A Progress Report on the Description of Geology and Chemical Composition of Soils in Illinois:
Cores 11 through 26

by

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

February 2003
Open-File Series 2003-1
Illinois State Geological Survey
Champaign, Illinois

PREFACE

This report is a progress report to present analytical chemistry data on soil cores 11 through 26 of an anticipated total of 137 soil cores. There are presently 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 persons from whose land the soil cores were collected. It will also be useful to others who are interested in learning about the chemical composition of soils in 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 expressed through 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. In addition, the ISGS conducts research on 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 gathered in the last 100 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," we 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 interpretation 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 increasing concerns from the general public about the chemical and mineralogical composition of soils in Illinois.

People are 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 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 from becoming confused, one should think of the soil profile as being superposed on the geologic material; there are both a geologic aspect and a soil aspect in the same volume of material.

For this study we are dealing with both the geologic and the soil science points of view. Therefore, we try to merge 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 by the two groups of scientists 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 existing 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 that could be followed. So we chose to build our own comprehensive database in a universal style that tries to avoid the professional biases of soil science and geology. Consequently, many parts of the project are still in a developmental stage. Eventually, we plan to incorporate all available data, but initially we chose to start building a database by selecting representative soils to be sampled and analyzed for their chemical and mineralogical composition. 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 soil cores 11 through 26 of an anticipated 137 coring sites from across the state. These cores were collected in southern Illinois during the fall of 1998. We currently have limited descriptions, identifications of the geologic units and soil horizons, and results from a suite of chemical determinations from a selection of 5 to 6 samples per core. Mineralogical data are not yet available for the samples, but we have speculated in the "Results and Discussion" section of this report on which minerals contain the various elements found in the soils. These speculations were based on available knowledge about the chemical compositions of the minerals that are generally 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 assess the health of the environment (including 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 determine the effects of soil composition (especially trace elements) on the health of 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 (2002) study were collected in northern Illinois, with minimal collection of samples from certain other parts of the state.

The purpose of this present research project is to determine the chemical and mineralogical characteristics of soils and underlying unconsolidated materials throughout 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 US 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 US, 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 valleys in the pre-glacial bedrock surface 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, Illinois, 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 deposits that are commonly over a hundred feet thick. The coarse deposits in former glacial meltwater channels form the excellent aquifers that supply water to many cities in Illinois.

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 covered most of the region. This lake remained for several thousand years and largely filled the lowlands with stratified silt and clay, which is over a hundred 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 deposition of the sediment. This information provides a valuable correlation tool for tracing the distribution of important deposits and finding the best aquifers.

The rise and fall of the meltwater in rivers according to the seasons exposed their very wide flood plains 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 during wind storms. 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 ten 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 vegetative cover. A large amount of loess came from deflation (wind erosion) of 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 reestablished.

In only a few places are wind-blown dune sand deposits significant. The largest of these are in Mason County; in the Green River Lowland of Whiteside, Henry, and Lee Counties; and in 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 byproducts 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 (Na, Ca, Fe, Mg, K, P, S, and many others) and producing byproducts (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 retain 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 and 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 general presence of an excess of limestone in the mixture, causing the youthful soils to be alkaline. After aging for a few thousand years, however, the excess limestone is leached from the upper horizons of the soils, which causes the pH to decrease into the neutral range.

In areas where the glacial sediment 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 incorporated large amounts of shale or fine-grained lake deposits, the resulting soil will be unusually fine-grained ("clayey") and rich in clay minerals.

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 by the water it caused a catastrophic flood that coursed down the river and overflowed the normal flood plain. The rising water transgressed onto the lower parts of the upland, and removed all the loose material down to solid 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). 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 the wind erosion. The loess deposits are thickest along the main river valleys (Figure 2), which provide further evidence that outwash- valleys were the main source areas 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-last glaciation, the Illinoian. 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 from 180,000 to 125,000 years ago. Previous estimates placed the older boundary back to 300,000 years ago, but we believe this age should be rejected because no glacial sediments in this age range 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 such distinguishing characteristics are those caused by soil formation. In geologic terms, the soil at the land surface, which 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 Quaternary deposits map of Illinois (Figure 3) shows the distribution of the major Quaternary units (Hansel and Johnson, 1996) 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 NE 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 loess 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 soils 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 when 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 five feet thick and underlain by the Roxana loess. In these situations the modern soil features (roots, and so forth) commonly extend through the Peoria and 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. The application of mineral fertilizers can maintain peak conditions if sufficient inputs are made, but 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, particularly 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 availablilty 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 are a major factor 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 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 airborne 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. For comparison, in fifteen samples of Illinois loessial surface soils the molybdenum content ranged from 0.75 to 6.40 mg/kg (Kubota, 1977). Prior to the "Clean Air Act," 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 µm diameter) generally causes metals 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 a sufficient amount of the solute will be adsorbed on soil particles to re-establish equilibrium. 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, the sugar dissolves. If the water is then allowed to evaporate, the sugar eventually separates from the solvent (water) as crystals; that is, the sugar 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 otherwise be soluble along with an insoluble precipitate (Fisher, 1961).

Oxidation and reduction refer to the transfer of one or more electrons between ions in solution. By definition, 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 generally is well-drained and aerated, and both iron and manganese precipitate, commonly 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. Several factors govern the reactions. These factors include: (1) temperature, (2) pH, (3) 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, (4) the degree of saturation of the soil by water, and (5) 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 we 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 ten 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 which we selected as sampling sites (see Figure 5).

The 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 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 two 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 was possible in the field. Samples were selected from the cores for chemical analysis on the basis of observable lithologic changes. Samples were dried at 50°C. The dried samples were then disaggregated to approximately <4mm size by passing them through a miniature jaw crusher with ceramic crushing surfaces. 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 in the samples [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), chromium (Cr), copper (Cu), nickel (Ni), lead (Pb), rubidium (Rb), strontium (Sr), vanadium (V), zinc (Zn), and zirconium (Zr)].

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 gram of the ignited sample was mixed with 5.4 g of a mixture of 1:1 lithium tetraborate:lithium metaborate and fused in a 95% platinum-5% gold crucible in the propane flame of a Claisse-bis® Fluxer. The melt was automatically poured into the crucible cover, which also served 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 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^2 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 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 sample by mixing the sample with aqua regia, a mixture of concentrated hydrochloric and nitric acids. The mercury dissolved from both inorganic and organic compounds is oxidized in the aqua regia to the mercurous ion (Hg^+) 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 passed through the reaction solution and 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 kg of soil.

Total Carbon and Inorganic Carbon

Total and inorganic carbon were determined in the less-than 208-µm samples using a Coulometrics Inc.® 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_2) gas. The generated CO_2 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_2.

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® 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 $(NaPO_3)_6$ was added to the bottle followed by 140 mL of deionized water. The blank contained $(NaPO_3)_6$ 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 µm to settle from the top 5 cm of the suspension. The shaken 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 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 μm (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 $(NaPO_3)_6$ 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 (H_2O_2). We also found it necessary to treat the upper two samples from each core with H_2O_2. Approximately 12 grams of <2 mm sample was weighed into a 500-mL beaker. Five mL of deionized water, 5 mL of 30% H_2O_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% H_2O_2 were added to the beakers at 15- to 20-minute intervals until a total of 30 mL of 30% H_2O_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 ten-gram portion of each H_2O_2-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 ±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 18, and depicted graphically in even-numbered Figures 10 through 40. The proportion of sand, silt, and clay in a sample was used to determine the textural class of the sample according to the definitions of 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 sixteen cores discussed in this report, the texture of the uppermost or surficial horizon (the plow layer in most cases) of twelve of the cores was silt loam, three were silty clay loams, and one was a sandy loam. The parent materials of the soils at locations 11, 17, 25, and 26 were alluvium, and were deposited as flood plain sediments. The core from location 12 was collected from the bed of an ancient lake, and is an example of a soil developed in the deposits left in a lake caused by glacial flooding. Core 13 was collected from an area of windblown sand, cores 18 and 19 were from prairie locations, and the remaining cores were collected from areas that were upland forests during soil development.

The clay contents of the subsamples of all the cores ranged from about 5% to 58%. The silt contents ranged from about 3% to 85%, and the sand from 1% to 92%. The soil texture classification, soil type, and developmental environment of the uppermost soil samples are listed in Table 19. In general, the clay content of the samples increased with depth, as shown in even-numbered Figures 10 through 40.

The silt size fraction of a soil is composed principally of quartz (SiO_2), with feldspar and carbonate minerals (where present), plus small amounts of heavy minerals, such as zircon ($ZrSiO_4$). Quartz and zircon are two of the most resistant minerals in soils along with rutile (TiO_2) and ilmenite ($FeTiO_3$) and other iron-bearing minerals. In addition, quartz and other silicate 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 20) of the sand-plus-silt size fraction with silicon. Quartz grains probably reached a size limit (terminal grade) below which they could not be ground by glacial action. The terminal grade for quartz is between about 31 and 62 µm (Dreimanis and Vagners, 1969, 1971). With artificial grinding, the terminal grade for quartz is about 16 to 32 µm (Gaudin, 1926), slightly finer than that observed by Dreimanis and Vagners, but still in the silt size range of 16 to 62 µm.

Results of Chemical Analyses

Table 20 lists the correlation coefficients between the various chemical constituents in the soil samples. The results of the chemical analyses of the subsamples from each of the 16 cores are presented in Tables 21 through 36 and Figures 21 through 52. Major and minor element contents as oxides (silicon through sulfur in the tables) are listed first, followed by trace element contents (barium through zirconium). Major elements are those whose contents are greater than 1%, minor elements are those whose contents are between 0.1 and 1%, and trace elements are present at less than 0.1%, or 1000 mg/kg. The contents of all major and minor elements, as oxides, are listed as weight percent; trace element contents 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 content of one constituent increases from one sample to another, the content of the second constituent increases also. If the correlation coefficient is negative, then as the content of the one constituent increases, the content of the other decreases.

For example, the correlation coefficient between alumina (Al_2O_3) and the clay-size fraction is 0.86 (Table 20). That is, in about 86% 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 particular mineral is present in a sample; they are merely suggestive.

The data from the 16 cores reported here, and from cores 1 through 10 (Dreher et al., 2002) are not sufficient to allow us 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 calculated correlation coefficients to highlight the relationships between constituents based on all samples analyzed from all 16 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 ($SiO_{2,}$ Brady and Weil, 1999) and that zirconium (Zr), which occurs principally in the mineral zircon ($ZrSiO_4$) in soils, also occurs predominantly in the silt-size particles. The correlations between silica and the sand-plus-silt size fraction (0.71) and between zirconium and the silt-size fraction (0.71) confirm these previous 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 SiO_2 is the major constituent of the silt- and sand-size fractions, we observed that as the amounts of the sand+silt fraction increased, the amount of SiO_2 in the samples also increased. That is, the correlation between Si and the sand+silt fraction was strong. As the silt and sand fractions increased, the clay-size fraction decreased and there was a negative correlation (-0.78) 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 (0.86).

The clay-size fraction was correlated with iron (0.85), magnesium (0.54), chromium (0.69), copper (0.78), rubidium (0.73), vanadium (0.91), and zinc (0.78). Therefore, these elements also correlated with each other. Iron and magnesium are commonly structural members, that is, they are necessary parts of certain clay minerals, although these elements also occur as exchangeable ions on clay minerals. Iron and magnesium are commonly found in illite. 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, rubidium, vanadium, and zinc, and others, such as nickel, 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 coefficient between lead (Pb) and phosphorus (as P_2O_5) in cores 11 through 26 is 0.78, which suggests that there is a chemical association between these two elements in soils from the sixteen cores. In cores 1 through 10, however, the correlation coefficient was only 0.43. On an individual core-by-core basis, 16 of the total of 26 cores had positive correlation coefficients of 0.5 or greater. This suggests that there might be a chemical association between Pb and P. Two minerals have been found in soils in trace amounts, pyromorphite [$Pb_5(PO_4)_3Cl$] and plumbogummite [$PbAl_3(PO_4)_2(OH)_5 \cdot H_2O$] (Nriagu, 1974). The correlation coefficient between Pb and Al was low (0.19) for all 26 cores, so that might rule out any consistent occurrence of plumbogummite in the soil samples.

Zinc was correlated in the sixteen cores with both Pb (0.65) and P_2O_5 (0.57). Zinc phosphate, such as zinc pyromorphite [$Zn_3(PO_4)_2 \cdot 4H_2O$] was observed in mine waste soil that had been amended with Na_2HPO_4 (Cotter-Howells and Caporn, 1996).

Soil pH Soil pH is a measure of the acidity or hydrogen ion (H^+) 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):

$$Al^{3+} + H_2O = AlOH^{2+} + H^+$$
$$AlOH^{2+} + H_2O = Al(OH) + H^+$$
$$Al(OH) + H_2O = Al(OH)_{3s} + H^+$$
$$Al(OH)_{3s} + H_2O = Al(OH) + H^+$$

If small amounts of gibbsite, $Al(OH)_3$, or clay minerals in the soil dissolve to contribute Al^{3+} to the soil solution, the Al^{3+} becomes hydrolyzed to $AlOH^{2+}$, which results in the addition of H^+ to the solution, which results in greater acidity. If additional H^+ enters the soil solution, the reaction between $AlOH^{2+}$ and Al^{3+} is driven to the left and H^+ is consumed in the formation of H_2O and Al^{3+}. Aluminum, therefore, serves as an acidity buffer for pH between values of about 4.5 and 5.5 (Sparks, 1995). Above pH 5.5, the solubility of Al^{3+} 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^+ and basic cations of the soil exchange-complex buffer the pH, (2) atmospheric carbon dioxide (CO_2) 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^+ 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):

$CO_{2\ gas} = CO_{2\ aq}$
$CO_{2\ aq} + H_2O = H_2CO_3$
$H_2CO_3 = HCO + H^+$
$HCO = CO^- + H^+$

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 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^-), 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 sixteen cores ranged from 3.86 (very acidic) to 7.78 (moderately alkaline), with a median value of 5.85. Of the 94 samples, 82 had a pH between 4 and 7. The approximate pH range of most soils found globally is 3.0 to 8.5 (Baas Becking et al., 1960). As shown in Figure 8 and Tables 21 through 36, the pH value in twelve of the sixteen cores became more acidic with increasing depth to about 4.0 feet, then became more alkaline with increasing depth. Calcareous till lies under the loess at several of the 16 coring locations. Calcium carbonate in the till imposes an alkaline pH on the deeper samples from these cores.

Means and Ranges of Element Contents The mean and range of the element contents determined in the uppermost samples from the sixteen cores in this portion of the project are compared in Table 37 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 (1984) 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 oxyhydroxides, and soil organic matter. The pH of the soil solution is also very important in determining the solubility and availability of various elements.

Silicon According to Kabata-Pendias (2001), quartz, or SiO_2, is the most resistant common mineral in soils. Likely, a large portion of the quartz in the source rocks for the glacial deposits in Illinois was originally sand-sized material. Glacial transport would have reduced the size of the quartz grains by grinding them to a size that approached the terminal grade in the silt range. Grinding to terminal grade does not appear to have reached completion in the natural setting, but had there been increased transport distance there would have been an increase in grinding and a subsequent increase in the amount of coarse silt (31 to 62 µm) would have been favored (Dreimanis and Vagners, 1971). As already stated, experimental tests by Dreimanis and Vagners (1971) indicated a terminal grade for quartz between 31 and 62 µm (0.031 to 0.062 mm). 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).

The average silica (SiO_2) content of samples from core 12 was 70.43%, the second lowest average SiO_2 content of the sixteen cores. This core was the only one of the sixteen that was collected from ancient lacustrine (lake bottom) sediments. Lacustrine sediments generally contain more clay-size particles than other types of sediments and less sand. The samples from core 16 appeared similar to those from core 12 in terms of average SiO_2, sand, and clay-size contents, but core 16 was collected from an upland location. The other upland soil samples contained from 73 to 78% SiO_2.

Five cores (11, 17, 22, 25, and 26) were collected from alluvial (floodplain) locations. The SiO_2 content of cores 11, 17, 22, and 25 did not vary appreciably with depth. In core 26, the SiO_2 content was relatively constant throughout the four upper samples, at about 83%, but decreased sharply in the bottom two samples to 75% and 60%, respectively.

The SiO_2 content in core 25 also decreased with depth, but not as drastically as in core 26. We note that the bottom two samples from both cores were from ancient buried soils (paleosols) in which the clay content was also higher than in the upper four samples from each of the cores. The SiO_2 content in the upper four samples was also relatively constant.

The relatively constant SiO_2 content with depth is not surprising, because all source materials in a watershed are blended in the formation of alluvium, and alluvial soils generally are not well developed. Periodic flooding provides a rapid addition of fresh sediment to the soil surface. In cores 25 and 26 the buried soil apparently had time to develop prior to being covered by the surficial alluvium, thus the observed differences between the upper four samples and the lower two in these two cores.

Aluminum The aluminum content of soils is mostly 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 profile of Al_2O_3 content versus depth was generally a mirror image of the SiO_2 profile. The aluminum content also was nearly constant in cores from the alluvial soils (cores 11, 17, 22, 25, and 26). The Al_2O_3 in cores 25 and 26 increased in the two bottom samples where the cores penetrated paleosols.

Iron Iron in soils occurs principally as oxyhydroxides, most commonly as goethite (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 oxyhydroxides 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_2O_3 content in these cores tended to pass through a maximum with increase in depth, an indication of downward migration.

The iron content of samples from the upland forest soils (cores 14-16, 20, 21, 23, and 24) typically was higher than in other soil environments, although the lacustrine soil, core 12, had the highest average Fe_2O_3 of all 16 cores.

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.

Calcium Calcite ($CaCO_3$) and dolomite [$CaMg(CO_3)_2$] 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 content of CaO 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 CaO content increased in deeper samples because the calcareous till underlying the soil was penetrated. This is especially evident from the profiles for CaO in cores 12 and 16.

Calcium may be present in soils because of human activities such as the use of limestone or dolomite in road building or for agricultural liming to reduce soil acidity. The latter activity might explain the high CaO content (8.73%) in the surface sample of core 19.

Magnesium Sedimentary minerals such as dolomite are probably the principal sources of magnesium in Illinois soils. The data for the 16 cores generally showed increasing MgO content with depth, and a decrease in MgO in the deepest samples analyzed. As with CaO, the MgO content of the deepest samples from cores 12 and 16 was higher than in samples from shallower in the profile, because of penetration of the underlying till. Core 19 had elevated MgO content in the uppermost sample, also possibly a result of agricultural liming. Once released from its source mineral, Mg compounds are somewhat soluble in water and Mg leached from high in the soil profile to deeper positions. This is evidenced by maxima in the MgO content at depths of 1.5 to 4 feet, as Mg is adsorbed by the soil exchange complex. Cores 22, 25, and 26 are exceptions.

Sodium Sodium-rich feldspar is probably the principal source of sodium in soils (Sparks, 1995). Sodium minerals generally are easily weathered, and once released from the parent mineral, sodium is quite mobile. Sodium forms many water-soluble compounds in soils, and is, therefore, easily leached from the soil column. The trends for the concentration of Na_2O in the sixteen cores were not consistent between cores. In some cores the concentration of Na_2O decreased with depth (core 26), in others it increased (core 13), and in others it passed through a maximum (core 12) or a minimum (core 11). In other cores (15, 21, 22, and 24) the Na_2O content varied irregularly with depth. A small portion of the Na present in the parent materials likely was incorporated onto exchangeable positions on clay minerals, such as smectites, but most Na probably leached from the soil to the groundwater system.

Titanium The sources of titanium in soils are oxides, such as rutile (TiO_2), and ilmenite ($FeTiO_3$) (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). In almost every core the TiO_2 content increased to depths of one to two feet then decreased. Only in cores 22, 25, and 26 was there no increase in the TiO_2 content just below the surface, but a steady increase in content with depth was observed.

Phosphorus The content of phosphorus in soils is low, which makes the identification of phosphorus-bearing minerals difficult. Apatite [$Ca_5(F,Cl,OH)(PO_4)_3$] has been identified in the silt-size fraction of some soils and might be the principal source of phosphorous, 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.

The P_2O_5 content in the cores generally decreased or remained relatively constant with depth. Where the P_2O_5 content of the uppermost sample or two was much higher than the lower samples, there might have been a recent application of P-containing fertilizer.

Manganese The principal source minerals for manganese in soils are amphiboles, pyroxene, biotite, and rhodonite ($MnSiO_3$) (Sparks, 1995). Although manganese occurs in the bulk of the soil as coatings on other minerals (Kabata-Pendias, 2001), it is also commonly found concentrated in nodules (concretions) accompanied by iron. These nodules seem to form in soil horizons that periodically become waterlogged so that reducing conditions prevail, and the soil then dries, restoring 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). However, the lack of correlation between MnO_2 and Fe_2O_3 in the samples from these cores suggests no such intimate association.

In most of the cores the MnO_2 content generally decreased with depth. One possible reason for this is that manganese is sensitive to oxidation and reduction. Oxidized species, such as MnO_2, precipitate where oxygen is readily available, as it normally would be near the surface of the soil, and decrease in content with depth. The solubility of manganese increases as the pH and Eh (the oxidation-reduction potential) decrease (the soil becomes more acidic and less oxidizing) (Lindsay, 1979). Under conditions of increasing acidity and decreasing oxidation potential, manganese would be somewhat easily leached.

In a few cores (12, 14, 15, 16, and 18), however, the MnO_2 content increased immediately below the surface to a depth of 1 to 2 feet, then decreased, and in other cores (12, 14, and 16) the behavior of manganese was opposite to that described above. Possibly in these cores the Mn was present in colloidal-sized particles that migrated downward from the surface before becoming lodged.

Sulfur The probable source of sulfur in these soils was the oxidation of the minerals pyrite and marcasite (they have the same chemical composition, 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 content of SO_3 in the sixteen cores was either below (cores 11 and 12) or near the analytical detection limit. The SO_3 content was uniform both with depth and from core to core. Sulfate forms compounds of low solubility with both barium and strontium.

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 by rainfall that transported sulfur dioxide from the atmosphere to the soil (Hoeft, 1986).

Barium Micas and feldspars are sources of barium in soils. These minerals contain potassium which is commonly replaced by barium, because of the similar size of the two atoms (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 ($BaSO_4$, Allen and Hajek, 1989) and hollandite [$Ba(Mn^{4+},Mn^{2+})_8O_{16}$, McKenzie, 1989]. Barium also 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. As mentioned above, barium also reacts with sulfate to form the sparingly soluble barium sulfate.

The barium content ranged from 253 to 1030 mg/kg. There were no consistent trends in the concentration of Ba with depth in the profiles of the sixteen cores.

Chromium Chromium is generally present in soils as Cr^{3+} and this is responsible for the element's relative insolubility and immobility in soils (Kabata-Pendias, 2001), because the Cr^{3+} 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 ($FeCr_2O_4$) in the parent material (Kabata-Pendias, 2001), and industrial fallout, such as dust and industrial contaminants.

Chromium contents were below the detection limit of 5 mg/kg in some of the samples of cores 13, 16, 18, 19, 22, and 26. In all but four of the cores the Cr content was at its maximum in the B horizon, commonly at the top of the horizon. The maximum Cr content was 169 mg/kg. Connor et al. (1957) noted similar behavior in podzols developed on glacial drift in New Jersey and Pennsylvania. Although Cr is relatively immobile in soils, because it forms strong bonds with clay minerals, it still can 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 was adsorbed by colloidal-sized clay minerals in the A horizon, the Cr would be transported downward with the colloidal particles. This hypothesis is supported by the previously mentioned correlation between Cr and clay-sized particles.

Copper Copper readily forms complex compounds with organic molecules, especially of the porphyrin type, but it also is adsorbed readily 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). Copper is rendered immobile as a result of any of these reactions and its concentration in the soil profile does not vary (Kabata-Pendias, 2001).

The depth profiles for copper were similar to those for chromium, but the copper contents were lower and their ranges were smaller. The Cu content ranged from <5 to 48 mg/kg.

Mercury The most common natural source of mercury in rocks is the mineral cinnabar, HgS, but this mineral is seldom found in detrital material, such as soils and sediments (Kabata-Pendias, 2001). Most mercury in soil is inherited from the parent materials. Mercury is also deposited on the soil surface as a result of human activities, such as combustion of coal or manufacturing processes.

Competing mechanisms of leaching, sorption, and volatilization determine the fate of mercury in soils. Organic matter (humic material) in soils has a greater capacity to adsorb mercury than the inorganic soil components (Yin et al., 1997), except that sulfide (S^{2-}) and sulhydryl groups (SH^-) in soils have high affinities for mercury, and may form mercuric sulfide (Barnett et al., 1997). Once sorbed, mercury may be desorbed slowly by soil solutions that contains little or no mercury (Yin et al., 1997). If mercuric sulfide is formed, this would tend to fix the mercury in position; however, if the mercury does not react with sulfur, it may be leached slowly downward in the soil profile. In contrast to these mechanisms of mercury retention, elemental mercury and methylated mercury compounds are easily volatilized. Methylated mercury has been shown by Rogers (1976, 1977) to be produced abiotically by humic substances in soils.

A clay-rich soil produced the most methylmercury, followed by a loam, followed by a sandy soil (Rogers 1976, 1977). The organic matter content of the soils followed the same order: clay>loam>sand.

In a few cores the mercury followed the pattern of the zinc profile (cores 14, 15, 17, 18, and 20). In several cores (11, 14-16, 18, 19, 21, 23, 24, and 26) the mercury content reached a maximum value (27 to 78 µg/kg) in the B horizon. This suggests downward migration of mercury and sorption by clay minerals or iron and manganese oxyhydroxides. The mercury content in core 12 was nearly invariant through the profile. In cores 20, 22, and 25, the mercury content of the core was highest at the surface, which suggests that not only might there have been an external source of mercury, such as industrial fallout or the application of a mercury-containing chemical, but also that the mercury was bound by soil organic matter with little downward leaching. The range of mercury contents was <2 to 92 µg/kg.

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 typically is 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).

It was not unusual for the nickel content in the sixteen cores to be below the analytical detection limit of 5 mg/kg in the upper samples of a core and to increase to detectable levels in samples from lower in the core. The maximum Ni content was 43 mg/kg. In a few cores (11, 15, and 16) the nickel content followed that of chromium. In several cores the highest Ni content was observed in the B horizon. Nickel is relatively stable in the soil solution. The concentration profiles for some cores indicate that Ni was probably leached downward during weathering and accumulated in the lower portions of the profile, following the clay minerals and iron oxyhydroxides.

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

In about half the cores the trend of the Pb content, which ranged from 11 to 72 mg/kg, followed that of copper. In many cores, the lead content of the soil was greatest at the surface and decreased with depth to the B horizon, where the content increased, then again decreased with depth. This behavior is indicative of input to the surface samples, probably by products from the combustion of leaded gasoline in vehicles and machinery, and from coal combustion. It appears that lead was leached from the surface and accumulated in the B horizon by adsorption on clay minerals and iron and manganese oxyhydroxides. Organic matter is also known to strongly adsorb lead, as might be reflected in those surface samples that contained elevated lead contents.

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 is 75 mg/kg and the range is 45 to 100 mg/kg. The Rb content in the sixteen cores fell within these ranges.

For most cores, the Rb content was greatest in the upper part of the B horizon. Cores 16, 17, 22, and 25 were exceptions to this observation. In core 26, the Rb content increased with depth through the core. The minimum Rb content was 37 mg/kg and the maximum was 191 mg/kg.

Strontium Strontium generally is associated with soil organic matter, but may also precipitate, under alkaline conditions, as strontianite ($SrCO_3$), and 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).

The Sr contents of these cores did not vary appreciably, falling in the range of 69 to 255 mg/kg. In general, the Sr content was in the range of 100 to 200 mg/kg. Even though the statistical correlation of Sr with Ca in these cores was very low, the shapes of the Ca depth trends were reflected in the Sr depth trends. The Sr trends, however, lacked the degree of change from sample to sample exhibited by the Ca trends.

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 (Butler, 1953, 1954). Vanadium also may form complex compounds with soil organic matter, for example in porphyrin-type compounds (Kabata-Pendias, 2001). In addition to the rocks in the parent materials, V can be contributed to soils by industrial processes such as smelting or combustion of coal or oil.

The V content in the soil (range from 37 to 172 mg/kg) generally was greatest in the upper B horizon. The trend of V content with depth paralleled that of Zn and/or Rb in all but four (16, 17, 22, and 25) of the sixteen cores.

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.

In these 16 cores the zinc content (range from 33 to 139 mg/kg) was generally greatest in the B horizon. In two cores (25 and 26), however, the Zn content increased through the depth of the cores.

Zirconium Zirconium generally occurs in soil as the very stable mineral zircon ($ZrSiO_4$). Zircon is very resistant to weathering and zirconium is only very slightly mobile in soils. The Zr content of soils is generally inherited from the parent materials (Kabata-Pendias, 2001).

The minimum Zr content was 98 mg/kg and the maximum was 489 mg/kg. The Zr content in 13 of the 16 cores was greater in the upper one or two samples than in those from deeper positions in the soil profiles, which suggests a greater degree of weathering of minerals other than zircon at the surface of the soil than lower in the profile. The content of Zr in the surface samples would also be elevated if Zr-containing mineral particles were carried to and deposited on the soil surface by wind.

Carbon Most carbon in the samples was combined in organic residues from biological material. Plant residues are typically plowed into the upper portions of the soil column after harvest each year. Burrowing animals leave waste behind, small insects, worms, and microorganisms die in the soil. The remains are incorporated into the soil column. In each core except cores 12 and 16 the highest total carbon content occurred inn the uppermost sample and it was almost entirely organic carbon. The range of total carbon content was 0.05 to 3.38 %, inorganic carbon content ranged from 0.01 to 2.31 %, and organic carbon ranged from 0.01 to 2.52 %. The total carbon content in cores 12 and 16 followed the same pattern as in most of the other cores except that the highest total carbon content occurred in the lowermost sample and was mostly inorganic, or carbonate carbon. A secondary maximum was observed in the total C content of core 15 in the fourth sample from the top. It is noted that the clay-size content and the content of trace elements associated with clay minerals also reached maxima in the same sample. The inorganic carbon in most of the cores was low and relatively invariant with depth, however, in a few cores (19, 20, 23, and 25) there were noticeable variations in carbon content. Limestone might have been applied to the field prior to collecting core 19, which resulted in relatively high inorganic carbon and calcium contents in the uppermost sample.

