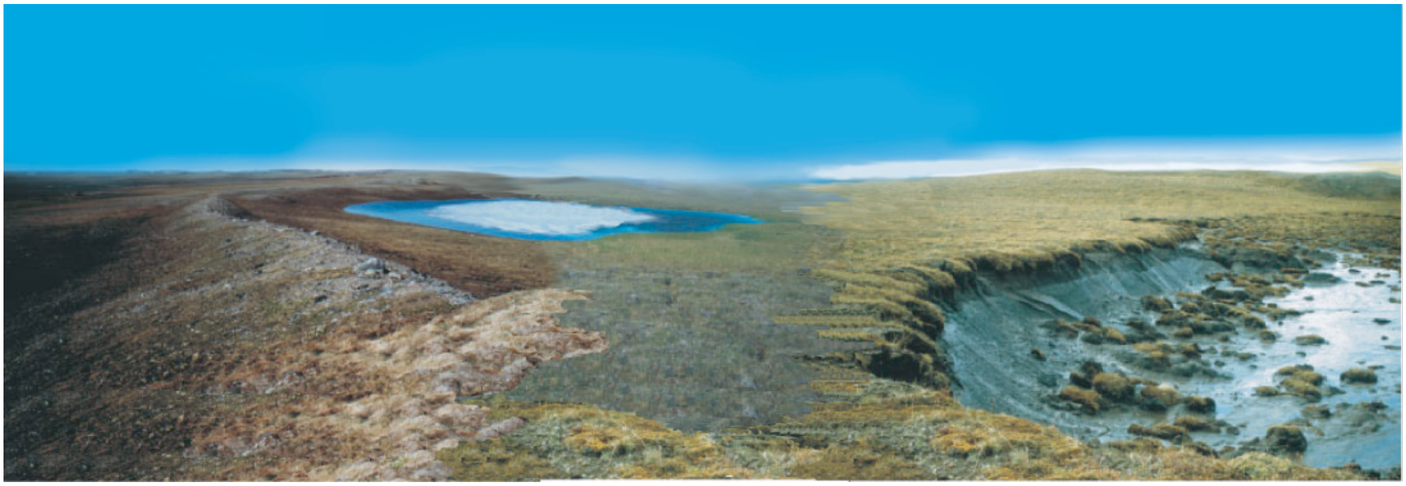
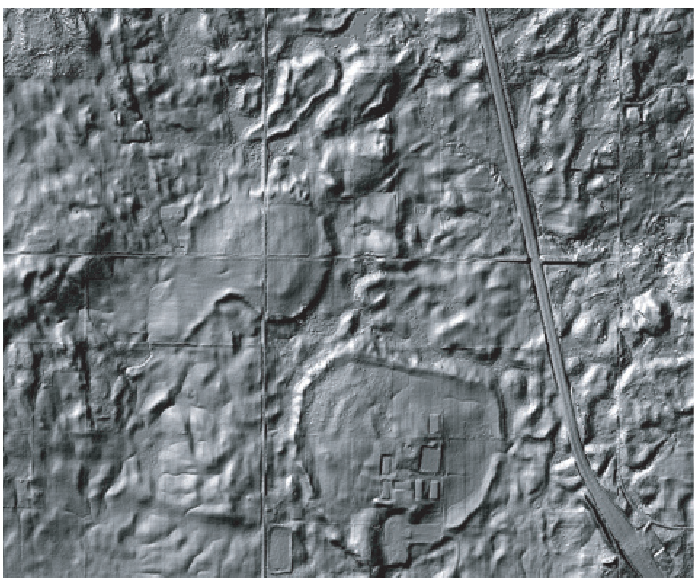
State of Illinois
Rod. R. Blagojevich, Governor
Illinois Department of Natural Resources
Illinois State Geological Survey

Deglacial History and Paleoenvironments of Northeastern Illinois

54th Midwest Friends of the Pleistocene Field Conference





May 16–18, 2008
DeKalb, Illinois

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Midwest Friends of the Pleistocene

Meetings of the Midwest Friends of the Pleistocene*

1 1950 Eastern Wisconsin

2 1951 Southeastern Minnesota
H.E. Wright Jr. and R.V. Ruhe

3 1952 Western Illinois and eastern Iowa
P.R. Shaffer and W.H. Scholtes

4 1953 Northeastern Wisconsin
F.T. Thwaites

5 1954 Central Minnesota
H.E. Wright, Jr., and A.F. Schneider

6 1955 Southwestern Iowa
R.V. Ruhe

7 1956 Northwestern lower Michigan
J.H. Zumberge and W.N. Melhorn

8 1957 South-central Indiana
W.D. Thornbury and W.J. Wayne

9 1958 Eastern North Dakota
W.M. Laird and others

10 1959 Western Wisconsin
R.F. Black

11 1960 Eastern South Dakota
A.G. Agnew and others

12 1961 Eastern Alberta
C.P. Gravenor and others

13 1962 Eastern Ohio
R.P. Goldthwait

14 1963 Western Illinois
J.C. Frye and H.B. Willman

15 1964 Eastern Minnesota
H.E. Wright, Jr. and E.J. Cushing

16 1965 Northeastern Iowa
R.V. Ruhe and others

17 1966 Eastern Nebraska
E.C. Reed and others

18 1967 South-central North Dakota
Lee Clayton and T.F. Freers

19 1969 Cyprus Hills, Saskatchewan and Alberta
W.O. Kupsch

20 1971 Kansas and Missouri Border
C.K. Bayne and others

21 1972 East-central Illinois
W.H. Johnson, L.R. Follmer and others

22 1973 West-central Michigan & east-central Wisconsin
E.B. Evenson and others

23 1975 Western Missouri
W.H. Allen and others

24 1976 Meade County, Kansas
C.K. Bayne and others

25 1978 Southwestern Indiana
R.V. Ruhe and C.G. Olson

26 1979 Central Illinois
L.R. Follmer, E.D. McKay III and others

27 1980 Yarmouth, Iowa
G.R. Hallberg and others

28 1981 Northeastern lower Michigan
W.A. Burgis and D.F. Eschman

29 1982 Driftless Area, Wisconsin
J.C. Knox and others

30 1983 Wabash Valley, Indiana
N.K. Bleuer and others

31 1984 West-central Wisconsin
R.W. Baker

32 1985 North-central Illinois
R.C. Berg and others

33 1986 Northeastern Kansas
W.C. Johnson and others

34 1987 North-central Ohio
S.M. Totten and J.P. Szabo

35 1988 Southwestern Michigan
G.J. Larson and G.W. Monaghan

36 1989 Northeastern South Dakota
J.P. Gilbertson

37 1990 Southwestern Iowa
E.A. Bettis III and others

38 1991 Mississippi Valley, Missouri and Illinois
E.R. Hajic W.H. Johnson and others

39 1992 Northeastern Minnesota
J.D. Lehr and H.C. Hobbs

40 1993 Door Peninsula, Wisconsin
A.F. Schneider and others

41 1994 Eastern Ohio and western Indiana
T.V. Lowell and C.S. Brockman

42 1995 Southern Illinois and southeast Missouri
S.P. Esling and M.D. Blum

43 1996 Eastern North Dakota & northwestern Minnesota
K.I. Harris and others

44 1998 North-central Wisconsin
J.W. Attig and others

45 1999 North-central Indiana & south-central Michigan
S.E. Brown, T.G. Fisher and others

46 2000 Southeast Nebraska and Northeast Kansas
R.D. Mandel and E.A. Bettis III

47 2001 NorthwestW Ontario and Northeast Minnesota
B.A.M. Phillips and others

48 2002 East-Central Upper Michigan
W.L. Loope and J.B. Anderton

49 2003 Southwest Michigan
B.D. Stone, K.A. Kincare and others

50 2004 Central Minnesota
A.R. Knaeble, G.N. Meyer and others

51 2005 North-central Illinois
E.D. McKay III, R.C. Berg and others

52 2006 Northwest-Central North Dakota
L. A. Manz

53 2007 East-Central Wisconsin
T. S. Hooyer

54 2008 Northeastern Illinois
B. B. Curry

* No meetings were held in 1968, 1970, 1974, 1977, and 1997.

On cover: *Top image is digitally joined pictures of two ice-walled lakes in the Tamyr Peninsula, Russia (photos courtesy of Helena Alexanderson). The shaded relief map is of rimmed ice-walled lake plains on the Tinley Moraine near Wadsworth, Illinois (LiDAR data courtesy of the Lake County Department of Information and Technology, GIS/Mapping Division). The flower is* Dryas integrifolia *by Pierre Brousseau from the Web site of Environment Canada, Imagier Flowers 4 Gallery.*

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Deglacial History and Paleoenvironments of Northeastern Illinois

54th Midwest Friends of the Pleistocene Field Conference:

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Edited by Brandon Curry

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CONTENTS

Acknowledgements

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INTRODUCTION

The surficial geology of northeastern Illinois encompasses Paleozoic shale, dolomite, limestone, sandstone, and coal that are covered by as much as 120 m of Pleistocene diamicton, sand and gravel, and silt and clay. We will focus our trip through landscapes formed during the last glaciation (Wisconsin Episode). The study area is framed on the north and west by the Bloomington Morainic System, and to the south by the Illinois and Des Plaines rivers (Figures 1 and 2). Throughout this area, landforms of glacial origin dominate the landscape. In most places, moraines that were formed largely by subglacial processes are partly mantled by supraglacial sediment, such as lake sediment, flowed till, and ice-contact sand and gravel of ice-walled lake plains. In some areas, the glacial landforms are pockmarked by kettle basins caused by collapse by melting ice and groundwater sapping. More rare on the landscape, but nevertheless important to the glacial history, are valleys that slash across several moraines. The steep slopes of the valley walls and large-scale fluvial landforms attest to the erosive power of floods caused by proglacial lakes that overtopped their moraine dams. In other areas, anastomosing valleys with underfit streams appear to have been formed by subglacial drainage. Low areas on the landscape of either fluvial or supraglacial origin were partly filled with sediment, at first by resedimented silty loess, and subsequently by post-glacial biogenic deposits of peat and marl.

Technological advancements over the past 20 years or so have improved our precision and presentation in mapping, and tightened our chronological control. Recently acquired shaded relief maps using LiDAR (Light Detection and Ranging) or high-resolution photogrammetric techniques have increased our awareness of subtle landforms and interest in mapping and investigating the near-surface geology of this area. New computer software and technology, in many cases aided by geophysical data, helps us to visualize three-dimensional relationships among geologic bodies and speculate on their genesis. Isotope and trace element assay of very small samples is now possible with mass spectrometers. For example, radiocarbon assay with accelerator mass spectrometry (AMS) uses samples containing as little as about 10 micrograms of carbon (a few spruce needles or seeds of terrestrial plants). Similarly, we can determine the $\delta^{18}O$ values of biogenic carbonate samples as small as 50 micrograms (e.g., 2 to 10 ostracode valves about one millimeter across). These data may be used in a variety of paleoenvironmental reconstructions, such as in the assessment of water source in proglacial lakes (was the water derived from primarily meltwater or precipitation?). Our new AMS ages not only allow revision of our local chronologies, but we are in much better position to compare our events with well-dated global records of climate change. For mapping in this region, these collective computer and lab technologies have aided, along with old-fashioned coring, near-surface geophysics, and airphoto interpretation, in the identification of subtle ice-walled lake plains, and the age of tundra plant fossils contained in the laminated sediment therein.

Northeastern Illinois has long been the focus of research of sediments of the last glaciation. The findings of applied studies have contributed to our evolving understanding of the “pure science” of glacial sedimentology and stratigraphy. These applied investigations have involved topics such as water supply, water quality, aggregate resources, wetland evaluation, climate change, landfill siting, and landfill leachate migration. As urban growth continues in this region, applied studies will remain as an important means of assuring the public’s health and safety. The Northern Illinois Planning Commission projects that from 2000 to 2030 the population of the six largest counties in northeastern Illinois will increase from about 8.1 to 10.0 million (an increase of 25%; Figure 3). This doesn’t count Kendall County, where two of our stops are located, that is currently the fastest growing county in the United States (Keen, 2008). The current population is about 100,000… up 77% since the year 2000.

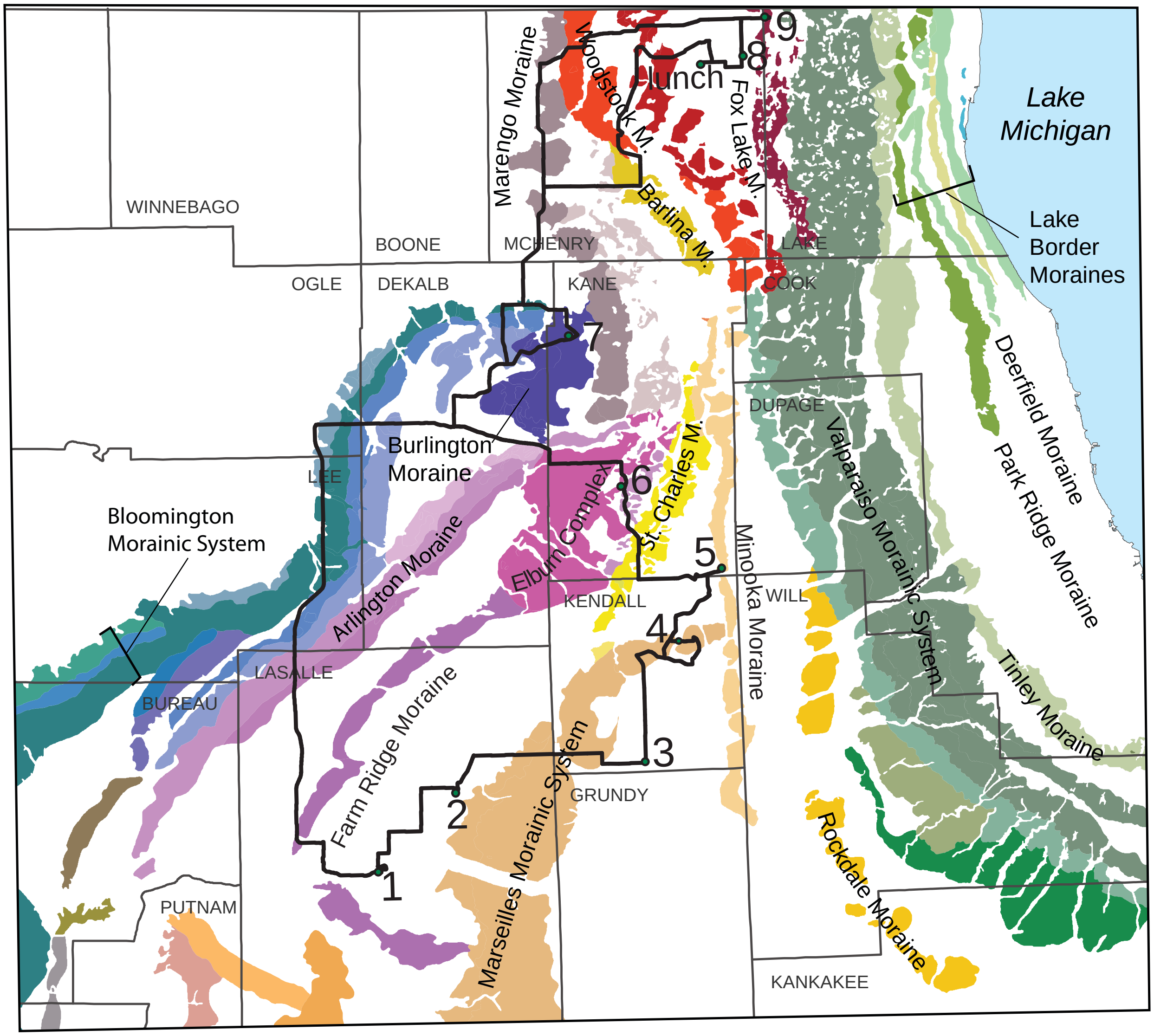


Figure 1 Moraines and morainic complexes of the last glaciation in northeastern Illinois, highlighting the study area and field trip route.

The glacial succession at Fermi National Laboratory, Batavia, Illinois, was among the first to be shown to be comprised of sediment layers with unique physical properties that may be traced in 3 dimensions (Landon and Kempton, 1971). Particle-size distribution, moisture content, and Atterberg Limit data from this study's cores provided validation that a well-characterized stratigraphic framework was necessary for meaningful future applied studies (Kempton, 1981; Graese et al., 1988; Berg et al., 1999).

For deposits of the last glaciation in northeastern Illinois, the current stratigraphic framework is a reclassification by Hansel and Johnson (1996) of the independent multiple classification system of Willman and Frye (1970). Included among the former's most important features and contributions:

1) Stratigraphic upgrades and revisions (i.e., Wedron Formation = Wedron Group; Tiskilwa Till Member = Tiskilwa Formation, Wadsworth Till Member = Wadsworth Formation, and

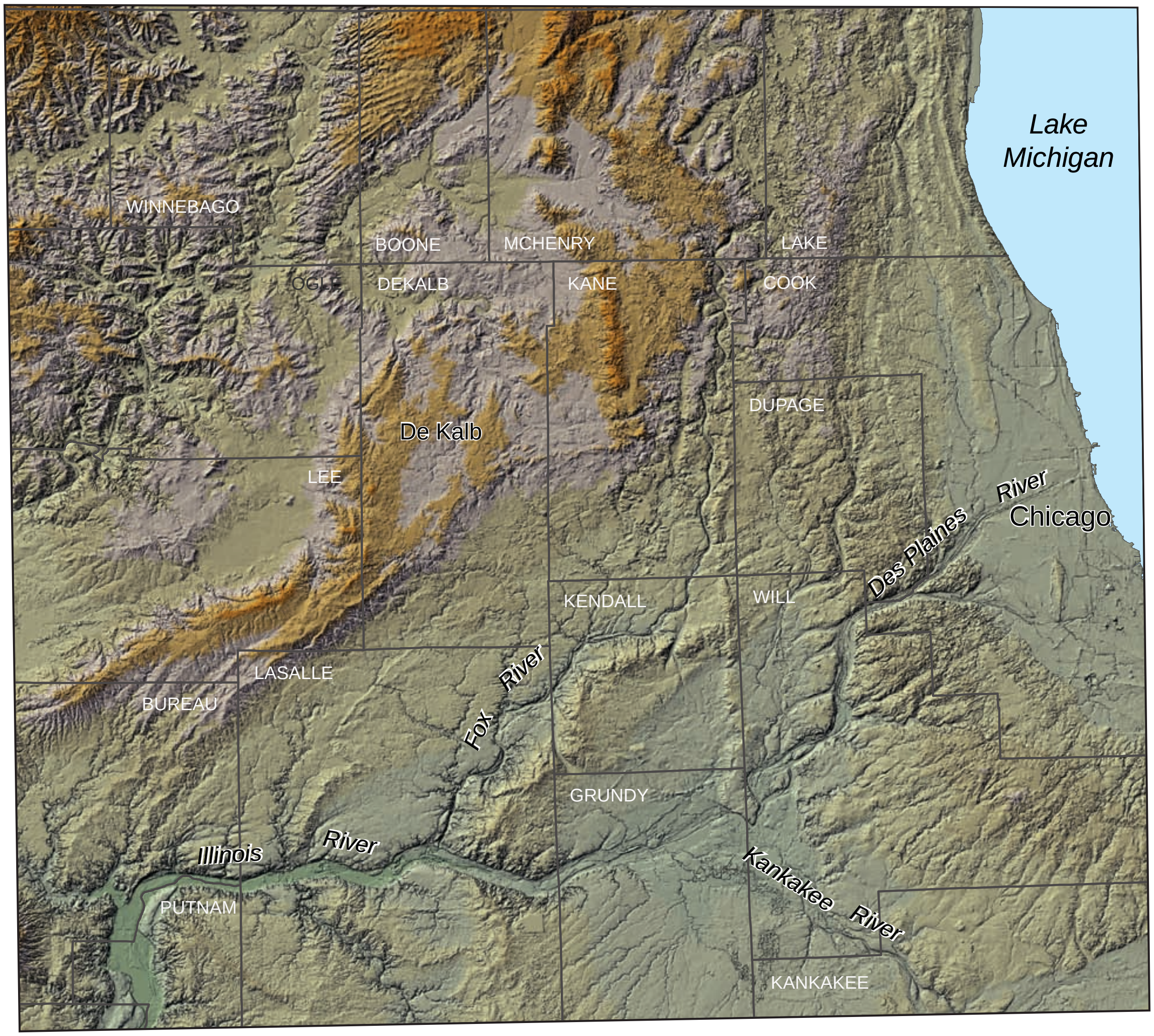


Figure 2 Painted relief map of northeastern Illinois; modified from Luman et al. (2003)

establishment of the Lemont Formation, comprised of the Batestown, Yorkville, and Haeger Members).

2) Definition of the Mason Group, comprised of units of sorted sediment that in many places interfinger (tongue) with units of the Wedron Group.

3) The replacement of chronostratigraphic (time) units with diachronic units based on the lithostratigraphic reclassification outlined above (Hansel and Johnson, 1992; Hansel et al., 1997).

4) Abolishment of the genetic terms “till” and “loess” in the names of stratigraphic units.

5) Organization of pertinent radiocarbon ages by stratigraphic unit and location. Many ages were previously unpublished or in the “gray literature”.

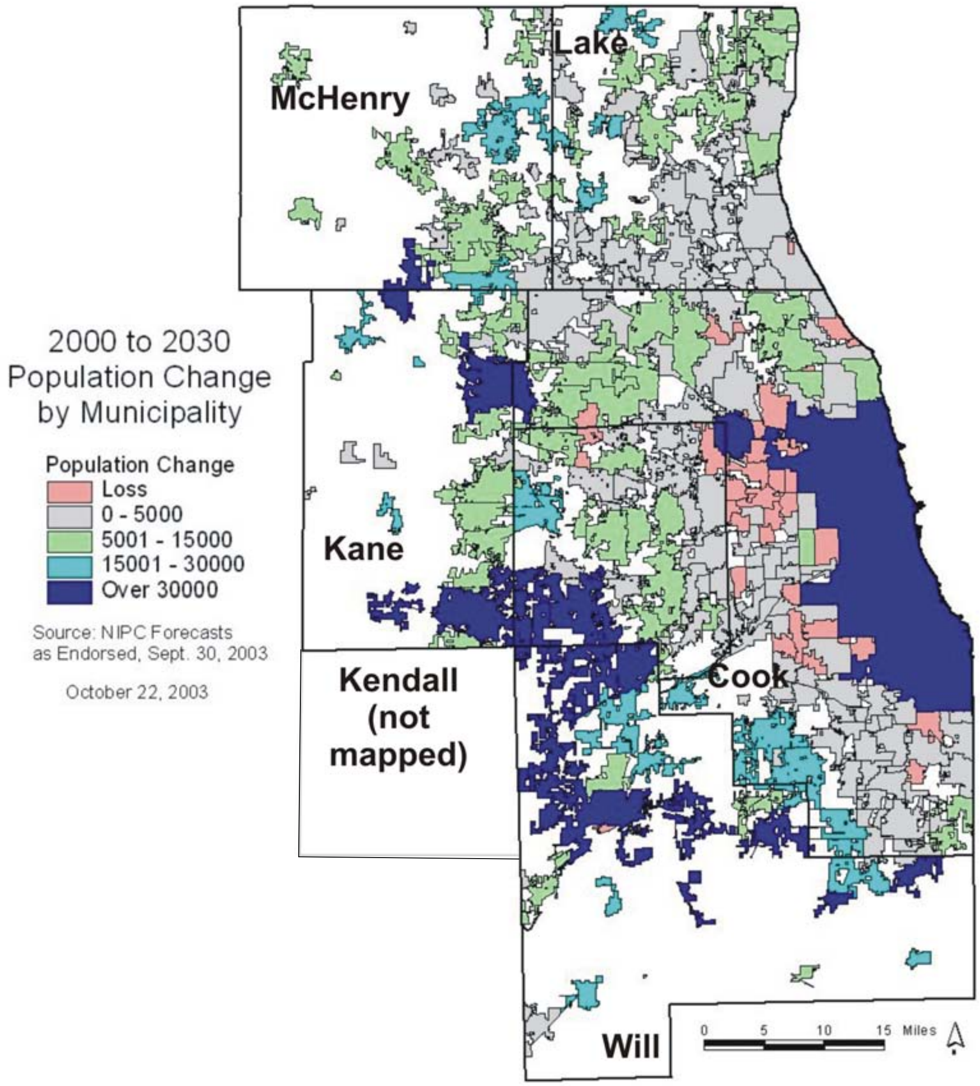


Figure 3 Projected growth areas in northeastern Illinois (from the Northern Illinois Planning Commission (NIPC)). Kendall County, not part of NIPC, is presently the fastest growing county in the United States (Keen, 2008).

Presently, the ISGS is engaged in an ambitious program of mapping the surficial geology of the entire state at a scale of 1:24,000 (ISGS, 1992; Bhagwat, 2001; Figure 4). Begun in earnest in the late 1990's, geologic mapping has been funded by federal and state programs (STATEMAP, EDMAP, and CGLGMC (The Central Great Lakes Geologic Mapping Coalition)) and by various county agencies, particularly in Kane, Lake, McHenry, DuPage, DeKalb, and Kendall counties. From start to finish, production of one 1:24,000 scale surficial geology map in northeastern Illinois costs, on average, about $250,000. The dollar amount is noteworthy because mitigation of emerging environmental and health issues, often dealing with water quality or water availability, may be addressed with geologic maps. Having up-to-date and accurate surficial geology maps and three-dimensional geology maps can potentially save taxpayers roughly ten times the cost of the original map (Bhagwat and Berg, 1992; Bhagwat and Ipe, 2000).

Today's applied studies are able to utilize the stratigraphic framework in applied studies by rendering the geology in three dimensions via digital models, cross sections, structure contour

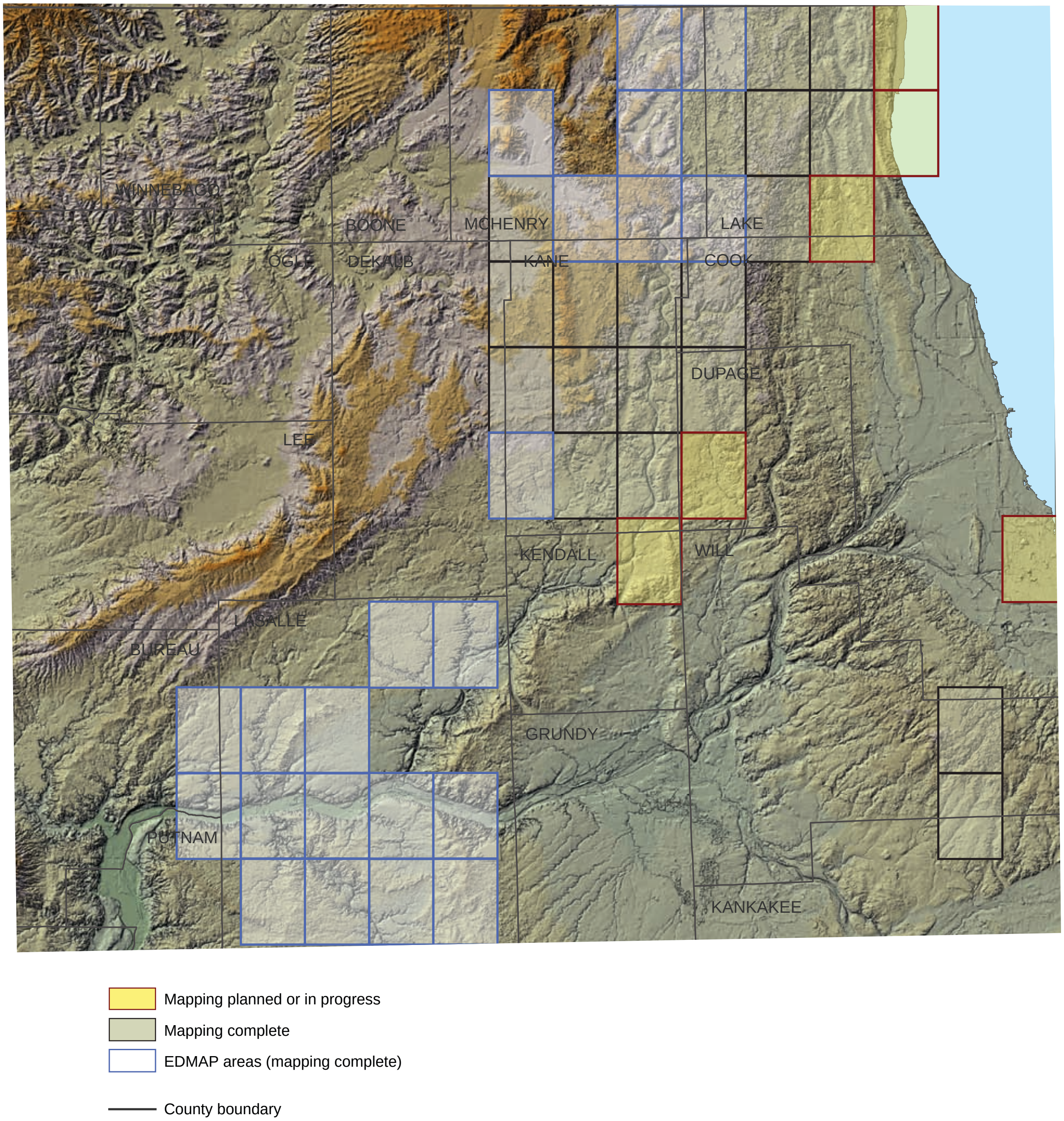


Figure 4. Location of quadrangles with published surficial geology maps.

maps, and other visualizations. Given key stratigraphic points in the area of interest, geologists may now explore issues such as the variability of channel sinuosity, width, and depth given an interpreted environment of deposition. The three-dimensional model may be rendered for use in digital flow models by hydrogeologists to evaluate groundwater availability. In Kane County, for example, the stratigraphic framework and unit architecture broadly defined by detailed core studies and mapping (Kempton et al., 1987a,b; Curry et al., 1988; Grimley and Curry, 2001, 2002) was harnessed to code description of the materials noted in about 30,000 water well logs and other observations (Dey et al., 2007a). Isopach and structure contour maps were produced for

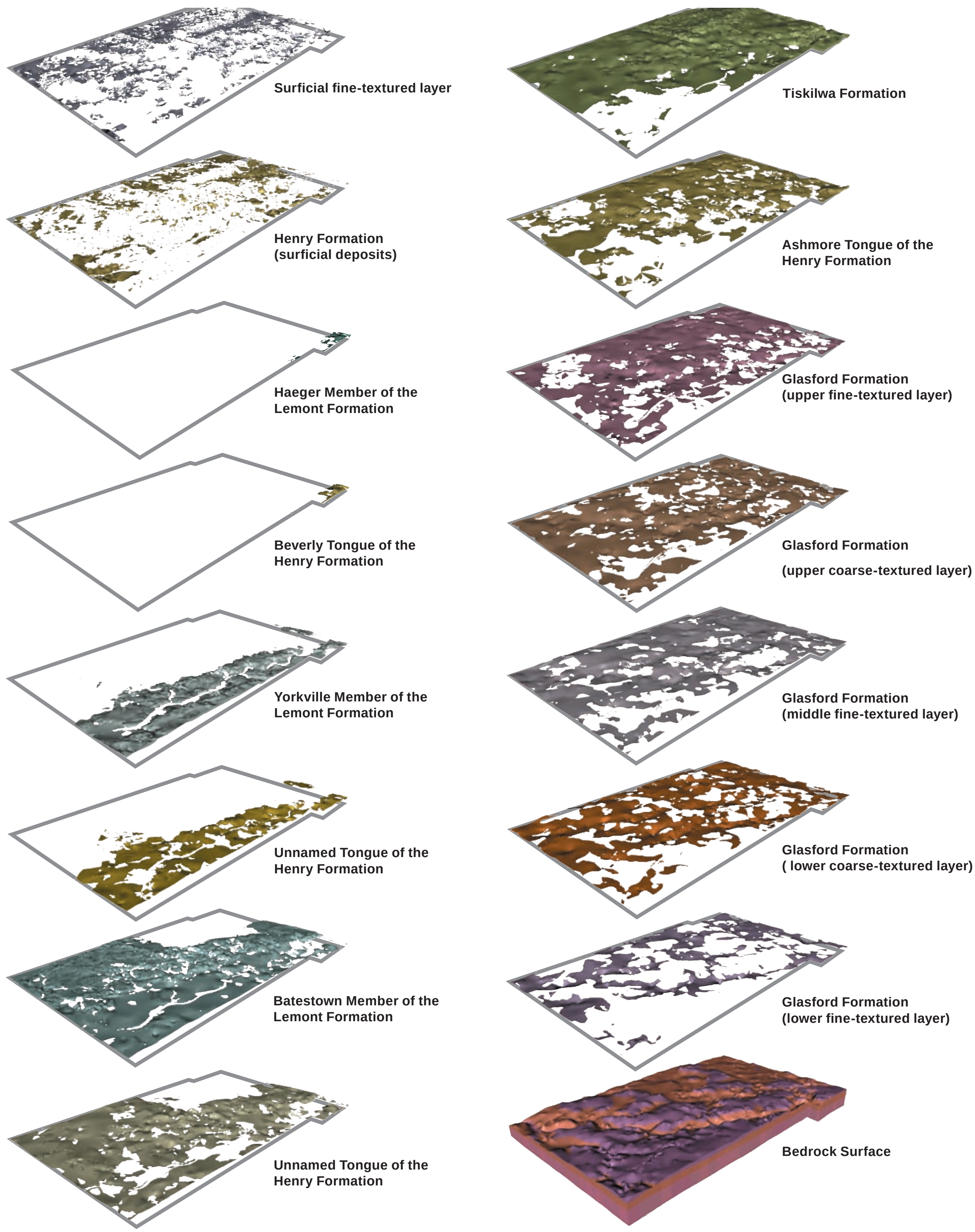


Figure 5. Stacked three dimensional model of glacial drift units in Kane County (Abert et al., 2007)

11 glacial drift units and three bedrock units (Dey et al., 2007b, c, d) as well as an aquifer sensitivity map (Dey et al., 2007e). The 14 layers were digitally stitched together to form a 1:100,000 scale three-dimensional model of the geology of Kane County (Abert et al., 2007; Figure 5). The Illinois State Water Survey recently adapted this model for groundwater flow by converting it to a 3-D array of square grid cells about 200 m across with a minimum thickness, employed in areas of absence, of 0.3 m. The groundwater flow model of the glacial drift and uppermost bedrock consists of 18 layers and approximately 1.5 million grid cells. The groundwater flow model was calibrated using stream flow measurements and by numerous water level measurements from wells open to the various aquifer units, including a "synoptic" measurement of water levels from over 1000 wells during a 60-day period (Locke and Meyer, 2007). Ultimately, the model will permit planning officials in Kane County to evaluate water availability under various scenarios of climate change to help them plan for future growth.

Objectives

Our discussions will focus on the deglacial history of northeastern Illinois based on identification of key landforms and their constituent sediment and fossils. We will examine the glacial landscape of northeastern Illinois and discuss the physical characteristics of the sediment units that we have identified and mapped. The succession and unit characteristics are the basis for our stratigraphic framework; our mapping is influenced further by interpretation of the environment of deposition of the units. We will also highlight the geomorphology the area by referring often to a shaded relief maps based on either 10-m DEM data (McGarry, 2000), LiDAR data (Lake County Department of Information and Technology, 2004), or 30-m DEM data (Luman et al., 2003).

Physical Setting

Northeastern Illinois is located in a region of varied geomorphologic character (Figure 2). Several named moraines that formed during the last glaciation occur in upland areas. We will be traveling across the Burlington, Marengo, Woodstock, and Ransom moraines, and the Bloomington and Marseilles Morainic Systems (Figure 1). The hummocky or ridge-like topography of the moraines contrasts with intervening tracts of lakebeds and the valleys of waterways. Most river valleys were formed by meltwater erosion, in some cases, clearly by the catastrophic release of meltwater.

Previous Investigations

During the late 19th century, the Geological Survey of Illinois reported on many aspects of northeastern Illinois geology (e.g., Bannister, 1870, 1882). At that time water was being obtained from major waterways such as the Fox and Kishwaukee Rivers and from relatively shallow sand and gravel alluvial aquifers. Building stone was mined from dolomite that cropped out along major rivers (Conover, 1884). Ancient lake bed sediments supplied clay for manufacture of brick and decorative terra cotta; peat was mined for fuel and soil conditioning.

Between World Wars I and II, information gained from the drilling of water wells had become a major source of geologic information (Udden, 1914) as well as test borings for determination of dolomite resources (Thwaites, 1923, 1927; Palmer, 1933). Evidence was emerging for the distribution of buried bedrock valleys, filled and covered by more recent glacial sediments, and the character of the deposits within those valleys (Leighton, 1925; Horberg, 1953).

In the late 1950's, geologists at the ISGS began to recognize that glacial deposits are traceable in the subsurface and their relationships could be classified based on stratigraphic principles, like rocks of Paleozoic age. The physical characteristics of glacial deposits such as particle-size distribution, clay mineralogy, and clast lithology began to be routinely analyzed on core subsamples, especially diamict units. Hackett (1960) and Hackett and Hughes (1965) implemented drilling programs designed specifically to acquire cores of glacial deposits (Lund, 1965a,b, 1966; Reed, 1972, 1975). Landon and Kempton (1971) characterized the glacial drift sequence during siting of the Fermi National Accelerator Laboratory at Batavia. Later, efforts to bring the U.S. Department of Energy's Superconducting Super Collider (SSC) to the region included a detailed examination of the stratigraphy, groundwater resources, and engineering characteristics of geologic materials through yet another drilling program in Kane County and surrounding areas (Kempton et al., 1987a,b; Curry et al., 1988; Graese et al., 1988; Vaiden et al., 1988; Harza Engineering Co., 1988). Wickham et al. (1988) also described the glacial geology of the area, and was based on the Master's thesis of the senior author.

Recent hydrogeological investigations have used a combination of geophysical techniques, especially seismic refraction, in addition to test borings and pumping tests, to characterize sediments and groundwater resources in buried bedrock valleys (e.g., Gilkeson et al., 1987; Fitzpatrick et al., 1992; Larson and Orozco, 1991, 1992; Larson et al., 1992a,b) and deeper bedrock aquifers (Visocky et al., 1985). Heigold (1990) analyzed and interpreted seismic refraction and reflection data that had been collected chiefly for the SSC study.

The first surficial geologic map of Kane County emphasized locating sand and gravel aggregate resources (Leighton et al., 1930). A regional geologic mapping program in northeastern Illinois (at a scale of 1:100,000), augmented with data from test-borings, was cosponsored by Northeastern Illinois Planning Commission and ISGS in the 1970s (Gilkeson and Westerman, 1976; Kempton et al., 1977). A regional sand and gravel aggregate map (scale: 1:100,000) was updated by Masters (1978). As an outgrowth of the geologic investigations for the SSC, maps at a scale of 1:62,500 were published of Kane County's drift aquifers (Vaiden and Curry, 1990a), bedrock topography (Vaiden and Curry, 1990b), drift thickness (Erdmann et al., 1990), stack-units (to a depth of 15 m; Curry, 1990a), Tiskilwa Formation isopach (Curry, 1990b), and surface slopes (Schneider et al., 1990).

The surficial geology of McHenry County was mapped at a scale of 1;100,000 (Curry et al., 1997). Published prior to the present mapping program, discussed below, McHenry County was selected for this early mapping effort because of high population growth rates and abundant shallow sand and gravel deposits (ISGS, 1992). Thirteen borings were sampled; particle-size determinations and clay mineral analysis confirmed the stratigraphic framework that included seven aquifers (Curry, 1995; Curry et al., 1997). An aquifer sensitivity map, mapped to a depth of 15 m, and the first of its kind at a scale of 1:100,000, verified that a significant portion of the county was underlain by shallow sand and gravel aquifers. A series of piezometric surfaces for each aquifer were mapped based on "synoptic" measurement of about 700 water wells over a 15 day period (Meyer, 1998). This experience led to the Kane County mapping projects, discussed above.

Recent work has been programmatically driven by STATEMAP (National Cooperative Geologic Mapping Program) and the Central Great Lakes Geologic Mapping Coalition (CGLGMC). The latter program formed in 1999 (Berg et al., 1999) as a partnership of the Illinois, Indiana, Michigan, Ohio, and United States geological surveys. The partnership utilizes the collective technical and scientific expertise of the agencies to map the glacial deposits in three dimensions of the participating states. The collaboration focuses on delivering earth science information to support understanding of societal and geologic hazards (see www.greatlakesgeology.org). Both STATEMAP and CGLGMC have provided resources needed to do geologic mapping and as a consequence, have led us to many of the stops on this trip. For work in Lake County, the ISGS has combined the efforts of these two programs to reach the goal of mapping the geology in three dimensions. Maps of glacial deposits at the surface are produced via STATEMAP, but the same areas are targeted with CGLGMP funds for additional deep exploration of glacial deposits, determination of facies relationships, and understanding the near-surface landform-sediment relationships. Mapping in Lake County has been underway since 2000. The project includes an extensive drilling program; more than 150 boreholes have been drilled including 64 with continuous cores to rock drilled since 2002. The ISGS plans on coring 20 to 30 new boreholes in Lake County over the next two years focusing on the Tinley and Lake Border moraines.

GEOLOGIC FRAMEWORK

Bedrock Geology

Paleozoic bedrock occurs beneath the glacial drift in northeastern Illinois ranging in age from Cambrian siliciclastic rocks to Pennsylvanian sandstone, siltstone, shale and coal (Figure 6). Distribution of rocks at the bedrock surface partially reflects the attitude of the bedrock units as well as post-Paleozoic erosion and incision of bedrock valleys as much as 115 m deep.

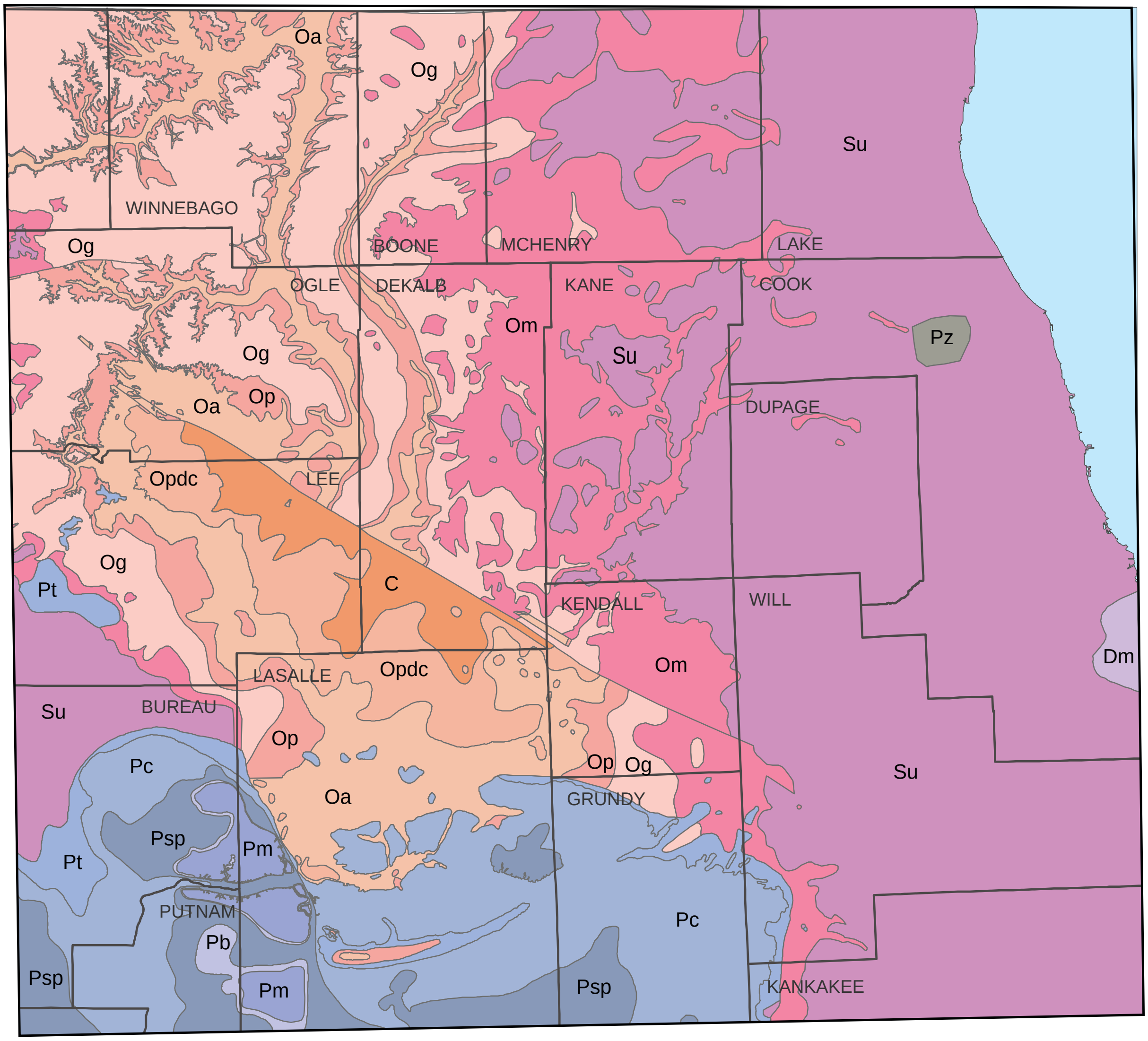
Northeastern Illinois is located at the intersection of the Wisconsin and the Kankakee Arches. The Kankakee Arch separates the Illinois Basin to the south from the Michigan Basin to the northeast (Figure 7). In much of northeastern Illinois, bedrock strata dip to the southeast approximately 0.1° to 0.2° (Graese, 1991). Small amplitude folds and minor faults are superposed upon this regional dip. Other major structural features in northeastern Illinois include the Des Plaines disturbance (Nelson, 1995), the Sandwich Fault Zone (Kolata et al., 1978), and La Salle Anticlinorium (Figure 7; Kolata, 2005). Faults of the Sandwich Fault Zone are nearly vertical with net northward downthrow as much as 240 m. The northern part of the La Salle Anticlinorium occurs near the stops on Day 1 at Wedron and Buffalo Rock. Near these sites, this structure is manifest as an angular unconformity between beds of the Ordovician St. Peter Sandstone that dip as much as 45 degrees in contact with nearly horizontal beds of shale and siltstone of the Pennsylvanian Tradewater Formation.

Bedrock Units Underlying Glacial Drift

St. Peter Sandstone. Both promontories at Buffalo Rock and Starved Rock State Parks are formed of St. Peter Sandstone 30 to 60 m thick. We will observe this unit where it is mined at the Wedron Sand Pits (Stop 2). The unit is composed of weakly cemented, well-rounded, nearly pure quartz sand (99.85% SiO_2). In general, more than 90% of the unit is finer than 0.5 mm and coarser than 0.15 mm with the mode occurring between 0.5 and 0.2 mm (medium sand). The St. Peter Sandstone is used for melting sand (used for containers, panes, fiberglass, and other products), foundry sand, oil and gas extraction, construction sand, fillers in products such as paint, grout, caulk, pool filter sand, and play sand (Keith and Kemmis, 2005).

Galena and Platteville Groups. Carbonate rocks of the Galena and Platteville Groups typically consist of pale yellowish brown, fine- to medium-grained, medium-bedded dolomite, and one region of limestone (Graese, 1991). The beds are generally 15 to 30 cm thick, wavy, and are separated by thin (1 – 2 mm), commonly stylolitic shale laminae. In some intervals the carbonate rocks contain chert nodules. The combined thickness of the Galena and Platteville Groups ranges from about 90 to more than 105 m where overlain by the Maquoketa Group. We will examine karst features formed in limestone of the Galena Group at Stop 3.

Maquoketa Group. Above the dolomite of the Galena Group is the thinly bedded dolomite, minor limestone, and shale of the Maquoketa Group. The full thickness of the Maquoketa in northeastern Illinois ranges from about 70 to 50 m. In most places in northeastern Illinois, facies changes within the Maquoketa (Graese, 1991) preclude recognition of the formations identified by Kolata



Pennsylvanian System

Pm	Mattoon Fm	shale, sandstone, coal,] limestone
Pb	Bond Fm	
Psp	Shelburn-Patoka Fms., undivided	
Pc	Carbondale Fm	
Pt	Tradewater Fm	

Devonian System

Dm — Muskatatuck Gp; shale

Silurian System

Su — Silurian Fms, undifferentiated; dolomite

Ordovician System

Om	Maquoketa Gp; dolomite and shale
Og	Galena Gp; dolomite, some limestone
Op	Plattville Gp; dolomite, argillaceous
Oa	Ancell Gp; St. Peter Sandstone
Opcd	Prairie Du Chien Gp; dolomite, shale, sandstone

Cambrian System

C — Cambrian formations, undifferentiated, including Potosi Fm (dolomite) and Mt. Simon Sandstone

Pz — Paleozoic System, undifferentiated.

Figure 6 Bedrock geology of northeastern Illinois (Kolata et al., 2005).

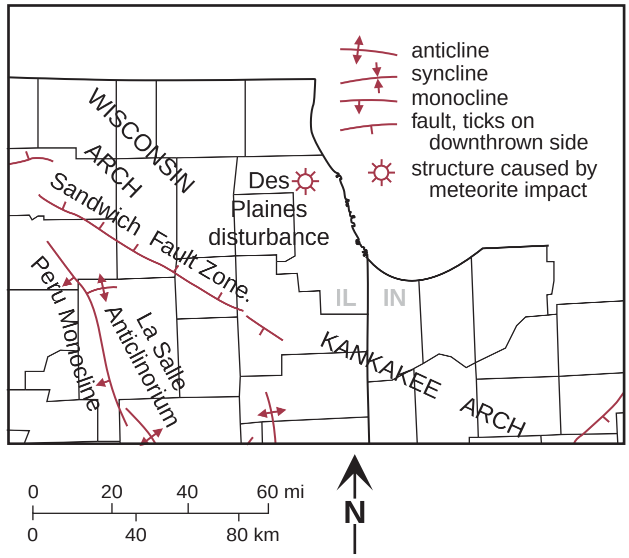


Figure 7 Major structural features associated with the bedrock of northeastern Illinois (from Nelson, 1995).

and Graese (1983). Although we will not see outcrops of this unit, it was readily comminuted by glaciers to silt and clay-sized particles. These fines were entrained in the glacial bed, later to be deposited as till, or redeposited as supraglacial or paraglacial debris, such as proglacial lake sediment.

Silurian-age rocks at or near the bedrock surface in northeastern Illinois primarily are composed of light gray, fine-grained dolomite and limestone with occasional wavy, greenish gray shale laminae. The oldest Silurian unit is typically the Elwood Formation, which is a slightly argillaceous, light gray, fine-grained dolomite; chert nodules and layers as much as 7 cm thick are present in most places. This unit is about 7 to 10 m thick where it has not been eroded. The Elwood Formation grades upward into the non-cherty, relatively pure dolomite of the Kankakee, Joliet, Sugar Run, and Racine formations with a combined thickness of about 115 m (Kolata, 2005).

Devonian-age rocks have recently been discovered in Illinois near the Indiana border, but they will not be seen on the field trip. Soft, easily eroded Devonian black shale subcrop beneath lake sediment and glacial drift in the Lake Michigan basin (Wold et al., 1981), and comprise as much as 20% of the coarse sand fraction of some grey till units (Willman and Frye, 1970). Microscopic amber discs (spores) of the Devonian green algae *Tasmanites* are resilient to glacial comminution and were concentrated in fine-grained diamicton and proglacial lake sediment.

