A Progress Report on the Description of the Geology and Chemical Composition of Soils in Illinois: Cores 127 through 137

by

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PREFACE

This is a progress report to present analytical chemistry data on soil cores 127 through 137 of a total of 137 soil cores. This report is not a final interpretation of the data collected. It 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 geological background and 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, the solution of environmental problems, as well as serving general educational needs about earth science.

The ISGS has a large collection of data gathered in the last 100 years for 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 available data, and forming a new interpretation. When the available information is insufficient, however, then a new research project must be conducted to address the need. This project was begun in 1998 to address the increasing interest of the general public in the chemical and mineralogical composition of soils in Illinois. To date, six open-file reports on results from samples of cores 1 through 126 have been published (Dreher et al., 2002, 2003a, 2003b, 2004a, 2004b, 2004c).

People are often confused about the differences between soil science and geology. The scopes of these subjects overlap and are interrelated; the two disciplines present 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 superimposed on the geologic material; there is both a geologic aspect and a soil aspect of 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 both 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 additional studies to fill gaps in the database following a priority plan determined by needs.

This report presents basic data acquired from soil cores 127 through 137 of a total of 137 coring sites from across the state. These cores were collected in central Illinois during the fall of 2003. We currently have descriptions, identifications of the geologic units and soil horizons, and results from a suite of chemical determinations from a selection of 6 samples per core. Mineralogical data are not available for the samples, but we have made what we believe to be educated guesses in the "Results and Discussion" section of this report about which minerals contain the various elements found in the soils. These educated guesses 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 with which to assess the health of the environment (including assessment of contamination of soils) and to aid 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 (Zhang and Frost, 2002), 94 samples of soil were collected from 54 counties in Illinois. 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 project is to determine the chemical and mineralogical characteristics of soils and their underlying unlithified parent 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 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

Most of 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). Although large-scale glaciation started earlier in Asia, the glaciers in North America did not grow large enough 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 2).

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 southward. 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 sediments formed by the glaciers in one way or another, 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 thicker deposits filled the pre-glacial bedrock valleys with drift up to 350 feet thick. Much of the drift is a homogenous mixture of fine-grained material with pebbles and a few boulders. This type of deposit was originally called boulder-clay and is now commonly called till. Interbedded with till in thick sequences of drift is outwash, which is commonly composed of stratified beds of sand and gravel, with some layers of silt and clay in places.

Beyond the margins of the continental glaciers, meltwater flowing from the glaciers coursed down 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 sediment. The surfaces of these deposits in many places 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 sediments laid down in former glacial meltwater channels form the excellent aquifers that supply water to many cities in Illinois.

The streams that joined the master meltwater rivers south of the glacial margins were flooded by the rising water in the master rivers during the major glacial events. This caused lakes to form in

the lower reaches of the tributaries. One of the larger flooded valleys 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 flood most of that region. This lake remained for several thousand years and largely filled the lowlands with stratified silt and clay; the sediment is over a hundred feet thick above the original channel of the Saline River. These deposits differ from typical 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 was laid down 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 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 the Mississippi's 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 1). 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 more than 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 in Illinois 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 the eastern Kankakee County area. During the peak of the last glaciation favorable conditions allowed "sand seas" to form. When the climate returned to warmer conditions 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 present 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 in rock particles in the sand or loess to be slowly altered, releasing ions (Na, Ca, Fe, Mg, K, P, S, and many others) and forming products (clay minerals) that cause the cation exchange capacity (CEC) of the developing soil to increase over time. The silt-size particles in loess generally are more altered or weathered than the coarser grains in the dune sands and, because loess deposits can retain significantly more water than dune sand, the water remains available to plants long after a rainfall.

The chemical composition of the Midwestern glacial deposits is near optimum for plant growth. In general, the 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 materials 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 allows 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 other 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 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 of the water and scope of the erosion caused by the Kankakee Flood (Willman and Frye, 1970) may seem incredible, but they 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 floodplains provide direct evidence for the wind erosion. The loess deposits are thickest along the bluffs of the main river valleys (Figure 1), which provides further evidence that river floodplains were the main source areas. In Illinois, at least four distinct deposits of loess have been verified in many 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 many 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 [Peoria Silt (Hansel and Johnson, 1996)] occurred during the last glaciation from about 25,000 to 10,000 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 glacial record 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 55,000 and 25,000 years ago.

In a few places in southern Illinois, the Peoria and Roxana silts 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 their physical and mineralogical properties. The most important such distinguishing characteristics are those caused by soil formation. To geologists, 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 units were once at the ground surface and exposed to active soil formation. Some other buried glacial deposits (for example, tills or outwash deposits) also have paleosols developed in their upper parts. These buried soils are important markers for mapping the distribution of certain units. The classification of the major Quaternary geologic units is based on these key markers (Follmer, 1982)

The Quaternary Deposits Map of Illinois (Figure 2) shows the distribution of the major Quaternary units (Lineback, 1981) as they would appear if they were not buried by loess deposits. The deposits of the last glaciation are named after Wisconsin. In Illinois the deposits of this most recent glaciation are mostly limited to the 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 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 commonly missing, also. At such sites, the soils are formed in other glacial deposits or bedrock.

The distinctive 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 biological traces) commonly extend though 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 called it Wisconsinan loess after the time interval during which it was deposited.

<u>Soil Development.</u> 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.

<u>Parent Material.</u> The mineralogical and chemical composition of the parent material governs the chemical composition of the soil (see Figure 3). 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 reconstituted by weathering processes into new minerals. Most elements that were present in trace concentrations in the 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).

<u>*Climate.*</u> The term "climate" encompasses both temperature and the long-term availability of water. The higher the temperature of the developing soil, the faster the soil will develop through chemical and biological processes. The availability of water is important in determining the types of plants and organisms that can 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), or passed into the groundwater.

<u>Topography.</u> 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).

<u>Organisms</u>. 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 (Ashman and Puri, 2002; 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 their absence (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 factories or power generating 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 μ m diameter) generally causes metals to move downward through the soil, but capillary action can also 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 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 solutes are attracted to the surfaces of the particles because of differences in electrical charge, and tend to become sorbed on the particle surfaces to the point of equilibrium with the concentration of the solute in the soil water.

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 processes whereby of solid materials enter into or separate 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 removed 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). Some 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, in 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 4).

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. All landowners we contacted in 2003 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 2000. Cores were collected at 25 locations during October 2000, after harvest was completed. The black square symbols on the map in Figure 4 indicate the approximate locations of the sampling sites. The core number, the county in which the core was collected, and the final depth of the core are recorded in Table 1.

A Giddings^{®1} 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 apparent 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), and manganese (Mn)] 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- μ m sieve. The ground samples were dried at 110°C overnight, and then ignited in platinum-rhodium crucibles at 1000°C for one hour to determine 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

¹The use of trade names does not constitute endorsement by the Illinois State Geological Survey.

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 in the following manner for the determination of trace elements, which generally are too diluted to be determined from the fused disk samples.

A portion of the 208-µm 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 agitated 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/in² 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 algebraic calculation to content in the original soil sample. The method detection limit is approximately 3 μ g of mercury per kg of soil.

Total Carbon and Inorganic Carbon

Total and inorganic carbon were determined in the 208- μ m 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 (CO₂) gas. The generated CO₂ 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 the CO_2 that was adsorbed by the ethanoldiamine

Soil pH

Five grams of the oven-dried $(50^{\circ}C) < 2 \text{ mm}$ soil 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 an oven-dried (50 $^{\circ}$ 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 (NaPO₃)₆ was added to the bottle followed by 140 mL of deionized water. A blank sample containing only (NaPO₃)₆ and deionized water also was prepared. 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 endto-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 28°C for 3 hours and 18 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 for settling at twominute 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 fractions had been withdrawn, the remaining contents of each bottle were poured through a 3.5-inch diameter, 62 μ m (No. 230) stainless steel sieve to separate the sand-size particles from the silt- and clay-size particles. Each bottle was thoroughly rinsed with deionized water. Successive rinsates were poured through the sieve until no particulate matter could be observed in the bottle. The sand was rinsed several times to remove all silt-size and smaller particles.

The sand was then backwashed from the sieve with deionized water into a beaker, and the sand was quantitatively filtered through a numbered and weighed circle of Whatman No. 41 filter paper. Each 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. The silt content of the samples was calculated by subtracting the percentages of clay plus sand from 100%.

Organic matter in a soil 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 (H₂O₂). We also found it necessary to treat the upper two samples from each core with H₂O₂. Ten grams of <2 mm sample was weighed into a 500-mL polymethlypentene bottle, then the bottle, cap, and soil sample were weighed and the weight was noted. Five mL of deionized water, 5 mL of 30% H₂O₂ and one drop of glacial acetic acid were added to the bottle in a fume hood and the bottles were loosely capped. When the reaction subsided, the bottles were transferred to a covered water bath at 60°C. Additional 5-mL aliquots of 30% H₂O₂ had been added to each bottle. The loosely capped bottles were allowed to stand in the covered water bath overnight. The bottles were allowed to cool to room temperature and then weighed. The amount of water required to bring the total amount of solution to 150 grams was calculated, then that amount of deionized water was added. Soil textures then were determined as described.

RESULTS AND DISCUSSION

Soil Texture

The sand, silt, and clay contents of each sample from each core are listed in Tables 2 through 12, and depicted graphically in odd-numbered figures 7 through 27. 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 11 cores discussed in this report, the textures of the uppermost or surficial horizon (the plow layer in most cases) of three of the cores were silt loam, seven were silty clay loam, and one was loam, as indicated in tables 2 through 12. As shown in table 13, the parent material of the soil at location 130 was alluvium; that is, the sediment was deposited on a flood plain by a stream. The soils of cores 128, 129, 131, and 133 developed in forest, and the soils of cores 127, 130, 132, and 134 through 137 developed on prairies. Cores 129, 131, and 133 were from upland forest locations, the soil of core 127 developed in lake bottom (lacustrine) sediments, the soil from core 128 developed in windblown (aeolian) sand, the

soil of core 130 developed in river bottom (alluvial) sediments, and the remaining 5 cores were collected from areas that were upland prairies during soil development.