CONCLUSIONS

The contents of several trace elements, including chromium, copper, rubidium, vanadium, and zinc were strongly correlated with the content of clay-sized particles. Each of these metals is easily sorbed by various clay minerals. As the colloidal-sized clay minerals migrate downward through the soil column, any elements attached to the clay particles are also transported.

The sand-plus-silt content is an indicator of the amount of silica in a sample. There is a similar relationship between clay and alumina content.

Indications of differences in chemical composition between modern and ancient buried soils were seen in cores 25 and 26, in which the silica content was higher and the alumina content lower in the samples from the modern soil than in those from the buried soil.

As more cores are analyzed and more data become available, we expect that some regional trends in soil composition may emerge. Presently, the data are too few to make such inferences reliably. With more data we may be able to distinguish between various mechanisms for the addition and removal of material. For example, we noted the possible input of Pb content to the surface from particulate matter carried and deposited by wind versus the increase in content due to weathering.

REFERENCES

Allen, B. L. and B. F. Hajek, 1989, Mineral occurrences in soil environments, *in* Dixon, J. B. and S. B. Weed, eds., Minerals in Soil Environments, Second Edition, SSSA Book Series: 1, Soil Science Society of America, Madison WI, 1244 p.

Baas Becking, L. G. N., I. R. Kaplan, and D. Moore, 1960, Limits of the natural environment in terms of pH and oxidation-reduction potentials, J. Geol., v. 68, p. 243-284.

Barnett, M. O., L. A. Harris, R. R. Turner, R. J. Stevenson, T. J. Henson, R. C. Melton, and D. P. Hoffman, 1997, Formation of mercuric sulfide in soil, Environ, Sci, Technol., v.31, p. 3037-3043.

Birkeland, P. W., 1999, Soils and Geomorphology, Third Edition, Oxford University Press, New York, 430 p.

Brady, N. C. and R. R. Weil, 1999, The nature and properties of soil, Twelfth Edition, Prentice Hall, Upper Saddle River, NJ, 881 p.

Butler, J. R., 1953, The geochemistry and mineralogy of rock weathering: (1) The Lizard area, Cornwall, Geochim. Cosmochim. Acta, v. 4, p. 157-178.

Butler, J. R., 1954, The geochemistry and mineralogy of rock weathering: (2) The Normark area, Oslo, Geochim. Cosmochim. Acta, v. 6, p. 268-281.

Connor, J., N. F. Shimp, J. C. F. Tedrow, 1957, A spectrographic study of the distribution of trace elements in some podzolic soils, Soil Science, v. 83, no. 1, p. 65-73.

Cotter-Howells, J. and S. Caporn, 1996, Remediation of contaminated land by formation of heavy metal phosphates, Appl. Geochem. v. 11, p.335-342.

Darnley, A. G., A. Bjöklund, B. Bølviken, N. Gustavsson, P. V. Koval, J. A. Plant, A. Steenfelt, M. Taichid, X. Xuejing, R. G. Garrett, and G. E. M. Hall, 1995, A Global Geochemical Database for Environmental and Resource Management, Recommendations for the International Geochemical Mapping Final Report of IGCP Project 259, UNESCO Publishing, Paris, France, x + 122 p.

Dreher, G. B., L. R. Follmer, and Y. Zhang, 2002, A progress report on the chemical composition of soils in Illnois: Cores 1 through 10, Illinois State Geological Survey Open-File Series 2002-2, 83 p.

Dreimanis, A. and U. J. Vagners, 1969, Lithologic relation of till to bedrock, *in* Wright, H. E., ed., Quaternary Geology and Climate, National Academy of Science Publication 1701, Washington, DC, p. 93-8.

Dreimanis, A. and U. J. Vagners, 1971, Bimodal distribution of rock and mineral fragments in basal till, *in* Goldthwait R. P., ed., Till: A symposium, Ohio State University Press, p. 237-250.

Esser, K. B., J. G. Buckheim, and P. K. Helmke, 1991, Trace element distribution in soils formed in the Indiana Dunes, U.S.A., Soil Sci., v. 150, p. 340-350.

Fehrenbacher, J. B., J. D. Alexander, I. J. Jansen, R. G. Darmody, R. A. Pope, M. A. Flock, E. E. Voss, J. W. Scott, W. F. Andrews, and L. J. Bushue, 1984, Soils of Illinois, Bulletin 778, University of Illinois at Urbana-Champaign College of Agriculture, 85 p.

Fisher, R. B., 1961, A Basic Course in the Theory and Practice of Quantitative Chemical Analysis, Second Edition, W. B. Saunders Company, Philadelphia, 501 p.

Follmer, L. R., 1982, The geomorphology of the Sangamon surface: its spatial and temporal attributes, *in* Thorn, C. E. ed., Space and Time in Geomorphology, Allen and Unwin, p. 117-146.

Follmer, L. R., 1996, Loess studies in central United States: evolution of concepts, Engineering Geology, v. 45, p. 287-304.

Fritz, J. S. and G. H. Schenk, Jr., 1966, Quantitative Analytical Chemistry, Allyn and Bacon, Inc., Boston, 516 p.

Gaudin, A. M., 1926, An investigation of crushing phenomena, Trans. Amer. Inst. Min. Metall. Petrol. Engin., v. 73, p. 253-316.

Goldschmidt, V. M., 1954, Geochemistry, Oxford University Press, London, G. B., 730 p.

Hansel, A. K. and W. H. Johnson, 1996, Lithostratigraphic reclassification of deposits of the Wisconsin Episode, Lake Michigan Lobe area, Illinois State Geological Survey Bulletin 104, 116 p.

Hassett, J. J., 1989, Soil chemistry: Equilibrium concepts, Course notes, Soils 307, University of Illinois at Urbana-Champaign.

Hassett, J. J. and W. L. Banwart, 1992, Soils and Their Environment, Prentice-Hall, New Jersey, 424 p.

Helling, C. S., G. Chesters, and R. B. Corey, 1964, Contribution of organic matter and clay to soil cation-exchange capacity as affected by the pH of the saturating solution, Soil Sci. Soc. Am. Proc. v. 28, p. 517-520.

Hildebrand, E. E. and W. E. Blume, 1974, Lead fixation by clay minerals, Naturwissenschaften, vol. 61, p.169 ff, as reported in Kabata-Pendias, A. and H. Pendias, 1992, Trace Elements in Soils and Plants, Second Edition, CRC Press, Boca Raton, FL, 365 p.

Hoeft, R. G., 1986, Plant response to sulfur in the Midwest and Northeastern United States, Agronomy, vol. 27, p. 345-356.

Hornick, S. B., D. E. Baker, and S. B. Guss, 1976, Crop production and animal health problems associated with high soil molybdenum, *in* Chappell, W. R. and K. K. Petersen, eds., Molybdenum in the Environment, Volume 2, The Geochemistry, Cycling, and Industrial Uses of Molybdenum, Marcel Dekker, Inc, New York, p. 665-684.

Imbrie, J., and K. P. Imbrie, 1979, Ice Ages: Solving the Mystery, Enslow Publishers, Short Hills, NJ, 224 p.

Indorante, S. J., L. R. Follmer, R. D. Hammer, and P. G. Koenig, 1990, Particle-size analysis by a modified pipette procedure, Soil Sci. Soc. Am. J., v. 54, p.560-563.

Jackson, M. L. and G. D. Sherman, 1953, Chemical weathering of minerals in soils, Adv. Agron., v. 5, p. 219-318.

Jenny, H., 1941, Factors of Soil Formation: A System of Quantitative Pedology, McGraw-Hill Book Company, New York, 281 p.

Jones, R. L., 1986, Barium in Illinois surface soils, Soil Sci. Soc. Am. J., v. 50, p. 1085-1087.

Kabata-Pendias, A., 1980, Heavy metal sorption by clay minerals and oxides of iron and manganese, Mineral. Pol., vol. 11, p. 3 ff, as reported in Kabata-Pendias, A. and H. Pendias, 1992, Trace Elements in Soils and Plants, Second Edition, CRC Press, Boca Raton, FL, 365 p.

Kabata-Pendias, A., 2001, Trace Elements in Soils and Plants, Third Edition, CRC Press, Boca Raton, 413 p.

Killey, M. M., 1998, Illinois' Ice Age Legacy, Illinois State Geological Survey GeoScience Education Series 14, 66 p.

Kubota, J., 1977, Molybdenum status of United States soils and plants, *in* Chappell, W. R. and K. K. Petersen, eds., Molybdenum in the Environment, v. 2, Marcel Dekker, New York, p. 555-581.

Lindsay, W. L., 1972, Zinc in soils and plant nutrition, Advances in Agronomy, v. 24, p. 147-186.

Lindsay, W. L., 1979, Chemical Equilibria in Soils, John Wiley and Sons, New York, 449 p.

Lindsay, W. L., P. L. G. Vlek, and S. H. Chien, 1989, Phosphate minerals *in* Dixon, J. B. and S. B. Weed, eds., Minerals in Soil Environments, Second Edition, SSSA Book Series: 1, Soil Science Society of America, Madison WI, 1244 p.

Lineback, J. A., 1981, Quaternary Deposits of Illinois (Map), Illinois State Geological Survey

Lineback, J. A., L. R. Follmer, H. B. Willman, E. D. McKay, J. E. King, F. B. King, and N. G. Miller, 1979, Wisconsinan, Sangamonian, and Illinoian stratigraphy in central Illinois, Illinois State Geological Survey Guidebook 13, 139 p.

Luxmoore, R. J., ed. 1994, Factors of soil formation: A fiftieth anniversary retrospective, SSSA Special Publication Number 33, Soil Science Society of America, Inc, Madison, WI, 160 p.

Mattigod, S. V. and A. L. Page, 1983, Assessment of metal pollution in soils, *in* Thornton, I., Applied Environmental Geochemistry, Academic Press, New York, p. 355-394.

McKeague, J. A. and R. J. St. Arnaud, 1969, Pedotranslocation: Eluviation-illuviation in soils during the Quaternary, Soil Science, v. 107, p. 428-434.

McKenzie, 1989, Manganese oxides and hydroxides, *in* Dixon, J. B. and S. B. Weed, eds., Minerals in Soil Environments, Second Edition, SSSA Book Series: 1, Soil Science Society of America, Madison WI, 1244 p.

Nriagu, J. O., 1974, Lead orthophosphates–IV: Formation and stability in the environment, Geochim. Cosmochim. Acta, v. 38, p. 887-898.

Paton, T. R., G. S. Humphreys, and P. B. Mitchell, 1995, Soils: A New Global View, Yale University Press, New Haven, 213 p.

Putman, B. R., I. J. Jansen, and L. R. Follmer, 1988, Loessial soils: Their relationship to width of the source valley in Illinois, Soil Science, v. 146, p. 241-247.

Severson, R. C. and H. T. Shacklette, 1988, Essential elements and soil amendments for plants: Sources and use for agriculture, U. S. Geological Survey Circular 1017, Washington, DC, 48 p.

Shacklette, H. T. and J. G. Boerngen, 1984, Element concentrations in soils and other surficial materials of the conterminous United States, U.S. Geological Survey Professional Paper 1270, U. S. Geological Survey, Washington, DC, 105 p.

Simonson, R. W., 1978, A multiple-process model of soil genesis, *in* Mahaney, W. C., ed., Quaternary Soils, Geo Abstracts, Norwich, England, p. 1-25.

Soil Survey Division Staff, 1993, Soil Survey Manual, U. S. Department of Agriculture Handbook No. 18, U. S. Government Printing Office, Washington, DC, 437 p.

Sparks, D. L., 1995, Environmental Soil Chemistry, Academic Press, San Diego, 267 p.

Stevenson, I. L., 1964, Biochemistry of Soil, *in* Bear, F. E., ed., Chemistry of the Soil, Second Edition, American Chemistry Society Monograph Series, Reinhold Publishing Corp., New York, 515 p.

Weatherley, A. H., P. S. Lake, and S. C. Rogers, 1980, Zinc pollution and the ecology of the freshwater environment, *in* Nriagu, J. O., ed., Zinc in the Environment, Part I: Ecological Cycling, Wiley-Interscience, New York, p. 337-418.

White, M. L., 1957, The occurrence of zinc in soil, Economic Geology, v. 52, p. 645-651.

Wilding, L. P., N. E. Smeck, and L. R. Drees, 1977, Silica in soils: Quartz, cristobalite, tridymite, and opal; *in* Dixon, J. B. and S. B. Weed, eds., Minerals in Soil Environments, Soil Science Society of America, Madison, WI, 948 p.

Willman, H. B. and J. C. Frye, 1970, Pleistocene Stratigraphy of Illinois, Bulletin 94, Illinois State Geological Survey, 204 p.

Yin, Y., H. E. Allen, C. P. Huang, D. L. Sparks, and P. F. Sanders, 1997, Kinetics of mercury(II) adsorption and desorption on soil, Environ. Sci. Technol. v. 31, p. 496-503.

Yuan, T. L., N. Gammon, Jr., and R. G. Leighty, 1967, Relative contribution of organic and clay fractions to cation-exchange capacity of sandy soils from several soil groups, Soil Sci., v. 104, p. 123-128.

Zhang, Y. and J. K. Frost, 2002, Regional distribution of selected elements in Illinois soils, Illinois State Geological Survey, Environmental Geology 154.

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

Core Number	County Name	Final Depth of Core (ft)
11	Franklin	21.6
12	Hamilton	22.2
13	White	15.2
14	Washington	20.4
15	Washington	17.6
16	Monroe	14.5
17	Monroe	16.9
18	Clinton	17.5
19	Clinton	16.3
20	Marion	14.6
21	Wayne	15.2
22	Edwards	10.8
23	Wabash	21.8
24	Edwards	15.1
25	Wayne	22.8
26	Jefferson	9.8

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	8.0	-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	3.0
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	1.0	1	0.7	0.3
20	Lorenzo-Warsaw-Wea	0.7	1	0.7	0
21	Jasper-LaHogue-Selma	1.2	3	2.2	-1.0
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	1.0	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	1.0	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 11*

Sample	Depth Interval (ft)	Horizon	Clay (%)	Sand (%)	Silt (%)	Texture
11-1	0.0-0.7	Ap	17.36	7.06	75.58	silt loam
11-2	0.7-1.4	AE	16.72	5.41	77.87	silt loam
11-3	1.4-2.0	AE	13.52	7.38	79.10	silt loam
11-5	2.8-3.5	B	14.50	14.10	71.40	silt loam
11-8	5.0-5.7	B	15.56	19.04	65.40	silt loam
11-34	22.2-22.7	C	16.57	27.36	56.07	silt loam

*Percentages in Tables 3 through 18 are weight-percent.

Table 4. Texture of samples from core 12

Sample	Depth Interval (ft)	Horizon	Clay (%) (%)Interval	Sand (%)	Silt (%)	Texture
12-2	0.3-0.8	Ap	32.75	4.74	62.51	silty clay loam
12-3	0.8-1.4	B	38.67	3.87	57.46	silty clay loam
12-4	1.4-2.0	B	37.86	3.90	58.24	silty clay loam
12-6	2.8-3.4	B	39.40	3.18	57.42	silty clay loam
12-19	9.2-10.0	2B	24.76	5.64	69.60	silt loam
12-27	22.0-22.5	3C	31.62	21.73	46.65	clay loam

Table 5. Texture of samples from core 13

Sample	Depth Interval (ft)	Horizon	Clay (%)	Sand (%)	Silt (%)	Texture
13-1	0.0-0.61	Ap	7.84	68.39	23.77	sandy loam
13-2	0.61-1.38	B	12.40	60.67	26.93	sandy loam
13-3	1.38-2.0	B	19.92	57.53	22.55	sandy loam
13-4	2.0-2.8	B	18.60	64.53	16.87	sandy loam
13-5	10.5-13.0	C	5.02	92.05	2.93	sand

Table 6. Texture of samples from core 14

Sample	Depth Interval (ft)	Horizon	Clay (%)	Sand (%)	Silt (%)	Texture
14-1	0.0-0.15	Ap	14.24	7.31	78.45	silt loam
14-2	0.15-0.75	Ap	14.40	6.15	79.45	silt loam
14-3	0.75-1.4	Ap	16.22	4.39	79.39	silt loam
14-4	1.4-1.8	E	28.44	4.04	67.52	silty clay loam
14-6	2.8-3.65	B	32.68	3.86	63.46	silty clay loam
14-15	6.5-7.1	2B	22.96	12.08	64.96	silt loam

Table 7. Texture of samples from core 15

Sample	Depth Interval (ft)	Horizon	Clay (%)	Sand (%)	Silt (%)	Texture
15-1	0.0-0.9	A	18.44	1.49	80.07	silt loam
15-2	0.9-1.7	A	18.84	2.79	78.37	silt loam
15-3	1.7-1.9	E	–*	--	--	--
15-4	1.9-2.0	B	57.76	1.43	40.81	silty clay
15-13	6.8-8.8	B	28.06	10.28	61.66	silty clay
15-20	15.0-16.0	C	32.30	25.53	41.17	clay loam

*Insufficient sample available to make the determinations of clay, sand, and silt content.

Table 8. Texture of samples from core 16

Sample	Depth Interval (ft)	Horizon	Clay (%)	Sand (%)	Silt (%)	Texture
16-1	0.0-0.4	A	30.83	1.16	68.01	silty clay loam
16-2	0.4-1.0	E	38.64	1.25	60.11	silty clay loam
16-3	1.0-1.9	B	43.94	2.27	53.79	silty clay
16-4	1.9-2.8	B	32.88	0.96	66.16	silty clay loam
16-16	9.1-9.7	2A	31.90	13.46	54.64	silty clay loam
16-26	18.22-19.0	2B	15.63	31.64	52.73	silt loam

Table 9. Texture of samples from core 17

Sample	Depth Interval (ft)	Horizon	Clay (%)	Sand (%)	Silt (%)	Texture
17-1	0.0-0.5	A	35.31	16.04	48.65	silty clay loam
17-2	0.5-1.0	A	37.10	14.69	48.21	silty clay loam
17-3	1.0-1.5	A	38.82	16.55	44.63	silty clay loam
17-4	1.5-2.0	A	36.71	24.54	38.75	clay loam
17-16	7.9-10.0	C	13.22	54.04	32.74	loam

Table 10. Texture of samples from core 18

Sample	Depth Interval (ft)	Horizon	Clay (%)	Sand (%)	Silt (%)	Texture
18-1	0.0-0.3	A	11.16	4.50	84.34	silt loam
18-2	0.3-0.8	A	11.21	4.11	84.68	silt loam
18-3	0.8-1.1	A	11.14	4.46	84.40	silt loam
18-4	1.1-1.45	E	10.54	5.02	84.44	silt loam
18-10	3.2-4.0	B	32.58	1.92	65.50	silty clay
18-16	8.0-10.0	C	19.96	25.50	54.54	silt loam

Table 11. Texture of samples from core 19

Sample	Depth Interval (ft)	Horizon	Clay (%)	Sand (%)	Silt (%)	Texture
19-1	0.0-0.25	A	14.11	16.64	69.25	silt loam
19-2	0.25-0.6	A	15.40	5.50	79.10	silt loam
19-3	0.6-1.0	E	16.76	7.00	76.24	silt loam
19-4	1.0-1.35	B	40.49	2.00	57.51	silty clay
19-5	1.35-1.8	B	45.17	0.83	54.00	silty clay
19-13	5.6-7.0	2B	25.84	18.19	55.97	silt loam

Table 12. Texture of samples from core 20

Sample	Depth Interval (ft)	Horizon	Clay (%)	Sand (%)	Silt (%)	Texture
20-1	0.0-0.44	A	14.92	12.39	72.69	silt loam
20-2	0.44-0.7	E	15.28	7.33	77.39	silt loam
20-3	0.7-1.25	B	34.12	2.93	62.95	silty clay loam
20-4	1.25-1.95	B	43.72	1.52	54.76	silty clay
20-8	4.2-5.7	B	21.76	22.16	56.08	silt loam
20-11	7.6-8.0	C	18.86	25.66	55.48	silt loam

Table 13. Texture of samples from core 21

Sample	Depth Interval (ft)	Horizon	Clay (%)	Sand (%)	Silt (%)	Texture
21-1	0.0-0.4	A	22.68	5.68	71.64	silt loam
21-2	0.4-1.15	A	22.52	5.94	71.54	silt loam
21-3	1.15-2.0	E	18.84	9.08	72.08	silt loam
21-4	2.0-2.8	B	44.25	3.28	52.47	silty clay
21-7	4.6-5.8	B	30.38	7.60	62.02	silty clay loam
21-10	6.5-7.4	B	35.35	11.68	52.97	silty clay loam

Table 14. Texture of samples from core 22

Sample	Depth Interval (ft)	Horizon	Clay (%)	Sand (%)	Silt (%)	Texture
22-1	0.0-0.3	A	13.60	28.60	57.80	silt loam
22-2	0.3-1.25	A	9.68	42.22	48.10	loam
22-3	1.25-2.0	A	9.32	36.03	54.65	silt loam
22-4	2.0-2.8	A	8.24	46.76	45.00	silt loam
22-8	4.55-6.0	C	12.22	32.92	54.86	silt loam
22-14	9.5-10.0	C	13.68	71.92	14.40	sandy loam

Table 15. Texture of samples from core 23

Sample	Depth Interval (ft)	Horizon	Clay (%)	Sand (%)	Silt (%)	Texture
23-1	0.0-0.5	A	16.20	4.78	79.02	silt loam
23-2	0.5-0.9	A	14.92	2.65	82.43	silt loam
23-3	0.9-1.75	B	20.90	1.26	77.84	silt loam
23-4	1.75-2.3	A’	31.88	0.56	67.56	silty clay loam
23-8	5.0-6.0	C	21.84	0.96	77.20	silt loam
23-14	10.35-11.0	C	28.16	30.77	41.07	clay loam

Table 16. Texture of samples from core 24

Sample	Depth Interval (ft)	Horizon	Clay (%)	Sand (%)	Silt (%)	Texture
24-1	0.0-0.6	A	15.00	4.56	80.44	silt loam
24-2	0.6-1.2	A	21.20	1.74	77.06	silt loam
24-3	1.2-1.45	B	29.12	1.02	69.86	silty clay loam
24-4	1.45-2.0	B	29.35	1.00	69.65	silty slay loam
24-10	6.05-7.0	C	19.52	34.26	46.22	loam
24-16	9.35-9.9	2B	22.64	46.70	30.66	loam

Table 17. Texture of samples from core 25

Sample	Depth Interval (ft)	Horizon	Clay (%)	Sand (%)	Silt (%)	Texture
25-1	0.0-0.7	A	18.68	6.12	75.20	silt loam
25-2	0.7-1.0	C	19.03	6.01	74.96	silt loam
25-3	1.0-2.0	C	22.52	6.09	71.39	silt loam
25-4	2.0-2.9	C	27.76	5.36	66.88	silt loam
25-22	8.4-9.0	2C	37.86	3.70	58.44	silty clay loam
25-29	14.3-15.7	2C	35.68	10.46	53.86	silty clay loam

Table 18. Texture of samples from core 26

Sample	Depth Interval (ft)	Horizon	Clay (%)	Sand (%)	Silt (%)	Texture
26-1	0.0-0.65	A	11.88	11.72	76.40	silt loam
26-2	0.65-1.6	A	14.44	18.70	66.86	silt loam
26-3	1.6-1.9	E	16.14	19.90	63.96	silt loam
26-4	1.9-2.8	B	13.62	28.60	57.78	silt loam
26-8	5.0-5.7	2C	34.98	19.82	45.20	silty clay loam
26-15	8.0-9.1	2D	24.02	2.38	73.68	silt loam

Table 19. Soil texture, soil type, and developmental environment

Core Number	Texture	Soil Type*	Soil Association	Developmental Environment
11	silt loam	Bonnie	Haymond-Petrolia-Karnak	alluvial, forest
12	silty clay loam	Zipp	Markland-Colp-Del Rey	lacustrine, forest
13	sandy loam	Alvin	Oakville-Lamont-Alvin	windblown sand
14	silt loam	Bluford	Ava-Bluford-Wynoose	upland, forest
15	silt loam	Bluford	Ava-Bluford-Wynoose	upland, forest
16	silty clay loam	Marine	Alford-Muren-Iva	upland, forest
17	silty clay loam	Riley	Lawson-Sawmill-Darwin	alluvial, prairie
18	silt loam	Cowden	Oconee-Cowden-Piasa	prairie
19	silt loam	Hoyleton	Hoyleton-Cisne-Huey	prairie
20	silt loam	Ava	Ava-Bluford-Wynoose	upland, forest
21	silt loam	Wynoose	Ava-Bluford-Wynoose	upland, forest
22	silt loam	Belknap	Haymond-Petrolia-Karnak	alluvial, forest
23	silt loam	Hosmer	Hosmer-Stoy-Weir	upland, forest
24	silt loam	Hosmer	Hosmer-Stoy-Weir	upland, forest
25	silt loam	Bonnie	Haymond-Petrolia-Karnak	alluvial, forest
26	silt loam	Bluford	Ava-Bluford-Wynoose	alluvial, forest

*Designations of soil types in this report are provisional and are subject to change after more detailed examination of the cores. Soil names were those of the map unit in which cores were collected.

Table 20. Correlation coefficients for constituents of cores 11 through 26 (C.I. = 95%)

	Depth	SiO_2	Al_2O_3	Fe_2O_3	K_2O	CaO	MgO	Na_2O	TiO_2
Depth	1.00								
SiO_2	-0.18	1.00							
Al_2O_3	0.16	-0.86	1.00						
Fe_2O_3	0.09	-0.83	0.90	1.00					
K_2O	0.07	-0.54	0.56	0.43	1.00				
CaO	0.29	-0.39	-0.08	-0.06	0.03	1.00			
MgO	0.37	-0.83	0.59	0.58	0.49	0.59	1.00		
Na_2O	-0.05	0.16	-0.13	-0.29	0.16	0.01	-0.07	1.00	
TiO_2	-0.17	-0.24	0.36	0.27	0.17	-0.23	-0.05	-0.13	1.00
P_2O_5	-0.19	-0.40	0.22	0.17	0.48	0.22	0.34	0.20	-0.15
MnO	-0.27	-0.06	-0.10	0.06	0.06	0.07	0.02	0.05	0.11
SO_3	0.12	-0.28	0.21	0.05	0.54	0.26	0.38	0.67	-0.31
Ba	-0.10	-0.42	0.43	0.30	0.59	0.00	0.30	0.38	0.22
Cr	0.08	-0.72	0.77	0.72	0.36	0.06	0.64	-0.03	0.01
Cu	0.07	-0.76	0.84	0.77	0.44	-0.04	0.51	-0.20	0.22
Hg	-0.23	-0.25	0.24	0.33	0.04	-0.06	0.11	-0.06	0.05
Ni	0.13	-0.35	0.43	0.41	-0.03	-0.02	0.34	-0.06	-0.02
Pb	-0.23	-0.38	0.29	0.27	0.30	0.08	0.25	0.02	-0.04
Rb	0.21	-0.74	0.87	0.70	0.63	-0.04	0.49	-0.24	0.39
Sr	0.14	-0.21	0.16	0.00	0.53	0.25	0.32	0.70	-0.36
V	0.02	-0.83	0.91	0.89	0.45	-0.07	0.55	-0.15	0.39
Zn	0.09	-0.83	0.83	0.76	0.62	0.08	0.63	-0.14	0.11
Zr	-0.37	0.52	-0.52	-0.48	-0.33	-0.19	-0.56	0.14	0.48
LOI	0.02	-0.74	0.34	0.36	0.18	0.69	0.68	-0.22	0.07
Moisture	-0.03	-0.64	0.73	0.72	0.22	-0.09	0.45	-0.06	0.16
Tot C	-0.21	-0.22	-0.21	-0.18	-0.01	0.61	0.25	-0.03	-0.11
In C	0.32	-0.36	-0.10	-0.07	-0.02	0.96	0.62	-0.08	-0.19
Org C	-0.07	0.12	-0.10	-0.16	0.01	-0.01	-0.10	0.19	0.01
Clay	0.06	-0.78	0.86	0.85	0.38	-0.04	0.54	-0.18	0.25
Sand	0.32	0.43	-0.43	-0.38	-0.28	0.07	-0.18	-0.11	-0.83
Silt	-0.39	0.02	-0.07	-0.12	0.07	-0.05	-0.14	0.23	0.75
Sand+Silt	-0.02	0.71	-0.77	-0.82	-0.23	0.05	-0.46	0.19	-0.22
pH	0.34	-0.08	-0.08	-0.06	0.24	0.36	0.27	-0.01	-0.38

Table 20 continued. Correlation coefficients for constituents of cores 11 through 26 (C.I. = 95%)

