Illinois soils in the primary minerals, particularly feldspars and micas (Sparks 1995), and the clay mineral illite, but it can also be held as an exchangeable ion on the soil exchange complex. Because feldspars and micas are quite resistant to weathering, K is not commonly found at high concentrations in the soil solution. In fact, Severson and Shacklette (1988) estimated that 90 to 98 percent of the K in soils is unavailable to plants.

In general, the concentration of K in the ten cores increased with depth, suggesting downward migration. Cores 1, 4, and 6 have slightly higher concentrations of K than the other seven cores.

**Calcium** Calcite ( $\text{CaCO}_3$ ) and dolomite [ $\text{CaMg}(\text{CO}_3)_2$ ] are common sources of calcium in soils, however, not all soils contain calcite or dolomite. Calcium can also be held as an exchangeable ion on the soil exchange complex. The concentration of Ca in most of the cores was highest in the surface sample and quickly decreased to nearly constant values in samples from greater depths. In some cores the Ca concentration increased in deeper samples because the calcareous till underlying the soil was penetrated. Core 8 was anomalous in its Ca concentration. This core was collected in an abandoned road bed near a creek. It is likely that crushed limestone was used in building the former road, residual portions of which were included in the cored sample. We also observed particles of coal and cinders in the uppermost sample of this core.

**Magnesium** Sedimentary minerals such as dolomite are probably the principal sources of magnesium in Illinois soils. Magnesium has leached downward in the soil profile, as the data for the ten cores generally showed increasing concentrations of Mg with depth in the cores, and generally a decrease in Mg in the deepest samples analyzed. Because Mg compounds are easily soluble in water, Mg very likely could have leached from the soil profile into underlying groundwater. In cores 8 and 10 the concentration of Mg decreased with depth in the profile.

**Sodium** Sodium-rich feldspar is probably the principal source of sodium in soils (Sparks 1995). Sodium is quite mobile and forms many water-soluble compounds in soils. The trends for the concentration of Na in the 10 cores were not consistent between cores. In some cores the concentration of Na decreased with depth, in others it increased, and in others it passed through a maximum. A small portion of the Na present in the parent materials likely has been incorporated onto exchangeable positions on clay minerals, such as smectites, but most Na probably leached out of the soil.

**Titanium** The sources of titanium in soils are oxides, such as rutile ( $\text{TiO}_2$ ), and ilmenite ( $\text{FeTiO}_3$ ) (Kabata-Pendias 2001), neither of which is easily weathered. These minerals occur nearly undecomposed in soils. Titanium presents no environmental concerns in soils (Kabata-Pendias 2001). The concentrations of Ti in the cores were relatively uniform with depth in the

soil profile.

**Phosphorus** The content of phosphorus in soils is low, which makes the identification of phosphorus-bearing minerals difficult. Apatite [ $\text{Ca}_5(\text{F,Cl,OH})(\text{PO}_4)_3$ ] has been identified in the silt-size fraction of some soils and might be the principal source of P, but most inorganic phosphate in soils occurs in the clay-size fraction (Lindsay et al. 1989). Phosphorus-containing fertilizers are the most common source of phosphorus in agricultural and residential soils.

There were no clear trends of P concentration with depth in the 10 cores.

**Manganese** The principal source minerals for manganese in soils are amphiboles, pyroxene, biotite, and rhodonite ( $\text{MnSiO}_3$ ) (Sparks 1995). Although Mn occurs in the bulk of the soil as coatings on other minerals (Kabata-Pendias 2001), it commonly is concentrated in nodules accompanied by iron. These nodules seem to form in soil horizons that periodically become waterlogged, that is, reducing conditions prevail, then the soil dries, providing oxidizing conditions (McKenzie 1989). In some soils a microscopic layered structure of alternating bands of iron-rich and manganese-rich material have been observed (McKenzie 1989).

In seven of the 10 cores there was a general trend of decreasing Mn concentration. One possible reason for this is that Mn is sensitive to oxidation and reduction. Oxidized species, such as  $\text{MnO}_2$ , precipitate when oxygen is readily available, as would normally be the case near the surface of the soil, and decrease in content with depth. The solubility of Mn increases as the pH and Eh (oxidation-reduction potential) decrease (the soil becomes more acidic and less oxidizing) (Lindsay 1979). Under conditions of increasing acidity and decreasing oxidation potential, Mn would be somewhat easily leached.

**Sulfur** The probable source of sulfur in soil was the oxidation of pyrite and marcasite (both minerals have the same chemical composition,  $\text{FeS}_2$ ). Pyrite in Illinois soils was oxidized long ago, and the S now is in the form of sulfate, as in gypsum, or as organic sulfur compounds. Biological activity is responsible for producing organic sulfur compounds (Stevenson 1964).

The concentration of S in the 10 cores was generally uniform both with depth and from core to core. Secondary sources of S are fertilizers and, perhaps more importantly, fallout of particles from emissions generated by coal-fired power plants (Hoeft 1986). As clean-air legislation has increasingly forced power plants to reduce the concentration of S in their flue gases, fertilizer will likely become a more important source of S for growing crops. Prior to enactment of clean-air legislation much of the crops' need for S was satisfied as rainfall transported sulfur dioxide from the atmosphere to the soil (Hoeft 1986). In most cores (1, 2, 4-8, and 10) the concentration of S in the uppermost samples was higher than in samples from lower in the cores. This upholds the possibility that S was being supplied by atmospheric fallout of industrial particulates.

**Barium** Common sources of barium in soils are micas and feldspars, because barium commonly replaces potassium in minerals (Kabata-Pendias 2001). Barium is strongly adsorbed on clay minerals in soils and, therefore, is not very mobile. Barium has been found in soils as barite

(BaSO<sub>4</sub>, Allen and Hajek 1989) and hollandite [Ba(Mn<sup>4+</sup>,Mn<sup>2+</sup>)<sub>8</sub>O<sub>16</sub>, McKenzie 1989]. Barium is concentrated in manganese and phosphorus concretions, and is specifically adsorbed on oxides and hydroxides. Fertilizer can be a secondary source of Ba in agricultural soils to which granular fertilizer has been applied, and in such cases, a steadily increasing load of Ba in the surface horizon is expected.

There were no consistent trends in the concentration of Ba with depth in the profiles of the 10 cores.

**Chromium** Chromium is generally present in soils as Cr<sup>3+</sup> and this is responsible for the element's relative insolubility and immobility in soils (Kabata-Pendias 2001). The Cr<sup>3+</sup> ion is readily adsorbed by clay minerals and oxyhydroxides, such as goethite. As a result of its normally low solubility and strong sorption, Cr generally is not available to plants. The principal sources of Cr in soils are minerals such as chromite (FeCr<sub>2</sub>O<sub>4</sub>) in the parent material, and industrial fallout, such as dust and industrial contaminants.

Chromium was at its maximum concentration in the B horizons of the modern soils in eight of the 10 cores. Although Cr<sup>3+</sup> is relatively immobile in soils, because it forms strong bonds with clay minerals, it can still migrate downward in the soil profile. McKeague and St. Arnaud (1969) suggested that clay-sized particles migrate downward from the A horizon and accumulate at the top of the B horizon. If the Cr<sup>3+</sup> was adsorbed by colloidal-sized clay minerals in the A horizon, the Cr would be transported downward as the colloidal particles migrated downward. This hypothesis is supported by the previously mentioned correlation coefficients.

**Copper** Copper readily forms complex compounds with organic molecules, especially of the porphyrin type, but it also is readily adsorbed by clay minerals and iron and manganese oxyhydroxides. Copper precipitates as sulfides and carbonates under reducing conditions and as hydroxides under alkaline conditions (Kabata-Pendias 2001). As a result of any of these reactions Cu is rendered immobile, and its concentration does not vary much in the soil profile (Kabata-Pendias 2001).

We found also that the concentration of copper did not vary appreciably with depth, however, it is noted that in six of the 10 cores the concentration of Cu was higher in the uppermost sample of the profile than in the next deeper sample. In the other four cores the opposite was true.

**Nickel** The major source of nickel in soils is the parent material, however, a possible external source is airborne particulate from coal combustion. The highest Ni content is typically found in loamy soils. The Ni that is most available to plants is probably that which is associated with the iron oxyhydroxides. (Kabata-Pendias 2001).

The concentration of Ni in the ten cores did not follow any general trend with depth, but the maximum concentration in six of the ten cores was in the B horizon. This suggested downward migration of Ni, probably through leaching.

**Lead** Lead is adsorbed by clay minerals, iron and manganese oxyhydroxides, and soil organic matter. Illite was observed by Hildebrand and Blume (1974) to be a better sorbent for Pb than other clay minerals, but Kabata-Pendias (1980) did not observe that property of illite.

The concentration of Pb in the ten cores was generally highest in the surface samples and decreased to rather constant concentrations in the samples from below the surface. This suggested some external source of Pb, such as in fertilizers, or more likely, in atmospheric fallout from internal combustion engine exhaust particles or emissions from coal-burning power plants.

**Rubidium** Because the radius of the rubidium ion (1.49 Å) is approximately the same as that of the potassium ion (1.33 Å), Rb commonly substitutes for K in K-feldspar (Kabata-Pendias 2001). However, Rb is not as mobile in the soil as K, due to the stronger affinity of Rb to sorb on clay minerals and iron oxyhydroxides than K (Goldschmidt 1954). As the soil develops, therefore, Rb concentrations are expected to remain relatively stable whereas K would decrease.

Shacklette and Boerngen (1984) noted that the mean concentration of Rb in alluvial soils of the U.S. is 100 mg/kg, with a range of 55 to 140 mg/kg, and for loess and soils on silt deposits the mean and range are 75 and 45 to 100 mg/kg, respectively. In this study the Rb concentrations were higher in the alluvial and lacustrine soils of cores 1, 4, and 6 than in the other cores.

Rubidium concentrations in seven of the ten cores were at their maximum value in the B horizon. In one core (core 3), the concentration increased with depth in the profile. In core 4, the maximum occurred in the AB horizon, a transition zone between the A and B horizons. In core 8, Rb was at a minimum concentration in the 3A horizon, a horizon in an ancient, buried soil. These behaviors again suggested leaching of Rb to the B horizon.

**Tin** Tin concentrations were either less than or very near the detection limit of 5 mg/kg. The values reported here might be in error, that is, too high, because Shacklette and Boerngen (1984) reported the mean concentration of Sn in loess and soils on silt deposits in the U.S. as 1.1 mg/kg with a range of 0.3 to 1.8 mg/kg, and for alluvial soils the mean and range they reported were 1.7 and 0.3 to 4.2 mg/kg, respectively. We observed concentrations of less than 5 to 8 mg/kg in the ten cores.

**Strontium** Strontium generally is associated with soil organic matter, but may also precipitate, under alkaline conditions, as strontianite ( $\text{SrCO}_3$ ). Strontium is commonly associated with calcium geochemically. Strontium is easily mobilized during weathering of soils, especially in acidic oxidizing environments, but it is quickly incorporated in clay minerals and strongly bound by soil organic matter (Kabata-Pendias 2001). In the ten cores, the concentration of Sr exhibited a maximum concentration in seven of the ten cores. The maximum in cores 3, 5, 7, and 9 occurred in the B horizon and in cores 6 and 8 in the 2B horizon. This pattern suggested downward migration of Sr. Strontium concentrations in cores 1 and 4 did not vary appreciably.

**Vanadium** During weathering vanadium can be mobile, depending on the host minerals. Once

freed by weathering, V tends to be incorporated in clay minerals or iron oxides. Vanadium may also form complex compounds with soil organic matter, for example in porphyrin-type compounds (Kabata-Pendias 2001).

As in the case of Rb, the concentrations of V in the cores of the alluvial and lacustrine soils, cores 1, 4, and 6, were higher than in the other cores, which was probably related to the greater clay-size content of these three cores. In cores 1, 4, and 6, the average concentration of V was greater than 110 mg/kg, whereas in the other seven cores it was less than 95 mg/kg. In seven cores the concentration of V increased with depth, reaching a maximum in the B horizon, again suggesting downward leaching.

**Zinc** Zinc is strongly adsorbed by clay minerals and soil organic matter, and, therefore, is commonly present in higher concentrations in surface horizons than at deeper positions in the soil profile (Kabata-Pendias 2001). The atmospheric input of Zn from industrial fallout may be higher than its loss from the soil profile by leaching (Kabata-Pendias 2001). This lack of balance between input and output results in an accumulation of Zn in the surface horizon. Zinc is tightly bound by clay minerals and soil organic matter (Lindsay 1972) and by iron oxyhydroxides (White 1957). Therefore, the availability of Zn to plants is low.

As with other elements that are correlated with the clay-size fraction, the concentration of Zn was at a maximum in the B horizon of 7 cores. It was at its maximum concentration in the 2B horizon of core 2. Again, this suggested downward migration of Zn.

**Zirconium** Zirconium generally occurs in soil as the very stable mineral zircon ( $\text{ZrSiO}_4$ ). Zircon is very resistant to weathering and is only very slightly mobile in soils. The Zr content of soils is generally inherited from the parent materials (Kabata-Pendias 2001).

Zirconium concentrations were higher in the samples of the surface horizon than those from deeper positions in the soil profiles of this project, most likely because the silt-size fraction is concentrated in the surface samples compared with the clay-size fraction. Because zircon was found in the silt-size fraction, it is reasonable that where the silt-size fraction was most abundant, Zr would also be detected at greater concentrations. It may also be noteworthy that Zr concentrations in the cores from the alluvial and lacustrine soils, cores 1, 4, and 6, were lower than those observed in the remaining cores.

## CONCLUSIONS

In typical fashion, the maximum concentrations of certain trace elements, including chromium, rubidium, strontium, and zinc, occurred in the B horizon of several cores. The concentration of silica appeared to serve as an indicator of the amounts of sand- plus silt-sized material, and the concentration of alumina as an indicator of clay-size content in the soil samples.

Samples from cores 1, 4, and 6 had higher clay-size contents than the other seven cores. Cores 1 and 4 were collected in floodplains of rivers, and core 6 was collected from an area that had been



occupied by ancient Lake Saline. The parent materials of these soils were sediments from streams or a lake. The parent material for the other seven cores was loess.

As more cores are analyzed and more data are available, regional trends in soil composition may emerge. Presently, the data are too few to make such inferences reliably

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Table 1. Core number, county name, and depth of core.

Core Number	County	Depth of Core (ft)
1	Alexander	24.9
2	Union	19.7
3	Johnson	13.0
4	Pope	13.0
5	Pope	9.1
6	Saline	16.0
7	Williamson	17.0
8	Jackson	19.0
9	Perry	17.3
10	Perry	16.0

Table 2. Distribution of soil associations in Illinois and soil cores collected in this project

No.*	Soil Association Name	% Coverage	Number of Cores	% of Cores	% Coverage Minus % Cores
1	Port Byron-Joy	0.2	0	0	0.2
2	Tama-Muscatine-Sable	4.6	11	8	-3.4
3	Tama-Ipava-Sable	8.5	13	9.4	-0.9
4	Herrick-Virden-Piasa	2.9	3	2.2	0.7
5	Oconee-Cowden-Piasa	1.7	3	2.2	-0.5
6	Hoyleton-Cisne-Huey	4.2	9	6.5	-2.3
7	Winnebago-Durand-Ogle	0.2	0	0	0.2
8	Broadwell-Waukegan-Pillot	0.5	1	0.7	-0.2
9	Catlin-Flanagan-Drummer	5.9	9	6.5	-0.6
10	Wenona-Rutland-Streator	0.4	0	0	0.4
11	Plano-Proctor-Worthen	5.2	3	2.2	3
12	Saybrook-Dana-Drummer	3.4	5	3.6	-0.2
13	Griswold-Ringwood	0.3	1	0.7	-0.4
14	Varna-Elliott-Ashkum	2.7	3	2.2	0.5
15	Symerton-Andres-Reddick	0.5	1	0.7	-0.2
16	Swygert-Bryce-Mokena	1.5	2	1.4	0.1
17	Clarence-Rowe	0.3	0	0	0.3
18	Harco-Patton-Montgomery	0.3	0	0	0.3
19	Martinton-Milford	1	1	0.7	0.3
20	Lorenzo-Warsaw-Wea	0.7	1	0.7	0
21	Jasper-LaHogue-Selma	1.2	3	2.2	-1
22	Sparta-Dickinson-Onarga	2.1	1	0.7	1.4
23	Channahon-Dodgeville-Ashdale	0.6	2	1.4	-0.8
24	Lawson-Sawmill-Darwin	6.5	6	4.3	2.2
25	Houghton-Palms-Muskego	0.2	0	0	0.2
31	Seaton-Timula	0.6	0	0	0.6
32	Fayette-Rozetta-Stronghurst	6.3	5	3.6	2.7
33	Alford-Muren-Iva	1	1	0.7	0.3
34	Clinton-Keomah-Rushville	7.9	12	8.7	-0.8
35	Hosmer-Stoy-Weir	3.4	6	4.3	-0.9
36	Ava-Bluford-Wynoose	6.7	8	5.8	0.9
37	Westville-Pecatonica-Flagg	0.4	1	0.7	-0.3
38	Middletown-Tell-Thebes	0.3	0	0	0.3
39	Birkbeck-Sabina-Sunbury	1.3	2	1.4	-0.1
41	St. Charles-Camden-Drury	1	1	0.7	0.3
42	Dodge-Russell-Miami	1.1	3	2.2	-1.1
43	Kidder-McHenry	0.2	0	0	0.2
44	Morley-Blount-Beecher	1.8	4	2.9	-1.1
45	St. Clair-Nappanee-Frankfort	0.4	0	0	0.4
46	Markland-Colp-Del Rey	0.8	1	0.7	0.1
48	Casco-Fox-Ockley	0.5	1	0.7	-0.2
49	Martinsville-Sciotoville	0.3	1	0.7	-0.4
50	Oakville-Lamont-Alvin	1.3	3	2.2	-0.9
51	Ritchey-New Glarus-Palsgrove	0.6	1	0.7	-0.1
52	Alford-Goss-Baxter	0.5	1	0.7	-0.2
53	Alford-Wellston	0.3	0	0	0.3
54	Hosmer-Zanesville-Berks	1.4	2	1.4	0
55	Grantsburg-Zanesville-Wellston	1.1	0	0	1.1
56	Derinda-Schapville-Eleroy	0.3	0	0	0.3
57	Haymond-Petrolia-Karnak	4.9	7	5.1	-0.2

\*Soil association numbers 26-30, 40, and 47 were not used by Fehrenbacher et al. 1984.

Table 3. Texture of samples from core 1.

Sample	Depth Interval (ft)	Horizon	Clay (%)	Sand (%)	Silt (%)	Texture
1-1	0.0-0.5	Ap	39.66	3.69	56.65	silty clay loam
1-2	0.5-1.0	AB	41.19	2.69	56.12	silty clay
1-3	1.0-1.5	BA	44.67	1.70	53.63	silty clay
1-4	1.5-2.0	B	45.68	1.51	52.81	silty clay
1-8	4.0-4.6	B	52.04	1.06	46.90	silty clay
1-11	8.0-8.4	B	55.03	0.88	44.09	silt loam

Table 4. Texture of samples from core 2.

Sample	Depth Interval (ft)	Horizon	Clay (%)	Sand (%)	Silt (%)	Texture
2-1	0.0-0.5	Ap	19.60	0.76	79.64	silt loam
2-2	0.5-1.4	B	25.63	0.90	73.47	silt loam
2-3	1.4-2.0	B	23.34	3.00	73.66	silt loam
2-4	2.0-2.6	E	22.22	3.84	73.94	silt loam
2-6	2.6-3.9	2B	25.44	3.06	71.5	silt loam
2-13	8.0-10.0	2A	17.12	0.62	82.26	silt loam
2-26	16.0-16.8	4B	23.65	0.60	75.75	silt loam

Table 5. Texture of samples from core 3.

Sample	Depth Interval (ft)	Horizon	Clay (%)	Sand (%)	Silt (%)	Texture
3-1	0.0-0.5	Ap	14.92	4.65	80.43	silt loam
3-2	0.5-1.3	B	19.04	3.13	77.83	silt loam
3-3	1.3-2.0	B	18.09	5.99	75.92	silt loam
3-4	2.0-3.1	B	23.44	5.18	71.38	silt loam
3-5	3.1-4.0	B	31.23	1.51	67.26	silty clay loam
3-10	6.6-7.3	B	22.22	2.82	74.96	silt loam
3-14	9.9-11.8	B	74.07	1.18	24.75	clay

Table 6. Texture of samples from core 4.



Sample	Depth Interval (ft)	Horizon	Clay (%)	Sand (%)	Silt (%)	Texture
4-1	0.0-0.5	A	35.76	1.81	62.43	silty clay loam
4-2	0.5-1.0	A	38.87	2.19	58.94	silty clay loam
4-3	1.0-1.5	AB	40.68	5.26	54.06	silty clay
4-4	1.5-2.0	B	41.14	5.21	53.65	silty clay
4-8	3.5-4.0	B	42.56	3.06	54.38	silty clay
4-12	5.5-6.0	B	38.22	1.62	60.16	silty clay loam

Table 7. Texture of samples from core 5.