The clay contents of the subsamples of all the cores ranged from about 0.6% to 48%, the silt contents ranged from about 4% to 78%, and the sand from 0.9% to 95%. The soil texture classification, soil type, soil association, and developmental environment of the uppermost soil samples are listed in Table 13. In each core the clay-size content passed through a maximum, generally in the B horizon, as shown in the odd-numbered figures 7 through 27.

The silt-size fraction of a soil is composed principally of quartz (SiO₂), with feldspar and carbonate minerals (where present), plus small amounts of heavy minerals, such as zircon (ZrSiO₄). Quartz and zircon are two of the most resistant minerals in soils along with rutile (TiO₂) and ilmenite (FeTiO₃), and some 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. 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 μ m (Dreimanis and Vagners, 1969, 1971). With artificial grinding, the terminal grade for quartz is about 16 to 32 μ m (Gaudin, 1926), slightly finer than that observed by Dreimanis and Vagners, but still in the silt size range of 16 to 62 μ m.

Results of Chemical Analyses

Table 14 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 11 cores are presented in Tables 15 through 25 and Figures 7 through 27 (provided at end of report). Major and minor element contents, reported as oxides (silicon through manganese 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, except for mercury, which is listed in μ g/kg.

Correlation Coefficients A correlation coefficient is a numerical description of the statistical 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 is likely to increase also. If the correlation coefficient is negative, then as the content of the one constituent increases, the content of the other is likely to decrease.

For example, the correlation coefficient between alumina (Al_2O_3) and the clay-size fraction is 0.87 (Table 14). That is, in about 87% of the samples, as the content of the clay-size fraction increased the alumina content also increased. Because two parameters are positively correlated does not necessarily mean that they are always present in the same ratio, only that they tend to

vary together. Therefore, correlation coefficients do not prove conclusively that any particular mineral is present in a sample; they are merely suggestive.

When the data from all 137 cores are considered in a final report we plan to study regional and state-wide trends, and differences between soil map units in terms of behavior of chemical elements in Illinois soils. For the present report we restrict our interpretations to cores 127 through 137. We calculated correlation coefficients to highlight the general relationships between constituents based on all samples analyzed from this set of cores.

Titanium oxide (TiO_2) was correlated (0.65) with the silt-size fraction, but it was also correlated (0.77) with the clay-size fraction. A possible explanation for this complex situation is that rutile (TiO_2) , ilmenite (FeTiO_3), and anatase (TiO_2) were all present in many of the soil samples. Rutile and ilmenite would have been inherited from the parent material and, because of their hardness, had a minimum size in the silt-size fraction. Anatase forms from the degradation of ilmenite; anatase crystals are very small and occur in the clay-size fraction (Milnes and Fitzpatrick, 1989). Therefore, titanium minerals in soils occur in both the silt-size and the clay-size fractions (Steinkoenig, 1914).

Clay minerals are important components of all soils. They occur principally in the clay-size fraction and contain Al as one of their major constituents. As already stated, aluminum (as Al_2O_3), demonstrated a strong positive correlation with the clay-size fraction (0.87).

The clay-size fraction was positively correlated with iron (0.78), potassium (0.67), titanium (0.77), copper (0.57), nickel (0.57), rubidium (0.85), and vanadium (0.89). The clay-size fraction was negatively correlated with sand-size (-0.78). Therefore, these elements also were correlated with each other. Iron is a common structural member, that is, it is a necessary part of certain clay minerals, although it may also occur as an exchangeable ion on clay minerals. Iron is commonly found in illite and in coatings of iron oxyhydroxides on other minerals (Wilding et al., 1977) and as concretions in many soils.

The other elements noted above, copper, nickel, rubidium, and vanadium 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, become trapped (occluded) within the clay mineral structure. The iron oxyhydroxide minerals are strong adsorbers of many trace elements, as well.

Calcium oxide (CaO) and magnesium oxide (MgO) were strongly correlated with each other (0.97) and with the inorganic carbon content (0.98 and 0.97, respectively). These correlations suggest the presence of calcite and dolomite in the soil samples. Titanium oxide and zirconium were correlated with each other (0.69). Each was also correlated with the silt-size fraction (0.65 and 0.57, respectively). The correlation between titanium oxide and zirconium was probably a result of their mutual correlation with the silt-size fraction and not because they form a particular mineral with each other.

Soil pH Soil pH is a measure of the acidity or hydrogen ion (H^+) concentration 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 the dissolution or decomposition of clay minerals, buffers the pH of the soil solution according to the following chemical reactions (Hassett, 1989):

$$\begin{split} Al^{3^{+}} + H_2O &= AlOH^{2^{+}} + H^+ \\ AlOH^{2^{+}} + H_2O &= Al(OH)_2^+ + H^+ \\ Al(OH)_2^+ + H_2O &= Al(OH)_{3s} + H^+ \\ Al(OH)_{3s} + H_2O &= Al(OH)_4^- + H^+ \end{split}$$

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, and this 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 H₂O 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 (CO₂) 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. (The soil exchange complex is that portion of mineral surfaces which is active in ion exchange.) 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.

In the pH range 5.5 to 7.2, weak acidic groups such as -AIOH 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⁺.

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):

$$CO_{2 gas} = CO_{2 aq}$$
$$CO_{2 aq} + H_2O = H_2CO_3$$

 $H_2CO_3 = HCO_3^- + H^+$ $HCO_3^- = CO_3^{2-} + H^+$

Above pH 7.2 the pH is controlled by the precipitation or dissolution of carbonate minerals such as calcite and dolomite. As calcite dissolves in the soil solution it releases carbonate ion (CO_3^{2-}) , which reacts with water to produce bicarbonate ions (HCO_3^{-}) , 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 11 cores ranged from 5.48 (mildly acidic) to 8.22 (moderately alkaline), with a median value of 7.28. Of the 66 samples, 26 had a pH between 5.48 and 7 and the remaining 40 samples had a pH greater than 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 5 and Tables 15 through 25, the pH value in 5 of the cores (129, 131, 134, 136, and 137) became more acidic (the pH value decreased with increasing depth to about 4 feet or less, then became more alkaline below this depth. Calcareous till or drift that underlies the modern soil was penetrated in all 11 cores. In each core the CaO, MgO and inorganic carbon contents increased sharply in the samples that were selected from the till or drift. Carbonate minerals in the till or drift impose an alkaline pH on the samples.

Means and Ranges of Element Contents The mean and range of the element contents determined in the uppermost samples from the 11 cores from this portion of the project are compared in Table 26 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.

Excursions of concentrations outside the ranges observed by Shacklette and Boerngen did not necessarily mean that the soil sample was deficient or contaminated in an element, nor that plants grown in that soil will absorb any of those elements in toxic amounts. Much depends on the particular minerals present in the soil and 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.

Cores 128 and 129 were collected in parks in urban areas. In both cases, it appears that fill material was placed above the existing modern soil. In both cases there was a buried A horizon or Oa horizon (core 129). An Oa horizon is a horizon that has an unusually elevated content of highly decomposed organic matter (13.98% in the case of core 129). Gas chromatography-mass spectrometry confirmed that all the organic matter in the sample from the Oa horizon was

derived from plant material.

Core 128 comprised much fine sand that had been placed above the original A horizon. More sand was observed below the buried A horizon. Although the soils at these two core locations are not normal, we included them in our discussion because they are part of the set of collected cores, and they represent the soil in place.

Silicon According to Kabata-Pendias (2001), quartz, or SiO₂, 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 during glacial transport, but had there been increased transport distance there would have been an increase in grinding and a consequent increase in the amount of coarse silt (31 to 62 μ m) would have been expected (Dreimanis and Vagners, 1971). As already stated, experimental tests by Dreimanis and Vagners (1971) indicated a terminal grade for quartz between 31 and 62 μ m (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).

There were no significant correlations between SiO_2 and sand, silt, or sand+silt, but silica was correlated with Na₂O, Ba, and Zr.

Generally, a correlation between SiO_2 and the sand-size fraction or the sand+silt fraction is expected. However, in these 11 cores no such correlation existed. Quartz remained the major component in the sands, however, feldspar, dolomite, and hornblende were observed visually to be significant components of the sands, perhaps more than in previously collected cores, in which the correlation between SiO₂ and sand or sand+silt was important.

Calcareous till was penetrated in all 11 cores. In most of the 11 cores, when the underlying calcareous till or drift was penetrated, the SiO_2 and the Al_2O_3 contents decreased as the CaO and MgO contents increased because of dilution of the SiO_2 and Al_2O_3 by dolomite [CaMg(CO₃)₂]. When the CaO and MgO contents increased sharply and the SiO_2 content decreased, it occurred in a sample from the BC, C, or CB horizon.

Aluminum The aluminum content of soils is mostly inherited from the parent materials (Kabata-Pendias, 2001). Above the calcareous till the Al_2O_3 content of the cores generally increased with depth. The depth profiles of Al_2O_3 followed the profiles of the clay-size content.

Above the calcareous till the Al_2O_3 depth profiles often (cores 128, 129, and 132-137) were mirror images of the SiO_2 profiles. When the CaO and MgO contents increased significantly, the Al_2O_3 content decreased because of dilution by dolomite. The Al_2O_3 content followed the claysize content closely, with a correlation coefficient of 0.87. **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 varied within the range of 1.16 to 8.27%. In all but one core (128) the Fe_2O_3 content passed through a maximum with increasing depth, generally in the B horizon, the clay-rich zone in the soil profile. The positive correlation between $Fe2O_3$ and clay-size fraction (0.78) affirms that iron oxyhydroxides tend to occur as coatings on clay particles.