Pennsylvanian-age shale, siltstone, sandstone, and coal occur along the Illinois River valley in the southern part of the study area. Here, the Pennsylvanian Tradewater and Carbondale Formations occur above an angular unconformity developed in Ordovician and Cambrian rocks that form the northern part of the La Salle Anticlinorium (Kolata, 2005; Figure 6).

Bedrock Topography

The regional bedrock surface is characterized by gently sloping highlands cut by steep-walled bedrock valleys (Figure 8). Elevations of the highest bedrock uplands in the study region range from about 210 to 250 m, and the lowest elevations, between about 120 to 135 m. Fluvial and glaciofluvial sand and gravel deposited in the bedrock valleys serve as important aquifers in the western suburbs that are reliant on groundwater, such as in McHenry and Kane county. The Troy, Rock, and St. Charles bedrock valleys are the largest buried bedrock valleys in this region (Figure 8). Maximum relief of the bedrock surface in the area is about 110 m. The overall slope of the bedrock valley floors is gradual. For example, the slope of the elevation of the St. Charles

Bedrock valley floor changes in Kane County from approximately 180 m above sea level to less than 150 m with a mean gradient of approximately 3 m/km.

GLACIAL DRIFT AND SURFICIAL DEPOSITS

The Pleistocene sediments of northeastern Illinois, as much as 120 m thick, were deposited during multiple fluctuations of the Laurentide Ice Sheet. For deposits of the last glaciation, moraine orientation, composition, and internal architecture (Mickelson et al., 1983) serve to characterize landforms associated with sublobes of the Lake Michigan lobe. The landscape is formed of unlithified diamicton, sand and gravel, silt, and silty clay, materials known collectively as glacial drift.

Methods

Lithostratigraphic units were identified in this study on the basis of (a) stratigraphic position, (b) physical and mineralogical characteristics of mappable units (including interpretations of downhole natural gamma-ray logs), and (c) identification of glacigenic successions.

Particle-size analyses. An important physical characteristic used to classify diamicton units in this region is particle-size distribution. Most analyses at the ISGS are determined by a hydrometer. The percentage of gravel (>2 mm) is calculated from the 15 to 40 gram samples. The weight of the gravel fraction is removed when calculating the relative percentages of sand, silt and clay. Particle-grain size categories of the < 2mm fraction include sand (2 mm to 0.063 mm), silt (0.063 to 0.004 mm), and clay (< 0.004 mm). Midwestern Quaternary geologists and soil scientists use agricultural textural terms such as “loam” and “silty clay loam” to classify the texture of the > 2 mm grain-size fraction (Buol et al., 2003).

Other physical characteristics commonly measured in applied and characterization studies, such as moisture content and Atterberg Limits, are strongly related to grain-size distribution (Curry, 1991), and aid in the identification of some stratigraphic units (Landon and Kempton, 1971; Wickham et al., 1988; Graese et al., 1988). For example, at Fermilab, a statistical analysis was done of a data set of more than 1,000 samples from about 50 borings used to characterize the glacial drift of the ca. 4 km^2 site. Not surprisingly, the analyses show that clay content has a strong positive linear correlation with moisture content and Atterberg Limit values (Curry, 1991). Linear regression equations developed from these relationships provide a quantitative basis for using moisture content, a cheap and readily obtained physical parameter, to aid in the classification of diamicton units. For example, loam to sandy loam diamicton typically has moisture contents that range from 9 to 12%, whereas silty clay diamicton has moisture contents ranging from 18 to 24% or more (Figure 9). The range in grain-size of diamicton in this data set (sandy loam to clay) is representative of all major diamicton units in northeastern Illinois

Semi-quantitative phase analyses, commonly referred to as clay mineral analyses, were done using oriented, aggregate, glycolated slides of the < 0.002 mm fraction (Wickham et al.,1988; Curry et al., 1999; Hughes et al., 1994). Clay minerals were separated into three groups and quantitatively calculated using peak height measurements (counts per second) at diagnostic d-spacings. The three groups include:

a) expandable clay minerals: all materials that expand to approximately 17Å (10^{-9} cm) when solvated with ethylene glycol;

b) illite: clay minerals with a 10Å basal spacing that do not expand when treated with ethylene glycol, and

c) kaolinite plus chlorite: all clay minerals with a 7.2Å basal spacing.

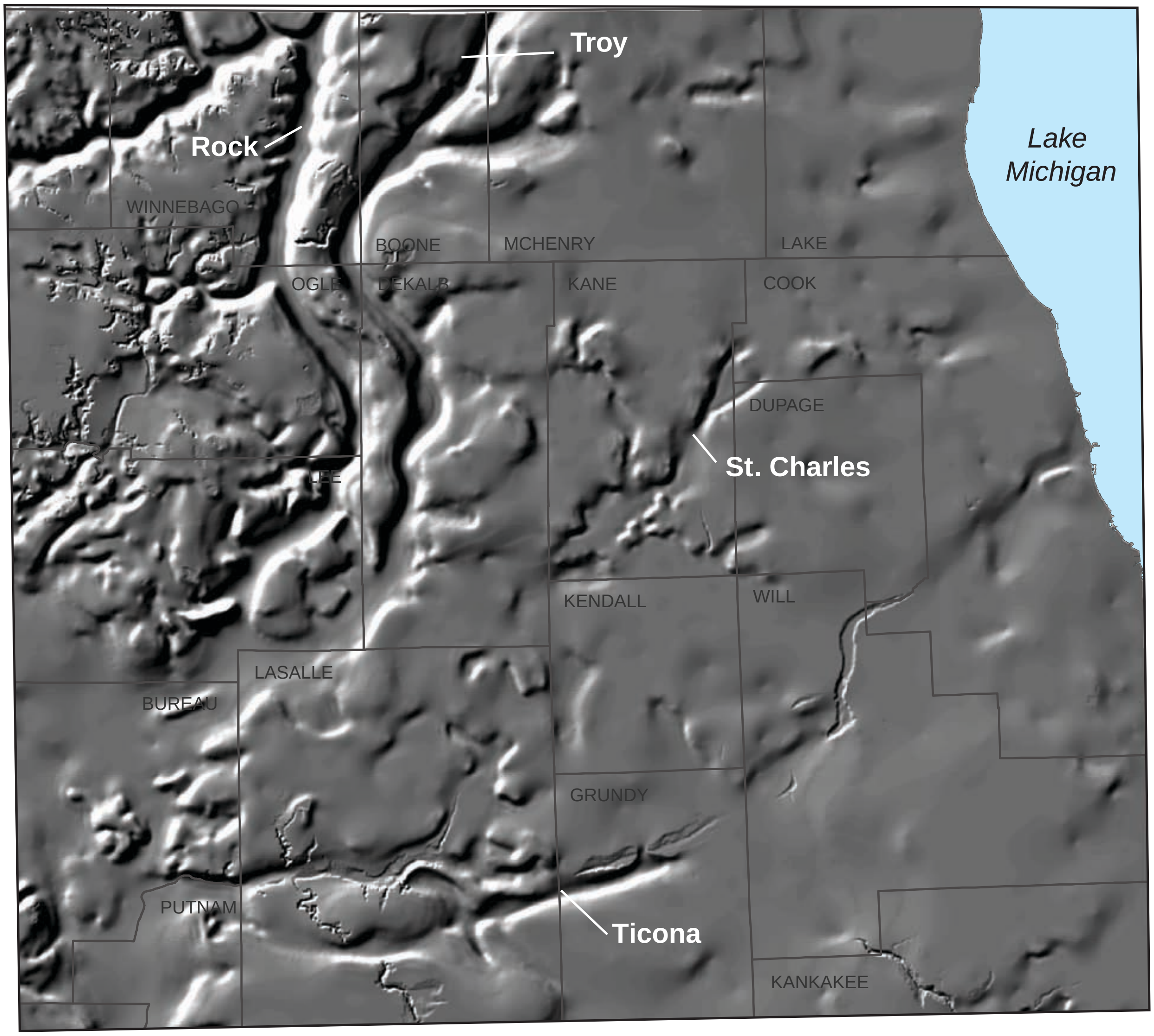


Figure 8 Shaded relief map of the bedrock topography of northeastern Illinois based on the map of Herzog et al. (1994) and data from Luman et al. (2003). Map courtesy of Curt Abert, ISGS. .

On the X-ray diffractograms, the spacings for the groups above are measured at 5.1, 8.8, and 12.4°2θ respectively; correction factors of 1.4, 4.0, and 1.8 are applied, respectively, in order to calculate relative percentages of each group (Hughes et al., 1994).

In more recent studies, the ISGS has been estimating kaolinite and chlorite percentages by applying the peak height ratio measured at 24.9°2θ and 25.1°2θ for kaolinite and chlorite, respectively, to the peak measured at 12.4°2θ (e.g., Curry and Grimley, 2006). We have found the individual measurements of kaolinite and chlorite to be useful in characterizing loess units in northeastern Illinois, but less useful for helping to differentiate among diamicton units of the Wedron Formation. Relative calcite and dolomite content of the < 0.002 mm fraction were measured by peak height (counts per second) on the XRD diffractograms. In some studies, we have estimated the relative content of quartz, plagioclase, and potassium feldspar from X-ray diffractograms (e.g., Curry and Grimley, 2006). Another useful parameter in characterizing stratigraphic units and identifying weathering horizons is the diffraction intensity ratio (DI). The DI is the ratio of the illite peak (8.8°2θ) with the compound kaolinite-plus-chlorite peak at (12.4°2θ). The ratio increases upwards through a weathering profile due to the loss of chlorite. Chlorite weathering is among the first reactions to take place in the soil profile (Droste, 1956), and occurs in many places before leaching of carbonate minerals is complete. Low-charge and high-charge vermiculite are products of this weathering process.

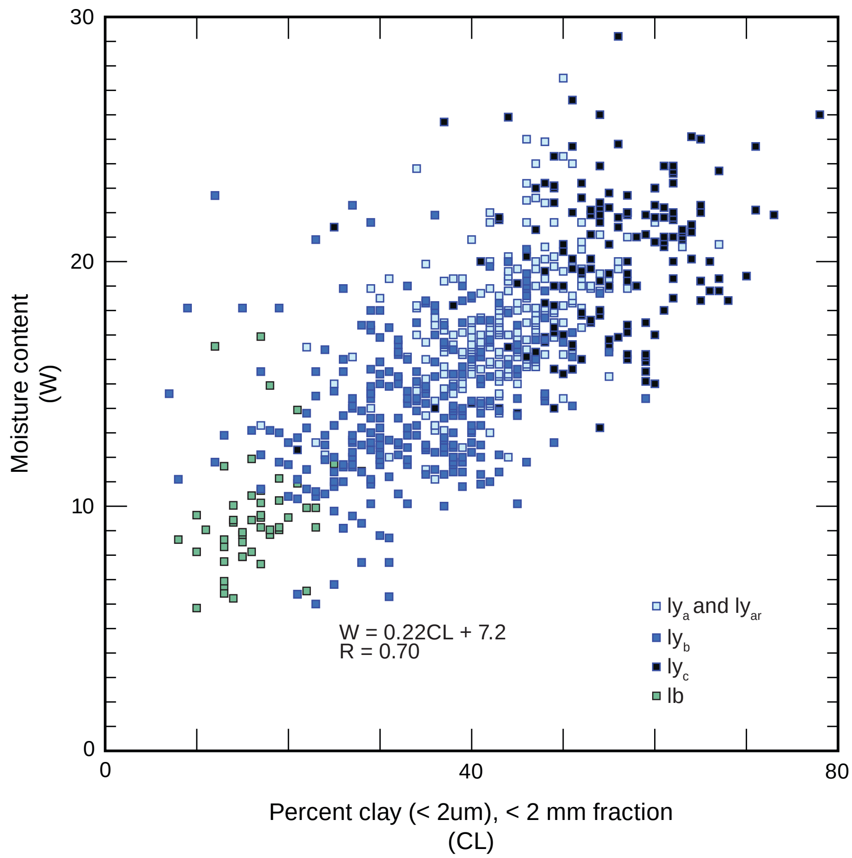


Figure 9. Relationship between clay content (< 0.004 mm), moisture content, plastic limit, and liquid limit (after Curry, 1991) based on data collected by the Army Corps of Engineers and the ISGS. Units lya, lyar, lyb, and lyc are facies of the Yorkville Member, and lb is the Batestown Member. See Table 2 for additional characterization data for these units..

Natural gamma-ray logging. Many lithostratigraphic units in the glacial drift in northeastern Illinois have characteristic down-hole natural gamma-ray "signatures" that facilitate mapping and interpretation of environments of deposition in three dimensions. The ISGS routinely obtains downhole natural gamma-ray logs for its deep boreholes. The logs reveal the natural radioactivity of the material surrounding the boring. In northeastern Illinois, the natural radioactivity is positively correlated to clay content. Clay minerals contain relatively high concentrations of the naturally occurring radioactive isotope of potassium, ^{40}K. Other sources of natural radioactivity include organic matter in paleosols (which under reducing conditions attracts and retains Uranium). The Batestown Member (Lemont Formation) is a relatively sandy diamicton unit that logs, in some areas, relatively high possibly due to the occurrence of ultramafic minerals (Curry et al., 1999). Natural gamma-ray logs may be used to differentiate diamicton from lacustrine successions and other deposits, and indicate the top of bedrock. Moreover, events may be interpreted from logging patterns, such as a gradual upsection shift to lower counts, indicative of decreasing clay content, and perhaps hinting of an advancing ice margin into a proglacial lake (Bleuer, 2004). Such interpretations are validated by examining cores and cuttings.

In summary, a mappable diamicton unit (formation) is characterized by its position in the stratigraphic succession, its relationship to moraines, color, and texture. Characterization may be refined further with mineralogical analyses, moisture content, and analysis of downhole natural gamma-ray logs. Several units have a characteristic frequency of inclusion of bodies of sorted sediment (Curry et al., 1999), discussed below.

Drift Thickness

Drift thickness in northeastern Illinois ranges from nil (where bedrock is exposed at the land surface) to about 120 m under the Marengo Moraine (Piskin and Bergstrom, 1975). Other moraines are formed of thinner drift from ca. 50 to 60 m thick. In general, the glacial drift thins to the south. Drift is absent along reaches of all major drainages such as the Illinois, Kankakee, Fox, and Des Plaines rivers.

Stratigraphic Succession

In the study area, there is evidence for two major advances of the Lake Michigan lobe with an intervening interglacial period when there was warmer climate and significant soil accretion, bioturbation, and weathering. The older Illinois Episode ice sheet and its meltwater deposited the Glasford and Pearl Formations, and the Teneriffe Silt from about 200,000 to 130,000 yr BP (Johnson, 1976; Curry and Follmer, 1992; Curry and Pavich, 1996; McKay, 2008). In northeastern Illinois, the Sangamon and Farmdale Geosols are superposed weathering profiles developed in late Illinois Episode glacial drift and Alton Subepisode (Wisconsin Episode) loess, respectively (Curry, 1989; Curry and Pavich, 1996). Michigan Subepisode (Wisconsin Episode) glacial sediment, dating from about 29,000 to 14,670 cal yr BP (24,000 to 12,500 C-14 yr BP), overlie the soil complexes, including sorted sediment of the Mason Group, and diamicton of the Wedron Group (Figure 10; Hansel and Johnson, 1992, 1996).

The thickness and general character of lithostratigraphic units are portrayed in cross sections across Kane County (Figures 11a,b) and McHenry County (Figure 11c). For deposits of the last glaciation, at the highest level of classification is the Mason and Wedron groups which differentiate sorted from very poorly sorted glacigenic sediment. The Mason Group is subdivided into deposits of laminated silt and clay (Equality Formation), uniform and/or weathered silt (Peoria and Roxana silts), and stratified or cross-bedded sand or sand and gravel (Henry Formation). Reflecting the importance of regionally important glaciogenic successions, several units are stratigraphic tongues which have the same rank as members. The inclusion of tongues in the new stratigraphic scheme allows for the repetition of units at the formation level at a single locality.

Banner Formation. The oldest named Quaternary lithostratigraphic unit in the region is correlated with the pre-Illinois Episode Lierle Clay Member of the Banner Formation (Willman and Frye, 1970). Of the more than 100 borings to bedrock, this unit was identified twice, once in northernmost McHenry County (Curry and Pavich, 1996), and once in northern Kendall County (Curry et al., 1988). At these sites, the Lierle Clay Member is one m-thick, and composed of gleyed, leached clayey diamicton and sesquioxide (organic iron - manganese) concretions. These features formed in an ancient pedogenic environment. The Lierle Clay Member also contains abundant clay minerals such as mixed-layer kaolinite/smectite, which is further indication of weathering. As it is elsewhere in the region, the Lierle Clay is considered part of the Yarmouth Geosol (Grimley et al., 2003).

Glasford Formation. Deposited during the Illinois Episode, the Glasford Formation is generally the oldest Quaternary unit observed in the study region. It generally occurs above bedrock and below sediment of the Mason or Wedron Groups. The Glasford Formation consists of generally gray to reddish brown diamicton with lenses of sand and gravel and local inclusions of fine sand, silt, and clay. The Glasford pinches out to the south and east. Several till members of the Glasford Formation have been traced from Boone and Winnebago Counties in the Rockford area to McHenry and Kane counties (Graese et al., 1988; Kempton et al., 1985).

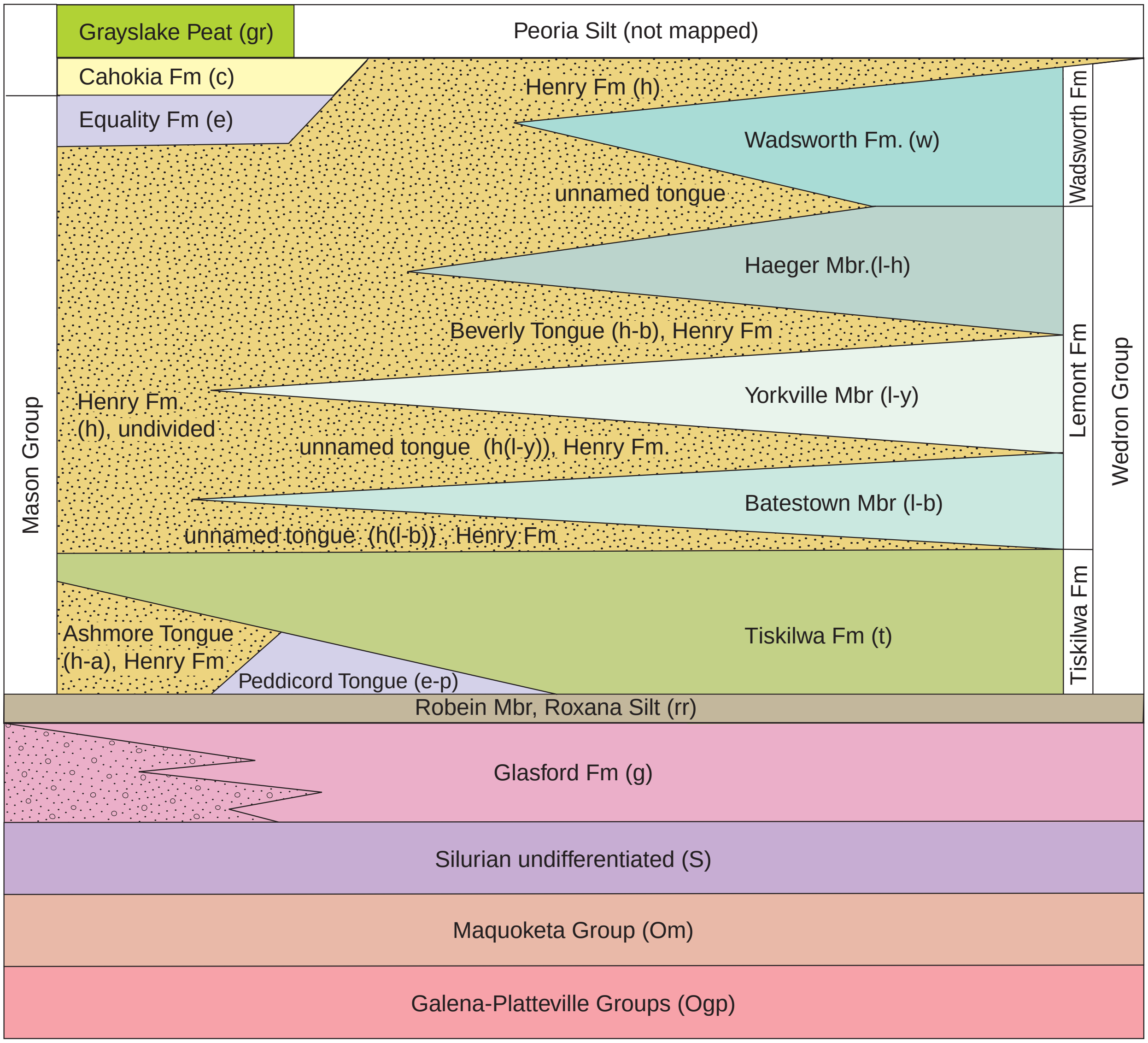


Figure 10 Quaternary stratigraphic framework of northeastern Illinois (after Curry et al., 1997).

Pearl Formation is composed of chiefly sand and gravel; *Teneriffe Silt* is primarily silt and clay. Like Glasford Formation deposits, the upper parts of these units possess features attributed to the Sangamon Geosol.

The Sangamon Geosol, a distinctive buried weathered horizon (Follmer, 1983) occurs in many places in upper Illinois Episode sediment where it may be overlain by accretionary Berry Clay or organic-rich Roxana Silt (Curry, 1989; Curry and Pavich, 1996; Curry et al., 1999; Jacobs et al., this volume (Stop 6)). Accretion gley deposits classified with the Sangamon Geosol are known as Berry Clay. Originally included as a member of the Glasford Formation by Willman and Frye (1970), the Berry Clay has been treated as a formation in areas where it is soft, fine-grained, smectite-rich, and distinct from the underlying hard, illite-rich loam diamicton (i.e., Curry et al., 1994).

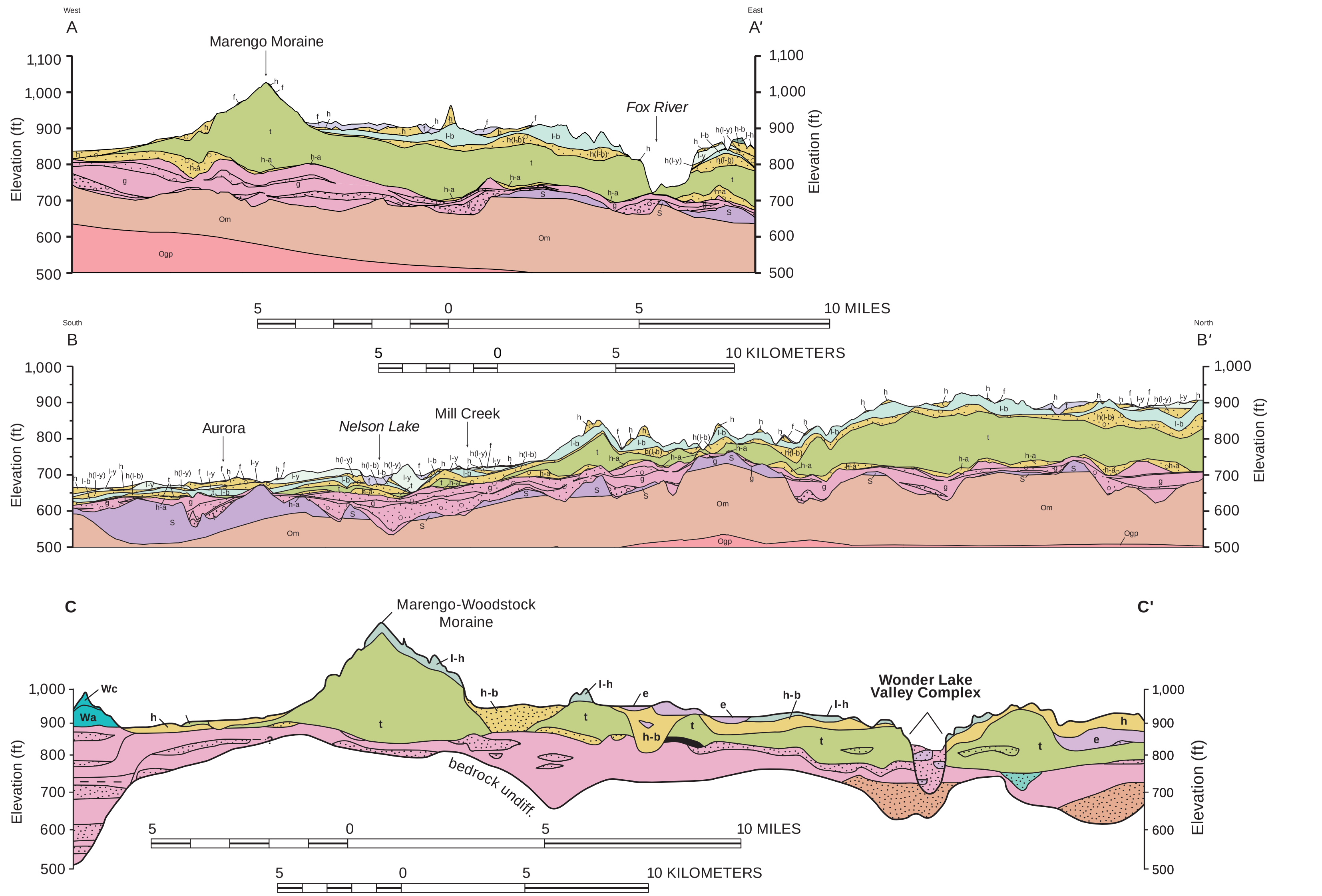


Figure 11 (A-A') East-west cross section of northern Kane County based on three-dimensional modeling (Dey et al., 2007a), (B-B') North-south cross section of central Kane County based on three-dimensional modeling (Dey et al., 2007a), (C-C' East-west cross section of northern McHenry County (after Curry et al., 1997).

DEPOSITS OF THE LAST GLACIATION

Mason Group

Occurring above the Sangamon Geosol, the Mason Group (Hansel and Johnson, 1996) is comprised of sorted sediment, including:

- ***Robein Member,*** *Roxana Silt.* Composed of leached, black, organic-rich silty clay and silt loam as much as about 1 m thick, the Robein Member ranges in age from about 55,000 yr BP to 29,000 cal yr BP. The lower bounding age is based, in part, on studies of the Roxana Silt in southwestern Illinois where it is thicker (McKay, 1979; Leigh and Knox, 1993). Most radiocarbon ages of this unit in northeastern Illinois are finite, and younger than 40,000 C-14 yr BP.

- ***Henry Formation.*** Composed of mostly sand, gravel, cobbles, and occasional coarser clasts, Henry Formation was deposited as outwash, littoral deposits, or in dunes. The largest mapped areas of Henry deposits, typically less than 15 m thick, form outwash plains and valley trains (Figure 12). Subunits of the Henry include (1) the Ashmore Tongue, which extends beneath the Tiskilwa Formation, (2) the Wasco facies, which forms most of the kames and eskers in the ice-stagnation topography of the Elburn Complex (Grimley and Curry, 2001) and (3) the Beverly Tongue which extends beneath the Haeger Member. Unnamed tongues of basal sand and gravel of the Batestown and Yorkville member of the Lemont Formation were mapped in Kane County by Dey et al. (2007a). Usually associated with deposits of the Wedron Group, at Stop 2 (the Wedron pit) we will observe an uncommon example of organic-rich sand and gravel facies of the Henry Formation that was synchronous with deposition of Roxana Silt.

The Beverly Tongue is regionally the thickest and most continuous subunit of the Henry Formation. Because thick near-surface deposits occur on the Valparaiso, West Chicago, and Woodstock moraines in areas of urban growth, the Beverly Tongue is mined for aggregate. Exposures of this unit are located at the Beverly Pit (Hansel et al., 1985), Meyers Pit (Curry et al., 1997), the Spring Lake pit (Stop 8), and the Thelen Pit (Stop 9). In many places the Beverly Tongue comprises three facies including 1) laminated fine sand and silt, 2) cross-bedded fine sand to coarse sand with gravel and small cobbles, and 3) very poorly-sorted coarse sand, gravel, cobbles, with either a silty sandy binder, or open framework. These facies roughly correspond to the distal, medial, and proximal facies, respectively, of Cobb and Fraser (1981) for deposits mapped near Crystal Lake, Illinois. In vertical section, the boundaries between facies typically are abrupt. In some parts of Lake and DuPage counties, the silt and fine sand facies may extend eastward beneath diamicton of the Wadsworth Formation to areas where sandy loam Haeger Member diamicton pinches out. In such areas, this material may be classified as an unnamed tongue of the Equality Formation (e.g., Stumpf, 2006). The Beverly Tongue is the primary component of the Valparaiso aquifer of Curry and Seaber (1990) and the related Carpentersville aquifer of Dey et al. (2007a).

- ***Equality Formation*** is composed of primarily finely stratified, laminated, or uniform fine-grained, moderately- to well-sorted clay, silt, and fine sand. The Equality has one named tongue, the Peddicord Tongue, which occurs beneath diamicton of the Tiskilwa Formation (Figure 10). Deposits of the Peddicord Tongue will be observed at Stop 6 (the LaFarge pit) and Stop 2 (the Wedron pit). Muds of the Equality and sandy deposits of the Henry unit intertongue in many places; how they are mapped and interpreted in cross section is “author’s choice”. The laminations, fine-grained texture, fossils, and landform association indicate that most deposits of the Equality Formation are of lacustrine origin. Throughout northeastern Illinois, surficial deposits of the Equality Formation are generally less than 8 m thick, but beneath some large lakes are as much as 15 m thick (Curry et al., 1999).

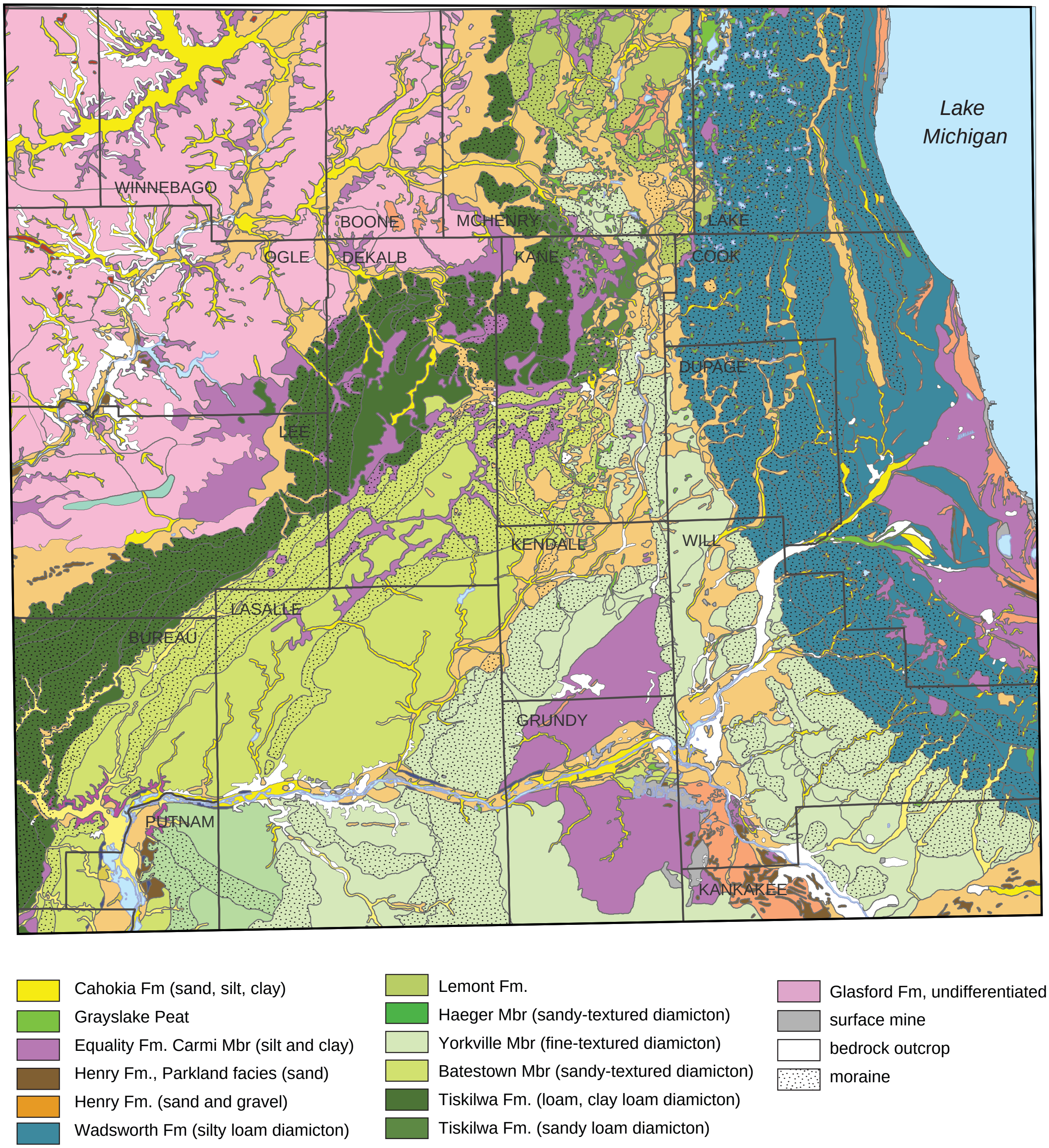


Figure 12 Simplified surficial geology map of northeastern Illinois (after Hansel and Johnson, 1996)

Deposits of lacustrine deposits of the last glaciation are extensive in northeastern Illinois as both surficial and subsurface deposits. A partial listing of lake types include: 1) large proglacial lakes, 2) slackwater lakes, 3) ice-walled lakes, 4) kettle lakes, and 5) postglacial lakes formed by beaver dams, log jams, etc., along low-gradient streams. The largest proglacial lakes in the study area were Glacial Lake Wauponsee (which formed between the retreating Peoria sublobe and the Marseilles Morainic System), Glacial Lake Milwaukee (ponded between the retreating Harvard sublobe and the Woodstock, Fox Lake, and West Chicago moraines), and Glacial Lake Chicago (which ponded between the Joliet sublobe and the Valparaiso Morainic System and Tinley Moraine). Each proglacial lake breached its morainal dams which led to downstream flooding, valley widening, and sculpting of large-scale features, such as Buffalo Rock (Stop 1). The largest slackwater lake deposits occur south of margin of the last glaciation in valleys tributary to large meltwater sluiceways, most importantly, the Wabash and Mississippi Rivers (i.e., Trent and Esling, 1995; Curry and Grimley, 2006). Less extensive slackwater lake deposits have been identified within the boundary of the last glaciation, and attain thicknesses of more than 15 m (Curry, 2007a). Ice-walled lake plains from the last deglaciation have been identified within the glacial margin as far south as about 40.1ºN latitude (Champaign, Illinois). Also known as DeKalb mounds, and originally interpreted as sediments filling melted pingos (Flemal et al., 1973), new evidence (fossil ostracodes, sedimentology, cross sections) points to a supraglacial ice-walled lake origin (Curry, 2008; Stop 7). In some areas, they form more than half of the mapped surficial sediment; the largest ice-walled lake complex is known as Glacial Lake Pingree located in north-central Kane County (Leighton et al., 1930; Willman and Frye, 1970; Grimley, 2005). Lake deposits associated with kettle basins are well known, and new mapping shows that many kettle lake basins are fringed by low terrace deposits formed of sediment containing ostracodes and plant macrofossils identical to those found in the deposits forming ice-walled lake plains (Curry, 2006). At Stop 5 at Mastodon Lake in Aurora, Illinois, we will focus our discussion on the sediments and fossils filling a large kettle typical of this region. Post-glacial lakes formed along streams have left an important but as of yet unstudied record. Limited exposures indicate that log jams and beaver dams were important features that led to their formation. Deposits of this type are usually mapped with the postglacial alluvium known as the Cahokia Formation, or if organic-rich, Grayslake Peat.

Sediment successions in kettles and other low-lying areas in northeastern Illinois share a distinctive change in lithology dating at 14,670 cal yr BP (12,500 C-14 yr BP). This age corresponds with the boundary identified in Greenland ice cores between the Older Dryas and Bölling Chronozones (Stuiver et al., 1995). Pre-Oldest Dryas sediment is silty gyttja rich in mineral matter (although in places it may contain fragments of coniferous wood, needles, etc.). The post-Oldest Dryas sediment has the opposite character and is comprised of 85% or more (by weight) biogenic peat and marl, with the remainder being silt and clay. The biogenic-rich materials have low bulk density and very high moisture contents that typically range from 100% to about 350%. In contrast, the pre-Older Dryas silty gyttja has moisture contents that range from about 35 to 50%.

We propose that the contact discussed above from the Brewster Creek site (Curry et al., 2007) serve as the material referent (type section) for the boundary between the Wisconsin and Hudson Episodes in Illinois. Radiocarbon and sediment characterization data for a paratype sections at Mastodon Lake (Stop 5) and the Oswego overflow channel (Stop 4) are forthcoming. Hansel and Johnson (1996) originally defined the Hudson Episode, but they did not designate a referent section in Illinois.

Fossils are an important component of all of the Equality deposits described above. Deposits of proglacial and slackwater lakes, and ice-walled lake plains all contain fossil shells of minute pelecypods (pillclams, *Pisidium* sp.), valves of ostracodes, head capsules of chironomids (midge larvae), elytra and eggcases of insects, cocoons of flatworms (*Tubellaria*), leaves and rootlets of aquatic plants, and the stems, branches, and leaves of tundra plants, especially bilberry (*Vaccinium ugilinosum* sp) and Arctic avens (*Dryas integrifolia;* Curry and Yansa, 2004). In addition to their paleoecological significance, the terrestrial plant material may provide meaningful radiocarbon ages. Other dating techniques may yield useful information, but the errors are, at this time in their technological development, larger than with AMS radiocarbon ages. For example, an optically stimulated luminescent age of about 15,000 yr BP will have, at best, about 5% error (≈ 750 yrs) whereas a radiocarbon assay of same-aged material will yield an error of about 0.3% (≈ 45 years). In most cases, calibration of radiocarbon ages increase the lab error, but for ages of the Michigan Subepisode and younger, the error increase will be less than about 0.6% (Reimer et al., 2004). An appreciable component of the organic matter in proglacial lake sediments is *Tasmanites* (fossil green algae). In addition to fragments of Pennsylvanian coal and other Paleozoic organics, occurrence of these cysts renders bulk samples of lake sediment unsuitable for radiocarbon dating.

● *Peoria Silt* is a ca. 1 m-thick mantle of silt loam and silty clay loam on most upland surfaces. It is generally absent on landforms that postdate 14,670 cal yr BP (12,500 C-14 yr BP) including most modern floodplains. Initially deposited as eolian silt (loess), Peoria Silt in northeastern Illinois has been significantly altered by modern soil development.

Wedron Group

As much as 95 m feet thick, the Wedron Group is composed of glacigenic diamicton and interbedded layers of sorted sediment that overlie the Glasford Formation, the Berry Clay and Roxana Silt, Ashmore Tongue of the Henry Formation, or Peddicord Formation of the Equality Formation (Figure 10). Previously classified as the Wedron Formation (Willman and Frye, 1970), five principal till units are mapped in the region: the Tiskilwa Formation, the Batestown, Yorkville, and Haeger Members of the Lemont Formation, and the Wadsworth Formation (Hansel and Johnson, 1996).

As much as 80 m thick, the Tiskilwa Formation is the thickest Wedron unit in northeastern Illinois (Wickham et al., 1988; Curry et al., 1997) forming the prominent Marengo Moraine, Burlington moraine (Curry, 2008), and Bloomington Morainic System (Figure 1); in places, it also forms the core of the St. Charles Moraine (Grimley and Curry, 2002; Curry, 2002). Diamicton of the Tiskilwa Formation is readily identified by its uniformity, thickness, and monotonic gamma-ray signature (Table 1; Figures 13 and 14). About 30 m of Tiskilwa Formation diamicton was identified in boring F-14 (Kempton et al., 1987a) located on the northwestern corner of the campus of Northern Illinois University. Although we will not see outcrops of the Tiskilwa Formation, it is the surficial glacigenic diamicton adjacent to and underlying the ice-walled lake plains at Stop 7.

The Lemont Formation embraces the Batestown, Yorkville and Haeger members. The Batestown Member was previously classified as the Malden Member (Hansel and Johnson, 1996), whereas the Yorkville and Haeger members have been retained from Willman and Frye (1970).

Batestown Member forms parts of the Elburn Complex where its diamicton laps onto and interfingers with kamic sand and gravel deposits of the Wasco facies of the Henry Formation (Grimley, 2005; Curry, 2006). The Batestown Member also forms the Arlington and Farm Ridge Moraines (Figure 1). The Batestown Member is noted for its abundant interbeds of sand and sand

Table 1 Summary of particle size of drift units; data from Graese et al. (1988).

Unit	Gravel (% of whole sample)	< 2-mm fraction Sand (%)	Silt (%)	Clay (%)
Cahokia Formation				
x	6	29	45	26
N	44	48	48	48
R	0-51	0-59	16-73	6-49
Grayslake Peat				
x	2	8	52	40
n	6	10	10	10
R	0-3	0-23	26-72	22-61
Richland Silt				
x	1	7	50	43
n	8	8	8	8
R	0-3	0-15	40-61	35-53
Equality Formation				
x	1	8	60	32
n	172	198	198	198
R	0-10	0-30	9-94	2-84
Henry Formation				
x	29	53	32	15
n	112	113	113	113
R	0-76	5-91	2-92	0-53
Wadsworth Formation				
x	6	14	43	43
n	54	54	54	54
R	0-15	0-39	16-81	21-70
Haeger Member				
x	21	38	49	13
n	27	27	27	27
R	5-41	16-53	39-65	5-24
Yorkville Member				
x	4	10	46	44
n	379	987	987	987
R	0-29	0-54	18-83	13-68
Batestown Member				
x	13	36	43	21
n	54	54	54	54
R	0-32	4-57	23-63	6-38
Tiskilwa Formation				

(ablation facies)				
x	17	43	39	18
n	41	43	43	43
R	5-70	16-62	18-54	8-37
Tiskilwa Formation				
x	7	35	38	27
n	315	315	315	315
R	0-25	4-52	28-71	6-45
Robein Member/ Sangamon Soil				
x	8	36	32	32
n	3	3	3	3
R	<1-16	30-44	26-38	17-40
Glasford Formation undivided				
x	11	38	36	26
n	71	77	77	77
R	<1-57	10-58	23-56	11-50

x = mean
n = number of samples
R = range

and gravel (Figures 13 and 14; Curry et al., 1999). In Kane County, the Batestown has a lower loam diamicton facies, and upper siltier diamicton facies (Curry et al., 1999).

As much as 30 m thick, the Yorkville Member forms the Marseilles Morainic Complex, and the Barlina, Minooka, Rockdale, and St. Charles Moraines. Noted for its gray color and overall fine matrix texture, three diamicton facies of the Yorkville Member have been identified at and adjacent to the Fermilab Accelerator Laboratory (Curry, 1991). Landon and Kempton (1971) first described the range of physical characteristics for the facies, but recent investigations have revised their stratigraphic interpretations, summarized below in Table 2. Its clay and silt-dominated texture results in relatively high natural gamma-ray counts (Figures 14 and 15).

The Haeger Member is composed of sandy loam diamicton with discontinuous lenses of sand and gravel that overlies the Beverly Tongue of the Henry Formation. These units form the Woodstock Moraine, the Fox Lake Moraine, and lower portions of the West Chicago Moraine south of Elgin. The unit is as much as 7 m thick where it crops out in large sand and gravel pits such as at Stop 8 (Spring Lake pit) and at the Beverly Pit near Elgin, Illinois (Hansel et al., 1985; Stumpf, 2007).The gamma-ray signature of Haeger diamicton is more variable than the sand and gravel deposits of the Beverly Tongue (Henry Formation) that typically underlie it (Figure 15).

The Wadsworth Formation forms the Valparaiso Morainic System, the upper part of the West Chicago Moraine, the Tinley Moraine, and the Lake Border Moraines (Hansel, 1981). The central concepts of the Wadsworth Formation are its high silt content, moderately high clay content, gray color, interbeds of silt and silty clay, lithologic variability, and association with the aforementioned moraines. Extensive characterization studies done for unpublished landfill reports at Mal-

Table 2 Selected physical properties of diamicton units identified by Landon and Kempton (1971) at Fermilab National Accelerator Laboratory. Units B, C, and D are facies of the Yorkville Member, and Unit E is the Batestown Member, both of the Lemont Formation

Unit	Number of. Samples	Gravel[-1]	< 2 mm fraction Sand	Silt	Clay	Moisture Content
B	341	5 ± 6	11 ± 6	47 ± 8	43 ± 8	17 ± 3
C	399	10 ± 9	20 ± 12	47 ± 9	33 ± 10	14 ± 3
D	210	4 ± 6	9 ± 6	36 ± 7	55 ± 9	20 ± 3
E	101	16 ± 11	44 ± 9	39 ± 14	16 ± 4	10 ± 2

-1 whole sample

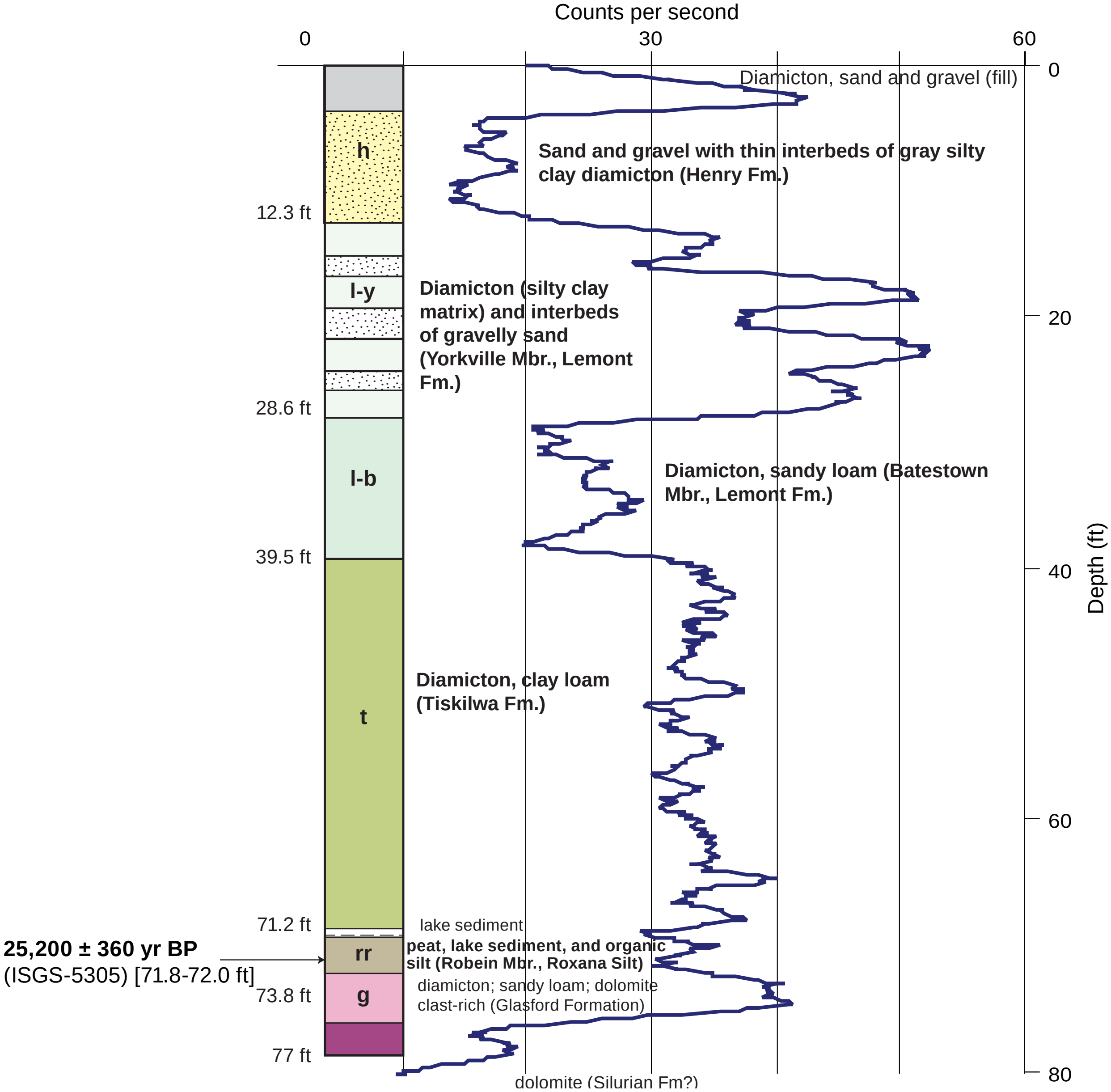


Figure 13 Schematic lithologic log and natural gamma-ray log of boring E-6, Elgin 7.5-minute quadrangle (Curry, 2007a).

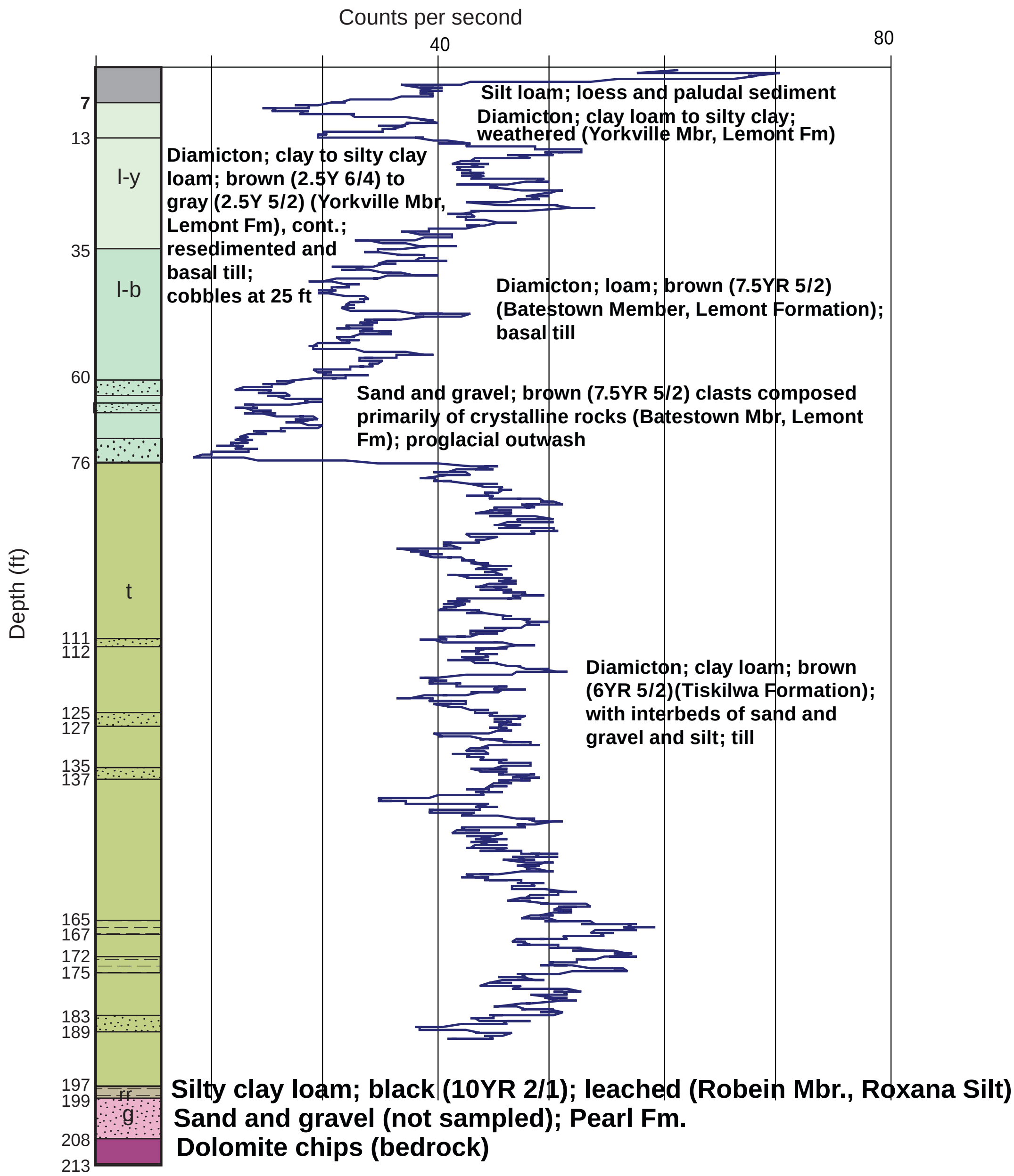


Figure 14 Schematic lithologic log and natural gamma-ray log of boring E-1, Elgin 7.5-minute quadrangle (Curry, 2007a).

lard Lake (Curry, 2007b; Jean Bogner, personal communication), and Blackwell Forest Preserve (Vagt, 1987). both located on the Valparaiso Morainic System, suggests that the Wadsworth is comprised of beds of fine-grained diamicton about 4 m thick separated in many places by thin beds of silt or loam about 1 m thick or less. Many natural gamma-ray logs of the Wadsworth Formation confirm its lithological variability (Figure 16).