Sample	Depth Interval (ft)	Horizon	Clay (%)	Sand (%)	Silt (%)	Texture
5-1	0.0-0.4	A	17.88	1.27	80.85	silt loam
5-2	0.4-0.7	E	13.90	2.80	83.30	silt loam
5-3	0.7-1.4	B	21.54	2.56	75.90	silt loam
5-4	1.4-2.15	B	27.36	1.92	70.72	silt loam
5-7	3.83-4.83	B	17.40	1.48	81.12	silt loam
5-14	8.0-8.5	3EB	20.58	21.33	58.09	silt loam

Table 8. Texture of samples from core 6.

Sample	Depth Interval (ft)	Horizon	Clay (%)	Sand (%)	Silt (%)	Texture
6-1	0.0-0.8	Ap	23.24	13.81	62.95	silt loam
6-2	0.8-1.5	B	46.58	5.54	47.88	silty clay
6-3	1.5-2.2	B	50.72	4.28	45.00	silty clay
6-4	2.2-3.2	B	48.42	4.24	47.34	silty clay
6-6	3.5-4.1	B	46.13	6.02	47.85	silty clay
6-20	12.4-13.1	2B	46.40	0.48	53.12	silty clay

Table 9. Texture of samples from core 7.

Sample	Depth Interval (ft)	Horizon	Clay (%)	Sand (%)	Silt (%)	Texture
7-1	0.0-0.6	A	17.68	3.10	74.18	silt loam
7-2	0.6-1.1	BA	24.76	1.06	70.46	silt loam
7-3	1.1-1.6	B	28.32	1.22	70.46	silt loam
7-6	2.3-3.1	B	27.34	0.52	72.14	silt loam
7-21	12.3-14.0	BE	15.42	5.46	79.12	silt loam

Table 10. Texture of samples from core 8.

Sample	Depth Interval (ft)	Horizon	Clay (%)	Sand (%)	Silt (%)	Texture
8-1	0.0-0.9	B	20.16	24.07	55.77	silt loam
8-2	0.9-1.3	C	16.28	10.94	72.78	silt loam
8-3	1.3-1.6	C	17.08	11.40	71.52	silt loam
8-6	3.3-3.9	2B	11.14	2.41	86.45	silt loam
8-12	8.0-8.7	3A	17.92	33.81	48.27	silt loam
8-22	13.3-13.9	4C	18.98	1.70	79.32	silt loam

Table 11. Texture of samples from core 9.

Sample	Depth Interval	Horizon	Clay (%)	Sand (%)	Silt (%)	Texture
9-1	0.0-0.8	Ap	15.44	5.55	79.01	silt loam
9-2	0.8-1.2	E	14.68	5.07	80.25	silt loam
9-3	1.2-1.7	E	19.00	4.85	76.15	silt loam
9-7	2.7-3.4	B	34.32	2.34	63.34	silty clay loam
9-12	6.2-6.8	B	19.50	8.26	72.24	silt loam
9-20	14.5-15.2	3B	28.55	33.20	38.25	clay loam

Table 12. Texture of samples from core 10.

Sample	Depth Interval	Horizon	Clay (%)	Sand (%)	Silt (%)	Texture
10-1	0.0-0.7	Ap	26.24	3.58	70.18	silt loam
10-2	0.7-1.4	EB	24.15	5.76	70.09	silt loam
10-3	1.4-2.0	EB	23.85	10.84	65.31	silt loam
10-5	2.7-3.5	B	25.06	15.42	59.52	silt loam
10-10	6.0-7.0	B	31.48	16.00	52.52	silty clay loam
10-16	11.5-12.0	BC	19.48	35.70	44.82	loam

Table 13. Soil texture, soil type, and developmental environment.

Core	Texture	Soil Type	Soil Association	Developmental
1	silty clay loam	Karnak	Haymond-Petrolia-Karnak	alluvial, forest
2	silt loam	Hosmer	Hosmer-Stoy-Weir	upland forest
3	silt loam	Hosmer	Hosmer-Stoy-Weir	upland forest
4	silty clay loam	Hurst	Markland-Colp-DelRey	lacustrine, forest
5	silt loam	Hosmer	Hosmer-Stoy-Weir	upland forest
6	silt loam	Hurst	Markland-Colp-DelRey	lacustrine, forest
7	silt loam	Ava	Ava-Bluford-Wynoose	upland forest
8	silt loam	Belknap	Haymond-Petrolia-Karnak	alluvial, forest
9	silt loam	Stoy	Hosmer-Stoy-Weir	upland forest
10	silt loam	Hoyleton/Dar	Hoyleton-Cisne-Huey	upland prairie

Table 14. Correlation coefficients for constituents of cores 1 through 10 (C.I. = 95%).

	Depth	SiO <sub>2</sub>	Al <sub>2</sub> O <sub>3</sub>	Fe <sub>2</sub> O <sub>3</sub>	K <sub>2</sub> O	CaO	MgO	Na <sub>2</sub> O	TiO <sub>2</sub>
Depth	1.00								
SiO <sub>2</sub>	-0.03	1.00							
Al <sub>2</sub> O <sub>3</sub>	0.07	-0.71	1.00						
Fe <sub>2</sub> O <sub>3</sub>	0.12	-0.65	0.79	1.00					
K <sub>2</sub> O	-0.09	-0.67	0.73	0.61	1.00				
CaO	0.09	-0.58	-0.11	-0.08	0.04	1.00			
MgO	0.22	-0.84	0.63	0.51	0.69	0.50	1.00		
Na <sub>2</sub> O	0.00	0.51	-0.46	-0.53	-0.15	-0.10	-0.23	1.00	
TiO <sub>2</sub>	-0.32	-0.14	0.45	0.31	0.45	-0.40	-0.02	-0.33	1.00
P <sub>2</sub> O <sub>5</sub>	0.03	-0.26	0.15	0.31	0.35	0.09	0.11	-0.10	0.09
MnO	-0.39	-0.01	-0.05	0.03	0.11	-0.10	-0.13	-0.23	0.36
SO <sub>3</sub>	-0.13	-0.33	-0.26	-0.20	-0.21	0.78	0.04	-0.11	-0.37
Ba	-0.25	0.08	0.06	0.00	0.22	-0.21	-0.03	0.40	-0.06
Cr	-0.05	-0.47	0.61	0.50	0.32	-0.01	0.38	-0.39	0.24
Cu	-0.09	-0.73	0.69	0.77	0.75	0.12	0.57	-0.38	0.29
Hg	-0.28	0.10	-0.17	-0.08	-0.12	-0.07	-0.23	-0.10	0.24
Ni	0.02	-0.58	0.67	0.62	0.58	-0.01	0.47	-0.56	0.41
Pb	-0.42	-0.20	-0.08	0.11	0.06	0.21	-0.08	-0.22	0.11
Rb	-0.12	-0.75	0.89	0.77	0.86	-0.02	0.64	-0.51	0.52
Sr	0.06	0.06	-0.12	-0.23	0.06	0.11	0.00	0.27	-0.16
V	-0.01	-0.65	0.89	0.83	0.65	-0.11	0.56	-0.37	0.28
Zn	-0.13	-0.68	0.58	0.65	0.54	0.20	0.39	-0.40	0.17
Zr	-0.17	0.81	-0.81	-0.76	-0.64	-0.22	-0.72	0.44	-0.06
LOI	-0.12	-0.82	0.24	0.24	0.29	0.85	0.58	-0.47	-0.08
Moisture	0.07	-0.48	0.68	0.45	0.35	-0.05	0.38	-0.38	0.07
Tot C	-0.25	-0.48	-0.19	-0.15	-0.03	0.84	0.19	-0.24	-0.21
In C	0.11	-0.61	-0.08	-0.03	0.11	0.98	0.58	-0.13	-0.34
Org C	-0.40	-0.32	-0.21	-0.19	-0.10	0.60	-0.06	-0.26	-0.10
Clay	0.09	-0.74	0.91	0.74	0.58	0.03	0.70	-0.51	0.16
Sand	0.30	0.18	-0.40	-0.04	-0.49	0.19	-0.26	-0.07	-0.58
Silt	-0.26	0.60	-0.63	-0.69	-0.27	-0.15	-0.51	0.53	0.19
Sand+Silt	-0.09	0.74	-0.91	-0.74	-0.58	-0.03	-0.70	0.51	-0.16
pH	0.21	0.12	-0.43	-0.48	-0.28	0.39	0.00	0.18	-0.53

Table 14 continued. Correlation coefficients for constituents of cores 1 through 10 (C.I. = 95%).

	P <sub>2</sub> O <sub>5</sub>	MnO	SO <sub>3</sub>	Ba	Cr	Cu	Hg	Ni	Pb
P <sub>2</sub> O <sub>5</sub>	1.00								
MnO	0.46	1.00							
SO <sub>3</sub>	0.04	-0.07	1.00						
Ba	0.40	0.42	-0.20	1.00					
Cr	-0.09	-0.18	-0.01	-0.23	1.00				
Cu	0.42	0.06	0.04	0.14	0.43	1.00			
Hg	0.26	0.47	0.03	0.23	-0.24	0.08	1.00		
Ni	0.23	0.13	-0.04	-0.15	0.63	0.64	0.06	1.00	
Pb	0.43	0.62	0.30	0.28	-0.17	0.26	0.57	-0.03	1.00
Rb	0.30	0.18	-0.20	0.11	0.44	0.75	-0.09	0.65	0.14
Sr	0.16	-0.14	0.01	0.12	-0.04	-0.09	-0.16	-0.09	-0.13
V	0.23	0.06	-0.25	0.32	0.44	0.67	-0.10	0.48	0.07
Zn	0.53	0.34	0.16	0.36	0.16	0.70	0.19	0.37	0.52
Zr	-0.31	0.02	0.01	-0.10	-0.34	-0.67	0.17	-0.50	-0.10
LOI	0.20	0.11	0.71	-0.17	0.21	0.45	0.06	0.29	0.41
Moisture	0.14	-0.02	-0.08	0.17	0.44	0.48	-0.16	0.34	-0.01
Tot C	0.16	0.17	0.87	-0.15	-0.05	0.15	0.16	0.00	0.52
In C	0.06	-0.09	0.68	-0.25	0.00	0.15	-0.07	0.02	0.17
Org C	0.19	0.29	0.81	-0.07	-0.08	0.12	0.26	-0.01	0.62
Clay	0.09	-0.09	-0.15	0.05	0.57	0.63	-0.22	0.60	-0.05
Sand	0.02	-0.05	0.27	-0.22	-0.28	-0.28	-0.14	-0.32	0.07
Silt	-0.10	0.11	-0.02	0.08	-0.38	-0.43	0.29	-0.38	0.01
Sand+Silt	-0.09	0.09	0.15	-0.05	-0.57	-0.63	0.22	-0.60	0.05
pH	0.09	0.10	0.23	0.08	-0.36	-0.41	-0.08	-0.22	0.07

Table 14 continued. Correlation coefficients for constituents of cores 1 through 10 (C.I. = 95%).

	Rb	Sr	V	Zn	Zr	LOI	Moisture	Total C	Inorg C
Rb	1.00								
Sr	-0.11	1.00							
V	0.84	-0.18	1.00						
Zn	0.70	-0.11	0.74	1.00					
Zr	-0.81	0.03	-0.82	-0.74	1.00				
LOI	0.37	-0.03	0.21	0.52	-0.48	1.00			
Moisture	0.56	-0.09	0.65	0.48	-0.59	0.21	1.00		
Tot C	-0.01	0.05	-0.19	0.32	-0.10	0.88	-0.06	1.00	
In C	0.03	0.10	-0.09	0.17	-0.24	0.83	-0.09	0.78	1.00
Org C	-0.03	0.01	-0.21	0.34	0.00	0.74	-0.03	0.94	0.51
Clay	0.82	-0.16	0.87	0.58	-0.86	0.36	0.74	-0.06	0.05
Sand	-0.35	-0.05	-0.21	0.04	0.03	0.01	-0.21	0.15	0.14
Silt	-0.57	0.19	-0.71	-0.58	0.80	-0.35	-0.58	-0.03	-0.13
Sand+Silt	-0.82	0.16	-0.87	-0.58	0.86	-0.36	-0.74	0.06	-0.05
pH	-0.34	0.26	-0.36	-0.10	0.13	0.16	-0.13	0.32	0.33

	Org C	Clay	Sand	Silt	Sand+Silt	pH
Org C	1.00					
Clay	-0.12	1.00				
Sand	0.12	-0.23	1.00			
Silt	0.04	-0.82	-0.37	1.00		
Sand+Silt	0.12	-1.00	0.23	0.82	1.00	
pH	0.25	-0.25	0.34	0.04	0.25	1.00

Table 15. Chemical composition of samples selected from Core 1.

Subsample	1-01	1-02	1-03	1-04	1-08	1-11	Average
Lab. No.	R21826	R21827	R21828	R21829	R21830	R21831	
Depth Interval	0.0-0.5	0.5-1.0	1.0-1.5	1.5-2.0	4.0-4.6	8.0-8.4	
Horizon	Ap	AB	BA	B	B	B	
SiO <sub>2</sub> (%)	68.1	69.1	68.3	68.3	67.3	66.2	67.9
Al <sub>2</sub> O <sub>3</sub> (%)	13.4	13.8	14.2	14.4	15.3	15.4	14.4
Fe <sub>2</sub> O <sub>3</sub> (%)	4.90	4.68	4.75	4.79	5.17	5.53	4.97
K <sub>2</sub> O (%)	2.41	2.36	2.36	2.36	2.44	2.36	2.38
CaO (%)	1.00	0.81	0.76	0.77	0.77	0.89	0.83
MgO (%)	1.09	1.19	1.28	1.33	1.56	1.68	1.36
Na <sub>2</sub> O (%)	0.85	0.85	0.79	0.78	0.76	0.70	0.79
TiO <sub>2</sub> (%)	0.77	0.76	0.76	0.76	0.75	0.72	0.75
P <sub>2</sub> O <sub>5</sub> (%)	0.22	0.16	0.15	0.16	0.15	0.12	0.16
MnO (%)	0.18	0.18	0.10	0.14	0.08	0.09	0.13
SO <sub>3</sub> (%)	0.15	0.11	0.09	0.10	0.11	0.12	0.11
Barium (mg/kg)	701	732	675	644	703	665	687
Cadmium	<15	<15	<15	<15	<15	<15	<15
Chromium	59	62	66	66	73	66	65
Copper (mg/kg)	40	31	31	34	37	34	35
Mercury (µg/kg)	42	39	38	37	34	40	38
Nickel (mg/kg)	35	33	30	35	37	36	34
Lead (mg/kg)	30	23	22	22	23	20	23
Rubidium	114	116	119	120	130	114	119
Tin (mg/kg)	6	6	<5	<5	<5	<5	<5
Strontium	133	125	120	123	133	125	127
Vanadium	104	112	113	114	128	126	116
Zinc (mg/kg)	96	87	80	84	89	86	87
Zirconium	212	207	191	192	161	171	189
LOI (%)	6.59	5.56	6.01	5.72	5.25	5.82	5.83
Total Carbon (%)	1.73	0.91	0.74	0.75	0.47	0.37	0.83
Inorganic C (%)	0.06	0.02	0.02	0.03	0.03	0.03	0.03
Organic C (%)	1.67	0.89	0.72	0.72	0.44	0.34	0.80
pH	6.56	6.21	5.90	5.62	5.94	6.20	6.07

Table 16. Elemental composition of samples selected from Core 2.

Subsample	2-01	2-02	2-03	2-04	2-06	2-13	2-26	Average
Lab. No.	R21759	R21760	R21761	R21762	R21763	R21764	R21765	
Depth Interval	0.0-0.5	0.5-1.4	1.4-2.0	2.0-2.6	2.6-3.9	8.0-10.0	16.0-16.8	
Horizon	Ap	B	B	E	2B	2A	4B	
SiO <sub>2</sub> (%)	77.1	74.5	73.5	73.7	73.3	76.1	78.5	75.2
Al <sub>2</sub> O <sub>3</sub> (%)	9.40	11.50	12.10	11.70	12.20	10.70	10.90	11.21
Fe <sub>2</sub> O <sub>3</sub> (%)	2.98	4.18	4.69	4.77	4.59	3.63	3.11	3.99
K <sub>2</sub> O (%)	2.14	2.28	2.31	2.30	2.26	2.39	1.53	2.17
CaO (%)	0.76	0.47	0.31	0.31	0.41	0.97	0.24	0.50
MgO (%)	0.61	0.87	0.94	0.94	0.99	0.83	0.46	0.81
Na <sub>2</sub> O (%)	0.81	0.79	0.83	0.93	1.01	1.38	0.49	0.89
TiO <sub>2</sub> (%)	0.80	0.80	0.80	0.80	0.80	0.78	0.82	0.80
P <sub>2</sub> O <sub>5</sub> (%)	0.09	0.15	0.15	0.14	0.12	0.13	0.04	0.12
MnO (%)	0.10	0.08	0.11	0.09	0.10	0.10	0.02	0.09
SO <sub>3</sub> (%)	0.15	0.12	0.12	0.12	0.11	0.12	0.11	0.12
Barium (mg/kg)	603	573	592	567	628	634	349	564
Cadmium	<15	<15	<15	<15	<15	<15	<15	<15
Chromium	62	86	60	60	62	57	55	63
Copper (mg/kg)	27	30	32	31	33	27	25	29
Mercury (µg/kg)	28	31	41	42	42	20	28	33
Nickel (mg/kg)	27	46	31	29	30	31	32	32
Lead (mg/kg)	21	22	23	21	22	17	16	20
Rubidium	80	92	92	89	89	78	64	83
Tin (mg/kg)	5	6	<5	<5	6	5	<5	5
Strontium	100	102	104	114	127	179	67	113
Vanadium	71	92	98	94	95	73	66	84
Zinc (mg/kg)	43	52	55	53	58	33	24	45
Zirconium	469	346	328	352	358	384	435	382
LOI (%)	4.43	3.89	3.76	3.44	3.34	2.40	3.41	3.52
Total Carbon	1.17	0.46	0.32	0.20	0.14	0.19	0.14	0.37
Inorganic C (%)	0.06	0.02	0.02	0.02	0.02	0.02	0.01	0.02
Organic C (%)	1.11	0.44	0.30	0.17	0.12	0.17	0.13	0.35
pH	6.78	6.60	5.03	4.84	5.00	6.38	6.27	5.84



Table 17. Chemical composition of samples selected from Core 3.

Subsample	3-01	3-02	3-03	3-04	3-05	3-10	3-14	Average
Lab. No.	R21766	R21767	R21768	R21769	R21770	R21771	R21772	
Depth Interval	0.0-0.5	0.5-1.3	1.3-2.0	2.0-3.1	3.1-4.0	6.6-7.3	9.9-11.8	
Horizon	Ap	B	B	B	B	B	B	
SiO <sub>2</sub> (%)	78.6	78.2	77.5	77.1	71.7	75.0	59.6	74.0
Al <sub>2</sub> O <sub>3</sub> (%)	8.4	9.7	10.0	10.9	13.0	11.4	21.1	12.1
Fe <sub>2</sub> O <sub>3</sub> (%)	2.98	3.30	3.77	4.20	5.54	3.76	6.05	4.23
K <sub>2</sub> O (%)	1.80	1.90	1.92	1.97	1.98	2.26	1.88	1.96
CaO (%)	0.60	0.52	0.44	0.32	0.36	0.97	1.16	0.62
MgO (%)	0.42	0.49	0.56	0.69	0.95	0.84	1.96	0.84
Na <sub>2</sub> O (%)	0.93	0.94	0.95	0.91	0.89	1.23	0.21	0.87
TiO <sub>2</sub> (%)	0.84	0.85	0.86	0.87	0.84	0.82	0.85	0.85
P <sub>2</sub> O <sub>5</sub> (%)	0.09	0.05	0.06	0.07	0.09	0.33	0.05	0.11
MnO (%)	0.13	0.10	0.09	0.06	0.03	0.09	0.05	0.08
SO <sub>3</sub> (%)	0.11	0.12	0.11	0.11	0.12	0.12	0.11	0.11
Barium (mg/kg)	442	454	419	472	453	594	360	456
Cadmium	<15	<15	<15	<15	<15	<15	<15	<15
Chromium	54	89	82	58	103	62	114	80
Copper (mg/kg)	23	26	25	29	34	27	25	27
Mercury (µg/kg)	28	25	31	41	77	19	13	33
Nickel (mg/kg)	24	37	23	26	45	34	61	36
Lead (mg/kg)	23	19	20	20	20	17	13	19
Rubidium	64	72	72	79	85	90	122	83
Tin (mg/kg)	<5	6	5	6	<5	<5	5	5
Strontium	112	119	114	105	107	148	124	118
Vanadium	64	73	76	86	96	76	133	86
Zinc (mg/kg)	19	20	23	34	49	41	60	35
Zirconium	586	540	523	462	344	337	81	410
LOI (%)	4.19	2.82	2.80	2.45	4.12	2.69	6.62	3.67
Total Carbon	1.29	0.33	0.27	0.20	0.18	0.15	0.21	0.38
Inorganic C (%)	0.02	0.02	0.02	0.02	0.01	0.01	0.02	0.02
Organic C (%)	1.27	0.31	0.25	0.18	0.17	0.14	0.19	0.36
pH	6.24	6.56	5.24	4.22	4.15	6.04	6.61	5.58

Table 18. Chemical composition of samples selected from Core 4.