	P_2O_5	MnO	SO_3	Ba	Cr	Cu	Hg	Ni	Pb
P_2O_5	1.00								
MnO	0.24	1.00							
SO_3	0.51	0.15	1.00						
Ba	0.55	0.41	0.70	1.00					
Cr	0.20	-0.02	0.40	0.37	1.00				
Cu	0.28	-0.22	0.14	0.30	0.72	1.00			
Hg	0.34	0.22	-0.01	0.22	0.31	0.28	1.00		
Ni	-0.06	0.19	0.19	0.15	0.43	0.35	0.26	1.00	
Pb	0.78	0.05	0.17	0.32	0.21	0.38	0.37	-0.17	1.00
Rb	0.19	-0.30	0.11	0.30	0.64	0.82	0.03	0.20	0.28
Sr	0.47	0.06	0.95	0.57	0.37	0.10	-0.03	0.09	0.19
V	0.28	0.13	0.21	0.51	0.81	0.82	0.31	0.45	0.35
Zn	0.57	-0.09	0.29	0.46	0.68	0.86	0.33	0.23	0.65
Zr	-0.27	0.26	-0.32	-0.10	-0.52	-0.52	-0.13	-0.31	-0.27
LOI	0.46	0.11	0.08	0.16	0.37	0.41	0.20	0.09	0.42
Moisture	0.34	0.09	0.22	0.43	0.66	0.69	0.35	0.40	0.42
Tot C	0.53	0.19	0.03	0.02	-0.19	-0.07	0.14	-0.33	0.41
In C	0.10	0.03	0.13	-0.11	-0.01	-0.06	-0.11	-0.07	0.01
Org C	-0.09	-0.05	0.06	0.06	-0.20	-0.09	-0.10	-0.33	-0.09
Clay	0.26	-0.04	0.15	0.34	0.69	0.78	0.29	0.40	0.37
Sand	-0.12	-0.24	0.05	-0.42	-0.14	-0.30	-0.15	-0.06	-0.17
Silt	-0.03	0.28	-0.14	0.24	-0.32	-0.16	-0.04	-0.18	-0.05
Sand+Silt	-0.19	0.07	-0.04	-0.22	-0.64	-0.76	-0.31	-0.32	-0.35
pH	0.13	0.09	0.23	0.04	-0.06	-0.18	-0.02	0.07	-0.05

Table 20 continued. Correlation coefficients for constituents of cores 11 through 26 (C.I. = 95%)

	Rb	Sr	V	Zn	Zr	LOI	Moisture	Total C	Inorg C
Rb	1.00								
Sr	0.11	1.00							
V	0.76	0.11	1.00						
Zn	0.81	0.26	0.80	1.00					
Zr	-0.47	-0.29	-0.39	-0.61	1.00				
LOI	0.34	0.05	0.39	0.51	-0.27	1.00			
Moisture	0.52	0.13	0.84	0.68	-0.36	0.29	1.00		
Tot C	-0.15	0.07	-0.16	0.10	0.06	0.75	-0.11	1.00	
In C	-0.06	0.13	-0.10	0.02	-0.17	0.67	-0.14	0.57	1.00
Org C	-0.08	0.07	-0.12	-0.13	0.11	-0.09	-0.12	0.02	-0.03
Clay	0.73	0.09	0.91	0.78	-0.44	0.38	0.87	-0.15	-0.09
Sand	-0.36	0.11	-0.51	-0.27	-0.40	-0.30	-0.38	-0.10	0.07
Silt	-0.06	-0.18	-0.01	-0.19	0.71	0.08	-0.12	0.20	-0.02
Sand+Silt	-0.64	0.01	-0.83	-0.69	0.39	-0.38	-0.76	0.12	0.08
pH	-0.07	0.26	-0.23	0.01	-0.25	0.13	-0.22	0.23	0.29

	Org C	Clay	Sand	Silt	Sand+Silt	pH
Org C	1.00					
Clay	-0.12	1.00				
Sand	-0.07	-0.43	1.00			
Silt	0.15	-0.16	-0.82	1.00		
Sand+Silt	0.11	-1.00	0.43	0.16	1.00	
pH	0.08	-0.20	0.29	-0.19	0.20	1.00

Table 21. Elemental composition of samples selected from Core 11

Subsample	11-1	11-2	11-3	11-5	11-8	11-34	Average
Lab. No.	R22127	R22128	R22129	R22130	R22131	R22132	
Depth Interval (ft)	0.0-0.7	0.7-1.4	1.4-2.0	2.8-3.5	5.0-5.7	22.2-22.7	
Horizon	Ap	AE	AE	B	B	C	
SiO_2 (%)	81.40	81.80	82.40	81.70	82.00	82.30	81.93
Al_2O_3 (%)	8.00	8.30	7.90	8.30	8.90	7.60	8.17
Fe_2O_3 (%)	2.67	2.30	2.20	2.23	2.20	2.25	2.31
K_2O (%)	1.74	1.70	1.66	1.55	1.63	1.75	1.67
CaO (%)	0.44	0.40	0.32	0.26	0.31	0.58	0.39
MgO (%)	0.28	0.30	0.26	0.24	0.32	0.54	0.32
Na_2O (%)	0.99	1.00	0.99	0.95	1.01	1.09	1.01
TiO_2 (%)	0.79	0.90	0.87	0.85	0.82	0.68	0.82
P_2O_5 (%)	0.10	0.10	0.09	0.08	0.07	0.11	0.09
MnO_2 (%)	0.10	0.10	0.09	0.07	0.15	0.03	0.09
SO_3 (%)	ND*	ND	ND	ND	ND	ND	—
Barium	416	521	555	515	582	539	521
Chromium	47	12	15	32	8	9	21
Copper	24	21	22	26	18	20	22
Mercury (µg/kg)	25	29	24	26	30	26	27
Nickel	21	6	6	13	8	7	10
Lead	19	16	16	17	12	15	16
Rubidium	61	63	61	69	52	60	61
Strontium	125	104	97	94	100	144	111
Vanadium	69	71	69	79	64	70	70
Zinc	50	45	41	69	36	42	47
Zirconium	426	416	399	393	391	424	408
Total C (%)	1.39	0.70	0.63	0.55	0.15	0.49	0.65
Inorganic C (%)	0.05	0.04	0.04	0.03	0.03	0.03	0.04
Organic C (%)	1.34	0.66	0.59	0.52	0.12	0.46	0.62
pH	5.42	5.12	4.86	4.64	4.83	4.90	4.96

*Not Determined

Table 22. Elemental composition of samples selected from Core 12

Subsample	12-02	12-03	12-04	12-06	12-19	12-27	Average
Lab. No.	R22133	R22134	R22135	R22136	R22137	R22203	
Depth Interval (ft)	0.3-0.8	0.8-1.4	1.4-2.0	2.8-3.4	9.2-10.0	22.0-22.5	
Horizon	Ap	B	B	B	2B	3C	
SiO_2 (%)	73.00	70.60	69.20	71.50	76.70	61.60	70.43
Al_2O_3 (%)	11.30	12.60	12.60	13.40	11.40	13.50	12.47
Fe_2O_3 (%)	4.61	5.25	5.33	5.67	3.61	4.89	4.89
K_2O (%)	2.43	2.49	2.47	2.57	2.13	2.26	2.39
CaO (%)	0.81	0.78	0.75	0.78	0.77	5.48	1.56
MgO (%)	0.91	1.13	1.18	1.30	0.78	1.93	1.21
Na_2O (%)	0.91	0.81	0.80	0.84	1.13	0.80	0.88
TiO_2 (%)	0.74	0.74	0.72	0.75	0.80	0.80	0.76
P_2O_5 (%)	0.17	0.10	0.09	0.09	0.11	0.13	0.12
MnO_2 (%)	0.06	0.07	0.08	0.11	0.06	0.06	0.07
SO_3 (%)	ND	ND	ND	ND	ND	ND	—
Barium	458	493	517	515	442	633	510
Chromium	60	63	57	54	41	51	54
Copper	32	30	27	27	26	35	30
Mercury (µg/kg)	23	23	23	24	24	26	24
Nickel	16	15	13	17	12	11	14
Lead	21	21	21	22	19	21	21
Rubidium	100	107	107	109	91	113	105
Strontium	135	133	122	126	147	111	129
Vanadium	94	106	102	108	86	107	101
Zinc	80	86	88	90	68	102	86
Zirconium	333	298	294	294	400	197	303
Total C (%)	1.15	0.66	0.60	0.46	0.27	1.54	0.78
Inorganic C (%)	0.06	0.05	0.05	0.04	0.05	1.15	0.23
Organic C (%)	1.09	0.61	0.55	0.42	0.22	0.39	0.55
pH	6.31	6.48	6.50	6.68	7.05	7.78	6.80

Table 23. Elemental composition of samples selected from Core 13

Subsample	13-01	13-02	13-03	13-04	13-18	Average
Lab. No.	R21985	R21986	R21987	R21988	R21989	
Depth Interval (ft)	0.0-0.61	0.61-1.38	1.38-2.0	2.0-2.8	10.5-13.0	
Horizon	Ap	B	B	B	C	
SiO_2 (%)	84.50	84.28	80.72	80.19	86.99	83.34
Al_2O_3 (%)	6.00	7.07	8.65	9.09	5.97	7.36
Fe_2O_3 (%)	1.85	2.20	3.16	3.35	1.77	2.46
K_2O (%)	1.68	1.95	2.05	2.05	1.35	1.82
CaO (%)	0.79	0.55	0.57	0.62	0.75	0.66
MgO (%)	0.21	0.29	0.43	0.45	0.37	0.35
Na_2O (%)	0.92	0.90	0.91	0.95	1.26	0.99
TiO_2 (%)	0.40	0.50	0.49	0.42	0.19	0.40
P_2O_5 (%)	0.13	0.09	0.11	0.09	0.05	0.09
MnO_2 (%)	0.09	0.08	0.07	0.08	0.06	0.08
SO_3 (%)	0.01	0.01	0.01	0.01	0.02	0.01
Barium	436	477	484	468	286	430
Chromium	<5	<5	17	30	<5	12
Copper	24	20	25	24	21	23
Mercury (µg/kg)	92	23	<2	2	<2	24
Nickel	<5	<5	<5	7	19	8
Lead	32	14	15	16	11	18
Rubidium	41	55	69	66	37	54
Strontium	142	139	138	146	158	145
Vanadium	45	55	73	76	37	57
Zinc	71	42	51	55	41	52
Zirconium	266	278	247	254	99	229
Total C (%)	1.46	0.28	0.23	0.23	0.13	0.47
Inorganic C (%)	0.06	0.04	0.03	0.04	0.03	0.04
Organic C (%)	1.40	0.24	0.20	0.19	0.10	0.43
pH	5.97	6.44	6.46	6.58	6.72	6.43

Table 24. Elemental composition of samples selected from Core 14

Subsample	14-01	14-02	14-03	14-04	14-06	14-15	Average
Lab. No.	R21990	R21991	R21992	R21993	R21994	R21995	
Depth Interval (ft)	0.0-0.15	0.15-0.75	0.75-1.4	1.4-1.8	2.8-3.65	6.5-7.1	
Horizon	Ap	Ap	A	E	B	2B	
SiO_2 (%)	78.56	79.39	79.56	76.14	73.29	78.67	77.60
Al_2O_3 (%)	7.33	7.57	8.53	11.10	12.28	9.94	9.46
Fe_2O_3 (%)	3.09	2.95	3.28	4.31	5.23	3.40	3.71
K_2O (%)	1.75	1.75	1.88	1.93	1.99	1.75	1.84
CaO (%)	0.57	0.57	0.46	0.37	0.50	0.61	0.52
MgO (%)	0.35	0.35	0.47	0.78	0.92	0.55	0.57
Na_2O (%)	0.88	0.88	0.94	0.95	1.06	1.12	0.97
TiO_2 (%)	0.79	0.82	0.85	0.81	0.76	0.78	0.80
P_2O_5 (%)	0.14	0.12	0.11	0.10	0.10	0.07	0.11
MnO_2 (%)	0.35	0.39	0.34	0.09	0.06	0.06	0.21
SO_3 (%)	0.01	0.01	0.01	0.01	0.02	0.01	0.01
Barium	640	685	672	649	648	488	630
Chromium	21	35	34	68	88	38	47
Copper	23	20	19	25	34	25	24
Mercury (µg/kg)	20	29	31	41	13	14	25
Nickel	6	<5	<5	8	9	27	10
Lead	23	21	21	20	23	17	21
Rubidium	54	54	61	81	80	74	67
Strontium	105	116	109	116	155	127	121
Vanadium	73	78	87	108	113	87	91
Zinc	51	51	50	72	86	59	61
Zirconium	459	463	435	368	348	351	404
Total C (%)	1.97	1.23	0.39	0.25	0.22	0.16	0.70
Inorganic C (%)	0.05	0.03	0.06	0.04	0.05	0.03	0.04
Organic C (%)	1.92	1.20	0.33	0.21	0.17	0.13	0.66
pH	5.86	6.37	6.40	4.28	3.86	5.01	5.30

Table 25. Elemental composition of samples selected from Core 15

Subsample	15-01	15-02	15-03	15-04	15-13	15-20	Average
Lab. No.	R21996	R21997	R21998	R21999	R22000	R22001	
Depth Interval (ft)	0.0-0.9	0.9-1.7	1.7-1.9	1.9-2.0	6.8-8.8	15.0-16.0	
Horizon	A	A	E	B	B	C	
SiO_2 (%)	81.10	79.72	67.83	62.54	77.37	73.98	73.76
Al_2O_3 (%)	8.10	8.87	14.07	17.55	11.27	11.89	11.96
Fe_2O_3 (%)	2.10	3.48	7.06	7.87	2.75	5.41	4.78
K_2O (%)	1.52	1.59	1.46	1.40	1.84	1.87	1.61
CaO (%)	0.56	0.38	0.38	0.40	0.71	0.55	0.50
MgO (%)	0.27	0.34	0.83	1.12	0.58	0.76	0.65
Na_2O (%)	1.08	1.09	0.80	0.64	1.10	0.83	0.92
TiO_2 (%)	0.77	0.80	0.79	0.78	0.77	0.75	0.78
P_2O_5 (%)	0.09	0.08	0.09	0.10	0.09	0.09	0.09
MnO_2 (%)	0.16	0.18	0.06	0.07	0.03	0.13	0.10
SO_3 (%)	0.01	0.01	0.01	0.01	0.01	0.01	0.01
Barium	597	531	421	418	502	404	479
Chromium	15	33	99	122	45	57	62
Copper	22	18	40	43	27	24	29
Mercury (µg/kg)	27	18	50	77	30	30	39
Nickel	7	7	17	32	17	30	18
Lead	19	20	27	29	19	19	22
Rubidium	60	56	96	112	93	83	83
Strontium	115	106	84	74	125	96	100
Vanadium	71	87	136	152	92	99	106
Zinc	45	40	93	116	68	81	74
Zirconium	411	396	290	242	348	289	329
Total C (%)	0.98	0.54	0.69	0.81	0.23	0.10	0.56
Inorganic C (%)	0.04	0.05	0.06	0.05	0.05	0.05	0.05
Organic C (%)	0.94	0.49	0.63	0.76	0.18	0.05	0.51
pH	5.54	4.74		4.50	6.50	7.21	5.70

Table 26. Elemental composition of samples selected from Core 16

Subsample	16-01	16-02	16-03	16-04	16-16	16-26	Average
Lab. No.	R22002	R22003	R22004	R22005	R22006	R22007	
Depth Interval (ft)	0.0-0.4	0.4-1.0	1.0-1.9	1.9-2.8	9.1-9.7	18.2-19.0	
Horizon	A	E	B	B	2A	2B	
SiO_2 (%)	72.40	69.06	66.60	70.80	75.93	65.10	69.98
Al_2O_3 (%)	11.33	13.49	14.60	12.78	11.63	8.01	11.97
Fe_2O_3 (%)	4.63	6.33	6.41	4.96	3.70	2.76	4.80
K_2O (%)	2.17	2.04	2.04	2.07	1.93	1.94	2.03
CaO (%)	0.74	0.66	0.77	1.11	0.88	6.70	1.81
MgO (%)	0.90	1.20	1.51	1.38	0.85	3.56	1.57
Na_2O (%)	0.88	0.88	0.89	1.19	0.95	0.87	0.94
TiO_2 (%)	0.77	0.74	0.69	0.73	0.71	0.50	0.69
P_2O_5 (%)	0.13	0.13	0.15	0.15	0.09	0.09	0.12
MnO_2 (%)	0.11	0.14	0.50	0.26	0.12	0.05	0.20
SO_3 (%)	0.01	0.01	0.02	0.02	0.01	0.01	0.01
Barium	539	556	1015	798	640	354	650
Chromium	60	98	113	78	50	<5	67
Copper	29	34	33	24	23	22	27
Mercury (µg/kg)	36	37	55	62	36	9	39
Nickel	8	13	43	40	17	<5	21
Lead	24	24	19	16	16	18	19
Rubidium	83	82	62	61	77	66	72
Strontium	113	115	150	173	126	136	136
Vanadium	106	126	172	119	89	66	113
Zinc	74	87	81	78	58	64	74
Zirconium	359	297	254	317	281	209	286
Total C (%)	1.15	0.40	0.32	0.26	0.12	2.36	0.77
Inorganic C (%)	0.04	0.04	0.04	0.08	0.11	2.31	0.44
Organic C (%)	1.11	0.36	0.28	0.18	0.01	0.05	0.33
pH	6.45	5.18	5.78	7.10	6.89	7.78	6.53

Table 27. Elemental composition of samples selected from Core 17

Subsample	17-01	17-02	17-03	17-04	17-16	Average
Lab. No.	R22008	R22009	R22010	R22011	R22012	
Depth Interval (ft)	0.0-0.5	0.5-1.0	1.0-1.5	1.5-2.0	7.9-10.0	
Horizon	A	A	A	A	C	
SiO_2 (%)	70.36	69.92	70.20	71.95	78.50	72.19
Al_2O_3 (%)	12.22	12.58	12.90	12.53	10.03	12.05
Fe_2O_3 (%)	3.95	4.11	4.14	4.09	2.80	3.82
K_2O (%)	2.52	2.54	2.49	2.41	2.29	2.45
CaO (%)	1.16	1.10	1.06	1.01	1.15	1.09
MgO (%)	1.14	1.18	1.23	1.21	0.78	1.11
Na_2O (%)	1.14	1.00	1.03	1.08	1.57	1.16
TiO_2 (%)	0.59	0.60	0.59	0.57	0.46	0.56
P_2O_5 (%)	0.33	0.35	0.27	0.16	0.15	0.25
MnO_2 (%)	0.10	0.10	0.09	0.09	0.05	0.09
SO_3 (%)	0.02	0.02	0.02	0.02	0.03	0.02
Barium	791	845	827	804	843	822
Chromium	58	52	58	64	169	80
Copper	32	33	30	29	22	29
Mercury (µg/kg)	45	46	35	25	20	34
Nickel	6	6	11	9	7	8
Lead	72	60	38	22	18	42
Rubidium	93	100	97	95	73	92
Strontium	169	160	165	174	255	185
Vanadium	110	113	118	121	78	108
Zinc	138	139	111	91	63	108
Zirconium	217	197	188	213	323	228
Total C (%)	1.85	1.81	1.36	0.93	0.27	1.24
Inorganic C (%)	0.07	0.07	0.08	0.08	0.07	0.07
Organic C (%)	1.78	1.74	1.28	0.85	0.20	1.17
pH	6.36	5.98	5.78	5.94	6.62	6.14

Table 28. Elemental composition of samples selected from Core 18

Subsample	18-01	18-02	18-03	18-04	18-10	18-16	Average
Lab. No.	R22013	R22014	R22015	R22016	R22017	R22018	
Depth Interval (ft)	0.0-0.3	0.3-0.8	0.8-1.1	1.1-1.45	3.2-4.0	8.0-10.0	
Horizon	A	A	A	E	B	C	
SiO_2 (%)	82.18	83.29	83.82	84.03	71.80	81.54	81.11
Al_2O_3 (%)	7.02	7.04	7.36	7.36	12.70	9.00	8.41
Fe_2O_3 (%)	1.34	1.32	1.43	1.46	5.09	2.25	2.15
K_2O (%)	1.85	1.82	1.89	1.92	1.98	1.69	1.86
CaO (%)	0.92	0.75	0.61	0.55	0.84	0.70	0.73
MgO (%)	0.28	0.24	0.24	0.24	1.06	0.50	0.43
Na_2O (%)	1.23	1.25	1.26	1.32	1.17	1.08	1.22
TiO_2 (%)	0.70	0.71	0.75	0.76	0.73	0.59	0.71
P_2O_5 (%)	0.13	0.09	0.06	0.05	0.18	0.08	0.10
MnO_2 (%)	0.07	0.06	0.06	0.07	0.09	0.02	0.06
SO_3 (%)	0.01	0.01	0.01	0.01	0.02	0.01	0.01
Barium	579	560	635	550	1030	480	639
Chromium	6	7	<5	<5	71	44	23
Copper	19	20	20	19	32	25	23
Mercury (µg/kg)	18	25	16	15	55	27	26
Nickel	<5	<5	<5	<5	33	19	12
Lead	18	18	13	14	21	17	17
Rubidium	56	58	59	57	79	68	63
Strontium	135	135	137	138	153	137	139
Vanadium	56	56	59	64	113	72	70
Zinc	47	44	36	33	87	51	50
Zirconium	425	435	420	422	315	304	387
Total C (%)	1.50	0.96	0.70	0.40	0.05	0.18	0.63
Inorganic C (%)	0.06	0.03	0.01	0.02	0.01	0.01	0.02
Organic C (%)	1.44	0.93	0.69	0.38	0.04	0.17	0.61
pH	6.73	6.76	6.62	6.48	6.34	6.87	6.63

Table 29. Elemental composition of samples selected from Core 19

Subsample	19-01	19-02	19-03	19-04	19-05	19-13	Average
Lab. No.	R22019	R22020	R22021	R22022	R22023	R22024	
Depth Interval (ft)	0.0-0.25	0.25-0.6	0.6-1.0	1.0-1.35	1.35-1.8	5.6-7.0	
Horizon	A	A	E	B	B	2B	
SiO_2 (%)	66.88	81.15	78.30	70.10	68.60	77.81	73.81
Al_2O_3 (%)	6.12	7.52	8.86	13.70	14.66	10.30	10.20
Fe_2O_3 (%)	2.14	2.64	3.66	5.50	5.73	3.49	3.86
K_2O (%)	1.35	1.53	1.65	1.54	1.57	1.75	1.56
CaO (%)	8.73	0.71	0.42	0.42	0.48	0.79	1.92
MgO (%)	1.17	0.31	0.43	0.92	1.16	0.68	0.78
Na_2O (%)	0.83	1.04	1.08	0.87	0.95	1.18	0.99
TiO_2 (%)	0.59	0.77	0.80	0.81	0.79	0.68	0.74
P_2O_5 (%)	0.16	0.11	0.09	0.08	0.09	0.09	0.10
MnO_2 (%)	0.20	0.20	0.23	0.06	0.04	0.08	0.13
SO_3 (%)	0.02	0.01	0.01	0.01	0.01	0.01	0.01
Barium	461	557	550	468	558	516	518
Chromium	<5	29	37	88	95	47	50
Copper	18	17	17	32	40	24	25
Mercury (µg/kg)	28	29	26	37	35	32	31
Nickel	<5	<5	<5	17	17	14	10
Lead	23	24	24	24	22	17	22
Rubidium	48	55	56	90	99	74	70
Strontium	151	115	104	94	104	136	117
Vanadium	67	73	89	123	133	87	95
Zinc	50	46	44	83	96	58	63
Zirconium	354	451	389	309	274	312	348
Total C (%)	3.38	1.09	0.58	0.79	0.77	0.18	1.13
Inorganic C (%)	1.99	0.04	0.04	0.03	0.03	0.04	0.36
Organic C (%)	1.39	1.05	0.54	0.76	0.74	0.14	0.77
pH	6.92	6.85	5.23	4.68	4.80	7.54	6.00

Table 30. Elemental composition of samples selected from Core 20

Subsample	20-01	20-02	20-03	20-04	20-08	20-11	Average
Lab. No.	R22025	R22026	R22027	R22028	R22029	R22030	
Depth Interval (ft)	0.0-0.44	0.44-0.7	0.7-1.25	1.25-1.95	4.2-4.7	7.6-8.0	
Horizon	A	E	B	B	B	C	
SiO_2 (%)	79.29	80.46	79.22	69.67	80.39	82.14	78.53
Al_2O_3 (%)	7.75	8.44	8.06	14.44	9.17	8.39	9.37
Fe_2O_3 (%)	2.86	3.07	2.54	5.83	3.10	3.02	3.40
K_2O (%)	1.66	1.85	1.93	1.80	1.50	1.43	1.70
CaO (%)	0.92	0.30	0.69	0.35	0.48	0.45	0.53
MgO (%)	0.40	0.43	0.42	1.21	0.49	0.38	0.56
Na_2O (%)	0.95	1.01	0.98	0.83	0.91	0.86	0.92
TiO_2 (%)	0.78	0.86	0.80	0.78	0.75	0.74	0.79
P_2O_5 (%)	0.14	0.10	0.09	0.09	0.06	0.04	0.09
MnO_2 (%)	0.25	0.18	0.20	0.05	0.04	0.04	0.13
SO_3 (%)	0	0	0	0	0	0	0.01
Barium	617	588	604	509	598	414	555
Chromium	22	33	78	87	28	24	45
Copper	18	18	30	35	24	21	24
Mercury (µg/kg)	43	35	28	17	7	14	24
Nickel	4	<5	11	12	23	10	11
Lead	22	19	24	24	17	18	21
Rubidium	52	61	91	102	72	68	74
Strontium	100	106	123	104	106	96	106
Vanadium	77	85	118	128	83	79	95
Zinc	52	43	78	102	59	44	63
Zirconium	408.39	405.05	433.99	273.53	327.87	327.73	363
Total C (%)	1.64	0.27	0.28	0.26	0.17	0.18	0.47
Inorganic C (%)	0.13	0.03	0.03	0.03	0.03	0.03	0.05
Organic C (%)	1.51	0.24	0.25	0.23	0.14	0.15	0.42
pH	6.13	4.18	4.06	3.88	5.60	6.74	5.10

Table 31. Elemental composition of samples selected from Core 21

Subsample	21-01	21-02	21-03	21-04	21-07	21-10	Average
Lab. No.	R22031	R22032	R22033	R22034	R22035	R22036	
Depth Interval (ft)	0.0-0.4	0.4-1.15	1.15-2.0	2.0-2.8	4.6-5.8	6.5-7.4	
Horizon	A	A	E	B	B	B	
SiO_2 (%)	77.17	79.84	80.83	70.86	76.44	75.36	76.75
Al_2O_3 (%)	8.71	8.78	8.51	14.46	11.88	12.36	10.78
Fe_2O_3 (%)	2.71	2.74	2.93	4.72	2.83	3.62	3.26
K_2O (%)	1.56	1.57	1.63	1.55	1.87	1.66	1.64
CaO (%)	0.58	0.43	0.29	0.31	0.49	0.47	0.43
MgO (%)	0.40	0.37	0.33	0.73	0.65	0.60	0.51
Na_2O (%)	0.90	0.89	0.92	0.76	1.05	0.80	0.89
TiO_2 (%)	0.83	0.86	0.90	0.85	0.85	0.90	0.86
P_2O_5 (%)	0.15	0.07	0.08	0.08	0.08	0.09	0.09
MnO_2 (%)	0.04	0.03	0.04	0.02	0.02	0.04	0.03
SO_3 (%)	0.01	0.01	0.01	0.01	0.01	0.01	0.01
Barium	471	500	467	418	582	558	500
Chromium	29	22	29	100	52	58	48
Copper	27	23	23	32	30	29	27
Mercury (µg/kg)	37	28	23	48	23	29	31
Nickel	8	6	<5	27	15	26	15
Lead	25	20	18	24	18	21	21
Rubidium	72	69	71	107	89	108	86
Strontium	94	96	97	95	136	116	106
Vanadium	81	80	82	118	99	105	94
Zinc	57	46	43	83	64	65	60
Zirconium	358	366	380	285	340	308	339
Total C (%)	1.99	0.93	0.48	0.54	0.24	0.26	0.74
Inorganic C (%)	0.03	0.04	0.03	0.04	0.03	0.04	0.04
Organic C (%)	1.96	0.89	0.45	0.50	0.21	0.22	0.71
pH	5.14	4.64	4.45	4.21	4.57	5.12	4.69

Table 32. Elemental composition of samples selected from Core 22

Subsample	22-01	22-02	22-03	22-04	22-08	22-14	Average
Lab. No.	R22037	R22038	R22039	R22040	R22041	R22042	
Depth Interval (ft)	0.0-0.3	0.3-1.25	1.25-2.0	2.0-2.8	4.55-6.0	9.5-10.0	
Horizon	A	A	A	A	C	C	
SiO_2 (%)	79.53	84.82	86.20	87.12	84.04	84.38	84.35
Al_2O_3 (%)	6.47	5.98	6.00	5.38	7.58	6.74	6.36
Fe_2O_3 (%)	2.18	2.21	1.91	2.13	1.75	3.02	2.20
K_2O (%)	1.63	1.64	1.62	1.47	1.64	1.37	1.56
CaO (%)	0.84	0.48	0.28	0.22	0.43	0.33	0.43
MgO (%)	0.46	0.34	0.19	0.17	0.33	0.38	0.31
Na_2O (%)	0.75	0.67	0.66	0.53	1.01	0.59	0.70
TiO_2 (%)	0.66	0.60	0.64	0.57	0.66	0.30	0.57
P_2O_5 (%)	0.17	0.10	0.08	0.07	0.05	0.08	0.09
MnO_2 (%)	0.10	0.10	0.07	0.14	0.02	0.04	0.08
SO_3 (%)	0.01	0.01	0.01	0.01	0.01	0.01	0.01
Barium	435	398	424	324	450	253	381
Chromium	<5	<5	<5	<5	<5	19	7
Copper	20	17	19	18	19	22	19
Mercury (µg/kg)	43	30	25	24	16	16	26
Nickel	<5	<5	<5	<5	<5	13	6
Lead	20	18	16	15	15	16	16
Rubidium	53	51	55	46	58	59	54
Strontium	92	78	83	69	117	78	86
Vanadium	60	58	50	49	60	58	56
Zinc	57	46	38	35	40	53	45
Zirconium	361	350	370	324	352	155	319
Total C (%)	2.62	0.61	0.38	0.38	0.26	0.14	0.73
Inorganic C (%)	0.10	0.07	0.02	0.02	0.04	0.03	0.05
Organic C (%)	2.52	0.54	0.36	0.36	0.22	0.11	0.69
pH	6.58	6.77	6.66	6.74	6.36	6.80	6.65