Potassium Typically, potassium occurs in Illinois soils in important rock-forming 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, which means it occurs in a form that is not readily soluble.

The content of K_2O ranged between 1.94 and 3.66%. The greatest variation of K_2O content within a depth profile was 1.48% (2.18 to 3.66%) in core 137. The correlation of K_2O with the clay-size fraction was 0.67.

Calcium and Magnesium Calcite $(CaCO_3)$ and dolomite $[CaO \cdot MgO(CO_2)_2]$ are common sources of calcium in soils, but not all soils contain calcite or dolomite. Calcium can also be held as an exchangeable ion on the soil exchange complex. Sedimentary minerals such as dolomite are probably the principal sources of magnesium in Illinois soils.

In each core, the greatest CaO content occurred in the BC, C (core 127), or CD (core 130) horizon. The soil of core 127 developed in an ancient lake sediment. Calcite (CaCO₃) masses were observed in the soil from a depth of 1.6 feet to at least 4.0 feet. In core 128, the sand content increased abruptly at the base of the A horizon. Dolomite grains were observed in the sand, giving rise to the abrupt increases of CaO and MgO below the A horizon. In core 129, calcareous masses were observed in the second sample below the surface, which explains the increased contents of CaO and MgO at this level. Below that sample was located a buried Oa horizon with little CaO and MgO contents. The CaO and MgO contents increased again in the two lowermost samples as the core penetrated the underlying till.

Figure 28 shows a scatterplot of MgO content versus CaO content for all samples from cores 127 through 137. The dashed line through the points represents the theoretical composition of dolomite, that is, the weight-percent ratio of CaO/MgO is 1.39, or the molar ratio is 1. The solid

line represents the best-fit (linear regression) line through the points. The regression line is close to the theoretical line, which indicates that dolomite was the dominant Ca-Mg mineral in the 11 cores.

Sodium Sodium-rich feldspar is probably the principal source of sodium in soils that are not near either an ocean or a roadway on which de-icing salt is applied (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. 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.

The range of Na₂O content in the 11 cores was small, with only 0.76% between the minimum value of 0.31% and the maximum of 1.07%. In eight of the 11 cores (128-130, 132-134, 136, and 137) the Na₂O content attained a maximum value in the B horizon or below, suggesting downward leaching of soluble sodium-containing compounds. A small portion of the sodium present in the parent materials probably was incorporated into exchangeable positions on clay minerals, but most sodium probably was 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 commonly occur nearly undecomposed in soils. Titanium presents no environmental concerns in soils (Kabata-Pendias, 2001). The titanium content exhibited behavior similar to that of sodium. The range of TiO₂ content was less than 0.75 percent in all 11 cores, with a minimum of 0.14% and a maximum of 0.74%.

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.

In all but one core (129) the content of P_2O_5 was at its maximum in the surface soil sample and decreased in at least the next deeper sample. Possibly P_2O_5 was contributed to the surface soil by plowing plant debris back into the soil or by application of phosphate fertilizer to the soil. In core 129 the maximum P_2O_5 content occurred in the sample that also contained the maximum organic carbon content. The range of P_2O_5 content was 0.26%, from a minimum of 0.05% to a maximum of 0.31%. The correlation between P_2O_5 content and organic carbon was 0.81.

Manganese The principal source minerals for manganese in soils are amphiboles, pyroxenes, biotite $[K(Mg,Fe)_3(AlSi_3O_{10})(OH)_2]$ (in which Mn can replace iron to a limited extent), and rhodonite (MnSiO₃) (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 of MnO_2 (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 has been observed (McKenzie, 1989). However, the lack of correlation between MnO and Fe_2O_3 in the samples from these cores suggests no such intimate association.

In 5 cores (127, 129-131, and 137) the MnO content passed through a maximum in the B horizon or below. In the remaining 6 cores (128 and 132-136) the MnO content was greatest in the A horizon. There was a secondary maximum in MnO content, however, in cores 132, 133, and 135-137, in the B horizon or below. One possible reason for the subsurface maximum is that manganese is sensitive to oxidation and reduction. Oxidized species, such as MnO₂, precipitate where oxygen is readily available, as it normally would be near the soil surface, 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. The MnO profiles of cores 128, 130-132, and 136 were the inverse of the pH profiles for those cores.

Barium Micas and feldspars are sources of barium in soils. These minerals contain potassium, which is commonly replaced by barium because the two atoms are of similar size (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 (BaSO₄, 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 Ba content of six cores (128 and 133-137) was at its maximum in the A horizon. In the remaining five cores the Ba content achieved its maximum in the B horizon or below. The propensity of Ba to adsorb on clay minerals might explain the subsurface maxima in the Ba contents. Barium might have been carried downward as an adsorbed species on colloidal clay particles.

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 (FeCr₂O₄) in the parent material (Kabata-Pendias, 2001), and industrial fallout, such as dust and industrial contaminants.

The Cr content in the 11 cores ranged from 33 to 756 mg/kg, with an average of 114 mg/kg. The behavior of Cr in the depth profiles was frequently erratic, as in cores 128-130, 132, and 136, in which unusually high Cr contents were observed. The Cr content was not well correlated with any other constituent for which we analyzed, which suggests multiple modes of occurrence. The Cr content was at its maximum in the A horizon of cores 128 and 133 through 137, and in the B horizon or below in the remaining cores.

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.

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 sulfide and/or carbonate minerals 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 individual soil profiles did not vary appreciably (Kabata-Pendias, 2001).

The maximum Cu content in eight of the 11 cores occurred in the B horizon. In core 128 the maximum Cu content was in the A horizon, in core 129 it was in the Oa horizon, and in core 131 it was in the C horizon. The Cu content followed the clay-size content in cores 127, 130, 131, and 133 through 135. In cores 128 and 129 the Cu content followed the organic carbon content. The Cu content followed the Fe₂O₃ content in cores 127 and 130 through 135. The range of Cu content was 7.8 to 76 mg/kg, with an average of 31 mg/kg.

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.

The maximum Pb content in 8 cores (127, 128, 130, and 132-136) was in the surface sample. In each of two cores (129 and 137) there was a secondary maximum Pb content in the surface sample (a deeper sample in each of these two cores contained greater Pb content). In core 131 the surface sample contained 23 mg Pb per kg soil, and the next deeper sample contained 24 mg/kg. The Pb content in the surface sample of core 128 was 146 mg/kg, far greater than the two cores with the next greatest Pb content (cores 127 and 130, with 39 mg/kg each). The core at location 128 was collected at an urban park. About 75 feet west of the coring location was a railroad track, and about 115 feet north there was an old coal-burning power generating station that formerly generated power for the park. Particles of what appear to be furnace slag were observed in the sand grains from the surface sample and the next deeper sample. These particles could have come from stack emissions of coal-burning locomotives or the power plant, or from fill

material that apparently was placed at the site.

At the other 10 sites the accumulation of lead in the uppermost samples suggests that Pb was contributed by an external source, such as atmospheric fallout from industrial activities or the use of leaded fuel in vehicles that passed over or near the coring location. Lead was correlated with Zn (0.75) and mercury (0.72).

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). Although some of the mercury in soil is inherited from the parent materials, mercury probably is also deposited on the soil surface as atmospheric fallout of particulate matter from industrial processes or as a dissolved component of rainwater. Applications of phosphate-rock or sewage sludge as fertilizer are other possible sources of mercury.

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 (S²⁻) 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 contain 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 has not reacted 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.

The Hg content of the 11 cores was in the range of 7 to 115 μ g/kg (parts per billion, as opposed to parts per million for other trace elements). In cores 127, 128, 130, and 132 the maximum Hg content was in the A horizon. In core 129 the maximum was in the 2Oa horizon, and in cores 131 and 133 though 137 it was in the B horizon or below. Mercury was correlated with Cu (0.73), Pb (0.72), V (0.51), Zn (0.77), total carbon (0.65), and organic carbon (0.85). Mercury was negatively correlated with Ca) (-0.55), MgO (-0.53), inorganic carbon (-0.59), and pH (-0.61). These correlations suggest that Hg was associated with soil organic matter.

Nickel The major source of nickel in soils is the parent material, but, a possible external source is airborne particulate from industrial emissions. 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).

The Ni content ranged from 16 to 81 mg/kg, with an average of 38 mg/kg. In all cores except 128, the maximum Ni content occurred in the B horizon or below. In these cores the Ni content was more strongly correlated with Fe_2O_3 (0.71), than with clay-size content (0.59), although both

iron oxyhydroxides and clay minerals will adsorb Ni. Nickel was also correlated with Al_2O_3 (0.62), K_2O (0.59), Cu (0.66), Rb (0.62), and V (0.64). It was negatively correlated with pH (-0.57).

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 in the upper layers and increase downward due to leaching.

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 all 11 cores ranged from 51 to 139 mg/kg, within the ranges stated above. Rubidium was correlated with Al_2O_3 (0.75), Fe₂O₃ (0.68), K₂O (0.84), TiO₂ (0.59), Cu (0.52), Ni (0.62), V (0.72), Zn (0.53), and clay-size fraction (0.85). Rubidium was negatively correlated with the sand-size fraction (-0.59) and pH (-0.82). These correlations suggest that Rb was adsorbed by clay minerals and iron oxyhydroxides. The correlation between Rb and K₂O is consistent with the substitution of Rb for K in certain minerals.

Strontium Strontium generally is associated with soil organic matter, but it may also precipitate under alkaline conditions as strontianite ($SrCO_3$), and is commonly associated geochemically with calcium. 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 content varied between 87 and 188 mg/kg, with an average of 131 mg/kg. The variation of Sr content through the profiles was small; the greatest variation of 65 mg/kg occurred in core 128 and the least variation of 16 mg/kg occurred in core 133. Strontium was correlated with Na₂O (0.79).