Subsample	4-01	4-02	4-03	4-04	4-08	4-12	Average
Lab. No.	R21832	R21833	R21834	R21835	R21836	R21837	
Depth Interval	0.0-0.5	0.5-1.0	1.0-1.5	1.5-2.0	3.5-4.0	5.5-6.0	
Horizon	A	A	AB	B	B	B	
SiO <sub>2</sub> (%)	64.8	64.5	64.2	63.9	63.2	63.6	64.0
Al <sub>2</sub> O <sub>3</sub> (%)	14.80	15.60	16.30	16.50	17.30	17.00	16.25
Fe <sub>2</sub> O <sub>3</sub> (%)	5.93	6.12	6.44	6.62	7.04	6.84	6.50
K <sub>2</sub> O (%)	2.66	2.70	2.76	2.78	2.82	2.84	2.76
CaO (%)	0.55	0.49	0.50	0.31	0.30	0.35	0.42
MgO (%)	1.12	1.13	1.16	1.14	1.28	1.27	1.18
Na <sub>2</sub> O (%)	0.51	0.48	0.47	0.48	0.46	0.49	0.48
TiO <sub>2</sub> (%)	1.04	1.06	1.08	1.05	1.02	1.02	1.05
P <sub>2</sub> O <sub>5</sub> (%)	0.21	0.17	0.17	0.18	0.17	0.19	0.18
MnO (%)	0.24	0.25	0.23	0.23	0.14	0.13	0.20
SO <sub>3</sub> (%)	0.10	0.12	0.08	0.09	0.09	0.09	0.10
Barium (mg/kg)	563	567	514	534	573	569	553
Cadmium	<15	<15	<15	<15	<15	<15	<15
Chromium	68	68	71	70	73	74	71
Copper (mg/kg)	39	36	34	36	36	35	36
Mercury (µg/kg)	76	49	33	31	44	45	46
Nickel (mg/kg)	49	46	42	42	46	43	45
Lead (mg/kg)	38	28	27	27	26	25	29
Rubidium	138	149	152	151	148	147	148
Tin (mg/kg)	6	7	6	8	7	8	7
Strontium	96	104	95	93	96	102	98
Vanadium	117	121	127	121	125	128	123
Zinc (mg/kg)	138	113	101	103	101	96	109
Zirconium	196	189	193	186	187	191	190
LOI (%)	7.45	6.95	6.30	6.37	5.75	5.39	6.37
Total Carbon (%)	1.77	1.31	0.78	0.87	0.50	0.36	0.93
Inorganic C (%)	0.03	0.04	0.04	0.03	0.03	0.03	0.03
Organic C (%)	1.74	1.27	0.74	0.84	0.47	0.33	0.90
pH	5.81	5.73	4.48	4.32	4.32	4.42	4.85

Table 19. Chemical composition of samples selected from Core 5.

Subsample	5-01	5-02	5-03	5-04	5-07	5-14	Average
Lab. No.	R21773	R21774	R21775	R21776	R21777	R21778	
Depth Interval	0.0-0.4	0.4-0.7	0.7-1.4	1.4-2.15	3.83-4.83	8.0-8.5	
Horizon	A	E	B	B	B	3EB	
SiO <sub>2</sub> (%)	78.7	79.8	78.0	74.1	75.9	80.4	77.8
Al <sub>2</sub> O <sub>3</sub> (%)	8.10	8.60	10.00	11.90	10.80	9.00	9.73
Fe <sub>2</sub> O <sub>3</sub> (%)	2.40	2.59	3.43	4.62	4.46	3.07	3.43
K <sub>2</sub> O (%)	1.81	1.80	1.88	1.87	2.12	1.45	1.82
CaO (%)	0.43	0.26	0.24	0.25	0.40	0.41	0.33
MgO (%)	0.40	0.41	0.54	0.77	0.78	0.60	0.58
Na <sub>2</sub> O (%)	0.72	0.70	0.71	0.74	1.16	0.77	0.80
TiO <sub>2</sub> (%)	0.88	0.89	0.89	0.89	0.86	0.76	0.86
P <sub>2</sub> O <sub>5</sub> (%)	0.13	0.10	0.09	0.11	0.08	0.05	0.09
MnO (%)	0.26	0.24	0.16	0.09	0.04	0.07	0.14
SO <sub>3</sub> (%)	0.13	0.11	0.12	0.11	0.12	0.11	0.12
Barium (mg/kg)	534	515	493	401	431	333	451
Cadmium	<15	<15	<15	<15	<15	<15	<15
Chromium	47	50	55	60	53	67	55
Copper (mg/kg)	24	22	27	28	27	23	25
Mercury (µg/kg)	52	39	44	51	31	41	43
Nickel (mg/kg)	25	27	27	26	24	28	26
Lead (mg/kg)	26	24	21	22	20	15	21
Rubidium	76	78	85	87	75	61	77
Tin (mg/kg)	<5	6	<5	5	<5	5	5
Strontium	93	82	89	77	123	97	94
Vanadium	63	67	78	89	73	67	73
Zinc (mg/kg)	40	37	44	49	40	21	39
Zirconium	448	456	419	323	378	460	414
LOI (%)	5.14	3.53	3.39	3.75	2.62	2.44	3.48
Total Carbon (%)	1.84	0.85	0.49	0.33	0.15	0.20	0.64
Inorganic C (%)	0.02	0.02	0.01	0.02	0.01	0.02	0.02
Organic C (%)	1.82	0.83	0.48	0.31	0.14	0.18	0.63
pH	5.83	5.35	4.90	4.50	4.08	5.90	5.09

Table 20. Chemical composition of samples selected from Core 6.

Subsample	6-01	6-02	6-03	6-04	6-06	6-20	Average
Lab. No.	R21779	R21780	R21781	R21782	R21783	R21784	
Depth Interval	0.0-0.8	0.8-1.5	1.5-2.2	2.2-3.2	3.5-4.1	12.4-13.1	
Horizon	Ap	B	B	B	B	2B	
SiO <sub>2</sub> (%)	76.0	69.1	67.4	67.5	66.7	47.8	65.8
Al <sub>2</sub> O <sub>3</sub> (%)	9.50	14.50	15.50	15.30	15.20	13.70	13.95
Fe <sub>2</sub> O <sub>3</sub> (%)	4.18	5.60	5.68	5.60	5.85	5.40	5.39
K <sub>2</sub> O (%)	1.98	2.32	2.50	2.62	2.94	3.16	2.59
CaO (%)	0.94	0.47	0.28	0.30	0.55	9.57	2.02
MgO (%)	0.66	1.17	1.46	1.53	1.79	4.54	1.86
Na <sub>2</sub> O (%)	0.81	0.71	0.78	0.89	1.07	0.78	0.84
TiO <sub>2</sub> (%)	0.82	0.81	0.79	0.79	0.76	0.69	0.78
P <sub>2</sub> O <sub>5</sub> (%)	0.09	0.05	0.05	0.07	0.13	0.12	0.09
MnO (%)	0.10	0.04	0.03	0.04	0.11	0.07	0.07
SO <sub>3</sub> (%)	0.14	0.11	0.13	0.12	0.12	0.13	0.13
Barium (mg/kg)	431	389	550	622	590	406	498
Cadmium	<15	<15	<15	<15	<15	<15	<15
Chromium	53	72	75	71	73	64	68
Copper (mg/kg)	25	31	32	33	35	35	32
Mercury (µg/kg)	22	11	5	4	35	17	16
Nickel (mg/kg)	21	26	31	33	64	38	36
Lead (mg/kg)	32	21	22	21	23	18	23
Rubidium	87	123	132	130	134	131	123
Tin (mg/kg)	<5	<5	6	5	6	7	6
Strontium	111	100	101	115	127	170	121
Vanadium	82	115	128	119	123	101	111
Zinc (mg/kg)	36	55	71	79	104	58	67
Zirconium	307	213	170	190	174	84	190
LOI (%)	4.38	4.75	5.01	4.86	4.31	13.61	6.15
Total Carbon (%)	1.12	0.50	0.54	0.34	0.33	2.89	0.95
Inorganic C (%)	0.08	0.02	0.02	0.02	0.02	2.71	0.48
Organic C (%)	1.04	0.48	0.52	0.32	0.31	0.18	0.48
pH	6.85	4.46	4.03	4.74	6.34	8.06	5.75

Table 21. Chemical composition of samples selected from Core 7.

Subsample	7-01	7-02	7-03	7-06	7-21	Average
Lab. No.	R21785	R21786	R21787	R21788	R21789	
Depth Interval	0.0-0.6	0.6-1.1	1.1-1.6	2.3-3.1	12.3-14.0	
Horizon	A	BA	B	B	BE	
SiO <sub>2</sub> (%)	76.8	75.6	74.1	72.7	77.5	75.3
Al <sub>2</sub> O <sub>3</sub> (%)	8.30	11.00	11.30	12.40	11.10	10.82
Fe <sub>2</sub> O <sub>3</sub> (%)	2.87	3.91	4.31	5.12	3.25	3.89
K <sub>2</sub> O (%)	1.81	1.99	1.99	2.16	1.96	1.98
CaO (%)	0.90	0.35	0.28	0.43	0.36	0.46
MgO (%)	0.47	0.74	0.85	1.07	0.57	0.74
Na <sub>2</sub> O (%)	0.78	0.81	0.82	1.08	1.13	0.92
TiO <sub>2</sub> (%)	0.83	0.86	0.86	0.84	0.84	0.85
P <sub>2</sub> O <sub>5</sub> (%)	0.12	0.08	0.08	0.08	0.05	0.08
MnO (%)	0.20	0.08	0.05	0.05	0.03	0.08
SO <sub>3</sub> (%)	0.16	0.11	0.11	0.12	0.11	0.12
Barium (mg/kg)	552	514	492	508	389	491
Cadmium	<15	<15	<15	<15	<15	<15
Chromium	55	58	63	62	49	57
Copper (mg/kg)	24	28	29	33	25	28
Mercury (µg/kg)	60	52	59	34	23	46
Nickel (mg/kg)	27	26	26	28	23	26
Lead (mg/kg)	24	22	20	22	15	21
Rubidium	74	87	86	82	75	81
Tin (mg/kg)	<5	<5	<5	6	<5	<5
Strontium	93	95	93	127	106	103
Vanadium	64	92	94	90	69	82
Zinc (mg/kg)	39	40	44	52	19	39
Zirconium	442	352	341	324	313	354
LOI (%)	5.86	3.64	4.62	3.59	2.70	4.08
Total Carbon (%)	1.97	0.39	0.29	0.26	0.09	0.60
Inorganic C (%)	0.11	0.01	0.01	0.01	0.01	0.03
Organic C (%)	1.86	0.38	0.28	0.25	0.08	0.57
pH	6.48	5.38	4.14	3.74	5.83	5.11

Table 22. Chemical composition of samples selected from Core 8.

Subsample	8-01	8-02	8-03	8-06	8-12	8-22	Average
Lab. No.	R21838	R21839	R21840	R21841	R21893	R21894	
Depth Interval	0.0-0.9	0.9-1.3	1.3-1.6	3.3-3.9	8.0-8.7	13.3-13.9	
Horizon	B	C	C	2B	3A	4C	
SiO <sub>2</sub> (%)	53.4	72.1	79.0	81.0	79.0	76.3	73.5
Al <sub>2</sub> O <sub>3</sub> (%)	7.90	8.60	9.10	8.50	8.30	10.90	8.88
Fe <sub>2</sub> O <sub>3</sub> (%)	3.09	3.03	3.15	2.03	4.81	3.47	3.26
K <sub>2</sub> O (%)	1.54	1.86	1.92	2.05	1.64	2.24	1.88
CaO (%)	12.11	3.29	0.67	0.52	0.63	1.07	3.05
MgO (%)	1.11	0.57	0.50	0.44	0.57	0.90	0.68
Na <sub>2</sub> O (%)	0.58	0.88	0.95	1.09	0.87	1.39	0.96
TiO <sub>2</sub> (%)	0.56	0.71	0.73	0.75	0.60	0.79	0.69
P <sub>2</sub> O <sub>5</sub> (%)	0.15	0.08	0.06	0.15	0.19	0.20	0.14
MnO (%)	0.07	0.08	0.08	0.05	0.30	0.05	0.11
SO <sub>3</sub> (%)	0.50	0.23	0.09	0.09	0.15	0.11	0.20
Barium (mg/kg)	370	449	459	532	733	680	537
Cadmium	<15	<15	<15	<15	<15	<15	<15
Chromium	62	53	67	60	54	43	57
Copper (mg/kg)	31	26	24	24	25	26	26
Mercury (µg/kg)	40	38	29	21	19	25	29
Nickel (mg/kg)	24	24	26	25	24	9	22
Lead (mg/kg)	38	28	18	16	25	19	24
Rubidium	67	68	68	68	59	77	68
Tin (mg/kg)	5	5	5	<5	<5	<5	5
Strontium	179	131	119	144	152	201	154
Vanadium	70	60	64	53	91	100	73
Zinc (mg/kg)	110	69	23	20	64	68	59
Zirconium	255	408	405	401	333	393	366
LOI (%)	18.68	7.90	3.13	2.39	2.63	2.20	6.16
Total Carbon (%)	7.79	3.60	0.50	0.58	0.19	0.16	2.14
Inorganic C (%)	2.57	0.54	0.02	0.03	0.03	0.03	0.54
Organic C (%)	5.22	3.06	0.48	0.55	0.16	0.13	1.60
pH	7.28	7.44	7.34	7.31	7.40	7.41	7.36

Table 23. Chemical composition of samples selected from Core 9.

Subsample	9-01	9-02	9-03	9-07	9-12	9-20	Average
Lab. No.	R21895	R21896	R21897	R21898	R21899	R21900	
Depth Interval	0.0-0.8	0.8-1.2	1.2-1.7	2.7-3.4	6.2-6.8	14.5-15.2	
Horizon	Ap	E	E	B	B	3B	
SiO <sub>2</sub> (%)	78.7	81.4	78.6	71.5	78.9	75.7	77.5
Al <sub>2</sub> O <sub>3</sub> (%)	8.00	8.10	9.40	13.20	10.40	8.90	9.67
Fe <sub>2</sub> O <sub>3</sub> (%)	2.99	2.95	3.75	5.51	2.21	7.55	4.16
K <sub>2</sub> O (%)	1.72	1.74	1.78	1.97	2.02	1.44	1.78
CaO (%)	0.83	0.52	0.53	0.60	1.00	0.60	0.68
MgO (%)	0.37	0.35	0.46	0.97	0.62	0.37	0.52
Na <sub>2</sub> O (%)	1.02	1.07	1.00	1.05	1.31	0.45	0.98
TiO <sub>2</sub> (%)	0.76	0.82	0.82	0.76	0.75	0.59	0.75
P <sub>2</sub> O <sub>5</sub> (%)	0.21	0.10	0.10	0.12	0.10	0.21	0.14
MnO (%)	0.32	0.26	0.14	0.12	0.03	0.04	0.15
SO <sub>3</sub> (%)	0.09	0.10	0.12	0.10	0.12	0.10	0.11
Barium (mg/kg)	824	739	683	785	775	398	701
Cadmium	<15	<15	<15	<15	<15	<15	<15
Chromium	29	24	36	100	28	45	44
Copper (mg/kg)	25	19	22	28	25	33	25
Mercury (µg/kg)	104	68	56	22	61	37	58
Nickel (mg/kg)	6	<5	<5	16	17	35	14
Lead (mg/kg)	57	24	22	23	17	24	28
Rubidium	68	70	76	82	62	66	71
Tin (mg/kg)	8	<5	<5	5	<5	<5	6
Strontium	122	125	114	132	182	85	127
Vanadium	79	81	94	119	85	92	92
Zinc (mg/kg)	80	45	52	82	56	79	66
Zirconium	403	413	398	303	373	256	358
LOI (%)	4.60	2.29	2.94	3.78	2.20	3.50	3.22
Total Carbon (%)	1.71	0.31	0.29	0.18	0.17	0.17	0.47
Inorganic C (%)	0.05	0.03	0.03	0.03	0.03	0.02	0.03
Organic C (%)	1.66	0.28	0.26	0.15	0.14	0.15	0.44
pH	7.22	7.07	7.11	3.84	5.68	5.88	6.13

Table 24. Chemical composition of samples selected from Core 10.

Subsample	10-01	10-02	10-03	10-05	10-10	10-16	Average
Lab. No.	R21901	R21902	R21903	R21904	R21905	R21906	
Depth Interval	0.0-0.7	0.7-1.4	1.4-2.0	2.7-3.5	6.0-7.0	11.5-12.0	
Horizon	Ap	EB	EB	B	B	BC	
SiO <sub>2</sub> (%)	74.8	76.0	78.6	78.9	75.9	79.6	77.3
Al <sub>2</sub> O <sub>3</sub> (%)	11.10	10.80	10.30	9.80	11.70	9.00	10.45
Fe <sub>2</sub> O <sub>3</sub> (%)	4.05	3.68	2.98	3.51	3.98	3.67	3.65
K <sub>2</sub> O (%)	1.89	1.98	1.86	1.65	1.69	1.72	1.80
CaO (%)	0.79	0.83	0.70	0.63	0.65	0.54	0.69
MgO (%)	0.81	0.77	0.59	0.52	0.66	0.54	0.65
Na <sub>2</sub> O (%)	1.19	1.25	1.10	0.97	0.87	0.89	1.05
TiO <sub>2</sub> (%)	0.78	0.78	0.77	0.71	0.75	0.61	0.73
P <sub>2</sub> O <sub>5</sub> (%)	0.07	0.06	0.05	0.11	0.06	0.09	0.07
MnO (%)	0.07	0.08	0.06	0.06	0.09	0.02	0.06
SO <sub>3</sub> (%)	0.12	0.08	0.09	0.10	0.09	0.09	0.10
Barium (mg/kg)	738	699	579	481	494	422	569
Cadmium	<15	<15	<15	<15	<15	<15	<15
Chromium	61	45	42	36	77	36	50
Copper (mg/kg)	29	27	22	19	23	22	24
Mercury (µg/kg)	31	14	10	26	29	25	23
Nickel (mg/kg)	15	11	12	8	13	7	11
Lead (mg/kg)	22	19	17	17	21	18	19
Rubidium	78	81	81	75	92	72	80
Tin (mg/kg)	5	<5	<5	<5	<5	<5	<5
Strontium	151	163	145	116	114	105	132
Vanadium	103	99	92	90	98	80	94
Zinc (mg/kg)	62	62	57	53	62	59	59
Zirconium	369	344	357	318	294	307	332
LOI (%)	3.99	3.28	2.57	2.64	3.17	2.23	2.98
Total Carbon (%)	0.74	0.33	0.17	0.1	0.12	0.09	0.26
Inorganic C (%)	0.03	0.02	0.02	0.02	0.02	0.02	0.02
Organic C (%)	0.71	0.31	0.15	0.08	0.1	0.07	0.24
pH	5.56	6.92	6.75	7.12	6.74	7.04	6.69



Table 25. Means and ranges of elemental concentrations in the surficial samples from ten cores of Illinois soils compared with results for surface soils found by other researchers.