Postglacial Deposits

Cahokia Formation is composed of postglacial deposits in the floodplains and channels along modern rivers. Little work has been done to characterize the lithology and fossils characteristic of the Cahokia, and what physical attributes might be used to distinguish it from late glacial de-

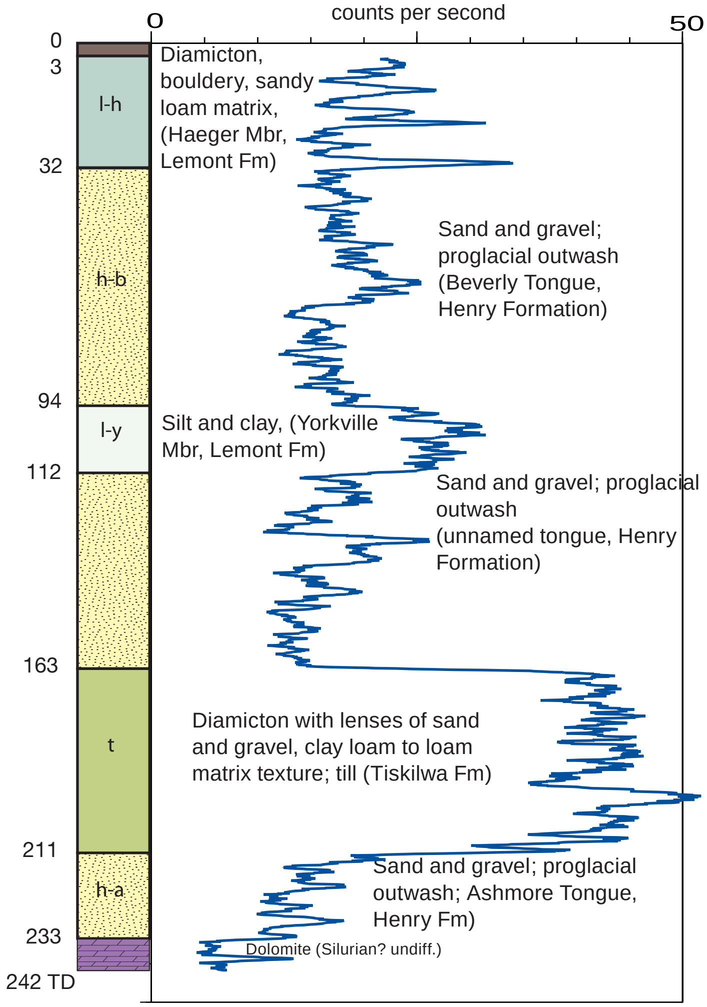


Figure 15 Schematic lithologic log and natural gamma-ray log of boring 35003, Chrystal Lake 7.5-minute quadrangle (Curry, 2005a).

posits of sand and gravel in valleys (Henry Formation) or slackwater lake and other lake deposits (Equality Formation). Sand and gravel deposits occur along active channels which grade to fossiliferous organic silt, silty clay, with pockets of woody debris, and containing discontinuous lenses of sand and gravel. The Cahokia Formation is typically less than 3 m thick in the study region.

Grayslake Peat is composed of peat and marl; the marl typically underlies the peat. Peat deposits in northeastern Illinois are typically composed of degraded emergent wetland plants such at cattails and sedges, and wood (such as larch). Moss is common at sites that were once fens where bicarbonate-charged groundwater emerged from seeps and springs.

Postglacial deposits of organic and marly gyttja (Equality Formation) occur in modern lakes and in the many reservoirs in the area. In natural lakes, these sediments grade laterally to deposits of marl and peat of the Grayslake Peat. Profundal post-settlement deposits in Crystal Lake, McHenry County, Illinois (maximum depth = 12 m), are black and odiferous, contain relatively abundant ragweed pollen (Eric Grimm, personal communication), and a species-poor ostracode fauna *(Cypria ophthalmica and Physocypria globula)* that tolerate summer anoxia by being rapid

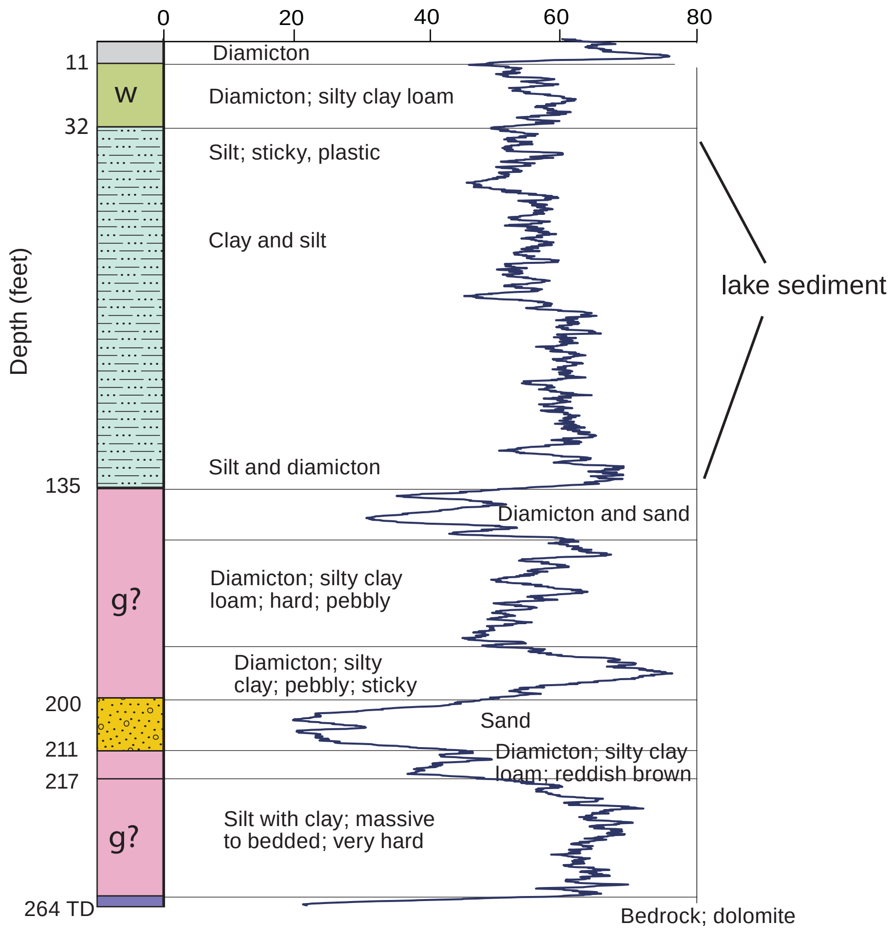


Figure 16 Schematic lithologic log and natural gamma-ray log of boring WAD 02-05, Wadsworth, 7.5-minute quadrangle.

swimmers. Underlying this "Anthropocene" sediment (Zalasiewicz et al., 2008) is gray, marly gyttja and marl with abundant ostracodes that live today in deep waters rich in dissolved oxygen year-round. These species are absent in the lake today. The post-settlement changeover to less species-rich assemblages is likely due to anthropogenic additions of limiting nutrients such as phosphorus and nitrogen that enhance primary production (diatom activity) in the lake and concomitant bacterial metabolism that result in anoxia, especially during the summer prior to fall turnover.

GLACIAL HISTORY

The glacial history of northeastern Illinois is based largely on the relative position of lithostratigraphic units and cross-cutting relationships among landforms and buried land surfaces (i.e., moraines, lake plains, flood channels, and paleosols) and their ages. The ISGS has adopted diachronic time classification for deposits associated with the last glaciation in northeastern Illinois (Figure 17; Hansel and Johnson, 1992, 1996; Johnson et al., 1997). This system replaced, or was offered as an alternative to, the chronostratigraphic scheme of Willman and Frye (1970). Both diachronic and chronostratigraphic systems use type sections and material referents to define units and establish events. An event in this context refers to either a point in time or span

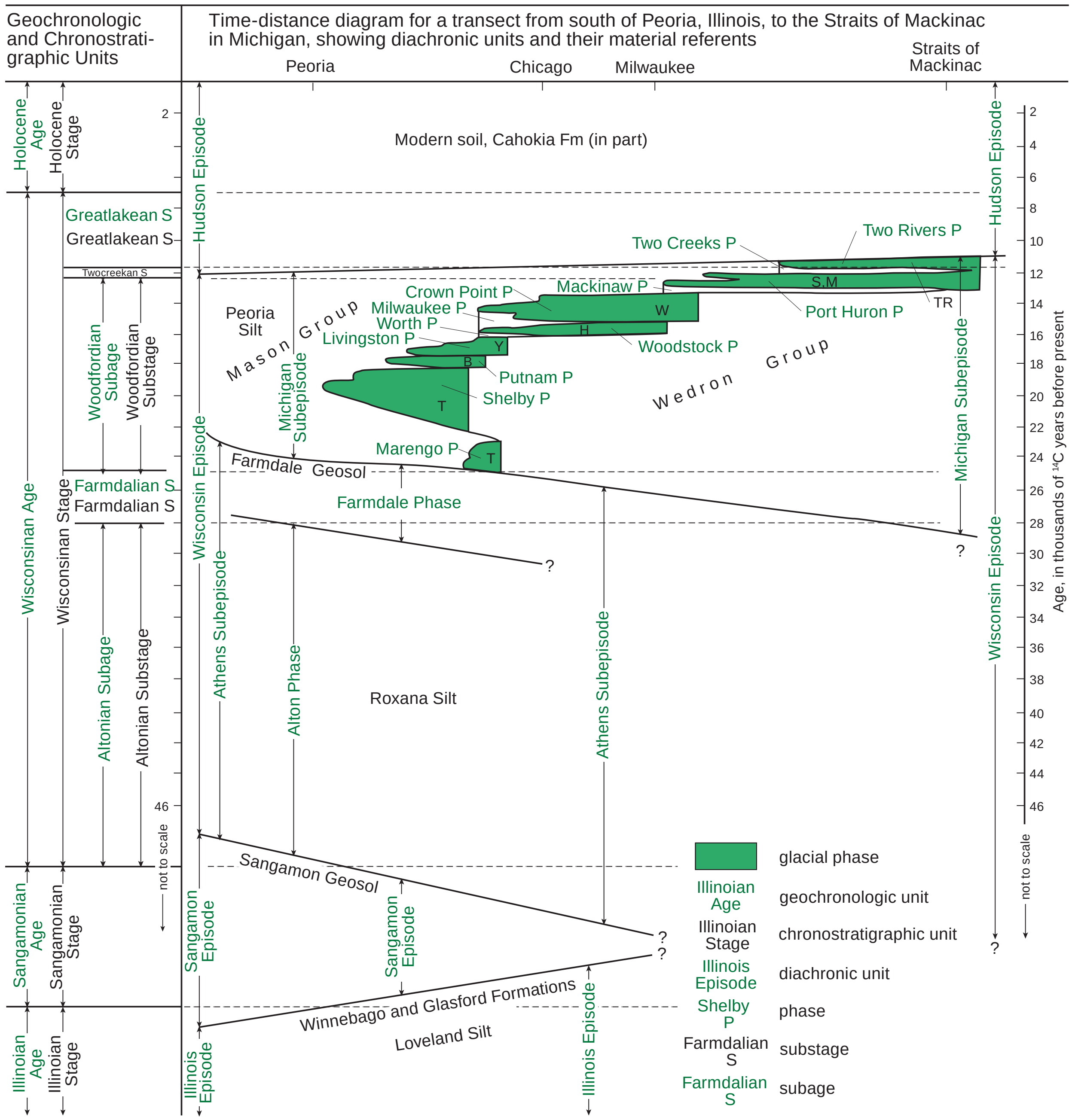


Figure 17 Geochronological, stratigraphic, and diachronic units in the Lake Michigan Lobe (Hansel and Johnson, 1996).

of time depending of the environment of deposition or other issues. An underlying principle in diachronic classification is that the time-lines between lithostratigraphic units from place to place are time-transgressive. In chronostratigraphic classification, the time boundaries are defined at a type section, and are invariant.

Pre-Illinois Episode (ca. 2,000,000-190,000 yr BP)

McKay et al. (2005) suggest that “because of correlation uncertainties and a paucity of age determination of oldest Quaternary deposits, glacial and interglacial episodes that predate the

Illinois Episode are … (known) collectively as the Pre-Illinois Episode.” As noted above, pre-Illinois deposits are rarely observed in cores from the study area, but they have been noted from a number of localities to the south and west. The “County Line Silt” is a unit from near Quincy, Illinois, that is magnetically reversed and ostensibly older than the Bruhnes Chron (ca. 700,000 yr BP; Miller et al., 1994). Magnetically reversed sediment is also described from western Indiana (Bleuer, 1976). In western Iowa and other places further west, volcanic ashes and their radiometric ages, and paleomagnetostratigraphy indicate that glaciation of north-central United States began before the Matuyama Chron (ca. 2.0 my; Roy et al., 2004). The age of the upper boundary of 190,000 is based on the age of the end of Marine Oxygen Isotope Stage 7 (Martinson et al. (1987) and supported by OSL ages (McKay, 2008), and Be-10 inventories (Curry and Pavich, 1996).

Illinois Episode (190,000 – 130,000 yr BP)

Most researchers concur that the Illinois Episode corresponds to Marine Oxygen Isotope Stage 6 (190,000 to 130,000 yr BP; Martinson et al., 1987). A Be-10 inventory of the Sangamon Geosol developed in the Glasford Formation in north-central McHenry County, Illinois, indicated an age of the latter deposit of 130,000 to 170,000 yr BP (Curry and Pavich, 1996). OSL ages of sand and gravel outwash that underlie the Kellerville Till (Glasford Formation) from cores sampled in the middle reaches of the Illinois River valley are in agreement with this assessment (McKay, 2008).

Sangamon Episode (130,000-55,000 yr BP)

The beginning of the Sangamon Episode, the penultimate interglaciation, is estimated to be 130,000 yr BP (Martinson et al., 1987). There is no referent section for the Sangamon Episode. The fossiliferous successions at Raymond and Pittsburg basins (Curry and Baker, 2000), and the paleosol-loess succession at Athens Quarry (Follmer et al., 1979; Curry and Follmer, 1992) are viable candidates, but are not well-dated.

The end of the Sangamon Episode, 55,000 years ago, is marked by the onset of loess deposition associated with the last glaciation (Wisconsin Episode). This age estimate comes with caveats, however. Based on extrapolation of uncalibrated radiocarbon ages, McKay (1979), Leigh and Knox (1993), and Curry and Follmer (1992) arrive at an age of ca. 55,000 C-14 yr BP. Investigation of the stable isotope profiles of U-series dated speleothems at Crevice Cave, Missouri, indicated the very same age of 55,000 yr BP (Dorale et al., 1998), but this age is neither extrapolated, nor does it need calibration. Moreover, the age is precise, with an error of about 500 yrs (less than 1%). This age, however, is not based on a material referent, but rather on an interpretation of vegetation change based on $\delta^{13}C$ values of the speleothems. Presently, the radiocarbon record is confidently calibrated to 24,000 C-14 yr BP (Reimer et al., 2004), and work is underway to calibrate the older portion of the record. Before it is, it is prudent not to attempt extrapolation of age of radiocarbon ages may require significant correction due to geomagnetic events such as the Mono Lake and Leschamps excursions (Grimm et al., 2006). At present, use of Dorale et al’s age of 55,000 yr BP for the end of the Sangamon Episode is recommended.

A note about calibration. In the remainder of this section, radiocarbon ages that were calibrated using Calib 5.02 (Reimer et al., 2004) will be denoted by “cal yr BP”, and uncalibrated radiocarbon years as “C-14 yr BP”. Radiocarbon ages greater than 24,000 C-14 yr BP were calibrated using CalPal online (Danzeglocke et al., 2008). Table 3 compiles the radiocarbon ages from the study region, and identified those ages that help to constrain or define diachronic boundary ages.

Wisconsin Episode

Athens Subepisode (55,000 yr BP-29,000 to 24,300 cal yr BP)

The end of the Athens Subepisode is marked by the age of the top of Robein Member of the Roxana Silt. In northeastern Illinois, the youngest age is about 29,000 cal yr BP (24,000 C-14 year BP; Curry and Yansa, 2004; Curry, 2007a). In areas of thin loess, such as the Lomax site in western Illinois, the age is as young as about 24,300 cal yr BP (20,350 C-14 yr BP; Curry, 1998). At Athens Quarry in central Illinois where the combined thickness of the Peoria and Roxana silts is about 6 m, the age is about 30,000 cal yr BP (25,000 C-14 yr BP; Curry and Follmer, 1992). This boundary is the most time-transgressive time boundary in the Illinois Pleistocene stratigraphic record, probably because it is so readily identified, occurs over a broad area, and contains abundant wood fragments for radiocarbon assay.

Michigan Subepisode (29,000-16,250 cal yr BP for till and outwash; 29,000-14,670 for loess)

The upper boundary of the Michigan Subepisode marks the end of the last glaciation in northeastern Illinois. The youngest known age (16,250 cal yr BP (13,650 ± 40 C-14 yr BP)) is of *Dryas integrifolia* leaves and stems from the base of an ice-walled lake succession on the Deerfield Moraine in Wadsworth, Illinois. The age is about 350 years younger than 14,000 C-14 yr BP indicated by Hansel and Mickelson (1985) and Hansel and Johnson (1992, 1996). The age is also about 240 years younger than the age of 13,890 ± 120 C-14 yr BP reported from wood fragments buried by sand and gravel in the Glenwood spit formed by Glacial Lake Chicago during the Glenwood I Phase. These dates confirm that the Lake Michigan lobe waxed and waned in Glacial Lake Chicago while it formed the Lake Border moraines.

In Illinois, the upper boundary of the the Michigan Subepisode with the Hudson Episode is the age of the top of Peoria Silt (Hansel and Johnson, 1996). In many places, this corresponds to ground surface, but in most river valleys and depressions, the Peoria Silt was redeposited by slopewash, or by suspension and settling in water. Investigations at Mastodon Lake (Stop 5) and Brewster Creek (Curry et al., 2007) indicate that the rate of loess resedimentation in shallow lakes abruptly slowed at about 14,670 cal yr BP (12,500 C-14 yr BP). The boundary is demarcated by terriginous silt loam below and biogenic marl above. The age of this boundary coincides with the boundary between the Oldest Dryas and Bölling Chronozones (Stuiver et al., 1995). Moreover, this age roughly corresponds with the drop in the level of Glacial Lake Chicago from the Glenwood to the Calumet levels (Hansel and Mickelson, 1988).

Burlington Phase(?), Michigan Subepisode, age unknown. Originally mapped as part of the Elburn Complex by Willman and Frye (1970), a well-defined ridge in northwestern Kane County and northeastern DeKalb County has been informally dubbed the Burlington moraine by Curry (2008). On its northern end, the crest of the Burlington moraine appears to be truncated by the Marengo Moraine near the village of Hampshire (Figure 18). The Burlington moraine is thus older than the Marengo Moraine. Both ridges are formed of diamicton of the Tiskilwa Formation (Curry, 2008). Several radiocarbon ages establish that the Marengo Moraine formed by 24,000 C-14 yr BP (see below), but as of yet, there are no radiocarbon ages of Robein Member below the Burlington Moraine to verify the cross-cutting relationship described above.

Marengo Phase, Michigan Subepisode (29,000 to ≈ 27,200 cal yr BP). The earliest glacial deposits of the last glaciation include proximal lacustrine sediment and outwash. In many places, these sediments were overridden during the Marengo Phase by the Harvard sublobe, which deposited diamicton and sorted sediment of the Tiskilwa Formation (Figure 19). The age of the advance is documented at the LaFarge sand pit (Stop 7) by radiocarbon ages of a stump dated about 28,970 cal yr BP (24,000 ± 390 C-14 yr BP (ISGS- 2108; Curry et al., 1999)), and

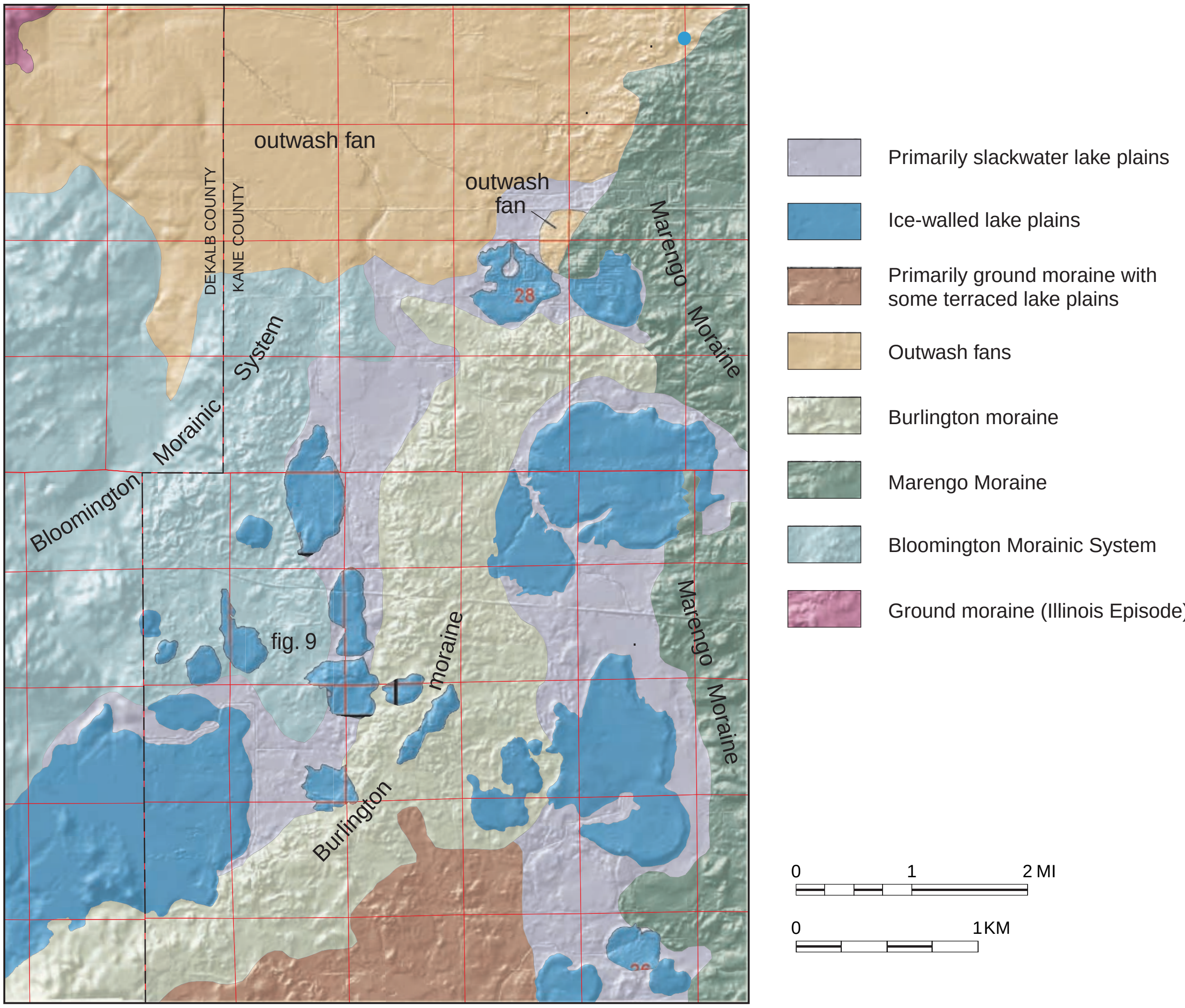


Figure 18 Shaded relief map of 2-m DEM of a portion of the Hampshire 7.5-minute Quadrangle showing truncation of the Burlington moraine against the Marengo Moraine.

at a quarry near Elgin, a wood date of about 28,990 cal yr BP (24,000 ± 270 C-14 yr BP; Curry and Yansa, 2005). The Harvard sublobe was severed from the main body of the Lake Michigan lobe by the Princeton Sublobe at about 27,200 cal yr BP (21,370 C-14 yr BP). This age is based on two dates on wood fragments sampled from a tundra-like plant fossil assemblage discovered at Wedron Quarry of 21,460 ± 470 C-14 yr BP (ISGS-1486; Garry et al, 1990) and 21,370 ± 240 C-14 yr BP (ISGS-2484; Hansel and Johnson, 1996).

A note about radiocarbon ages associated with ice-walled lakes. Peppered throughout the discussion of the age of phases of the Michigan Subepisode are radiocarbon ages of tundra plants from laminated silt and very fine sand that form the core of ice-walled lake plains. The upper 1 to 4 m of the laminated sediment are weathered and do not contain reliably dateable material. Therefore, the youngest radiocarbon ages of ice-walled lake plain deposits provide minimum landform ages. The oldest radiocarbon ages give minimum ages of when ice deposited the underlying diamicton. Detailed radiocarbon dating of a single core from the iec-walled lake plain visited at Stop 1 suggests that sediment accumulation rates are non-linear and should not be used to estimate the age of the top of the lake deposit (Stop 1, this volume).

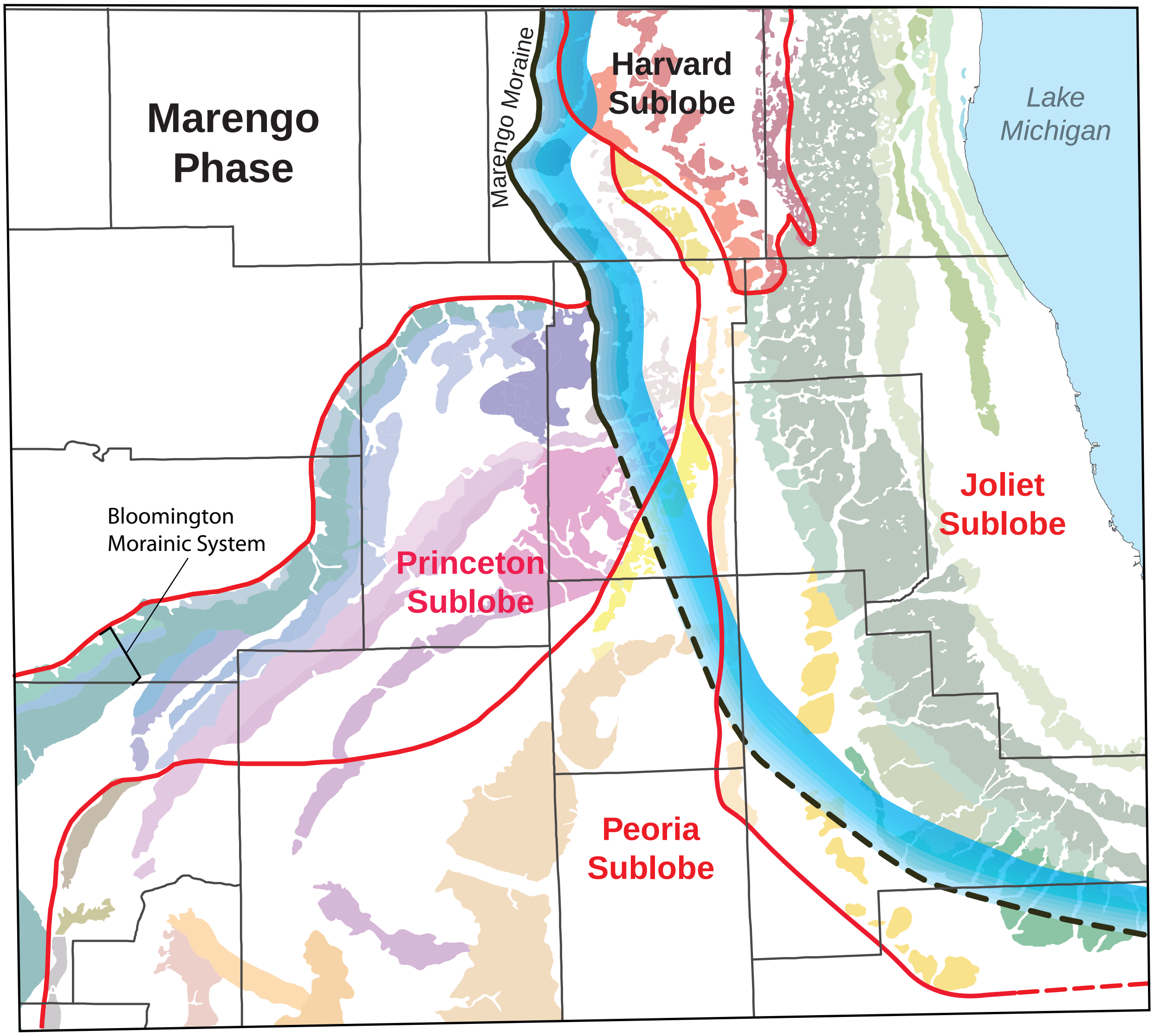


Figure 19 Moraine map showing the location of the ice margin during the Marengo Phase (29,000 to ≈ 27,200 cal yr BP). Red lines show the boundaries between the sublobes of the Lake Michigan lobe..

Shelby Phase, Michigan Subepisode (≈ 27,200 to 24,000 cal yr BP). After the Marengo Moraine was formed by the Harvard sublobe, the Princeton sublobe advanced into the southern part of the study area forming the moraines of the Bloomington Morainic System, including possibly the Burlington moraine discussed above. The change in flow direction of the lobe, as interpreted by moraine orientations was attributed by Wickham et al (1988) to interaction of the Lake Michigan lobe with the Huron-Erie lobe (Figures 20 and 21). The Northern Illinois University campus is built on ground moraine of this age (Kempton et al., 1987b).

Putnam Phase (≈ 24,000 to ≈22,000 cal yr BP). After melting back to about the Kane-DuPage county boundary, the Princeton sublobe readvanced into central Illinois to form a series of moraines, the most prominent of which is the Arlington Moraine (Figure 20). Radiocarbon ages as young as 15,250 C-14 yr BP of fossils contained in ice-walled lake plains located to the west near Hampshire indicate that the Harvard sublobe stagnated during the Putnam Phase and

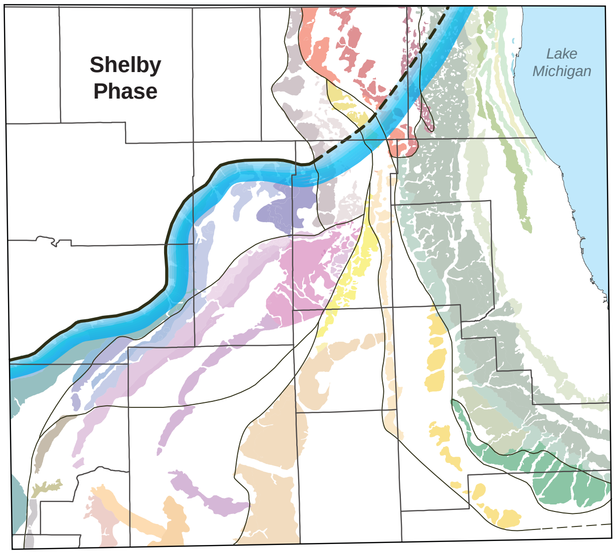


Figure 20 Moraine map showing the location of the ice margin during the the Shelby Phase (≈27,200 to 24,000 cal yr BP).

early Livingston Phase, described below. Presently, there are no reliable ages from materials associated with the Putnam Phase.

Livingston Phase (≈22,000 to ≈20,000 cal yr BP). Although at present there are no ages that define the boundaries of the Livingston Phase, there are dates of about 20,750 cal yr BP from two localities that indicate when there was a significant change in the flow direction of the Lake Michigan lobe. During the early part of the Livingston Phase, the Princeton sublobe formed the St. Charles Moraine and possibly the Barlina Moraine, followed by formation of the Marseilles Morainic System by the Peoria Sublobe (Figure 23). After these sublobes retreated to an unknown position, the ice readvanced as the Joliet sublobe to form the Rockdale, Minooka, and possibly Barlina moraines (Figure 24). This change in flow behavior is attributed to cessation of interaction of the Lake Michigan lobe with the Huron-Erie lobe, a condition that had begun during the Shelby Phase (Figure 21). The age in the change in glacier behavior is about 20,750 cal yr BP (17,540 ± 130 yr BP; ISGS-A-0021). The age comes from assay of tundra plant fossils recovered from lacustrine sediment sandwiched by till of the Yorkville Member below that forms the St. Charles Moraine, and overlying sorted sediment associated with the Minooka Moraine (Curry et al., 1999; Curry and Yansa, 2004; Figure 24). The 20,750 cal yr BP age is confirmed by an age of 20,970 cal yr BP (17,760 ± 60 C-14 yr BP; UCI-AMS-26260) of tundra plant fossils recovered from the upper part of a large ice-walled lake plain deposit on the Marseille Morainic Complex near the Newark Channel (Figure 23). The interval that is dated here is buried by about 4 m of unfossiliferous, pedogenically altered lacustrine sediment; the age thus provides a minimum age for the Marseille Morainic System.

Woodstock Phase (≈20,000 to 18,150 cal yr BP). During the Woodstock Phase, the coalesced Harvard and Joliet sublobes formed the Woodstock Moraine and the lower part of the West Chicago Moraine. Extensive outwash plains associated with these moraines were deposited during this time (Figure 25). There are no ages to constrain the beginning of the Woodstock Phase. The end of the Woodstock Phase is limited by ages of about 17,640 cal yr BP (14,860 ± 110 C-14 yr BP (ISGS-A-0165)) of fossils collected at the base of a kettle fill on the Woodstock Moraine in Crystal Lake (Curry, 2005), and an age of 17,950 cal yr BP (14,780 ± 50 C-14 yr BP; UCI-AMS-46831) on the eaves and stems of *Dryas* from laminated silts forming an ice-walled lake plain near Streamwood, Illinois (Figure 25).

Kankakee Torrents. Ekblaw and Athy (1925) were the first to describe features such as boulder-topped alluvial successions, and large transverse bars in the Illinois River valley attributed to large meltwater discharge. The legacy of a large flood has held the interest of Quaternarists ever since, but curiously there has been no concentrated study directed at understanding this event or related events, in part because of the paucity of dateable material in the flood-related sediment. We are beginning to piece together fragments of the history of the "Kankakee Torrent", and one conclusion seems inescapable: there was more than one flood, and that the youngest flood was among the largest.

Three erosional channels cut across the Marseilles Morainic System located near the communities of Oswego, Newark, and Marseilles (Figure 26). The Marseilles channel is occupied today by the Illinois River. A remarkable 9-m thick succession of fossiliferous lacustrine sediment, marl, and peat occurs below the floor of the Oswego channel (Stop 4). Four AMS C-14 age of tundra plant fossils are the basis for the minimum age of channel erosion; the weighted mean age is about 18,920 cal yr BP (15,690 ± 35 yr BP). A younger flood is interpreted from a well-dated slackwater lake deposit located in the lower Illinois River valley near the town of Havana. The oldest age, on fine organic material from near the base of the deposit, is 18,030 cal yr BP (14,830 ± 50 C-14 yr BP; Hajic, 2007). This age is similar to several ages that post-date retreat of the Harvard sublobe during the Woodstock Phase, and the onset of the Milwaukee Phase, discussed below. An investigation of isostatic rebound indicates synchronous initial formation of the three channels (Stop 4).

Milwaukee Phase (< 18,100 and > 17,300 cal yr BP). During the Milwaukee Phase, the coalesced Harvard and Joliet sublobes melted back to as far north as about Milwaukee, Wisconsin, depositing lacustrine sediment attributed to Glacial Lake Milwaukee (Schneider and Need, 1985). Later, the lacustrine sediment was buried by glacigenic diamicton of the Wadsworth Member of Illinois and Oak Creek Member of Wisconsin during the Crown Point Phase, discussed below. Correlative lacustrine sediment comprises a large proportion of the glacial drift succession in northeastern

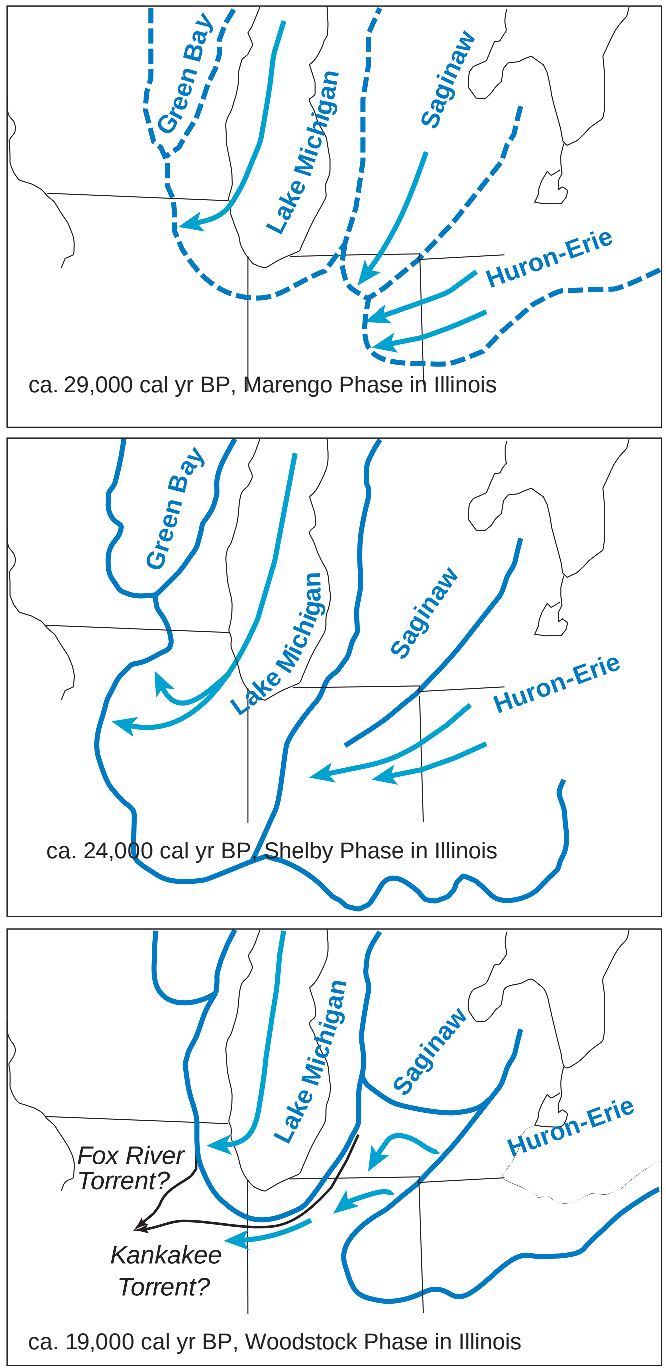


Figure 21 Major lobes of the south-central part of the Laurentide Ice Sheet at about 29,000, 24,000 and 19,000 cal yr BP showing how interaction with the Huron-Erie lobe impacted the flow direction of the Lake Michigan lobe (from Curry et al., 2007). During this period of interaction (which began at about 29,000 cal yr BP during the Shelby Phase), the west-flowing Erie lobe abutted against the south-flowing Lake Michigan lobe. The interaction caused the Lake Michigan lobe to flow in a more westerly direction resulting in arcuate moraines with similar orientations as the end moraine of the Erie lobe (Iroquois Moraine). As the Erie and Lake Michigan lobes melted back during the latter part of the Livingston Phase, the interaction ceased such that the subsequent ice advance that formed the younger Minooka Moraine was parallel to the modern southwestern shoreline of Lake Michigan.

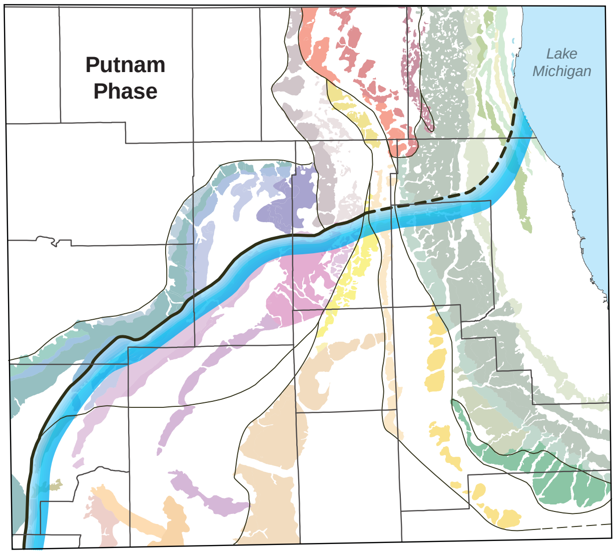


Figure 22 Moraine map showing the location of the ice margin during the Putnam Phase (≈ 24,000 to ≈ 22,000 cal yr BP)

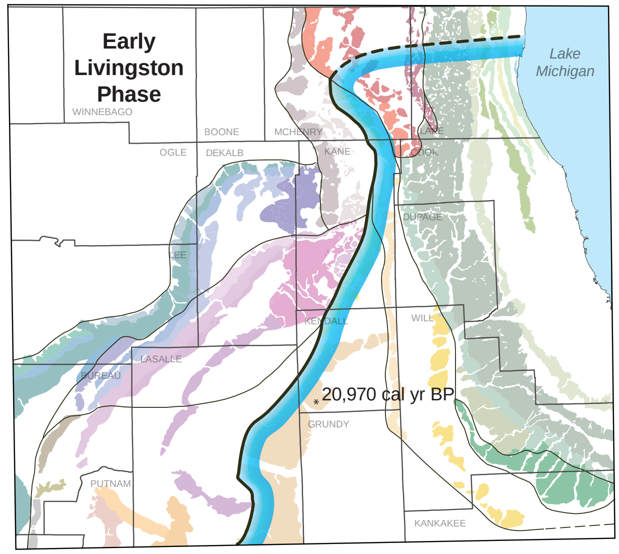


Figure 23 Moraine map showing the location of the ice margin during the early Livingston Phase (≈22,000 to 20,750 cal yr BP)

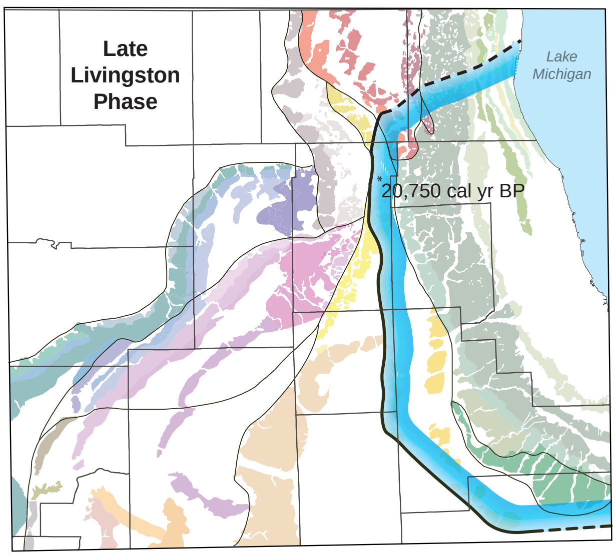


Figure 24 Moraine map showing the location of the ice margin during the late Livingston Phase (20,750 to ≈20,000 cal yr BP)..

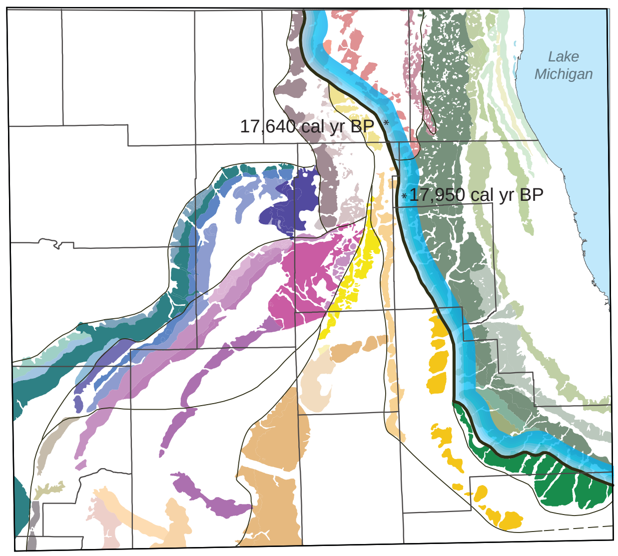


Figure 25 Moraine map showing the location of the ice margin during the Woodstock Phase (≈20,000 to 18,150 cal yr BP).

Illinois. Some of the earliest lacustrine sediment related to Glacial Lake Milwaukee will be visited at Stop 9. Here, a proglacial delta deposited sorted sediment into a small basin formed as the ice margin melted back and stalled at the position of the Fox Lake Moraine. The bounding ages of the Milwaukee Phase are not known, but are approximated by a basal age of about 18,100 cal yr BP (14,900 C-14 yr BP), the age of tundra fossils in ice-walled lake sediments covering the Woodstock Moraine, and an upper age of about 17,300 cal yr BP, the bottom age of an ice-walled lake plain succession on the Tinley Moraine, discussed below.

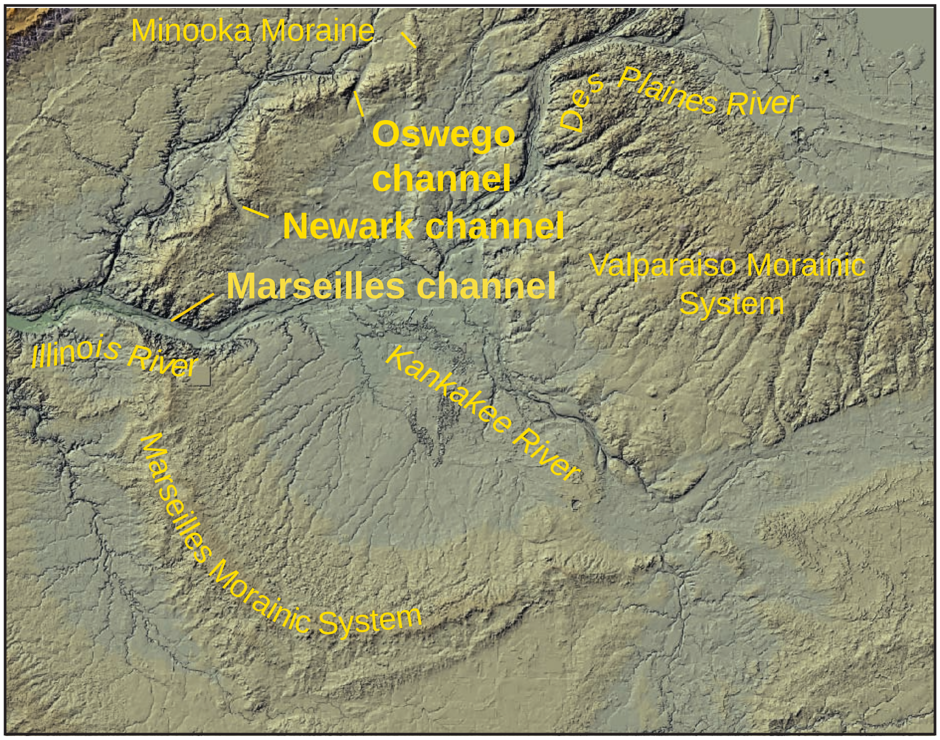


Figure 26 Shaded relief map of 30-m DEM of northern Illinois showing location of the Oswego, Newark, and Marseilles channels across the Marseilles Morainic System.

Crown Point Phase (>17,300 and <16,700 cal yr BP). During the Crown Point Phase, the Joliet sublobe formed the Valparaiso Morainic System, and subsequently the Lake Border Moraines, including the Tinley, Park Ridge, and Deerfield moraines. (Figure 27). Radiocarbon ages from the top and bottom of two ice-walled lake plain deposits located on the Tinley Moraine and the Deerfield Moraine provide the chronology of the Crown Point Phase (Figure 28). The dates associated with the Tinley Moraine ice-walled lake plain deposit are 17,290 cal yr BP (14,420 ± 40 C-14 yr BP; UCIAMS-26264), 16,820 cal yr BP (14,110 ± 35 C-14 yr BP; UCIAMS-26263), and 16,580 cal yr BP (14,070 ± 40 C-14 yr BP; UICAMS-26262). Fossil *Dryas* stems and leaves from the base of a succession of fossiliferous lake sediments on the Deerfield Moraine yielded an of about 16,250 cal yr BP (13,650 ± 40 C-14 yr BP; UICAMS-46829), confirming that the Lake Border moraines were formed during the existence of Glacial Lake Chicago, discussed below.