Table 33. Elemental composition of samples selected from Core 23

Subsample	23-01	23-02	23-03	23-04	23-08	23-14	Average
Lab. No.	R22043	R22044	R22045	R22046	R22047	R22048	
Depth Interval (ft)	0.0-0.5	0.5-0.9	0.9-1.75	1.75-2.3	5.0-6.0	10.35-11.0	
Horizon	A	A	B	A’	C	C	
SiO_2 (%)	76.03	73.48	78.15	73.17	74.00	76.80	75.27
Al_2O_3 (%)	7.90	12.43	9.64	12.46	11.39	10.87	10.78
Fe_2O_3 (%)	2.85	4.59	3.18	5.04	4.94	4.63	4.20
K_2O (%)	1.92	1.82	2.06	2.14	2.60	1.50	2.01
CaO (%)	1.61	0.30	0.49	0.50	0.65	0.44	0.66
MgO (%)	0.84	0.91	0.54	0.95	0.98	0.57	0.80
Na_2O (%)	0.91	0.92	0.91	0.80	1.11	0.64	0.88
TiO_2 (%)	0.78	0.81	0.84	0.83	0.80	0.74	0.80
P_2O_5 (%)	0.13	0.09	0.08	0.09	0.12	0.05	0.09
MnO_2 (%)	0.13	0.06	0.16	0.08	0.14	0.07	0.10
SO_3 (%)	0.01	0.01	0.01	0.01	0.01	0.01	0.01
Barium	521	511	633	633	618	402	553
Chromium	13	10	26	68	47	51	36
Copper	22	20	24	31	30	22	25
Mercury (µg/kg)	30	30	36	33	41	32	34
Nickel	<5	10	10	11	15	12	11
Lead	19	21	18	23	20	20	20
Rubidium	55	58	70	95	76	78	72
Strontium	121	95	115	106	135	96	111
Vanadium	70	69	83	109	91	92	86
Zinc	55	57	60	91	80	58	67
Zirconium	456	294	393	317	338	373	362
Total C (%)	1.99	1.12	0.50	0.33	0.22	0.17	0.72
Inorganic C (%)	0.34	0.05	0.04	0.03	0.04	0.03	0.09
Organic C (%)	1.65	1.07	0.46	0.30	0.18	0.14	0.63
pH	6.88	6.80	6.67	6.02	6.20	6.84	6.57

Table 34. Elemental composition of samples selected from Core 24

Subsample	24-01	24-02	24-03	24-04	24-10	24-16	Average
Lab. No.	R22049	R22050	R22051	R22052	R22053	R22054	
Depth Interval (ft)	0.0-0.6	0.6-1.2	1.2-1.45	1.45-2.0	6.05-7.0	9.35-9.9	
Horizon	A	A	B	B	C	2B	
SiO_2 (%)	81.73	79.83	74.58	73.50	80.94	81.93	78.75
Al_2O_3 (%)	7.27	9.12	11.65	12.20	8.97	9.07	9.71
Fe_2O_3 (%)	2.50	3.14	4.83	5.14	3.29	3.28	3.70
K_2O (%)	1.88	2.03	2.10	2.15	1.45	1.38	1.83
CaO (%)	0.38	0.36	0.45	0.43	0.34	0.23	0.37
MgO (%)	0.32	0.45	0.80	0.89	0.46	0.40	0.55
Na_2O (%)	0.78	0.75	0.70	0.83	0.63	0.43	0.69
TiO_2 (%)	0.88	0.91	0.89	0.89	0.72	0.75	0.84
P_2O_5 (%)	0.11	0.07	0.08	0.09	0.06	0.06	0.08
MnO_2 (%)	0.16	0.10	0.05	0.05	0.03	0.04	0.07
SO_3 (%)	0.01	0.01	0.01	0.01	0.01	0.01	0.01
Barium	584	581	638	556	382	324	511
Chromium	ND	28	56	69	39	40	46
Copper	ND	24	30	33	27	22	27
Mercury (µg/kg)	34	34	54	78	10	27	40
Nickel	ND	6	7	7	21	9	10
Lead	ND	19	23	25	19	16	20
Rubidium	ND	73	96	97	74	68	82
Strontium	96	97	95	97	95	76	93
Vanadium	ND	80	106	109	86	78	92
Zinc	ND	55	78	83	62	48	65
Zirconium	489	426	343	338	343	362	383
Total C (%)	1.13	0.55	0.39	0.32	0.18	0.18	0.46
Inorganic C (%)	0.04	0.04	0.04	0.04	0.03	0.03	0.04
Organic C (%)	1.09	0.51	0.35	0.28	0.15	0.15	0.42
pH	5.42	6.02	6.00	5.48	5.72	6.42	5.84

Table 35. Elemental composition of samples selected from Core 25

Subsample	25-01	25-02	25-03	25-04	25-22	25-29	Average
Lab. No.	R22055	R22056	R22057	R22058	R22059	R22060	
Depth Interval (ft)	0.0-0.7	0.7-1.0	1.0-2.0	2.0-2.9	8.4-9.0	14.3-15.7	
Horizon	A	C	C	C	2C	2C	
SiO_2 (%)	78.41	78.55	77.41	75.46	68.65	67.01	74.25
Al_2O_3 (%)	9.85	10.07	10.70	11.88	14.05	16.02	12.09
Fe_2O_3 (%)	3.22	3.25	3.66	3.99	5.66	6.12	4.32
K_2O (%)	2.08	2.07	2.10	2.15	2.70	2.54	2.27
CaO (%)	0.37	0.29	0.29	0.31	0.93	0.45	0.44
MgO (%)	0.43	0.41	0.54	0.64	1.28	1.05	0.73
Na_2O (%)	1.02	1.03	1.02	0.96	0.88	0.78	0.95
TiO_2 (%)	0.91	0.91	0.90	0.89	0.84	0.89	0.89
P_2O_5 (%)	0.13	0.11	0.10	0.10	0.13	0.07	0.11
MnO_2 (%)	0.15	0.10	0.10	0.07	0.19	0.12	0.12
SO_3 (%)	0.01	0.01	0.01	0.01	0.01	0.01	0.01
Barium	519	501	555	515	733	763	598
Chromium	45	ND	59	50	71	85	62
Copper	19	ND	25	27	27	27	25
Mercury (µg/kg)	39	34	21	12	35	22	27
Nickel	ND	ND	<5	8	15	12	10
Lead	18	ND	19	20	18	19	19
Rubidium	71	ND	84	97	96	122	94
Strontium	115	116	117	114	141	130	122
Vanadium	87	ND	93	97	116	127	104
Zinc	50	ND	57	71	79	105	72
Zirconium	365	366	361	334	254	223	317
Total C (%)	0.52	0.26	0.20	0.20	0.32	0.20	0.28
Inorganic C (%)	0.05	0.03	0.04	0.04	0.10	0.04	0.05
Organic C (%)	0.47	0.23	0.16	0.16	0.22	0.16	0.23
pH	4.19	4.05	3.98	4.00	7.42	7.34	5.16

Table 36. Elemental composition of samples selected from Core 26

Subsample	26-01	26-02	26-03	26-04	26-08	26-15	Average
Lab. No.	R22061	R22062	R22063	R22064	R22065	R22066	
Depth Interval (ft)	0.0-0.65	0.65-1.6	1.6-1.9	1.9-2.8	5.0-5.7	8.0-9.1	
Horizon	A	A	E	B	2C	2D	
SiO_2 (%)	83.06	83.62	82.68	83.64	75.66	60.20	78.14
Al_2O_3 (%)	6.76	7.36	7.91	7.27	11.89	19.94	10.19
Fe_2O_3 (%)	1.74	2.22	2.69	2.79	3.75	6.07	3.21
K_2O (%)	1.56	1.50	1.53	1.49	1.58	3.33	1.83
CaO (%)	0.75	0.39	0.33	0.30	0.36	0.33	0.41
MgO (%)	0.28	0.27	0.29	0.24	0.59	1.36	0.51
Na_2O (%)	0.92	0.85	0.86	0.83	0.71	0.66	0.81
TiO_2 (%)	0.76	0.74	0.74	0.68	0.72	1.02	0.78
P_2O_5 (%)	0.13	0.06	0.06	0.07	0.07	0.14	0.09
MnO_2 (%)	0.14	0.08	0.07	0.08	0.02	0.04	0.07
SO_3 (%)	0.01	0.01	0.01	0.01	0.01	0.01	0.01
Barium	457	445	410	403	547	798	510
Chromium	<5	9	17	15	79	122	41
Copper	19	20	20	22	31	49	27
Mercury (µg/kg)	28	23	19	27	8	16	20
Nickel	7	<5	<5	<5	20	27	12
Lead	25	16	19	18	19	20	20
Rubidium	49	54	59	54	102	191	85
Strontium	107	97	98	88	105	141	106
Vanadium	58	62	73	72	102	149	86
Zinc	43	40	45	38	79	139	64
Zirconium	437	455	439	412	278	188	368
Total C (%)	1.01	0.33	0.28	0.24	0.17	0.57	0.43
Inorganic C (%)	0.07	0.03	0.03	0.04	0.03	0.04	0.04
Organic C (%)	0.94	0.30	0.25	0.20	0.14	0.53	0.39
pH	6.74	5.05	4.38	4.36	4.12	6.80	5.24

Table 37. Means and ranges of elemental contents of the uppermost samples from Illinois soil cores 11 through 26 compared with results for surface soils as determined by other researchers

	This Work		Shacklette and Boerngen(2)	
Element	Mean	Range	Mean	Range
Silicon (%)	36.39	31.0-39.5	NR†	29-45**
Aluminum (%)	4.37	3.18-6.47	NR	<0.07-8.5**
Iron (%)	1.94	0.94-3.24	2.1(1)	0.1-4.0**
Potassium (%)	1.52	1.12-2.09	1.7(1)	0.22-2.25**
Calcium (%)	0.90	0.26-6.24	0.40(1)	0.3-1.5**
Magnesium (%)	0.33	0.13-0.71	0.30(1)	0.005-1.25**
Sodium (%)	0.70	0.56-0.91	NR	0.6-1.25**
Titanium (%)	0.44	0.24-0.55	0.41	0.05-1.0
Phosphorus (%)	0.060	0.04-0.14	0.065(1)	0.013-0.68**
Manganese (%)	0.11	0.03-0.27	0.052	0.005-0.15
Sulfur (%)	0.0046	<0.001-0.017	0.08(1)	<0.08-0.5**
Barium (mg/kg)	533	416-791	675	200-1500
Chromium (mg/kg)	26	<5-60	55	10-100
Copper (mg/kg)	23	18-32	25	7-100
Mercury (µg/kg)	30	<2-92	70***	20-360***
Nickel (mg/kg)	7	<5-21	17	5-30
Lead (mg/kg)	25	18-72	19	10-30
Rubidium (mg/kg)	63	41-100	75	45-100
Strontium (mg/kg)	120	92-169	305	20-1000
Vanadium (mg/kg)	75	45-110	87	20-150*
Zinc (mg/kg)	61	43-138	58.5	20-109
Zirconium (mg/kg)	383	216-489	NR	NR

†NR = not reported
***Values for soils on glacial till, U.S., Shacklette and Boerngen, 1984.
**Average concentration in Illinois soils; from Shacklette and Boerngen, 1984.
*values for loamy and clay soils, U.S.; from Shacklette and Boerngen, 1984.
(1)values for humid region soils; from Severson and Shacklette, 1988.

(2)values for loess and silty soils, U.S., unless noted otherwise; from Shacklette and Boerngen, 1984.

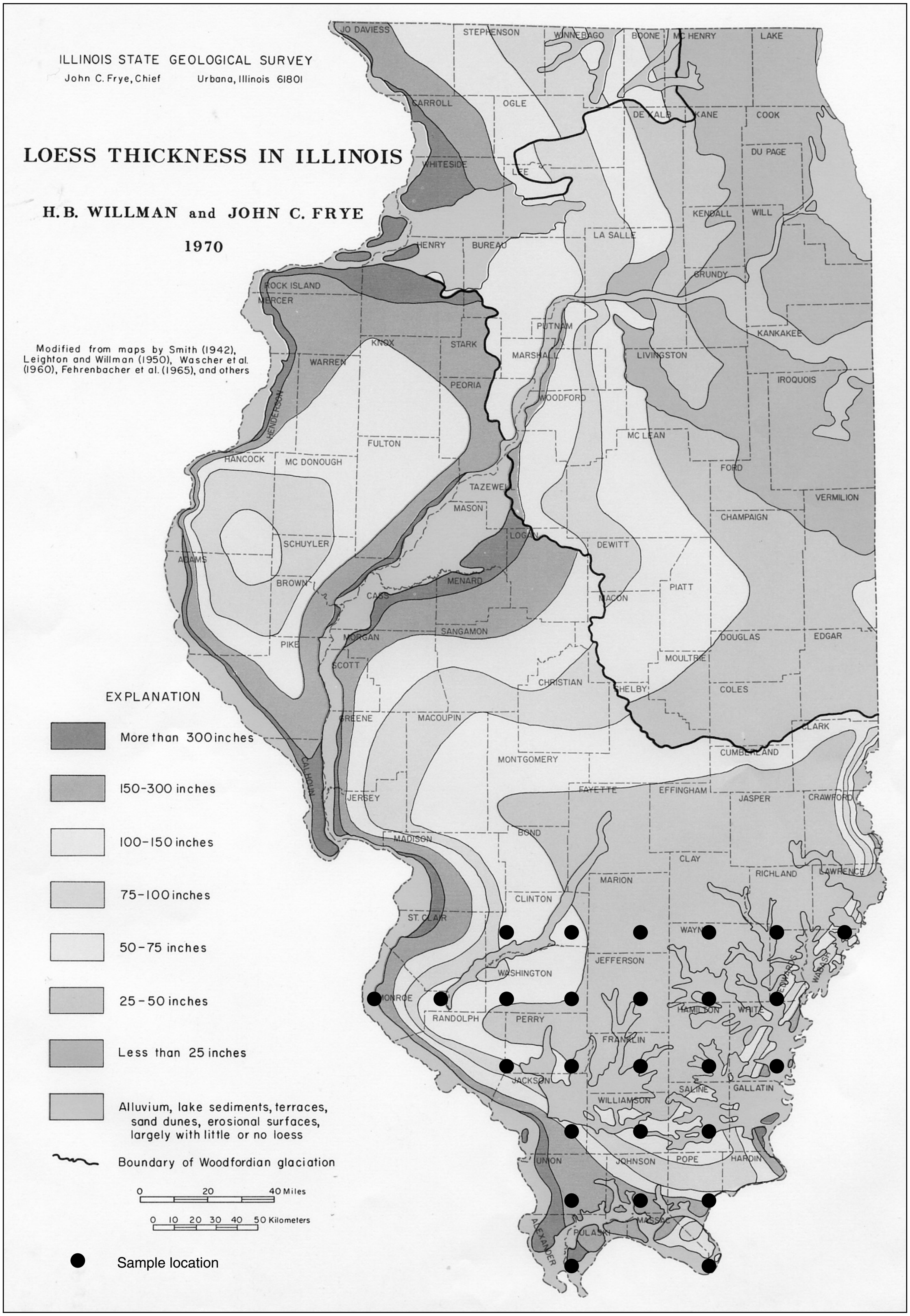


Figure 1 Sample locations related to loess thickness in Illinois (after Willman and Frye 1970).

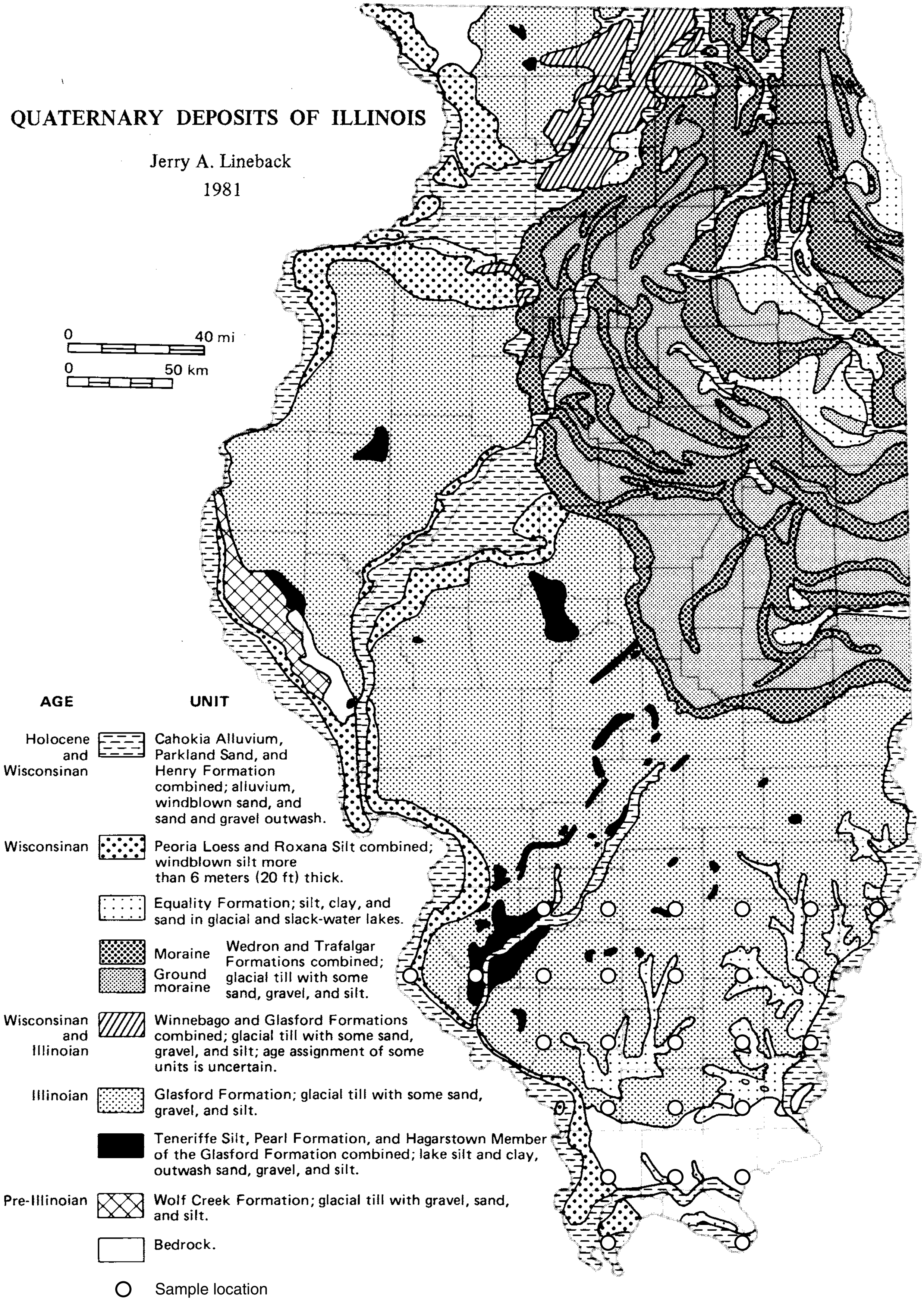
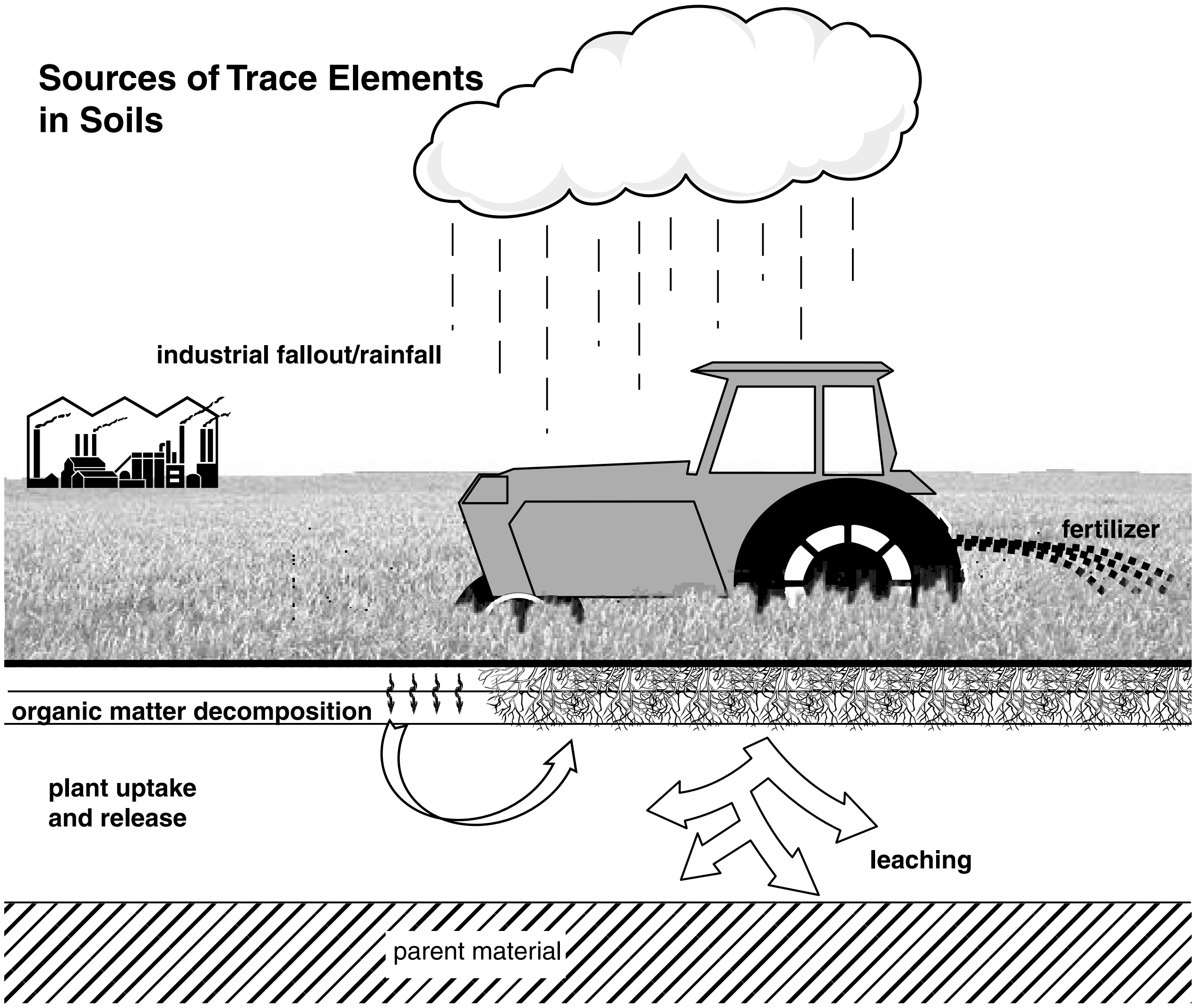


Figure 2 Sample locations related to Quaternary deposits in Illinois (after Lineback 1981).



Reactions
  - precipitation $\rightleftharpoons$ dissolution
  - adsorption $\rightleftharpoons$ desorption
  - fixation $\rightleftharpoons$ release (biota)
  - complexation $\rightleftharpoons$ release
  - gaseous phase $\rightleftharpoons$ soil (e.g. selenium)

variables
  - pH
  - Eh
  - kinetics
  - biological activity
    - decomposition
    - respiration
  - organics on clays, oxides

Figure 3 Schematic diagram of various input and ouput functions that affect trace elements, reactions that affect soil chemistry, and variables that affect reactions.

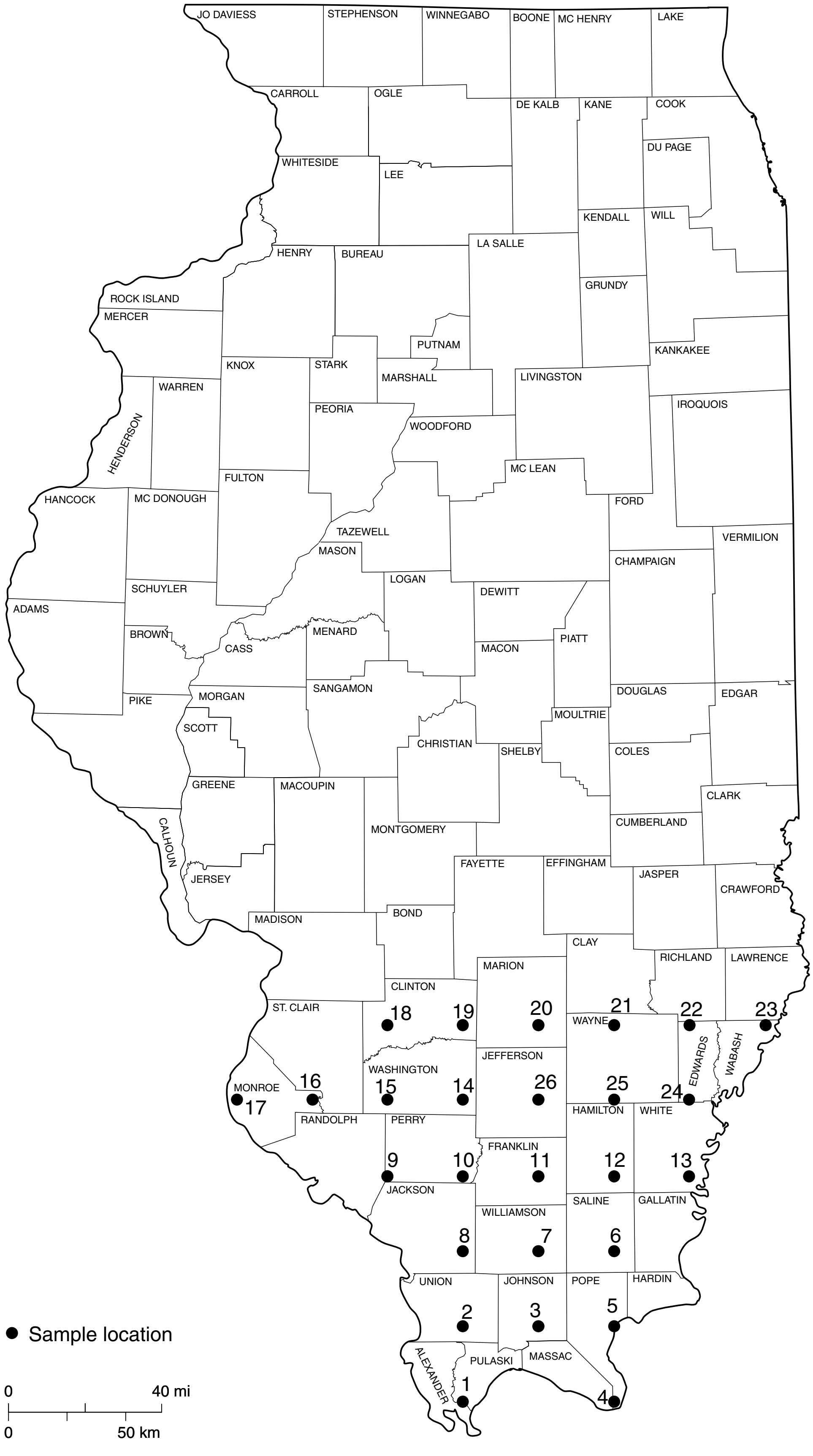


Figure 4 Locations of cores collected in 1998 at intersections of grid lines. Grid lines are 20 miles apart.

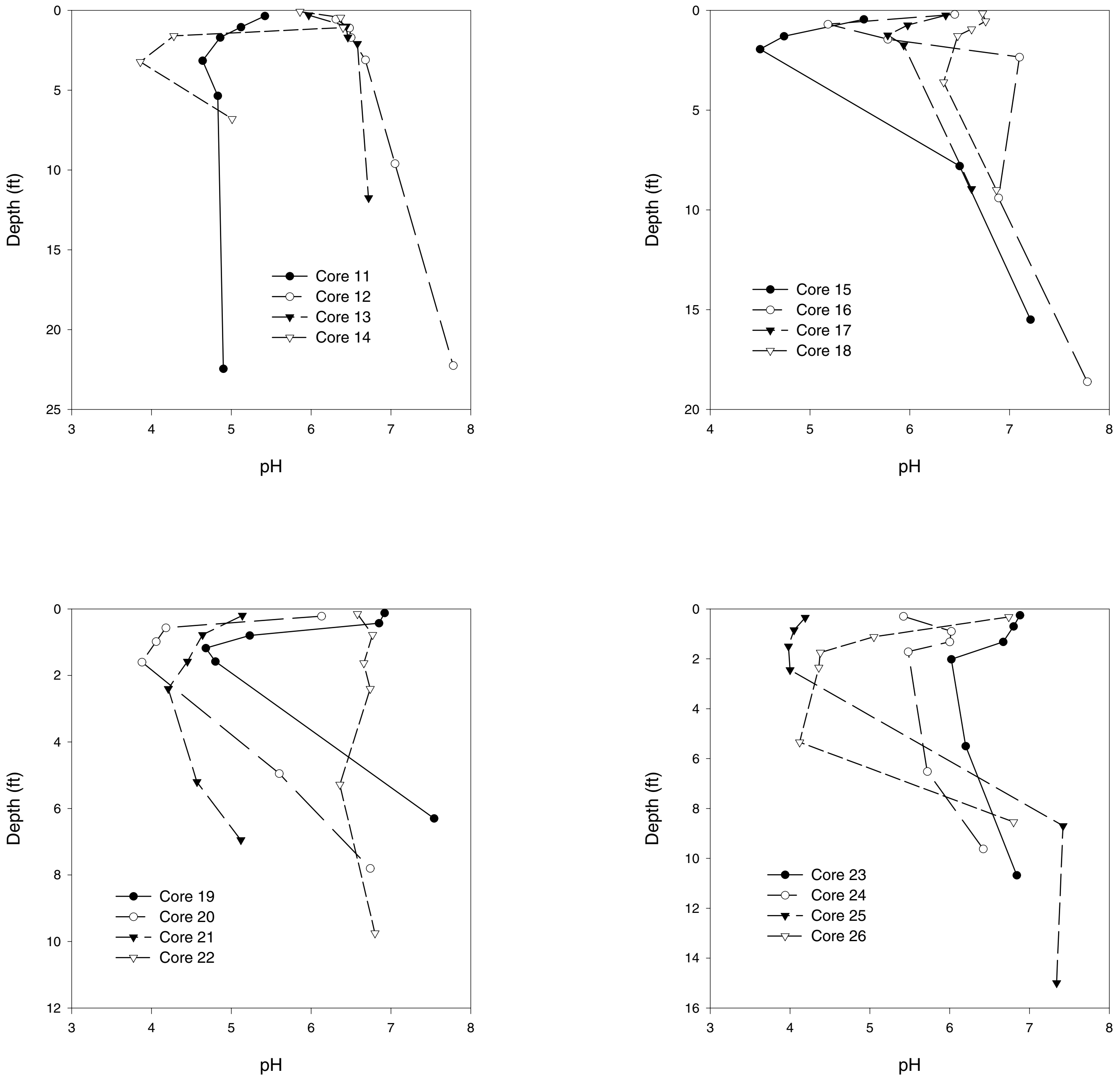


Figure 5. Soil pH versus depth.