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 range of V content in the 11 cores was 18 to 125 mg/kg, with an average of 87 mg/kg. In all but one core (128) the V content was at its maximum in the B horizon or below. In every core except 128 and 130, the V depth profile generally followed the clay-size depth profile. Vanadium in core 128 followed the silt-size depth profile. The general parallelism between V and the clay-

size content suggests that V was adsorbed by clay minerals. Vanadium was correlated with Al_2O_3 (0.94), Fe_2O_3 (0.86), TiO_2 (0.90), Ba (0.51), Cu (0.65), Ni (0.64), Rb (0.72), Zn (0.61), Hg (0.51), and clay size fraction (0.89). Vanadium was negatively correlated with sand-size fraction (-0.79) and pH (-0.87). The correlations between V and Fe_2O_3 and clay-size fraction indicate associations of V with iron oxyhydroxides and clay minerals.

Zinc Zinc is strongly adsorbed by clay minerals and soil organic matter and, therefore, is commonly present in higher concentrations in the B horizon of 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). The imbalance between input and output can result in a short-term accumulation of Zn in the surface horizon, but formation of soluble mobile species results in downward movement of Zn to the B horizon, where it may be tightly bound by clay minerals, soil organic matter (Lindsay, 1972) and iron oxyhydroxides (White, 1957). Consequently, the availability of Zn to plants is low.

The Zn content of the 11 cores was in the range of 20 to 209 mg/kg, with an average of 79 mg/kg. The maximum Zn content occurred in the surface sample of core 128, which may have received an external contribution of furnace slag, and thereby may have contributed excess zinc to the soil. If this sample is excluded, the maximum Zn content in the remaining 10 cores was 160 mg/kg and the average decreased to 77 mg/kg.

The maximum Zn content occurred in the A horizon in five cores (127, 128, 132, 134, and 136), in the B horizon or below in five cores (130, 131, 133, 135, and 137), and in the 2Oa horizon of core 129. Zinc was correlated with Al_2O_3 (0.51), Fe_2O_3 (0.62), TiO_2 (0.54), Cu (0.81), Pb (0.75), Rb (0.53), V (0.61), and Hg (0.77), which suggest that Zn may have been more commonly associated with iron oxyhydroxides than with clay minerals or organic matter.

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 (Hodson, 2002). The Zr content of soils is generally inherited from the parent materials (Kabata-Pendias, 2001).

The range of Zr contents was 62 to 427 mg/kg, with an average of 242. The Zr content was correlated with SiO_2 (0.62), TiO_2 (0.69), Ba (0.80), and silt-size fraction (0.57). The Zr content was negatively correlated with MgO (-0.66), inorganic carbon (-0.67), sand-size fraction (-0.50), and pH (-0.76). In cores 127 and 132-136 the maximum Zr content was in the surface sample, whereas in the remaining cores the Zr content passed through a subsurface maximum.

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 and their remains are incorporated into the soil column.

In all cores except core 129, the maximum organic C content occurred in the uppermost sample

and decreased with depth. In core 129 there was a buried organic-rich horizon at a depth of approximately four feet. This core was collected at an urban park. Fill material might have been applied to a peaty area to make the land accessible by patrons. The organic carbon content of the peaty layer was 13.98 percent, the highest observed during this study of 137 cores.

Each of the cores penetrated the underlying calcareous till, as evidenced by a sharp increase in inorganic (carbonate) C and the parallel increases in CaO and MgO. The inorganic carbon content typically varied from less than 1 percent above the till to 2 to 6 percent in the till.

CONCLUSIONS

The content of Al_2O_3 was correlated with clay-size content. Because aluminum, in combination with silica in aluminosilicates, is an integral unit of clay minerals, this correlation was expected. The correlation between Fe₂O₃ and clay-size content suggests that clay minerals are coated with iron oxyhydroxides. K₂O was correlated with clay-size content. Potassium is a structural component of the clay mineral, illite. Copper, rubidium, vanadium, zinc, and mercury were also correlated with clay-size content, suggesting adsorption of these trace elements on clay minerals. Copper, nickel, rubidium, vanadium, zinc, and mercury were also correlated with Fe₂O₃ content, an indication that these trace elements were sorbed by iron oxyhydroxides. Iron oxyhydroxides in soil.

The calcareous or dolomitic till or limestone beneath the surficial loess was penetrated by all 11 cores, and in samples selected from the till, sharp increases in CaO, MgO, and inorganic C and decreases in SiO_2 and Al_2O_3 were observed.

Titanium dioxide was correlated with both the silt-size and clay-size fractions, indicating that titanium-bearing minerals occurred in both size fractions. For example, ilmenite and rutile, being resistant to grinding, may have occurred in the silt-size fraction, and anatase, a secondary titanium oxide mineral of small particle size could have occurred in the clay-size fraction.

Some of the mercury content of the soils was probably inherited from the parent materials, but additional amounts probably resulted from atmospheric fallout from industrial sources or other external sources, such as fertilizer application.

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Core Number	County Name	Final Depth of Core (ft)
127	Will	20.0
128	Cook	20.0
129	DuPage	20.0
130	Cook	20.0
131	Lake	20.0
132	McHenry	20.0
133	Kane	20.0
134	DeKalb	17.5
135	Kane	20.0
136	Kendall	13.17
137	Will	20.0

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

Sample	Depth Interval (ft)	Horizon	Clay (%)	Sand (%)	Silt (%)	Texture
127-1	0.0-0.8	А	37.67	6.15	56.18	silty clay loam
127-3	1.1-1.6	Bgtj	46.16	4.59	49.25	silty clay
127-4	1.6-2.2	BCk	42.03	7.27	50.29	silty clay
127-5	2.2-2.8	BCk	42.39	7.74	49.43	silty clay
127-6	2.8-3.2	BCk	42.83	4.59	52.29	silty clay
127-7	3.2-4.0	С	42.63	2.54	54.62	silty clay

Table 2. Texture of samples from core 127*

*Percentages in Tables 2 through 12 are weight-percent.

Table 3. Texture of samples from core 128

Sample	Depth Interval (ft)	Horizon	Clay (%)	Sand (%)	Silt (%)	Texture
128-1	0.0-0.6	A1	17.91	42.98	39.11	loam
128-2	0.6-1.1	A2	26.85	25.86	47.29	loam
128-3	1.1-2.0	Bwcalc	0.76	93.77	5.19	sand
128-4	2.0-2.6	Bwcalc	0.40	95.42	4.00	sand
128-7	3.4-3.7	Ab2	5.20	78.30	16.48	loamy sand
128-9	4.0-4.6	C2	2.56	89.24	8.28	sand

Table 4. Texture of samples from core 129

Sample	Depth Interval (ft)	Horizon	Clay (%)	Sand (%)	Silt (%)	Texture
129-01	0.0-0.6	Bw	30.44	15.50	54.06	silty clay loam
129-04	2.6-3.6	С	28.66	16.33	55.01	silty clay loam
129-05	3.6-4.2	2Oa	32.36	5.86	61.78	silty clay loam
129-06	4.2-4.8	2Bg	30.70	1.08	68.22	silty clay loam
129-07	4.8-5.3	3Bg	24.24	13.04	62.72	silt loam
129-11	6.6-7.3	3C22	28.34	17.76	53.90	silty clay loam

Sample	Depth Interval (ft)	Horizon	Clay (%)	Sand (%)	Silt (%)	Texture
130-1	0.0-0.5	А	35.44	4.30	60.26	silty clay loam
130-3	1.2-2.0	B1	39.35	1.96	58.69	silty clay loam
130-5	2.2-2.8	B2g	39.74	1.61	58.65	silty clay loam
130-8	3.9-4.2	CBg	38.07	2.22	59.71	silty clay loam
130-12	5.1-5.8	CD	27.48	3.48	69.04	silty clay loam
130-13	5.8-6.4	D	28.96	2.94	68.10	silty clay loam

Table 5. Texture of samples from core 130

Table 6. Texture of samples from core 131

Sample	Depth Interval (ft)	Horizon	Clay (%)	Sand (%)	Silt (%)	Texture
131-1	0.0-0.3	A+C1	33.74	18.31	47.95	silty clay loam
131-3	1.1-1.5	C3	48.16	15.72	36.12	clay
131-4	1.5-2.0	2CB	42.84	3.31	53.85	silty clay
131-5	2.0-2.6	2C	45.36	3.33	51.31	silty clay
131-6	2.6-3.4	2C	46.26	2.56	51.18	silty clay
131-7	3.4-4.0	2C	46.32	0.94	52.74	silty clay

Table 7. Texture of samples from core 132

Sample	Depth Interval (ft)	Horizon	Clay (%)	Sand (%)	Silt (%)	Texture
132-1	0.0-0.6	Ap	24.79	12.05	36.16	silt loam
132-2	0.6-1.4	Btj	30.10	22.04	47.86	clay loam
132-3	1.4-2.0	Btj	34.91	4.74	60.35	silty clay loam
132-4	2.0-2.4	2Bt	27.07	42.69	30.24	clay loam
132-5	2.4-2.9	2BC	14.08	40.79	45.13	loam
132-6	2.9-3.2	2C1	7.34	14.30	78.36	silt loam

Sample	Depth Interval (ft)	Horizon	Clay (%)	Sand (%)	Silt (%)	Texture
133-1	0.0-0.9	Ap	17.91	15.20	66.89	silt loam
133-2	1.1-1.7	Btj	27.13	18.06	54.81	silty clay loam
133.3	1.7-2.1	2Bt	24.85	25.24	49.91	loam
133-4	2.1-2.5	2BC	20.62	34.71	44.67	loam
133-5	2.5-3.3	2C1	12.44	32.52	55.04	silt loam
133-6	3.3-4.0	2C2	9.17	37.60	53.23	silt loam