Glenwood Phase (<16,700 and >14,670 cal yr BP). During the Glenwood Phase, the Lake Michigan lobe began toretreat from what is now the mainland of Illinois. Glacial Lake Chicago formed between the ice margin and the mainland, depositing littoral, coastal, and lacustrine sediment (Figure 29). During an early part of the Glenwood Phase, Glacial Lake Chicago eroded the Chicago Outlet across the Tinley Moraine and Valparaiso Morainic System. The level of Glacial Lake Chicago stabilized at about 195 m, an elevation controlled in part by

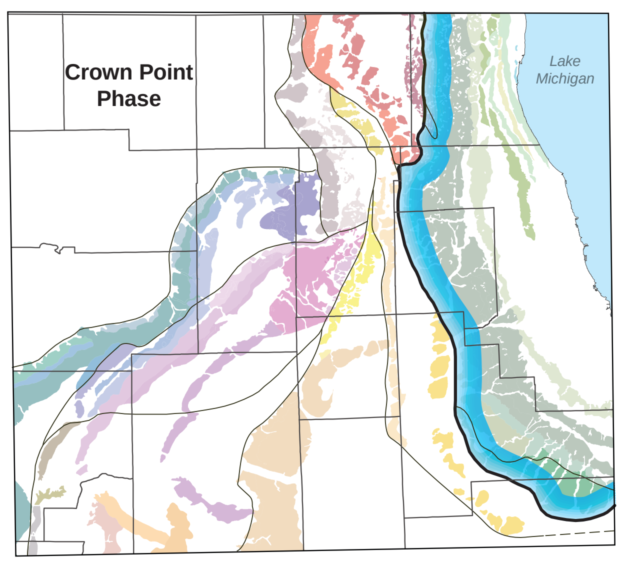


Figure 27 Moraine map showing the location of the ice margin during the Crown Point Phase (≈17,300 to ≈16,700 cal yr BP)

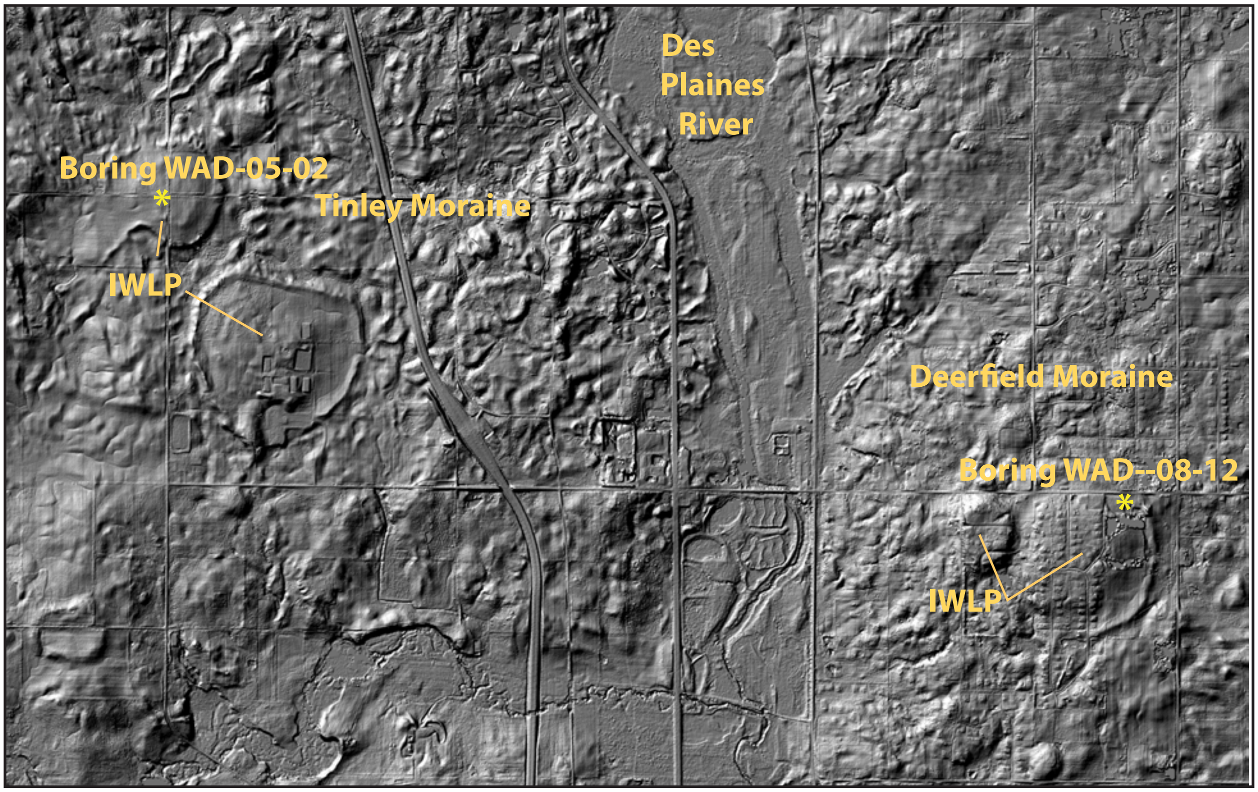


Figure 28 Shaded relief map of LiDAR data from Lake County showing portions of the Tinley and Deerfield moraines in Old Mill Creek and Wadsworth, Illinois. Radiocarbon ages from boring WAD-05-02 on the Tinley Moraine include 17,290 cal yr BP, 16,820 cal yr BP, and 16,580 cal yr BP. The radiocarbon age from boring WAD-12-08 on the Deerfield Moraine is 16,250 cal yr BP.

the bedrock-defended floor of the Chicago outlet (Bretz, 1951). The Glenwood Phase includes a two high stands known as the Glenwood I and Glenwood II levels (Figure 30). Of importance to this discussion are littoral deposits formed during the Glenwood I level. Wood buried by sand and gravel of the Glenwood spit located in the southern part of the lake include ages of about 16,240 and 16,310 cal yr BP (Hansel and Johnson, 1992). These ages are chronologically consistent with the youngest age of about 16,770 cal yr BP from the ice-walled lake plain deposits on the Tinley Moraine described above. Ice later readvanced into Glacial Lake Chicago several times to form the Lake Border Moraines; the basal age of 16,250 cal yr BP associated with the ice-walled lake plain deposit on the Deerfield Moraine is in harmony with this interpretation.

Glenwood I and II levels were separated by the intra-Glenwood (Mackinaw) low-water phase that dates at about 13,470 ± 130 C-14 yr BP (ISGS-1378; Monaghan and Hansel, 1990). During this time, the Lake Michigan lobe retreated north to the point that the Chicago outlet was abandoned in favor of drainage to the north across the Straits of Mackinaw or Indian River lowland in northern Michigan. During the Glenwood II lake phase, ice readvanced during the Port Huron glacial phase causing drainage to revert back through the Chicago outlet. At about 14,670 cal yr BP (12,500 C-14 yr BP), the level of Glacial Lake Chicago dropped to the Calumet level at about 175 m (Figure 30). This drop in lake level was attributed by Bretz (1951) to lowering of the level of the Chicago outlet, but other hypotheses have been advocated (Hansel and Mickelson, 1988; Colman et al., 1994; Chrzastowski and Thompson, 1994) such as a changes in discharge attributed to climate change (Hansel and Mickelson, 1988). With respect to the findings of recent investigations on the mainland, it is notable that drop from the Glenwood II to Calumet I levels

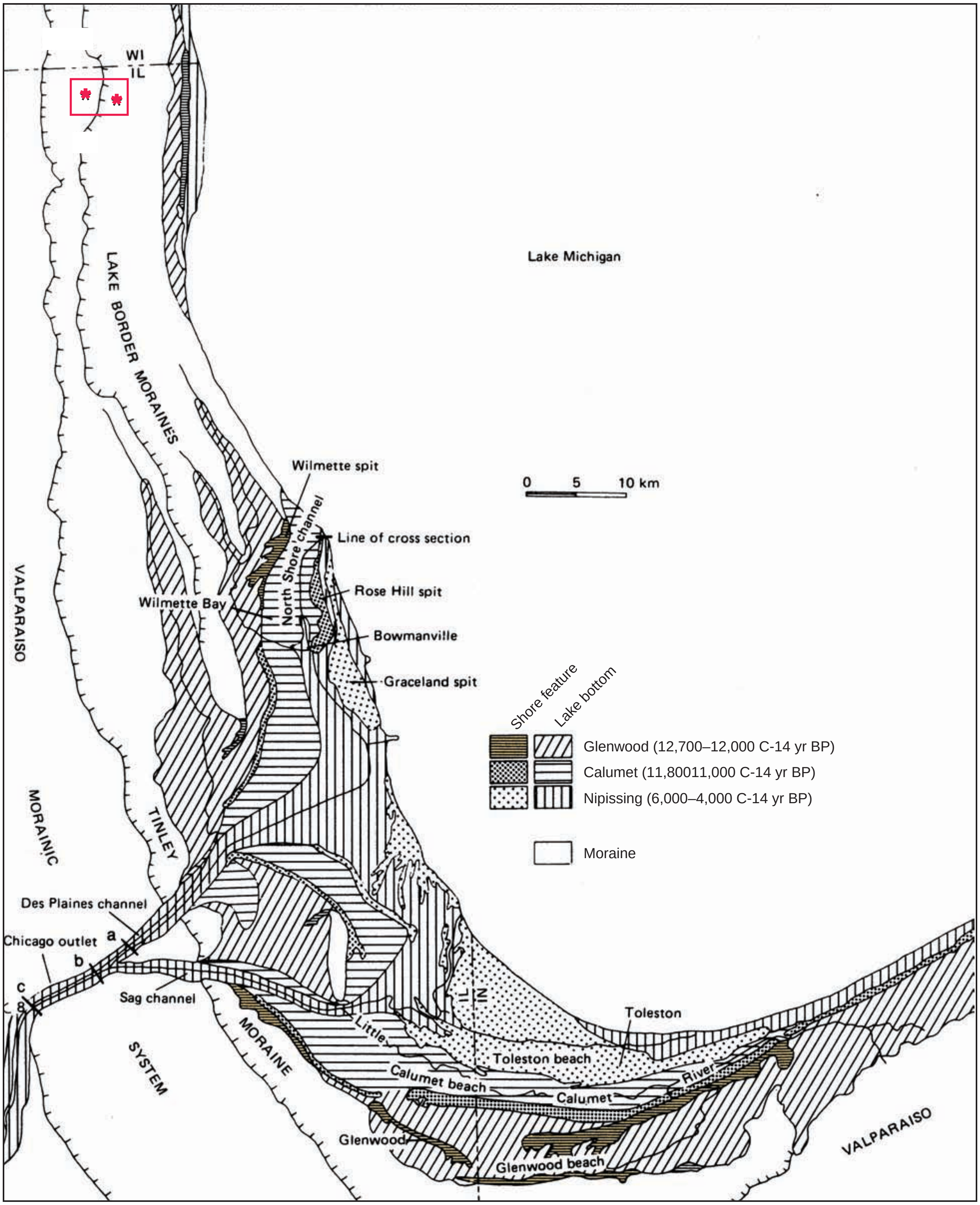


Figure 29 Map of Glacial Lake Chicago and its relationship with the Lake Border Moraines (after Hansel and Mickelson, 1988). The location of Figure 28 and the two dated ice-walled lake plains are shown in the red box.

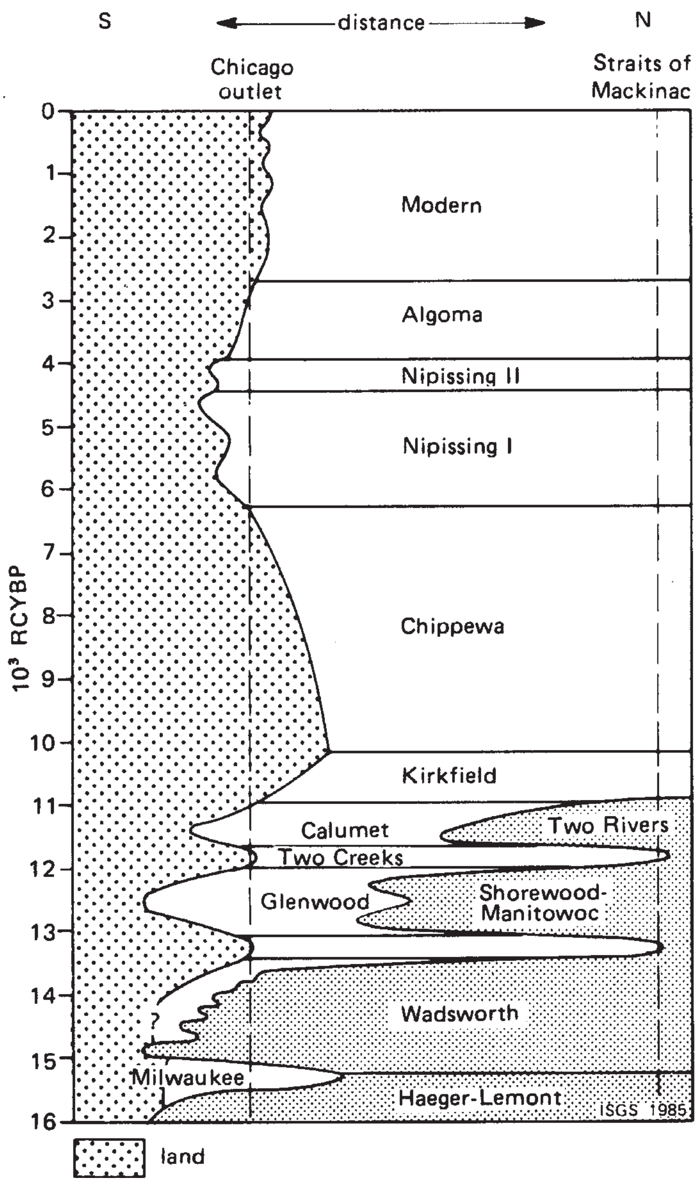


Figure 30 Glacial and lake phases associated with Glacial Lake Chicago (after Hansel and Mickelson, 1988).

corresponds in time with when loess accumulation rates slowed considerably in shallow lakes in northeastern Illinois (discussed below).

Keller Farm Phase [new](in study area, post-Michigan Subepisode) (20,900 to 14,670 cal yr BP). A new phase is proposed for the time when Peoria Silt was deposited. In the time-distance diagrams of Hansel and Johnson (1992; 1996), Peoria Silt is shown to have been deposited during the Michigan Subepisode (Figure 17). By this definition, the material should be related in some fundamental way to sediment comprising the Wedron Formation. However, the Peoria is mineralogically distinct from sediment of the Wedron Formation, with the former containing primarily expandable clay minerals, and the latter, illite and chlorite (Graese et al., 1988). The particle-size distribution is likewise distinct from deposits represented by the Wedron Formation; the former is primarily well-sorted silt loam to silty clay loam; the latter, matrix-supported diamicton (Table 1). Mason et al. (2007) show that the Peoria Silt in eastern Nebraska is partly, if not mostly, derived from non-glacial alluvium derived from bedrock occurring in the High Plains. Hence, there are neither compositional nor textural, and in at least eastern Nebraska, genetic reasons to have the Michigan Subepisode embrace the Peoria Silt as a material referent.

The well-dated section at Keller Farm, Illinois (Wang et al., 2003) is suggested for the material referent section for the Keller Farm Phase. In northeastern Illinois, data from core BC-1 at Brewster Creek is a proposed paratype section (Curry et al., 2007). In this core, the lower boundary is marked by a change in lithology from glacigenic diamicton or outwash below to sorted, silt-rich sediment above. In depressions and on some slopes, the upper boundary is marked by silty, terriginous sediment below, and biogenic marl and peat above (Figure 31). The terriginous-poor material was deposited during the Hudson Episode, discussed below. Several radiocarbon ages of spruce needles picked from of the contact at the Brewster Creek site indicates that the change in lithology occurred at 14,670 cal yr BP (ca. 12,500 C-14 yr BP; Figure 31; Curry et al., 2007). The upper boundary is ground surface in most upland situations.

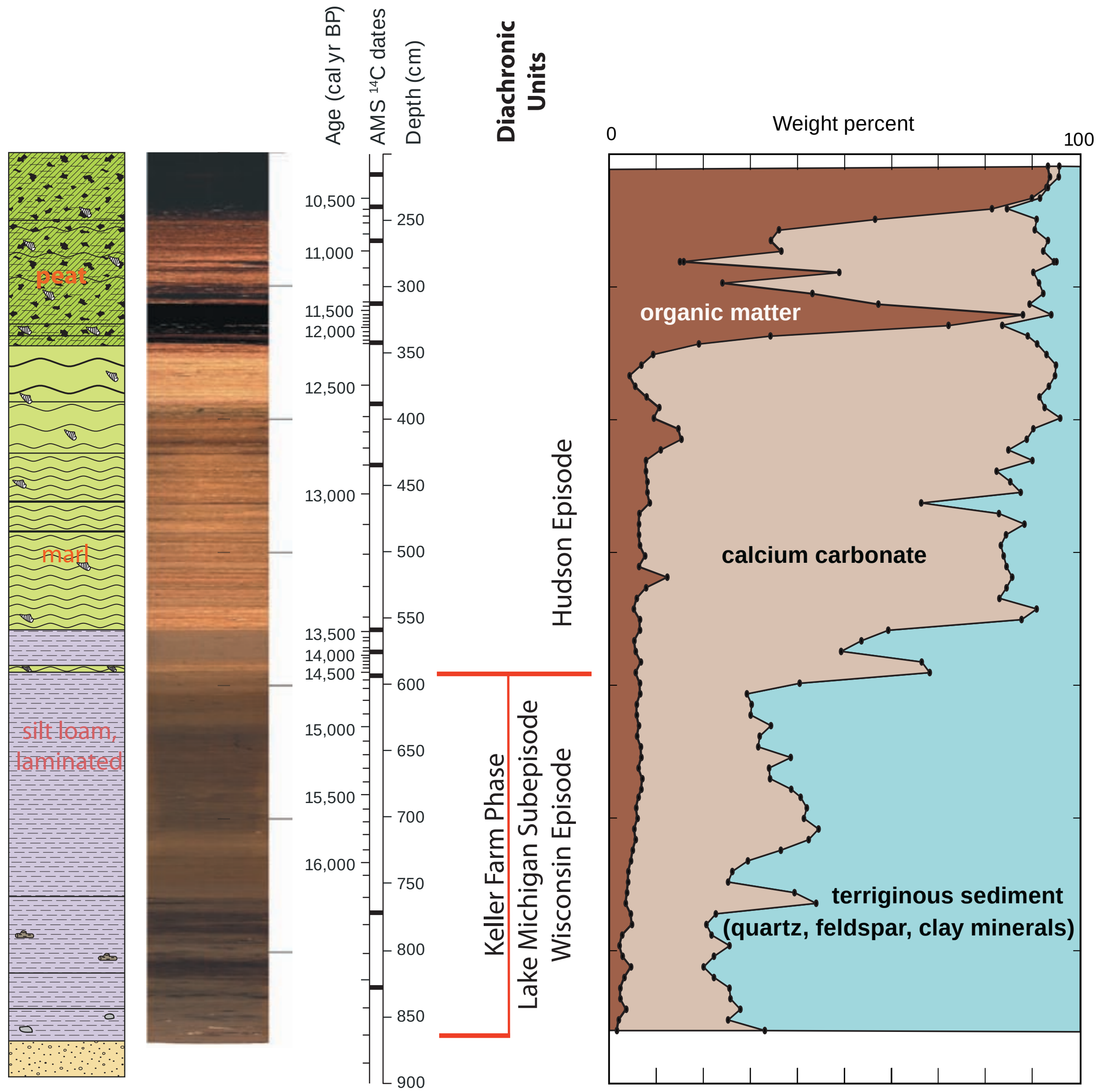


Figure 31 Downcore loss-on-ignition data and radiocarbon ages from core BC-1 at the Brewster Creek site showing the lithologic contact demarcating the terriginous-rich silt of the Wisconsin Episode (Michigan Subepisode) below and the biogenic marl and peat-rich material of the Hudson Episode. From Curry et al. (2007).

The results of recent work in northeastern Illinois poses an additional dilemna regarding diachronic classification. A glacial “phase”, as used by Hansel and Johnson (1996) is the time associated with deposition of a material referent (i.e., a lithostratigraphic unit). Radiocarbon dating of fossils in ice-walled lake sediment indicates that the latter stages of sedimentation associated with stagnating ice may overlap in time with initial formation of an adjacent moraine composed of another material referent. For example, the sediment comprising ice-walled lake plains of the DeKalb region date from about 20,400 to 18,600 cal yr BP (17,250 to 15,150 C-14 yr BP; Curry, 2008; Stop 7); this time corresponds with the Livingston and Woodstock Phases when the Yorkville and Haeger Members were respectively deposited. The lakes occur beyond the mapped limits of these units, and the age of their sediments transgresses multiple glacial phases. Hence, they should have their own unique name and referent sections. Since phases were not intended to overlap (North American Commission of Stratigraphic Nomenclature, 1983; Hansel and Johnson, 1996), a special case is necessary for ice-walled lake plain deposits. To avoid a proliferation of new diachronic names, it is proposed that the ice-walled lake deposits in Illinois were deposited during the early Keller Farm phase.

Hudson Episode (14,670 cal yr BP – present). The Hudson Episode began when loess-related sedimentation ceased after the last glaciation. As discussed above, this time boundary at the Brewster Creek site coincides with a material change from silt loam to sediment that is dominated by biogenic marl and peat. At this site, the lake was relatively shallow (< 7 m deep). The sediment record from deeper lakes (17-20 m deep) in northeastern Illinois (e.g. Crystal Lake in SE McHenry County) are complicated by layers of sandy sediment probably derived from turbidites that delivered littoral sediment to profundal zones by deep, wind-driven waves during storms (Curry, unpublished data).

The deglacial and postglacial history of northeastern Illinois is known from pollen and ostracode records from cores sampled from Nelson and Mastodon lakes (Kane County), Crystal Lake (McHenry County), and Brewster Creek (Curry et al., 2007). A synopsis of the late Pleistocene and Holocene vegetation history begins with replacement of tundra vegetation by a dense boreal forest at about 16,700 cal yr BP (data from Nelson Lake and from Stop 5). This coniferous forest was dense and comprised almost exclusively of white and black spruce, and persisted to about 14,670 cal yr BP. During the transition from the last glacial to present interglacial (14,670 cal yr BP to about 10,700 cal yr BP), deciduous trees became more important, especially black ash. The dominance in the pollen record of this wetland-loving tree and spruce indicates that northeastern Illinois remained wet during the transition. At Brewster Creek, spruce pollen persisted to about 10,500 cal yr BP, about 650 years longer than at Crystal Lake (Gonzales and Grimm, unpublished data). The nature of the remainder of the Holocene vegetation record has been investigated in Nelson Lake (Nelson et al., 2006) and in Crystal Lake to about 8,000 cal yr BP (Gonzales, unpublished data). The early Holocene deciduous forest included abundant oak, elm, and black ash, suggesting conditions somewhat wetter than present. Pollen-assemblages and the $\delta13C$ values and accumulation rates of charcoal reveal that the prairie initially developed from about 9,000 to 7,300 cal yr BP. There was a brief return to wetter conditions from 7,300 to about 6,200 cal yr BP after which there was full expansion of the prairie (Nelson et al., 2006). Tree stands were comprised chiefly of oak, hickory, and walnut, and persisted to when the area was settled by Europeans in the early-mid 1800's.

The modern landscape largely reflects the glacial legacy presented above although the postglacial landscape has been smoothed somewhat by infilling of depressions and low areas with alluvium and biogenic sediment. Since the encroachment of settlers more than 150 years ago in northeastern Illinois, many trees have been removed, swampy soils have been drained for agricultural development, and large amounts of sand and gravel have been extracted for aggregate used in the construction of roads and foundations. Lakes, both natural and man-made, have filled in quickly with sediment derived from agricultural practices and urban improvements. As the population of northeastern Illinois grows, the glacial deposits have found new significance as both a receptacle for waste disposal and as a groundwater resource, in addition to continued use in agriculture and as an aggregate resource.

Table 3 Comprehensive list of radiocarbon ages. Site locations shown on Figure 32.

No.	Lab number	Site or boring (depth, cm)	Material dated	C-14 age (BP)	±	Cal year (BP) sigma 1	Cal year (BP) mean	Cal year (BP) sigma 1
15	Mastodon Lake, kettle, variable lithology, Stop 5							
	CAMS-116407	Base peat, trench, island	bulrush seeds	9115	35	10254	10210	10390
	CAMS-110992	museum specimen	Collagen	10430	40	12340	12120	12610
	UCIAMS-19321	museum specimen	Collagen	10190	25	11770	11900	12030
	CAMS-110994	museum specimen	Collagen	10980	60	12920	12850	13050
	UCIAMS-19329	museum specimen	Collagen	11130	30	12940	13030	13110
	CAMS-110995	museum specimen	Collagen	11320	50	13200	13110	13290
	ISGS-5657	Top of marl, main cut	wood	11520	90	13370	13210	13600
	ISGS-5790	Base of marl, main cut	wood	12360	130	14400	13980	14930
	ISGS-5633	Silt unit, trench on island	wood	13600	70	16190	15800	16610
	ISGS-5656	Silt unit, main cut	wood	13710	70	16330	15960	16750
	ISGS-5655	Silt unit, main cut	wood	14130	70	16850	16420	17260
14	Brewster Creek core BC-1, kettle, variable lithology							
	CAMS-105794	216 cm (peat)	Larch needles	9185	40	10244	10342	10487
	CAMS-105795	240 (peat)	Larch needles	9335	40	10421	10549	10671
	CAMS-105796	256 (peat)	Larch needles	9525	40	10686	10860	11082
	CAMS-116405	314 (peat)	Larch needles	9925	35	11236	11312	11596
	CAMS-111247	341.5-345 (peat)	Larch needles	10555	40	12395	12430*	12738
	CAMS-116406	389-390 (marl)	Larch needles	10495	35	12246	12530†	12678
	CAMS-111248	435 (marl)	bulrush seeds	10980	35	12854	12909	12974
	CAMS-111409	557-559 (marl)	Larch needles	11605	45	13319	13440	13598
	CAMS-111410	575-576 (marl)	Larch needles	12055	45	13793	13907	14023
	CAMS-111411	592-594 (silt)	Larch ndls + charcoal	12495	45	14247	14614	14940
	CAMS-105797	772 (silt)	*Picea* needles	13750	40	16034	16380	16761
	CAMS-105798	828 (silt)	*Picea* needles	13870	60	16148	16526	16909
13	Deerfield Moraine (Lake Border Moraine), ice-walled lake plain, fine sand							
	UCIAMS-46829	WAD 08-12 (base)	stems, leaves	13650	40	16050	16250	16430
12	Glacial Lake Glenwood spit, driftwood (sand)							
	ISGS-1549	Lynwood Res.	wood	13870	170	16240	16530	16810
	ISGS-1649	Tinley Park	wood	13890	120	16310	16550	16790
11	Tinley Moraine, IWLP, tundra plants, silt							
	UCIAMS-26262	WAD 05-02 342-366	stems, leaves	14070	40	16580	16770	16970
	UCIAMS-26263	WAD 05-02 384-392	stems, leaves	14110	35	16640	16820	17020
	UCIAMS-26264	WAD 05-02 692-695	stems, leaves	14420	40	17100	17290	17510
10	Sleepy Hollow oxbow, organic matter, silt							
	ISGS-2054	30766; 402	wood	13670	140	16010	16270	16520
9a	Cranberry Lake (West Chicago Moraine), ice-walled lake plain, tundra plant fossils, silt							
	UCIAMS-46831	STR 05-01 439-451	leaves	14780	50	17830	17950	18050
	UCIAMS-26265	STR 05-01 698-716	stems, leaves	14860	40	17960	18080	18150
9b	Nancy Drive kettle, (Woodstock Moraine); tundra fossils, silt							
	ISGS-A-0143	37144; 738	stems	14610	110	17940	18140	18240
	ISGS-A-0165	37144; 841	stems	14860	110	17450	17640	17950

West Chicago Moraine,								
9c	Higgins/Sutton Rd intersection (Streamwood Quad), plant fossils in 3-6 m thick beds							
	UCIAMS-46835	Higgins/Sutton #3: 150	*Picea* needles	13910	45	16370	16570	16770
	UCIAMS-46834-1	Higgins/Sutton #2: 243	stems, leaves	14880	80	17982	18100	18210
8	Havana, IL, slackwater lake, beyond glacial margin, organic matter, silt							
	B-207033	EMQ-33; 853	organics	14400	50	17080	17270	17500
	B-207031	EMQ-32; 1298	organics	14830	50	17890	18030	18130
7	Oswego overflow channel lake; tundra plant fossils, silt, Stop 4							
	UCIAMS-26256	KC2 853-884	stems, leaves	15750	45	18910	18960	19000
	UCIAMS-26257	KC2 884-896	stems, leaves	15660	60	18860	18910	18960
	UCIAMS-26258	KC3 853-896	stems, leaves	15670	140	18810	18920	19010
	UCIAMS-26259	KC3 899-945	stems, leaves	15470	110	18720	18800	18870
	mean of four ages above			15690	35	18880	18920	18960
6	Fox River Stone Company, delta; tundra plant fossils, silt							
	OxA-W814-13	FX-A-411	stems, leaves	17540	130	20480	20700	20880
5	De Kalb Mounds (ice-walled lake plains), tundra plant fossils, silt, Stop 7							
	UCIAMS-23773	23513; 428	stems, leaves	15150	45	18170	18590	18340
	UCIAMS-23772	35696; 415	stems, leaves	15740	150	18750	18970	19280
	UCIAMS-23765	35155; 497	stems, leaves	17290	140	20070	20420	20850
	UCIAMS-23768	35696; 268	stems, leaves	15125	45	18130	18560	18360
	UCIAMS-23770	35696; 335	stems, leaves	17090	190	19850	20220	20690
	UCIAMS-23769	35696; 579	stems, leaves	17250	60	20130	20370	20620
	OxA-W917-11	35527; 610	stems, leaves	16700	90	19560	19850	20060
	OxA-W917-9	35527; 759	stems, leaves	17610	270	20120	20820	21610
4	Marseilles Morainic Complex; ice-walled lake plain; tundra plant fossils, silt							
	UCIAMS-26260	KC5 524-579	stems, leaves	17760	60	20780	20970	21150
	UCIAMS-26261	KC5 782-792	stems, leaves	18210	60	21510	21680	21910
3	Shelbyville Morainic Complex; proglacial lake,wood; silt, (Hansel and Johnson 1996)							
	ISGS-2918	Charleston Quarry	wood	19340	180	22660	23010	23290
	ISGS-2842	Charleston Quarry	wood	19980	150	23740	23930	24140
	ISGS-2593	Charleston Quarry	wood	20050	170	23800	24010	24230
2	Marengo Moraine (NE Illinois); proglacial lake,wood; silt							
	ISGS-5632	Bluff City Quarry	wood	24700	270	28420	28850	29290
	ISGS-2108	Feltes Pit	stump	24000	390	28370	28870	29370
	ISGS-1830	Feltes Pit	wood	24360	430	28570	29170	29760
	ISGS-3982	Feltes Pit	wood	24670	220	29170	29620	30080
	ISGS-4055	Feltes Pit	wood	25820	310	30440	30850	31270

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1 Wedron, Pit 7, Stop 2 (previously published ages)							
ISGS-1486	Garry et al., 1990	wood, stems	21460	470	25070	25780	26500
ISGS-2484	Hansel and Johnson, 1996	wood	21370	240	25120	25540	25950
ISGS-863	Liu et al., 1986	wood	24370	310	28620	29150	29680
ISGS-862	Liu et al., 1986	wood	24900	200	29570	29880	30200
1 New ages							
UCIAMS-46836	5.02 m	Picea needles	25010	150	29700	29980	30250
ISGS-6216	5.03 m	wood	25100	140	29790	30040	30290
ISGS-6213	5.34 m	wood	25260	190	29860	30130	30410
ISGS-6212	6.50 m	wood	25490	160	30010	30370	30730
ISGS-6210	8.25 m	wood	25900	220	30540	30910	31280
ISGS-6211	8.75 m	wood	28670	380	32600	33130	33650

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STOP 1: Buffalo Rock State Park

Large floods and rapid deglaciation of the Lake Michigan Lobe and environs

Edward Hajic and Brandon Curry

Introduction and Purpose

Buffalo Rock is a bedrock-defended topographic high isolated within the upper Illinois River valley (Figure 1). In plan view, Buffalo Rock has a streamlined erosional form that approaches the shape of a lemniscate loop, a shape that minimizes fluid drag. Such streamlined forms elevated above bedrock or till valley floors are referred to as erosional residuals, a common geomorphic feature resulting from catastrophic outburst floods along major valleys that drained the Laurentide Ice Sheet (c.f. Kehew and Lord, 1986; 1987; Baker, 1978). Buffalo Rock formed during the Kankakee Torrent (Ekblaw and Athy, 1925; Athy, 1928), the term used for one or more floods of catastrophic magnitude that coursed down the Illinois Valley, so named for evidence of torrential flooding in the Kankakee River subbasin of the Illinois River drainage basin. Hajic (1990; Hajic and Johnson, 1989) concluded that the bedrock valley trench of the upper Illinois River, and many other macro-geomorphic features in the upper Illinois Valley region and along the length of the Illinois Valley resulted from one or more catastrophic floods of the Kankakee Torrent. The purpose of this stop is to review the evidence for catastrophic flooding that cut the upper Illinois River valley bedrock trench, and accounts for a range of large-scale geomorphic features along the valley and within lake basins drained by the flooding (Hajic, 1990). Collectively, floods of the Kankakee Torrent and the resulting suite of mega-geomorphic features just post-date what will be discussed at most stops on the fieldtrip, but there is some interesting temporal synchronicity between retreat and disintegration of the Lake Michigan lobe and onset of catastrophic flooding.

Catastrophic Flood Geomorphology of the Upper Illinois Valley

Kehew and Lord (1986) described the suite of large-scale geomorphic features resulting from catastrophic glacial lake outbursts. In addition to the bedrock valley trench, functionally an inner flood channel, these features include an outer erosional zone, large-scale bars, spillways, erosional residuals, underflow fans at the heads of lake basins, and incised lake outlets. The entire suite of geomorphic features, as well as associated sediments, was identified in and along the Illinois valley (Hajic, 1990) (Figure 2).

The outer erosional zone is believed to be the product of initial scour by floodwater where there is no well-defined pre-existing channel large enough to transport the flow (Kehew, 1982). Uplands between end moraines from the Marseilles Moraine to the bend at Hennepin show evidence of oriented scour and channelization to varying degrees (Figure 3a). North of the valley, where best expressed, shallow anastomosing channels influence the course of many first order tributaries and several higher order tributaries. Contrasting tonal patterns on high altitude aerial photography reflect subtle channels and erosional patterns. Maximum altitude of scour is about 195.1 m (640 ft). Channels were cut into thin drift away from the valley and into bedrock along the valley. The lip of the bedrock valley illustrates various degrees of convexity. Locally, multiple benches are eroded into the lip, and were cited as evidence supporting a step-wise incision of the upper Illinois valley (Willman and Payne, 1942). Several streamlined deposits of generally poorly sorted sand and gravel exhibiting steep foreset cross-stratification occur upon scoured upland surfaces near the valley trench. These represent bars deposited in the outer erosional zone as a late phase of the scouring, perhaps as flow was concentrated and incision of the in-

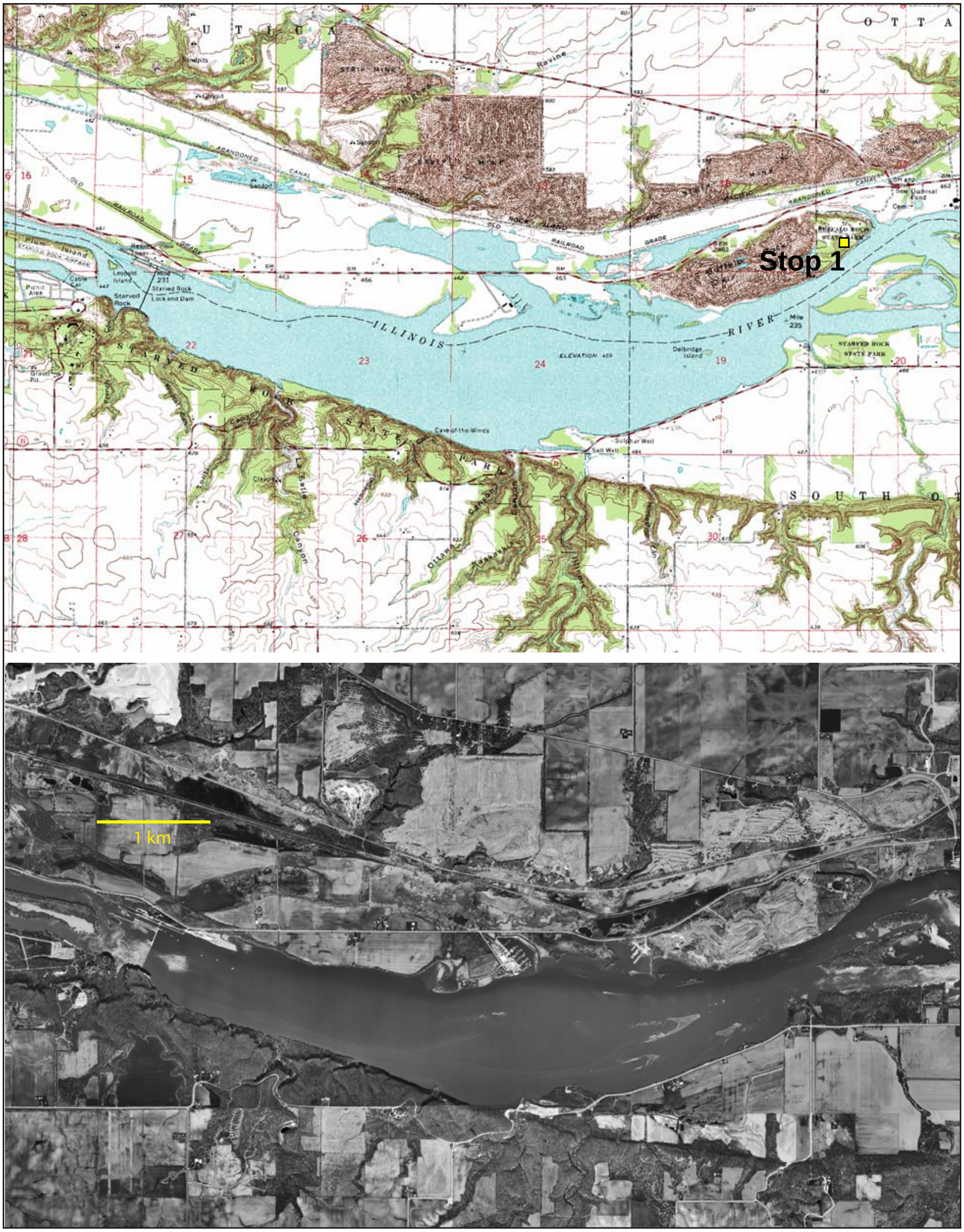


Figure 1 Location of Stop 1 at Buffalo Rock State Park on the Starved Rock, Illinois, 7.5-minute Quadrangle ner channel began. Previously, these features were interpreted as deltas deposited periodically in “Lake Illinois” as the Lake Michigan Lobe advanced and retreated (Fisher, 1925; Willman and Payne, 1942). Because these deposits are the only cited evidence for a “Lake Illinois”, the existence of such a lake is unlikely. Down valley from Hennepin, there is no obvious outer erosional zone, but there was a pre-existing valley (Ancient Mississippi Valley). However, there are a few saddles along the margin of the uplands between 158.5 and 170.7 m (520 and 560 ft) that may have been scoured at the time of flooding

During flood flow, the inner channel forms with widening and deepening of a preferred longitudinal groove. Once initiated, cutting of the inner channel is thought to progress rapidly because of the inefficiency of flows with large wetted perimeters and potential effects of armoring of the eroding surface (Shepherd and Schumm, 1974; Kehew, 1982). The upper Illinois Valley exhibits classic morphology of spillways formed by catastrophic discharge (Kehew and Lord, 1987). It is a deep trench (typically 53 m [175 ft] deep west of the Morris basin) with vertical valley walls that maintain a uniform width of about 2 km (1.2 mi) along the length of the upper valley (Figure 3b). Morphology is independent of lithology as carbonates, sandstone and unlithified glacial drift were excavated with little, if any, effect on channel course. Citing the straightness of the valley walls, Barrows (1910) and Cady (1919) suggested large floods from the Chicago Outlet, a feature we now know to post-date the Kankakee Torrent, cut the valley walls.

Typical of spillway channels, sedimentation is minimal because of the erosive power of the initial, relatively sediment fee, glacial lake discharge. There are no large-scale bar forms preserved within the upper Illinois bedrock trench, but poorly sorted gravel, including house-sized blocks of till, underlies a terrace just upstream of the spillway downstream of Morris (Figure 3c). Six pendant and alcove bars, most with marginal channels, are preserved in the upper-middle Illinois Valley between Hennepin and Peoria (Figure 3d). Immediately preceding flooding, the Illinois Valley downstream of Hennepin was at a relatively high level (Hajic, 1990), filled with early Late Wisconsin outwash over the distinctly pinkish Sankoty Sand Member. Stratigraphy and sedimentology underlying one of the bars suggests not all of this pre-flood outwash was scoured from the valley south of Hennepin (Kemmis et al., 2005).

Several spillways through moraines, in addition to the major breaches in moraines through which the Illinois and Vermilion Rivers flow, formed in response to flooding and cutting of the inner flood channel (upper valley trench). Narrow spillways at altitudes similar to the maximum altitude of upland scour cut Farm Ridge Moraine on either side of the inner flood channel (Figure 3e). A spillway through the Minonk Moraine directed proglacial lake waters to the upper-middle Illinois valley via Sandy Creek (Hajic, 1990). Two spillways through the Marseilles Moraine, namedthe Oswego and Newark channels at Stop 4, channeled floodwater down the Fox River Valley. A small spillway through the Farm Ridge Moraine connected the Streator-Pontiac basin to the Ottawa basin. These smaller spillways cover a range of altitudes and some of the higher ones may be related to drainage of smaller ice-marginal lakes.

With the exception of Buffalo Rock, erosional residuals in the upper Illinois trench are less pronounced on the bedrock valley floor, in part simply having less relief, and in part being somewhat obscured by later sedimentation. For example, streamlined residuals of low relief are located slightly east-southeast and west of Buffalo Rock (Figure 3f). Starved Rock, the famous Illinois landmark on the south side of the upper Illinois Valley, just south of Buffalo Rock, appears to be an erosional residual abandoned before it became completely detached from the uplands (Figure 3f). The town of Joliet is situated on an eroded diamicton remnant that could be considered an erosion residual. Two similar forms lie directly north of Joliet. Over a dozen well-

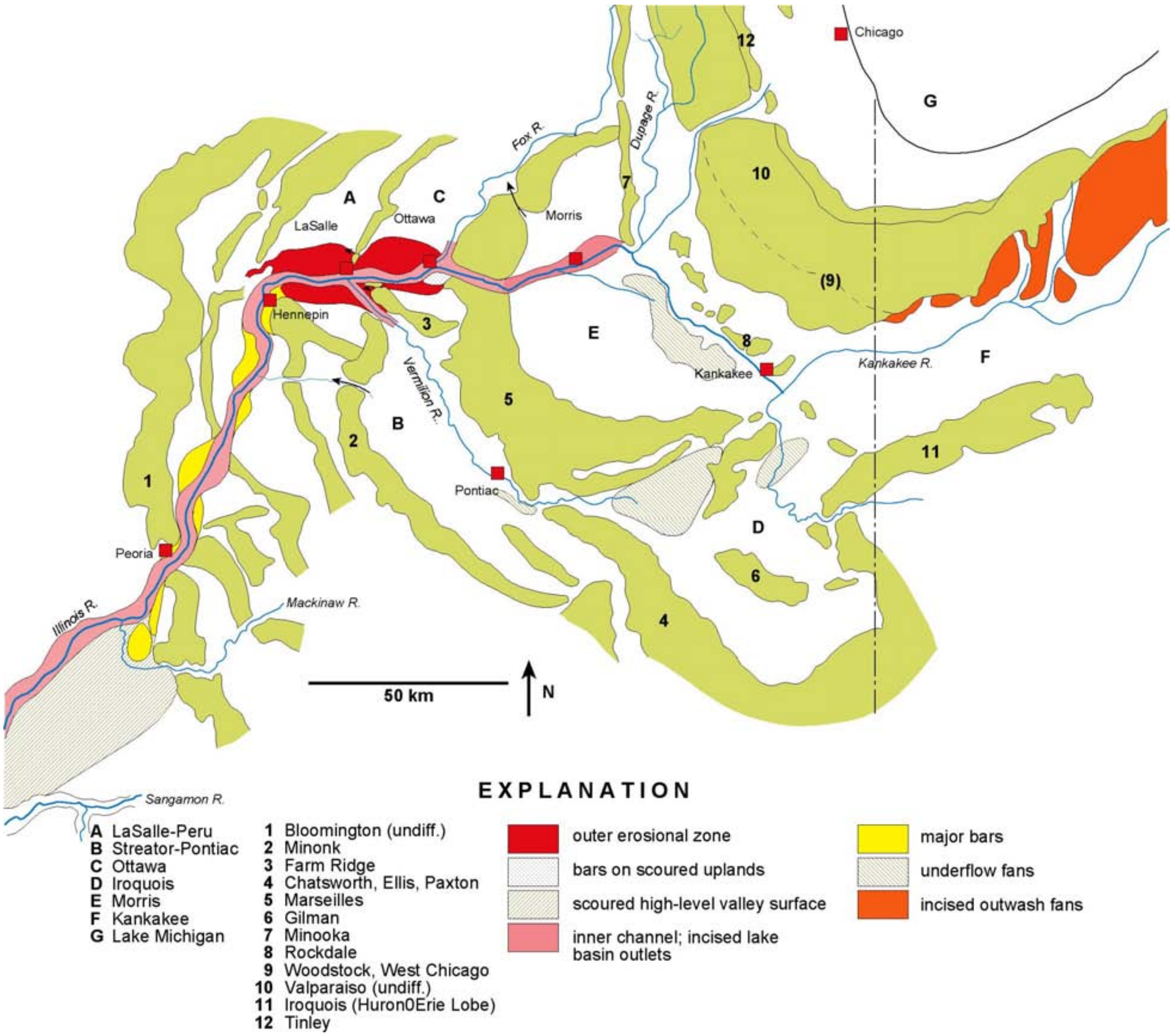


Figure 2 Macro-geomorphic features related to catastrophic flooding of ‘Kankakee Torrent’ floods, upper and upper-middle Illinois River Valley.

expressed erosional residuals occur in the lower-middle Illinois Valley, in the inner flood channel, between Peoria and Beardstown.

Sand and gravel bodies in glacial lakes at the mouths of former spillways that lack delta foresets or a delta foreset scarp are interpreted as underflow fans deposited when lakes were receiving catastrophic discharge from basins upstream (cf. Fenton et al., 1983; Kehew and Clayton, 1983). Sand bodies that probably represent underflow fans occur in the Streator-Pontiac, Iroquois, and Morris basins just downstream of lake inlets (Figures 2 and 3g).

Outburst floods from glacial lake basins tend to entrench outlet areas below glacial lake plains (cf. Kehew and Clayton, 1983). East of the Marseilles Moraine and the Morris basin outlet, the

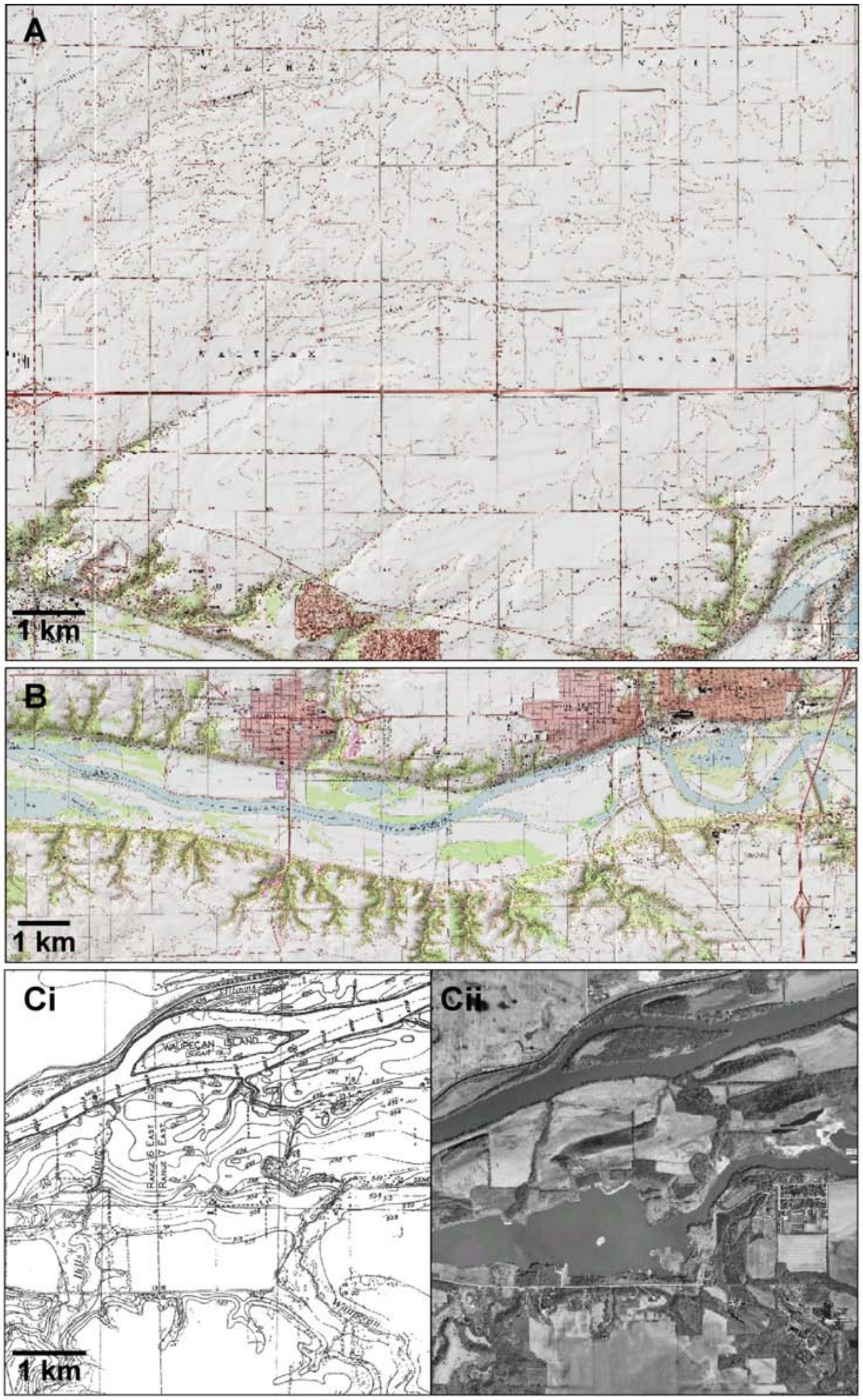


Figure 3 (A) Example of outer erosional zone between Farm Ridge Moraine and Utica to the west, and the Marseilles Moraine and Ottawa to the east. (B) Inner flood channel (upper Illinois River bedrock valley) between Spring Valley to the west and La Salle to the east. (C) 1944 topographic map (Ci) of a high terrace immediately upstream of the Illinois Valley gap in the Marseilles Moraine and downstream of Morris, and modern aerial imagery (Cii) showing mined terrace. “Islands” and protrusion into the mined area are enormous blocks of till.