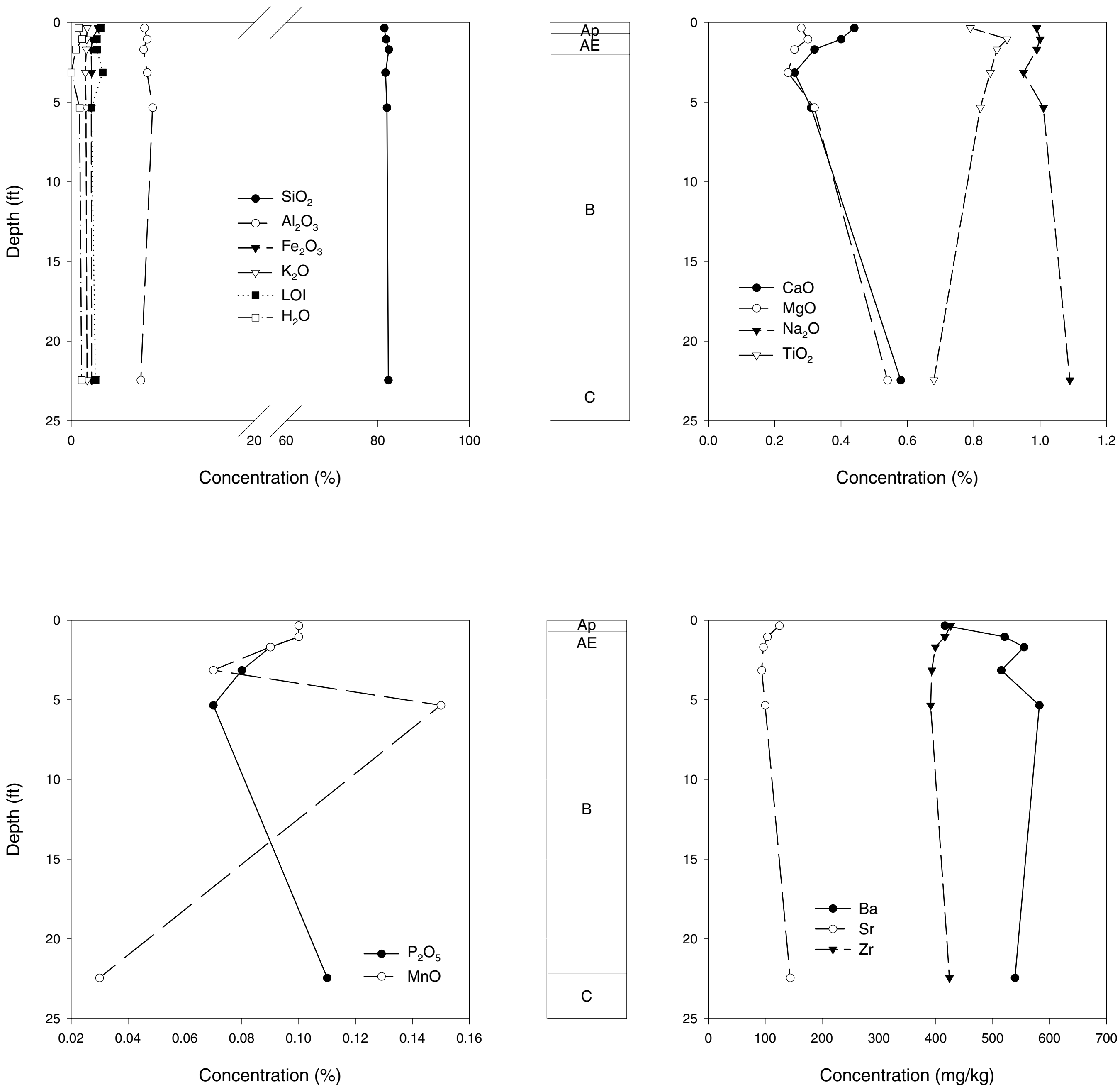

Core 11

Figure 6. Elemental concentrations versus depth in core 11.

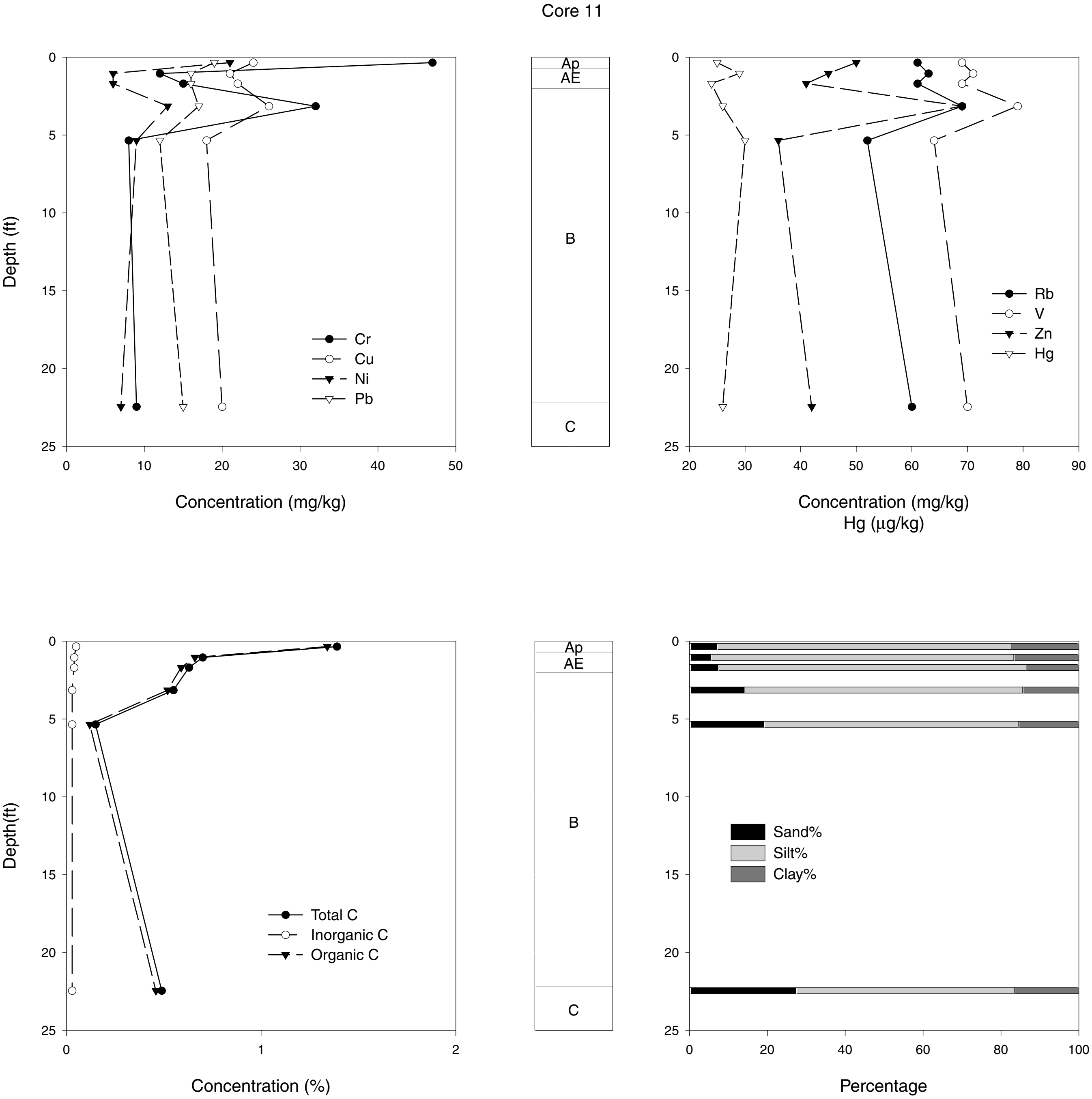


Figure 7. Elemental concentrations versus depth in core 11.

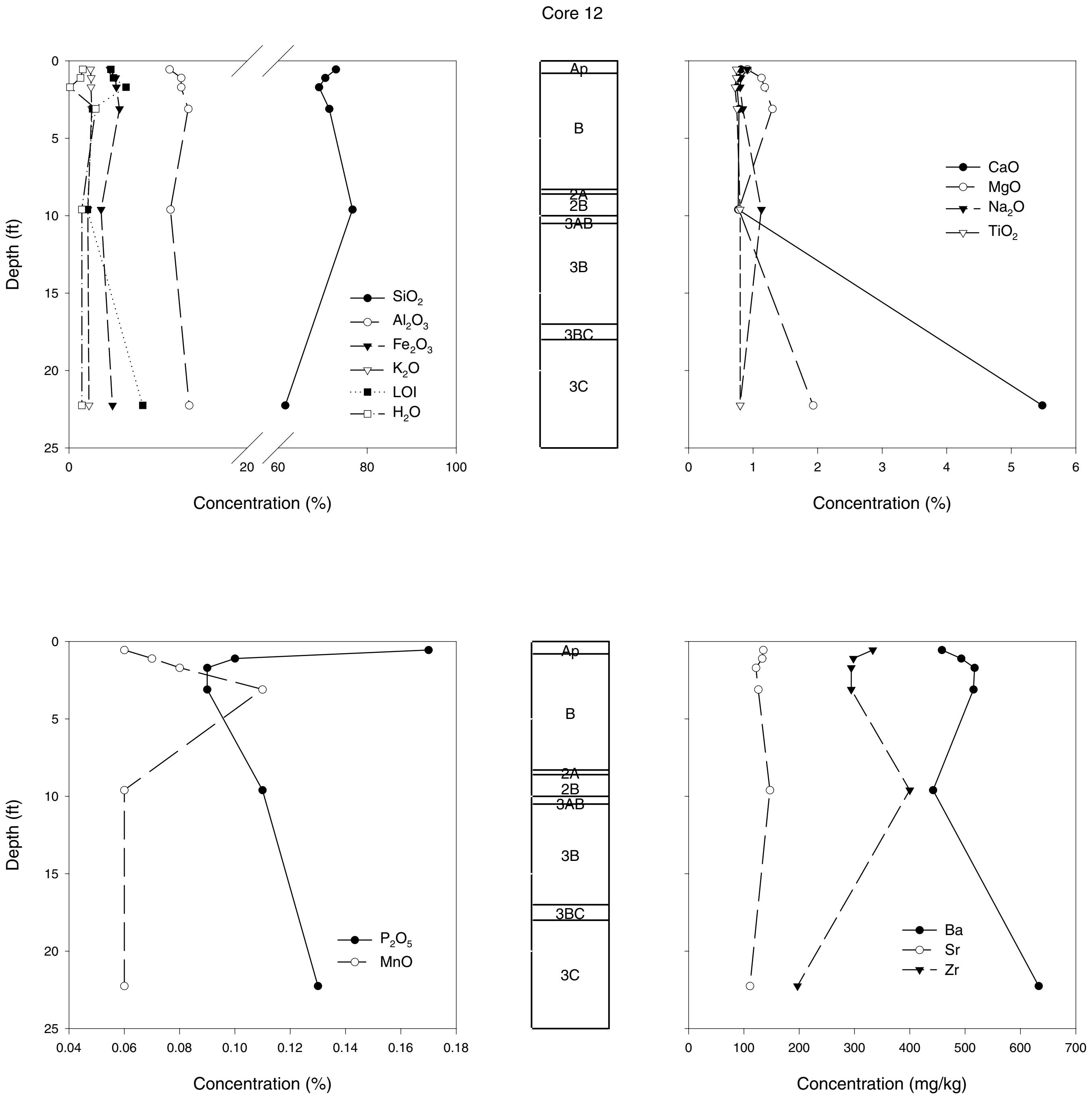


Figure 8. Elemental concentrations versus depth in core 12.

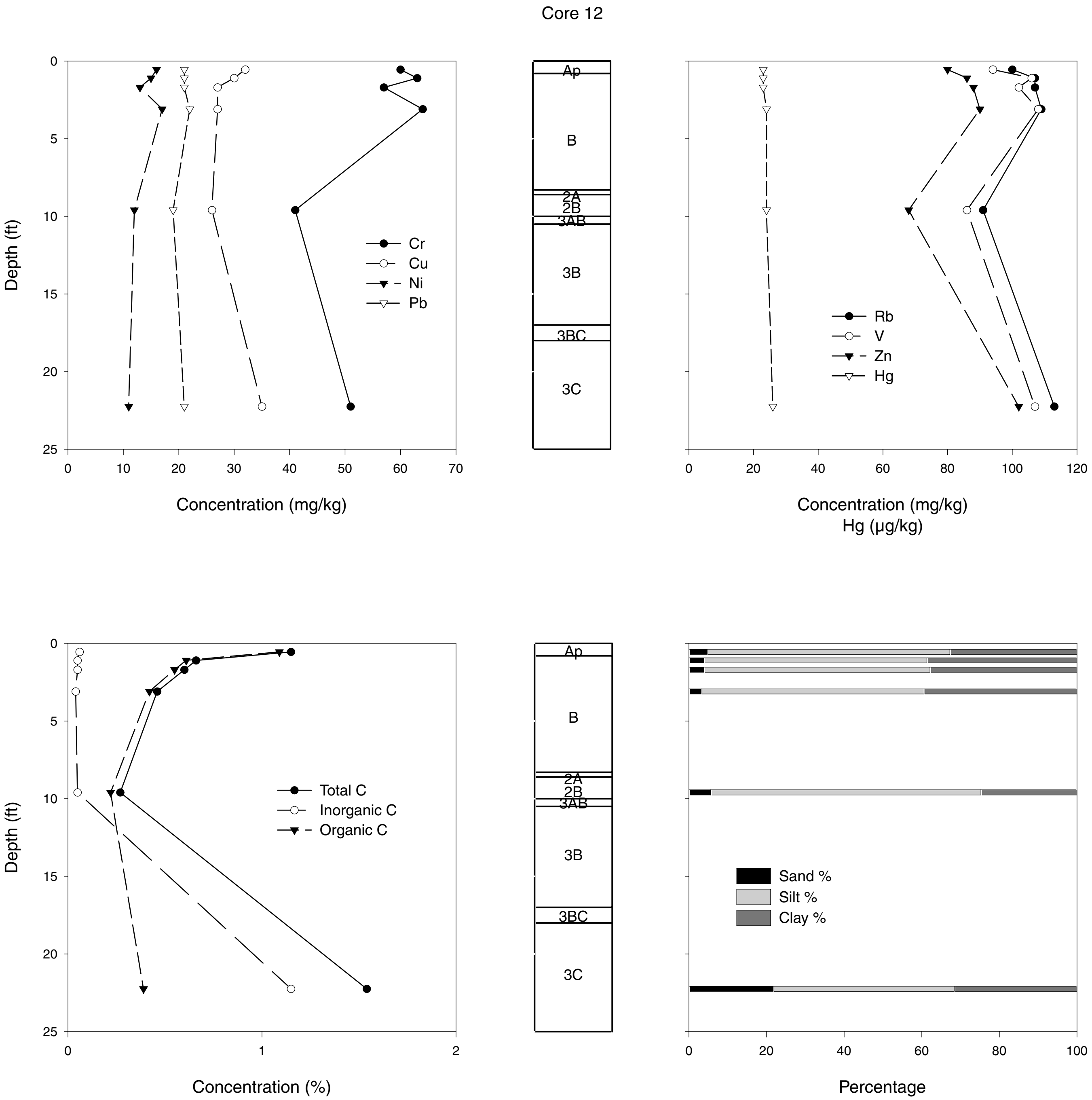


Figure 9. Elemental concentrations versus depth in core 12.

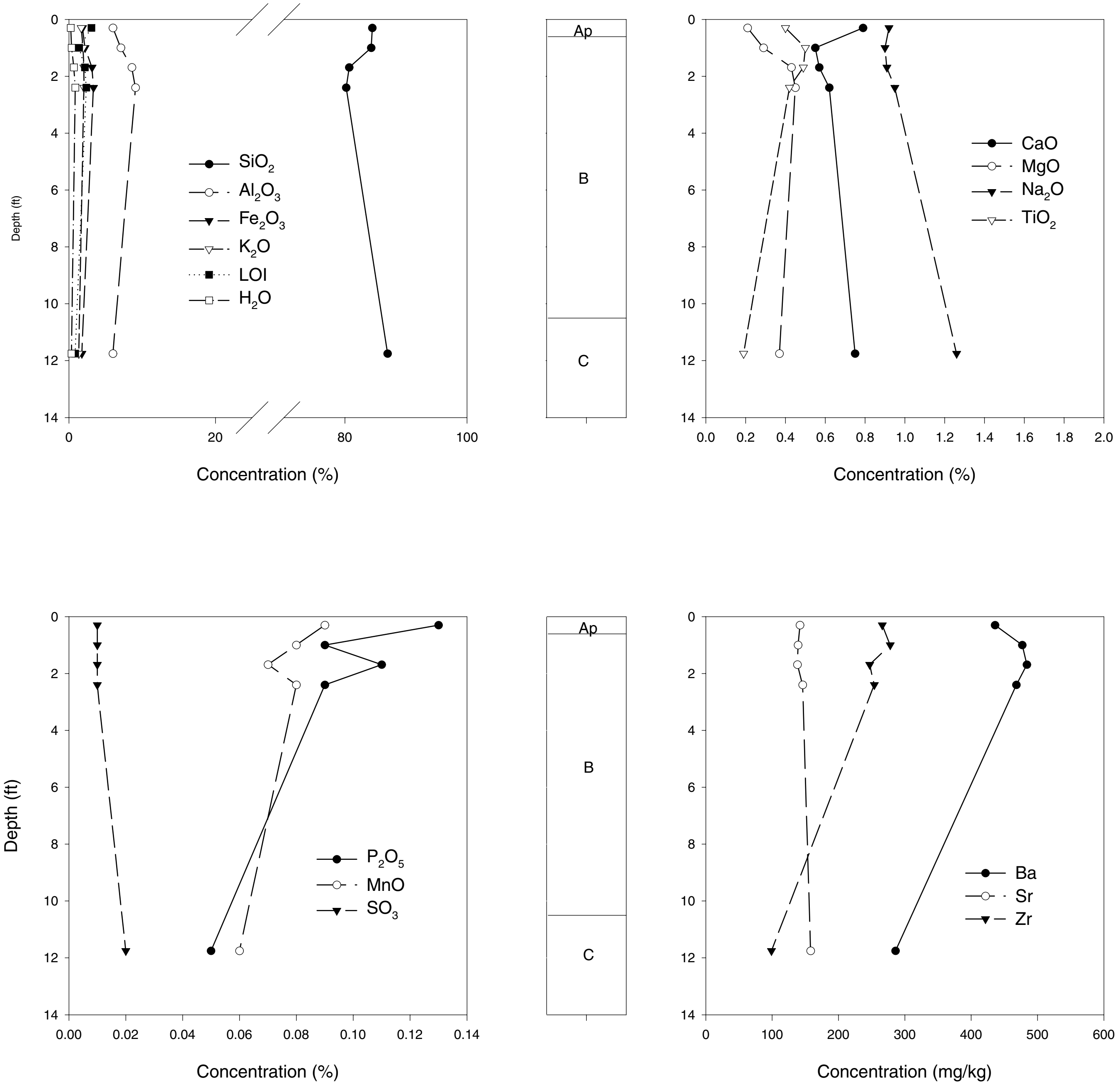
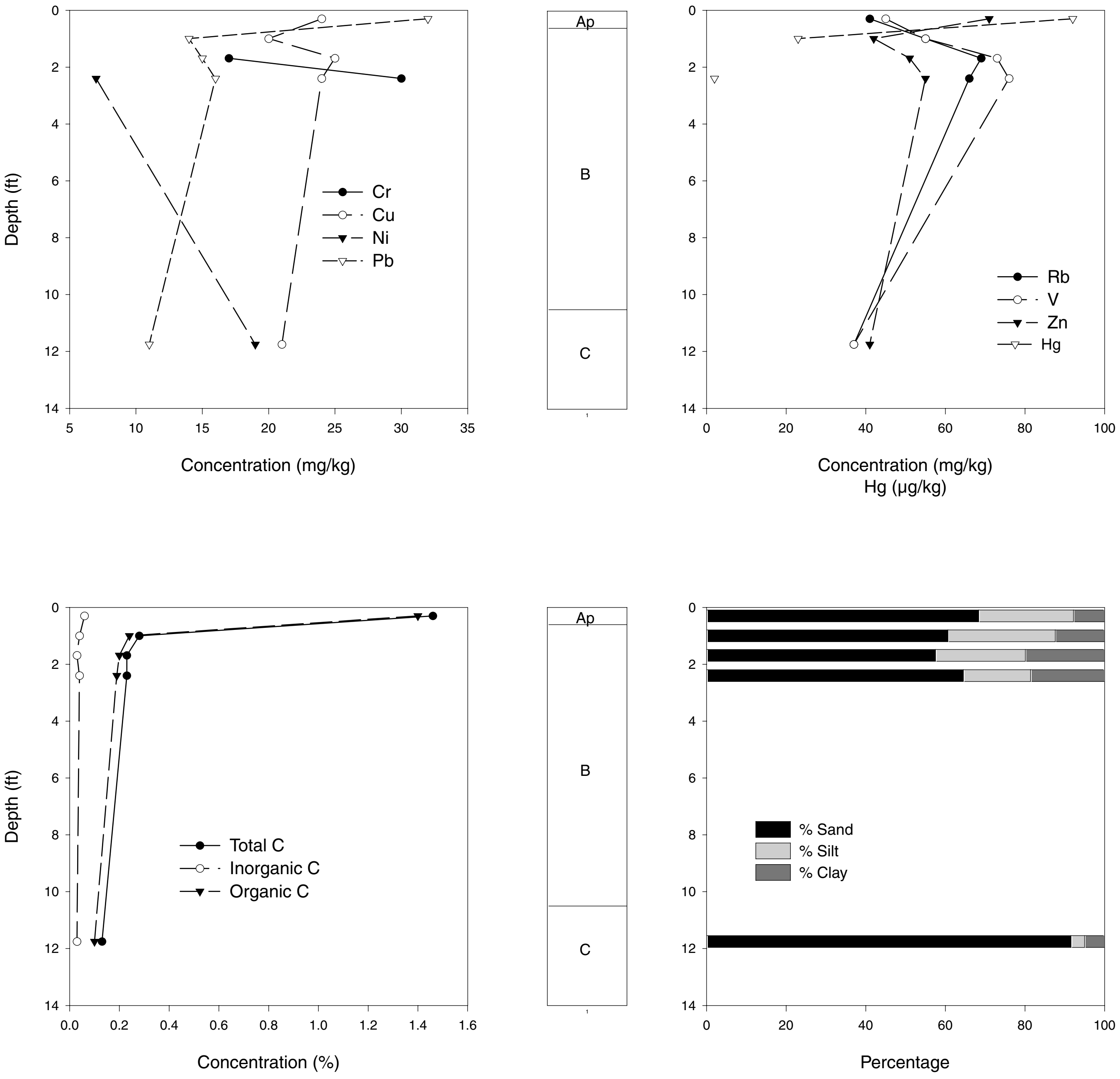

Core 13

Figure 10. Elemental concentrations versus depth core 13.



Core 13

Figure 11. Elemental concentrations versus depth in core 13.

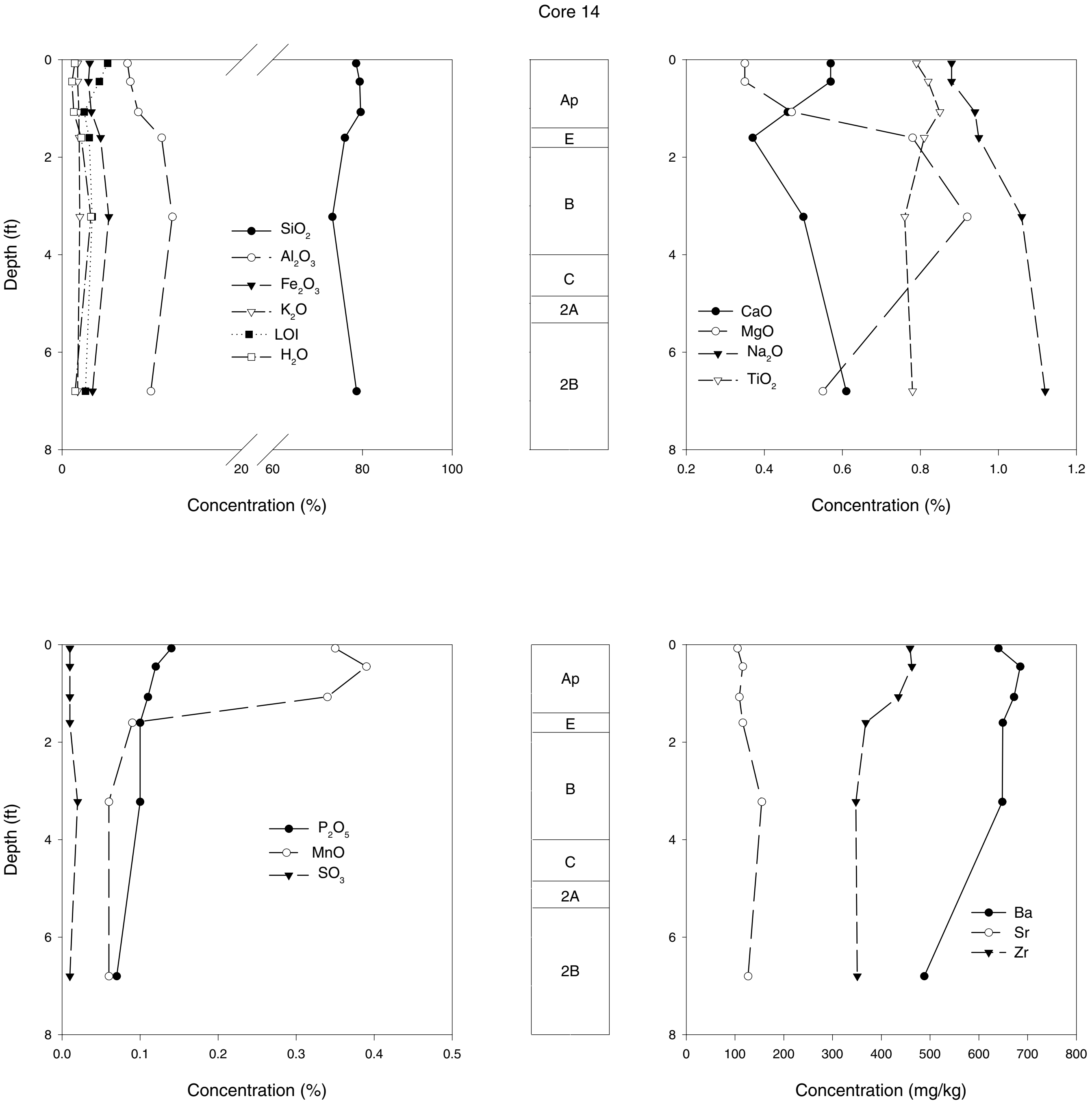
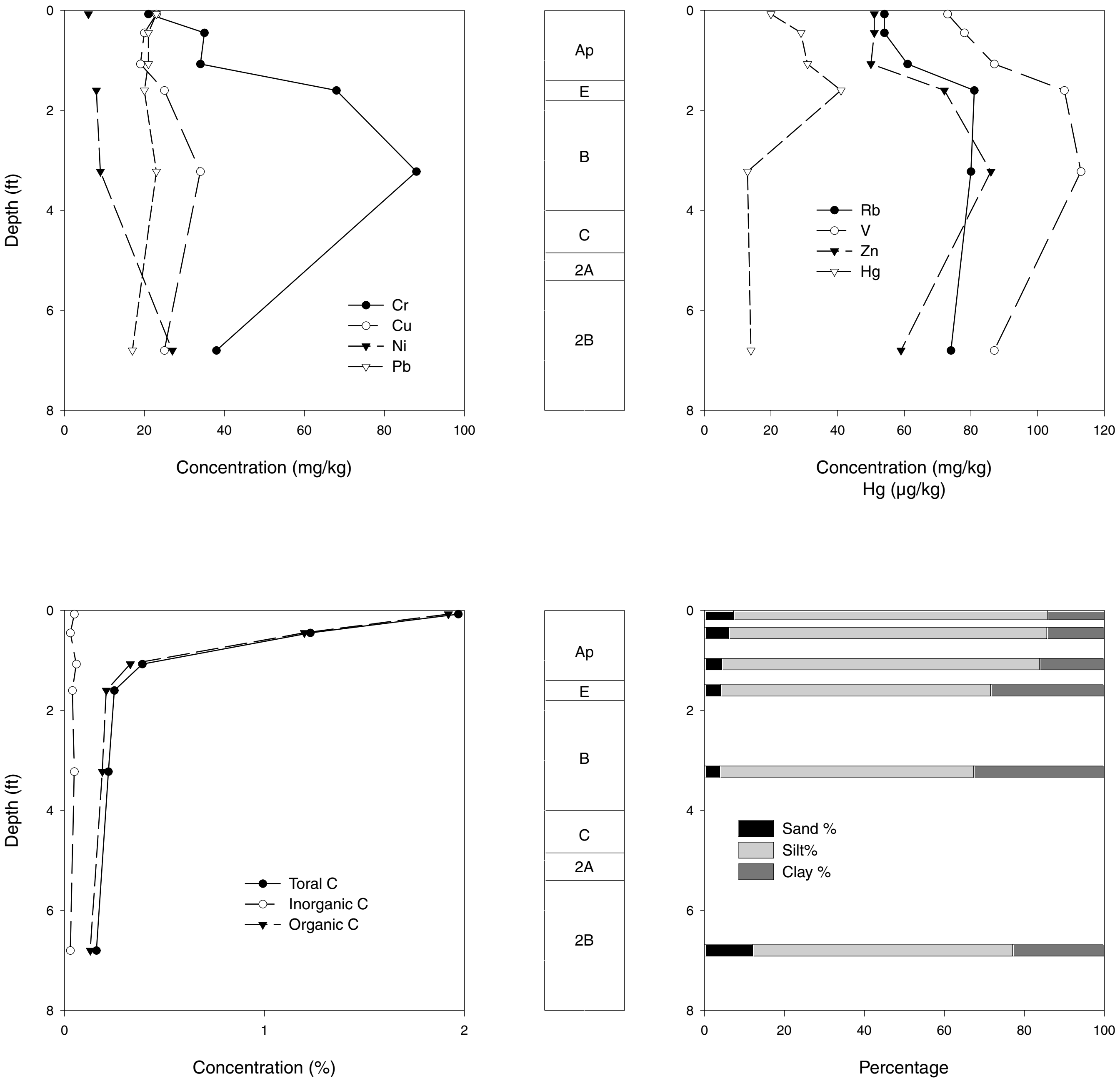


Figure 12. Elemental concentrations versus depth in core 14.