Table 8. Texture of samples from core 133

Table 9. Texture of samples from core 134

Sample	Depth Interval (ft)	Horizon	Clay (%)	Sand (%)	Silt (%)	Texture
134-1	0.0-0.7	Ap	25.13	3.98	70.89	silt loam
134-2	0.7-1.4	B1	31.70	1.63	66.67	silty clay loam
134-3	1.4-2.0	B2tj	33.90	1.60	64.50	silty clay loam
134-4	2.0-2.6	B3t	32.11	5.02	62.87	silty clay loam
134-5	2.6-3.3	B4t	27.21	22.92	49.87	clay loam
134-6	3.3-4.0	С	17.89	36.35	45.76	loam

Table 10. Texture of samples from core 135

Sample	Depth Interval (ft)	Horizon	Clay (%)	Sand (%)	Silt (%)	Texture
135-1	0.0-1.0	Ap	28.77	4.11	67.12	silty clay loam
135-2	1.0-1.5	Btj	37.96	14.56	47.48	silty clay loam
135-3	1.5-2.0	Btj	34.37	37.34	28.29	clay loam
135-6	2.0-2.4	2Bt	44.17	20.20	35.63	clay
135-5	2.4-3.0	2CB	26.88	20.29	52.83	silt loam
135-6	3.0-4.0	2C	19.46	25.20	55.34	silt loam

Sample	Depth Interval (ft)	Horizon	Clay (%)	Sand (%)	Silt (%)	Texture
136-1	0.0-0.8	Ap1	27.06	3.84	69.10	silty clay loam
136-4	1.5-2.0	Btj	34.26	1.43	64.31	silty clay loam
136-6	2.6-3.1	Btj	32.29	6.28	61.43	silty clay loam
136-8	3.6-4.5	2Bt	24.80	51.67	23.53	sandy clay loam
136-9	4.5-4.9	2C1	16.26	44.82	38.92	loam
136-10	4.9-5.4	2C2	16.91	36.00	47.09	loam

Table 11. Texture of samples from core 136

Table 12. Texture of samples from core 137

Sample	Depth Interval (ft)	Horizon	Clay (%)	Sand (%)	Silt (%)	Texture
137-1	0.0-1.0	Ap	31.82	6.44	61.74	silty clay loam
137-2	1.0-1.5	B1t	39.97	3.70	56.33	silty clay loam
137-3	1.5-2.0	B2tj	35.76	7.22	57.02	silty clay loam
137-4	2.0-2.6	B2tj	37.38	13.02	49.60	silty clay loam
137-5	2.6-3.4	B3t	38.90	11.94	49.16	silty clay loam
137-6	3.4-4.0	2C	37.02	11.40	51.58	silty clay loam

Core Number	Texture	Soil Type*	Soil Association	Developmental Environment
127	silty clay loam	Martinton	Martinton-Milford	lacustrine, prairie
128	loam	Alvin	Oakville-Lamont-Alvin	aeolian, forest
129	silty clay loam	Markham	Marley-Blount-Beecher	upland, forest
130	silty clay loam	Mundelein	Plano-Proctor-Worthen	alluvium, prairie
131	silty clay loam	Grays-Markham	St. Charles-Camden-Drury	upland, forest
132	silt loam	Ringwood	Griswold-Ringwood	upland, prairie
133	silt loam	Millbrook	St. Charles-Camden-Drury	upland, forest
134	silt loam	Flanagan	Catlin-Flanagan-Drummer	upland, prairie
135	silty clay loam	Saybrook	Saybrook-Dana-Drummer	upland, prairie
136	silty clay loam	Plattville	Channahon-Dodgeville-Ashdale	upland, prairie
137	silty clay loam	Ashkum	Varna-Elliott-Ashkum	upland, prairie

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

*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.

	Depth	SiO ₂	Al_2O_2	Fe ₂ O ₂	K ₂ O	LOI	Moisture	CaO	MgO
Depth	1								
SiO ₂	-0.48	1							
Al_2O_3	-0.32	0.20	1						
Fe ₂ O ₃	-0.17	0.01	0.83	1					
K ₂ O	0.09	-0.40	0.54	0.59	1				
LOI	0.42	-0.89	-0.49	-0.33	0.05	1			
Moisture	-0.20	0.12	0.52	0.38	0.01	-0.07	1		
CaO	0.54	-0.88	-0.54	-0.36	0.18	0.84	-0.45	1	
MgO	0.52	-0.87	-0.51	-0.28	0.20	0.81	-0.47	0.97	1
Na ₂ O	-0.14	0.77	-0.13	-0.43	-0.58	-0.58	0.01	-0.55	-0.60
TiO ₂	-0.43	0.25	0.85	0.70	0.29	-0.44	0.57	-0.57	-0.54
P_2O_5	-0.36	0.14	0	-0.07	-0.24	0.06	0.32	-0.26	-0.31
MnO	-0.45	0.22	0.38	0.49	0.19	-0.36	-0.03	-0.35	-0.27
Ва	-0.49	0.75	0.58	0.27	-0.20	-0.73	0.41	-0.85	-0.85
Sr	0.12	0.39	-0.20	-0.48	-0.41	-0.26	0.10	-0.20	-0.33
Zr	-0.43	0.62	0.39	0.23	-0.31	-0.60	0.34	-0.68	-0.66
Cr	0.05	0.32	-0.21	-0.35	-0.32	-0.20	-0.05	-0.13	-0.17
Cu	-0.02	-0.13	0.54	0.66	0.35	0.06	0.68	-0.27	-0.24
Ni	0.09	-0.06	0.62	0.71	0.59	-0.13	0.43	-0.25	-0.21
Pb	-0.32	0.07	0.16	0.22	-0.06	0.01	0.22	-0.25	-0.27
Rb	-0.17	-0.25	0.75	0.68	0.84	-0.04	0.32	-0.09	-0.10
V	-0.23	0.07	0.94	0.86	0.47	-0.34	0.62	-0.46	-0.44
Zn	-0.18	0	0.51	0.62	0.26	-0.08	0.44	-0.34	-0.32
Hg	-0.31	0.23	0.41	0.46	-0.10	-0.13	0.63	-0.55	-0.53
Tot C	0.12	-0.35	-0.25	-0.11	-0.30	0.83	0.50	0.06	-0.07
In C	0.39	-0.72	-0.30	-0.11	0.42	0.48	-0.61	0.98	0.97
Org C	-0.09	0.05	-0.07	-0.04	-0.46	0.48	0.74	-0.44	-0.54
Sand	0.14	0.19	-0.67	-0.57	-0.37	0.02	-0.54	0.14	0.14
Silt	-0.07	-0.18	0.32	0.25	0.05	0.09	0.37	-0.04	-0.04
Clay	-0.18	-0.17	0.87	0.78	0.67	-0.13	0.52	-0.19	-0.19
Sand+Silt	0.18	0.13	-0.84	-0.75	-0.62	0.15	-0.53	0.21	0.22
pН	0.39	-0.62	-0.43	-0.39	0.27	0.57	-0.42	0.77	0.76

Table 14. Correlation coefficients for constituents of cores 127 through 137 (C.I. = 95%)

Table 14, continued

	Na ₂ O	TiO ₂	P_2O_5	MnO	Ba	Sr	Zr	Cr	Cu
Na ₂ O	1								
TiO ₂	-0.08	1							
P_2O_5	0.07	0.35	1						
MnO	-0.18	0.50	0.36	1					
Ba	0.54	0.69	0.38	0.35	1				
Sr	0.79	-0.18	0.03	-0.49	0.23	1			
Zr	0.40	0.69	0.43	0.47	0.80	0.13	1		
Cr	0.53	-0.37	-0.17	-0.28	0.03	0.49	-0.17	1	
Cu	-0.34	0.47	0.16	0.01	0.12	-0.20	0.01	-0.27	1
Ni	-0.32	0.37	-0.08	0.22	0.12	-0.26	-0.11	-0.10	0.66
Pb	0.01	0.20	0.34	0.04	0.16	-0.03	0.07	-0.12	0.48
Rb	-0.47	0.59	0.08	0.29	0.11	-0.33	-0.08	-0.32	0.52
V	-0.25	0.90	0.12	0.39	0.51	-0.24	0.43	-0.35	0.65
Zn	-0.19	0.54	0.35	0.18	0.28	-0.14	0.14	-0.27	0.81
Hg	0.01	0.47	0.41	0.18	0.44	-0.10	0.35	-0.15	0.73
Tot C	-0.23	-0.10	0.68	-0.23	-0.15	-0.15	-0.05	-0.18	0.54
In C	-0.45	-0.39	-0.45	0.15	-0.80	-0.24	-0.67	-0.07	-0.48
Org C	0.02	0.10	0.81	-0.28	0.26	-0.02	0.29	-0.12	0.70
Sand	0.34	-0.86	-0.34	-0.32	-0.37	0.20	-0.50	0.50	-0.45
Silt	-0.15	0.65	0.40	0.22	0.32	-0.05	0.57	-0.46	0.25
Clay	-0.47	0.77	0.06	0.28	0.27	-0.31	0.17	-0.36	0.57
Sand+Silt	0.44	-0.78	-0.12	-0.32	-0.28	0.30	-0.21	0.35	-0.53
рН	-0.32	-0.59	-0.39	-0.50	-0.72	-0.07	-0.76	0.06	-0.26