Element	This Work		Shacklette and Boerngen <sup>(2)</sup>	
	Mean	Range	Mean	Range
Silicon (%)	34.0	25.0-36.8	NR <sup>†</sup>	29-45**
Aluminum (%)	5.23	4.16-7.86	NR	<0.07-8.5**
Iron (%)	2.54	1.68-4.15	2.1 <sup>(1)</sup>	0.1-4.0**
Potassium (%)	1.64	1.28-2.21	1.7 <sup>(1)</sup>	0.22-2.25**
Calcium (%)	1.35	0.31-8.66	0.40 <sup>(1)</sup>	0.3-1.5**
Magnesium (%)	0.42	0.22-0.67	0.30 <sup>(1)</sup>	0.005-1.25**
Sodium (%)	0.61	0.38-0.88	NR	0.6-1.25**
Titanium (%)	0.48	0.34-0.62	0.41	0.05-1.0
Phosphorus (%)	0.060	0.030-0.096	0.065 <sup>(1)</sup>	0.013-0.68**
Manganese (%)	0.13	0.052-0.25	0.052	0.005-0.15
Sulfur (%)	0.065	0.036-0.20	0.08 <sup>(1)</sup>	<0.08-0.5**
Barium (mg/kg)	576	370-824	675	200-1500
Cadmium (mg/kg)	<5	—	0.27	0.13-0.55*
Chromium (mg/kg)	55	29-68	55	10-100
Copper (mg/kg)	29	23-40	25	7-100
Mercury (µg/kg)	36	4-104	70	20-360
Nickel (mg/kg)	25	6-49	17	5-30
Lead (mg/kg)	31	21-57	19	10-30
Rubidium (mg/kg)	85	64-138	75	45-100
Tin (mg/kg)	<5	<5-8	NR	NR
Strontium (mg/kg)	119	93-179	305	20-1000
Vanadium (mg/kg)	82	63-117	87	20-150*
Zinc (mg/kg)	66	19-138	58.5	20-109
Zirconium (mg/kg)	369	196-586	NR	NR

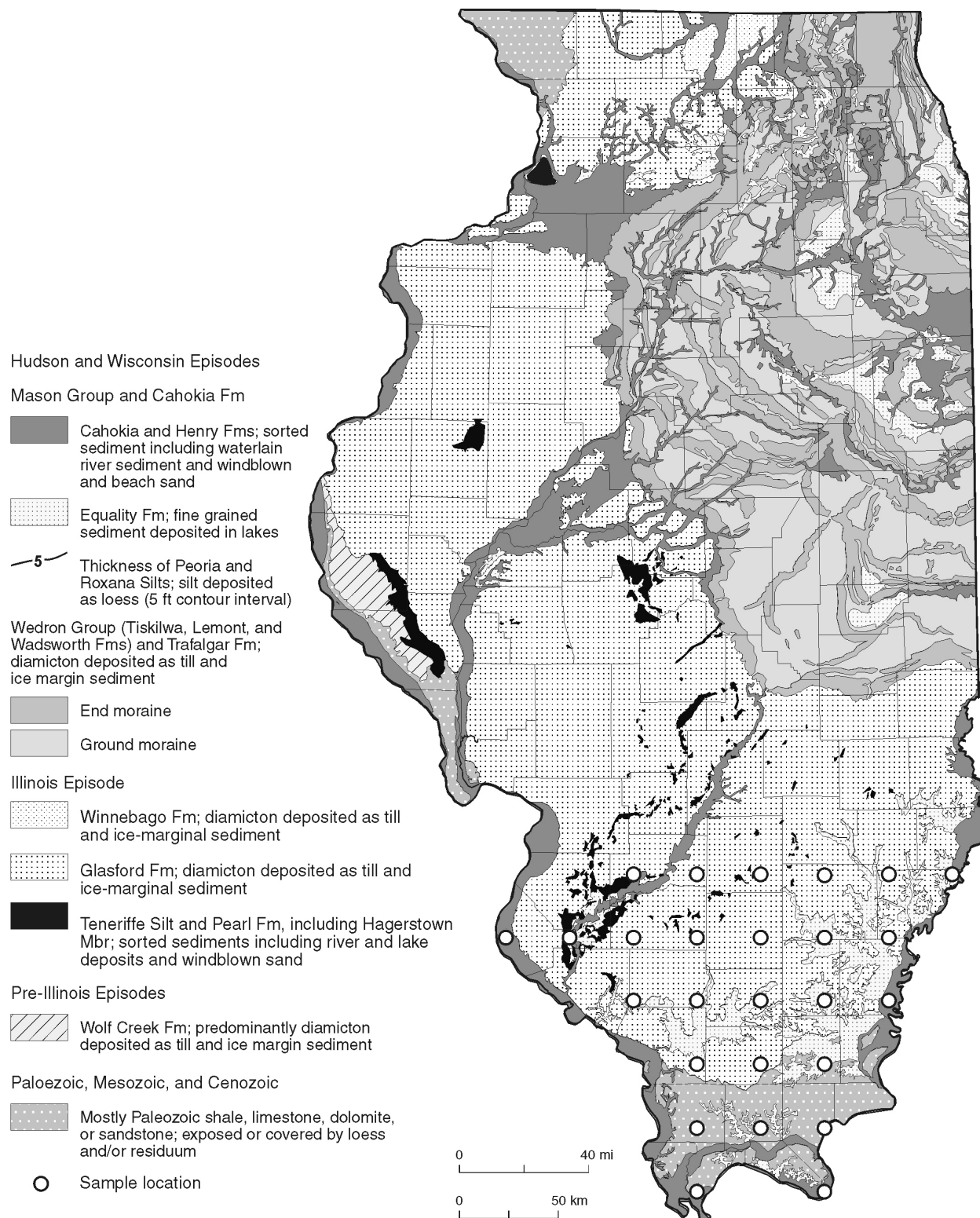
<sup>†</sup>NR = not reported

\*\*Average concentration in Illinois soils; from Shacklette and Boerngen, 1984.

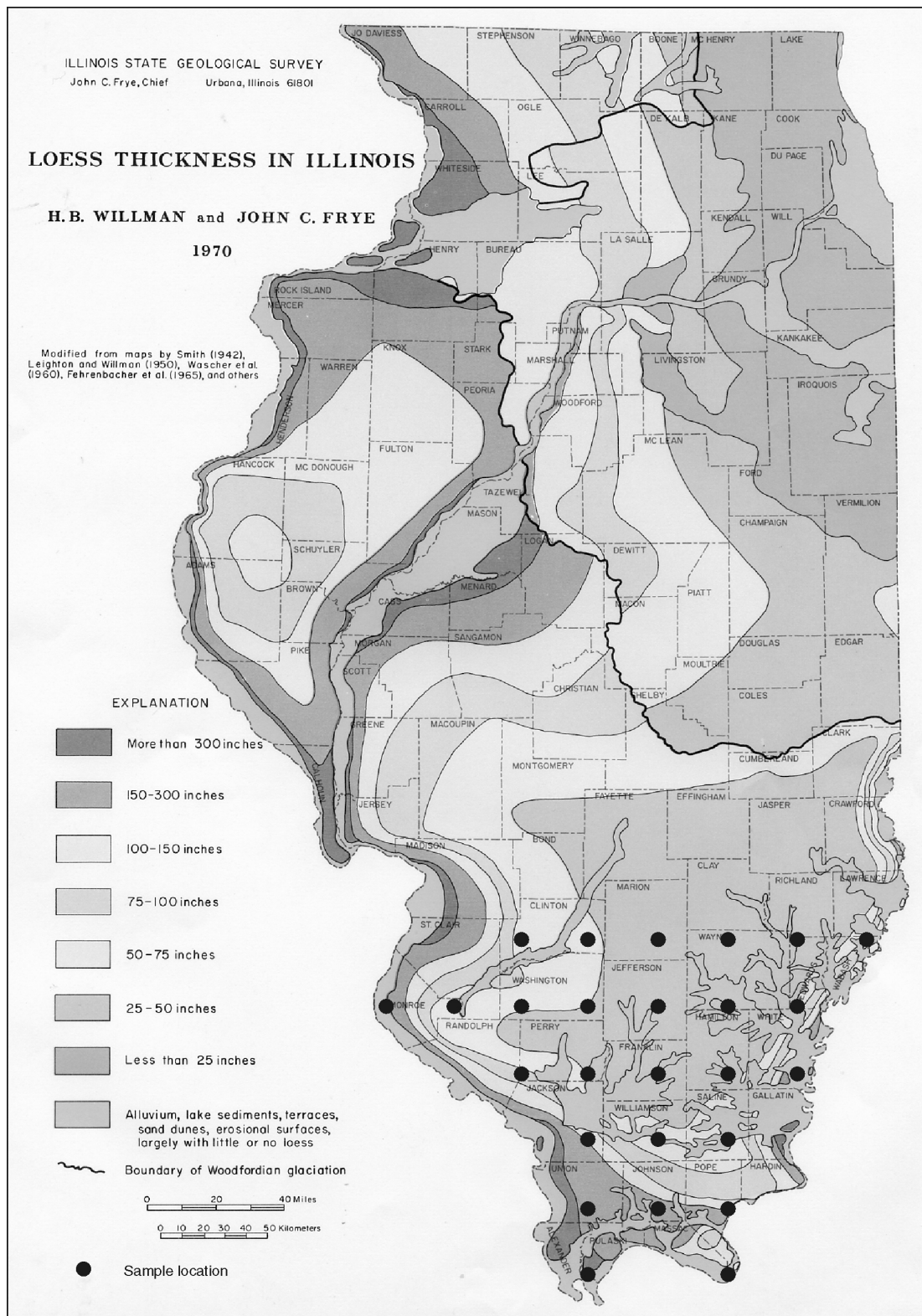
\*Values for loamy and clay soils, U.S.; from Shacklette and Boerngen, 1984.

<sup>(1)</sup>Values for humid region soils; from Severson and Shacklette, 1988.

<sup>(2)</sup>Values for loess and silty soils, U.S., unless note otherwise; from Shacklette and Boerngen, 1984.



**Figure 1** Sample locations related to the glacial episodes in Illinois.



**Figure 2** Sample locations related to loess thickness in Illinois.

# QUATERNARY DEPOSITS OF ILLINOIS

Jerry A. Lineback  
1981

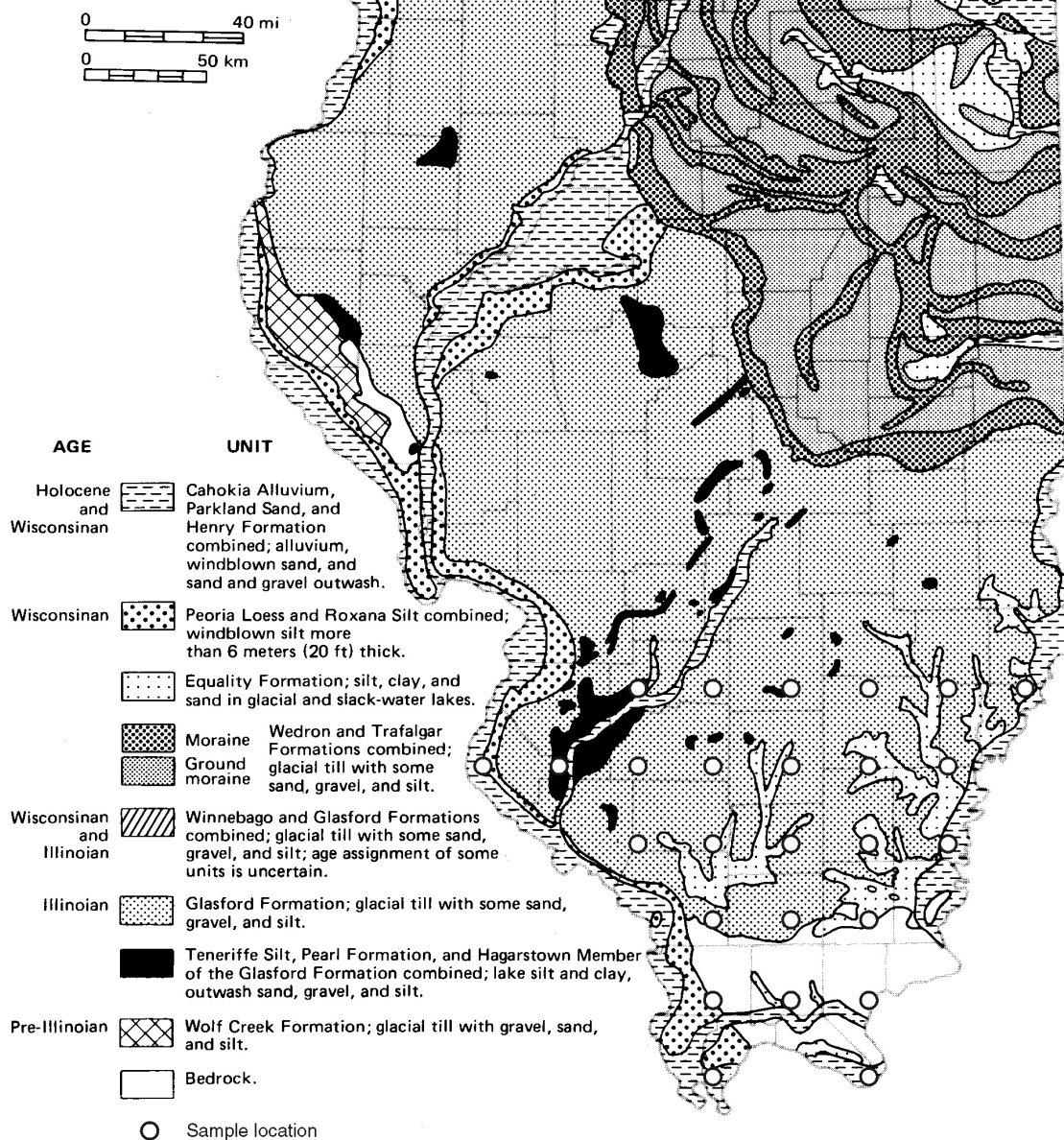
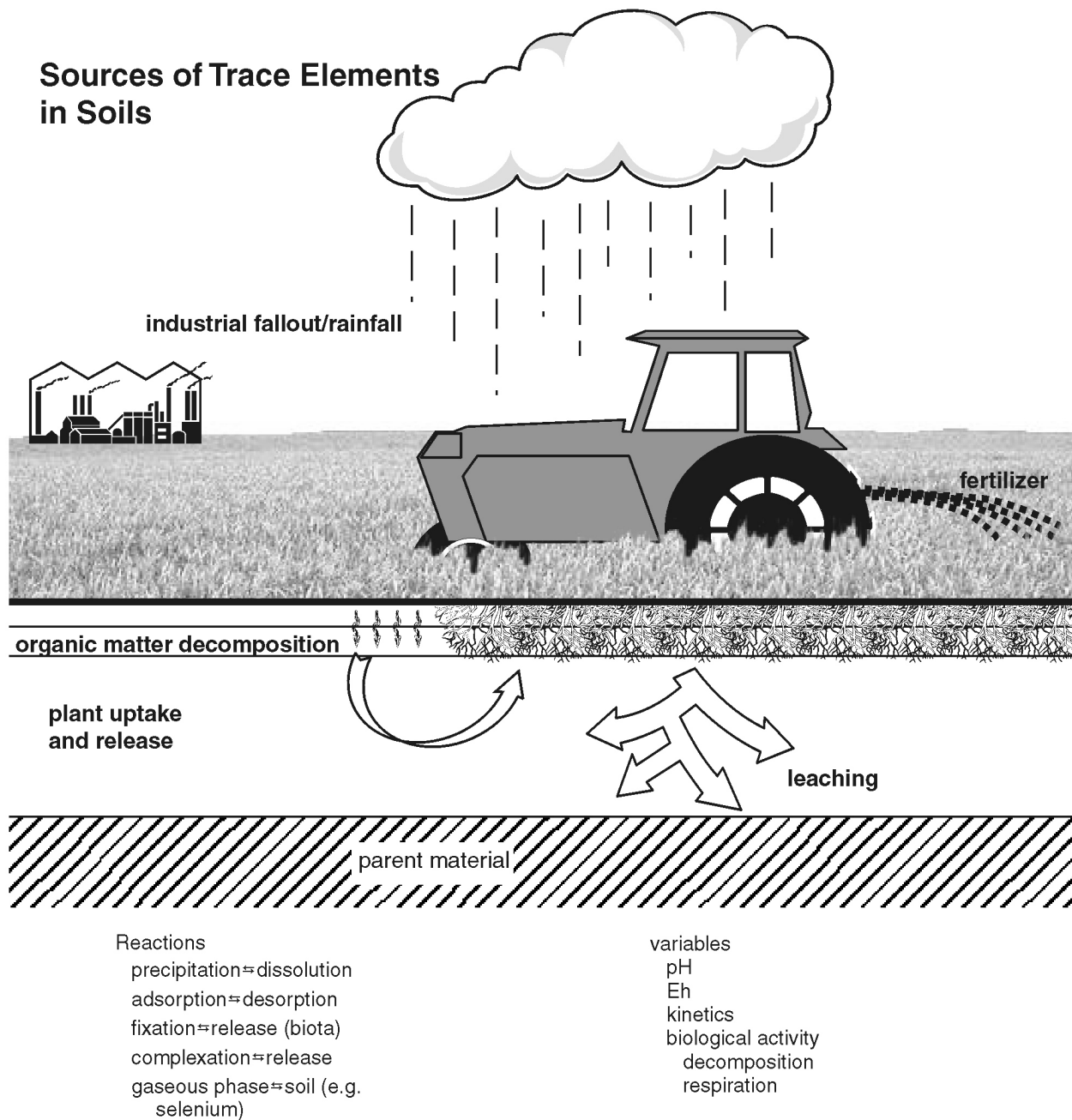
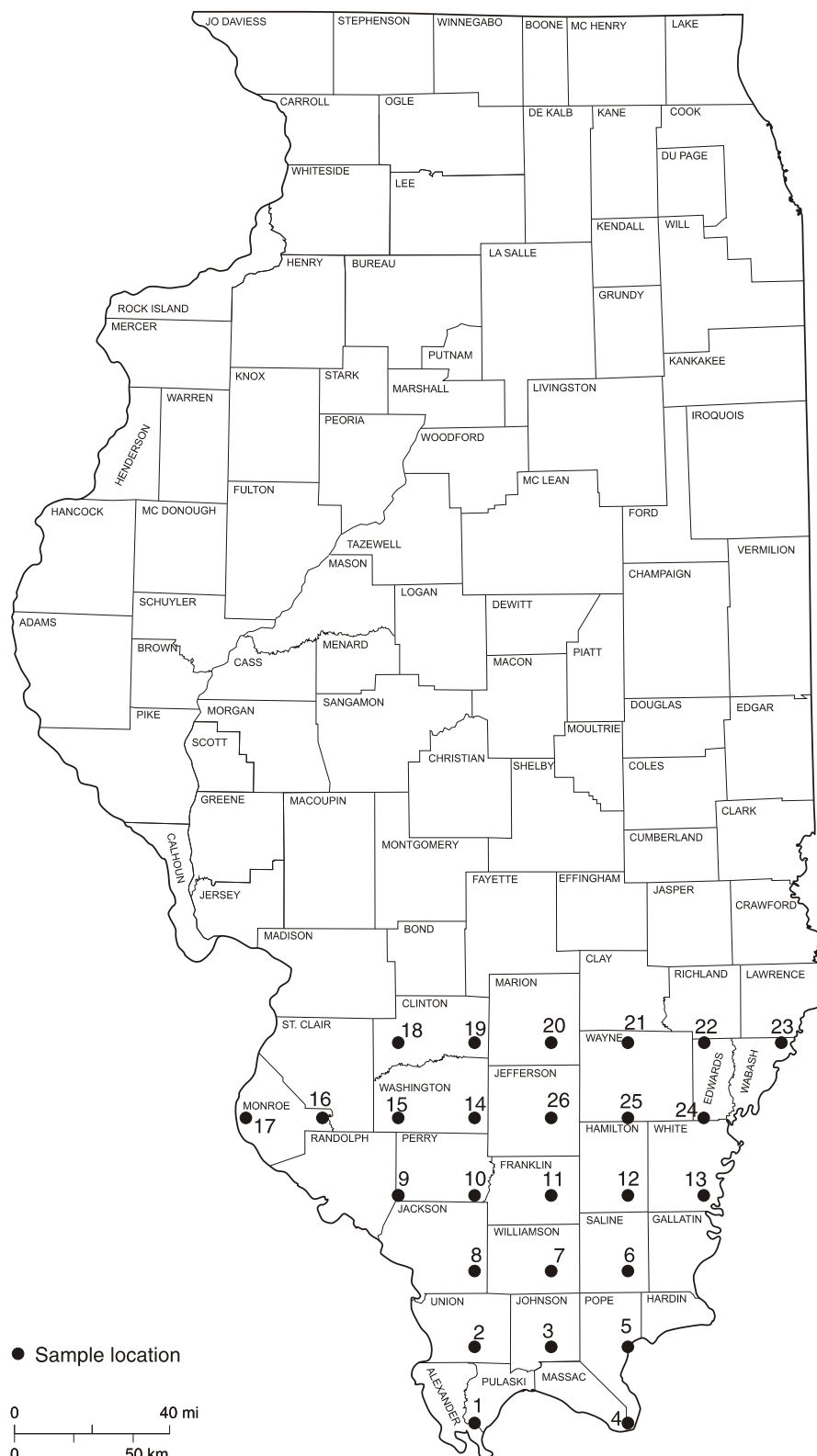


Figure 3 Sample locations related to Quaternary deposits in Illinois.