Core 14

Figure 13. Elemental concentrations versus depth in core 14.

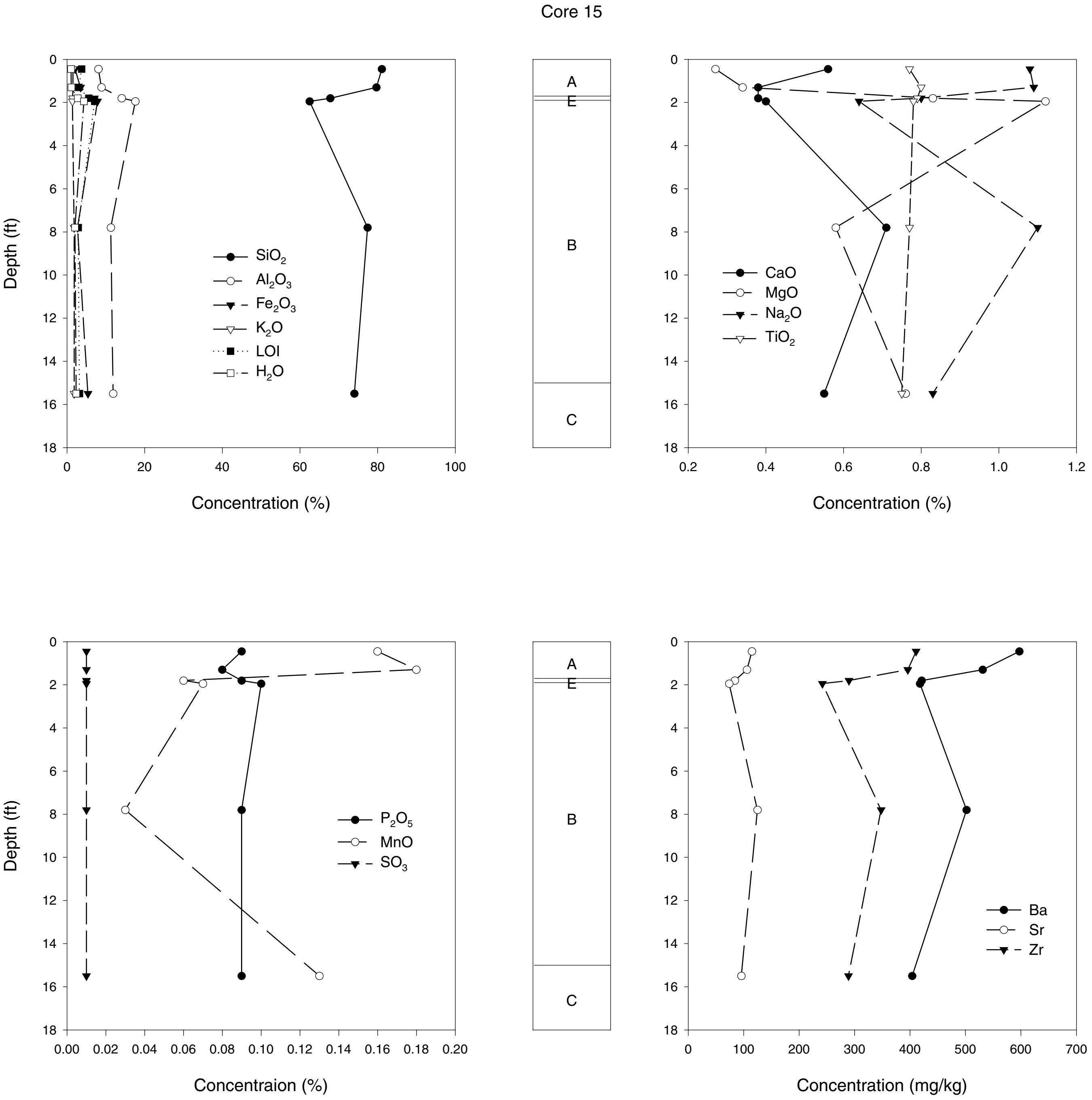


Figure 14. Elemental concentrations versus depth in core 15.

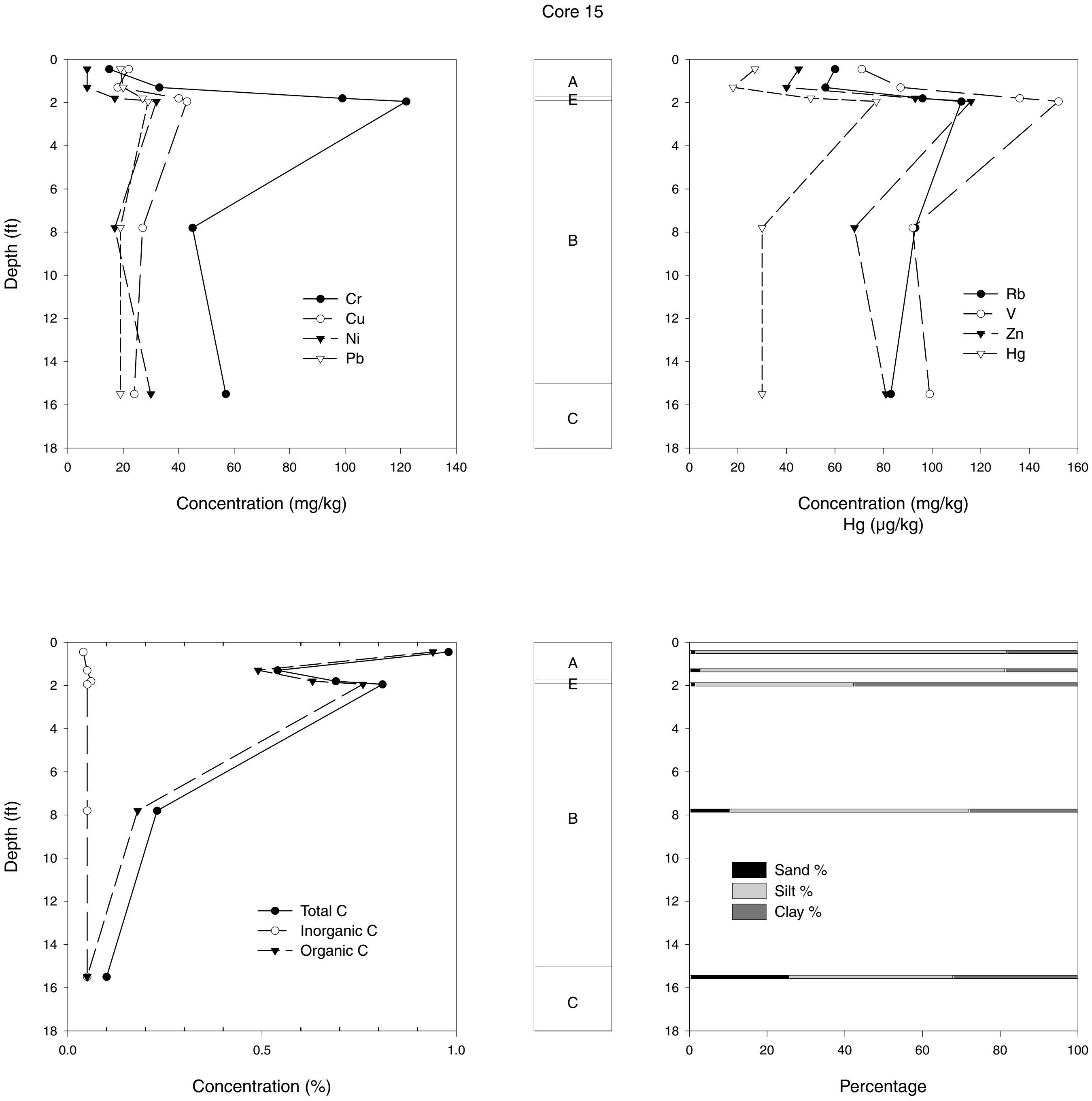


Figure 15. Elemental concentrations versus depth in core 15.

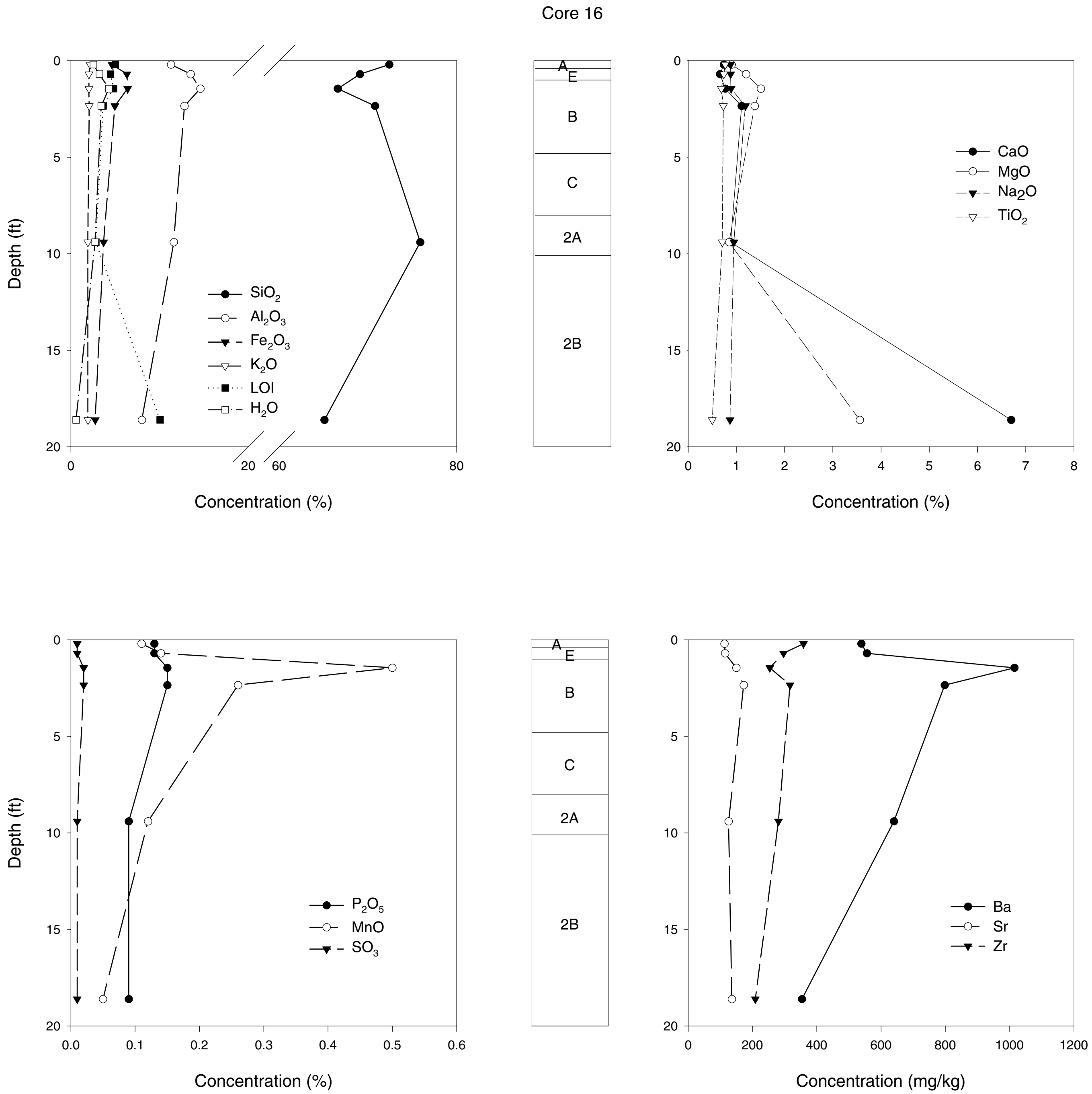


Figure 16. Elemental concentrations versus depth in core 16.

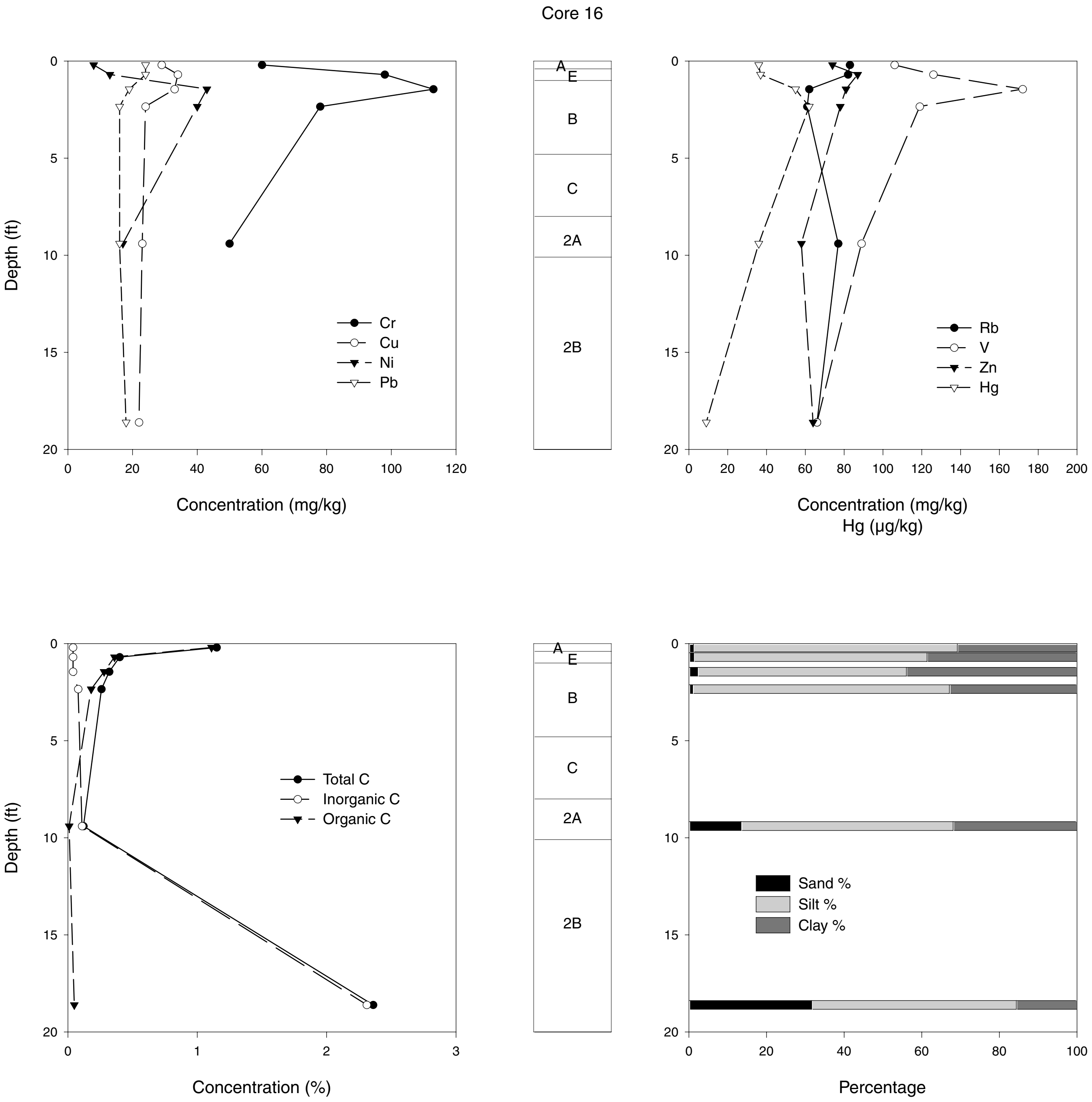


Figure 17. Elemental concentrations versus depth in core 16.

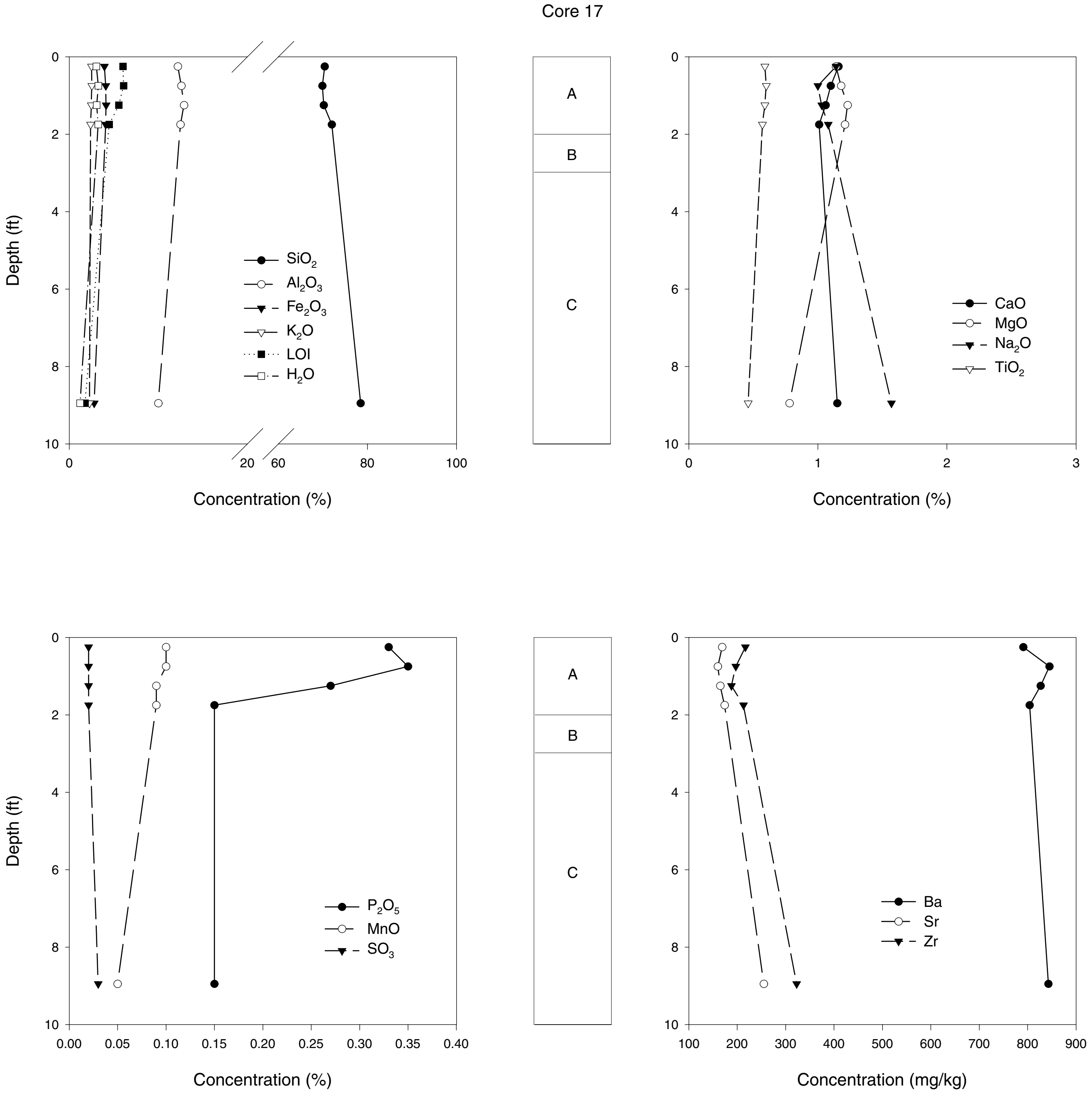


Figure 18. Elemental concentrations versus depth in core 17.

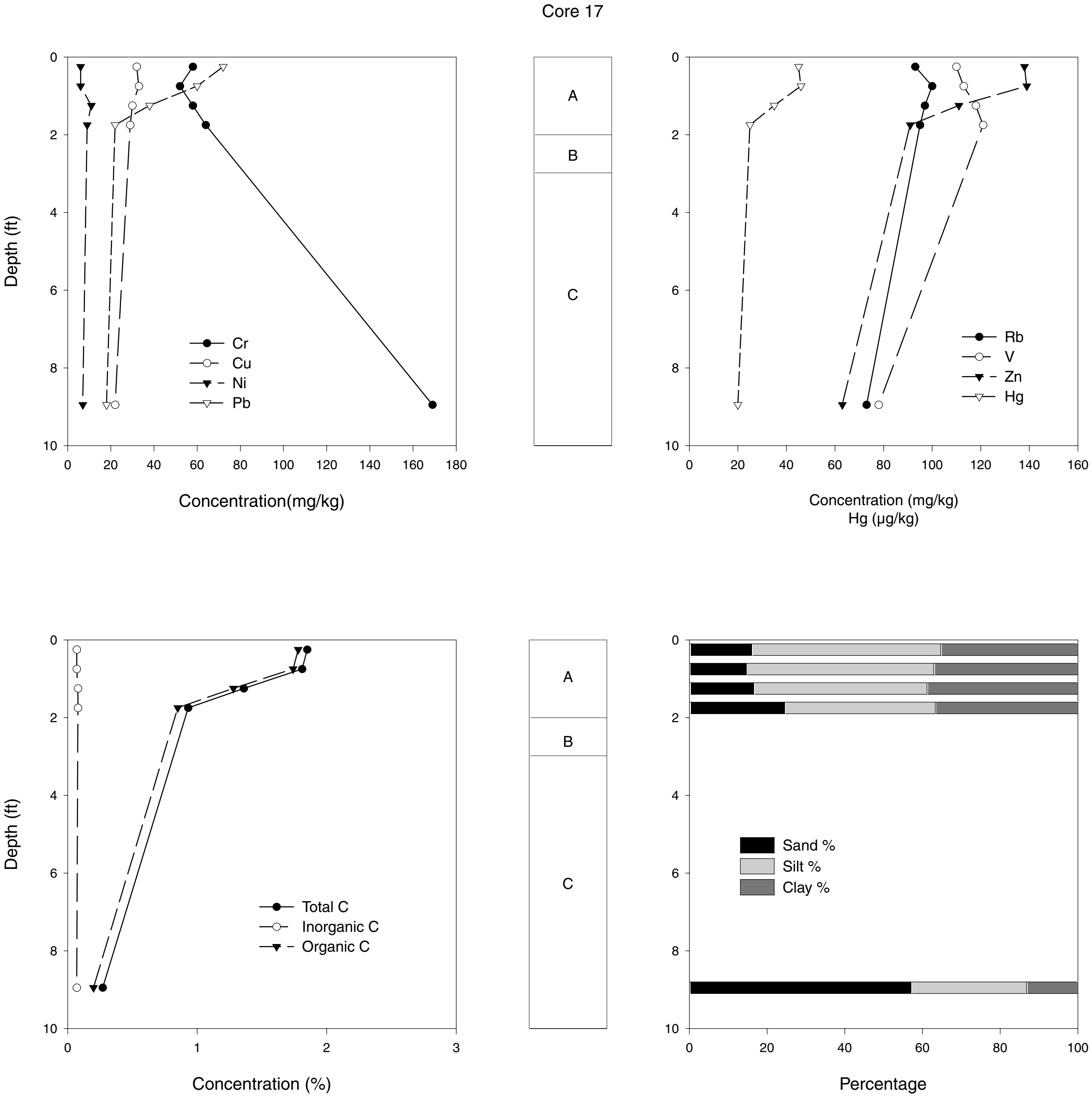


Figure 19. Elemental concentrations versus depth in core 17.

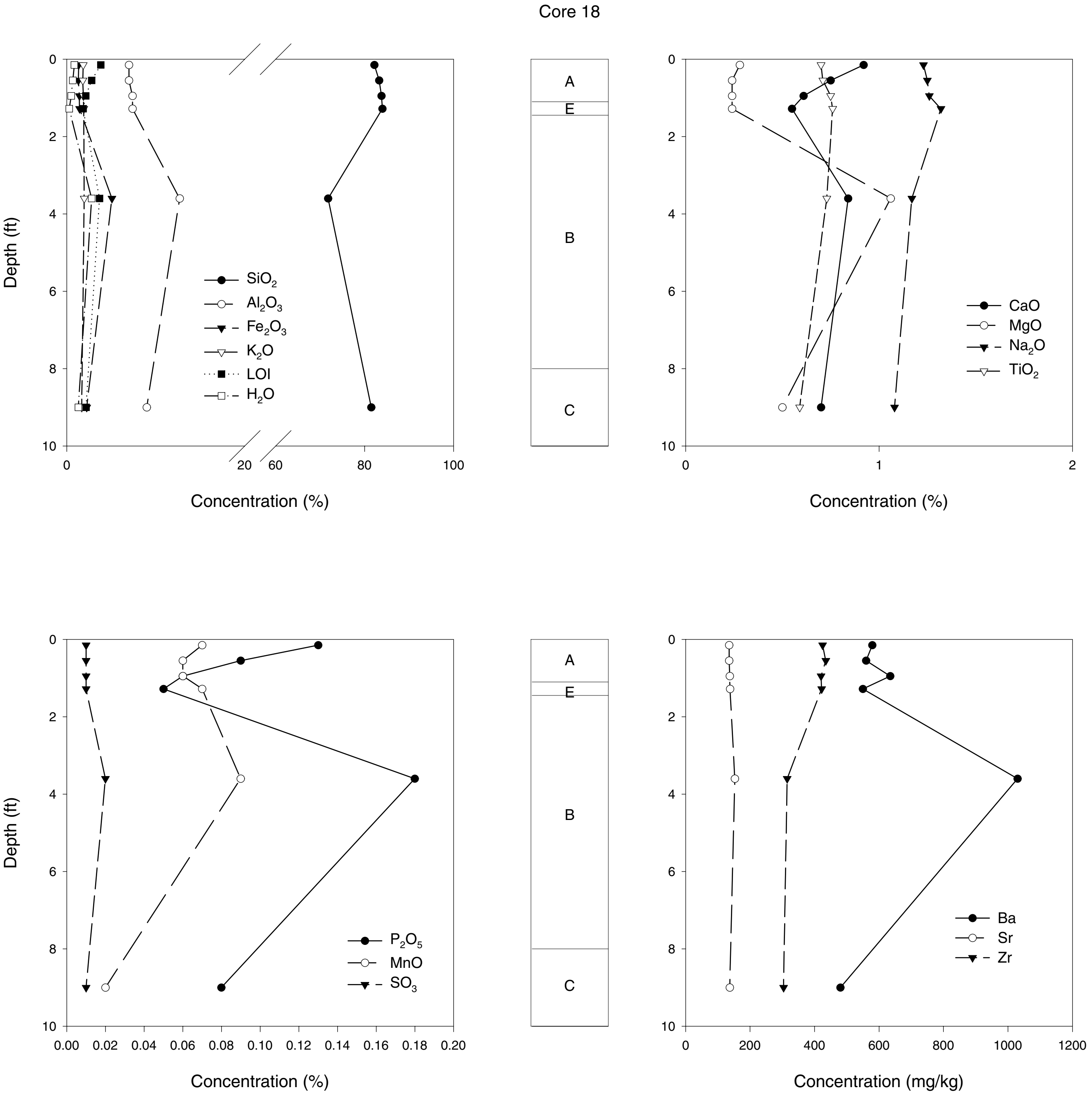


Figure 20. Elemental concentrations versus depth in core 18.