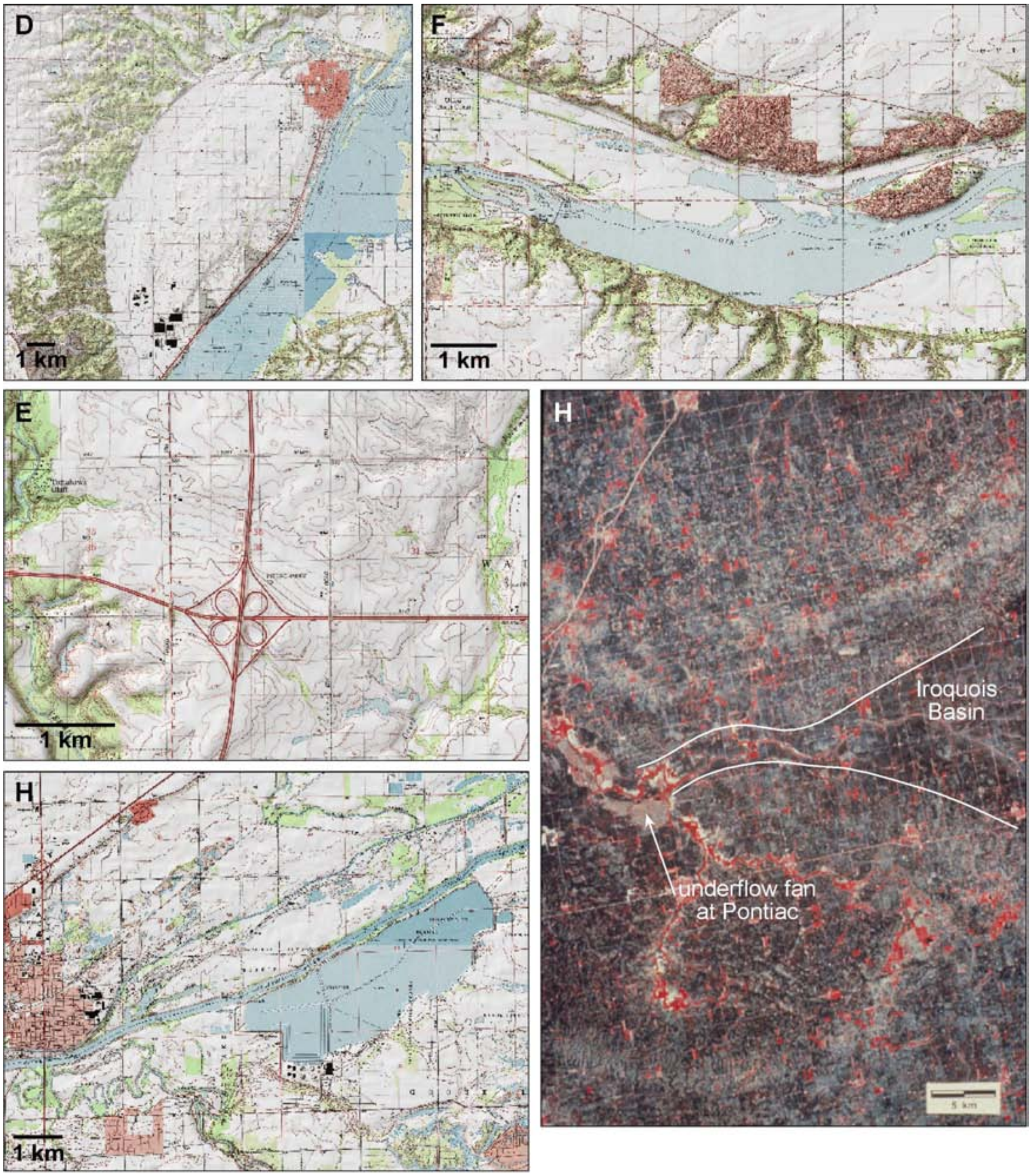


Figure 3 continued (D) Alcove bar with marginal channel at Chillicothe in the upper-middle Illinois Valley. (E) Spillway through the Farm Ridge Moraine north of the Illinois Valley trench. (F) Buffalo Rock, a strong example of an erosional residual, a more subtle erosional residual to the west of Buffalo Rock with a sand cover, and Starved Rock, a failed erosional residual (G) Landsat imagery illustrating lighter tone of underflow fan at the head of the Streator – Pontiac Basin. (H) Incised lake outlet in the Morris Basin behind the Marseilles Moraine.

Illinois Valley is entrenched only 12.2 to 6.1 m (40 to 20 ft) deep where it represents the incised outlet of a lake that drained catastrophically from the Morris Basin (Figure 3h). It is assumed this was the principal spillway and watercourse for the Kankakee Torrent. The lowest reach of the Vermilion River Valley is the outlet of the Streator-Pontiac basin. It is entrenched somewhat, but not to the extent of the upper Illinois Valley.

Stop 2a: Wedron Silica Company Quarry

Stratigraphic succession of glacial deposits at Wedron, IL

Jason Thomason

Background of Studies at Wedron

Not only an important source for silica sand (Keith and Kemmis, 2005), the Wedron Quarry has been an important resource for exposures of Quaternary units in Illinois since the turn of the century. The first mention of this site in ISGS archives is 1907, when geologist G.H. Cady observed that the quarry "produced a capacity of 7 car loads (of silica sand) per day" at a price of about $1 per ton. The Wedron Quarry was also important for early publications of the geography and geology of the Illinois River Valley (Sauer, 1916) and geology and economic resources of the St. Peter Sandstone (Lamar, 1927).

Extensive study of the glacial deposits at Wedron began in the 1930s when distinguished ISGS geologist H.B. Willman examined the drift exposures in the quarry to help reconstruct the history of glaciation in Illinois. For the next three decades, with his colleagues M.M. Leighton, J.C. Frye, and J.N. Payne, Willman related layers of glacial sediment at Wedron with similar deposits

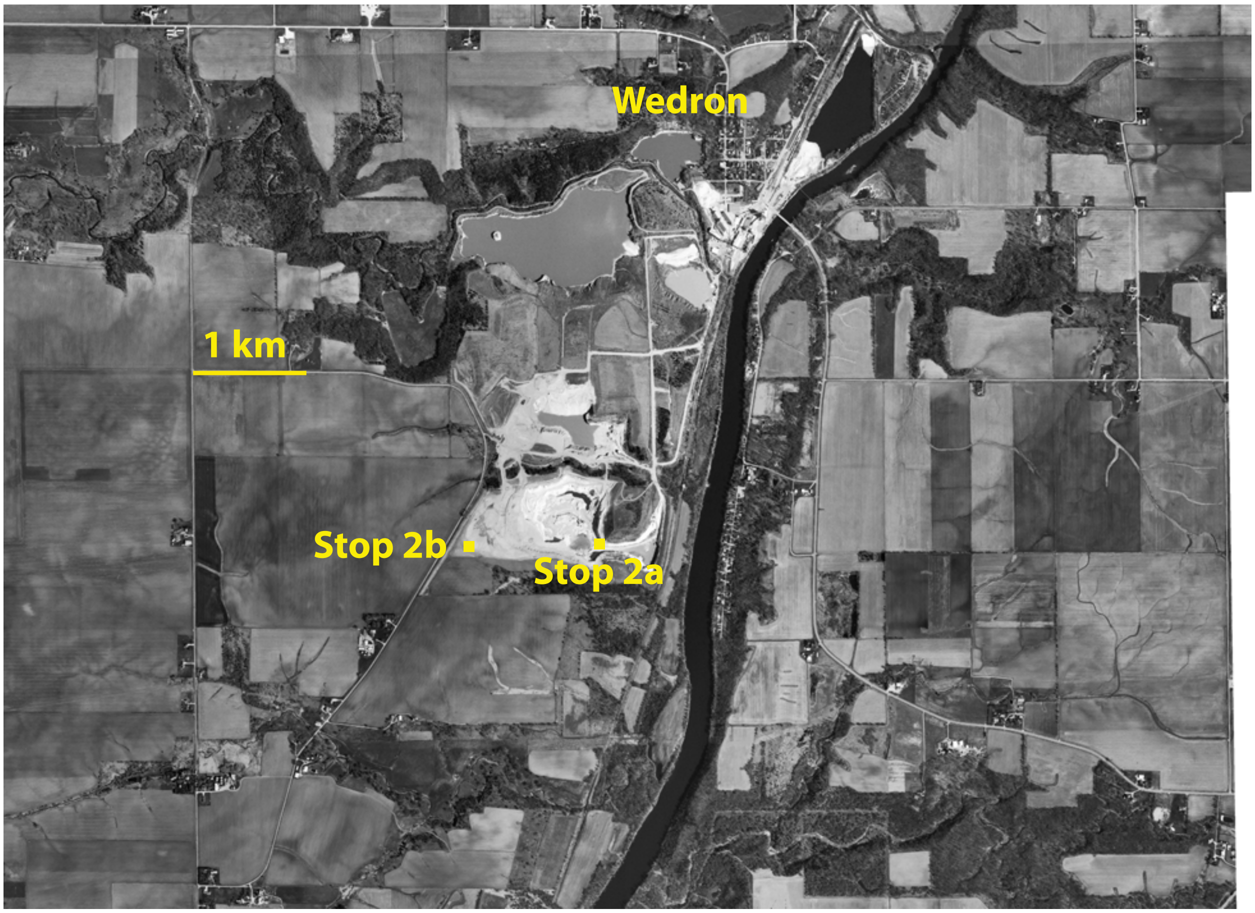


Figure 1 Location of Stops 2a and 2b at the Wedron Silica quarry on the digital orthophotoquad map of the Wedron, Illinois, 7.5-minute map.

found in end moraines to the west (Willman and Payne, 1942; Leighton and Willman, 1953; Willman and Frye, 1970). University of Illinois Professor W.H. Johnson regularly visited Wedron Quarry with his glacial geology students in the 1970s through the 1990s. In the 1980s, he and ISGS geologist A.K. Hansel began stratigraphic studies at Wedron Quarry, which serves as the type section of the deposits of the last glaciation. In the 1990s, Johnson and Hansel undertook research at Wedron Quarry on glacial sedimentation processes and ice sheet dynamics of the Lake Michigan lobe (Johnson and Hansel, 1990; 1999; Hansel and Johnson, 1996). Their studies, sponsored in part by a National Science Foundation (NSF) grant, initiated further NSF-sponsored research at Wedron by professors and students from the University of Maine, the University of Cincinnati, Oregon State University, and Iowa State University.

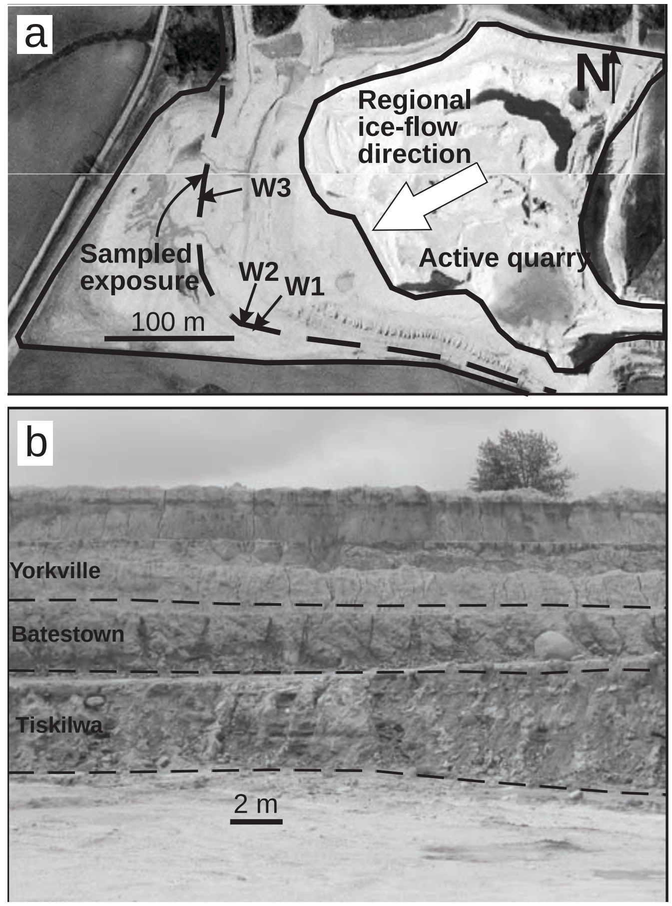


Figure 2 (a) Oblique aerial view of the Wedron silica quarry and profiles W1-W3. (b) Wedron exposure of the Batetown member and members deposited by earlier (Tiskilwa) and later (Yorktown) advances.

Quaternary units

Stratigraphic sections are currently exposed and accessible along the south and west walls of Pit 7 (Fig. 1b, excerpt from Thomason and Iverson, in review). The general succession of deposits is similar to those described in other pits throughout the quarry (Johnson and Hansel, 1990; Johnson et al., 1985), but local variability exists. Pre-Wisconsin deposits are not preserved at this location. However, most complete exposures reveal diamicton units of the first three Wisconsin Episode glacial advances into north-central Illinois: named Tiskilwa, Batestown, and Yorkville, respectively. Multiple glacigenic facies are associated with each Wisconsin advance. Lacustrine and fluvial facies of fine sand, silt and clay often underlie till facies. Beneath the lowermost till bed (Tiskilwa), boudinage structures are commonly preserved within fine silty sand of the Peddicord Formation, which is also often enfolded into the base of the overlying till. Till beds are distinguished by differences in lithology (grain size, color, mineralogy) and are commonly composed of multiple subfacies (e.g. Batestown). The Batestown till is underlain by massive clay lacustrine deposits which likely served as local sediment supply for a fine-grained facies of the till. Furthermore, subglacial channel deposits are sometimes found within and between till beds, which typically have erosional contacts at their bases. Fluvial facies of stratified sand and gravel are in places found beneath the uppermost till facies (Yorkville) and are marked by a remarkably sharp upper boundary with the till. A think blanket of loess composes the uppermost portion of the section.

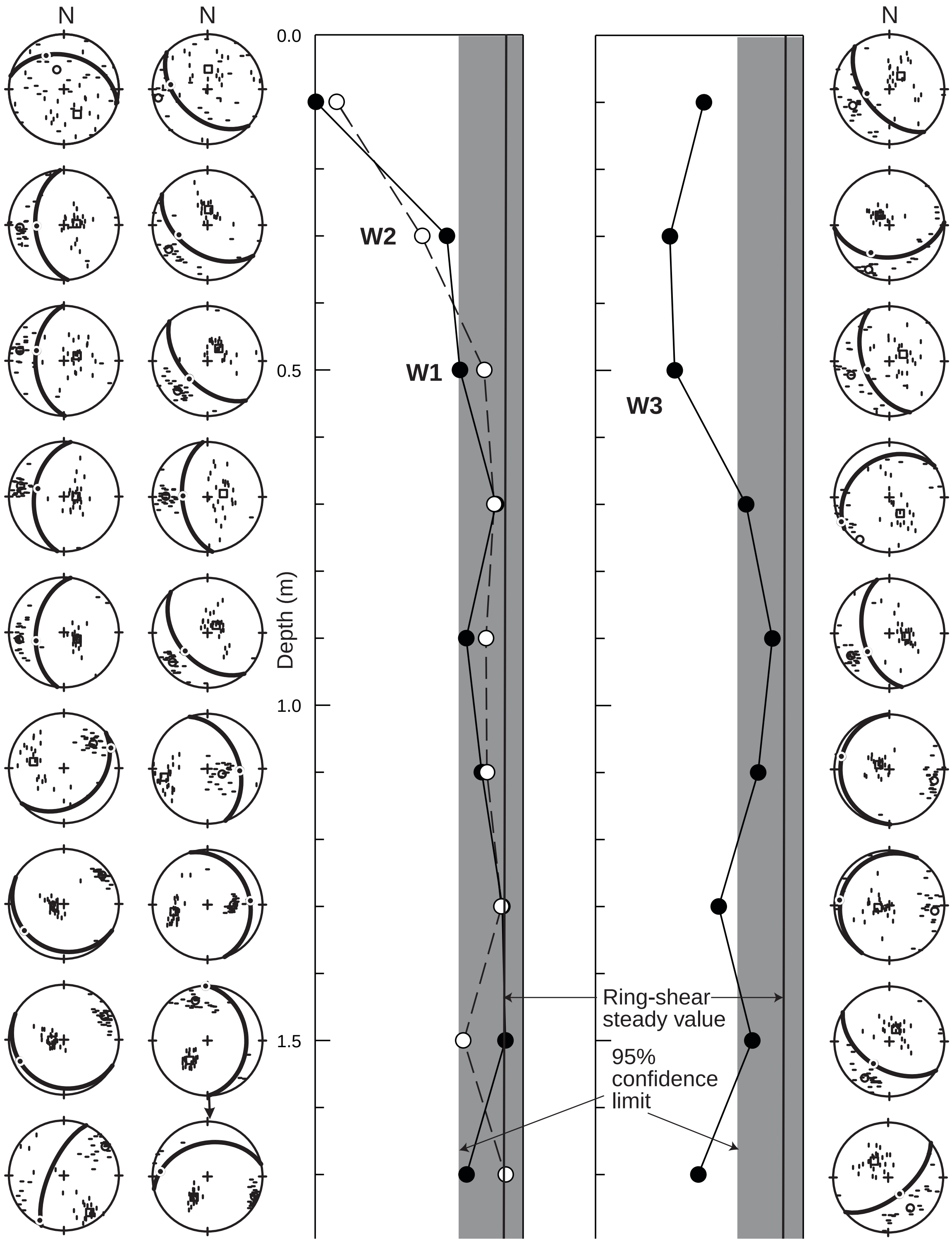


Figure 3 AMS fabric strengths and directions as a function of depth at W1-3. Horizontal and vertical marks in lower-hemisphere stereoplots are k_1 and k_3 orientations, respectively. Their respective eigenvectors (V_1) are shown with open circles and squares. Great circles indicate planes of shear, as inferred by assuming only simple shear and directly applying relationships from ring-shear experiments (see text). Solid dots within those planes indicate the direction of shear. W1 and W2 were about 2 m apart; W3 was ~ 100 m away.

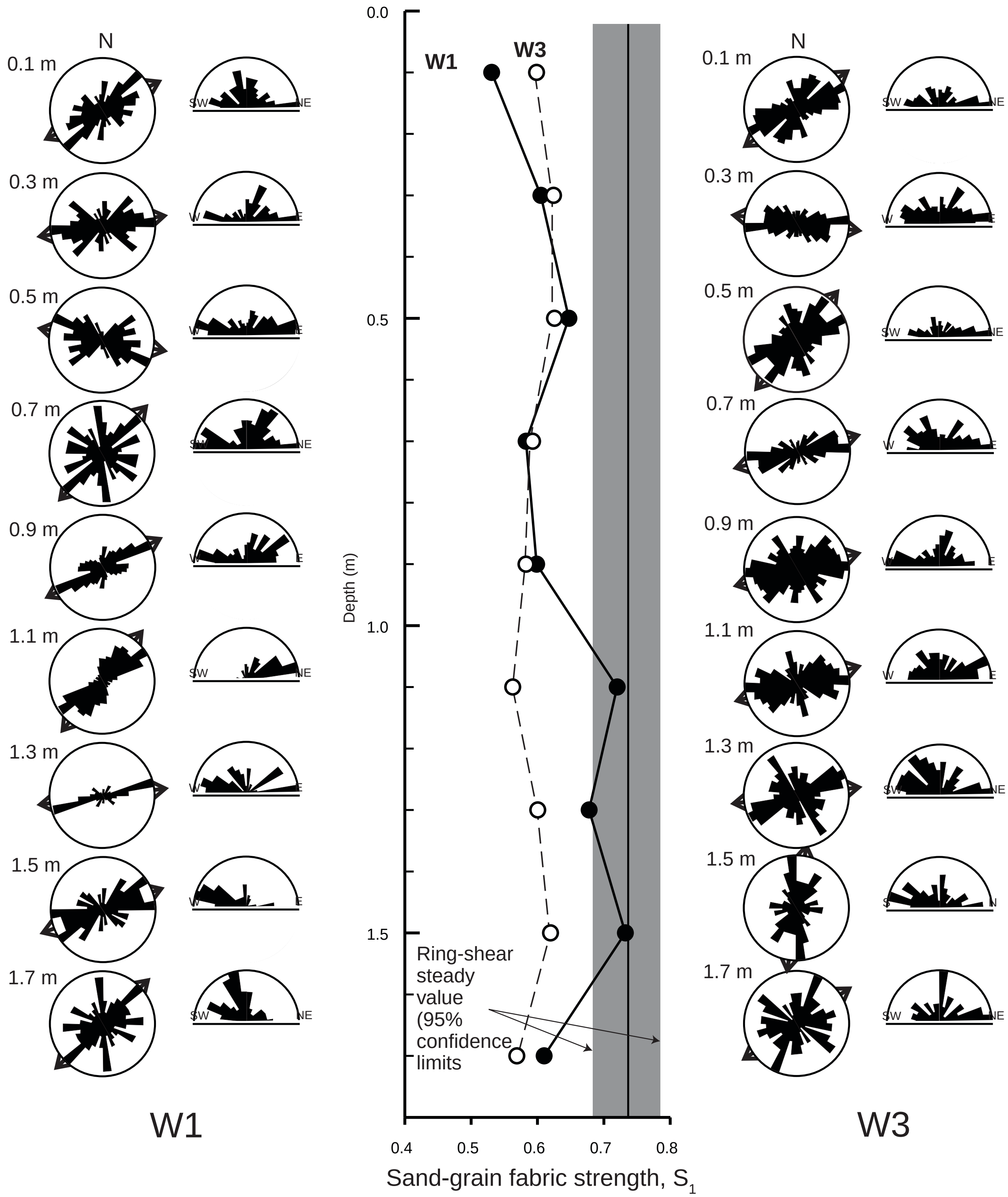


Figure 4 Sand-grain long-axis fabric strength and direction as a function of depth at W1 and W3. Full-circle diagrams display particle measurements in a horizontal plane. Triangular tick marks at circle perimeter indicate orientations of V_1 (azimuths). Half-circle diagrams display particle measurements in a vertical plane parallel to V_1. Two-dimensional S_1 values are based on orientations of particles in vertical thin sections, to allow direct comparison with measurements from ring-shear experiments.

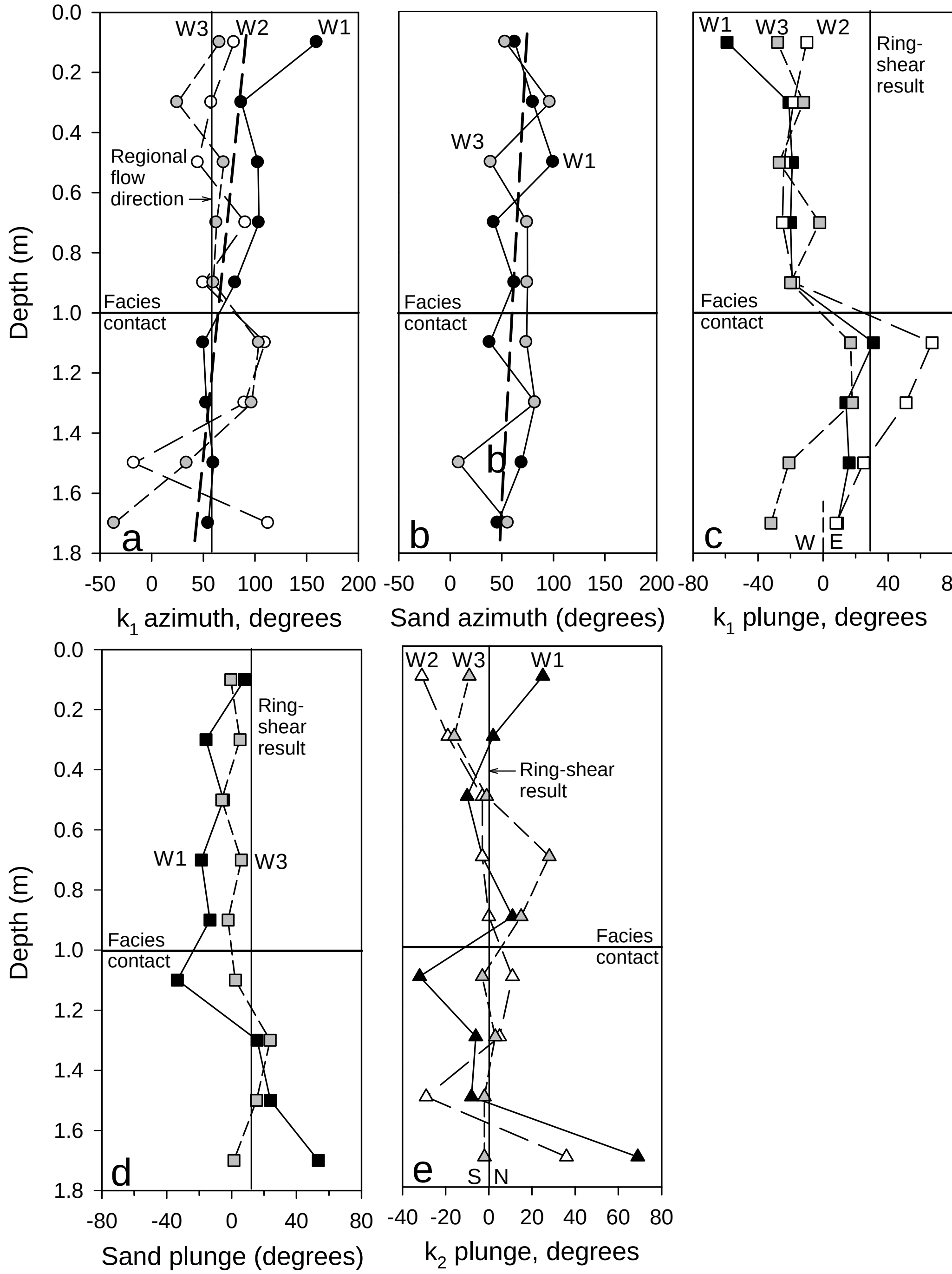


Figure 5. (a) Azimuths (V_1 orientations) of k_1 fabrics at W1-3. Dashed, bold line is a regression of the data. The regional flow direction was assumed to be the clast-fabric direction measured by Hansel and Johnson (1996). (b) Azimuths of sand-grain fabrics at W1 and W3. (c) Plunges of k_1 fabrics at W1-3 and ring-shear result. (d) Plunges of sand grain fabrics at W1 and W3 with ring-shear result (e) Plunges of k_2 at W1-3 compared with the result from ring-shear tests.

Discussion

Based on the remarkable preservation of stratigraphic relationships at other Wedron pits, the facies changes throughout succession were interpreted to be a result of major changes in sedimentary environment from subglacial to subaerial, and vice versa. Till facies at Wedron were thought to be largely deposited by lodgement (deposition directly from glacial ice) with some local subglacial transport and deposition (Johnson and Hansel, 1990). However, studies later suggested that deformation of sediments in the subglacial environment may have contributed largely to ice flow mechanisms, sediment transport, and geomorphic processes of the Lake Michigan Lobe (Boulton, 1996a,b; Jenson et al, 1996). Thus, the sequences at Wedron were later studied and interpretations included the possibility of sediment transport and geomorphic contributions by subglacial deformation (Johnson and Hansel, 1999). Figures 3, 4, and 5 includes excerpts of a paper aimed at investigating the degree of deformation within the Batestown till (Thomason and Iverson, in review, Journal of Glaciology). Analyses with various till fabric-measurement techniques suggest that subglacial deformation was likely, but it was shallow and spatially heterogeneous.

Things to see at this stop:

1) Completeness of glacial sequences associated with ice advances.

2) Nature of contacts between lithologic units.

3) Homogeneity within a given till facies.

4) Differences in character between till units (color, texture, etc.)

5) Boudinage/deformation structures within stratified sand below the Tiskilwa till

Stop 2b: Wedron Silica Company Quarry

Sedimentological and paleontological evidence for evolving lacustrine environments prior to the last glacial maximum, Wedron Southeast Section

Nicole Fox, Jason Thomason, Brandon Curry, David Grimley

Introduction

Our objective at this stop is to examine a 9 m-thick succession of fossiliferous lacustrine and barren proglacial lacustrine sediment at Pit 7 of the Wedron Silica Company quarry in La Salle County, Illinois (41.41ºN, -88.79ºW). Dating from about 31,000 to 30,000 cal yr BP (26,000 to 25,000 C-14 yr BP), modern analogs of ostracodes, gastropods, and spruce needles and cones hint that the fossiliferous part of the succession was deposited in a shallow but deepening lake under conditions that were somewhat drier and much cooler than today. Our results contrast with another study at Wedron Quarry. Garry et al. (1999) investigated the fossil beetles and plant macrofossils from a pocket of colluvium resting on bedrock (St. Peter Sandstone) and buried by diamicton of the last glaciation (the Tiskilwa Formation). Dating from about 25,120 cal yr BP (21,370 ± 240 C-14 yr BP; ISGS-2484; Hansel and Johnson, 1992), the assemblage was interpreted to be tundra or tundra-like, containing some boreal elements. Regional radiocarbon evidence indicates that this age probably just predates arrival of the Princeton sublobe (Lake Michigan lobe), the last glacier to advance across this region. Hence, our study, and the aforementioned study of Garry et al. (1999) provide snapshots of paleoenvironmental conditions prior to the arrival of the last glaciers with about 4,400 years of record missing between them.

The Wedron quarry is well-known for studies of glacial stratigraphy and sedimentology (Willman and Frey, 1970; Johnson and Hansel, 1990; Stop 2a), but notable paleontological studies have been done as well. Gastropods from correlative lacustrine sediment at the Wedron Quarry and other sites were the focus of investigation by Leonard and Frye (1960) and Willman et al. (1971). Radiocarbon ages obtained in these early studies ranged from about 28,000 to 25,000 C-14 yr BP. Additional radiocarbon ages have been obtained at Wedron from stratigraphic horizons pertinent to our investigation including an age of 28,000 cal yr BP (24,900 ± 200 C-14 yr BP; ISGS-862) of wood in leached organic-rich silt (Robein Member, Roxana Silt), and 28,500 cal yr BP (24,370 ± 310 C-14 yr BP; ISGS-863) of wood in overlying calcareous lacustrine sediment (Liu et al., 1986).

Our study differs from these earlier studies by including analysis of the ostracode fauna, particle-size distribution, and clay mineralogy of the sediment succession. Sedimentological and mineralogical data were collected to verify stratigraphic correlations and aid in the interpretation of depositional environments. Paleontological data was collected to provide a basis for our paleoenvironmental and paleoclimatic interpretations.

Methods

The section at Pit 7, freshly excavated and benched by normal quarry activity, was described in detail. Samples were collected every 0.5 m for analyses of particle-size distribution (PSD), semi-quantitative phase analyses (i.e., clay mineralogy), and fossil assemblages. PSD analyses were performed in a lab at the Illinois State Geological Survey using standard sieve and hydrometer techniques. Data are reported as percentage of gravel in the whole sample, and percentage sand (2.00 to 0.062 mm), silt (0.062 to 0.004 mm), and clay (less than 0.004 mm) in the matrix (all material less than 2.00 mm). Clay minerals were analyzed using X-ray diffraction of ethylene glycol-solvated aggregate slides of the < 2−µm size fraction. Peak heights were measured from diffractograms using a Scintag Model XPH-103 diffractometer. We used the peak intensity correction factors shown in Hughes et al. (1994). About 100 grams of material were examined at each interval for gastropods. Samples were softened in tap water, and then wet sieved. Shells were identified using the keys of Leonard and Frye (1960) and the Leonard gastropod collection housed at the Illinois State Geological Survey. The numbers of shells were also tabulated. Ostracodes were extracted and identified from ca. 20 gram subsamples of the material used for gastropods. The material was pretreated in boiling water and wet sieved under a shower spray of water. Species were identified by using the web site NANODe (Forester et al., 2006) and the reference articles of Delorme (1970a,b,c; 1971). Preliminary paleoenvironmental reconstructions based on presence/absence ostracode data were done using the range method of Smith and Forester (1994) using the ca. 600 sites from the United States in NANODe. This method uses the sort function in computer spreadsheets to identify the best-fit sites in NANODe.

Description

The Wedron Southeast section overlies weakly cemented St. Peter Sandstone (Ordovician) which is hydraulically mined at the Wedron quarry (Keith and Kemmis, 2005). Starting at the base, the section begins with approximately 1 m of medium to coarse sand and fine to medium gravel with common conifer wood fragments. Much of the clastic component is sand reworked from the subjacent bedrock. The sorted sediment occurs in channels that are 0.5-3 m wide and <0.5 m deep. Coarse fragments are locally imbricated, and the sands, in places, cross-bedded. The sand and gravel unit has an abrupt and planar upper contact with organic-rich silt about 3

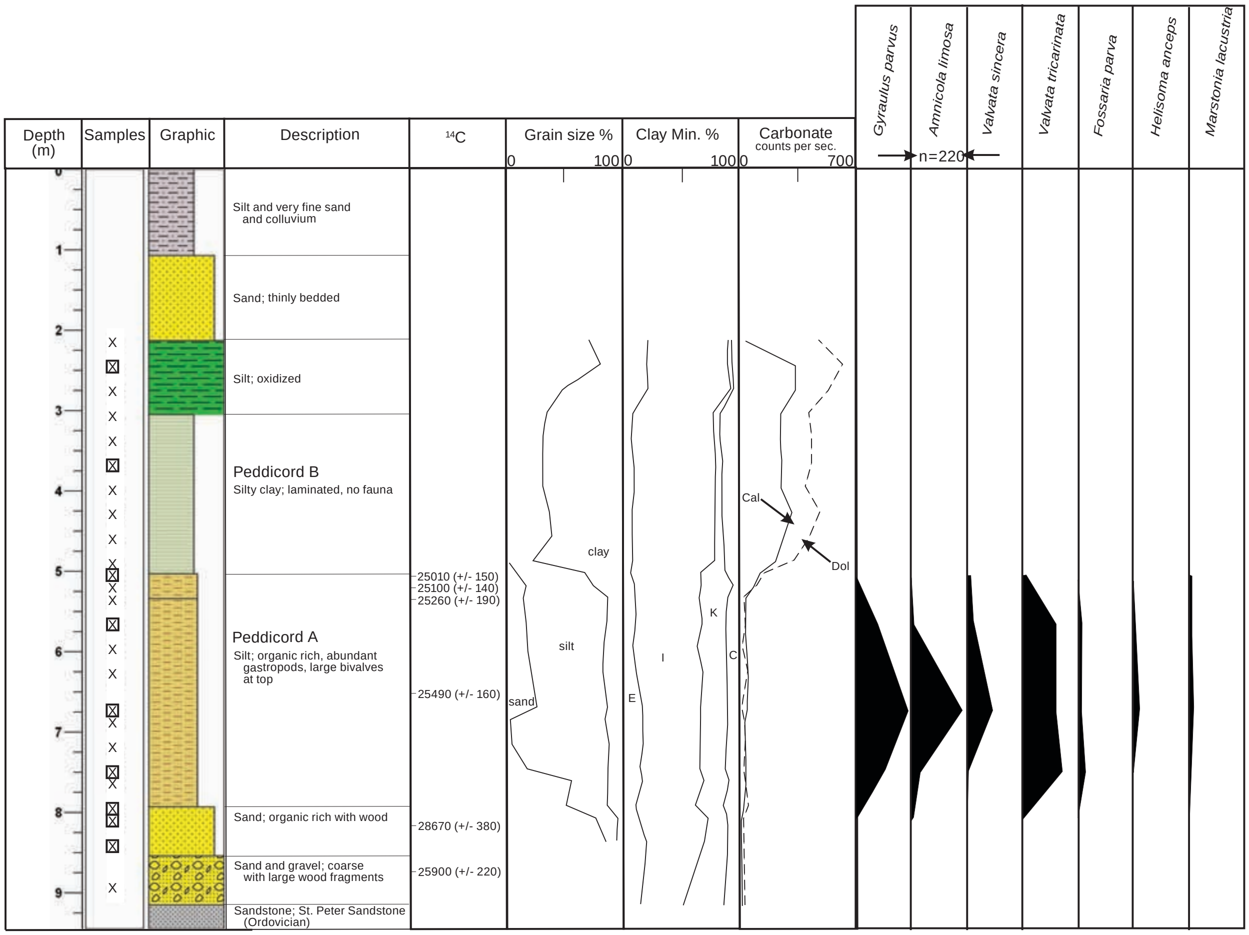


Figure 1 Stratigraphic column, radiocarbon ages, < 2 mm grain size distribution, < 2 mm clay mineral proportions, and gastopod assemblage of lacustrine sediments at the Wedron Southeast Section. Mineralogical data include cumulative expandables (E), illite (I), kaolinite (K), and chlorite (C). Gastropod data were collected from bulk samples (800-1000 grams), and a majority of V.sincera was juvenile.

m thick. The grayish-brown organic silt is weakly stratified and contains abundant aquatic gastropods, lacustrine ostracodes, plant macrofossils (terrestrial and aquatic), and small wood fragments. The upper 0.25 m of this unit contains fewer gastropods and ostracodes, is grayer, and contains abundant large bivalves (shells > 5 cm long; Fig. 1). Above the silt unit is approximately 2 m of laminated silty clay and silt. In places, the beds in the lower 1 m are rhythmically bedded with grayish-red clay 1-3 cm thick and laminae of pale yellow silt. This unit is barren of fossils, and is calcareous. The contact between the fossiliferous silt and barren silty clay is abrupt and planar, but in some places the lower few cm of the latter unit contain few small wood fragments, ostracodes, and needles. Above the laminated red clay are beds of oxidized silt, and fine sand 0.8 to 1.2m thick into which the modern soil is developed. In another short-lived section about 300 m to the northwest of our main section, the laminated clays are about 4 m thick and overlain by about 5 m of sand and gravel (Ashmore Tongue, Henry Formation) and an unknown thickness of reddish-brown diamicton (the Tiskilwa Formation). This study focuses on the brown silt and red clay units.

Sedimentology and stratigraphy

The channels and coarse texture of the basal unit indicate a fluvial origin, and is correlated with an unnamed tongue of the Henry Formation. The fossils and laminations in the organic silt and barren silty clay clearly point to a lacustrine origin. Because these sorted muds are buried by diamicton of the Tiskilwa Formation, they are correlated to the Peddicord Tongue of the Equality Formation (Hansel and Johnson, 1996). This unit was originally named the Peddicord Formation by Willman et al. (1971). The two subunits of the Peddicord are differentiated as Peddicord A (the fossiliferous silts) and Peddicord B (the barren silty clay). The weathered material is likely colluvium that we correlate to the Peyton Formation. The sand and gravel unit observed to the northwest is the Ashmore Tongue of the Henry Formation, and the diamicton, the Tiskilwa Formation (Stop 9b). In many places in the Wedron pit, the Tiskilwa or Peddicord units overlie leached, pedogenically modified, organic-rich silt known as the Robein Member (Roxana Silt; Johnson et al., 1999). This latter unit will be examined in detail at Stop 7.

Results

Clay mineral percentages and particle-size distributions in Peddicord A (Table 1) are similar to that of Robein Member (Roxana Silt) analyzed from previous studies at Wedron (Johnson et al., 1985; 1999). The absence or low quantity of calcite and dolomite in the clay fraction of Peddicord A is notable given the abundance of mollusk shells and ostracode valves.

Peddicord B has more than 50% more clay than Peddicord A, and contains more illite. The clay mineral assemblage is similar to that of the Peddicord clay as well as that of the overlying Tiskilwa Formation (Johnson et al.,1985; 1999), Radiocarbon ages of wood and needles from Peddicord A span less than 1,000 years, from about 30, 910 – 30,000 cal yr BP (25,900-25,010 C-14 yr BP; Radiocarbon Table, p 44). These ages are within the range of ages obtained from earlier studies in and around the Wedron area (Leonard and Frye, 1960; Willman et al., 1971), with the exception of the younger ages of subtundra fossils of about 25,500 cal yr BP (21,340 C-14 yr BP; Garry et al., 1990; Hansel and Johnson, 1992).

Fossil gastropod shells and pill clams (*Pisidium* sp.) are most abundant between depths of 8-5 m in Peddicord A (Fig. 1). Fifteen gastropod species were identified. The most abundant species include *Valvata tricarinata*, *Amnicola limosa*, and *Gyraulus parv*us, with fewer *Valvata sincera*, *Marstonia lustrica*, *Fossaria parva*, and *Helisoma anceps*. The large bivalve noted from the top of Peddicord A is *Pyganodon grandis* (Giant Floater). Ostracodes identified in Peddicord A include *Fabaeformiscandona rawsoni*, *F. caudata*, *Candona acuta*, *Cypridopsis vidua*, *Candona inopinata, Cyclocypris laevis, Candona distincta, Cyclocypris ampla*, and *Physocypria globula*.

Discussion

The collective evidence of radiocarbon ages and mineralogy indicates that unit Peddicord A was deposited in a lake adjacent to an upland in which a soil was forming in loessial parent material, the Robein Member of the Roxana Silt. The clay mineralogy of Peddicord A (our analyses) and of the Robein Member (earlier analyses) reveals little high-charge vermiculite and abundant illite, unlike several samples of the Robein Member from the LaFarge pit at Stop 7. We attribute the differences in mineralogy to provenance of the source outwash. In the case of the Robein from Stop 7, the silt was likely derived from Rock River outwash sourced from Wisconsin, whereas the Robein at this stop was derived from ancient Mississippi River outwash sourced from the upper Mississippi River basin. This scenario requires that there was little or no winds from the north which would have homogenized the clay mineral proportions from these two sites.

The paleoenvironmental data in the lake succession represents the final 1,000 year period of a ca. 4,000 year period during which the Farmdale Geosol developed in northeastern Illinois, roughly from 32,500 to 28,850 cal yr BP (28,000 to 24,000 C-14 yr BP; Curry, 1989; Curry et al., 1999). The switch to tundra conditions prior to approach of the Princeton Sublobe is not known, but occurred between the youngest known age of Peddicord A of about 29,150 cal yr BP (24,370 ± 310 C-14 yr BP; Liu et al., 1986) and the age of 25,540 cal yr BP (21,370 ± 240 C-14 yr BP) of Hansel and Johnson (1992).

The molluscan assemblage thrives today on the aquatic vegetation in permanent lakes in the Great Lakes region such as in Minnesota (Baker, 1935; Dawley, 1947), Wisconsin (Baker, 1928) and southern Ontario (Clarke, 1981). Indeed, we found abundant charophyte oogonia (charophytes are a green algae common to hard-water lakes and ponds), as well as the seeds and leaves of aquatic plants, including *Zanichellia palustris* and *Naja* sp.(Kantrud et al., 1989). Two gastropod species, *Martstonia lustrica* (boreal Marstonia) and *Valvata sincera* (boreal turret snail), though not common, are more typical of northern regions, with southern limits in Minnesota, Wisconsin, and Massachusetts (Baker, 1928; Clarke, 1981). *Valvata tricarinata, Helisoma anceps, Amnicola limosa* and *Gyraulus parvus* are cosmopolitan North American species.

The decrease in gastropod shells and occurrence of the large bivalve *Pyganodon grandis* in the upper part of unit Peddicord A indicates a change to high sediment accumulation rates and greater water depth (Kevin Cummings, INHS, personal communication). We attribute the paucity of fossils in Peddicord B to heightened rates of sediment accumulation, dilute water (which would have dissolved any biogenic carbonate), and lower productivity, in part due to shorter annual ice-free lake conditions and cooler temperatures as the glacier approached.

Based on the modern analog analysis using NANODe (Forester et al., 2006), the paleolacustine environment was similar to lakes that occur today in northwestern Iowa (Lake Okiboji, Storm Lake, Goose Lake) and southwestern Minnesota (Bloody Lake, Lake Mina). These lakes occur in an area where the average mean annual precipitation is roughly 700 mm/yr, about 200 mm/yr less than the Wedron area today. In terms of effective/net moisture, the analog lakes occur in a region with negative values between -100 to -200 mm/yr (because evaporation exceeds precipitation); the modern effective moisture/ net moisture of the Wedron area is positive 50 mm/yr (Winter and Woo, 1988). The average mean annual temperature of the modern analog lakes is about 6.5ºC, about 3ºC cooler than the Wedron region today. The modern analog lakes are hard-water lakes, meaning that more than 50% of their dissolved anions were bicarbonate ions (HCO3-). *Fabaeformiscandona rawsoni* and the aquatic plant *Zannichellia palustris* are common to lakes in the prairie pothole region of the North American prairies (Smith, 1993; Stewart and Kantrud, 1972).

The character of the environmental shift from boreal to tundra-like conditions is not captured in the section we analyzed. Instead, there is a hiatus of about 4,400 years, an unconformity marked by the contact between Peddicord A and B.

The collective lacustrine ostracode and gastropod fauna, coupled with the presence of spruce needles and wood, suggest that the Peddicord A unit was deposited in a lake with a water depth within the photic zone such that aquatic vegetation was abundant. The modern analog lakes indicated by the ostracodes occur near the prairie-forest tension zone. Plant macrofossils, particularly of needles, indicate that spruce trees grew around the margin of the lakes. The southernmost occurrence of spruce in Minnesota occurs at a latitude of about 43ºN well within the range of the northernmost analog lakes such as Lake Mina (45.89ºN, -95.45ºW). Ironically, the ostracode *F. rawsoni* and aquatic plant *Z. palustris* indicate that the upland vegetation was prairie-like. Pollen analyses may bear this out, but we note that the relative abundance of boreal tree pollen in lake records is strongly dependent upon the size or shoreline development of the lake. In areas with riparian boreal trees and upland prairie plants, smaller lakes, or lakes with many embayments and islands will have a higher tree/prairie pollen ratio compared to larger lakes with smooth shorelines (Grüger, 1972).

Peddicord lake deposits occur as high as ~163 m (535 feet) elevation in both the Wedron and Morris sections. This relationship suggested to Willman et al. (1971) the presence of a large Farmdale Phase slackwater lake in the Ticona Valley and its tributaries. The latter is a buried bedrock valley, only partially filled with pre-Wisconsin Episode deposits, tributary to the ancient Mississippi River valley (see Figure 8 of prologue, p 13). Aggradation of the ancient Mississippi River valley in response to glaciation of the upper Midwest would have served to impound the Ticona Valley. As Wisconsin Episode glaciers approached the site, sediment accumulation rates and water depth increased. The provenance of the lake sediment changed from eolian and pedogenic to fresh glacial debris under colder conditions that marked the eventual arrival of Michigan Subepisode glaciers.

Table1 Comparison of sedimentological data between the current study and previous studies at Wedron.

Study	Sediment	Mean %		
		E	I	K+C
Stop 10	Peddicord A	13.1	48.4	18.5
Johnson et al., 1985	Robein silt	20.2	59.4	20.4
Stop 10	Peddicord B	11.5	64.6	12.5
Johnson et al., 1985	Peddicord clay	7.3	73.1	19.6
Wickham et al., 1988	Tiskilwa diamicton	8.2	68.0	23.0
		sand	silt	clay
Stop 10	Peddicord A	28.7	55.0	16.3
Johnson et al., 1985	Robein silt	12.9	58.1	28.9
Stop 10	Peddicord B	0.3	41.6	58.1
Johnson et al., 1985	Peddicord clay	1.2	9.8	89.0

Questions for discussion:

1) What processes/factors may have contributed to systematic changes of biota abundance with depth? Is the missing 4,400 years of sedimentation between the two types of organic deposits at Wedron due to erosion or slow deposition rates? What is the nature of the sharp contact between the brown silt (Peddicord A) and red clay (Peddicord B) units?

2) What does the presence of *Pyganodon grandis*, giant floater, indicate about the paleoenvironment? The death position of *Pyganodon grandis* is such that the shells lie flat on the ancient lake bottom instead of perpendicular to that surface like most large bivalves. What does this suggest about the locomotion and living strategy of this species? In what kinds of lacustrine environments would this be advantageousfor this species?

Stop 3: Central Quarry

Exceptionally Well-preserved Paleokarst and Pennsylvanian Cavefills

Roy E. Plotnick, Fabien Kenig, Andrew Scott, Ian Glasspool

Introduction

At this stop we will view a recently discovered middle Paleozoic paleokarst and Lower Middle Pennsylvanian cave fill. This site contains a diverse plant assemblage in the poorly consolidated infill, much of which is superbly preserved as charcoal meso- and macrofossils and includes what is possibly the earliest North American conifer. The site is also important for our understanding of sub-Absaroka erosion and the earliest stages of Absaroka deposition on the northern edge of the Illinois Basin.

The paleokarst was discovered in 2004 during a UIC graduate paleontology class field trip. Investigations are still ongoing, and include detailed paleontological and biogeochemical studies. WE ASK THAT COLLECTING BE KEPT TO A MINIMUM AND THAT ANY NOTEWORTHY FINDS BE REPORTED TO US. ALSO NOTE THAT THE SLOPES ARE UNSTABLE AND THERE IS A STRONG POSSIBILITY OF ROCK FALLS. PLEASE USE CAUTION!!

Sloss (1964) recognized that the Phanerozoic stratigraphic record of the North American craton is marked by large bodies of sedimentary rock, now termed “Sloss sequences”, bounded above and below by extensive unconformities. For example, the Absaroka sequence is comprised of cratonal sediments deposited from the beginning of the Pennsylvanian through the Permian. These rocks are separated by a deep and widespread erosional surface from the underlying Devonian-Mississippian Kaskaskia sequence.

The erosional surface below the Absaroka sequence is one of the most extensive and best-known unconformities in mid-continent North America, with maximum relief on the order of 120 m (Collinson et al., 1988). This unconformity can be recognized worldwide and has been tied to a major eustatic sea-level drop during the late Mississippian, resulting from the initiation of widespread Southern Hemisphere glaciation in Gondwana (Veevers and Powell, 1987). Extensive karst is developed on this surface in the western United States (Palmer and Palmer, 1989) and is particularly well-known from the Black Hills. Palmer and Palmer termed this the “post-Kaskaskia” paleokarst. In general, rocks under the unconformity in the eastern United States are clastic, and thus paleokarst features are much less common (Palmer and Palmer, 1989).

In northwestern Illinois, paleokarst features are known from Devonian limestones of the Rock Island area (Leary, 1981; Leary and Trask, 1985) including solution widened joints (“cutters”), caves, sinkholes and solution channels. These features are filled and overlain by clastic sediments. Some localized channel fills contain an abundant flora at their base with plants that grew on well-drained soils above the water table (Leary, 1981). This flora may be as old as Late Mississippian. Other channel fills contain strata of oldest Pennsylvanian age (Morrowan = lower Bashkirian; Figure 1) belonging to the Caseyville Formation. These rocks also contain an abundant flora, with drier climate elements in the lower parts, followed by a more typical wetland (mire) flora.