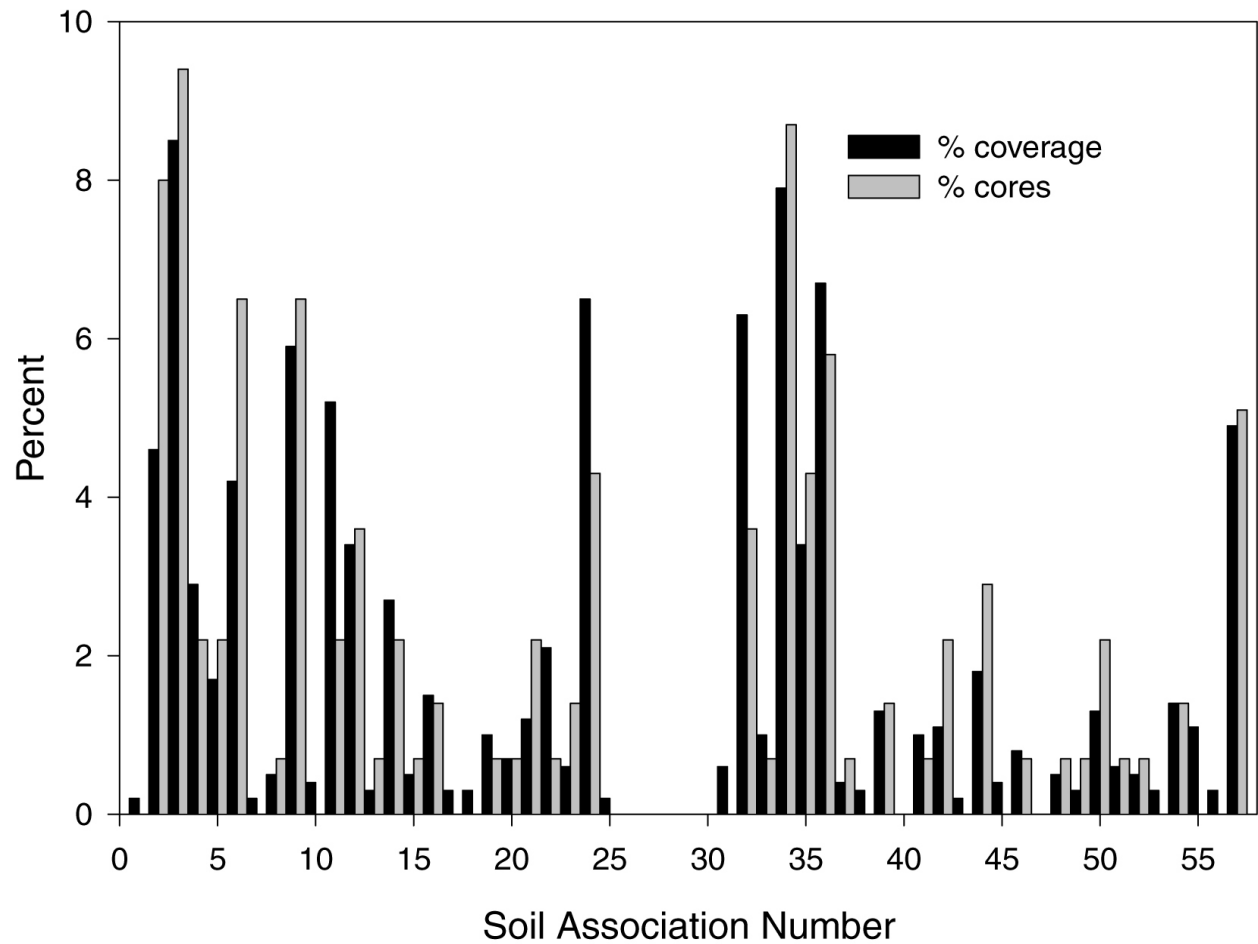
## Sources of Trace Elements in Soils



**Figure 4** Schematic diagram of various input and output functions that affect trace elements, reactions that affect soil chemistry, and variables that affect reactions.



**Figure 5** Locations of cores collected in 1998.



**Figure 6** Percentage of cores to be collected in each soil association (light-colored bars) compared with the percentage of areal coverage in Illinois of each soil association (dark-colored bars).

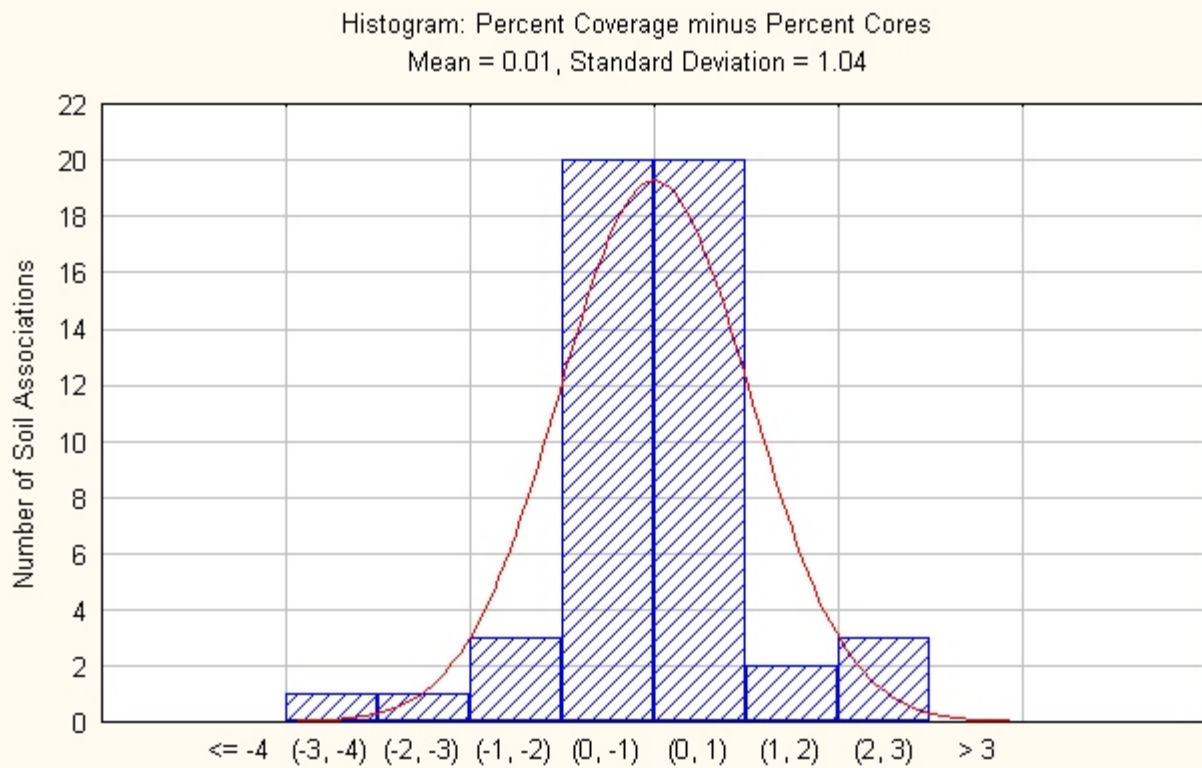


Figure 7. Distribution of differences, percentage of areal coverage minus percentage of cores collected in each soil association. Mean = 0.01, sd = 1.04



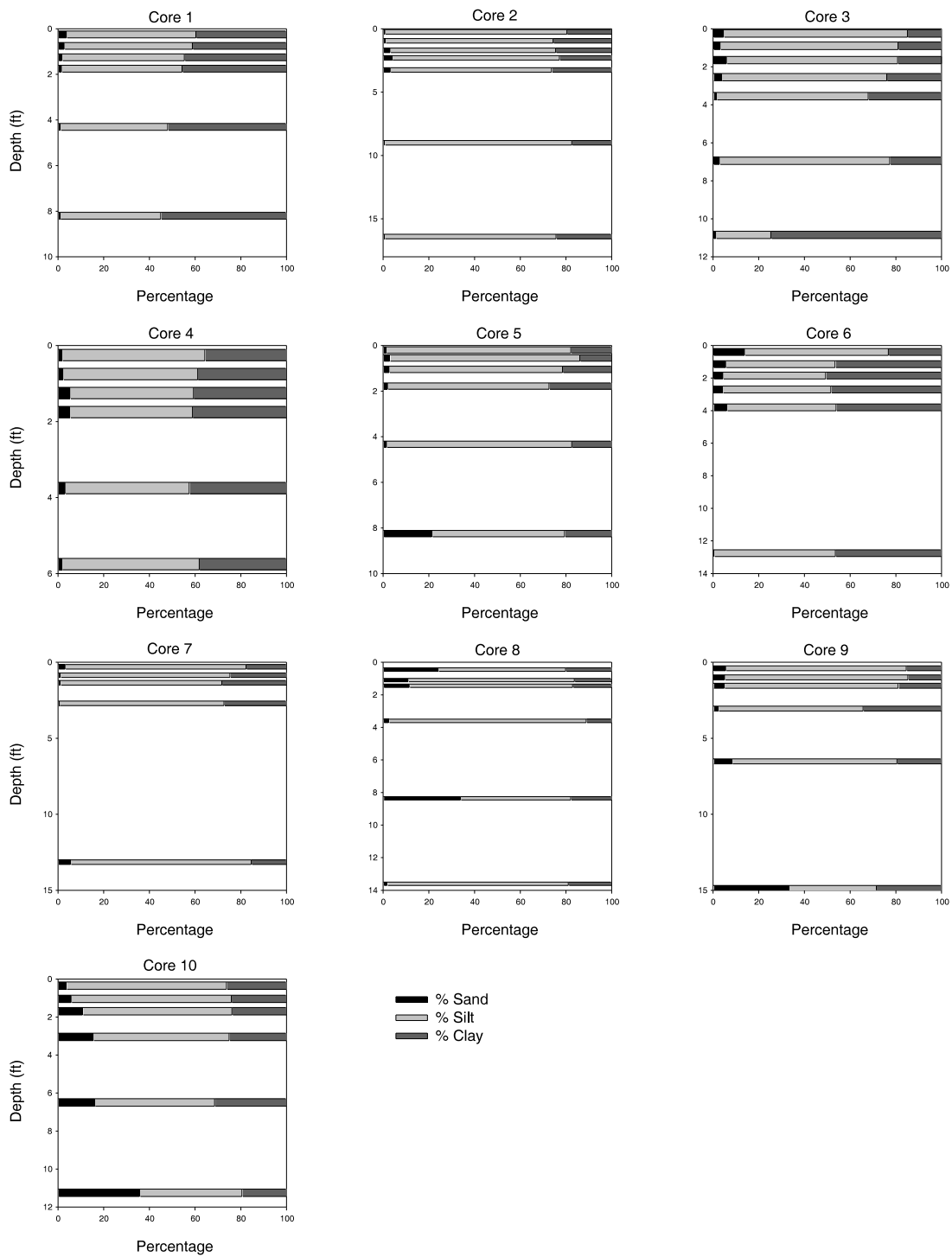


Figure 8. Sand, silt, and clay contents of soil samples

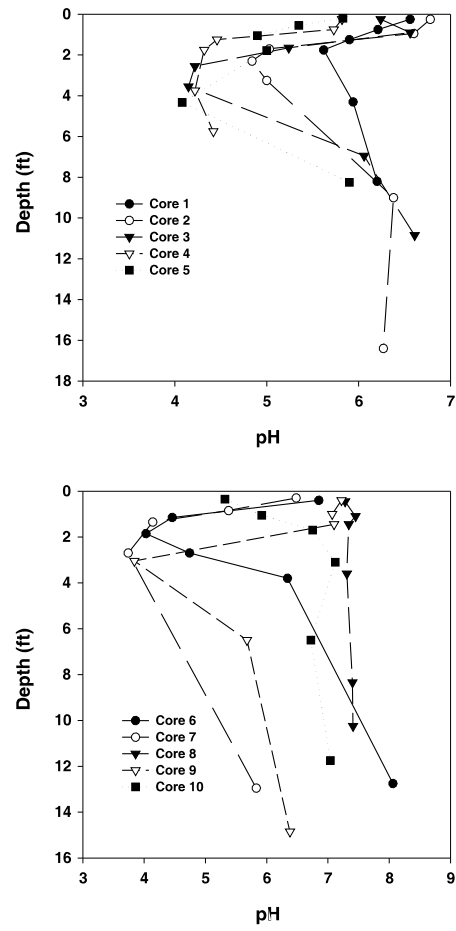


Figure 9. Soil pH versus depth.

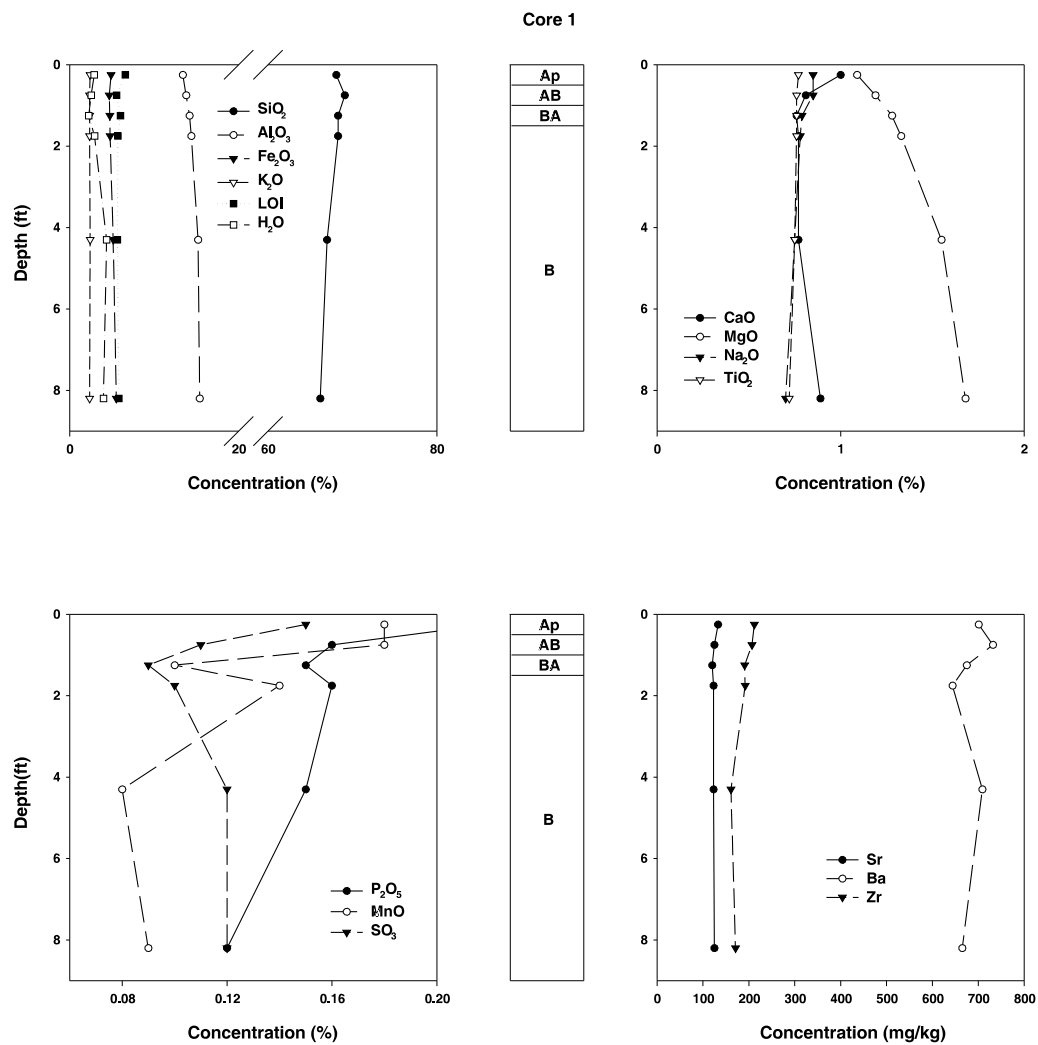


Figure 10. Elemental concentration versus depth in core 1.

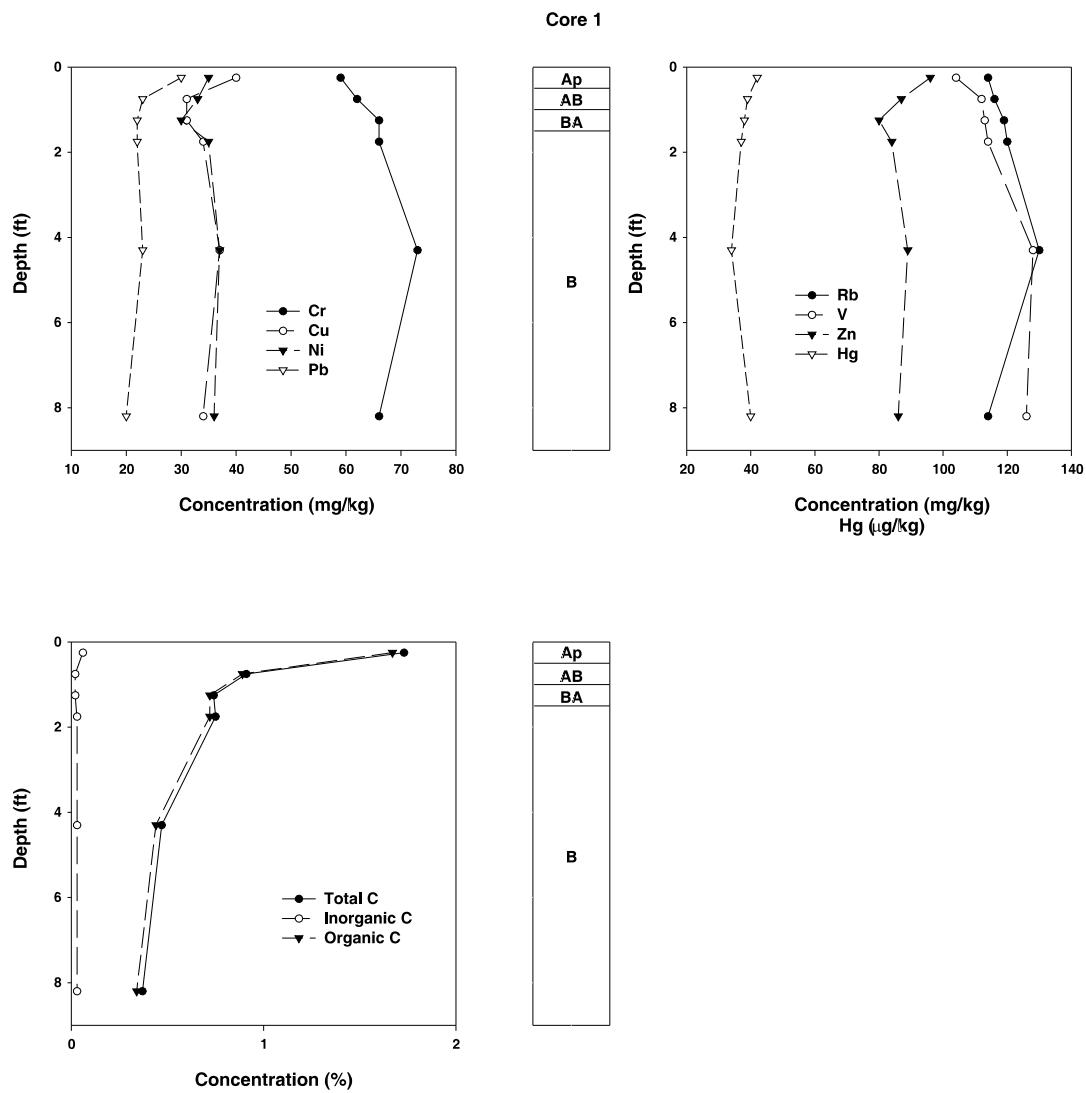


Figure 11.Elemental concentrations versus depth in core 1.

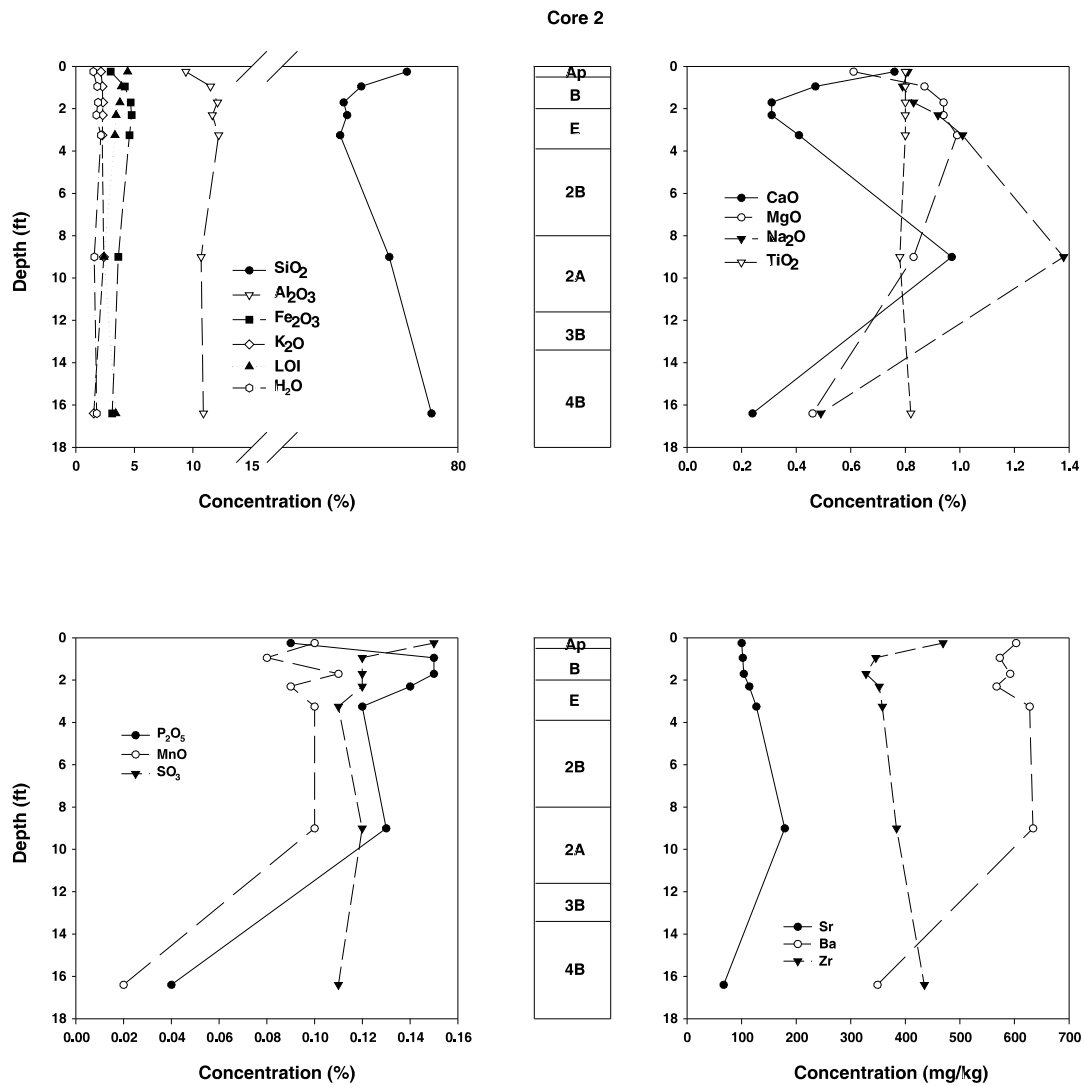


Figure 12. Elemental concentration versus depth in core 2.