Table 14, continued

	Ni	Pb	Rb	V	Zn	Hg	Tot C	In C	Org C
Ni	1								
Pb	0.14	1							
Rb	0.62	0.19	1						
V	0.64	0.21	0.72	1					
Zn	0.48	0.75	0.53	0.61	1				
Hg	0.40	0.72	0.18	0.51	0.77	1			
Tot C	0.13	0.36	-0.17	0	0.28	0.65	1		
In C	-0.24	-0.34	0.02	-0.28	-0.49	-0.59	-0.01	1	
Org C	0.23	0.48	-0.16	0.14	0.48	0.85	0.87	-0.50	1
Sand	-0.28	-0.09	-0.59	-0.79	-0.45	-0.31	-0.07	0.06	-0.09
Silt	-0.03	0.05	0.22	0.48	0.28	0.22	0.19	-0.13	0.23
Clay	0.59	0.10	0.85	0.89	0.48	0.27	-0.09	0.03	-0.09
Sand+Silt	-0.57	-0.09	-0.82	-0.87	-0.48	-0.30	-0.46	-0.03	0.09
pН	-0.16	-0.20	0.01	-0.48	-0.32	-0.61	0.09	0.75	-0.77

	Sand	Silt	Clay	Sand+Silt	pН	
Sand	1					_
Silt	-0.86	1				
Clay	-0.78	0.38	1			
Sand+Silt	0.77	-0.34	-0.96	1		
pН	0.33	-0.30	-0.19	0.23	1	

Subsample	127-1	127-3	127-4	127-5	127-6	127-7	Average
Lab. No.	R23534	R23535	R223536	R23537	R23538	R23539	
Depth Interval (ft)	0.0-0.8	1.1-1.6	1.6-2.2	2.2-2.8	2.8-3.2	3.2-4.0	
Horizon	А	Bgtj	BCk	BCk	BCk	С	
SiO ₂ (%)	67.09	64.51	55.61	52.09	52.81	53.23	57.56
Al ₂ O ₃ (%)	12.79	14.95	13.38	12.57	13.07	13.26	13.34
Fe ₂ O ₃ (%)	4.56	5.72	4.85	4.38	4.53	4.54	4.76
K ₂ O (%)	2.64	3.25	3.26	3.22	3.39	3.48	3.21
CaO (%)	1.18	1.37	6.05	8.51	7.39	6.84	5.22
MgO (%)	1.49	2.19	3.98	4.17	4.40	4.48	3.45
Na ₂ O (%)	0.70	0.68	0.61	0.54	0.57	0.59	0.62
TiO ₂ (%)	0.74	0.74	0.62	0.57	0.59	0.60	0.64
P ₂ O ₅ (%)	0.11	0.09	0.10	0.09	0.10	0.10	0.10
MnO (%)	0.09	0.11	0.07	0.07	0.08	0.08	0.08
Barium	645	672	599	523	500	589	588
Chromium	71	77	72	63	63	60	68
Copper	28	31	27	27	27	27	28
Mercury (µg/kg)	36	36	24	22	21	20	27
Nickel	31	60	42	39	44	42	43
Lead	39	22	20	18	18	19	23
Rubidium	119	131	124	120	127	130	125
Strontium	121	112	125	122	132	123	123
Vanadium	104	121	109	101	102	99	106
Zinc	97	84	77	72	78	80	81
Zirconium	264	215	170	148	158	158	186
Total C (%)	2.97	0.93	2.59	3.13	2.99	2.96	2.60
Inorganic C (%)	0.11	0.23	2.06	2.81	2.78	2.59	1.76
Organic C (%)	2.86	0.70	0.53	0.32	0.21	0.39	0.84
pН	6.88	7.52	8.06	8.16	8.18	8.16	7.83

Table 15. Elemental composition of samples selected from Core 127

Subsample	128-1	128-2	128-3	128-4	128-7	128-9	Average
Lab. No.	R23540	R23541	R23542	R23543	R23544	R23545	
Depth Interval (ft)	0.0-0.6	0.6-1.1	1.1-2.0	2.0-2.6	3.4-3.7	4.0-4.6	
Horizon	A1	A2	Bwcalc	Bwcalc	Ab2	C2	
SiO ₂ (%)	64.18	71.18	72.72	71.81	69.73	68.76	69.73
Al ₂ O ₃ (%)	9.55	10.81	6.74	6.58	7.26	6.76	7.95
Fe ₂ O ₃ (%)	4.35	3.90	1.19	1.16	1.42	1.28	2.22
K ₂ O (%)	1.97	2.20	2.00	1.97	1.96	1.94	2.01
CaO (%)	2.75	1.36	4.83	5.23	4.53	6.17	4.15
MgO (%)	1.92	1.26	2.98	3.29	2.98	3.67	2.68
Na ₂ O (%)	0.79	0.71	1.07	1.05	0.99	1.00	0.94
TiO ₂ (%)	0.46	0.59	0.14	0.15	0.18	0.16	0.28
P ₂ O ₅ (%)	0.16	0.12	0.05	0.05	0.09	0.05	0.09
MnO (%)	0.06	0.06	0.04	0.04	0.04	0.04	0.05
Barium	546	630	461	491	486	467	514
Chromium	107	123	187	358	756	373	317
Copper	53	33	9	8	10	9	20
Mercury (µg/kg)	115	46	8	7	16	7	33
Nickel	34	36	16	28	31	21	28
Lead	146	43	11	9	12	10	39
Rubidium	80	93	54	52	60	54	66
Strontium	145	117	172	182	165	159	157
Vanadium	74	86	21	18	31	24	42
Zinc	209	96	21	20	20	20	64
Zirconium	188	261	63	64	69	62	118
Total C (%)	7.07	3.21	2.33	2.39	3.67	3.66	3.72
Inorganic C (%)	0.80	0.18	2.00	2.22	1.85	2.62	1.61
Organic C (%)	6.27	3.03	0.33	0.17	1.82	1.04	2.11
pН	7.28	7.28	8.00	8.13	8.08	8.17	7.82

Table 16. Elemental composition of samples selected from Core 128

Subsample	129-1	129-4	129-5	129-6	129-7	129-11	Average
Lab. No.	R23744	R23745	R23746	R23747	R23748	R23749	
Depth Interval (ft)	0.0-0.6	2.6-3.6	3.6-4.2	4.2-4.8	4.8-5.3	6.6-7.3	
Horizon	Bw	С	2Oa	2Bg	3Bg	3C22	
SiO ₂ (%)	58.36	53.87	51.06	65.96	62.94	51.75	57.32
Al ₂ O ₃ (%)	10.42	9.88	10.42	11.98	9.54	9.21	10.24
Fe ₂ O ₃ (%)	4.71	4.54	3.56	4.18	3.46	5.15	4.27
K ₂ O (%)	2.83	2.85	1.98	2.51	2.22	3.02	2.57
CaO (%)	4.40	6.75	2.65	1.71	5.09	8.26	4.81
MgO (%)	3.20	4.14	1.32	1.52	3.71	5.33	3.20
Na ₂ O (%)	0.56	0.56	0.55	0.89	0.89	0.59	0.67
TiO ₂ (%)	0.56	0.51	0.54	0.66	0.55	0.47	0.55
P ₂ O ₅ (%)	0.12	0.09	0.21	0.13	0.13	0.09	0.13
MnO (%)	0.07	0.08	0.03	0.04	0.04	0.08	0.06
Barium	495	405	559	735	548	366	518
Chromium	70	110	58	78	136	82	89
Copper	38	36	76	43	27	30	42
Mercury (µg/kg)	56	39	109	54	29	22	52
Nickel	38	42	57	67	47	38	48
Lead	28	23	38	21	21	20	25
Rubidium	101	97	87	88	73	98	91
Strontium	122	129	125	170	171	113	138
Vanadium	90	83	102	107	91	74	91
Zinc	99	87	107	93	81	94	94
Zirconium	208	181	198	305	314	148	226
Total C (%)	5.16	5.27	14.16	4.37	3.14	4.41	6.09
Inorganic C (%)	1.59	2.51	0.18	0.15	1.86	3.42	1.62
Organic C (%)	3.57	2.76	13.98	4.22	1.28	0.99	4.47
pН	7.65	7.65	6.32	6.82	7.57	7.91	7.32

Table 17. Elemental composition of samples selected from Core 129

Subsample	130-1	130-3	130-5	130-8	130-12	130-13	Average
Lab. No.	R23750	R23751	R23752	R23753	R23754	R23755	
Depth Interval (ft)	0.0-0.5	1.2-2.0	2.2-2.8	3.9-4.2	5.1-5.8	5.8-6.4	
Horizon	А	B1	B2g	CBg	CD	D	
SiO ₂ (%)	65.08	69.50	71.00	69.66	51.27	51.97	62.68
Al ₂ O ₃ (%)	11.18	12.19	12.38	12.96	9.88	10.02	11.49
Fe ₂ O ₃ (%)	4.43	4.65	5.12	3.95	4.32	4.50	4.51
K ₂ O (%)	2.51	2.30	2.60	2.79	2.59	2.69	2.59
CaO (%)	1.50	1.58	1.42	1.83	8.17	7.24	4.05
MgO (%)	1.38	1.57	1.64	1.61	3.99	4.30	2.62
Na ₂ O (%)	0.68	0.78	0.86	0.95	0.67	0.64	0.78
TiO ₂ (%)	0.63	0.67	0.68	0.70	0.49	0.50	0.61
P ₂ O ₅ (%)	0.21	0.11	0.10	0.09	0.10	0.10	0.10
MnO (%)	0.06	0.04	0.03	0.03	0.07	0.07	0.05
Barium	574	615	644	697	424	462	568
Chromium	84	113	306	188	76	71	151
Copper	43	61	43	34	33	36	41
Mercury (µg/kg)	48	41	46	42	27	27	37
Nickel	36	36	30	46	38	37	37
Lead	39	22	22	22	19	21	21
Rubidium	113	87	94	102	94	98	95
Strontium	148	168	168	188	173	156	171
Vanadium	97	99	94	109	91	93	97
Zinc	114	122	122	102	90	91	105
Zirconium	244	270	278	283	182	182	239
Total C (%)	5.39	2.28	0.96	1.63	4.08	3.91	2.57
Inorganic C (%)	0.13	0.10	0.11	0.17	2.79	2.72	1.18
Organic C (%)	5.26	2.18	0.85	1.46	1.29	1.19	1.39
рН	6.48	7.12	7.41	7.70	7.24	7.52	7.40