Evidence for a karst surface of top on Silurian dolomitic bedrock in northeastern Illinois was reviewed by Panno et al. (1997). A number of Pennsylvanian sinkhole and fissure fill deposits have also been described from these dolomites (Bretz, 1940). The best known are in the Lehigh Stone Quarry about 17 km west of Kankakee. Bretz described sinkhole deposits at this quarry

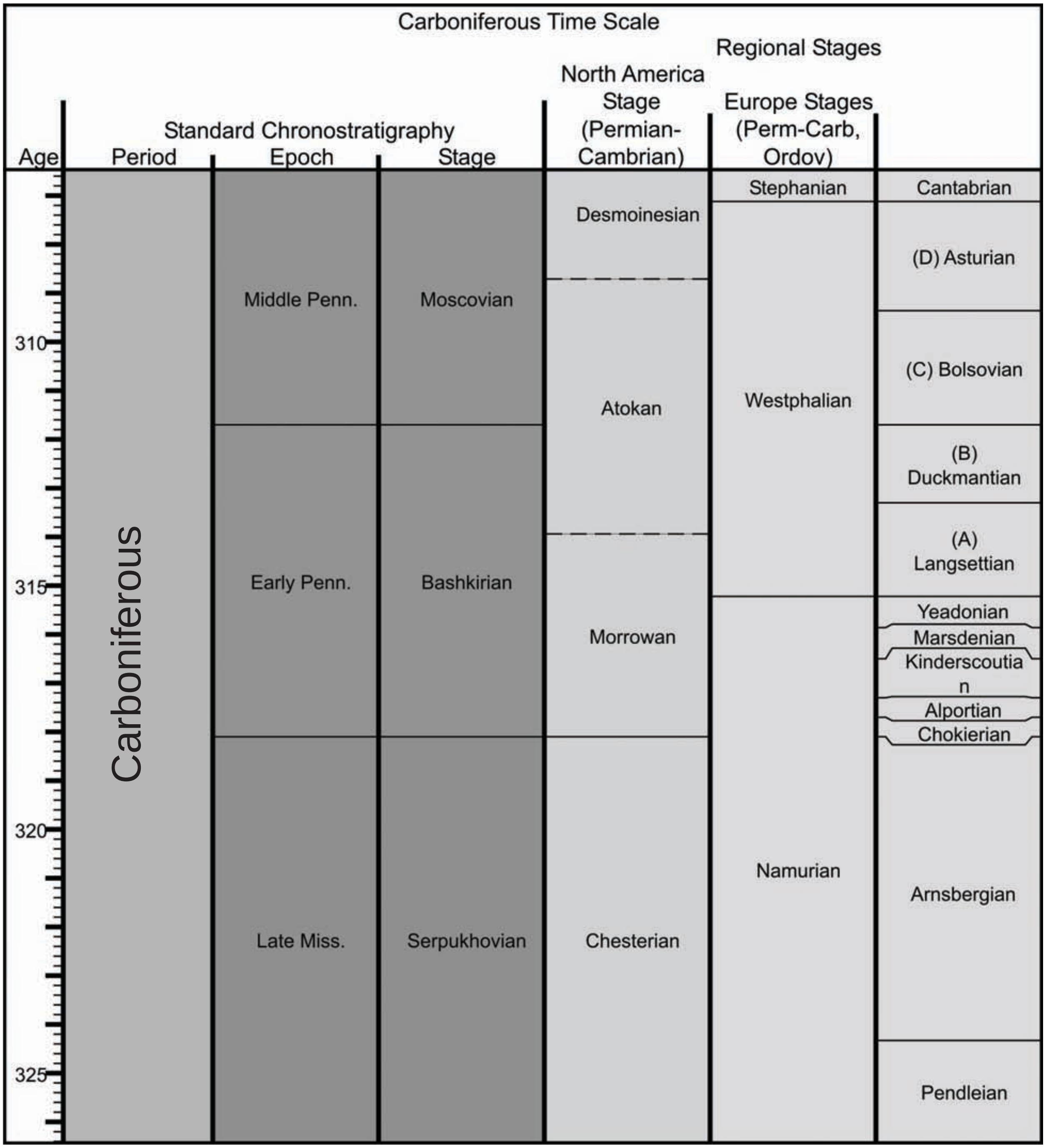


Figure 1 Carboniferous stratigraphic nomenclature. The inferred stratigraphic position of the Central Quarry site is indicated by the large "C" in the North American Stage column. Figure produced with TS-Creator (http://www.stratigraphy.org).

as being predominantly shale with pyrite and plant fragments. Plant remains include leaves, stems, spores, charcoal, and pyritized wood. We have recently revisited this quarry. Quarry operations over the last 70 years have not only exposed additional of sinkhole deposits, but also filled caves lower in the section.

The Site

The paleokarst and associated fill we will see are located in the Central Limestone quarry, Kendall County, Illinois (Sec. 28, T35N,R7E; N 41.48 W 88.44), about 80 km southwest of Chicago. The host rocks for the paleokarst are Late Ordovician limestones of the Dunleith Formation, Galena Group. A detailed section of the quarry was published by Willman and Kolata (1978) who described styolites and calcite-filled cavities near the top of the sequence. We have noted extensive sulfide mineralization (pyrite and/or marcasite) along fractures in the limestone. The limestone is somewhat fossiliferous, including brachiopods *(Rafinesquina, Protozyga)*, trilobites *(Illaenus)*, crinoids, bryozoans, snails *(Hormotoma)* and receptaculitid algae ("sunflower corals"). Trace fossils are also common in some layers.

Central Quarry is located about 5 km north of the mapped edge of the Pennsylvanian outcrop belt in northern Illinois (Figure 2). The nearest Pennsylvanian outcrops are small outliers of poorly constrained age near Morris. Test borings for a proposed solid waste disposal facility have revealed undated Pennsylvanian sandstones and shales 3 km to the east. Structurally, the locality lies on the northern edge of the Illinois Basin, about 65 km east of the crest of the LaSalle Anticlinorium and about 10 km southwest of the Sandwich Fault Zone, on the upthrown side of the fault (Kolata et al., 1978).

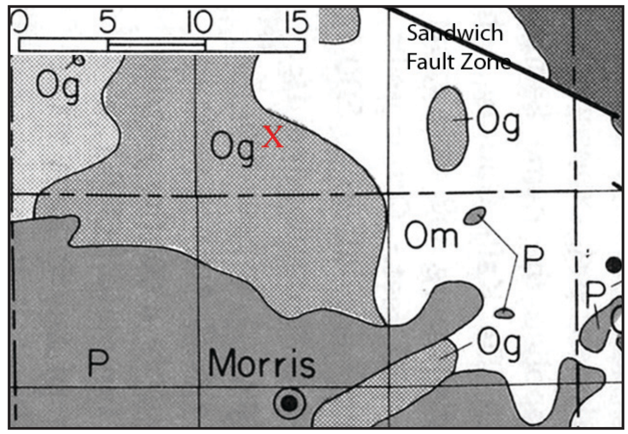


Figure 2 Local bedrock geology. X marks the location of the Central Quarry. Sandwich Fault Zone is to the north. Scale is in kilometers. Og = Ordovician Galena Group; Om = Ordovician Maquoketa Group; P = Pennsylvanian. Based on Kolata et al. (1978).

The Caves and Cavefills

The ancient caves and associated fills are located in the upper part of the section, along an approximately 300 m exposure along the southern wall of the quarry. The smallest features are about one meter in cross section. The best preserved filled cave is approximately 10 m high and 18 m across (Figure 3). Unpublished boring data for a new quarry suggest that the paleokarst extends at least 1 km to the south.

The ancient caves are generally semi-circular in cross-section, with an irregular flat upper surface, including what may be pendants produced by differential solution of the roof rock (Figure 4). These profiles are typical of cave passages formed at the water table (A. Palmer, pers. comm.) A number of the lateral margins are noticeably smoothed. There is no evidence of speleothems. Although there is some collapse breccia at the base of some of the caves, most of the fill is allochthonous. Overall, this suggests the caves were an open system at the time of filling.

The cave fills are variable in composition and are very poorly lithified, disaggregating readily when rewetted. Being roofed, they were subject only to minor compaction due to settling and dewatering. Acidic groundwater infiltration has likely removed any potential carbonate cements (A. Palmer, pers. comm.). Due to rapid erosion of exposed fill, the visual appearance of the various caves is constantly in flux; many of the smaller features are relatively ephemeral.

The fills show evidence of soft-sediment deformation. In the largest features, the beds dip and pinch out towards the margin of the cave. Some of the smaller features show concentric layers surrounding the center of the fill, suggesting that these were passages that filled from the outside in (Figure 4).

Figure 3 Largest and best preserved of the caves and cave fills at Central Quarry.

Figure 4 Smaller cave fill showing possible roof pendant (arrow) and concentric filling.

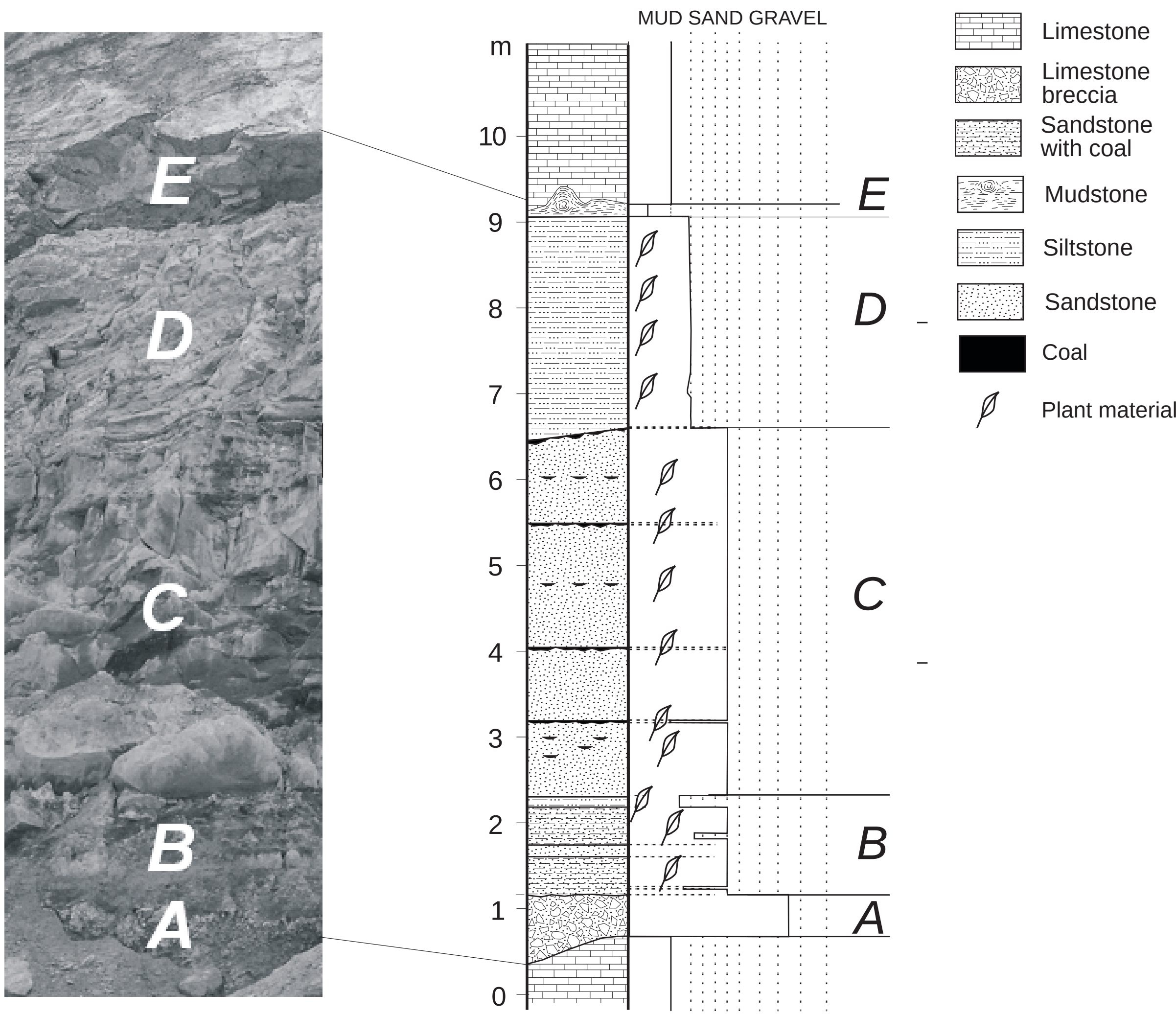


Figure 5 Stratigraphic section of the largest fill. We correlate unit C with the Babylon sandstone and unit E with the Cheltenham Clay. Both are in the Atokan age Tradewater Formation.

The largest and best preserved fill shows a distinct fining upward sequence, which can be subdivided into five facies (Figure 5). One or several of these units also occur in the smaller features. These units are, from the bottom up:

1) Roof collapse breccia; breccia is also embedded in other units elsewhere in the quarry;

2) Alternating layers of coarse sand and poorly-sorted coarse silt and fine sand with varying amounts of coaly shale. Small scale cross-bedding is observed at some places;

3) Coarse quartz sandstone, generally massive, with distinct euhedral overgrowths; occasional coaly lenses and thin coaly layers;

4) Poorly sorted, laminated grey siltstones, with appreciable amounts of sand and clay and abundant plant debris; small-scale cross beds are common.

5) Finely laminated grey clay, which often breaks conchoidally, with minor fine sand grains. The layers occasionally exhibit small normal faults (Figure 6). X-ray diffraction of the clay shows the presence of kaolinite and illite (Guggenheim, pers. comm.)

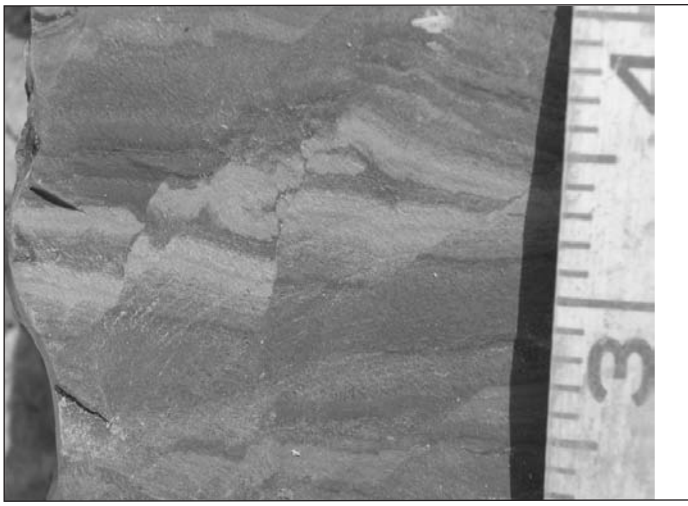


Figure 6 Small normal faults in laminated claystones.

Secondary sulfide minerals (pyrite or marcasite) are common throughout the fill. There is a possible angular unconformity between the laminated siltstone and the underlying sandstone (Figure 5). This suggests that there was a hiatus between the deposition of the two units, during which the lower units underwent some compaction.

The sedimentology and stratigraphy of the fill suggest several episodes of fluvial sedimentation. It has not been determined to what extent these may have caused additional erosion of the cave passages. The similarity to surface stratigraphy elsewhere in northern Illinois (see below) suggests a close connection between the karst system and the surface, probably through sinkholes or fissures that are not currently visible.

Paleontology

Fossil plants occur in all units of the fill, with the general exception of the grey clay. Megascopic plant remains consist of abundant small wood fragments, leaves, seeds, and rare small leafy twigs (Figure 7) . Most of these are preserved as charcoal and show remarkable anatomical and morphologic detail.

Many of the megascopic remains can be assigned to the extinct tree *Cordaites*, believed to closely related to conifers. There are also true conifers, which are probably related to walchian conifers, although detailed taxonomic assignment is difficult from isolated leaves and twigs (Figure 7).

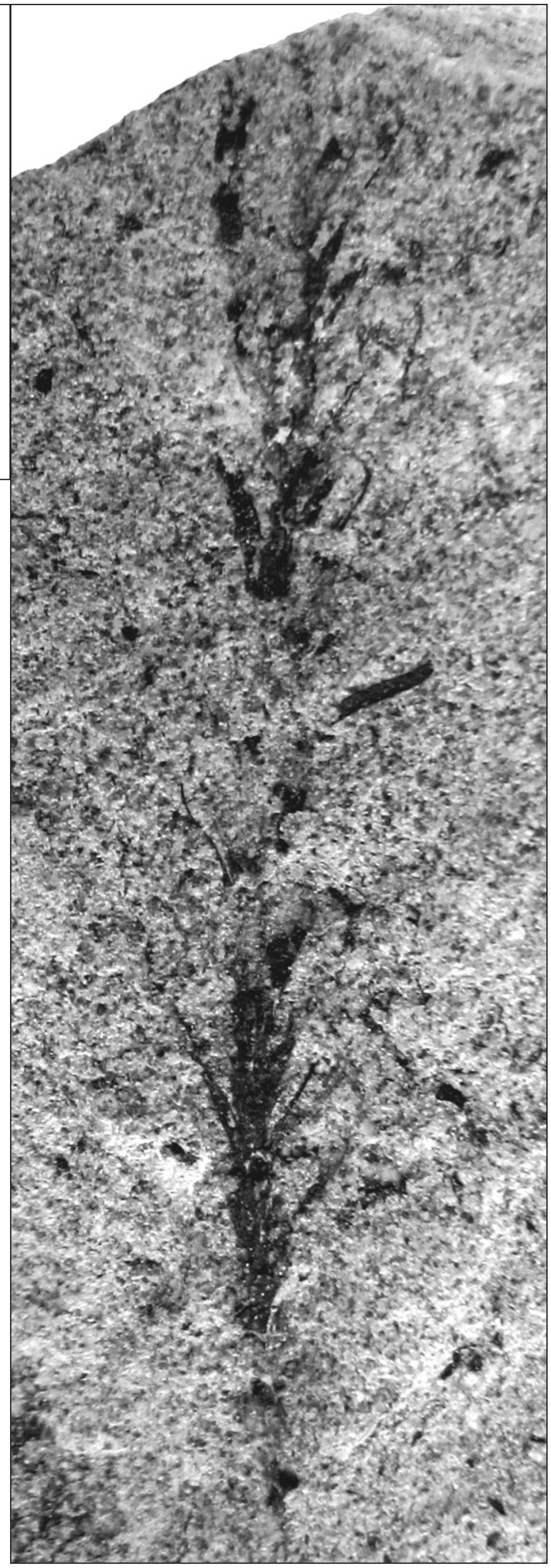


Figure 7 Small twig of walchian conifer. Length is about 2.5 cm.

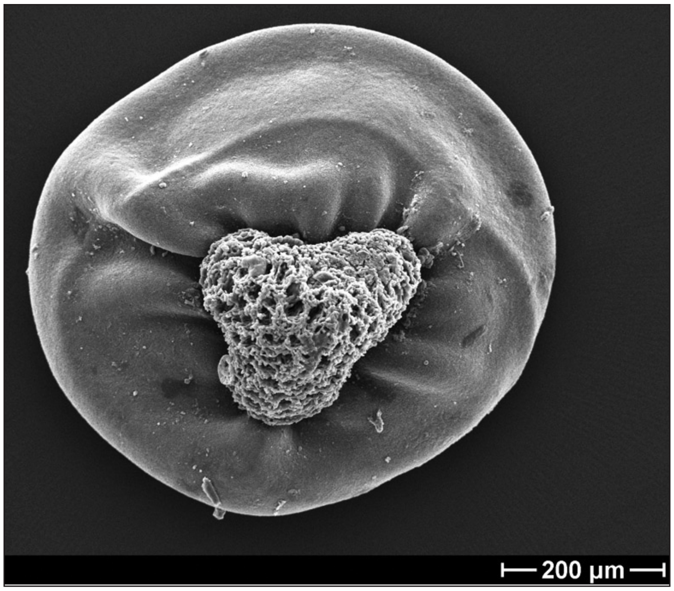


Figure 8 Micrograph of Cystoporites megaspore taken with a scanning electron microscope.

Microscopically, the conifer leaves possess stomata with overarching papillae; such features are usually considered adaptations to water stress (Hernandez-Castillo et al., 2001).

The diverse palynoflora includes lycopsid megaspores and miospores. Lycopsids are an ancient group of land plants that include the modern club mosses and quillworts. The megaspores include the taxa Cystosporites, Rotatisporites, Zonalesporites and *Pseudovalvisisporites* and show minimal compaction (Figure 8; Glasspool et al., in review).

The miospore assemblage indicates a parent vegetation dominated by herbaceous forms, particularly those belonging to the lycopsid groups Chaloneriaceae (Isoetales) and Selginellales. Miospores from arborescent lycopids are rarer, and include spores produced by the isoetalean Lepidocarpaceae and Sigillariaceae (Balme, 1995).

In addition to the lycopsid miospores, *Florinites* is also abundant. Pollen of this type has been reported *in situ* from *Cordaitanthu*s and further indicates the importance of the cordaite vegetation at this locality. Similarly, the occurrence of *Potonieisporites* type pollen further supports the leaf evidence for the presence of conifers.

Animal fossils are rare. We have fragmentary pieces of scorpion cuticle, but have found no bone or teeth.

Age of the Fill and Paleokarst

The cave deposits are correlated both litho- and biostratigraphically to the Pennsylvanian. In western Illinois, the base of the Pennsylvanian is in most places the Babylon Sandstone Member of the lower part of the Tradewater Formation (Wanless, 1962). This unit is also the earliest widespread Pennsylvanian unit in northwestern Illinois (Leary and Trask, 1985). Peppers (1996) placed this unit in Westphalian B (=Atokan, =late Bashkirian). The Babylon Sandstone is comprised of medium to coarse quartz, with prominent secondary euhedral overgrowths (Wanless, 1962). Based on this distinctive lithology, we suggest that the massive sandstones in the quarry correspond to this unit.

In most of northern Illinois, including nearby localities to the southeast and southwest of the Central Quarry , the lithologic unit below the Colchester Coal is the Cheltenham Clay (Wanless, 1962). This unit is known locally as the Goose Lake Clay in Grundy County and the Lowell Clay and Utica Clay in LaSalle County. Wanless (1939) mentioned that sandstones similar to the Babylon sandstone are found in places beneath the Utica and Lowell Clays in LaSalle County. The Cheltenham Clay ranges from the upper part of the Tradewater Formation into the lower part of the Carbondale Formation. Lithologically, it is grey, breaks conchoidally, and is dominated by kaolinite, with minor amounts of illite and abundant pyrite. There are also grains of fine sand, which may be derived from the underlying St. Peter sandstone. Based on similar lithology and mineralogy, we propose that the clay deposits at the top of the sequence in the quarry (unit E in Figure 5) with Cheltenham Clay.

This stratigraphic placement is consistent with the biostratigraphy provided by the miospores. With the assistance of Cortland Eble of the Kentucky Geological Survey, we have placed the cave deposits within the *Torispora secures – Vestispora fenestrate* (SF), and *Radiizonates difformis* (RD) miospore zones of Peppers (1996), which are Middle to late-Middle Atokan (=Westphalian C, =Moscovian) in age. In sum, the paleocave deposits can with confidence be placed in the Tradewater Formation and in the Atokan Stage (=Westphalian C, possibly as old as Westphalian B).

The age of cave formation is more difficult to constrain, since it can only be bracketed based on the age of the host rock and the fill as being mid-Paleozoic. The regional distribution of erosional features below Pennsylvanian deposits indicates that it is correlative to the “post-Kaskaskia” paleokarst and thus formed in the latest Mississippian or earliest Pennsylvanian.

Discussion

The paleocaves and associated fills are significant for our understanding of plant evolution and paleoenvironments in this portion of the Pennsylvanian of the mid-continent North America. First of all, based on our correlation, the conifers are probably the oldest known from North America. The previously oldest described North American material (Rothwell et al., 1997), have been dated as Upper Moscovian (= Westphalian D). Second, nearly all Pennsylvanian plants represent wetland (mire) deposits; i.e., the classic Pennsylvanian coal forest vegetation. Plants from drier and well-drained environments are relatively rare. The lycopsids in our cave fill are predominant in the lower part of the sequence and prefer wetter environments. Early conifers, which are found near the top, are thought to live in drier environments (Lyons and Darrah, 1989). This interpretation is supported by the over-arching papillae of the conifers, a stomatal structure that allows growth under relatively dry conditions.

The presence of abundant charcoal in the deposit indicates the existence of ancient wildfires and thus of at least occasional dry episodes supporting combustion (Scott and Glasspool, 2006). The presence of fire is also indicated by the presence of large polyaromatic hydrocarbons (PAHs). Lipid extracts are dominated by PAHs such as pyrene, benzo[a]pyrene, benzo[e]pyrene, benzo[g,h,i]perylene, and coronene, which are typical products of vegetation fires.

Finally, as summarized by (Parrish, 1998), karst formation is generally indicative of a humid climate. The karst features may be attributed to the onset of wet conditions at the inception of the earliest Pennsylvanian.

What to look for:

1) Distribution of ancient caves along the quarry wall.
2) Profiles and sizes of the cave passages.
3) Facies differences within and among the cave fills.
4) Distribution of fossil plant material within the fills.
5) Sedimentary structures within the fill layers.
6) Fossils within the surrounding bedrock.

Questions to ask while looking:

1) Why are the sediments so poorly indurated?
2) Why are the plant remains so fragmentary?
3) What is the relationship of cave formation to changes in the local water table?
4) Why is are sulfide minerals so common?
5) Why are bones and teeth absent?

What questions remain:

1) When did the caves form?
2) Were there multiple episodes of cave filling?
3) What is the relationship between surface sedimentation and deposition within the karst system?
4) Did enlargement of the caves occur contemporaneously with sediment infilling?
5) What is the climate signal from karst formation and cave filling?
6) What was the surface environment like at the time the caves formed and when they filled?

Acknowledgements:

Cortland Eble (Kentucky Geological Survey) provided invaluable help on dating the fill. John Shaw (Central Limestone Co.) has been extremely cooperative in granting access to the quarry. Jack Wood and Arthur Palmer contributed useful discussions on caves and paleokarsts. S. Gibbons is thanked for laboratory preparation and P. Goggin of the Royal Holloway Electron Microscope Unit for help with the SEM. S. Guggenheim performed the x-ray diffraction analysis of the clay samples. G. T. Ventura first suggested the cave nature of the deposits, Z. Mateo and A. Burkemper assisted with the granulometry. W. DiMichele provided numerous useful comments. The comments of C. P. Weibel and B. B. Curry greatly improved the manuscript.

Stop 4: Oswego Channel

The 15,770 C-14 yr BP Oswego channel and evidence of its synchronous formation with other overflow channels

Brandon Curry, Kevin Befus, James Clark

The Oswego channel breaches the Marseilles Morainic System between the villages of Yorkville and Oswego (Figure 1). The floor of Oswego channel is underlain by a succession of silty glacial lake sediment, gyttja, marl, and peat; this succession rests on dolomite bedrock. The age of the base of the channel fill is 18,900 ± 35 cal yr BP (15,710 ± 35 C-14 yr BP), and is the weighted mean of four AMS C-14 ages of fossil stems and leaves of *Vaccinium ugilinosum*. These are the only known ages in Illinois (or in Midwestern North America?) associated with a breached mo-

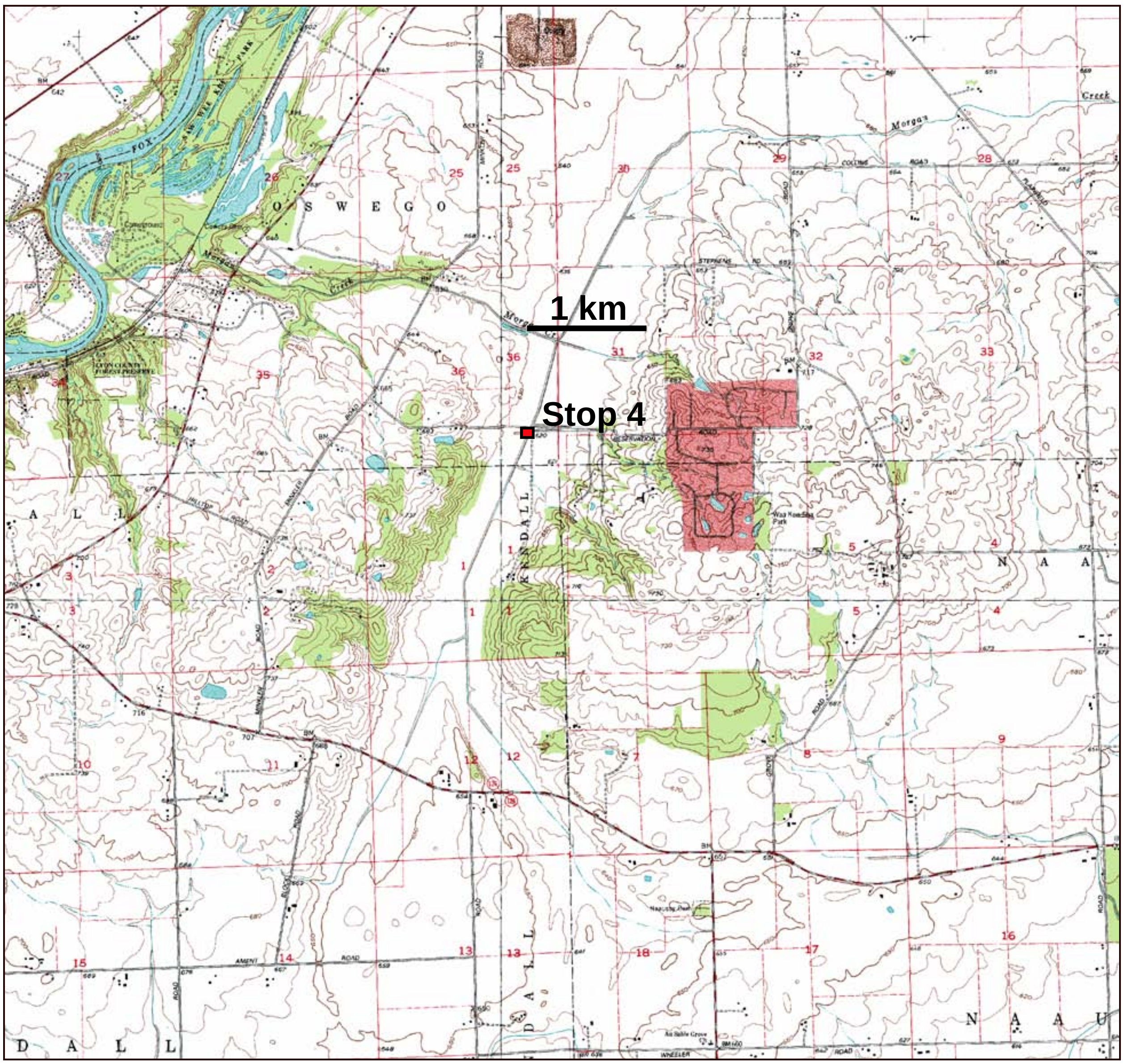


Figure 1 Location of Stop 4 and the Oswego overflow channel on parts of the Aurora South, Yorkville, Yorkville SE, and Plattville, Illinois, 7.5-Minute Quadrangles.

rainic dam at its source, and the discharge ("torrent") that ostensibly formed the channel. The ages are, however, about 900 years older than the oldest known slackwater lake sediment in the valley of the Illinois River near Havana, Illinois, attributed to large-scale flooding. The Havana site is just downstream of where the Illinois River crosses the Shelbyville Morainic System, the Wisconsin Episode terminal moraines of the Lake Michigan Lobe.

Glacial Lake Wauponsee is one of several proglacial lakes dammed by looping moraines of the Peoria and Decatur sublobes (Willman and Frye, 1970). The formation of the overflow channels across the Marseilles Morainic System was likely associated with overflow of Glacial Lake Wauponsee. North of the Illinois River in Kendall County, deposits attributed to Glacial Lake Wauponsee are as much as about 10 m thick, but are typically less than 3 m thick. The sediment is rhythmically bedded to uniform silty clay, silt loam, and silt. The rhythmically bedded sections appear "varved" and are comprised of centimeter-thick beds of silty clay intercalated with 1-3 mm layers of silt. No dateable fossils have been found in these deposits, although they should be amenable to optically stimulated luminescence dating. The elevation of the top surface of the surficial lake sediments are about one meters higher than the elevation of the overflow sill (196.9 vs. 195.7 m) consistent with channel formation by overflow of Glacial Lake Wauponsee.

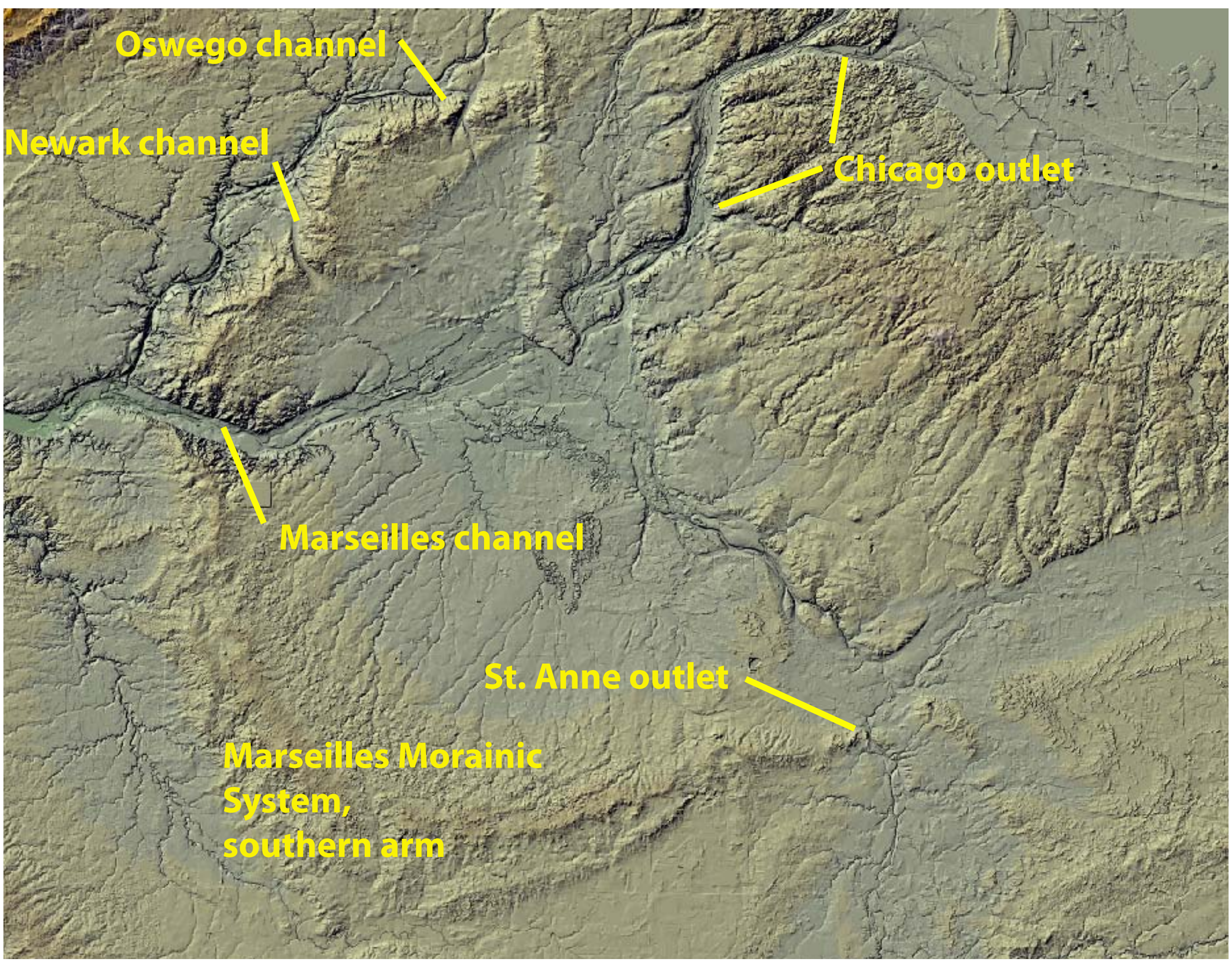


Figure 2 Location of the Oswego, Newark, and Marseilles channels, and St. Anne outlet across the Marseilles Morainic System. The base map is from Luman et al. (2003).

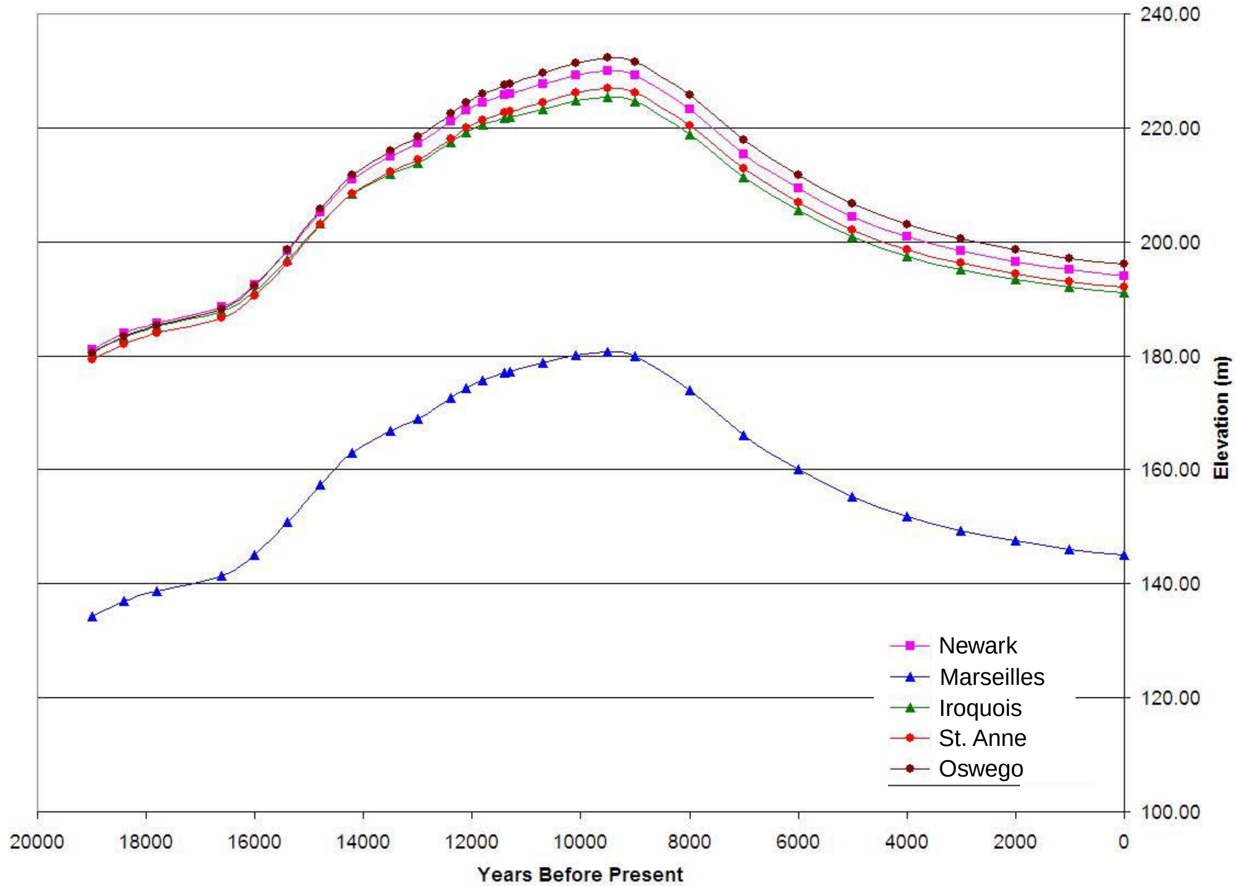


Figure 3 Time dependent elevations of Newark, Marseilles, and Oswego channels, and of the St. Anne and Iroquois outlets.

The Oswego channel is one of three that breach the Marseilles Morainic System; the others are the Newark and Marseilles channels (Figure 2). Near their heads, the elevations of the overflow sills of the Newark and Oswego channel are about 197.5 m and 195.7 m, respectively. The ancient floor of the Marseilles channel was subsequently eroded, notably by discharge that formed the Chicago outlet.

To explore the idea that Oswego channel formed first, followed by the Newark and Marseilles channels due to isostatic rebound, we used the model of Clark et al. (2008). Figure 3 shows the time dependent deformation of the three overflow channels. Note that during the time of interest, 19,000 to 18,000 yr BP, there is little differential rebound between the sites. Models of Glacial Lake Wauponsee at 19,000 and 18,000 yr BP (Figures 4 and 5) were done by (1) creating a ground surface DEM from 10-m data available from the USGS, (2) tilting the model according to isobases generated by the model of Clark et al. (2008) and , (3) modeling the extent of Glacial Lake Wauponsee. The lake was modeled to fit between the Joliet sublobe at about the position of the Minooka and Rockdale Moraines, and the Marseilles Morainic system (the Marseilles channel was artificially “plugged”). The lake levels shown in the figures are the highest levels before water began to drain across the Oswego channel. The aerial extent of Glacial Lake Wauponsee shown in Figures 4 and 5 are likely conservative because the evidence of the initial sills are not preserved in the channel valleys. It is interesting to note that the southern shores of Glacial Lake Wauponsee coincides with the approximate northern extent of the gigantic field of ice-walled lakes that occur on the southern arm of the Marseilles Morainic System. This sug-

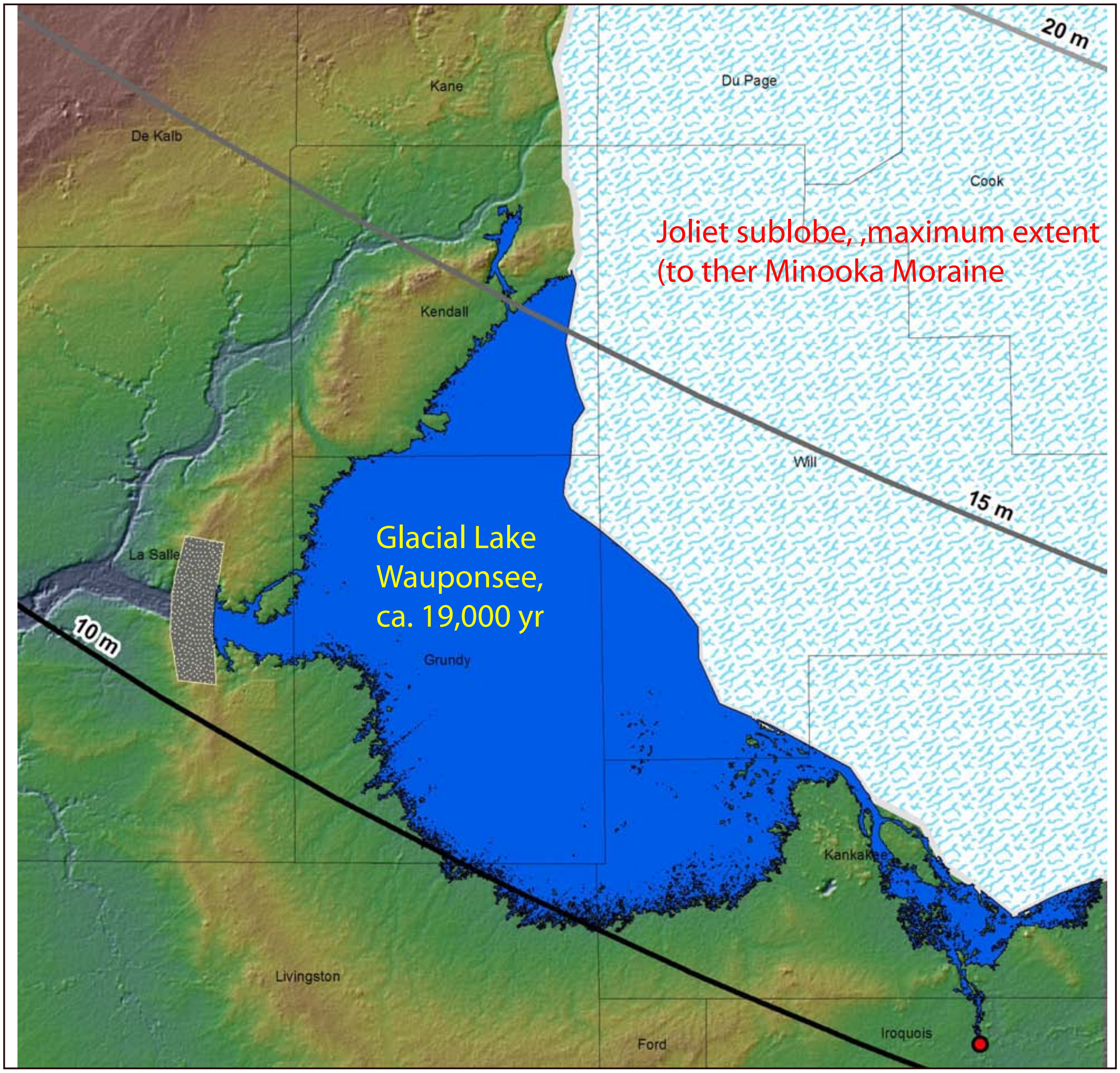


Figure 4 Predicted isobases (meters) of deformation relative to present for 19,000 yr BP, and extent of Glacial Lake Wauponsee. The position of the Joliet sublobe is inferred.

gests that some ice-walled lake plains may have been obliterated by Glacial Lake Wauponsee or are onlapped by its deposits.

The results of the deglacial isostatic rebound models indicate that the three channels that the Oswego, Newark, and Marseilles channels formed at the same time. This suggests that rising water in the proglacial lake exploited low passes of nearly equal elevation across the moraine. Multiple breaches across moraines are known from other moraines in Illinois, notably the Marengo and Rockdale moraines. The elevation of gaps (breaches) and diamicton thickness across the Marengo Moraine suggest that proto-channel formed while the moraine was being deposited (Curry et al., 1997). During deglaciation, meltwater exploited the low pass across the moraine,

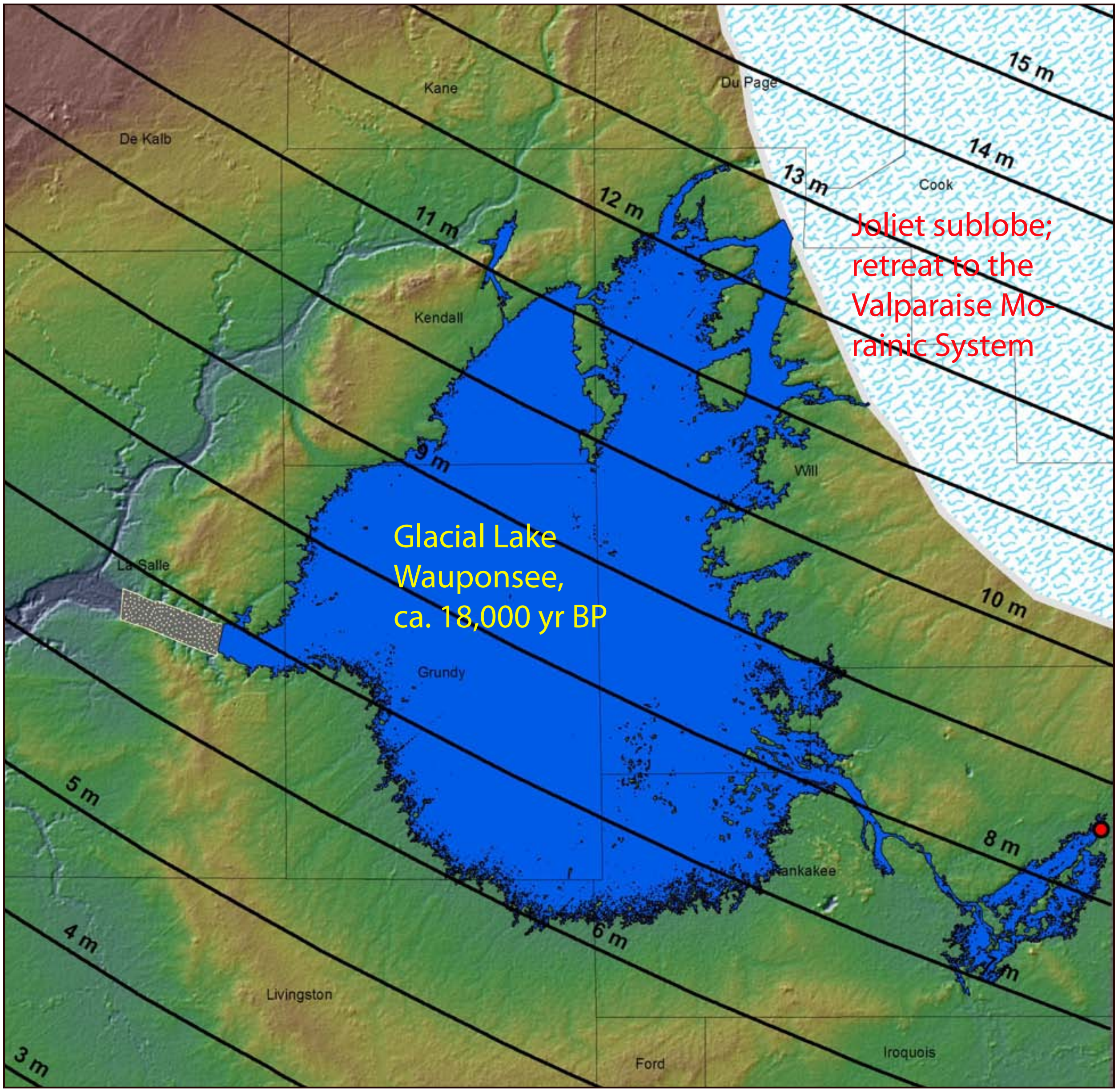


Figure 5 Predicted isobases (meters) of deformation relative to present for 18,000 yr BP.

and in many cases, modern drainage follows these drainage paths. In the case of the Marseilles Morainic System, the three channels across the Marseilles Morainic System may have been formed initially by subglacial drainage, and later were exploited for drainage by Glacial Lake Wauponsee.

There is little geomorphic evidence to suggest that the Oswego channel formed by one large discharge event, although high-quality shaded-relief maps of northern Kendall County may reveal large-scale structures covered or obscured by modern infrastructures. There is, however, geomorphic evidence for one or two large discharge events associated with formation of the Newark channel. Aerial photography reveals megaripples and younger distributary channels on an alluvial fan or delta where the Newark channel widens into the valley of the Fox River (Figure 6).

Figure 6 Porton of the Newark, Illinois, 7.5-minute Quadrangle (top) and digital orthophotoquad (USGS, 1983; bottom) showing maegaripples and younger distributary channels fan deposits leading from the mouth of the Newark channel towards the Fox River.

Sediment cores

Several sediment cores have been sampled along Reservation Road with a PowerProbe and CME 750 drill rig; three cores were sampled at Stop 4 where the road crosses Morgan Creek. Each core revealed about 8.5 m of lacustrine and paludal sediment resting on dolomite bedrock. The bedrock "high" in this region may explain the existence of the lake. Although it may have simply been eroded to its configuration during overflow, the northern lip of the basin may be sealed by the gently rising bedrock surface. The cored successions includes 3.41 m of gray silt loam with few interbeds of fine to coarse sand, 2.59 m of silty, organic gyttja, 0.79 m of marl, and 2.44 m of surficial peat. The moisture content of the lower silt unit is about 50%, which more than doubles in the lower part of the gyttja unit to more than 120%. One sample of shell-rich peat yielded a moisture content of 350%. Core recovery of marl and peat was less than 15%, whereas core recovery of the silts exceeded 100% (the latter occurs due to vertical expansion of the cored material during sampling). There is great potential to do a paleoecological study of paleovegetation and paleohydrological change during the topographic inversion that will be discussed more fully at Stop 7, in particular, of the changes that occurred when vegetation shifted from tundra to boreal forest. Ostracodes, plant macrofossils, and charcoal are especially abundant in the succession.

Future studies

Our work in this area indicates the following themes for future research:

1) Obtaining OSL ages on sediments of Glacial Lake Wauponsee to confirm the connection between the latter and formation of the overflow channel.

2) Understanding the nature of the transition from when vegetation changed from tundra to boreal forests. The topographic inversion that occurred from about 15,000 to 14,000 C-14 yr BP is not represented in either ice-walled lake deposits (too old) or in most kettle succession (too young).

3) Although we have tried at three localities, additional cores should be taken in the Newark channel to see if it has dateable material to confirm that it formed at about the same time as the Oswego channel. Our earlier cores showed thin debris (< 1 m) modified by the modern soil over silty clay till.

4) Determining discharge rates of Glacial Lake Wauponsee through geomorphic studies of the overflow channels.

5) Understanding the age and nature of the interconnection between Glacial Lake Wauponsee and other large proglacial lakes such as glacial lakes Iroquois and Pontiac, as well as the potential discharge from the receding Erie, Saginaw, and eastern Lake Michigan lobes via the Kankakee River.

6) Gain a better understanding of the slackwater records in the Illinois River valley such as the one near Havana, Illinois. Perhaps more earlier and more complete records of large floods occur in large valleys tributary to the Illinois, such as the Spoon or Sangamon rivers.

Stop 5: Mastodon Lake at Phillips Park, Aurora

History, Educational Outreach, Mastodons, Cosmic Dust?, and Geology

Introduction

The remains of mastodonts are known from marl and peat-filled kettle depressions located throughout continentally glaciated North America. Many of these fossils were deposited during the last glacial/interglacial transition, the ca. 2,400 year period known to many as the Bölling, Allerød, and Younger Dryas Chronozones. At this stop, three “experts” will give their perspective of recent work done at "the Aurora site" and other sites in the region. Brandon Curry will review the geology of the site, Jeffrey Saundeers will provide a literature review of the Aurora mastodonts, in addition to a picture-laden review of mastodonts in Illinois, and David Vorhees will discuss the history of excavations at the site, and his experiences with the "Aurora Mastodon Project", a successful education outreach program funded by the City of Aurora.