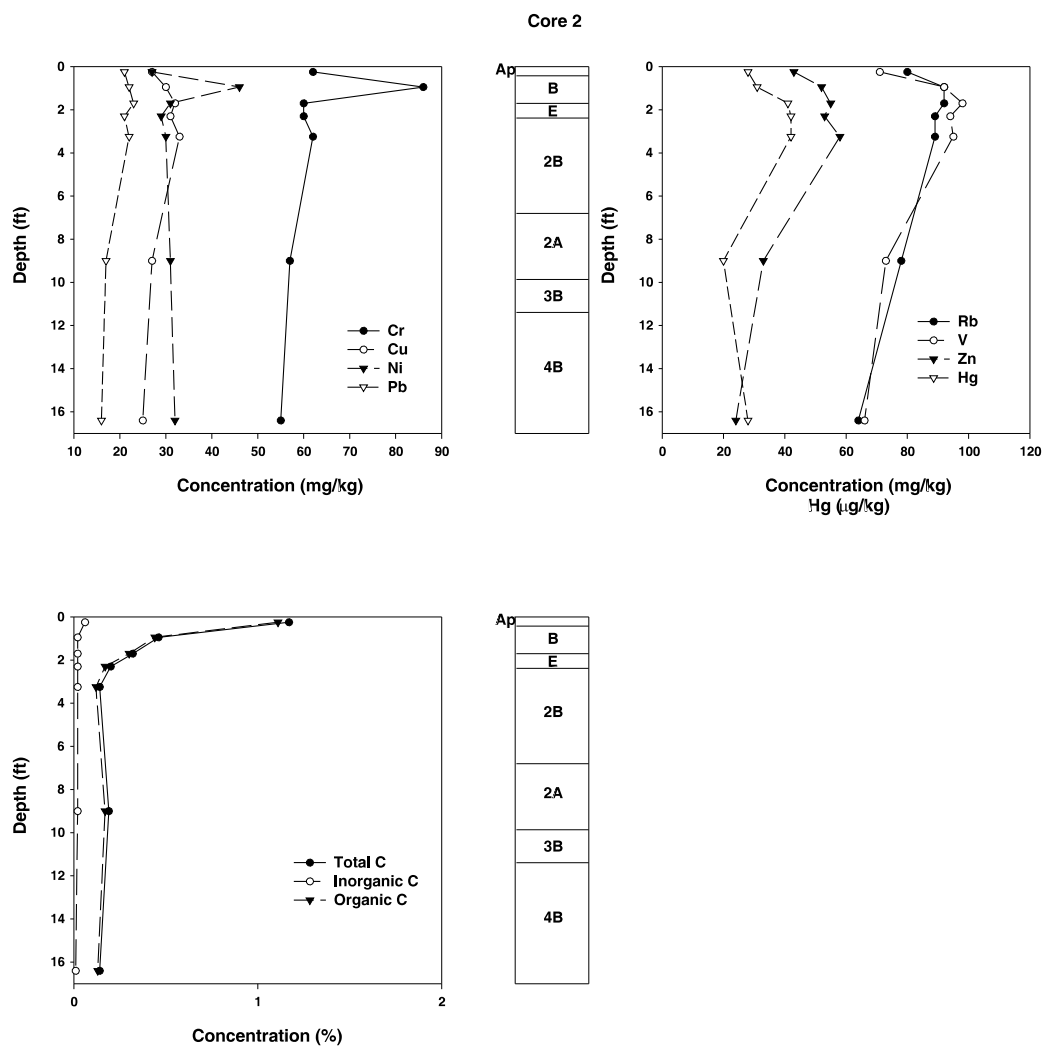


Figure 13. Elemental concentrations versus depth in core 2.

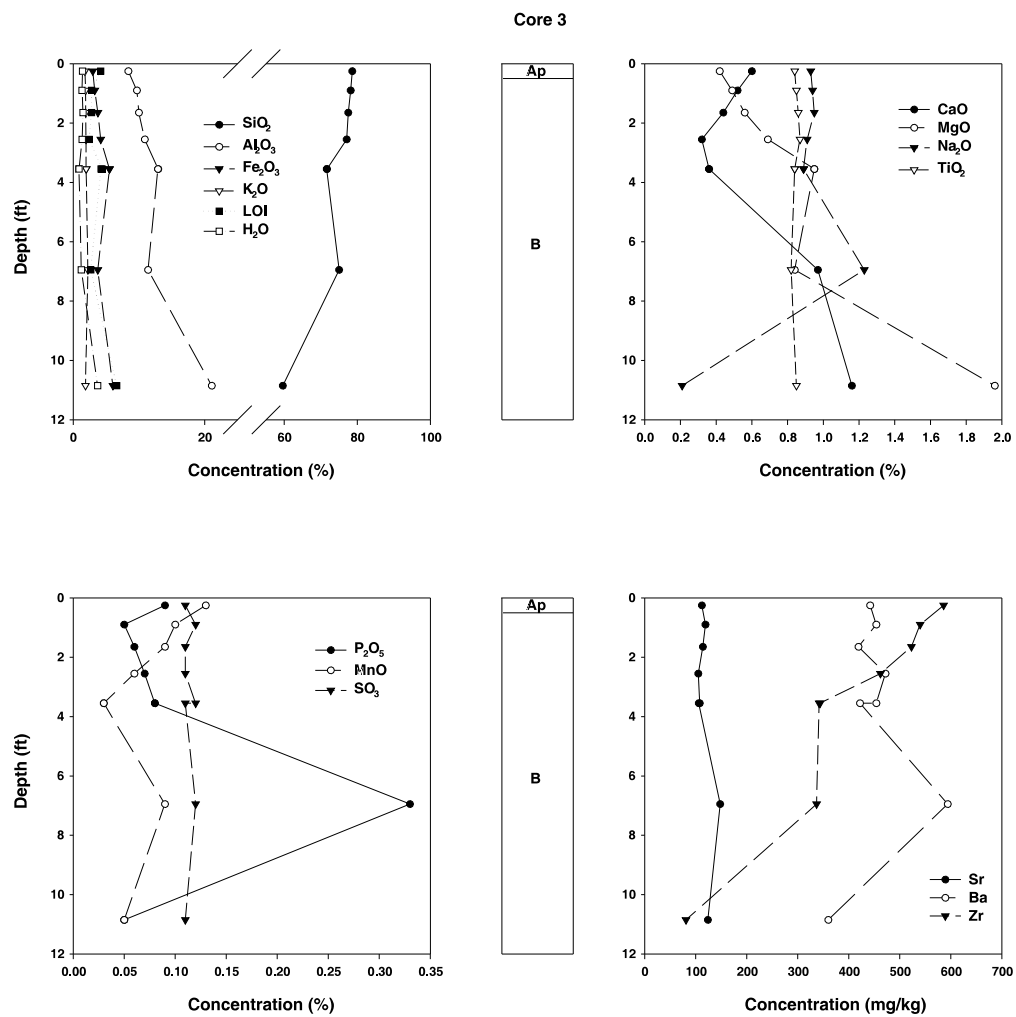


Figure 14. Elemental concentration versus depth in core 3.

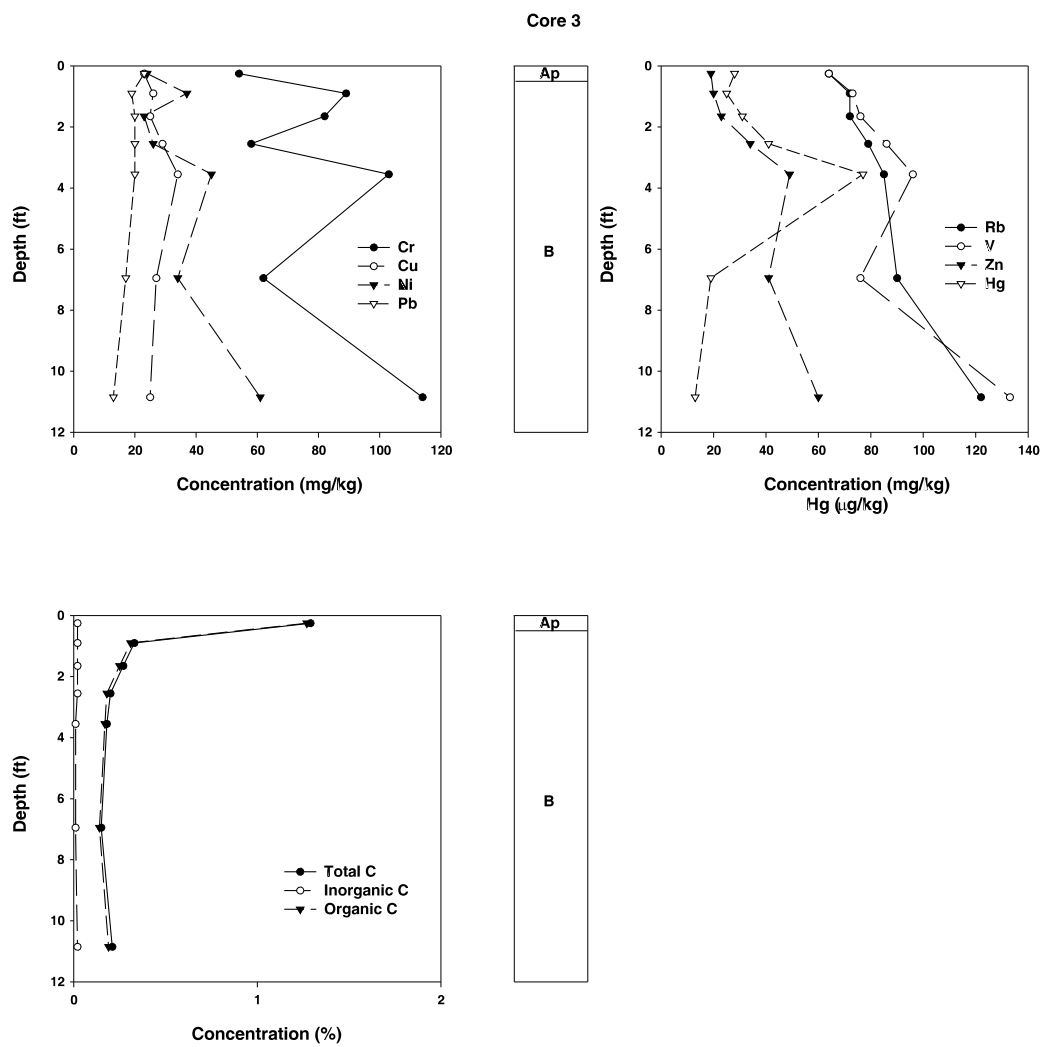


Figure 15. Elemental concentrations versus depth in core 3.



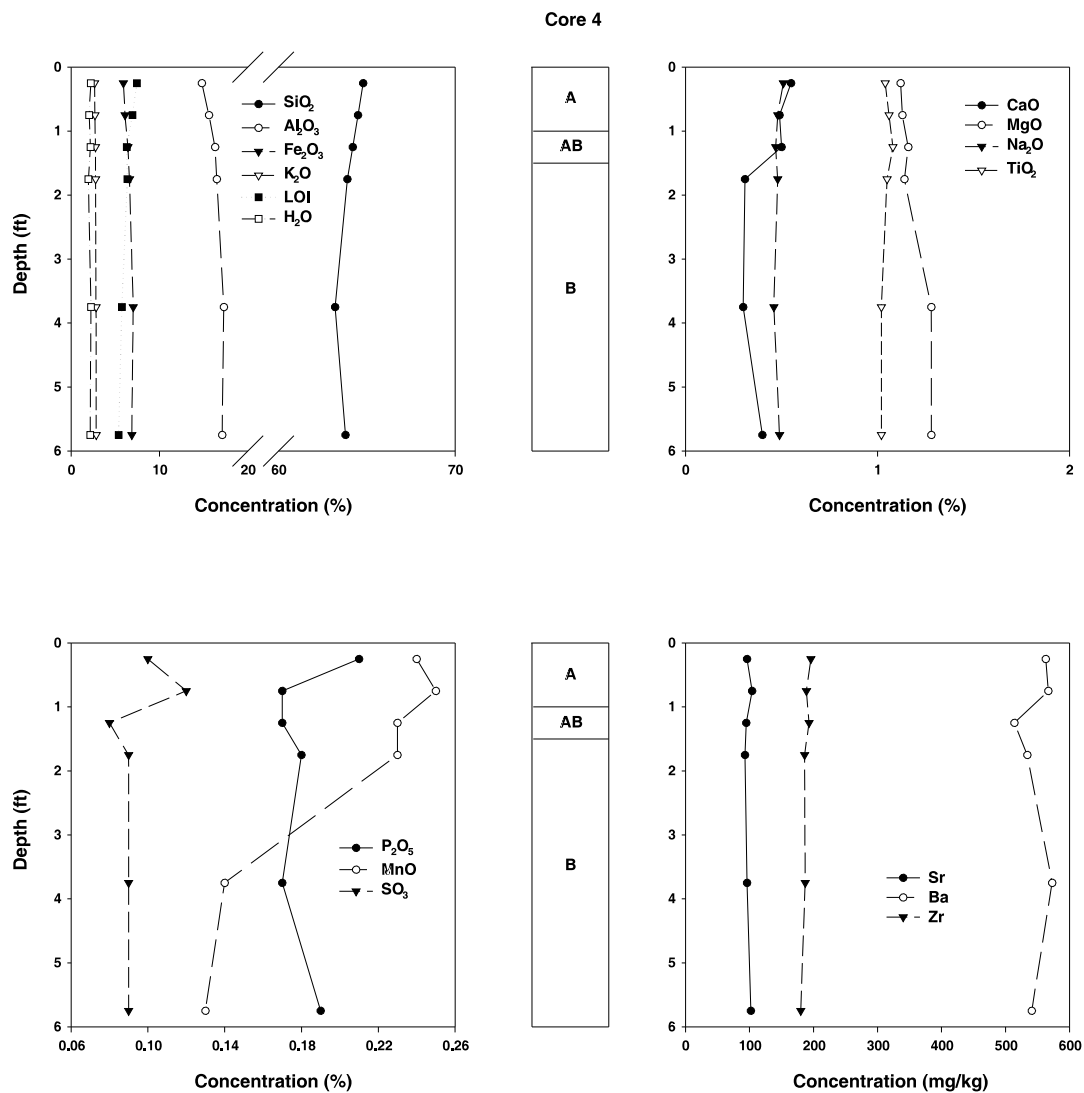


Figure 16. Elemental concentration versus depth in core 4.

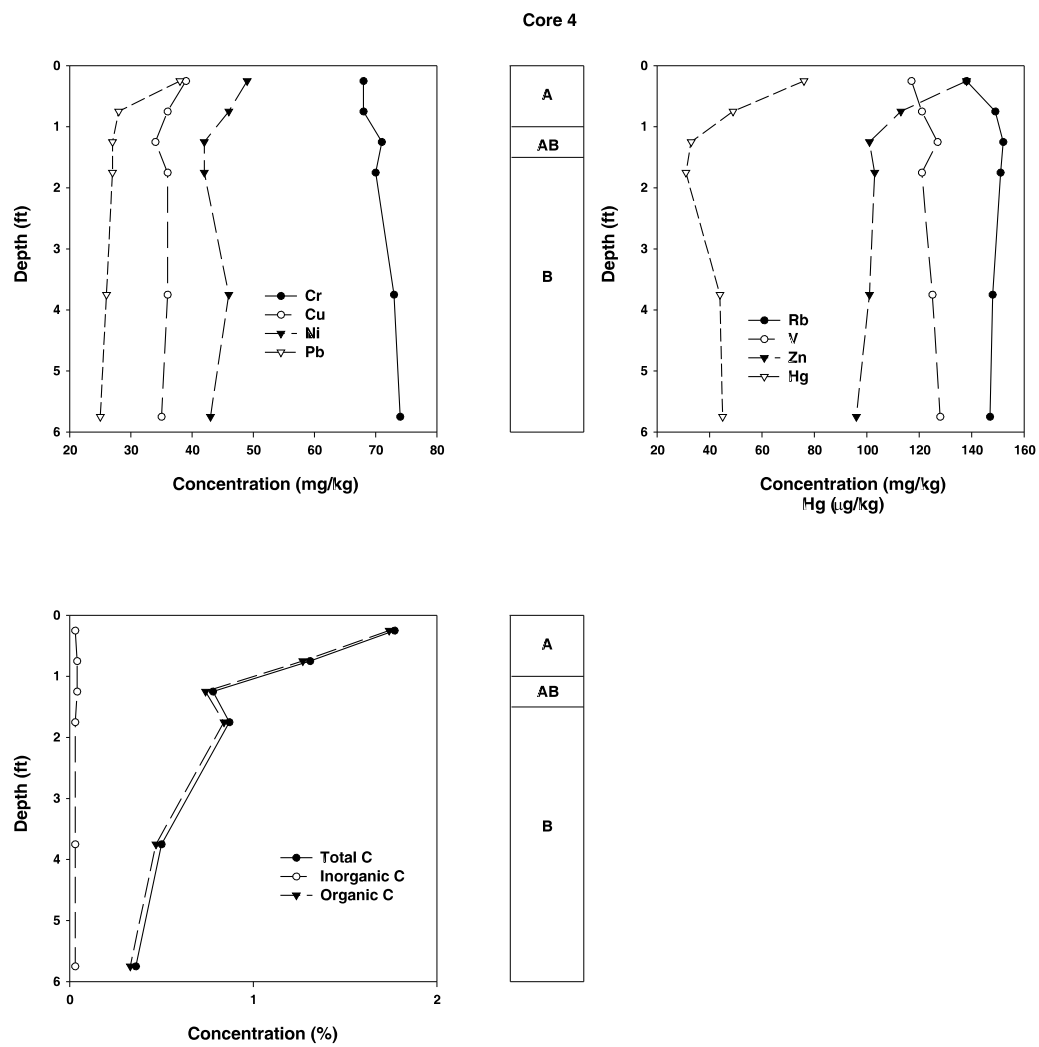


Figure 17. Elemental concentration versus depth in core 4.