Table 18. Elemental composition of samples selected from Core 130

Subsample	131-1	131-3	131-4	131-5	131-6	131-7	Average
Lab. No.	R23756	R23757	R23758	R23759	R23760	R23761	
Depth Interval (ft)	0.0-0.3	1.1-1.5	1.5-2.0	2.0-2.6	2.6-3.4	3.4-4.0	
Horizon	A+C1	C3	2CB	2C	2C	2C	
SiO ₂ (%)	58.92	60.25	50.91	43.86	44.79	44.68	50.57
Al ₂ O ₃ (%)	10.50	14.28	12.31	10.89	10.99	10.64	11.60
Fe ₂ O ₃ (%)	4.84	7.18	5.28	4.33	4.50	4.39	5.09
K ₂ O (%)	2.95	3.46	3.51	3.23	3.32	3.30	3.30
CaO (%)	5.35	2.32	6.99	11.75	11.05	11.26	8.12
MgO (%)	3.81	2.98	5.58	6.22	6.29	6.44	5.22
Na ₂ O (%)	0.56	0.44	0.38	0.31	0.36	0.38	0.41
TiO ₂ (%)	0.55	0.66	0.61	0.53	0.53	0.52	0.57
P ₂ O ₅ (%)	0.11	0.09	0.11	0.10	0.09	0.09	0.10
MnO (%)	0.12	0.17	0.09	0.08	0.08	0.07	0.10
Barium	412	454	390	373	425	408	410
Chromium	111	81	87	63	61	68	79
Copper	32	46	36	31	28	30	34
Mercury (µg/kg)	27	39	23	19	18	17	24
Nickel	43	71	44	41	42	36	46
Lead	23	24	20	16	17	16	19
Rubidium	106	139	109	108	110	109	114
Strontium	125	102	98	119	119	128	115
Vanadium	90	122	94	87	91	87	95
Zinc	81	82	60	59	61	59	67
Zirconium	188	204	151	127	127	136	156
Total C (%)	3.37	1.59	3.24	4.45	4.27	4.51	3.57
Inorganic C (%)	2.07	0.68	2.84	4.34	4.03	4.17	3.02
Organic C (%)	1.30	0.91	0.40	0.11	0.24	0.34	0.55
рН	7.92	7.57	8.16	8.18	8.19	8.17	8.03

Table 19. Elemental composition of samples selected from Core 131

Subsample	132-1	132-2	132-3	132-4	132-5	132-6	Average
Lab. No.	R23762	R23763	R23764	R23765	R23766	R23767	
Depth Interval (ft)	0.0-0.6	0.6-1.4	1.4-2.0	2.0-2.4	2.4-2.9	2.9-3.2	
Horizon	Ар	Btj	Btj	2Bt	2CB	2C1	
SiO ₂ (%)	73.83	73.49	70.66	75.34	58.84	52.04	67.37
Al ₂ O ₃ (%)	10.24	11.48	12.77	10.29	7.27	6.81	9.81
Fe ₂ O ₃ (%)	3.86	4.98	5.29	4.66	2.72	2.54	4.01
K ₂ O (%)	2.17	2.11	2.02	2.62	2.16	2.09	2.20
CaO (%)	1.02	1.01	1.00	1.08	7.70	10.12	3.66
MgO (%)	1.10	1.39	1.45	1.37	5.66	6.95	2.99
Na ₂ O (%)	0.73	0.77	0.71	0.68	0.57	0.67	0.69
TiO ₂ (%)	0.66	0.59	0.70	0.49	0.38	0.45	0.55
P ₂ O ₅ (%)	0.19	0.10	0.09	0.11	0.09	0.10	0.11
MnO (%)	0.13	0.10	0.12	0.08	0.06	0.07	0.09
Barium	644	651	728	568	416	398	568
Chromium	102	112	117	227	33	33	104
Copper	21	30	28	30	18	17	24
Mercury (µg/kg)	58	39	44	34	21	18	36
Nickel	36	42	39	47	20	16	33
Lead	24	18	21	17	11	12	17
Rubidium	82	70	74	77	58	51	69
Strontium	131	134	123	125	113	127	126
Vanadium	89	98	112	78	49	47	79
Zinc	90	62	71	58	38	34	59
Zirconium	383	307	378	202	152	229	275
Total C (%)	1.92	0.77	0.88	0.63	3.77	4.81	2.13
Inorganic C (%)	0.11	0.12	0.12	0.08	3.65	4.75	1.47
Organic C (%)	1.81	0.65	0.76	0.55	0.12	0.06	0.66
рН	6.14	7.07	6.63	7.15	8.14	8.20	7.22

Table 20. Elemental composition of samples selected from Core 132

Subsample	133-1	133-2	133-3	133-4	133-5	133-6	Average
Lab. No.	R23768	R23769	R23770	R23771	R23772	R23773	
Depth Interval (ft)	0.0-0.9	0.90-1.4	1.4-2.0	2.0-2.8	4.0-4.6	5.7-6.4	
Horizon	Ар	Btj	2Bt	2BC	2C1	2C2	
SiO ₂ (%)	76.86	73.48	74.33	75.62	59.08	56.81	69.36
Al ₂ O ₃ (%)	9.56	11.65	11.21	10.66	7.81	6.96	9.64
Fe ₂ O ₃ (%)	3.01	4.79	4.65	4.15	2.66	2.29	3.59
K ₂ O (%)	2.69	2.75	2.81	2.87	2.26	2.15	2.59
CaO (%)	0.99	0.95	1.02	1.15	7.46	8.67	3.37
MgO (%)	0.99	1.23	1.22	1.19	5.35	6.06	2.67
Na ₂ O (%)	0.82	0.78	0.83	0.88	0.72	0.69	0.79
TiO ₂ (%)	0.62	0.58	0.54	0.50	0.36	0.32	0.49
P ₂ O ₅ (%)	0.12	0.08	0.09	0.12	0.11	0.09	0.10
MnO (%)	0.11	0.09	0.11	0.10	0.07	0.07	0.09
Barium	668	635	617	613	482	458	579
Chromium	43	80	90	116	96	69	82
Copper	17	28	26	23	16	14	21
Mercury (µg/kg)	28	38	38	26	21	16	28
Nickel	21	29	33	36	30	18	28
Lead	20	19	17	17	13	11	16
Rubidium	84	84	82	79	62	54	74
Strontium	143	133	135	146	131	130	136
Vanadium	68	86	79	69	48	40	65
Zinc	51	58	56	52	38	32	48
Zirconium	410	370	337	341	245	225	321
Total C (%)	1.49	0.93	0.48	0.51	3.55	4.07	1.84
Inorganic C (%)	0.19	0.12	0.13	0.12	3.32	4.06	1.32
Organic C (%)	1.30	0.81	0.35	0.39	0.23	0.01	0.52
рН	6.23	6.66	6.92	7.25	8.16	8.22	7.24

Table 21. Elemental composition of samples selected from Core 133

Subsample	134-1	134-2	134-3	134-4	134-5	134-6	Average
Lab. No.	R23774	R23775	R23776	R23777	R23778	R23779	
Depth Interval (ft)	0.0-0.7	0.7-1.4	1.4-2.0	2.0-2.6	2.6-3.3	3.3-4.0	
Horizon	Ар	B1	B2tj	B3t	B4t	С	
SiO ₂ (%)	71.59	70.98	70.42	71.33	74.00	55.54	68.98
Al ₂ O ₃ (%)	10.52	12.72	13.22	12.54	11.15	7.82	11.33
Fe ₂ O ₃ (%)	3.70	4.91	5.28	5.19	4.82	2.92	4.47
K ₂ O (%)	2.09	2.04	2.04	2.14	2.56	2.38	2.21
CaO (%)	1.36	0.98	1.02	1.20	1.07	8.48	2.35
MgO (%)	1.28	1.34	1.45	1.52	1.34	5.99	2.15
Na ₂ O (%)	0.80	0.78	0.82	0.88	0.81	0.52	0.77
TiO ₂ (%)	0.72	0.72	0.70	0.68	0.59	0.38	0.63
P ₂ O ₅ (%)	0.26	0.10	0.11	0.11	0.11	0.08	0.13
MnO (%)	0.17	0.10	0.11	0.12	0.11	0.06	0.11
Barium	753	751	676	701	681	404	661
Chromium	122	109	382	94	144	82	156
Copper	23	23	28	28	28	20	25
Mercury (µg/kg)	34	50	55	39	40	18	39
Nickel	22	27	41	34	61	27	35
Lead	23	20	20	20	17	12	19
Rubidium	87	81	78	77	80	70	79
Strontium	130	141	141	142	134	112	133
Vanadium	84	103	112	107	88	51	91
Zinc	101	71	73	71	66	44	71
Zirconium	428	413	404	417	307	146	353
Total C (%)	2.55	1.04	0.69	0.52	0.47	3.96	1.54
Inorganic C (%)	0.23	0.08	0.09	0.08	0.14	3.94	0.76
Organic C (%)	3.32	0.96	0.60	0.44	0.33	0.02	0.95
рH	6.80	5.81	5.59	5.87	6.82	8.22	6.52