Geology

Brandon Curry

Mastodon Lake occupies a complex of interconnected kettle basins on the western flank of the Minooka Moraine (Figure 1) which is formed of clayey diamicton of the Yorkville Member, Lemont Formation. The west side of Mastodon Lake is nestled in sand and gravel of the Henry Formation (Willman and Frye, 1970; Hansel and Johnson, 1995). The Minooka Moraine was

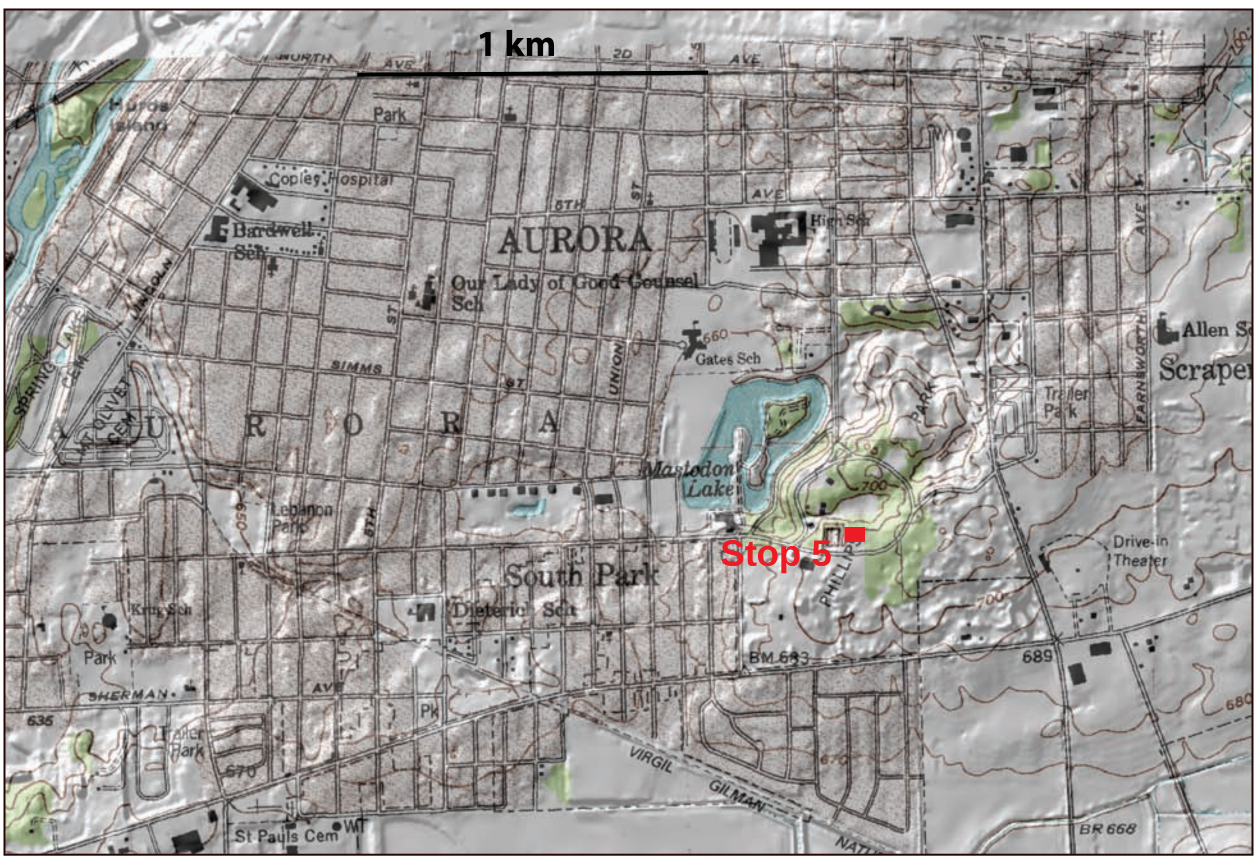


Figure 1 Location of Mastodon Lake in Phillips Park, Aurora, Illinois. The map is located on the Aurora South, Illinois 7.5-minute Quadrangle.

formed during the late Livingston Phase, between about 20,750 to ≈20,000 cal yr BP. At Mastodon Lake, radiocarbon ages from organics found above the Yorkville diamicton are much younger. Here wood fragments encased in the basal lacustrine silt from a backhoe trench on the large island yielded an age of 17,480 ± 180 cal yr BP (14,130 ± 70 C-14 yr BP).

The geologic investigation of Mastodon Lake included sampling eight 4.5-cm diameter cores with a PowerProbe. Many coring sites were located on terraces that were interpreted from the shaded relief map of 2-ft contour data (McGarry, 1990). Without exception, the terraces proved to be man-made benches composed of fill. It appears that the entire kettle basin has either been scooped out (to deepen lakes and ponds for recreational use), or has been stripped of peat and marl and replaced by fill composed of compacted diamicton, peat, and marl. The sediment cores indicated that a site just north of present-day Mastodon Lake has the thickest succession of silt, gyttja, marl, and lower peat. In late 2007, a continuous 10.9 m core was sampled at this site using a Livingstone piston corer. The core was sampled three years after the mastodon dig as part of a separate study of four lakes in northeastern Illinois to examine the pollen and ostracode records from the last glacial to interglacial transition. The study is projected to be completed by the year 2010. Among the four sites, Mastodon Lake was the last site that was cored, so we have the least information about it. The other three sites are located at Brewster Creek, Crystal Lake, and Nelson Lake. Eventually the core will be examined in great detail, with perhaps as many as 20 radiocarbon ages to help anchor a sediment accumulation model. The regional geology and paleoenvironments are discussed in a recently completed report on the geology, paleohydrology, and paleovegetation of the Brewster Creek site (Curry et al., 2007).

Stratigraphy

The contacts between the units described below are abrupt. The sediment succession recovered from the Phillips Park study includes, from base to top:

1) Diamicton, matrix supported, with a silty clay matrix or very poorly sorted sand and gravel outwash about 15 m thick

2) Organic-rich silt, uniform to laminated, about 0.5 m thick, that occurs on the basin margin where the student’s excavations took place. The unit contains large wood fragments, including logs, probably of spruce. Large wood fragments have yielded the following radiocarbon ages: 14,130 ± 35 C-14 yr B.P. (ISGS-5655), 13,710 ± 35 C-14 yr B.P. (ISGS-5656), and 13,600 ± 35 C-14 yr B.P. (ISGS-5633).

3) Gyttja, a deep-water facies of unit 2, the organic-rich silt. Gyttja is a coprogeneous lake sediment, and is as much as about 5 m thick. The deposit is bioturbated, uniform, or finely laminated. Initially a dark olive-green, the gyttja oxidizes within about 5 minutes to black. The texture is primarily silt.

4) Marl, composed of almost entirely biogenic carbonate, is as much as 3.1 m thick. Along the kettle margin, the marl is uniform, but towards the center of the kettle, the marl is banded. Some marl layers are fiber-rich. Mollusc shells, ostracode valves, micrite, and carbonate encrustations on charophyte oogonia and stems are the bulk of the biogenic carbonate.

5) Fibric peat with matted vegetation typical of wetlands is as much as 1.7 m thick. In many places, this unit often contains glass, fragments of clay pigeons, etc. A sample of basal, matted peat with abundant reed stems yielded an age of about 10,210 cal yr BP (9,115 ± 35 C-14 yr BP).

6) Fill as much as 2.3 m thick consisting of redeposited peat, marl, diamicton, bricks, mortar,

sinter, ash, glass, and other debris encased in blocks of unweathered diamicton, peat, or marl.

The most important aspects of the succession described above have yet to be explored. We would like to verify the hypothesis that the silt (gyttja)-to-marl transition occurred at the onset of the Bölling Chronozone, about 14,670 cal yr BP (12,500 C-14 yr BP). In addition, we are examining the core in some detail for geochemical, charcoal, and other evidence for a bollide explosion associated with the onset of the Younger Dryas at about 12,900 cal yr BP (11,000 C-14 yr BP (Firestone et al., 2007).

Annotated bibliography of paleontological resources for Aurora, Illinois

Jeffrey Saunders

Wilber, C. D. 1861. Mastodon giganteus. Transactions of the Illinois State Agricultural Society, with Notices and Proceedings of County Societies, and Kindred Associations. Volume IV.--1859-'60.

"The 'largest specimen' of this order, (Mastodon,) once lived near Aurora, where his remains were recently found [1850 fide Smith, see below; 1853 fide Anderson, 1905], in excavating for the track of the Chicago, Burlington and Quincy railroad. There were the tusks and seven teeth--all in a good state of preservation, the "tooth of time" having consumed all other vestiges. The teeth and tusks were found as near each other as when they were in the animal's head; from which we may conclude that he laid him down to die with much composure, and was allowed to sleep on quietly through the ages."

Anderson, N. C. 1905. A Preliminary List of Fossil Mastodon and Mammoth Remains in Illinois and Iowa. Augustana Library Publications, Number Five, Rock Island, Ill.

"Kane County. Aurora.--In 1870 tusks and several teeth of a mastodon were obtained from the superficial deposits of this county near Aurora when the excavation for the track of the Chicago, Burlington and Quincy railroad were made. These remains are in the Museum of Clark Seminary at that place (Illinois Geological Survey, Vol. IV, p. 113)."

Aurora.--In 1853, while extending the Burlington railroad south of Aurora, workmen found teeth and a tusk of a mastodon in a swamp on the edge of Fox River, where the Burlington repair shops at Aurora are located. The remains were presented to Jennings Seminary by an official of the road, Benjamin Hackney. (Reported by Mrs. Susan H. Quereau, Aurora.)"

Smith, C. R. 1935. Mastodon and other Remains at Aurora, Illinois. Science, April 19, 1935, Vol. 81, No. 2103, pages 379-380.

"Finding of mastodon parts and other material during recent months will contribute items of interest regarding the life of this vicinity in early post-glacial times. The finds were made by CWA workers while digging for an artificial lake in a swamp in Phillip's Park, which is located in the southeast part of Aurora, Illinois.

The mastodon parts consist of three skulls, one of which includes the lower jaw, three tusks, a femur, an ulna, a scapula, a number of ribs, several vertebra and a number of foot bones. Most of the material is in excellent preservation. E. S. Riggs, ... Field Museum of Natural history in Chicago, ... identified the species as being Mastodon americanus. There were also found in the same formation as the mastodon material three pairs of bird humeri and a portion of breast, all of the same species of bird. Identification has not yet been made of the bird specimens, but they are being examined by Professor L. A. Adams, ... the University of Illinois. The size suggests a bird possibly four feet in height.

The [fossil producing] deposit...is a bed of gray marl enclosed on three sides by hills of glacial till, ... situated a mile and a half east of the Fox River. Professor William E. Powers, ... Northwestern University, has examined the geological features of the locality, and believes that the marl represents a post-glacial lake which probably once connected with the river. A series of borings made in a north ...- south line across the marl bed revealed a maximum thickness of thirty feet. Overlying the marl was a layer of peat varying in thickness from two to five feet and over this about two feet of black muck which comprised the bottom of the modern swamp. The mastodon and bird skeletal parts were found in the upper three feet of marl, with the exception of the [mastodont] scapula which was in clay at the margin of the marl bed. This was the first specimen found, and obscurity of reports as to exact locality do not justify definite conclusions as to whether it differs in age from the rest of the specimens [thus 3 or 4 individuals represented].

A hemlock cone found in the cavity of one of the mastodon tusks has been identified by Dr. W. T. McLaughlin, ... Northwestern University, as being of the species, *Tsuga canadensis*. Several other cones found in the marl of the same vicinity, he considers to be of the same species. There were also found two cones which he considers to be apparently black spruce, *Picea mariana.*

Professor F. C. Baker, ... University of Illinois, ... identified twenty-one species of shells in a sample of the marl sent to him by Professor Powers. Baker reports that “it is, as far as climate is concerned, a cold-temperate fauna.” He also states that it is “quite like the marl fauna found a few years ago in the bottom of Green Lake, Wisconsin, which is certainly middle Wisconsin in age, not later.” He considers the cones of hemlock and spruce as further indication of a cold-temperate climate.

Another find of interest was a right femur of the giant beaver, *Castoroides ohioensis*, the specimen ... identified by Professor Adams. It was reported by workmen to have been found in the peat layer, but there is reason for believing that this may be an error and that the specimen was more likely in the marl.

In the peat layer quite a collection of mammal skeletal parts has been found and most of it examined by Professor Adams. Most frequently represented is the Virginia deer. A skull he has identified as that of a female elk. The most recent find in the peat layer is a skull apparently of a muskrat. This has not yet been studied in detail.

With the completion of the lake-digging project the finding of specimens has now come to an end. .... The City of Aurora is keeping the specimens on display in a museum at Phillip’s Park.”

Powers, W. E. 1935. Geological Setting of the Aurora Mastodon Remains. Transactions of the Illinois Academy of Sciences, Vol. 28:193-194.

“Since March, 1934, excavation for a municipal lake in Phillips Park, Aurora, has yielded remains of a surprising assemblage of now extinct animals. Among these remains are three skulls of the American mastodon, three tusks, a lower jaw, and other smaller parts of the same animal, together with bones belonging to the trumpeter swan, giant beaver, deer, elk, bear [this is the only mention of bear among the 1935 reports], and other animals. These bones have been carefully studied by professor Clarence R. Smith of Aurora College. Largely through his work, they have become known to the local public and have aroused large popular interest.

The area from which these remains were taken is a peat bog in the southeastern part of Aurora. Part of the bones came from the peat itself, but practically all the bones of the mastodon and trumpeter swan were found in the upper part of shell marl that underlies the peat.

The peat ....attains a maximum thickness of about six feet, and everywhere except at the edges it rests on shell marl. Beneath the peat is white to light gray marl known to attain a thickness of 30 feet."

Smith, C. R. 1935. Mastodon and Other Finds at Aurora. Transactions of the Illinois Academy of Sciences, Vol. 28:195-196.

"Finding the remains of mastodon and other forms of life in an old bog at Aurora has furnished a glimpse of that locality in a more or less fragmentary sequence from early post-glacial times down to the present. The locality is within the limits of Phillips Park in Aurora and about a mile and a half east of the Fox River. The bog itself is surrounded on three sides by hills of glacial till, and the formation consists of a deep deposit of gray marl surmounted by a layer of peat, in turn covered by a layer of black muck which has comprised the bottom of the modern swamp. The specimens were found by CWA workmen while excavating for an artificial lake.

The mastodon parts were found in the upper 3 feet of marl deposit and consist of three skulls (one including the mandible), three tusks, a scapula, an ulna, a femur, a number of vertebra, ribs, and several foot bones. In color the bones were brownish-yellow to brown and were in a good state of preservation except for the scapula and one skull which were very fragile. One skull measured 45 ½ inches long and 28 ½ inches wide. Two of the tusks appeared to constitute a pair, nearly white in color, of similar curvature and lengths–8 feet 2 inches and 8 feet 3 inches. The third tusk was not well preserved and broke into three pieces during exhumation. Mr. E. S. Riggs ... Field Museum of Natural History, ... Chicago, identified the species as Mastodon americanus.

For preservation treatment the mastodon bones were saturated with the following mixture: Varnish having 100 per cent bakelite base 70 cc., turpentine 18 cc., raw linseed oil 10 cc., and oil of wintergreen 2 cc. The use of bakelite varnish for fossils is not new; the other ingredients were added after experimentation by the author. It was hoped that the penetrating properties of oil of wintergreen would be an advantage with dense materials such as tusk ivory and teeth. The mixture is more efficacious than shellac in hardening fossil bone and in retarding the cracking as the bone dries and shrinks.

At the pulp end of the two better preserved tusks were a series of six or eight faint ridges or so-called "growth rings." Such rings have been mentioned by Kunz and others but their true significance is uncertain. However, if we entertain the possibility of their representing years of growth, the figures are interesting. The average width of the rings is 62 mm., and assuming uniformity for the entire length of the tusks, would represent the age of the animal as 41 years. Within each of the wide rings is a series of about seven narrower rings of differing widths, and in any two adjacent wide rings these secondary ring patterns are remarkably similar. This seems to support the theory that the wide rings represent yearly growth and that the secondary pattern represents a physiological response to change in weather, abundance of food, kind of food, etc., within the year. [It is ? noteworthy that only since 1990 have the implications of these observations by Smith, regarding in essence "annual and seasonal cycles in proboscidean dentin" become a topic of research–tuskology; see for example Fisher 1990].

In the north area of the marl deposit were found at about the same level as the mastodon parts, three pairs of bird humeri and breast portions which were identified by Dr. Alexander Wetmore, assistant secretary of the Smithsonian Institution, as being of the trumpeter swan (*Cygnus buccinator*). The species is now said to be almost extinct, being found only in parts of northern Canada, although according to Audubon and other early writers it was abundant in the Mississippi Valley a century ago. It is of no small interest to trace its residence here back to Pleistocene time.

A jaw and right femur have been identified as being of the giant beaver (*Castoroides ohioensis*), the jaw having been identified by Mr. E. S. Riggs ... and the femur by Professor L. A. Adams of the University of Illinois. The reports of workmen are obscure as to depth and other circumstances, but there is reason to believe that the specimens came from the upper marl layer.

In the peat layer above the marl, and consequently of later date, were found several horns and numerous bones of the Virginia deer, also skull and other parts of elk. Most of this material was submitted to Professor Adams who very kindly studied the collection and made identifications.

Buried in the black earth above the peat was a small skull identified as muskrat by Dr. S. H. McFarlene of Aurora College. Its condition suggests an age of perhaps a hundred years, thus bringing the sequence up to modern times. The black earth layer also produced the skeleton of a buffalo which does not show evidence of having been buried more than 30 to 50 years. The Custodian of Phillips Park is sure it could not be a captive buffalo which died in the park, his opinion being based on the skeleton having been found outside of the former park boundary. The finding of a buffalo skeleton, however, would not be out of harmony with the well known fact of their inhabiting this area in modern times.

A small cone found with the marl in the hollow end of one of the mastodon tusks has been identified by Dr. W. T. McLaughlin of Northwestern University as being of the species Tsuga canadensis, and throws some light on the vegetation of the time. Numerous sticks and logs of wood were found deep in the marl as well as in the upper layers. Some of these were twisted similar to cedar. The tapered end of one stick suggested that it may have been cut by a beaver.

The City of Aurora plans to keep the specimens. They are at present on display in a temporary Museum at Phillips Park, and an attempt is being made to promote interest in a better housing for the material."

Wetmore, A. 1935. A Record of the Trumpeter Swan from the Late Pleistocene of Illinois. Wilson Bulletin, XLVII, September, 1935, page 237.

"In material secured at Aurora, Illinois, by Professor Clarence R. Smith of Aurora College, forwarded to me for examination by Dr. L. A. Adams of the University of Illinois, I have identified humeri, a broken sternum, and part of the scapula of the Trumpeter Swan (*Cygnus buccinator*). According to Professor Smith [1935. Science 81, 380, see above] these specimens were found in a marl deposit underlying a peat bog above which was a layer of muck forming the bottom of a swamp in Phillip's Park in the southeast part of Aurora at a point a mile and a half east of the Fox River. They were obtained by C. W. A. workers during excavation of the swamp to make an artificial lake.

The swan remains were associated with bones of mastodon (*Mastodon americanus*) and giant beaver (*Castoroides ohioensis*) and are believed to have been deposited in the bed of a post-glacial lake. .... Dr. Adams writes me that mollusks of Pleistocene species were taken from the pneumatic foramina of the swan humeri.

The bones are distinguishable at a glance as those of the Trumpeter Swan. The sternum, while in fragments, shows the characteristic bulbous swelling projecting into the body cavity at the anterior end in addition to the swollen channel in which the trachea is folded. The humeri exhibit two sizes, possibly indicative of sexual difference in wing measurement, and are greater in size than the largest Whistling Swans. All of the specimens , which bear numbers 111, 734, B, C, D, F, H, J, and K, are fresh and clean in appearance, and are in good state of preservation.

Previously this swan has been reported from Pleistocene deposits in Oregon and Florida, the present being the first occurrence of it in the central portion of our country. The find is one of definite importance in view of the few reports of birds that have come from Pleistocene beds of the area in question.

The specimens have been returned to Professor Smith, ... and are preserved by the city of Aurora in a museum at Phillip's Park.

Smith, Clarence R. 1967 "The Mastodon Finds at Aurora" a note on reposit in the Illinois State Museum (September 20, 1967):

"The mastodon bones of 1934 were found in the upper three feet of a marl deposit which was surmounted by a 5-ft layer of peat which in turn was overlain by a layer of black muck which comprised the bottom of the modern swamp. A series of borings revealed the marl deposit to have an average thickness of about 15 feet, although at one spot a thickness of nearly 40 feet was found.

Dr. William E. Powers of Northwestern University has described the glacial setting of this bog as "between a high rolling ridge, the Minooka glacial moraine, on the east, and a flat terrace of gravel on the west. The gravel terrace stands 50 to 60 ft. above the Fox River and evidently marks a period when the Fox River valley was filled with gravel to the level of the present terrace top.

Interest in the finds of 1934 led to searching of old newspaper reports and interviewing farmers and old-time residents regarding previous finds. Altogether, there is now evidence of at least 18 individual mastodons within 20 miles of Aurora....The first find in the area was in Aurora in 1850 when a 10-foot tusk and several teeth were found. Other finds have been made not only in Aurora but also at Batavia, Bristol Station, Hinckley, Kaneville, Maple Park, Oswego, Plano and Yorkville. Another find not far from the Aurora area was made at Crystal Lake in December, 1958 [Smith, C. R. 1960. Elephants at Crystal Lake. Earth Science 13, 63-64.].

Also found at Aurora, associated with the mastodon finds of 1934, were bones of the Trumpeter Swan and Giant beaver."

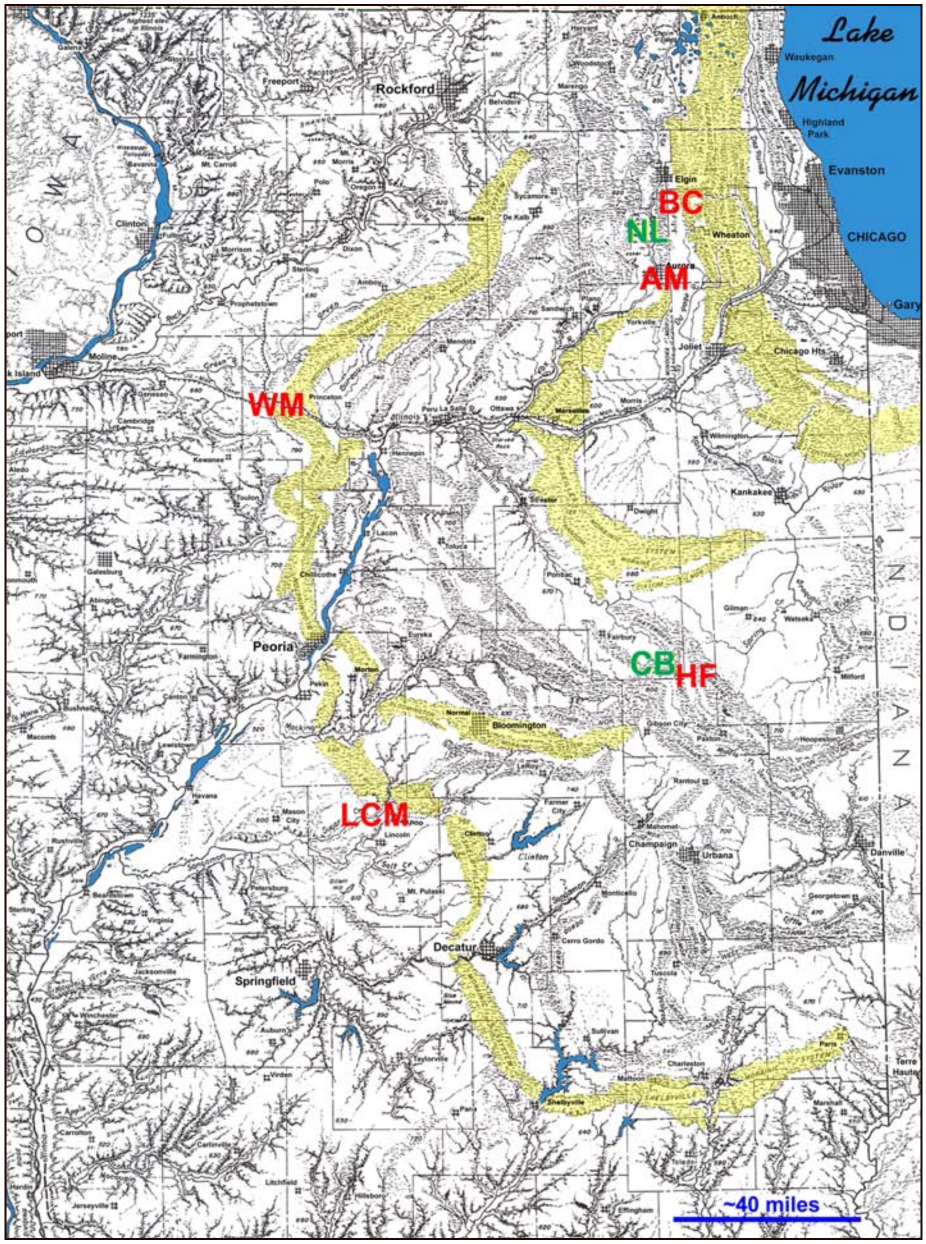


Figure 2 A map showing the relationship of recently dated proboscidean remains to primary glacial features and pollen profiles in northeastern Illinois. The locations of the Wyanet Mammoth (WM), Lincoln College Mammoth (LCM), and Hawthorne Farm (HF) Mastodont relative to the Chatsworth Bog (CB), pollen profile and of the Brewster Creek (BC), and Aurora mastodonts (AM), and the Nelson Lake pollen profile (NL) are indicated.

Figure 3 (Facing page) Aurora mastodonts “1" (above) and “2" (below) are two of four mastodonts (*Mammut americanum*) recovered beginning in March 1934 by Civil Works Administration workers digging what is now known as Mastodon Lake in Phillips Park, Aurora, Illinois. Renewed excavations in the summer of 2004 by James S. Oliver (Illinois State Museum), Brandon Curry (Illinois State Geological Survey), and David Voorhees (Waubonsee Community College) were supported by the City of Aurora Parks and Recreation Department. Although these excavations failed to uncover additional remains, they facilitated the two AMS dates shown here. Both dates are later than Brewster Creek *Mammut* and currently dated *Mammuthus* spp. As of now Mastodont 1 at Aurora is Illinois’ “last mastodont standing.”

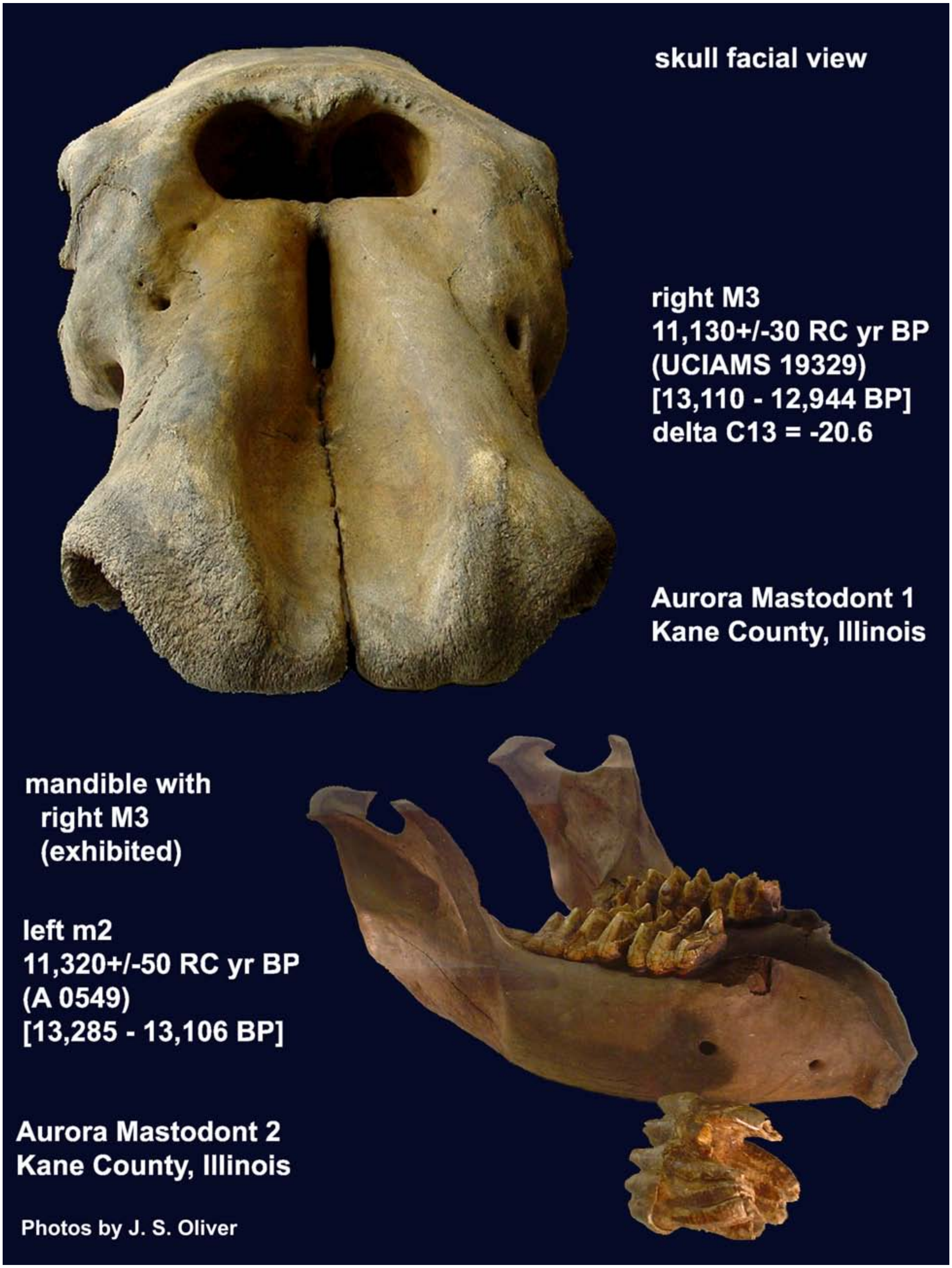

skull facial view
right M3
11,130+/-30 RC yr BP
(UCIAMS 19329)
[13,110 - 12,944 BP]
delta C13 = -20.6
Aurora Mastodont 1
Kane County, Illinois
mandible with
right M3
(exhibited)
left m2
11,320+/-50 RC yr BP
(A 0549)
[13,285 - 13,106 BP]
Aurora Mastodont 2
Kane County, Illinois
Photos by J. S. Oliver

Figure 4 The Brewster Creek mastodont *(M. americanum)* was discovered on the present day surface following a controlled burn in August 2005 by Daniel Terpstra while conducting wet-land restoration in Pratt's Wayne Woods, a part of the Forest Preserve District of Du Page County near Bartlett, Illinois.
Brewster Creek is one of four Chicago area sites currently being investigated by Brandon Curry of the ISGS, Eric Grimm of the ISM, and colleagues (the other sites are Nelson, Crystal, and Mastodon lakes: Curry, B. B, et al. 2006). Primarily for this reason, its recovery was of immediate interest that expedited its dating, funded by the Forest Preserve District of Du Page County and coordinated by Dr. Tom Stafford.

Although it is not a particularly late date for Illinois mastodonts, the age indicated for Brewster Creek–11,455+35 C-14 yr BP–postdates the youngest date (11,500+160 C-14 yr BP) provided by Guthrie (2004) for *Mammuthus primigenius* in mainland Alaska and the Yukon. This, coupled with the fact that the Aurora mastodonts are later, whereas no currently dated mammoths are, suggest the possibility that in Illinois *Mammut* americanum survived *Mammuthus* spp. in the latest Pleistocene–a view long ago nurtured by Clarence Smith: ".... It is well known that these gigantic beasts were at home in Northern Illinois in early post glacial times. Remains of eighteen individual mastodons have been found within twenty miles of Aurora where a major series of finds have been made in 1934. Many other finds have been made in Illinois.
A number of these are of written record and there is hardly a museum or collection of curios in the state which does not have a mastodon or mammoth tooth. In data so far accumulated by this author, specimens found in gravel deposits have represented the mammoth and those found in peat deposits have been mastodons. If more extensive evidence should follow this pattern it would indicate that the mammoth was the earlier inhabitant, possibly being here very early in the retreat of the Wisconsin glaciation if not actually disturbed by the last state of this glaciation. The mastodon on the other hand, where positive identification could be made, has not been found in the gravel deposits but in the marl and peat beds of several thousands of years later date. ...."[Smith, C.R. 1960. Elephants at Crystal Lake. Earth Science 13, 63-64.]

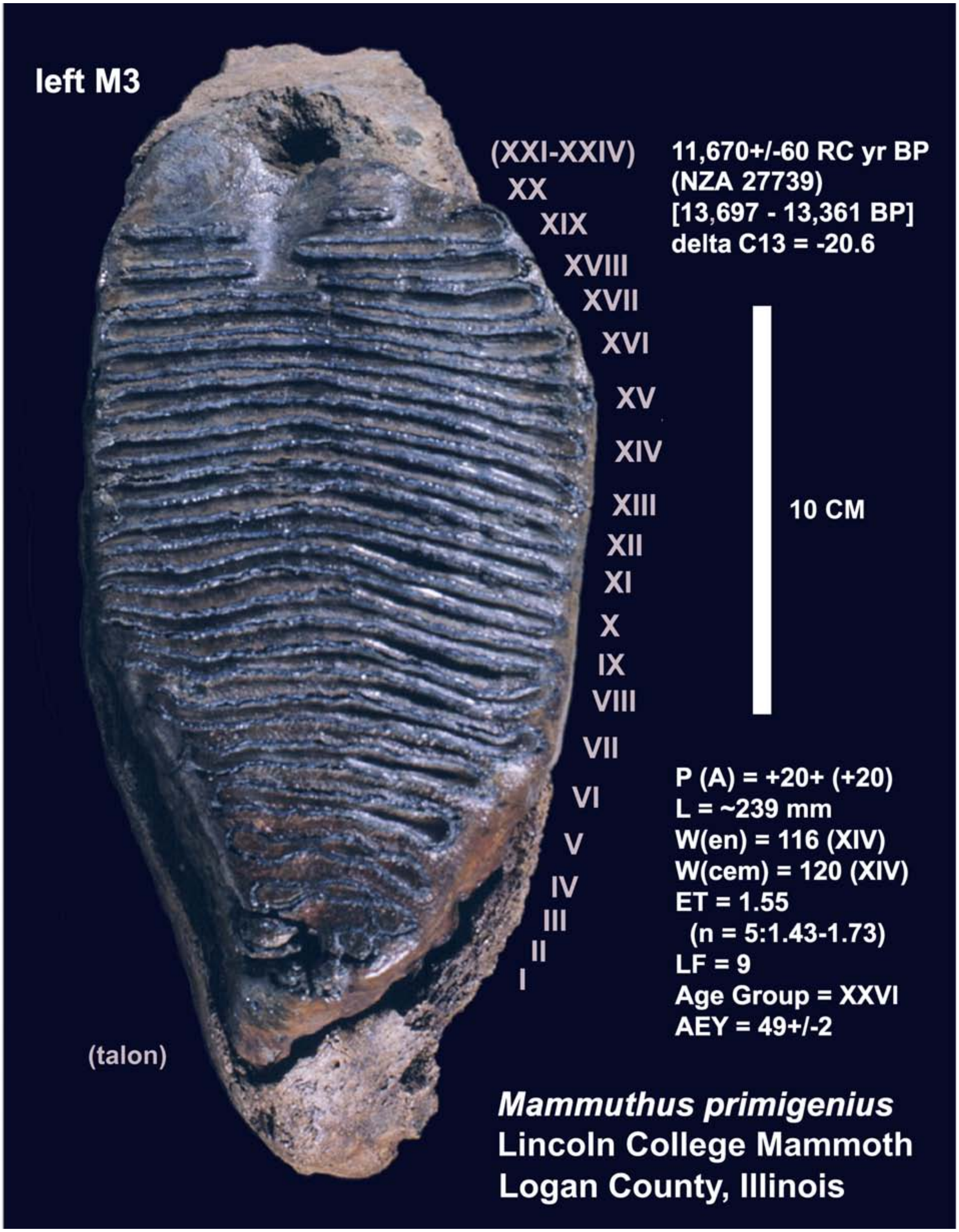


Figure 5 The left upper 3rd molar (with maxillary bone adhering, including the resorbing alveolus of the 2nd molar) of the Lincoln College Mammoth (*Mammuthus primigenius*), recovered in December, 2005 by Dr. G. Dennis Campbell, of Lincoln College. Its features of short length, wide width, thin enamel, and high ridge-plate compression clearly indicate a woolly mammoth is represented.

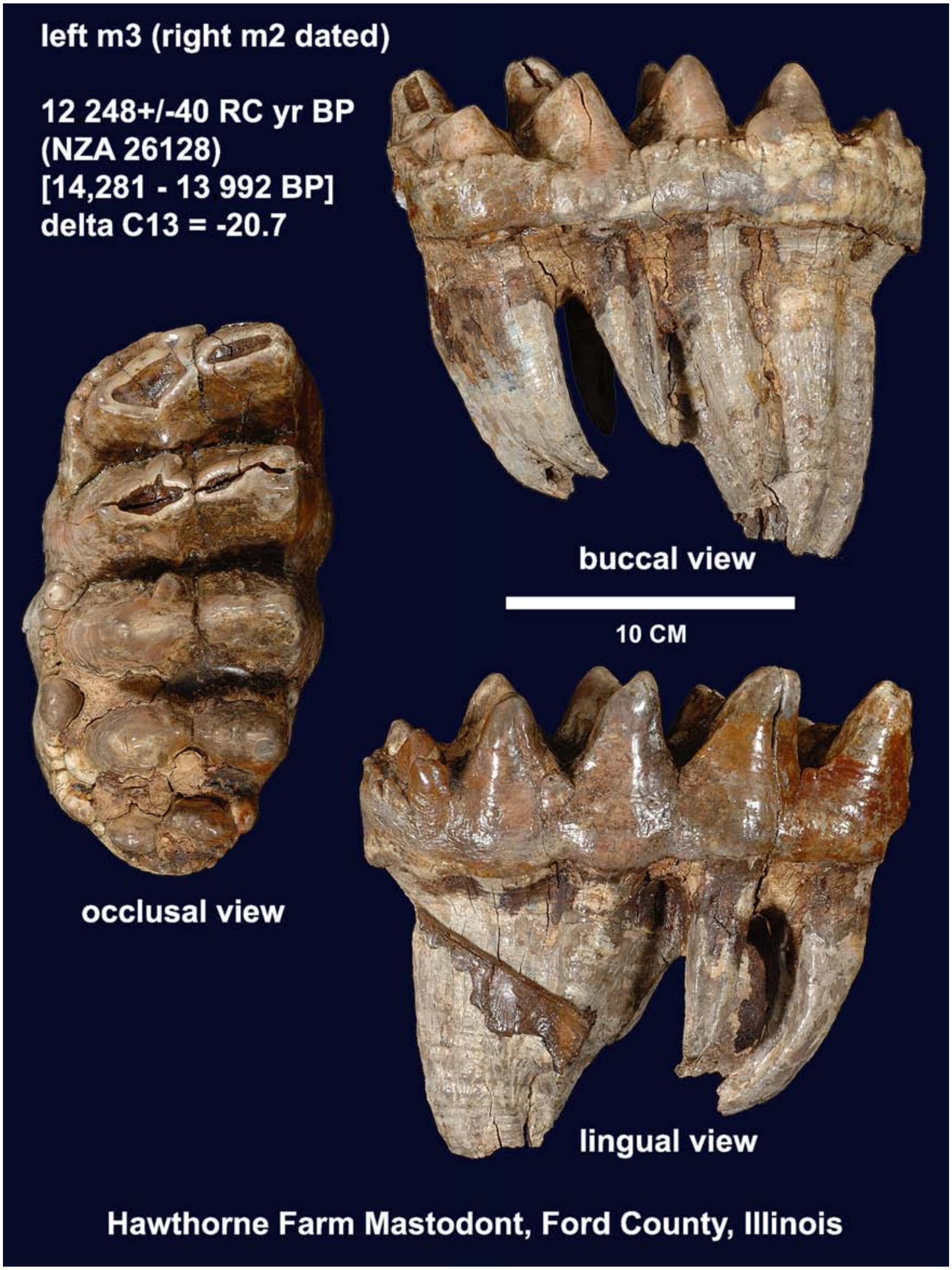


Figure 6 The left lower 3rd molar of the Hawthorne Farm Mastodont (*M. americanum*) from Ford County, recovered in 1956 and formerly in the collections of the University of Illinois Museum of Natural History that were transferred to the Illinois State Museum in 2005.
This was of dating interest because the tooth's small size (L = 191 mm, W [tritolophid] = 104 mm),and rugged cingulum on the buccal side of the crown were consistent with a previous model proposing a very late occurrence on these bases (King and Saunders 1984). But the date–12,248+40 C-14 yr BP–is relatively early.

Figure 7 The right lower second molar of the Wyanet Mammoth from Bureau County, Illinois. Reported recently by Pasenko and Schubert (2007), it is one of the few Illinois mammoths attributed in the literature to *Mammuthus jeffersonii*, the riparian or Jeffersonian species. An age of 11,170+140 C-14 yr BP was assigned to the tooth on the basis of loosely associated *Picea* wood. But the actual AMS dated age of the individual on the basis of dentine collagen–15,947+60 C-14 yr BP–was nearly 5000 years older, resulting in "a cautionary tale."

Figure 8 The left upper 3rd molar of the Painter Creek Mammoth, *Mammuthus primigenius*, from Hancock County, Illinois has been in the ISM collections since 1905. Expressing features of short length, wide width, thin enamel, and high ridge-plate compression the PCM is, like the LCM, clearly shown to be a woolly mammoth. The PCM was AMS dated to pursue the hypothesis that the appearance of the woolly mammoth south of 55 degrees N. latitude occurred only after the opening of the “ice free corridor.” The date–17,681+70 C-14 yr BP–demonstrated this hypothesis to be false.

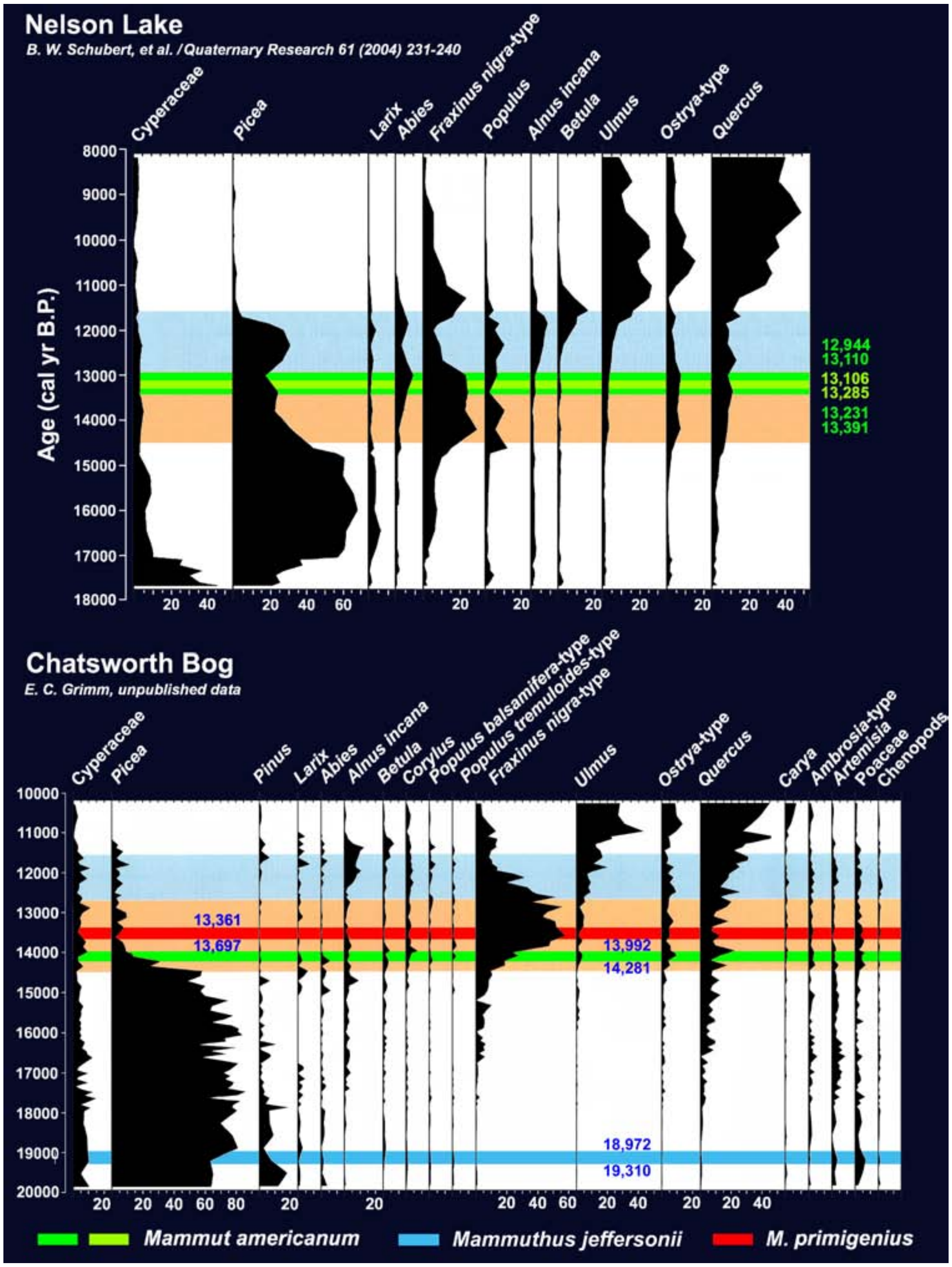


Figure 9 Pollen profiles and dated remains from a sediment core of Nelson Lake (above) with "pollen spectra" for the Brewster Creek and Aurora mastodonts, indicated in dark, light, and dark green, with calibrated dates shown at the right margin. The brown shading represents the Bølling/Allerød warm interval, the blue shading the Younger Dryas cold event. A mixed spruce (*Picea*), black ash (*Fraxinus nigra*), and oak (*Quercus*) forest is indicated for each mastodont represented.

Chatsworth Bog (below) with similar spectra for the Wyanet and Lincoln College mammoths and Hawthorne Farm mastodont.. When plotted against the dated pollen profile at Chatsworth Bog, the Hawthorne Farm mastodont is shown, like the former, to be occupying a forest of spruce, black ash, and oak during the Bølling warm interval.

The Lincoln College Mammoth is also shown to be occupying a forest during the succeeding Allerød warm interval, but a forest now with less spruce and dominated by black ash. Black ash represented at this percentage of relative abundance (55%) is non-analog today. It suggests, most conservatively, cool winters, warm and moist summers, and poorly drained soil. Less conservatively, the poorly drained soil supporting black ash in these amounts may have been developing in very wet mires. In any event a dry, arid and treeless landscape usually interpreted for *Mammuthus primigenius* is not indicated for the Lincoln College mammoth. Nor does this spectrum leave room for a refuge–a "locally buffered tundra-like opening"–in a regional forest. Rather it suggests that the woolly mammoth in the Southern Great Lakes, like recently shown for the Arctic fox in mid-latitude Europe (Dalén et al. 2007), did not track habitat, as the latter shifted northward.

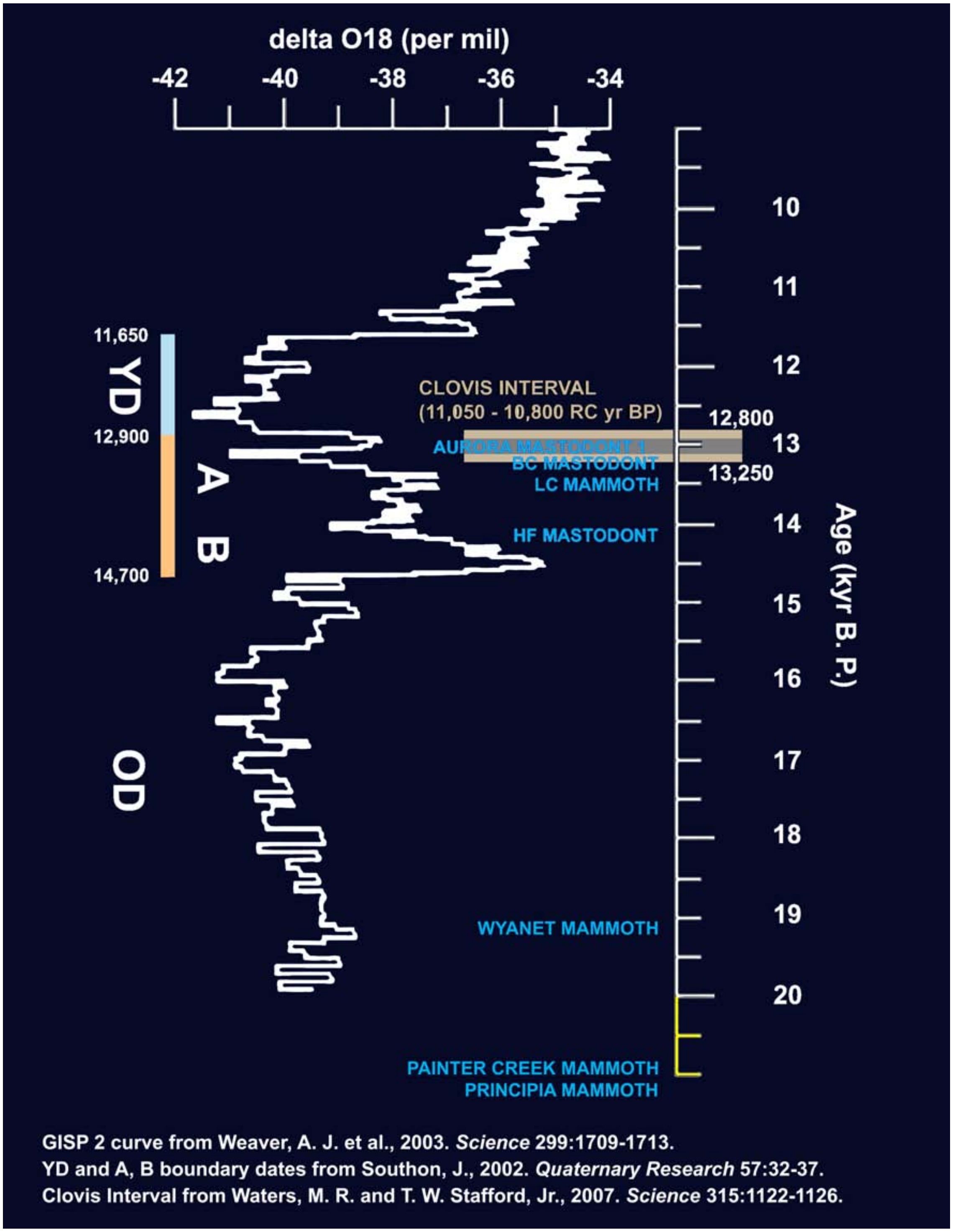


Figure 10 This slide elaborates on the former, plotting our dated sites along the GISP2 ice-core paleoclimatic time scale. It is noteworthy that the Brewster Creek and both Aurora mastodonts extend into the Clovis Interval reported by Waters and Stafford (2007). It is also noteworthy that Aurora Mastodont 1 extends nearly to the Allerod-Younger Dryas boundary, at 12,900 cal yr BP. This boundary coincides with the late Pleistocene extinction event, marking the final disappearance, through climatic change, human hunting, disease, or fireballs, of the North American (mainland) "Mammoth fauna."