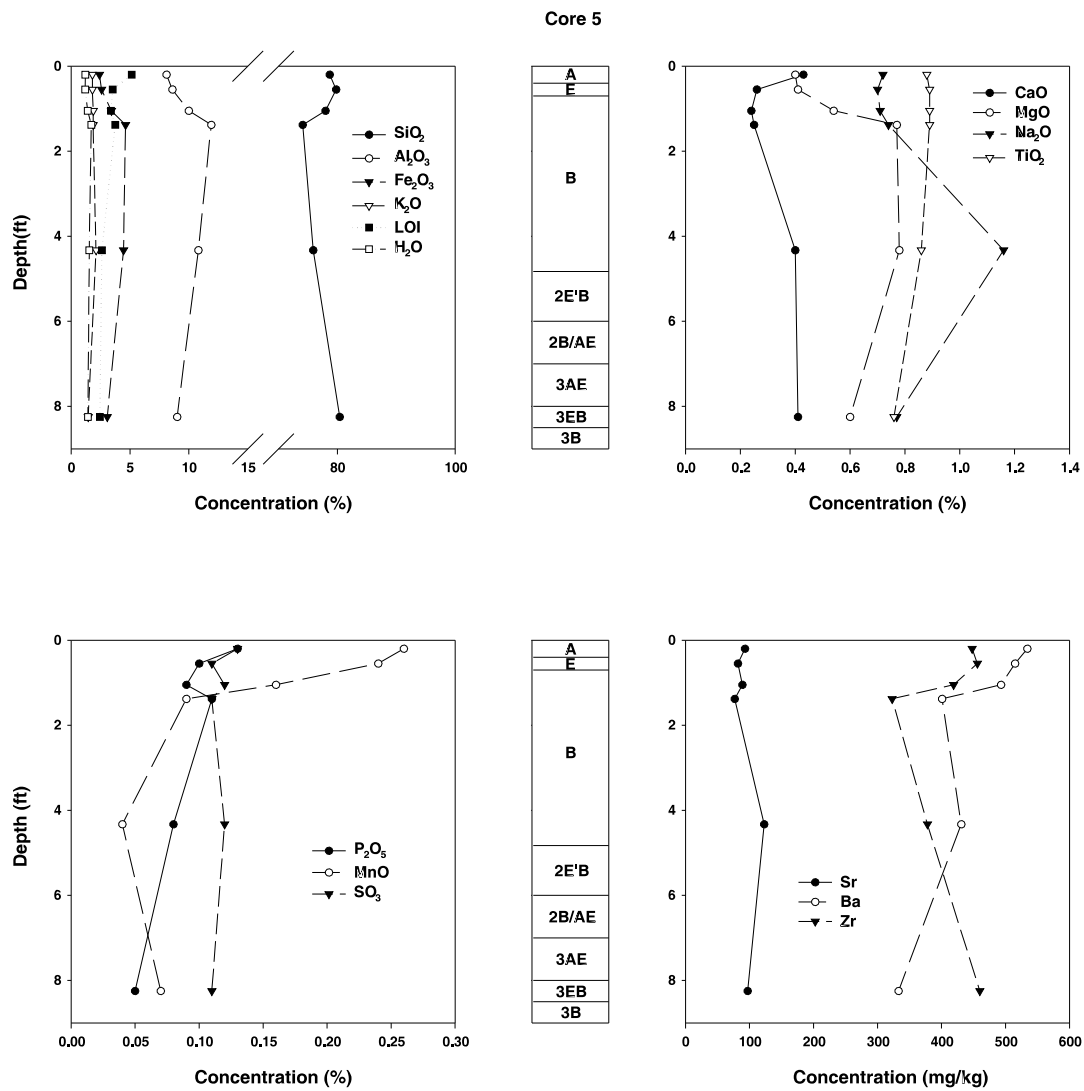


Figure 18. Elemental concentration versus depth in core 5.

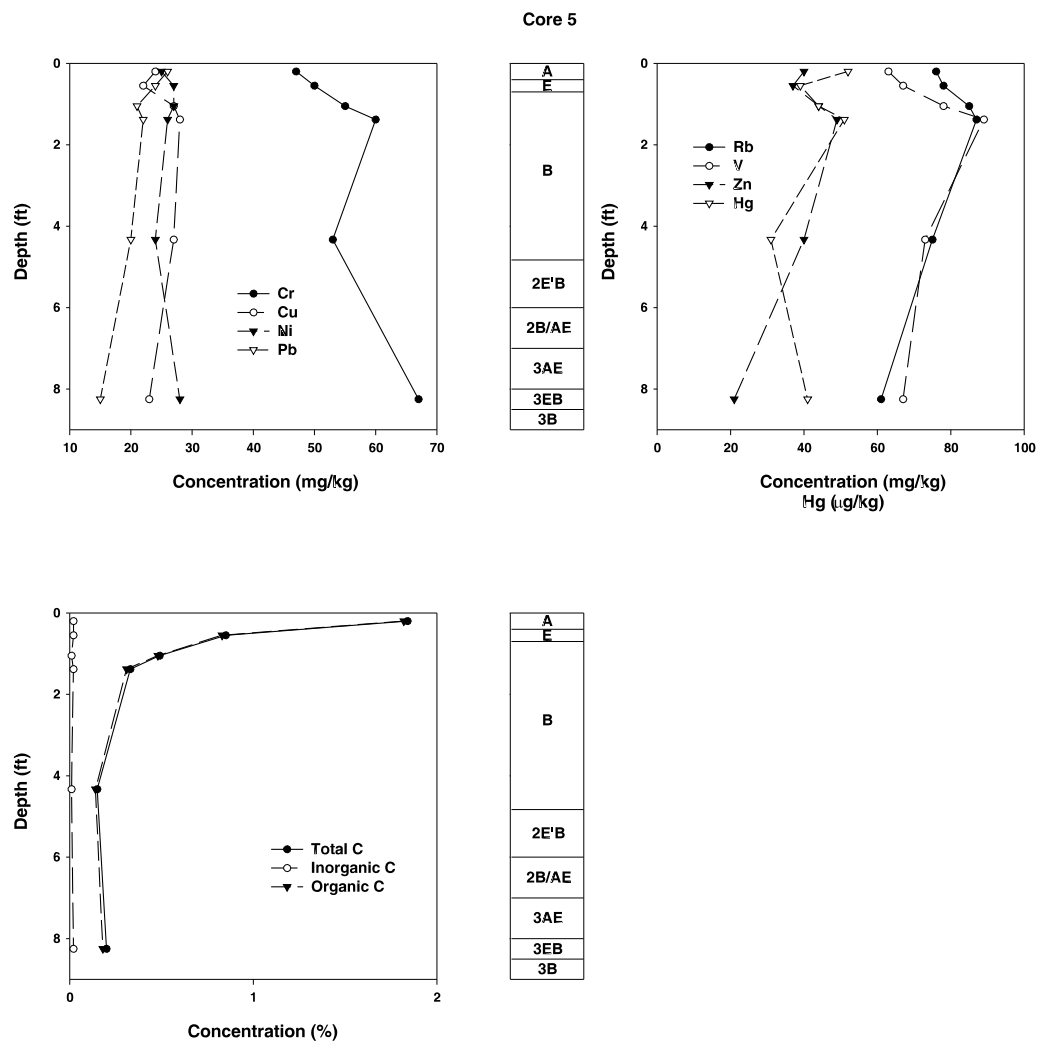


Figure 19. Elemental concentration versus depth in core 5.

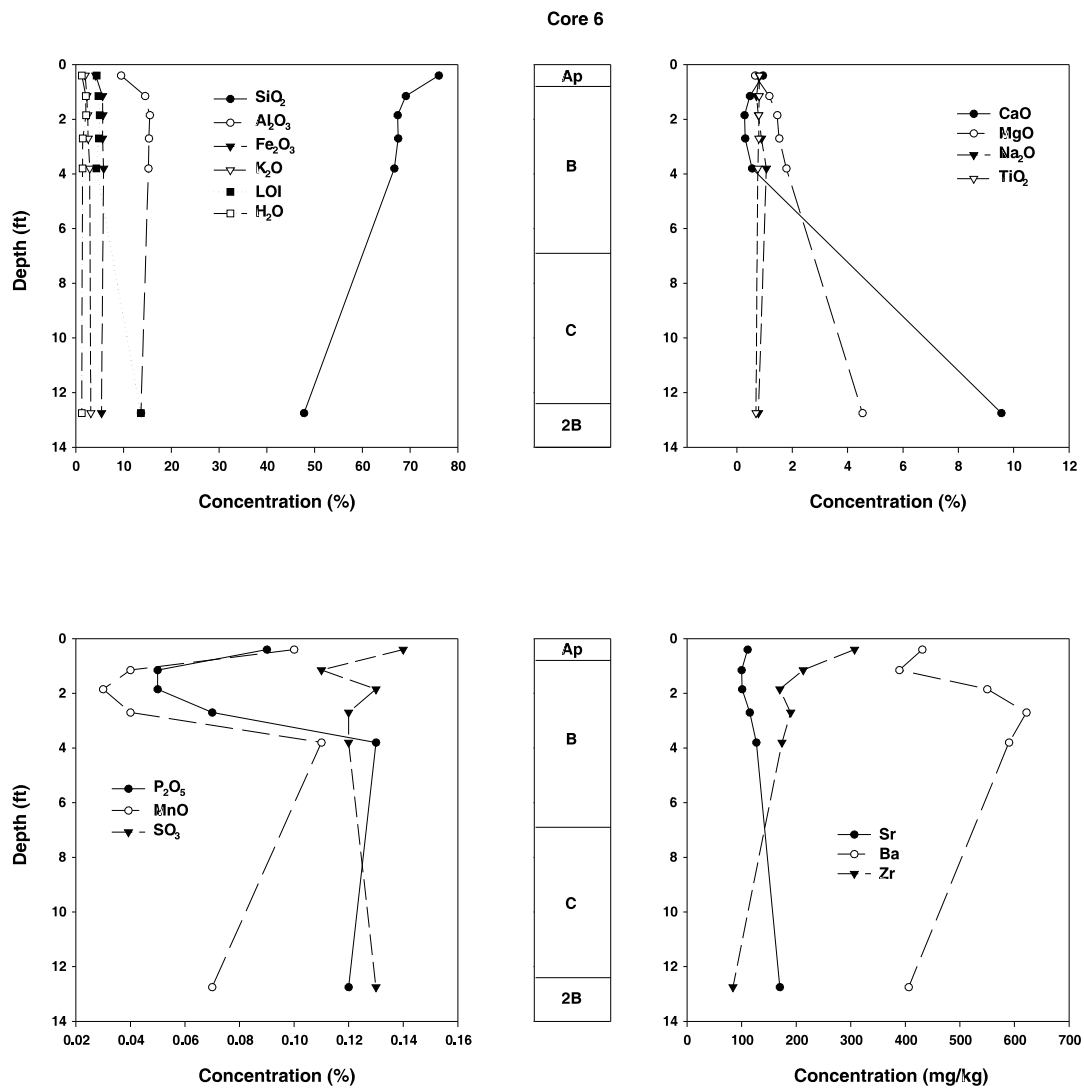


Figure 20. Elemental concentration versus depth in core 6.

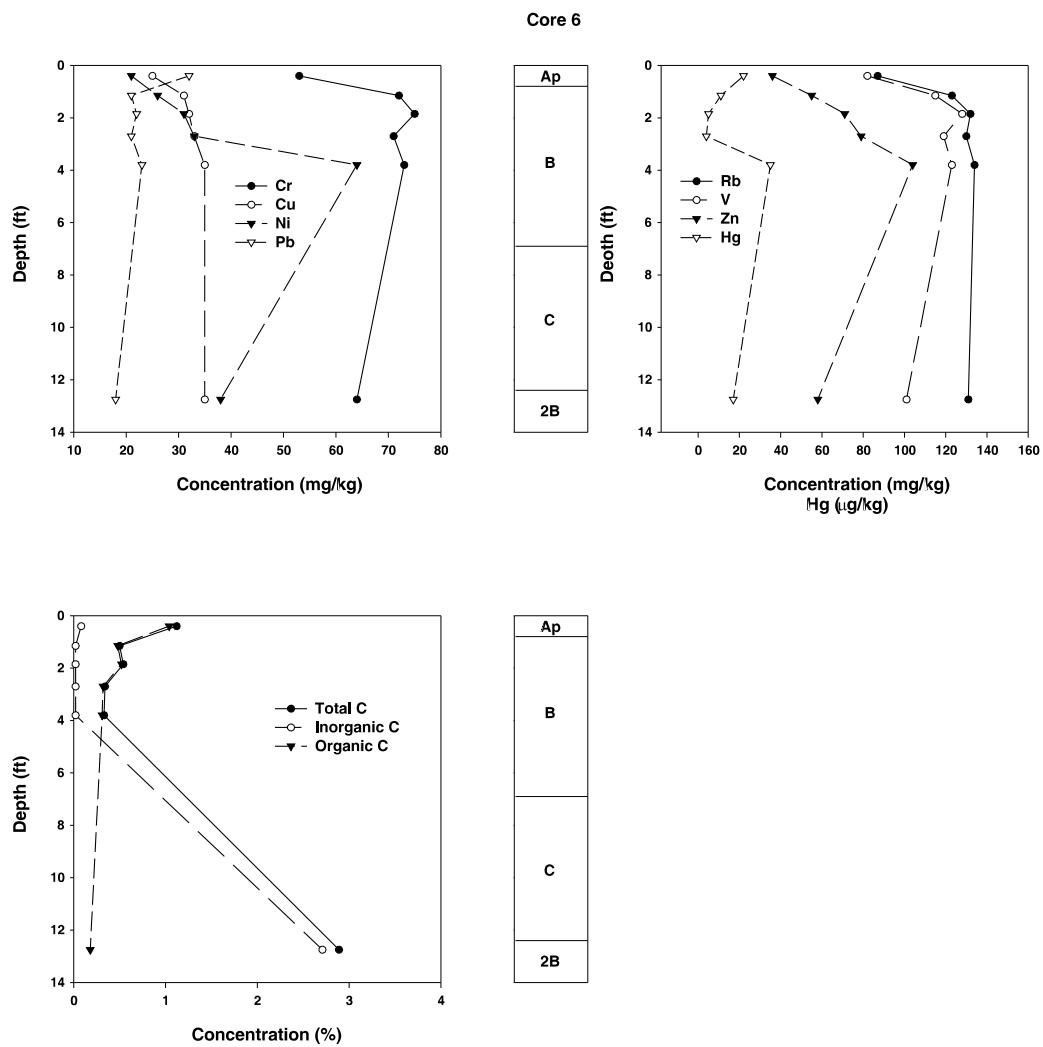


Figure 21. Elemental concentration versus depth in core 6.

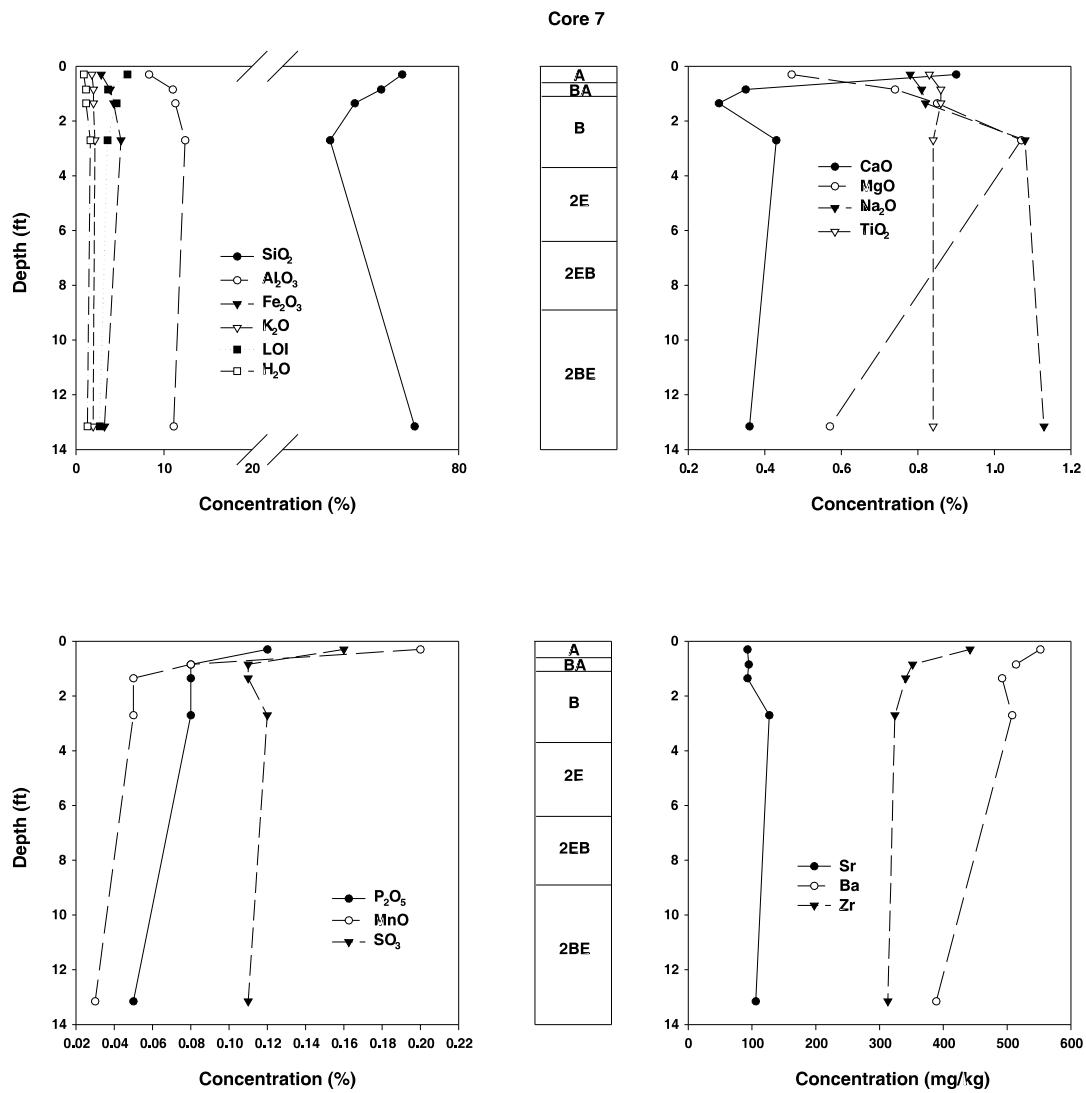


Figure 22. Elemental concentration versus depth in core 7.

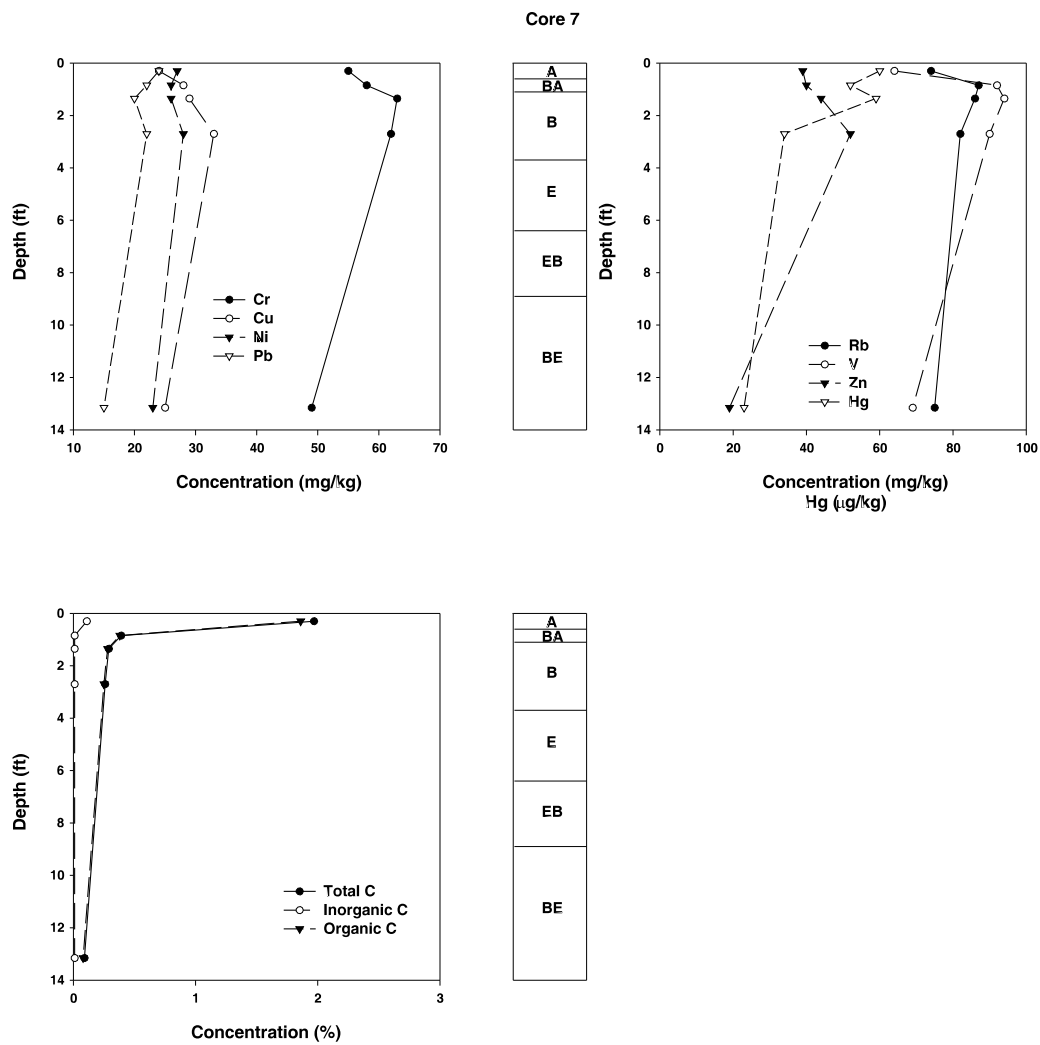
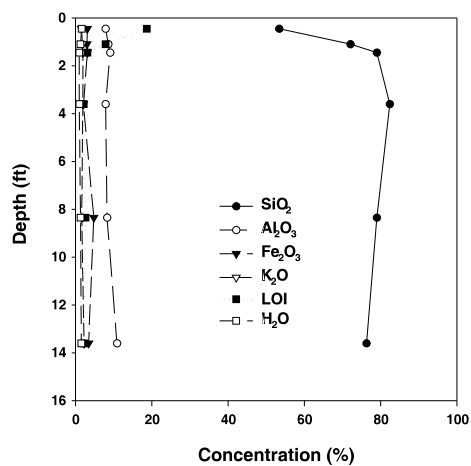


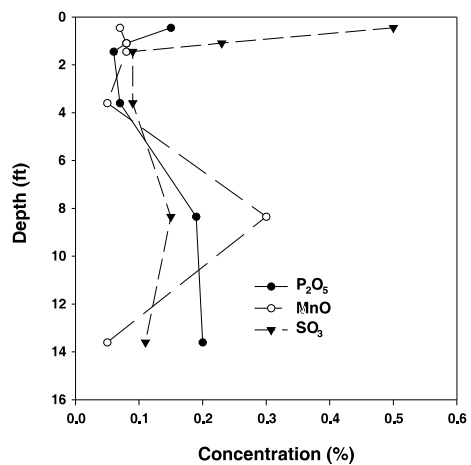
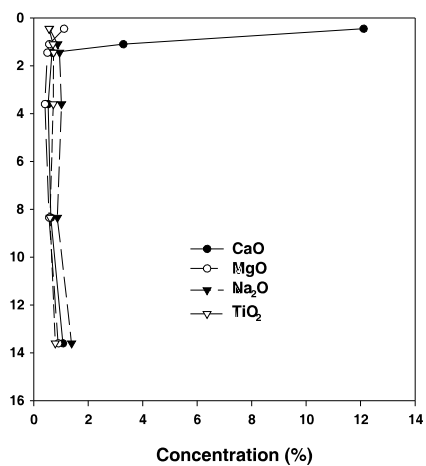
Figure 23. Elemental concentration versus depth in core 7.