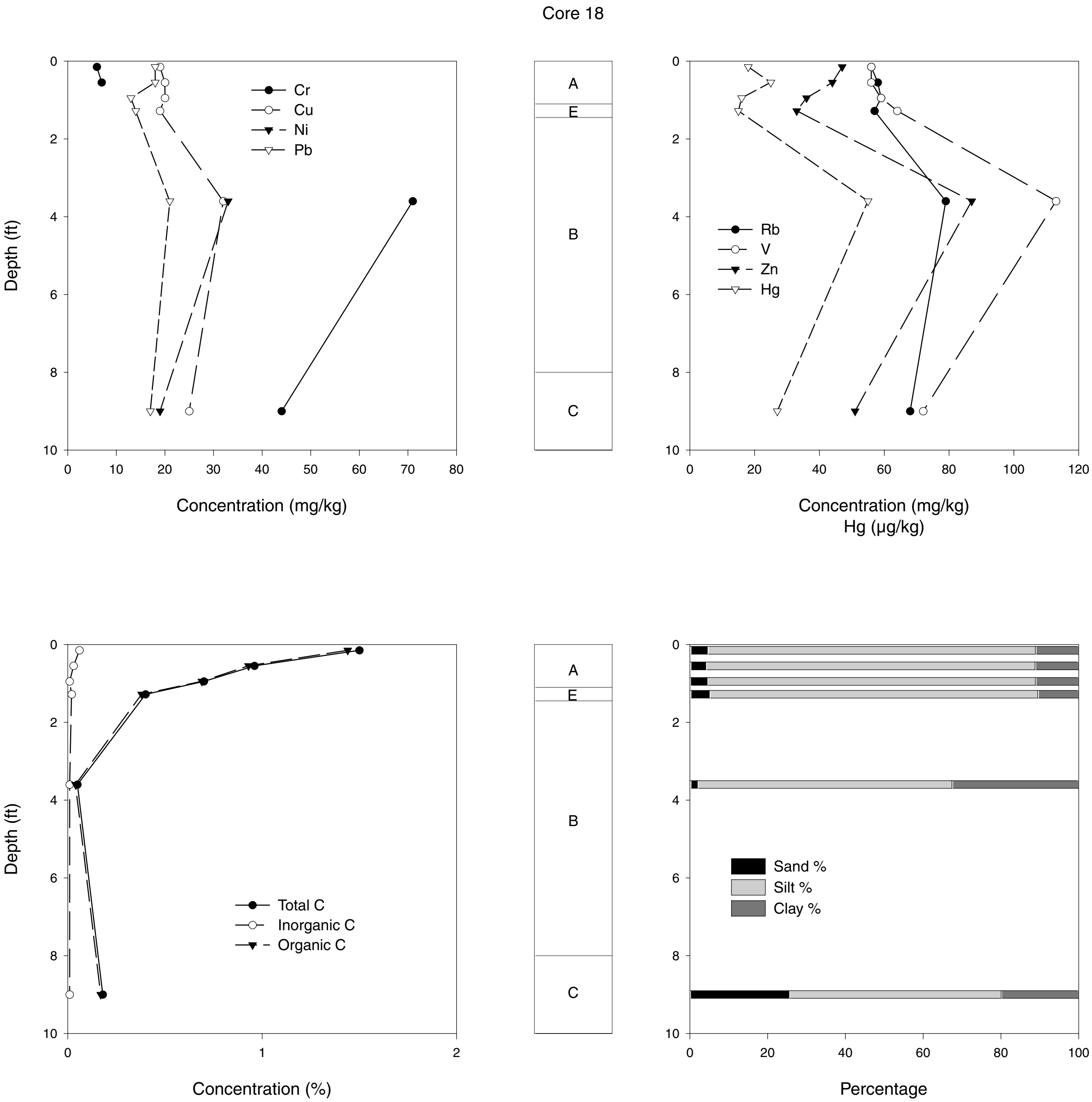


Figure 21. Elemental concentrations versus depth in core 18.

Core 19

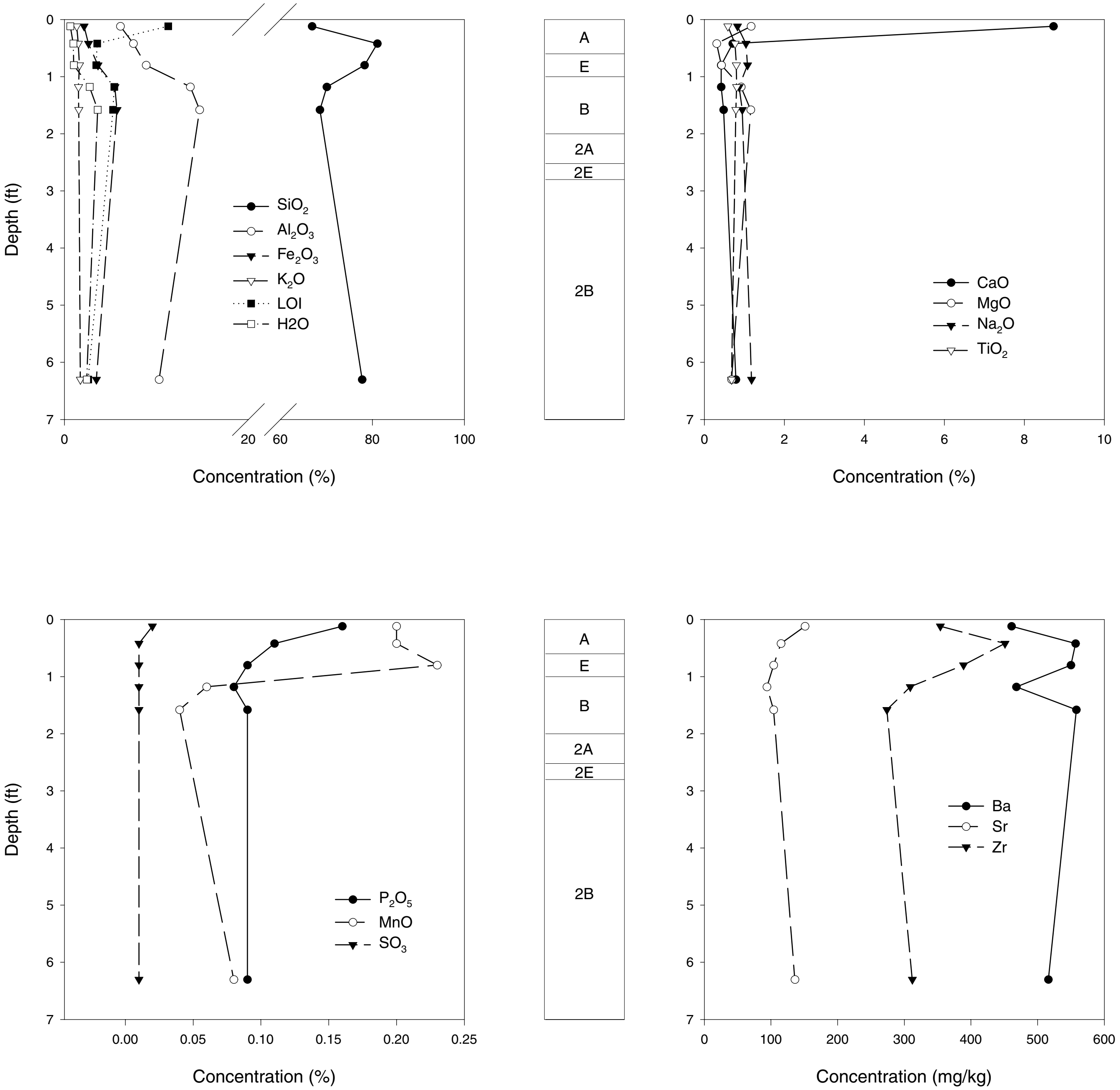


Figure 22. Elemental concentrations versus depth core 19.

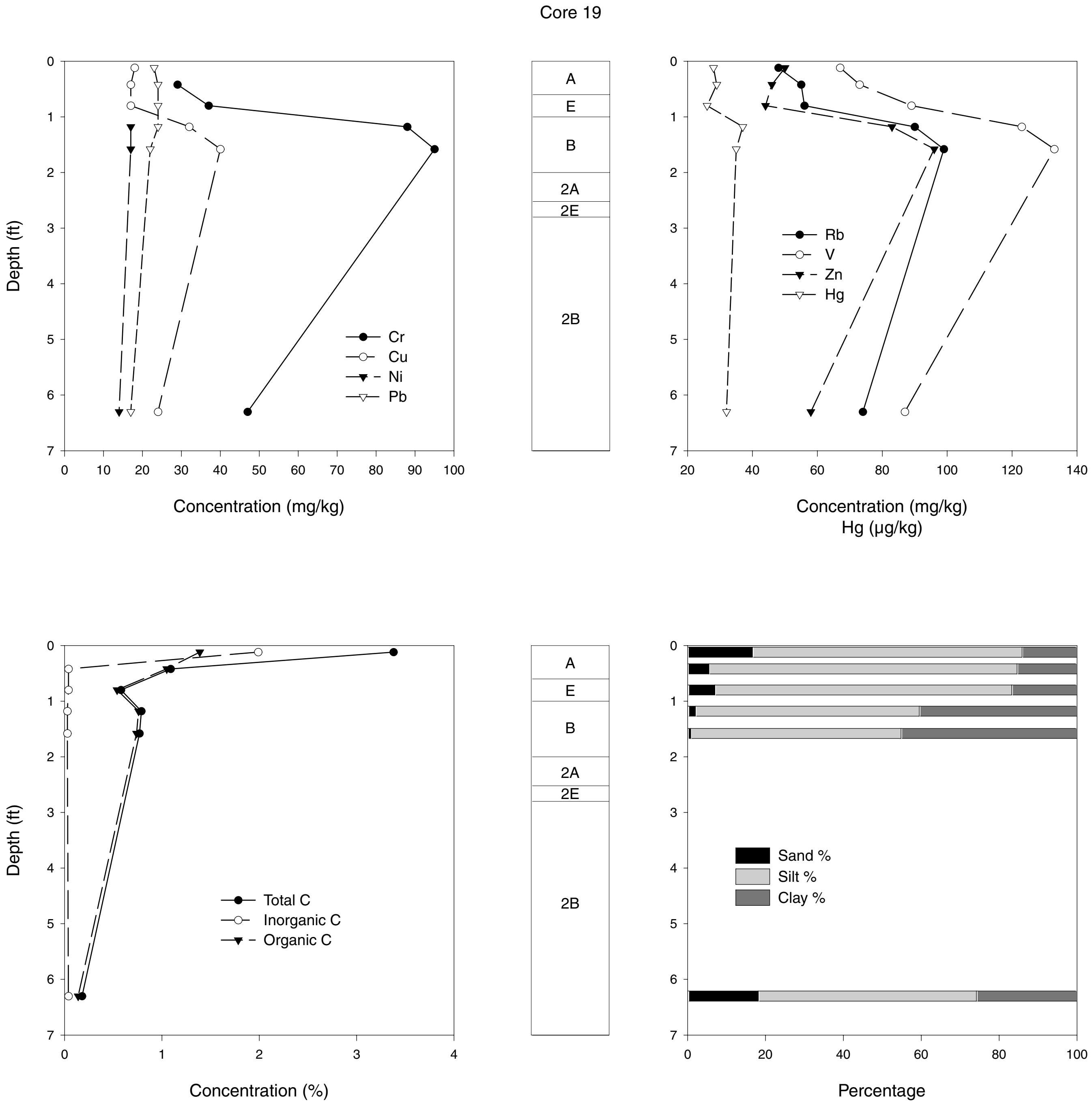


Figure 23. Elemental concentrations versus depth in core 19.

Core 20

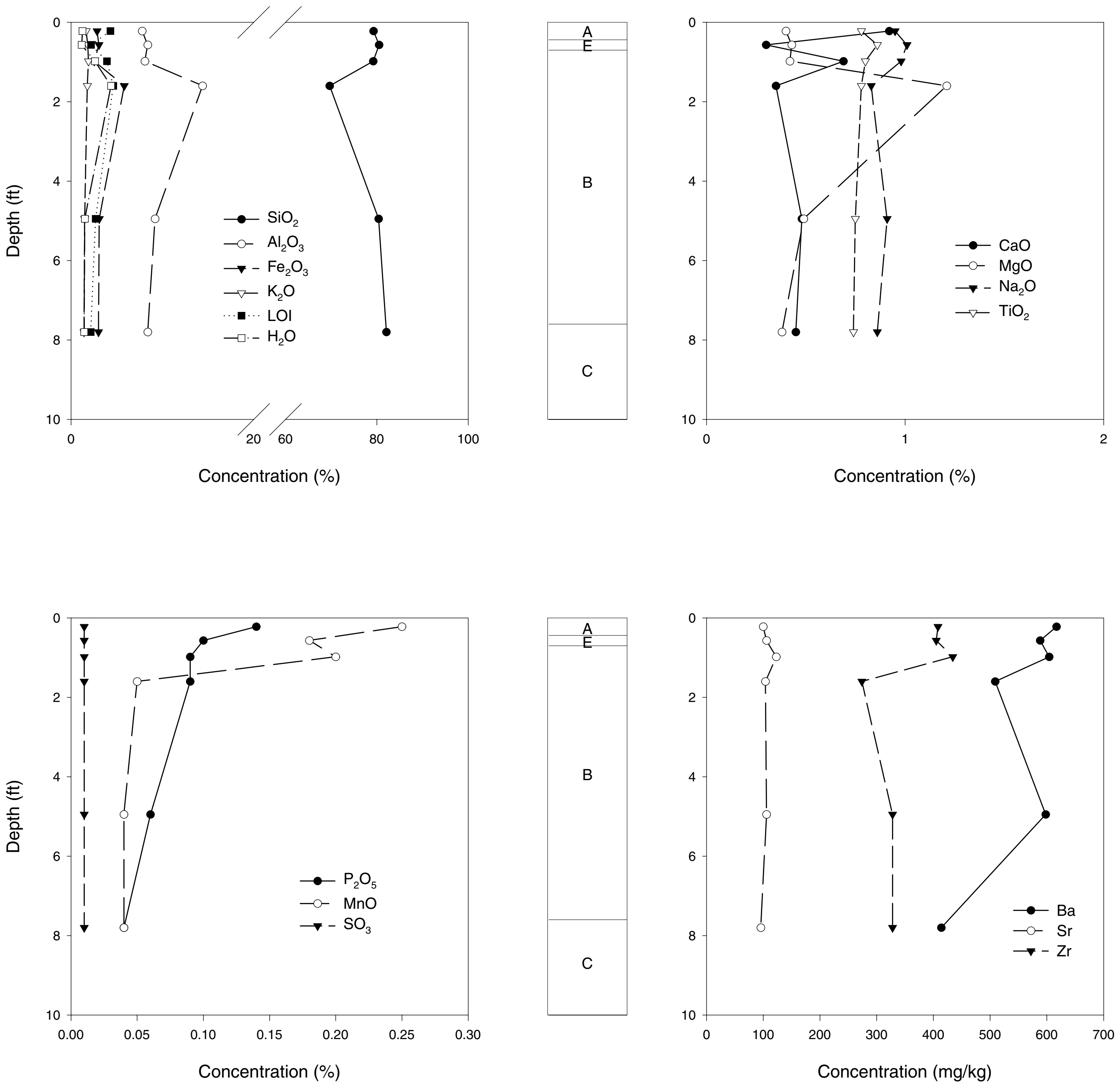


Figure 24. Elemental concentrations versus depth in core 20.

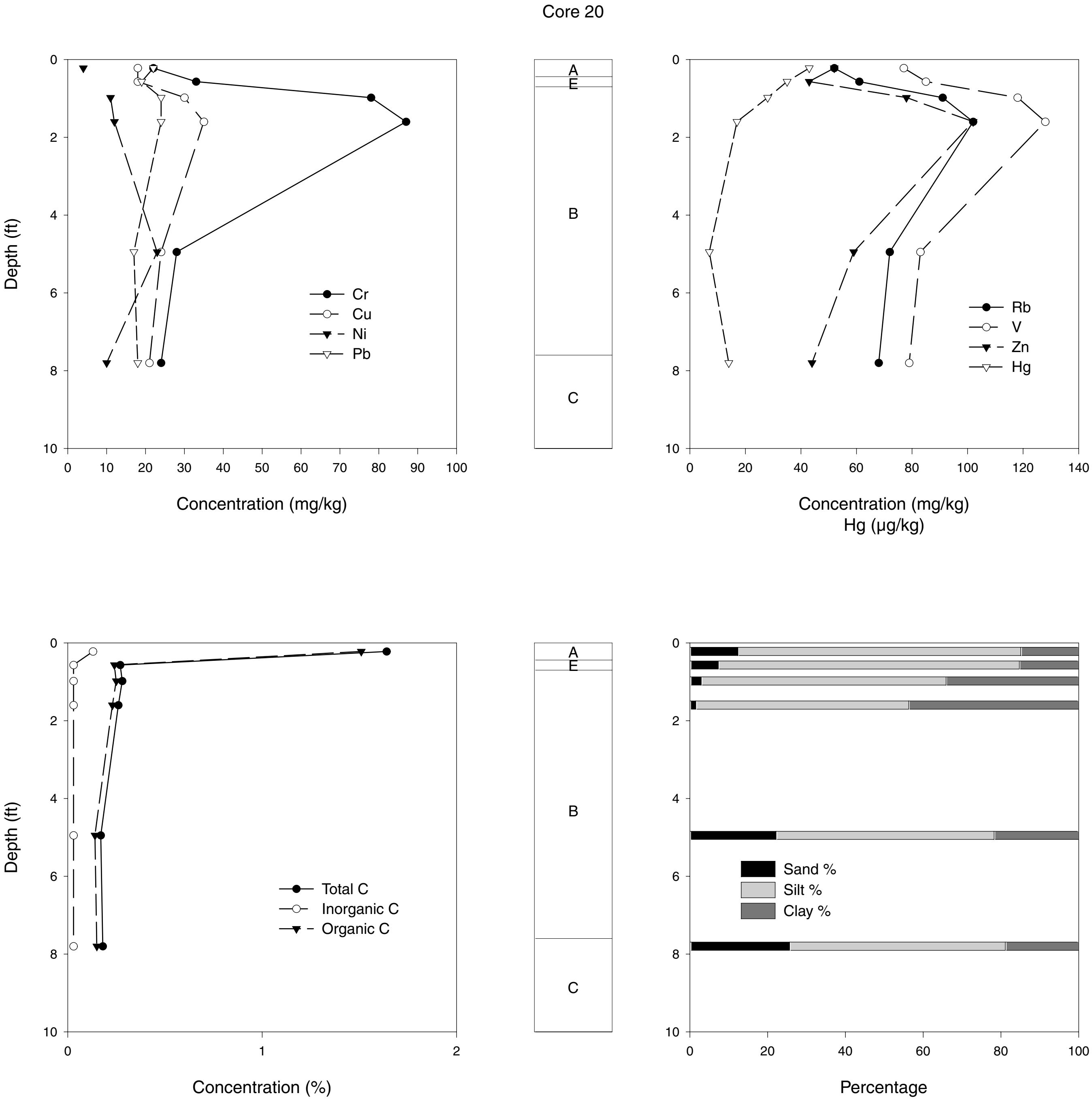


Figure 25. Elemental concentrations versus depth in core 20.

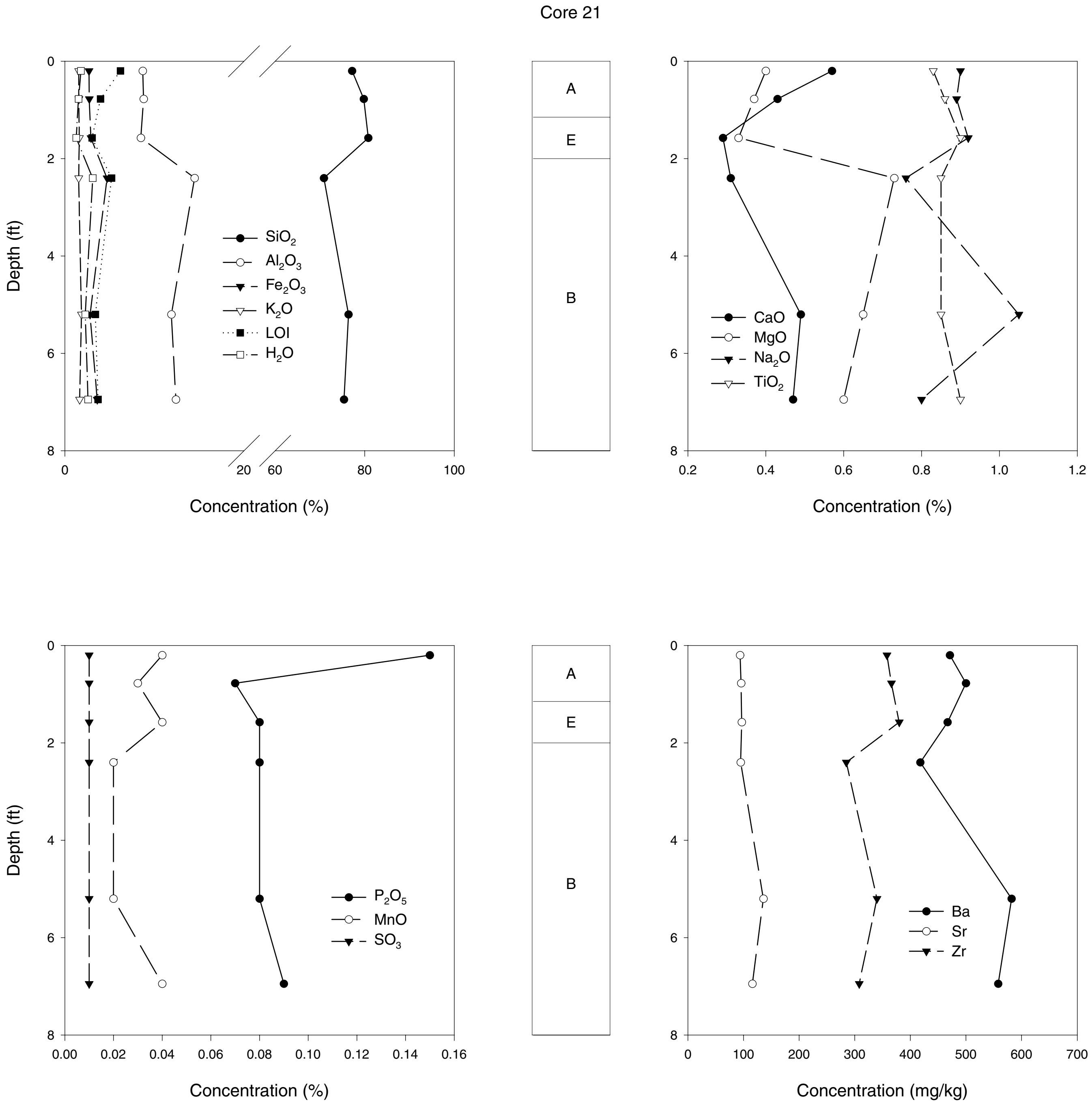


Figure 26. Elemental concentrations versus depth in core 21.

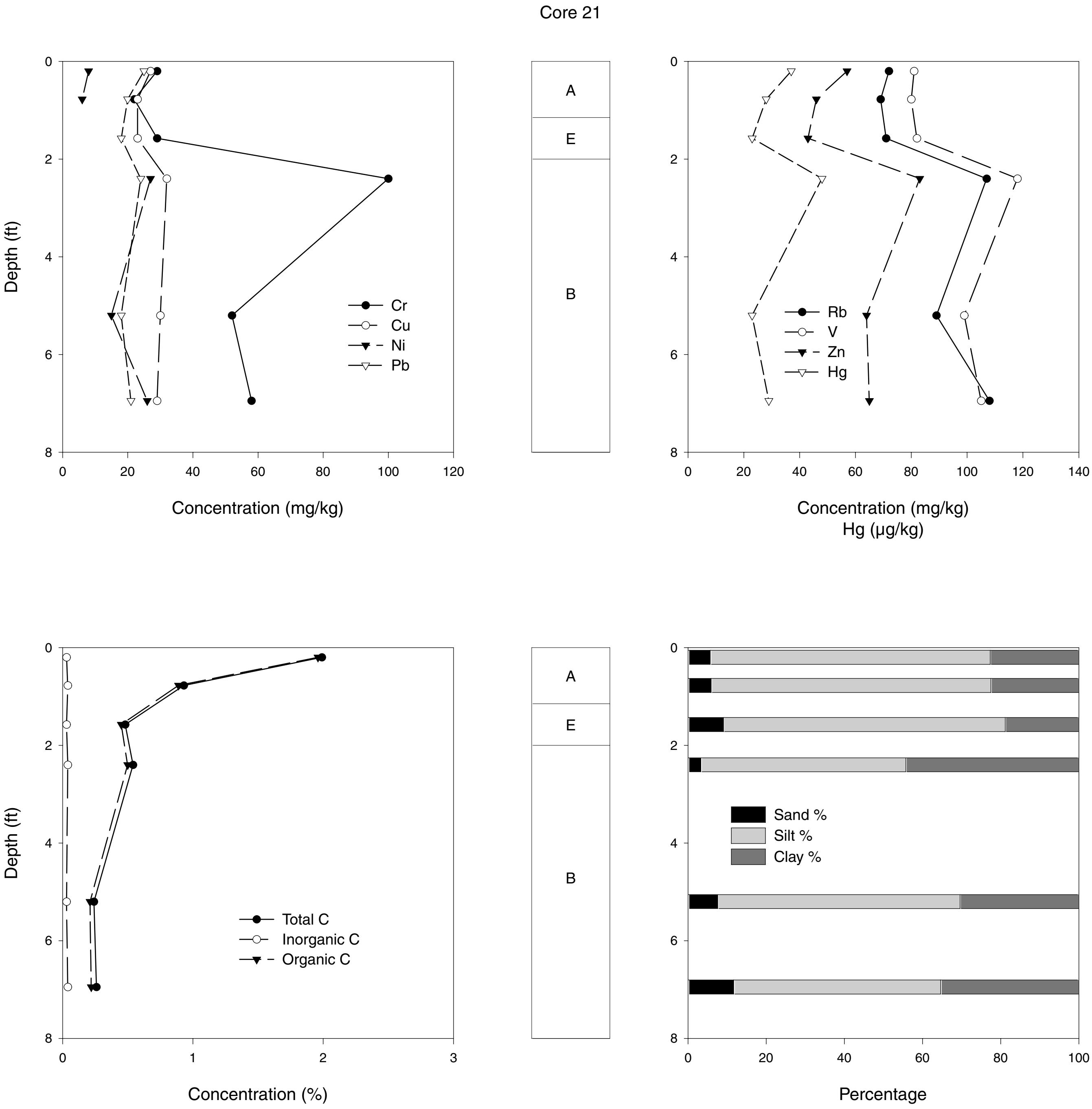


Figure 27. Elemental concentrations versus depth in core 21.

Core 22

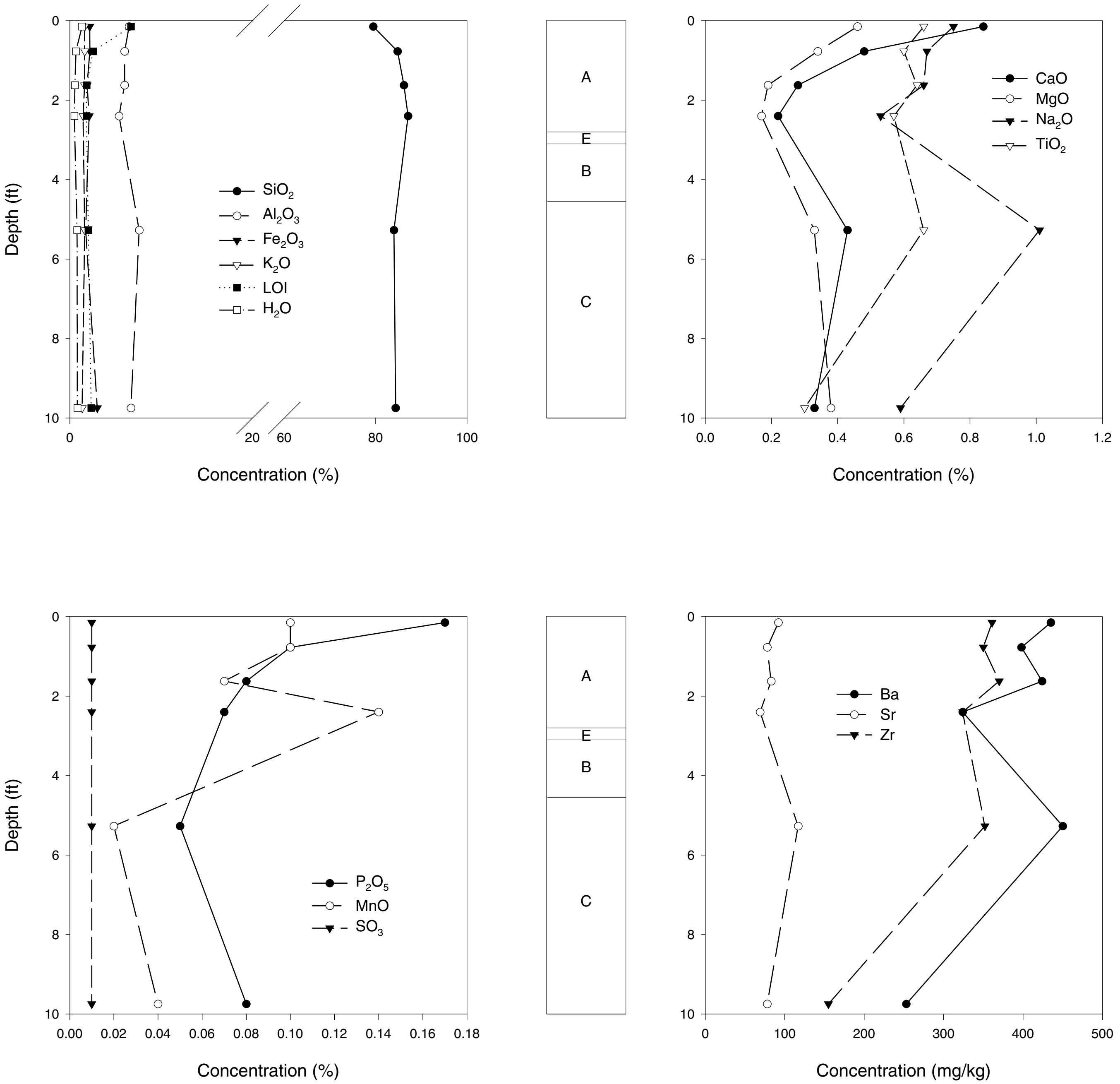


Figure 28. Elemental concentrations versus depth in core 22.