Table 22. Elemental composition of samples selected from Core 134

Subsample	135-1	135-2	135-3	135-4	135-5	135-6	Average
Lab. No.	R23790	R23791	R23792	R23793	R23794	R23795	
Depth Interval (ft)	0.0-1.0	1.0-1.5	1.5-2.0	2.0-2.4	2.4-3.0	3.0-4.0	
Horizon	Ар	Btj	Btj	2Bt	2CB	2C	
SiO ₂ (%)	72.36	69.01	70.01	63.86	49.29	40.21	60.79
Al ₂ O ₃ (%)	11.32	13.36	12.42	14.46	10.11	7.44	11.52
Fe ₂ O ₃ (%)	3.94	5.92	6.38	8.28	6.15	4.06	5.79
K ₂ O (%)	2.25	2.49	2.89	3.42	3.07	2.37	2.75
CaO (%)	1.03	0.93	0.85	0.97	7.71	12.30	3.97
MgO (%)	1.14	1.48	1.49	1.82	6.05	8.98	3.49
Na ₂ O (%)	0.80	0.60	0.57	0.46	0.40	0.33	0.53
TiO ₂ (%)	0.72	0.65	0.55	0.64	0.51	0.38	0.58
P ₂ O ₅ (%)	0.16	0.09	0.08	0.11	0.10	0.09	0.11
MnO (%)	0.14	0.12	0.12	0.14	0.11	0.10	0.12
Barium	720	627	595	620	405	365	555
Chromium	61	96	74	83	56	49	70
Copper	23	36	39	51	39	31	37
Mercury (µg/kg)	44	50	57	59	42	32	47
Nickel	27	53	50	81	47	37	49
Lead	24	22	26	31	27	19	25
Rubidium	89	94	99	122	98	73	96
Strontium	126	112	113	98	95	87	105
Vanadium	91	109	107	120	89	62	96
Zinc	77	89	101	138	103	86	99
Zirconium	402	262	189	192	138	111	216
Total C (%)	1.86	1.03	0.86	1.00	4.20	6.23	2.53
Inorganic C (%)	0.06	0.08	0.09	0.22	3.93	6.06	1.74
Organic C (%)	1.80	0.95	0.77	0.78	0.27	0.17	0.79
рН	6.38	6.56	6.54	6.83	6.90	8.02	6.87

Table 23. Elemental composition of samples selected from Core 135

Subsample	136-1	136-4	136-6	136-8	136-9	136-10	Average
Lab. No.	R23796	R23797	R23798	R23799	R23800	R23801	
Depth Interval (ft)	0.0-0.8	1.5-2.0	2.6-3.1	3.6-4.5	4.5-4.9	4.9-5.4	
Horizon	Ap1	Btj	Btj	2Bt	2C1	2C2	
SiO ₂ (%)	71.56	70.44	70.75	77.57	54.77	45.43	65.09
Al ₂ O ₃ (%)	9.88	13.09	12.87	9.35	7.66	7.17	10.00
Fe ₂ O ₃ (%)	3.45	5.11	5.60	4.45	3.53	3.45	4.27
K ₂ O (%)	2.18	2.29	2.45	2.27	2.25	2.21	2.28
CaO (%)	1.95	1.99	1.26	0.84	0.62	0.52	1.20
MgO (%)	1.03	1.41	1.38	1.20	6.24	7.96	3.20
Na ₂ O (%)	0.82	0.79	0.83	0.68	0.50	0.43	0.68
TiO ₂ (%)	0.69	0.74	0.66	0.45	0.37	0.38	0.55
P ₂ O ₅ (%)	0.31	0.10	0.10	0.10	0.10	0.11	0.14
MnO (%)	0.17	0.11	0.13	0.09	0.08	0.07	0.11
Barium	785	763	658	506	388	369	578
Chromium	62	64	120	59	47	41	66
Copper	24	23	32	26	26	22	26
Mercury (µg/kg)	50	51	46	34	25	21	38
Nickel	26	28	40	43	27	20	31
Lead	34	22	22	17	15	14	21
Rubidium	93	81	79	71	64	62	75
Strontium	135	122	143	126	104	107	123
Vanadium	79	113	107	76	57	56	81
Zinc	100	73	74	62	63	62	72
Zirconium	388	377	382	253	173	164	290
Total C (%)	3.55	0.83	0.56	0.56	4.14	5.59	2.54
Inorganic C (%)	0.47	0.13	0.10	0.12	3.86	5.25	1.66
Organic C (%)	3.08	0.70	0.46	0.44	0.28	0.34	0.88
рН	6.02	6.26	5.48	6.25	8.08	8.18	6.71

Table 24. Elemental composition of samples selected from Core 136

Subsample	137-1	137-2	137-3	137-4	137-5	137-6	Average
Lab. No.	R23802	R23803	R23804	R23805	R23806	R23807	
Depth Interval (ft)	0.0-1.0	1.0-1.5	1.5-2.0	2.0-2.6	2.6-3.4	3.4-4.0	
Horizon	Ар	B1t	B2tj	B2tj	B3t	2C	
SiO ₂ (%)	71.19	69.29	69.37	66.19	65.34	57.84	66.54
Al ₂ O ₃ (%)	11.10	13.43	13.39	14.61	14.74	12.96	13.37
Fe ₂ O ₃ (%)	4.10	5.50	5.90	7.02	7.44	5.19	5.86
K ₂ O (%)	2.32	2.18	2.35	3.42	3.66	3.30	2.87
CaO (%)	1.31	1.05	0.99	0.80	0.83	4.77	1.63
MgO (%)	1.33	1.51	1.49	1.65	1.74	4.07	1.97
Na ₂ O (%)	0.72	0.72	0.76	0.58	0.57	0.57	0.65
TiO ₂ (%)	0.71	0.73	0.70	0.72	0.70	0.60	0.69
P ₂ O ₅ (%)	0.18	0.08	0.08	0.07	0.08	0.10	0.10
MnO (%)	0.13	0.06	0.08	0.14	0.09	0.09	0.10
Barium	692	682	671	573	654	576	641
Chromium	73	71	90	82	84	66	78
Copper	23	31	37	58	61	38	41
Mercury (µg/kg)	29	39	46	61	61	32	45
Nickel	29	27	32	61	70	48	45
Lead	31	22	27	36	37	22	29
Rubidium	99	87	87	118	126	112	105
Strontium	129	131	123	105	104	117	118
Vanadium	90	117	117	126	124	98	112
Zinc	82	74	81	139	160	88	104
Zirconium	369	374	369	219	199	189	287
Total C (%)	2.20	0.93	0.68	0.88	0.91	2.37	1.33
Inorganic C (%)	0.17	0.13	0.14	0.11	0.11	1.98	0.44
Organic C (%)	2.05	0.80	0.54	0.77	0.80	0.39	0.89
рН	7.32	7.00	6.87	6.69	7.04	8.03	7.16

Table 25. Elemental composition of samples selected from Core 137

	This	s Work	Shacklette and Boerngen ⁽²⁾		
Element	Mean	Range	Mean	Range	
Silicon (%)	31.9	27.3-35.9	NR [†]	29-45**	
Aluminum (%)	5.63	5.05-6.77	NR	<0.07-8.5**	
Iron (%)	2.84	2.10-3.38	2.1 ⁽¹⁾	0.1-4.0**	
Potassium (%)	2.01	1.64-2.45	1.7 ⁽¹⁾	0.22-2.25**	
Calcium (%)	1.44	0.70-3.82	0.40 ⁽¹⁾	0.3-1.5**	
Magnesium (%)	1.02	0.58-2.30	0.30 ⁽¹⁾	0.005-1.25**	
Sodium (%)	0.53	0.42-0.61	NR	0.6-1.25**	
Titanium (%)	0.38	0.28-0.44	0.41	0.05-1.0	
Phosphorus (%)	0.07	0.05-0.14	0.065 ⁽¹⁾	0.013-0.68**	
Manganese (%)	0.08	0.05-0.13	0.052	0.005-0.15	
Barium (mg/kg)	630	412-785	675	200-1500	
Chromium (mg/kg)	83	43-122	55	10-100	
Copper (mg/kg)	30	17-53	25	7-100	
Mercury (µg/kg)	48	27-115	70***	20-360***	
Nickel (mg/kg)	31	21-43	17	5-30	
Lead (mg/kg)	39	20-146	19	10-30	
Rubidium (mg/kg)	96	80-119	75	45-100	
Strontium (mg/kg)	132	121-148	305	20-1000	
Vanadium (mg/kg)	87	68-104	87	20-150*	
Zinc (mg/kg)	100	51-209	58.5	20-109	
Zirconium (mg/kg)	316	188-428	NR	NR	

Table 26. Means and ranges of elemental contents of the uppermost samples from Illinois soil cores 127 through 137 compared with results for surface soils as determined by other researchers

[†]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.

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Figure 1. Loess thickness in Illinois (after Willman and Frye, 1970).



Figure 2. Quaternary deposits in Illinois (after Lineback, 1981).



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



Figure 4. Locations of soil cores collected from 1998 through 2003.



Figure 5. pH versus depth in cores 127 through 137, collected during 2003.



Figure 6. Element concentrations versus depth in core 127.



Figure 7. Element concentrations versus depth in core 127.



Figure 8. Element concentrations versus depth in core 128.



Figure 9. Element concentrations versus depth in core 128.



Figure 10. Element concentrations versus depth in core 129.



Figure 11. Element concentrations versus depth in core 129.



Figure 12. Element concentrations versus Depth in core 130.



Figure 13. Element concentrations versus Depth in core 130.


Figure 14. Element concentrations versus depth in core 131.



Figure 15. Element concentrations versus depth in core 131.



Figure 16. Element concentrations versus depth in core 132.



Figure 17. Element concentrations versus depth in core 132.



Figure 18. Element concentrations versus depth in core 133.



Figure 19. Element concentrations versus depth in core 133.



Figure 20. Element concentrations versus depth in core 134.



Figure 21. Element concentrations versus depth in core 134.



Figure 22. Element concentrations versus depth in core 135.



Figure 23. Element concentrations versus depth in core 135.



Figure 24. Element concentrations versus depth in core 136.



Figure 25. Element concentrations versus depth in core 136.



Figure 26. Element concentrations versus depth in core 137.



Figure 27. Element concentrations versus depth in core 137.



CaO versus MgO

Figure 28. Calcium oxide content versus magnesium oxide content of cores 127 through 137.