Core 8



A
B
C
2E
2B
2C
3A
3C
4B
4C



A
B
C
2E
2B
2C
3A
3C
4B
4C

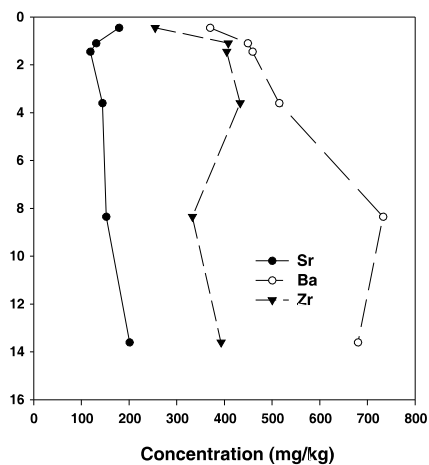


Figure 24. Elemental concentration versus depth in core 8.

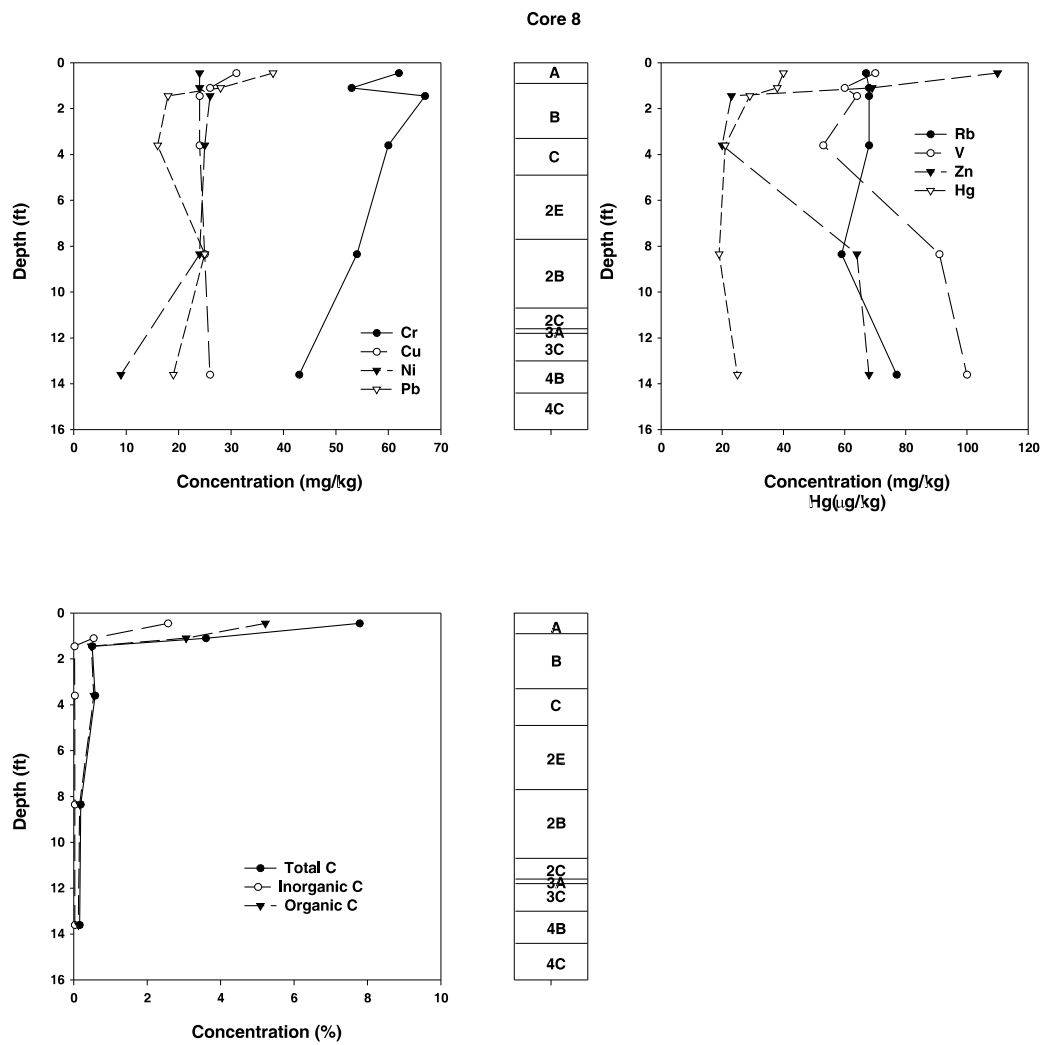


Figure 25.Elemental concentration versus depth in core 8.

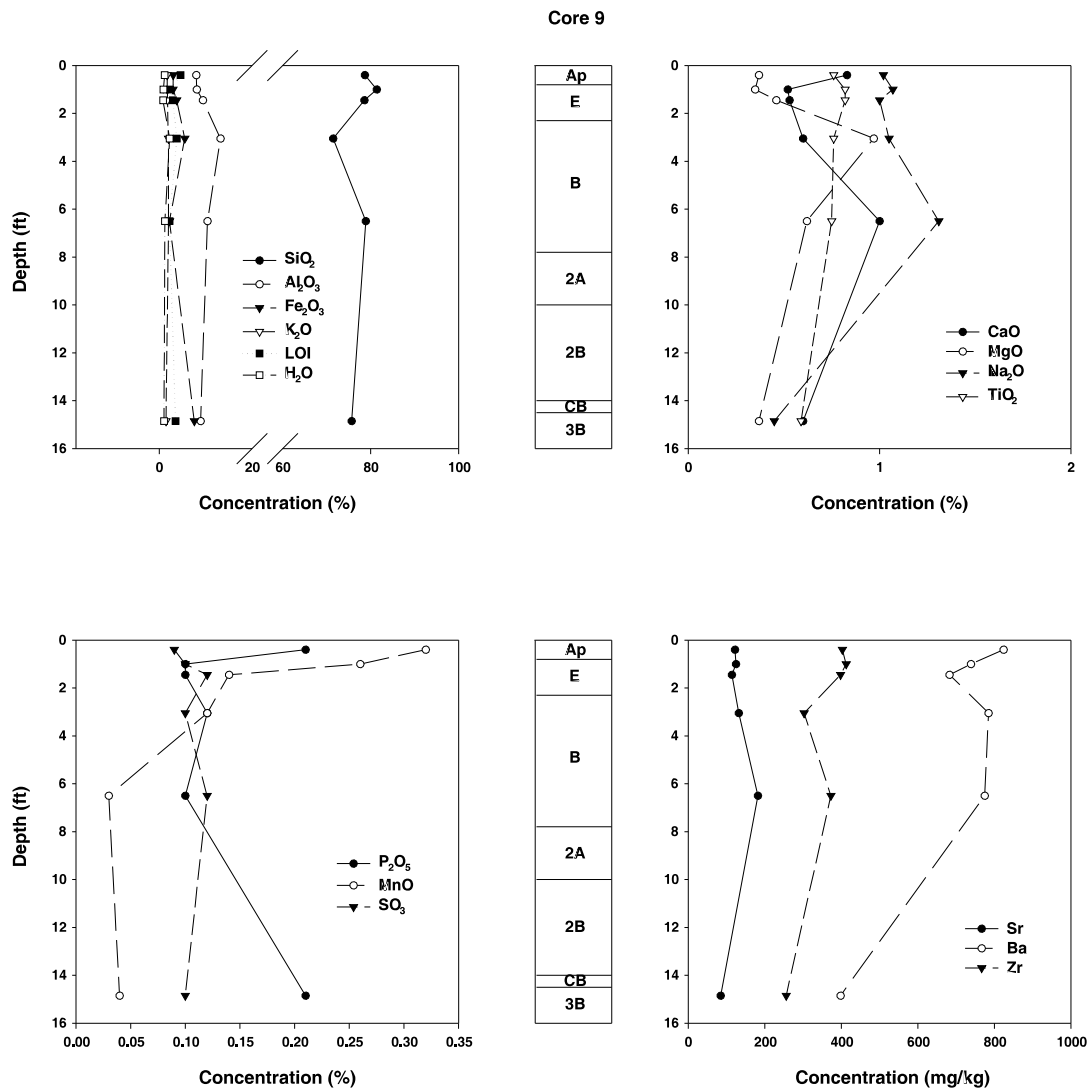


Figure 26. Elemental concentration versus depth in core 9.

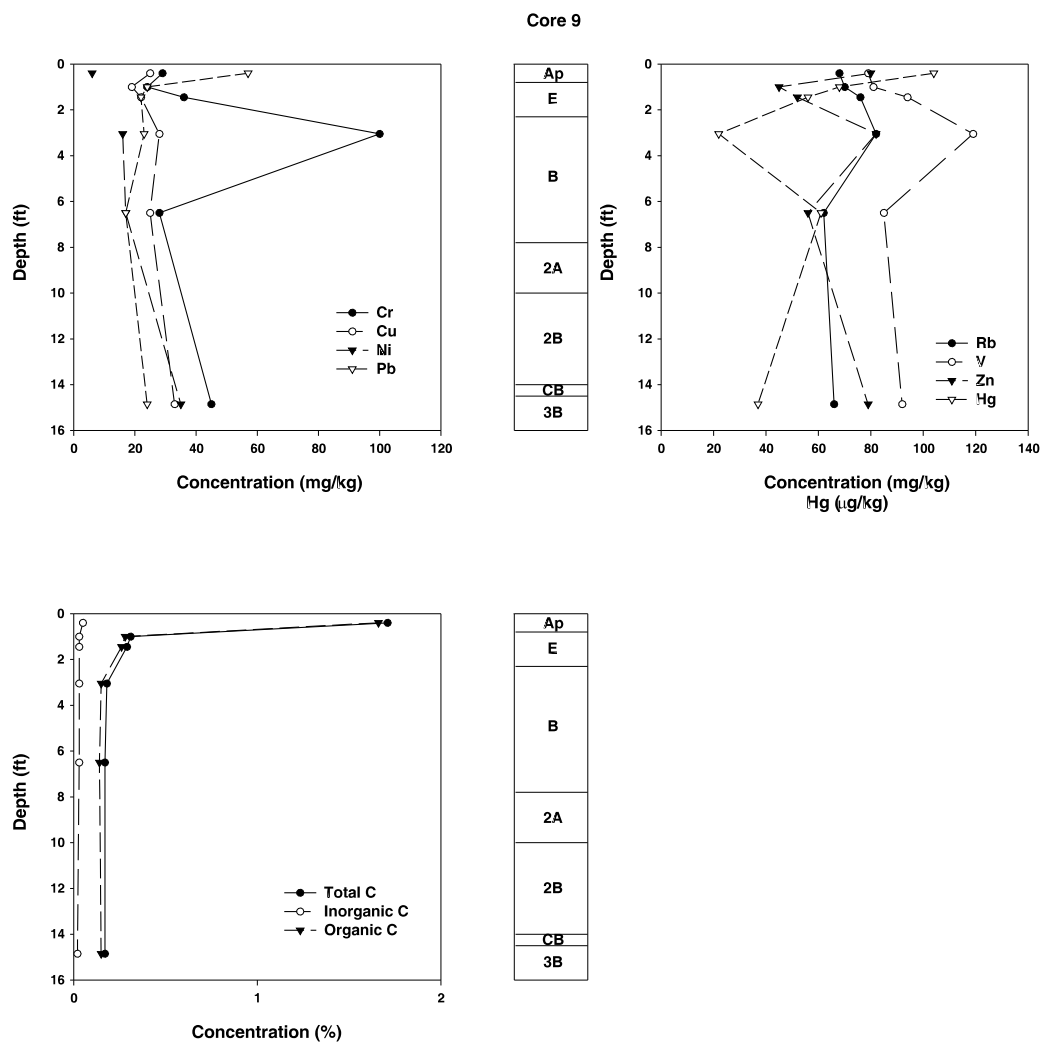


Figure 27. Elemental concentration versus depth in core 9.

Core 10

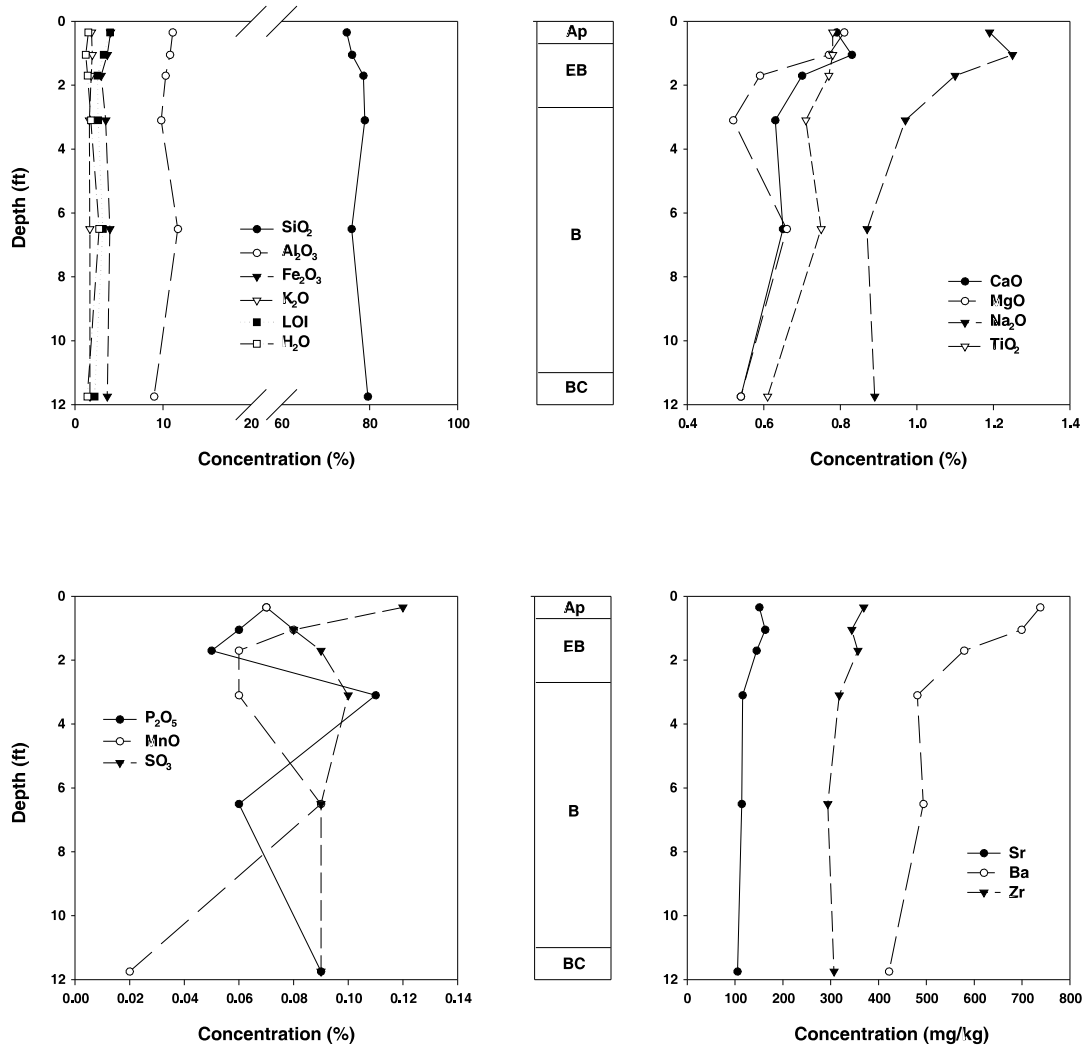


Figure 28. Elemental concentration versus depth in core 10.

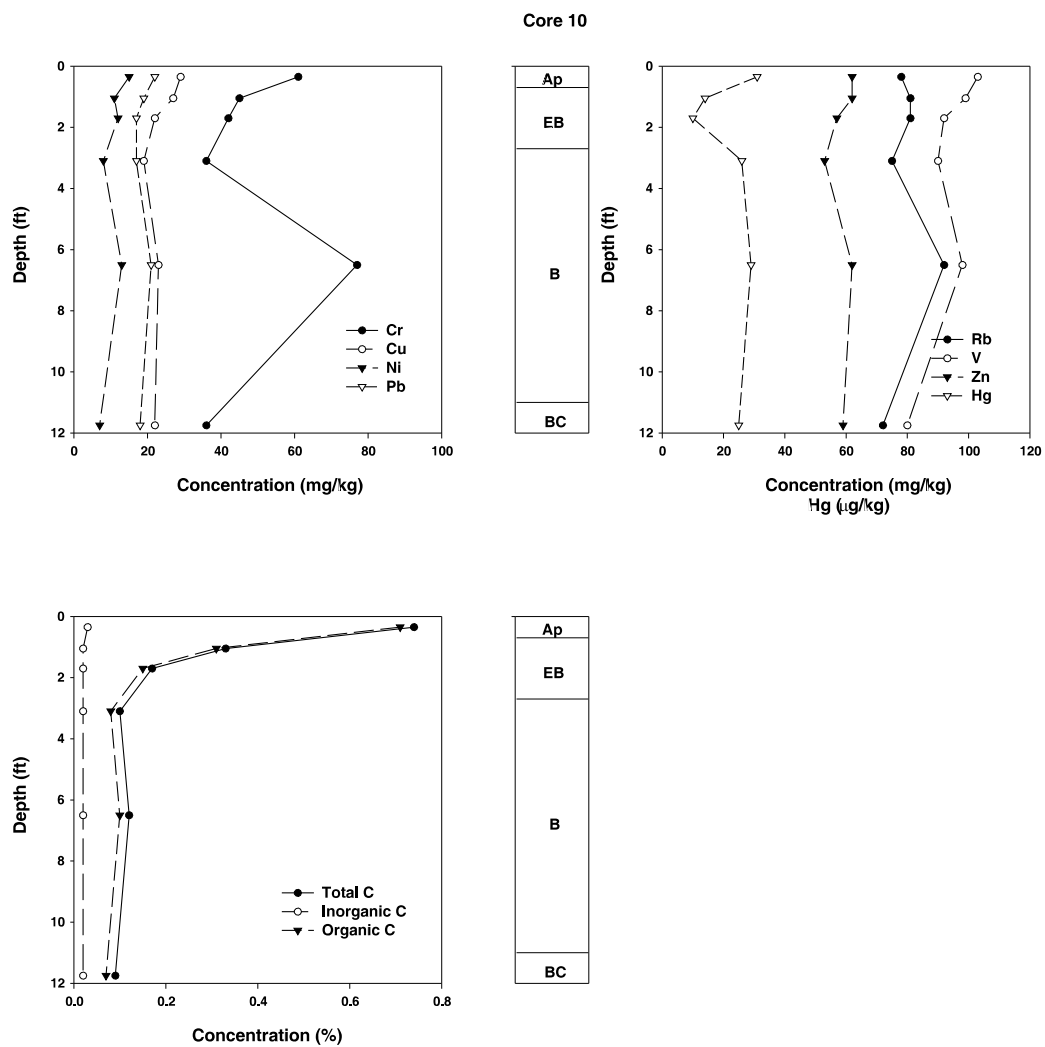


Figure 29. Elemental concentration versus depth in core 10.