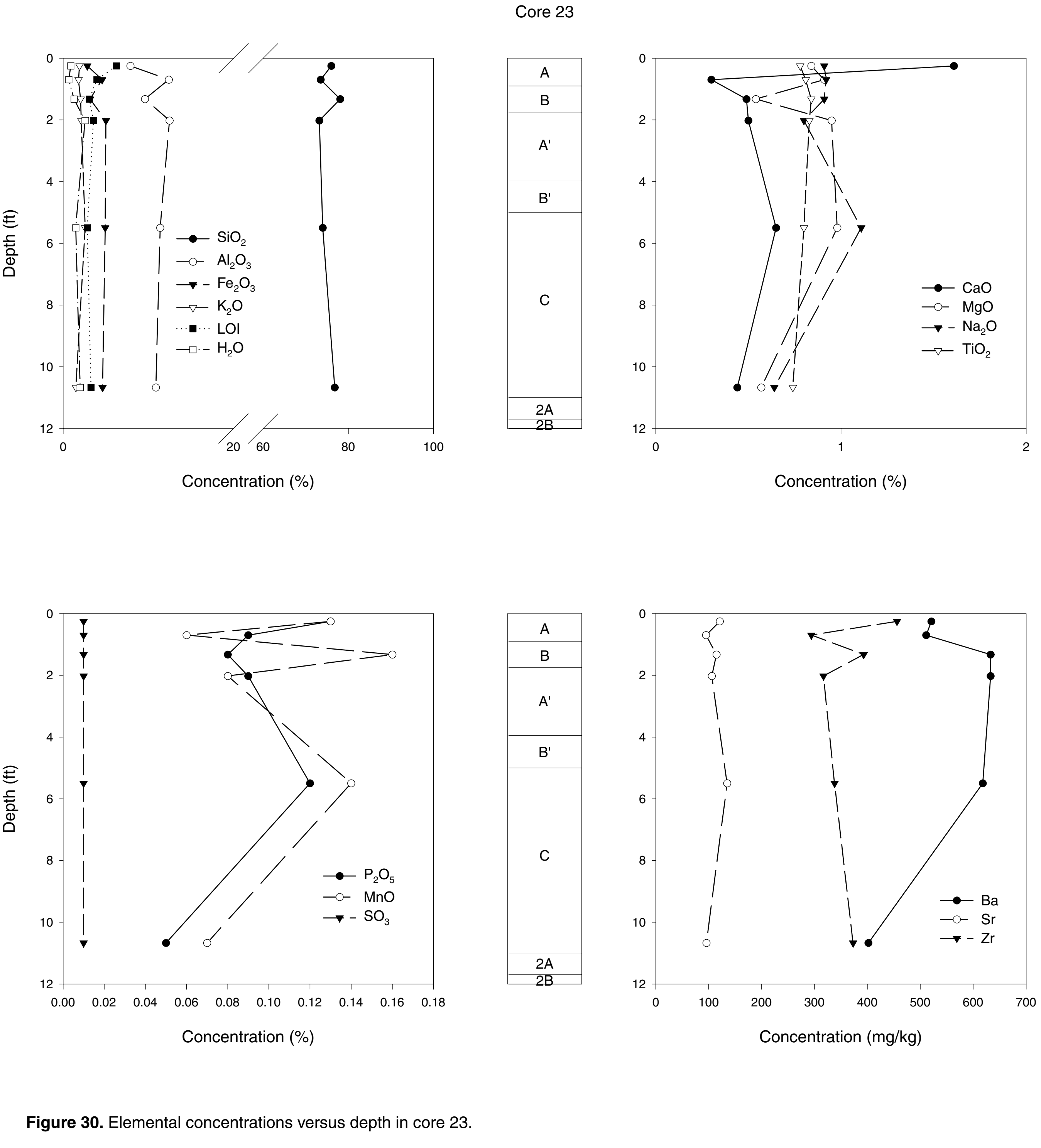


Figure 30. Elemental concentrations versus depth in core 23.

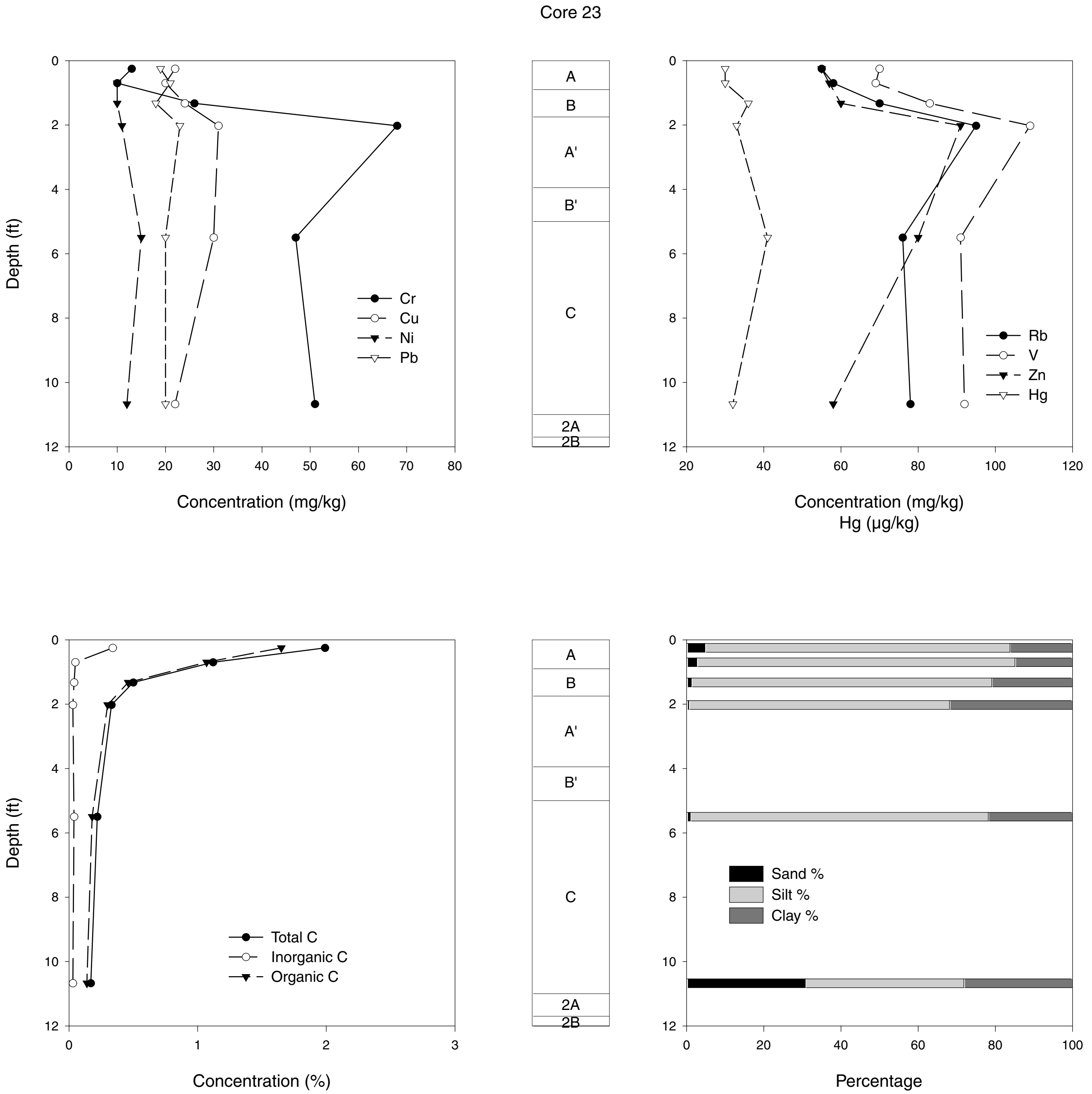


Figure 31. Elemental concentrations versus depth in core 23.

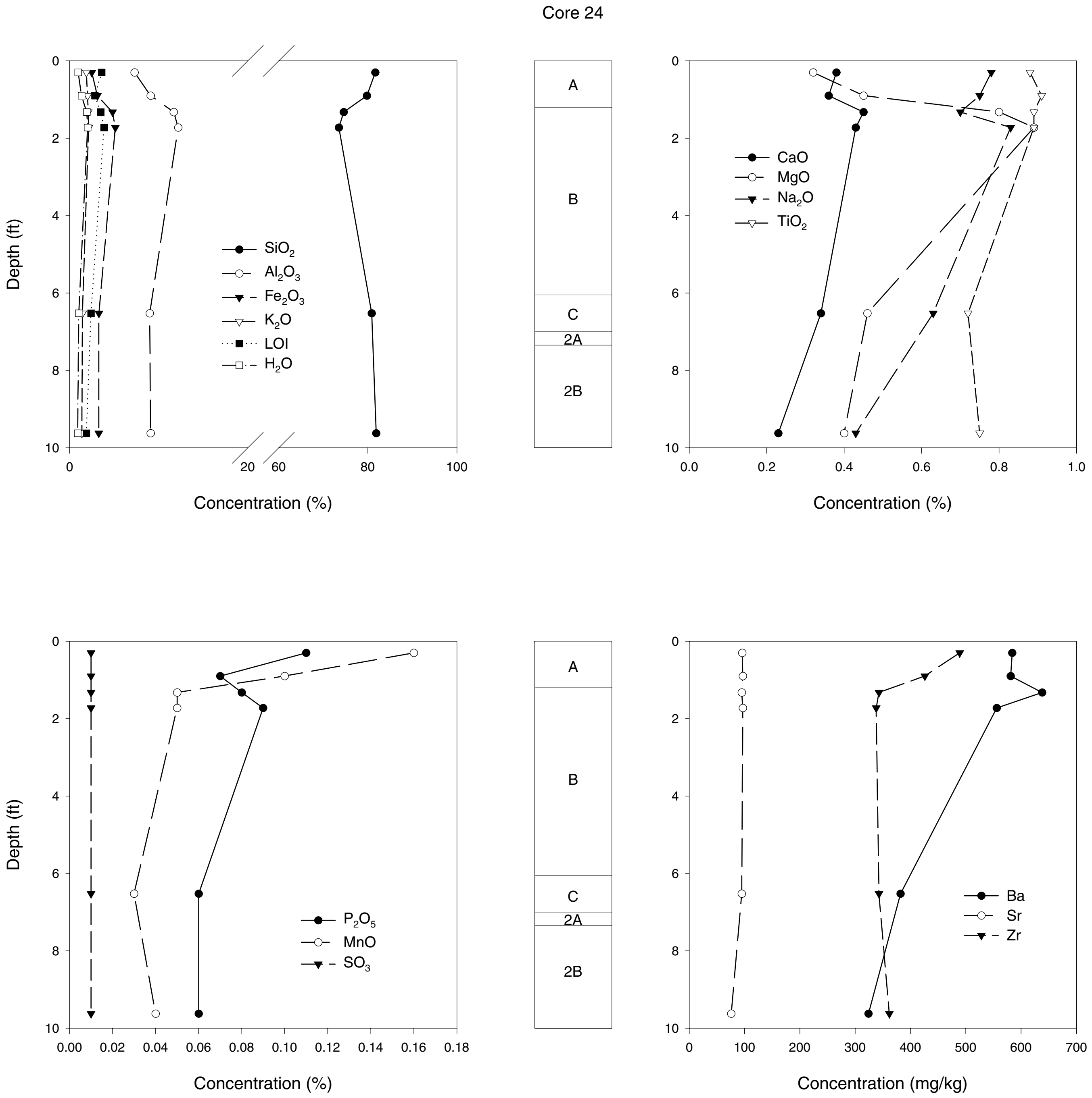


Figure 32. Elemental concentrations versus depth in core 24.

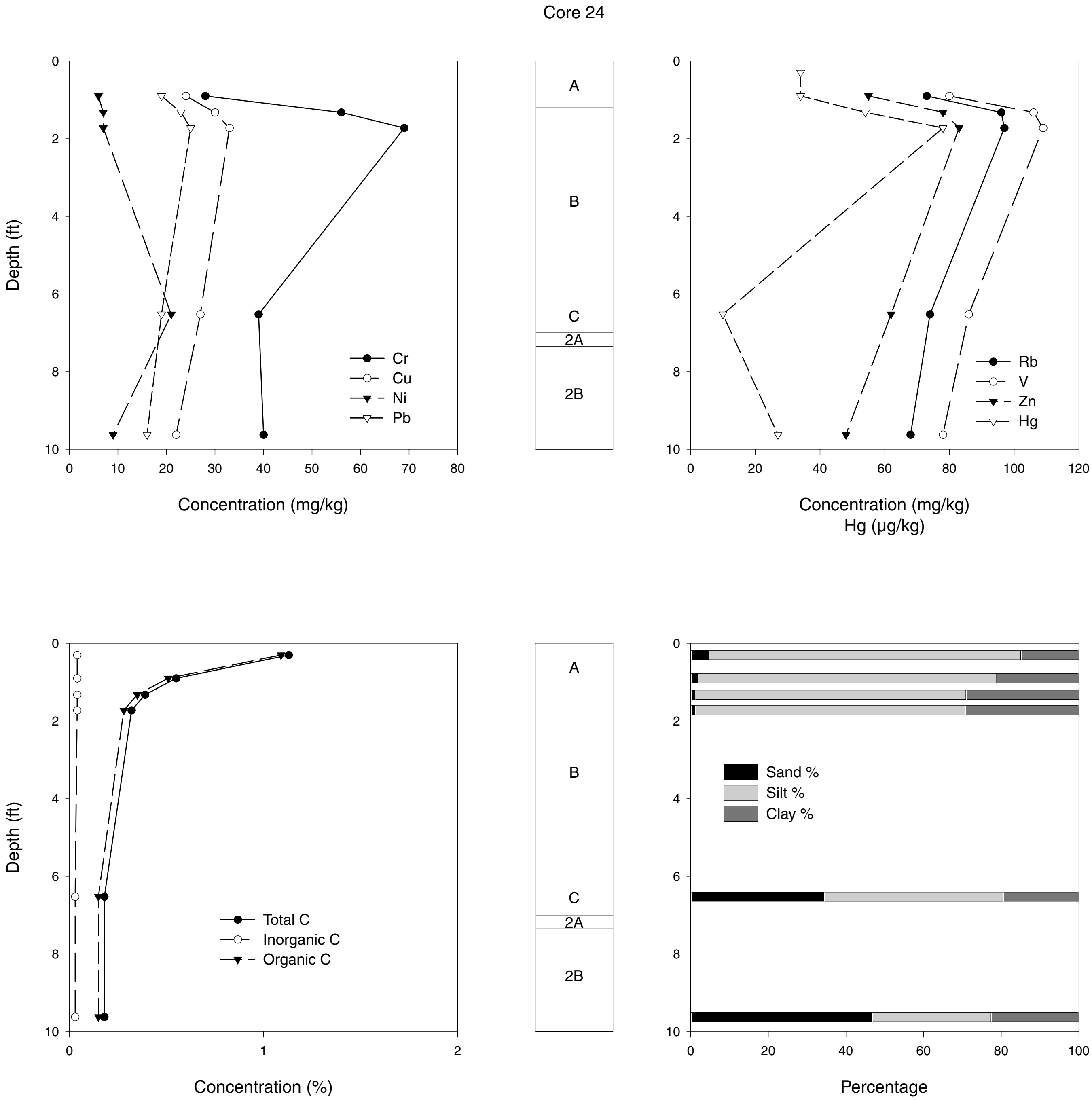


Figure 33. Elemental concentrations versus depth in core 24.

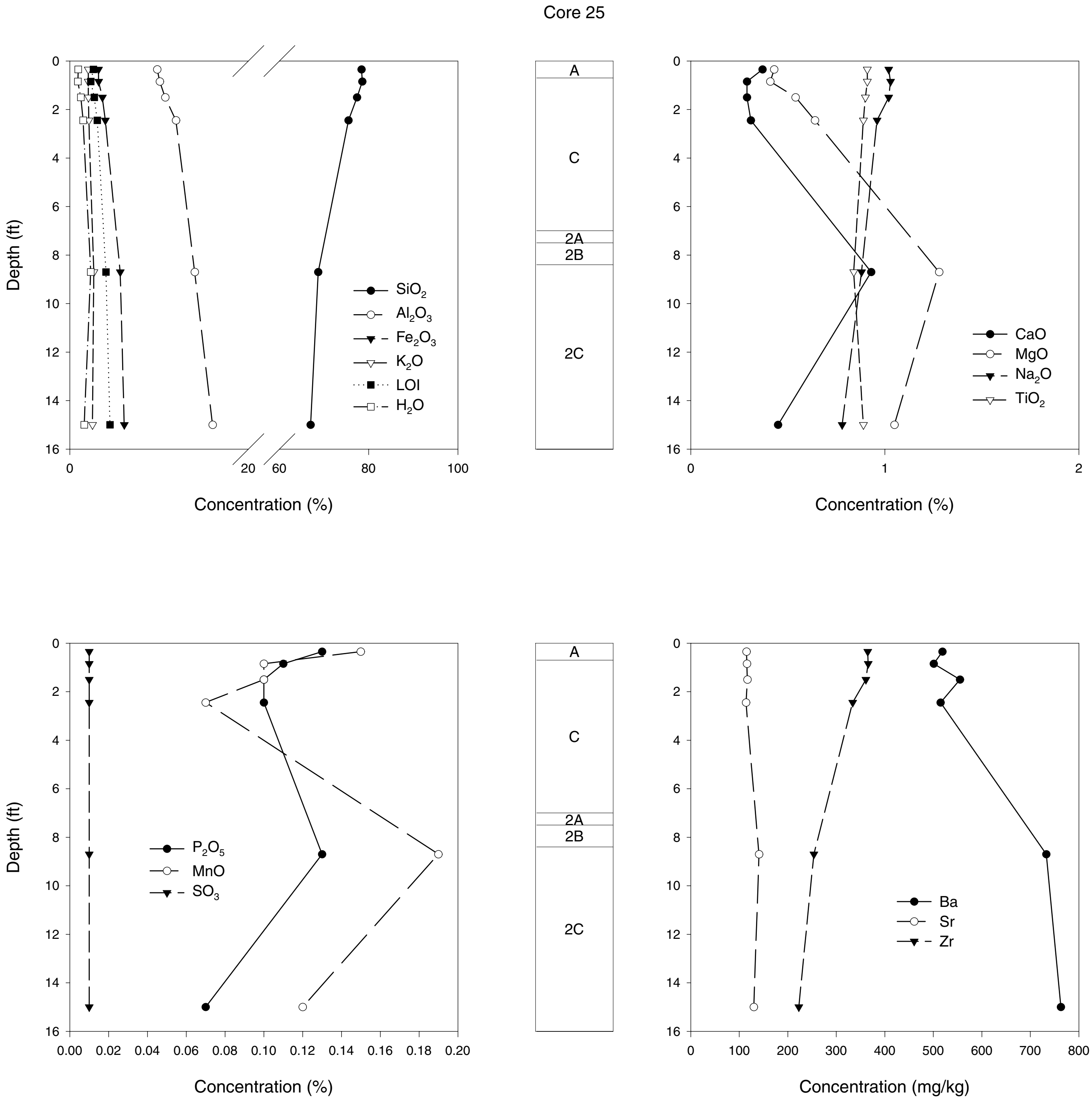

Core 25

Figure 34. Elemental concentrations versus depth in core 25.

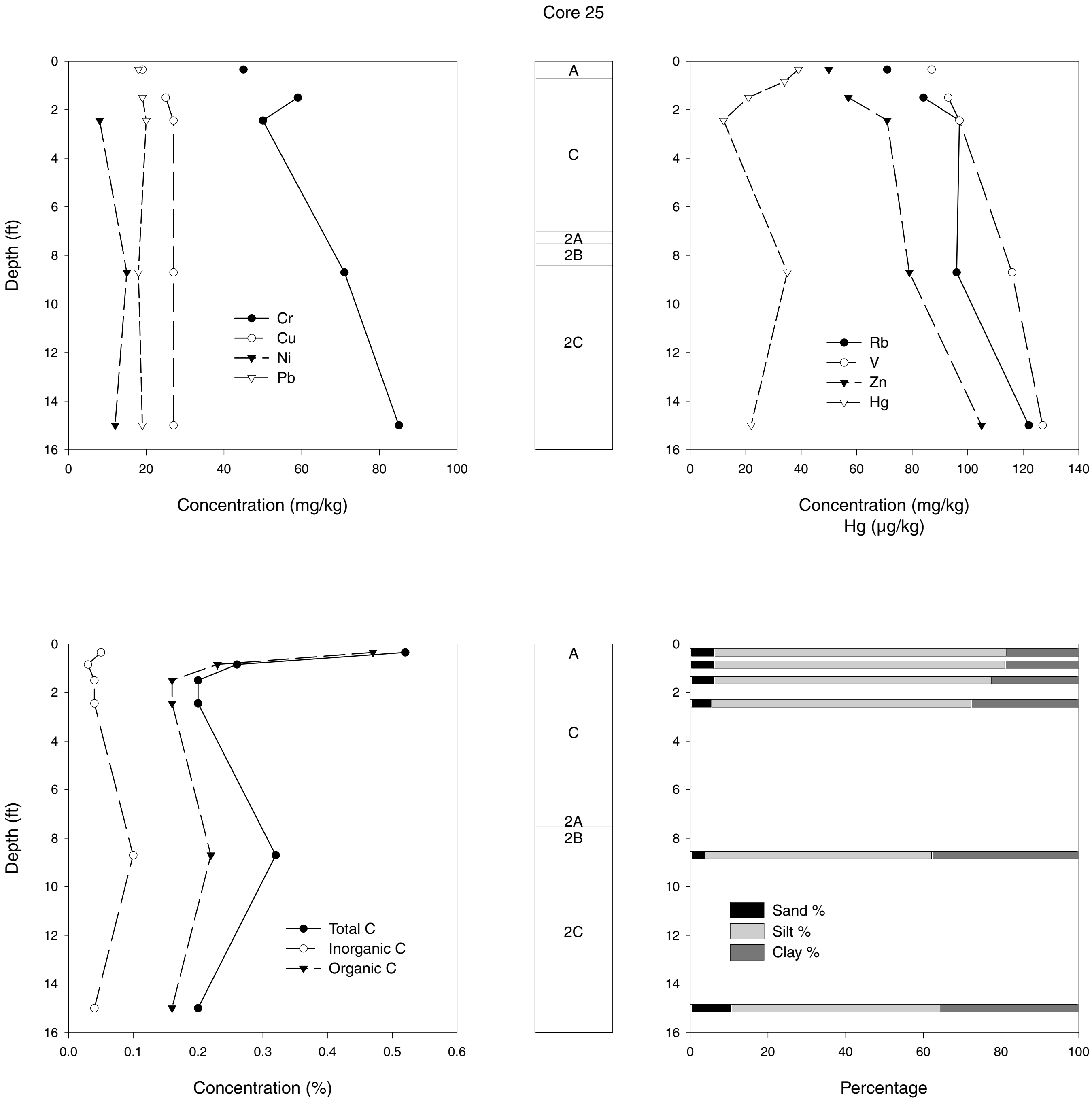


Figure 35. Elemental concentrations versus depth in core 25.

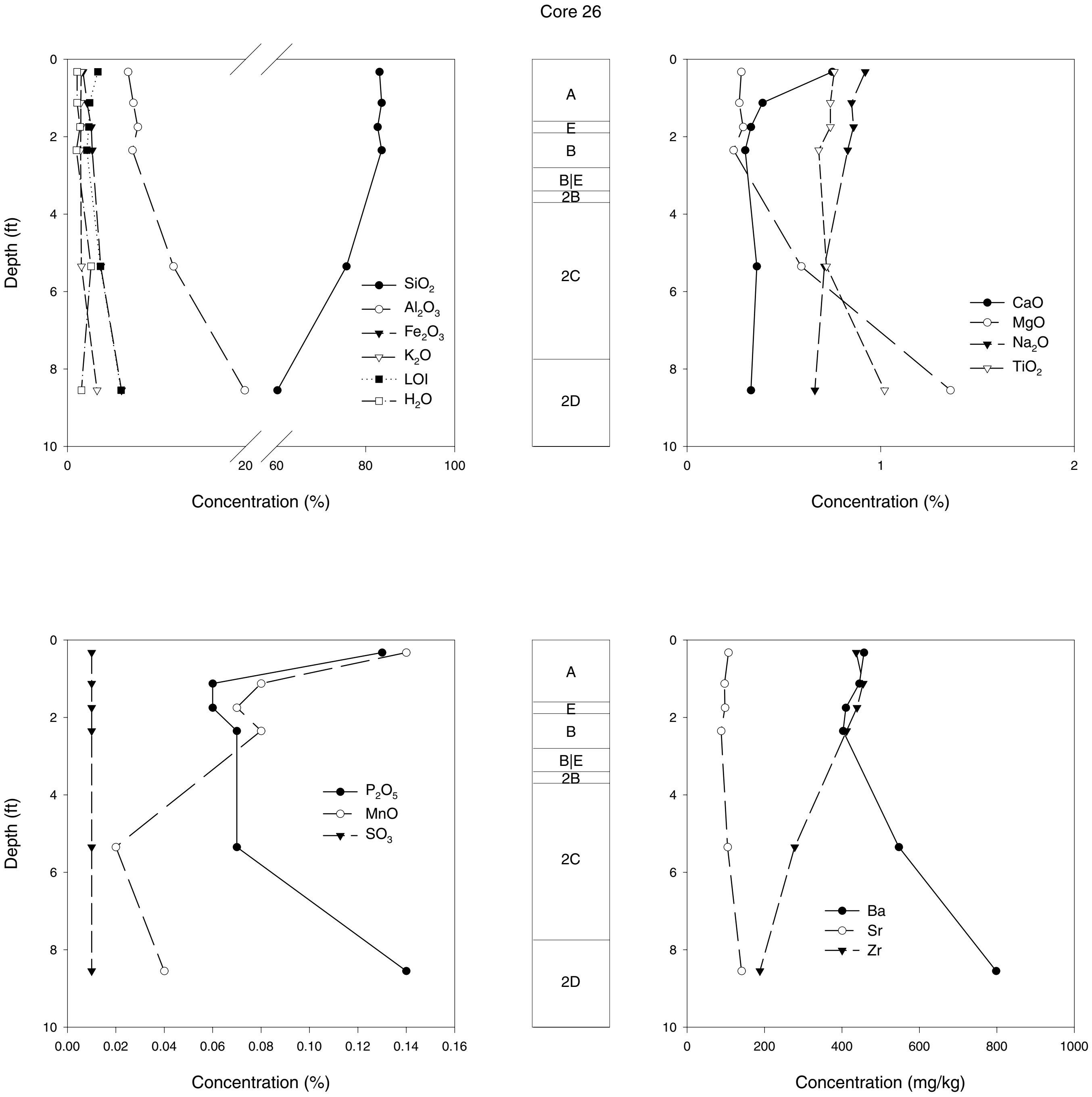


Figure 36. Elemental concentrations versus depth in core 26.

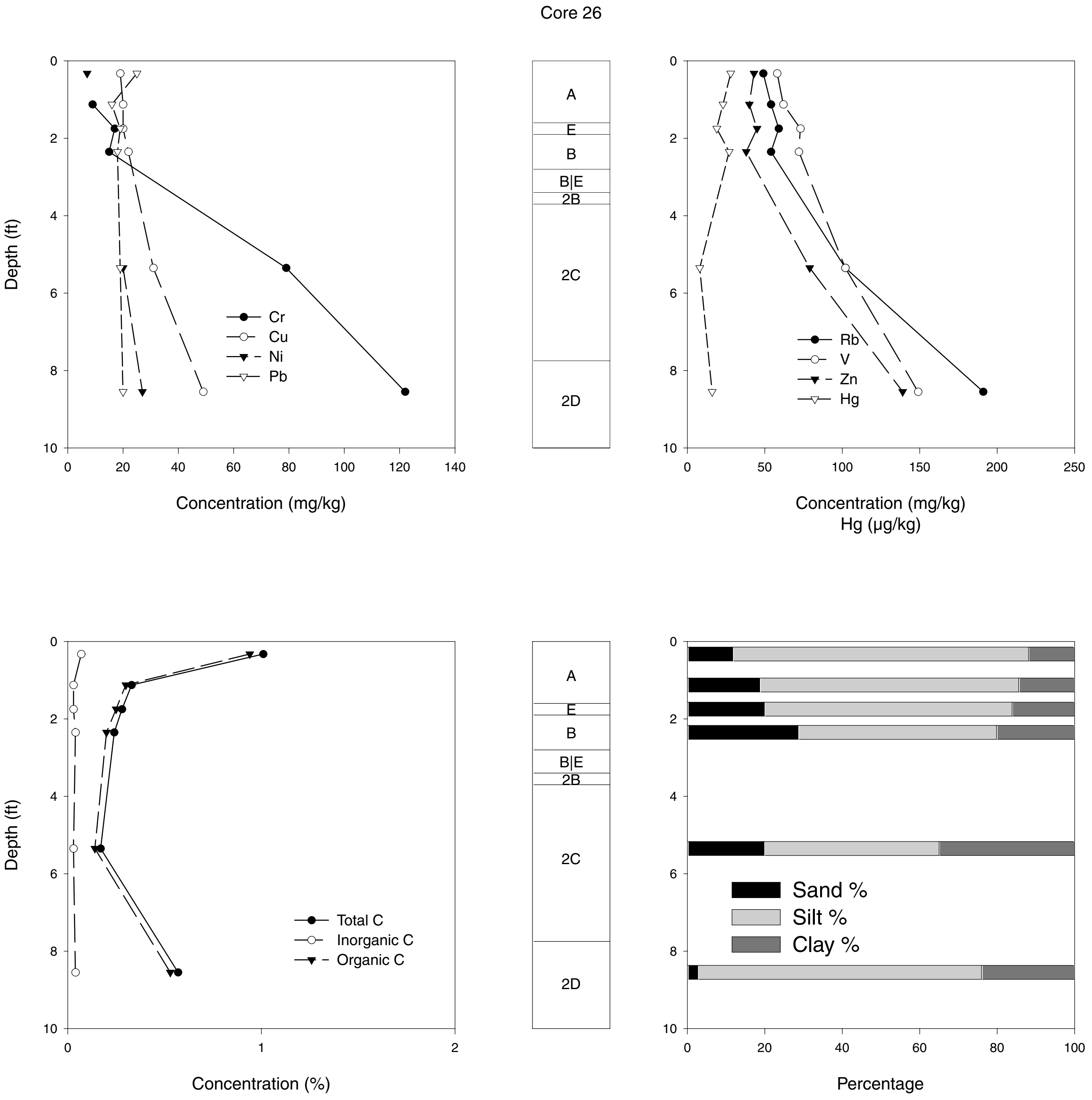


Figure 37. Elemental concentrations versus depth in core 26.