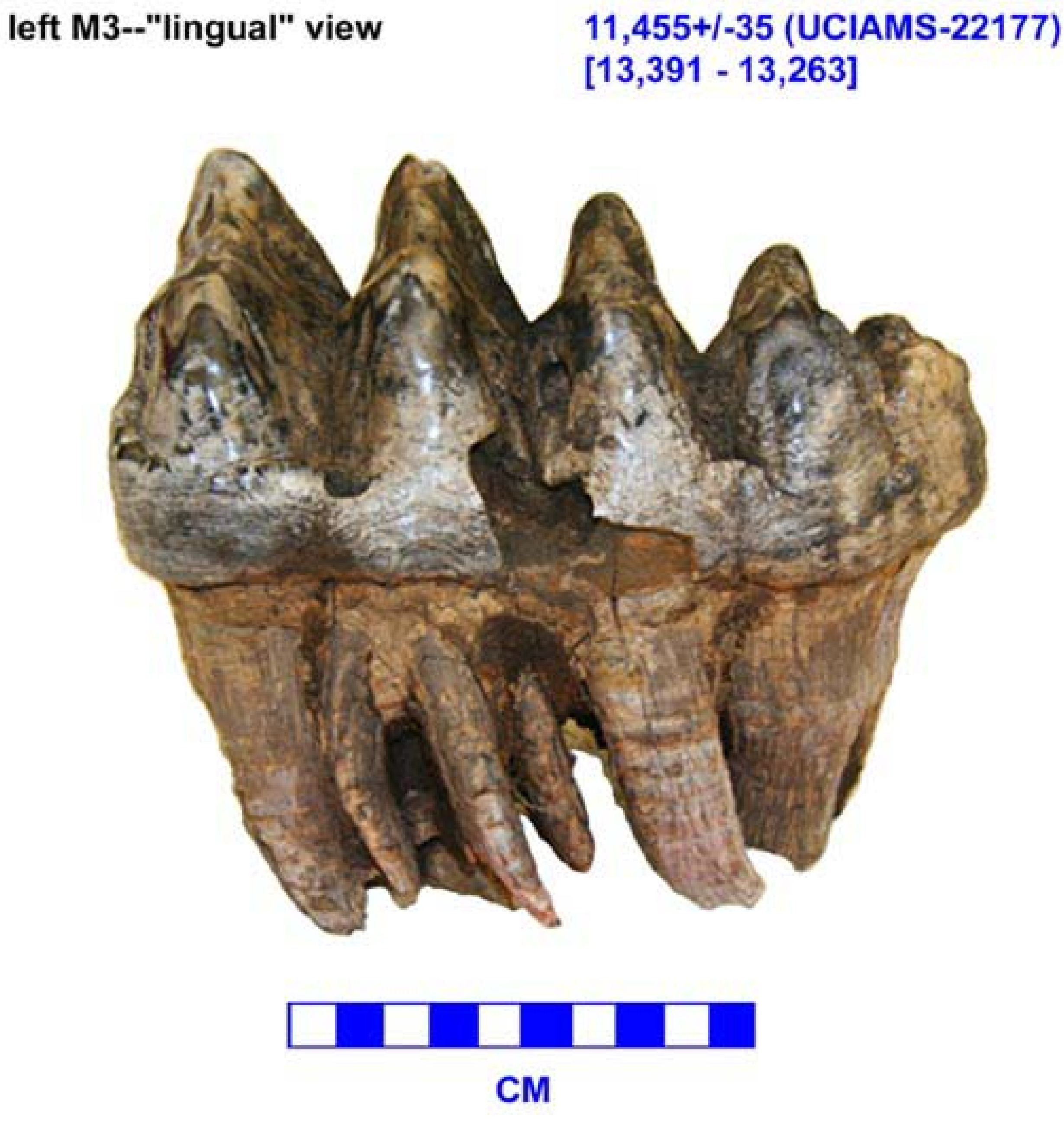


Figure 11 Brewster Creek Mastodont (*Mammut americanum*) DuPage County, Illinois

Aurora Mastodont Project 2004 – A successful outreach and educational experiment

David Voorhees

Background

In the late Spring of 2003, David Stover, the Mayor of Aurora, and Jim Pilmer, Director of Parks and Recreation for the City of Aurora (CoA) approached Waubonsee Community College (WCC) and its science faculty to see if they were up to a challenge.
In this meeting the Mayor described the 18 month Civil Works Administration (CWA) project that started in 1934 that consisted of 555 men with picks and shovels that were assigned to dig a municipal lake out of a swamp in Phillips Park, to be called Townsend Lake. During the initial months of the CWA project, the digging progressed without incident creating an island of spoil, presumably the north island of the islands in Mastodon Lake (Figure 12). In January of 1934, four mastodont bones (*Mammut americanum*) were discovered by CWA workman Joseph Gari, according to a Beacon News article. Work continued on digging the lake through 1935, recovering more mastodont bones. Ultimately three skulls, three tusks, a scapula, a mandible, three ribs, a femur, several articulated vertebrae and toes were recovered from the CWA project (on display in the Phillips Park Mastodon Gallery at this stop). A crude map was drawn showing the wide distribution of these finds, which is mainly along the eastern shore of Mastodon Lake (Figure 13). In addition to the mastodont remains, the right femur of a giant beaver (*Casteroides ohioensis*) and three pair of humeri from a trumpeter swan (*Cygnus buc-*

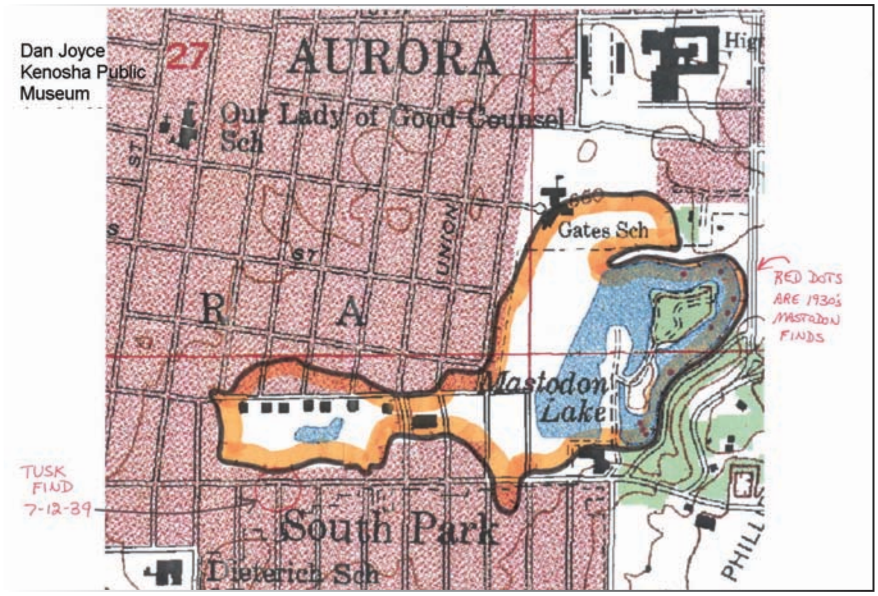


Figure 12 Map of the Phillips Park kettle showing Mastodon Lake. The 2004 AMP was along the eastern

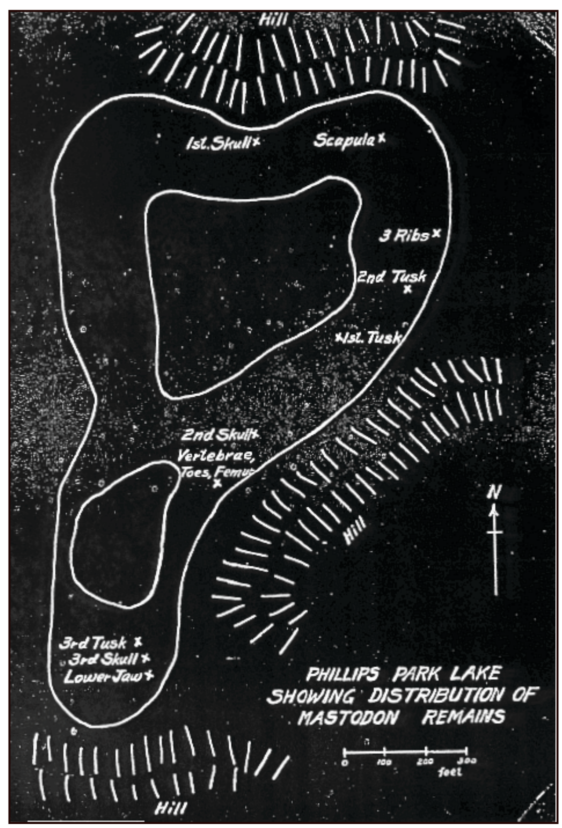


Figure 13 Map of the distribution of the CWA finds of 1934 – 1935 in Mastodon Lake

Figure 14 Aurora Mastodon Project (AMP) in June of 2004, showing some of the activity on the open squares with volunteers and students.

cinator) were recovered. These biologic remains were reportedly found in the upper three feet of gray marl that was overlain by two to five feet of peat, which in turn was overlain by two feet of black muck which was the bottom of the Modern swamp. The identification of the mastodon bones was confirmed by Professor Clarence R Smith of Aurora College (now Aurora University), as well as being of Pleistocene age. The trumpeter swan finds were identified by Professor L. A. Adams of the University of Illinois. Dr. F.C. Baker (UIUC) identified 20 and species of gastropods and pelecypods in the marl and proclaimed them consistent with an interpretation of cold-temperate climate of Middle Wisconsin age. This cold climate was supported by the occurrence of hemlock and spruce cones (*Tusga canadensis* and *Picea mariana*) in association with the mastodont remains (Powers, 1935; Smith, 1935). Remains of deer, elk and muskrat were found in the peat, and were probably of Modern age. The mastodont finds from the 1930's so enraptured the residents of Aurora at the time that the lake became known unofficially as 'Mastodon Lake', to which it is referred to today. These mastodont bones are now on permanent display in the Phillips Park Visitor Center and Mastodon Gallery.

During Mayor Stover's presentation he indicated that there is an oral tradition in Aurora that tells of mastodont bones that remain in Mastodon Lake. He then told the story of Ray Moses, who was one of the original CWA workers and subsequently a Superintendent of Phillips Park. Ray's son Ed, who was a boy during the CWA project, remembers his father telling of a fourth mastodont skull that was found near the end of the project. Since the CWA project monies were nearly gone, and time was short, it was reburied and left in the ground to be recovered by an interested scientific or educational agency. Other legends indicate that there is as much as a full skeleton still buried. Ed Moses also recalls that the fourth mastodont skull (or skeleton) was reburied near the 'big rock' along the eastern shore of Mastodon Lake (Figure 13). It was this (these) mastodont remains that Mayor Stover and Mr. Pilmer was challenging the science faculty of WCC to recover. All that were present at that meeting saw an exciting opportunity for scientific research, education, and collaboration.

The collaboration was started when WCC Earth Science / Geology Instructor David Voorhees contacted Dr. B. Brandon Curry of the Illinois State Geological Survey (ISGS) and Dr. Jeffrey Saunders of the Illinois State Museum (ISM). Both Curry and Saunders were familiar with the Aurora mastodonts and agreed to the project's potential. A budget was drawn up and presented to the CoA, who provided WCC and the ISM with a $60,000 grant to initiate, organize, and conduct an archeologically based excavation of the rumored mastodont remains, thus beginning the 2004 Aurora Mastodont Project (AMP). All four agencies were hopeful of encouraging community participation, creating an active learning environment and a model that would encourage all participants to engage in the scientific process (Figure 14). The ISM provided the paleontological and archeological expertise, WCC provided the educational and outreach expertise, the CoA provided significant logistical expertise, and the ISGS provided the geological expertise. The ISM also was able to provide a major contribution to the project by selecting Jim Oliver as the Field Director of the 2004 AMP.

Outreach activities

The 2004 AMP could not have been as successful as it was without the army of dedicated and enthusiastic volunteers, many of which returned countless times. A total of 212 volunteers participated during the 11 weeks of the project. Volunteers were asked to sign up for a minimum of 5 days, either in a row or in consecutive weeks. The volunteers were given the choice of working in the morning, the afternoon, or both. Volunteers were able to sign up by calling the Program Development and Alternative Learning Office of WCC (a total of 96) or by filling out a response window on the official AMP website (a total of 116). Upon receipt of the volunteer's response, a confirmation letter was sent out to each volunteer, listing the days and times they signed up for. Included in this confirmation letter were release forms which had to be signed prior to participating in the dig. There were on average 10 to 13 volunteers every day participating in the dig.

The experience and motivation for participation of the volunteers varied from a general interest in science and archeology (*viz.*, paleontology), to wanting to be a part of the same project that their grandfathers, who were part of the original CWA project. The ability to participate in a professionally run archeological dig in their own backyard was also irresistible to many. One of these volunteers expressed her thoughts on the Volunteers log of the dig website (dig.waubonsee.edu):

> *My grandfather was one of the hundreds who worked to dig Mastodon Lake. Somewhere there is a photo of him at the site leaning on a shovel. Thanks to Francis I now have a picture of myself at the dig leaning on a shovel. Someday, I hope to place those two photos side by side.*
>
> *I learned how to systematically trowel and level a square ten centimeters at a time. I practiced how to measure for level and experienced the frustration of water seeping, trickling, and/or flowing into my square. I encountered broken bottles, rocks, twigs, a mussel shell, and marl. I felt an ownership for square C14. After that week in June, it was my square.*

When I came back for one last day in July, Jim sent me to C4. There I did it all again, but this time I maneuvered around several pieces of spruce, trying my best to get rid of water, level the square, and protect those fragile pieces of history.

Although there is disappointment that the elusive skull has not been found, many people have been given the opportunity to learn scientific methods and participate in a historical undertaking. What a fantastic experience it has been for all of us! If the dig is reopened next summer, I will be back, and I will be joined by many of my fellow volunteers. We are hooked and someday a part of a mastodon will be recovered from the area. Until then I can only express my gratitude to Jim, Dave, Jane, and the many great people I worked with for allowing me to participate in this remarkable experience.

Many of the volunteers spent many days at the dig, performing countless hours of mundane, dirty, and physically difficult tasks that are the nature of scientific research. Most were surprised to learn how physically demanding the 'dirty side of science' can really be. The first phase of the excavation involved removing the Modern muck from the bottom of the newly drained lake bottom (Figure 15). The process was to fill a 5 gallon bucket of mud, haul it to the bank of the lake, dump it in a wheel-barrow, and then cart the wheel-barrow to the dump pile. It was this first phase that was a true eye-opener, which inspired one student enrolled in the WCC Special Topics class to pen the "Life cycle of mud" in her field notebook (Figure 16).

Figure 15 Removal of Modern muck behind newly placed coffer dam

Although an initial motivation for volunteering may have been an interest in science, the development of new friends and re-acquainting with old friends often became a secondary motivation. The excitement of actively engaging in the scientific process with new-found friends became so strong with one volunteer, Anita Weber,

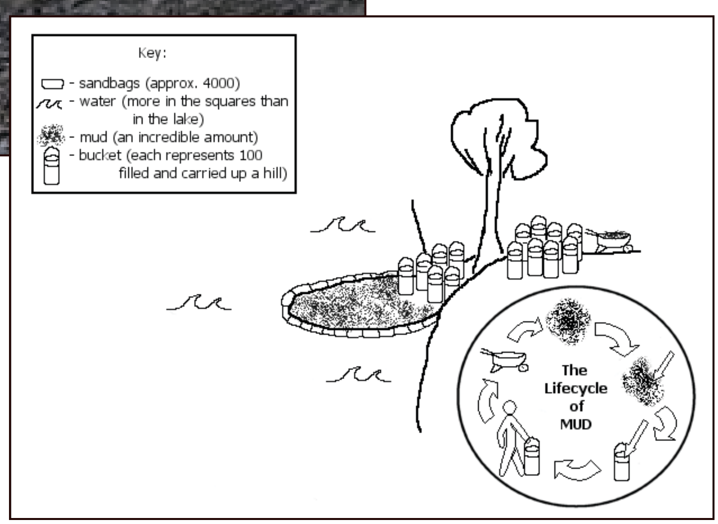


Figure 16 Sketch from a student's field notebook about removing Modern muck.

Figure 17 Field leader Anita Weber demonstrating screen washing.

that she returned each and every day for the entire 11 weeks, after initially volunteering for only 4 days. Her talents and enthusiasm naturally led her to assume a critical role as she supervised and organized the screen wash station for the entire summer (Figure 17). Some of her thoughts at the end of the AMP which she posted to the dig website:

> *I have so many good memories about the dig it is hard to narrow them down to fit on this page but I will try. I knew when I first heard about the dig I needed to get involved. And I am glad I did. I have met some really great people. Everyone was upbeat and optimistic the whole time. One day that stands out was when we got a good afternoon shower and I came down from the screen wash area to help cover the dig sites. Doreen, Jim and I had to spread the tarps over the sites so they wouldn't get too wet. We were sliding all over the place. Doreen and I were laughing through the whole process. Then we started packing up the equipment. I asked Doreen if she was having a good time, Which she answered with, "Yes, and are you?" I told her I was having a blast! Jim walked up and said, "You guys are sick!"*
>
> *I enjoyed our 'picnics in the park' when we could get together and talk about the progress of the dig, what we did over the weekend or what we had plans for the next day. This is probably where we learned more about what we did individually (in our real lives). Because I spend a majority of my time at the screen wash area, I didn't have a lot of time to 'socialize' with the dig crew so I looked forward to lunch.*
>
> *I was lucky enough to have both my mother and my daughter help with the project. My mom came out two to three days a week. She really enjoyed the experience. We had a lot of fun, even if she did have to keep me in line sometimes. My daughter came out for one week and she can't wait for school to start so she can share this experience with them. She is already working on her Science Fair project that is due in March 2005.*

At the end of the day sometimes Jim and Dave would come up to screen wash. Dave would make me laugh because he was always looking for a bone and found wood. When I would show him a bone, he would ask if I was sure. I think it was in the last two weeks when he finally found a bone all by himself. He was so happy!

Of course there were the water fights. You can't have eight hoses hooked up to a fire hydrant and not spray someone once in a while. You better make sure if you point a hose at someone, you have room to run.

I remember when we had a backlog of buckets in the garage and I was trying desperately to get them done. I could bring about nine buckets at a time up to the screen wash area on the club cart. I would just get started on dumping them when I would hear this sound. It was the sound of the red truck backing up to drop off a load of buckets. About 25 to 30 per load. I would dread this. I had dreams about it. I was haunted by red trucks. Then in the later weeks, when we were caught up, I saw the truck as 'job security'. It meant I still had something to do. The digging slowed and I could usually get the buckets up from the dig site throughout the day.

Now that it is over, I miss my friends, I miss the picnics in the park, my tan is fading, and I really want to see that red truck back up with another load of buckets. I guess I will have to wait until next year.

A boy scout's memories posted to the website:

I remember how hot and muddy the site was. I was one of the boy scouts working the screens, shaking them back and forth, looking for mastodon bones. Once I shook the screen too hard, fell backwards and nearly landed in the lake! That really made me laugh.

I learned how to bag and tag any item found and how important it is to document and about accuracy. I had fun being part of the group for this dig and most of all working alongside with my Dad, Terry Morrison.

The WCC Continuing and Professional Education Department created a course for (15) K-12 teachers that provided exposure to the scientific process of an archeological dig for 15 CPDU credits after participating at the AMP for three days. The WCC Community Education Department offered several *Paleontologist For A Day* programs for (89) K-8 students, and the *Bone Up On Mastodon's*, a four day course for grades 3-5 for10 students. Highlights of activities in the *Paleontologist for a Day Program* included making a glacier out of crushed ice from a snow cone machine, a special tour of dig site, and screen washing (Figure 18), where the Junior Paleontologists were able to actively participate in the scientific research of the dig, and get wet at the same time. The goal was to send them home happy, tired & covered in mud!

WCC's video production team created a video that was shown throughout the district on educational television to inform the community of the project and to recruit volunteers. WCC's Public Relations Department also distributed the video and press releases to television stations and newspapers. This began a flurry of press coverage, in which the AMP was ultimately featured in 35 newspaper articles, 3 radio interviews, and 5 television programs. In addition, WCC also created a comprehensive and interactive web page (http://dig.waubonsee.edu) to communicate the history of the site, the science and the scientists, and allow visitors to monitor the progress by viewing the Field Director's field log, and also to post their own impressions of the project. As of May 2008, this website is still live.

Figure 18 *Paleontologist For A Day* participants screenwashing matrix from Mastodon Lake.

Perhaps the most common outreach activity came after the common cry from a passer-by in the Park or a passenger on the Phillips Park Mastodon Express tram (Figure 19) of “Have ya found anything yet?”. To which we would always answer, “No, but…”, which is when we would use this ‘teaching moment’ to give the questioner an impromptu lecture on the scientific importance of what we were doing, what we had found, as well as general Pleistocene geology. These informal lectures were repeated throughout the summer to an estimated 2400 visitors to Phillips Park and the digsite.

Figure 19 Phillips Park Mastodon Express tram on one of its daily trips to the AMP digsite.

The WCC Special Topics class

WCC offered two sections of a credit class taught by David Voorhees entitled *ESC 296 Special Topics: Mastodont Excavation.* The first section was a four credit class that began on 18 May, and the second section was a three credit class that began on 12 June. The three credit class was scheduled to begin with the end of the semester of the surrounding Middle and High Schools to allow any interested science teachers to enroll. Both sections had one credit of lecture, and the remaining credits as 'laboratory' at the Phillips Park dig site.

A total of 15 students were enrolled in both sections, 10 in the four credit class and 5 in the three credit (none were teachers); 7 were present or future WCC students, 4 were high school (and 1 home-schooled), 3 were out of the WCC district, and one was a returning student who was a former student of Professor Clarence Smith of Aurora College. Of the 15 students registered for the 2 classes, 9 expressed interest in geology, paleontology or archeology, thus providing a pivotal event in their educational and scientific careers. As experienced educators are well aware, to teach is to know, and the students enrolled in the class were at the digsite enough so that they themselves became teachers to the new volunteers that showed up daily.

The one credit lecture component consisted of 16 hours of class time on the topics listed in Table 1, and consisted 25% of the students' grade for the course. Topics 9, 10 and 15 were presented by guest lecturers Dr. B. Brandon Curry (ISGS), Dan Ward (WCC), and Jim Oliver (ISM), respectively (Figure 20). The required texts for the course were Killey (1998) and Lange (2002). Upon completion of the lecture component of the class, the students were given a take-home essay exam and given 1 week to answer the ten questions. The essays that were submitted were of a quality higher than I expected, based upon my experience as a Community College science instructor. The combination of the experiential learning and active engagement at the dig site significantly reinforced the classroom material much more than routine library study.

Figure 20 Jim Oliver (ISM) discussing donated cow bones used in a taphonomy experiment using the wolves at the Phillips Park Zoo as part of the WCC Special Topics class.

The laboratory component comprised 75% of the students' final grade; 40% for the evaluation of the student's field notebook and 35% for a final report. The "labs" met immediately after the lecture, and were for 3 or 4 hours (depending upon whether the student was registered for the 3 or 4 credit class), during which the students learned about the activities of a paleontologist/geologist/archeologist during a ‘field season’ or dig. Tasks assigned to the students included those of the volunteers (digging in assigned squares, manning the screen wash), as well as sorting and cataloguing the recovered samples, and general geologic, stratigraphic and pedologic description of the walls of the squares and several test trenches. In both of these classes, the students immediately went from the classroom to the field with the newfound knowledge in geology, sedimentology, evolution, fossilization, mastodonts, and the glacial and paleoenvironmental history of Illinois.

The final report was to be written in the style of a paper publishable in the *Journal of Geoscience Education*, and the students were given a copy of their ‘Instructions to Authors’. In the final report the students were to answer 3 questions: (1) What did I learn about the geology, ecology, and biology of Phillips Park and the biologic components of Mastodon Lake?, (2) What techniques did I learn or observations did I make in the excavation process?, and (3) What are my comments and reflections of the class just completed?.

Several students commented on their experiences in the class as part of the third question of their final report. Several excerpts are provided below:

> ***High school student:*** *This class has given me some excellent resources, and a thorough background in field geology, paleontology, and archaeology, which will benefit me in the future to pursue a college degree and a career in paleontology. I have thoroughly enjoyed this experience and have learned a wealth of information in this class. This summer has helped me to confirm that I really do love field paleontology.*
>
> ***WCC student:*** *This class was an important experience for me as an aspiring archaeologist. I learned so much in this course that I never dreamed I would learn. Now I understand just how difficult it is to be an archaeologist, geologist or any kind of scientist for that matter. It is one thing hearing about it in a classroom and a whole new experience actually doing the excavating yourself.*
>
> ***Out of district Journalism major:*** *As I look back upon the summer, I realize I have learned a lot about scientific processes, paleontology, and the geology of Illinois – all without ever laying eyes on the fabled Mastodont of Phillips Park. Like a true paleontologist, I learned that not every excavation is successful. As I gazed across sites A and B on the last day of excavation, I saw the progress we made and enthusiasm we churned, even without finding what we came there to find. Even if all we produced was a large hole, we inspired people to visit the site and ask questions, to learn, and to think about the land we live on. Above ground, Aurora will bustle with the normal activity of modern Northern Illinois and a new generation of students and excavators has added a chapter to the legend of the Phillips Park Mastodont, while below ground, those mysterious bones await discovery for at least one more year.*
>
> ***WCC student:*** *As I reflect back on this course, I am very pleased with myself for taking it and gaining so much experience plus reigniting my passion for the monsters of the past. I have enjoyed an experience I wanted since I was 3 years old and that is working at a paleontological dig site. Why I have almost passed out in the heat, became ill in the cold rain, stung by ants, bitten by mosquitos and pestered by a sore body, but when I look at those ancient bones from big movers and shakers of the Pleistocene like the deer, buffalo, and carnivores I can’t help but smile and I’d do it all over again to enjoy that electrical excitement of holding in my very palm the jaw, tooth or vertebrae of an animal that actually was born, lived and died, not to mention shared life with or was one of the amazing monsters of the Illinois Pleistocene.*

High School student: *I enjoyed the dig and found it very worthwhile. I have always been interested in archaeology and being able to experience it firsthand was very beneficial, especially as I am at an age when considering what my future career will be is a main focus.*

I believe the lecture portion of the course was well taught. I feel very proficient in mastodonts and glacial history. It was challenging in the assignments, which made for a more thorough learning experience.

The actual excavation portion of the course was also challenging, but in a physical way. A few hours in the sun and carrying buckets can leave you exhausted. Exhaustion can make you sarcastic and cause you to create crude drawings (Figure G2). However, the knowledge gained in the mastodont dig far outweighs the physically tiring aspect. The experience itself outweighs the fact that the rumored mastodont skull has not been discovered.

Prior to the beginning of the actual dig at Mastodon Lake, an Honors student at WCC worked with Dr. Brandon Curry (ISGS) to learn about ostracodes and how to process cores for ostracode analyses. She extracted and studied the ostracodes in subsamples of cores taken as part of a reconnaissance study around Phillips Park. Her interpretation of the ostracode data suggested that Mastodon Lake was once a lake that eventually became a wetland.

Table 1 Topics covered in the lecture component of ESC 296

Number	Topic
1	Introduction to class and history of Phillips Park
2	How to be a Field geologist / archeologist
3	Minerals & rocks
4	Sedimentary rocks
5	Sedimentary facies
6	Glacial geology
7	Glacial sedimentology & geomorpholog
8	Illinois Quaternary history
9	Illinois Quaternary history
10	Evolution of *Mammut americanum*
11	Mastodont systematics & differences with other *Proboscideans*
12	Mastodont habitat & finds
13	Mastodont anatomy
14	Fossils & fossilization
15	Taphonomy

I would suggest that the Special Topics classes were extremely successful, as half of the students received an A, indicating very active involvement and participation. As an aside, I rarely have greater than about 10% of any of my traditional classes earning an A. The performance of the students in these classes clearly demonstrates the effectiveness of the active pedagogy of laboratories and field work in (geo)science education. In addition, a very effective component of the classes was application of concepts we talked about in the classroom almost immediately at the Phillips Park digsite. This course design provided the students with an understanding of the scientific rationale behind the seemingly menial and mundane tasks that incorporates much of the typical scientific research process.

Also, in these classes I was fortunate to develop a closer relationship with students that is not normally possible given the traditional semester schedule. In doing so, I was able to foster and develop individual interests of the students, especially those interested in the geosciences. I was truly impressed with the depth of interest in the geosciences, and in some cases, depth of knowledge, in students who may have otherwise gone unnoticed in the more traditional collegiate learning environment. These classes have also given me a profound sense of satisfaction to see the students becoming excited about a subject that I am personally passionate about, even (or especially) when they are covered in mud.

Conclusions

The 2004 AMP was an unqualified outreach and pedagogic success. The integration of outreach activities alongside formal classes was challenging, but extremely rewarding. Through the 2004 AMP, we were able to expose a significant portion of the community to the excitement and importance of science and geology at a time when the scientific literacy of the general public seems to be declining. Given the overwhelming response to the 2004 AMP, as well as the continued inquires four years later, there is a natural curiosity about science, geology and paleontology that is ready, willing, and able to be tapped and enriched, as long as, we as scientists and educators, can reach out to it.

Although it would be very difficult to replicate this initiative at other colleges, it is clear that this could serve as a model for the integration of multidisciplinary studies that would provide unique learning opportunities. Clearly, a direct replication would require the discovery of an extinct *Probiscidean*, although extremely rare, it is not unheard of. For example, the educational program developed by Dr. Janice Treworgy at Principia College (Treworgy, et al., 2006 and Treworgy, 2008) was used as a model for the formal classes in this project. Parts of the 2004 AMP could easily be replicated to study or answer a particular research question, which would integrate the classroom with the active learning in the field while collecting data to answer this question, using formally enrolled students or the general public. Applications of this model to other earth science and geology initiatives might include the intensive study of a watershed that is threatened with industrial or residential development.

I would like to thank Mr. Jim Pilmer and all of the Phillips Park employees who were extremely generous in giving of both their time and resources to the 2004 AMP. Their generosities were a significant contributor to the success of the project. There were also many other donors and contributors, too many to individually mention here, from local businesses and academia which made the 2004 AMP a little easier and productive.

Stop 6: LaFarge Sand and Gravel Pit

A buried catena of the Farmdale-Sangamon Geosol Complex, Elburn, Illinois

Peter Jacobs, Michael Konen, and Brandon Curry

Introduction

The Farmdale-Sangamon Geosol complex is historically significant because the last interglacial (LIG; Sangamon Episode) in North America was originally recognized by Leverett (1898) on the basis of what are now accepted as soil characteristics between till sheets (Follmer et al., 1979). Today, the Farmdale-Sangamon Geosol complex provides the most extensive paleopedological record of the last interglacial to glacial transition in the Midwest United States (Follmer, 1983; Curry and Follmer, 1992). The Sangamon Geosol forms the lower part, and the Farmdale Geosol, the upper part of the pedocomplex.

The type area of the Sangamon Geosol is in central Illinois, just beyond deposits marking the limits of Marine Isotope Stage (MIS) 2 glaciation, where the LIG landscape is preserved intact, buried by MIS 2 loess deposits (Follmer, 1978). Beyond the type area, the Sangamon Geosol is widely distributed across the Midwest and Great Plains, especially outside the limits of MIS 2 glaciation. Most geochronologic studies indicate that this geosol developed throughout MIS 5 and 4, continuing into MIS 3 in many cases (Curry and Pavich, 1996; Markewich et al., 1998; Karlstrom et al., 2007; Mason et al., 2007).

Profiles of the Sangamon Geosol occur in two genetic types: *in situ* and accretion gley (Follmer, 1978). *In situ* profiles occur most frequently (Willman, 1979), and are those profiles that formed in place by weathering and pedogenic reorganization of a geologic material such as till, outwash, or loess. The weathering and morphological characteristics vary with drainage status of the paleoprofile: brown, oxidized colors are typical on uplands or topographic highs, while dark gray colors prevail under poor drainage conditions. Nearly all *in situ* profiles are texture-contrast profiles with distinct horizonation. Most profiles are believed to include eluvial upper sola and illuvial (argillic) B horizons, however the texture contrast has been accentuated in most profiles by relatively slow aggradation and bioturbation of eolian silt, mainly during MIS 3 (Follmer, 1983). Known as the Roxana Silt, this loess was incorporated into the profile and led to the development of a very thick and silty upper solum with pedogenic characteristics associated with cooling climatic conditions during MIS 3. Only in sites within about 30 km of major outwash sluiceways was the loess accumulated rapidly enough so that sola of the Sangamon and Farmdale Geosols are separated by relatively unweathered loessial material, in some places with thin A/C couplets (Curry and Follmer, 1992). Profiles that display the aggradational genetic profile, normally with a relatively thin increment of Roxana Silt, formed partly in sediment deposited during MIS 3 are referred to as the Farmdale-Sangamon Geosol pedocomplex.

In situ profiles resemble modern Alfisols or Ultisols, and the degree of morphological expression varies northward across the type area (Follmer, 1983). In addition, *in situ* profiles across the Midwest are leached of carbonates, often to several meters, and the degree of mineral alteration increases with coarser parent material grain size (Brophy, 1959; Jacobs 1998a).

As a formal pedostratigraphic unit, the designation of a type catena (i.e., composite stratotype) is necessary to document and define the range in pedological characteristics (i.e., pedofacies) of the Sangamon Geosol. Follmer et al. (1990) designated exposures in a limestone quarry at Athens, Illinois as the type locality because of the accessibility, quality, and range of profile features exposed, from well drained *in situ* to accretion gley profiles. Other studies in the mid-continent demonstrated that variations in paleodrainage conditions and parent materials contribute to significant differences in profile morphology, weathering trends, and post burial modification (Hall, 1999; Jacobs, 1998b).

Here we report on the paleoenvironmental record of landscape and soil forming processes preserved in a paleo-hillslope that forms a catena of *in situ* profiles of the Farmdale-Sangamon Geosol complex that is preserved under Wisconsin Episode glacial sediments. Specifically, we investigate the effect of topography on morphological expression and weathering characteristics over the LIG to glacial transition. Our findings indicate that interglacial-scale pedogenesis along a hillslope is best associated with drainage-related morphological differences in profiles of both geosols, but the distinction is most pronounced in the Farmdale Geosol. Clay mineral and horizonation patterns appear be more strongly controlled by variations in parent material grain size and its stratification than topography.

Study Site

Location and significance

The catena we studied is located in a large gravel mine owned by Lafarge, Inc. (formerly known as the Feltes Sand and Gravel Company) south of Elburn, Illinois (Figure 1). The pit contains a thick sequence of Illinois and Wisconsin Episode outwash and diamicton (Figure 2; Curry et al., 1999). The site is located just west of the Elgin Bedrock Valley, a tributary of the St. Charles Bedrock Valley; both features contain important aquifers (Curry and Seaber, 2000; Curry et al., 2002; Dey et al., 2007a). This site is significant because it is among the most northerly catenas of the Sangamon-Farmdale complex, located 150 km north of the type area at Athens Quarry.

Stratigraphy and Age

The stratigraphic units exposed in our catena include a coarsening upwards > 2 m thick bed of dolomite-rich, cobbly sand and gravel capped by < 0.5 m of loamy diamicton of the Glasford Formation (MIS 6; Illinois Episode). The coarse texture and locally-derived dolomite of the Illinois Episode deposits probably reflect ice marginal deposition. The Sangamon Geosol is formed in these coarse-grained deposits, whereas the Farmdale Geosol formed in an eolian and colluvial silt (Robein Silt). *In situ* stumps that once grew in the Robein Silt yielded radiocarbon ages ranging from 24,000 to 25,820 C-14 yr BP (see Radiocarbon Table, p. 43). These trees were killed when they were inundated by a proglacial lake dammed by ice or sediment associated with the Lake Michigan Lobe during the last glaciation. The lake sediment (Peddicord Tongue, Equality Formation) protected the catena from erosion by Wisconsin Episode glacier ice and meltwater that ultimately deposited about 10 m of sediments.

Methods

The contacts between depositional and pedostratigraphic units exposed along the paleo-hillslope sequence were surveyed and mapped along a 40 m transect. Four soil profiles were described and sampled in detail to assess slope effects on Farmdale-Sangamon pedogenesis.

In the field, the elevation of soil horizon and stratigraphic breaks were all surveyed from a common bench mark using a theodolite and stadia rod. Because of modest slump and wash of the excavation face, we chose to survey profiles at even intervals where we could scrape the surface of the profile clean. For the profiles selected for detailed analysis, we excavated back into the face several decimeters to a meter or more. Soils were described in the field in pedologic detail using NRCS terminology (Soil Survey Staff, 1993). Samples were collected from each

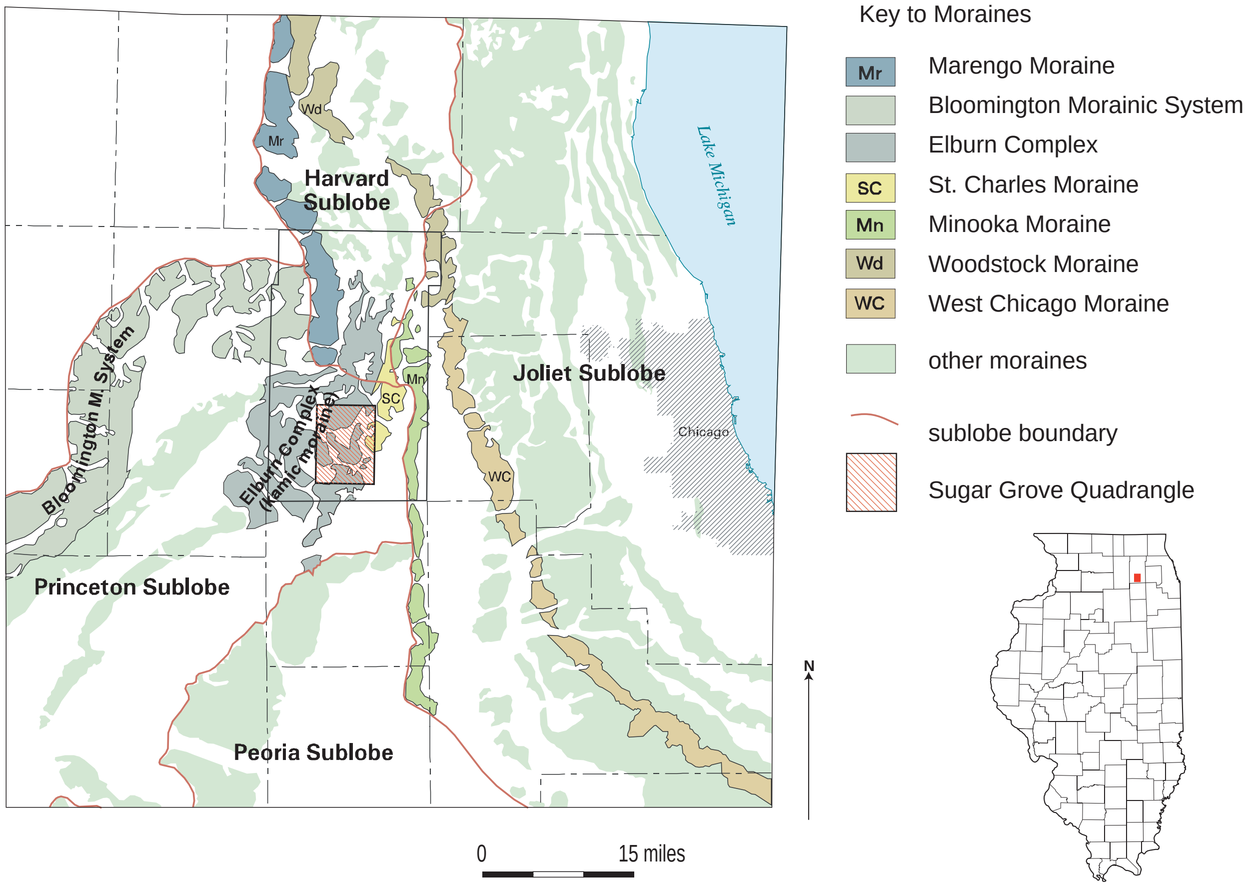


Figure 1 Major moraines of the last glaciation, northeastern Illinois (after Willman and Frye, 1970; Hansel and Johnson, 1996). The location of the Sugar Grove 7.5-Minute Quadrangle is shown; the LaFarge Pit is located in the northwestern corner of the map

genetic horizon described, and horizons thicker than 20 cm were subdivided to maintain close-interval sampling. In addition to bulk samples for laboratory analysis, each horizon was sampled for bulk density using a metal tube (Blake and Hartge, 1986), and in selected horizons intact oriented samples were collected for thin sections.

Laboratory analyses presented here include particle size analysis determined by wet sieving and pipet (Kilmer and Alexander, 1949). Organic carbon (OC) was determined by loss on ignition (Konen et al., 2002) and total C and N using a CNH analyzer. Inorganic carbon, believed to be mostly calcium and magnesium carbonate was calculated by difference.

Semi-quantitative mineralogy of the < 2−µm size fraction was determined by X-ray diffraction using ethylene glycol-solvated aggregate slides and a Scintag Model XPH-103 diffractometer. Peak heights for common phyllosilicates, carbonates, and silicates in the clay were measured at fixed 2Θ positions. Relative abundance was corrected by the peak intensity factors in Hughes et al. (1994). The relative proportion of two mineral suites was determined: 1) clay minerals (e.g., expandable clays, illite, kaolinite, and chlorite) and 2) dolomite plus the silicate minerals quartz, potassium feldspar, and plagioclase. The percentage of kaolinite and chlorite was determined by calculating the ratio of the intensity (in counts per second) of the 2^{nd}−order kaolinite peak at 24.9º 2Θ versus the 4^{th}−order chlorite peak at 25.1º 2Θ and applying that ratio to the relative intensity of the compound 1^{st} −order kaolinite/2^{nd} −order chlorite peak at 12.4º 2Θ.

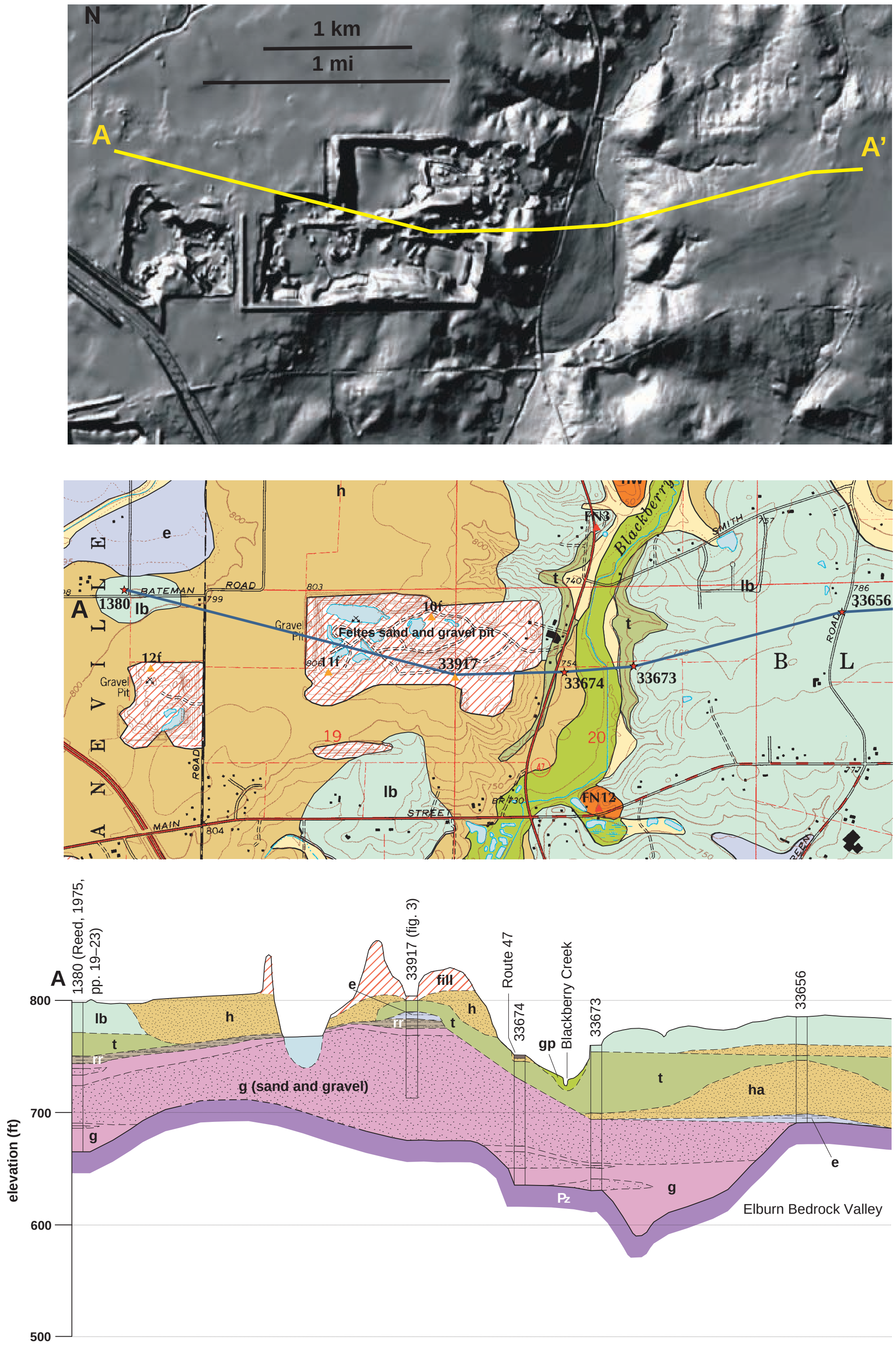


Figure 2 The LaFarge sand and gravel pit, Elburn, Illinois. (A) Shaded relief map of 10-m DEM (after McGarry, 2000) showing line of section for cross section A-A'. (B) Portion of surficial geology map of the Sugar Grove Quadrangle (Curry et al., 2002). (C) Cross section A-A' (Curry et al., 2002). For the surficial geology map and cross section, Pz = Paleozoic bedrock, undifferentiated, g = Glasford Formation, e = Equality Formation, ha = Ashmore Tongue, Henry Formation, t = Tiskilwa Formation, and lb = Batestown Member, Lemont Formation.

Oriented thin sections, 25 x 45 mm and 30 µm thick were prepared by Spectrum Petrographics, Inc., Winston, Oregon. Micromorphological observations were made to interpret soil forming processes and environmental history and are largely non-quantitative. Terminology follows that in Stoops (2003).

Results

Hillslope topographic profile:

Topographic relief on the Sangamon paleosurface was approximately 0.75 m along the 40 m transect (Figure 3). The western end of the transect is lowest and is quite flat until about 17 m east of the starting point. Between 17 and 24 m is a backslope that rises 0.64 m, a 9% slope, to a nearly flat surface on the eastern end of the transect. Profile 23 is located on the flat topographic low and is herein referred to as the low profile. The middle profile, Profile 20, is located on the backslope. The upper profile, Profile 17, is located on the higher flat surface. Profile 18A is located at the head of a gulley cut into the high flat surface near Profile 17 and was used for some morphological measurements and for collection of thin sections. Recognition of the top of the Sangamon paleogeomorphic surface was based on characteristics such as the sand-silt ratio, coarse fragments, color, and mesoscale soil fabric (klumpen fabric of Follmer, 1998), and is supported by laboratory analyses. In contrast, total topographic relief on the Farmdale paleosurface is only 0.45 m. In nearly all cases there was no indication of erosion of the Farmdale surface and contact with the overlying Equality Formation lake sediment is apparently conformable, based on horizonation characteristics such as the darkest topsoil colors and the occurrence of woody detritus.

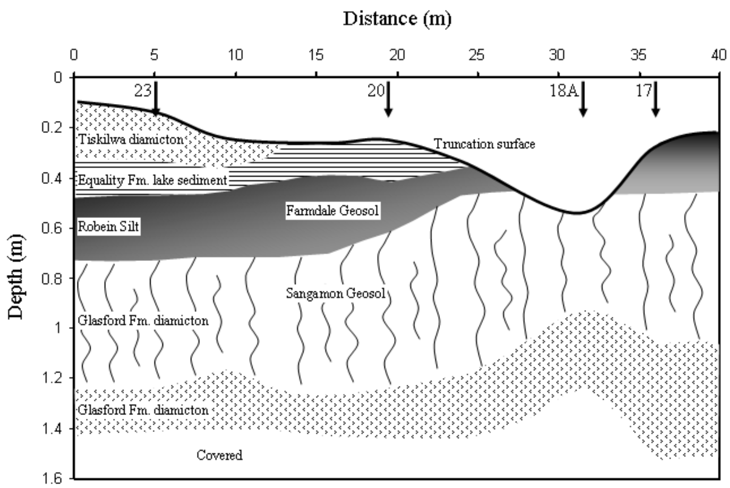


Figure 3 Schematic cross section of paleo-hillslope exposure at LaFarge Sand and Gravel showing stratigraphy and soil geomorphology. Location and identification of profiles is shown at arrows.

The overall topography of the study profile is an irregular surface that is attributed to sedimentological processes associated with ice stagnation. The actual mode of hillslope formation is not clear from the outcrop, but since we observed no faulting or deformation in the Sangamon Geosol or underlying sediment, we believe the irregular surface topography originated from debris flow rather than collapse caused by melting of buried ice.

Physical, chemical, and mineralogical characteristics of the sedimentary units

Analytical characteristics of each lithologic unit and soil horizon are reported in Table 1. The calcareous Glasford Formation diamicton (soil C horizons) is loamy, with 50-60% sand, 30-40% silt, and 10-15% clay in the fine earth fraction. Coarse fragments are mostly dolostone and limestone and are greater than 20% of the bulk soil material. Due to small number of samples, variation in grain size characteristics along the hillslope can not be tested statistically, but we did observe less sand in topographic lows, which may reflect the deposition processes that created the original topography. Clay content is relatively invariable, so as sand content decreases, silt

Table 1 Soil horizonation and depths, along with selected physical and chemical data of profiles at the La-Farge Sand and Gravel pit.

Horizon	Upper depth (cm)	Lower depth (cm)	Gravel %	Sand % 2-0.053mm	Silt % 0.053-0.002mm	Clay % <0.002 mm	USDA Texture Class	Geometric Mean 2-0.002 mm fraction (μm)	Bulk density whole soil (g/cm³)	pH 1:1 H_2O	carbon %	carbon %	Sample No
Profile 17													
Till	+10	0	12.6	55.5	31.0	13.5	sl	79		7.9	3.26		1565
Ab1	0	17	0.3	23.8	66.9	9.3	sil	24	1.74	7.7	0.21	0.26	1566
	18	35	0.4	32.3	59.0	8.8	sil	32	1.71	7.6	0.18	0.30	1567
Ab2	35	42	1.1	40.6	51.9	7.5	sil	40	1.75	7.5	0.18	0.19	1568
	42	50	1.1	41.9	48.4	9.6	l	43	1.82	7.4	0.12	0.16	1569
2AEb	50	60	0.5	42.5	47.1	10.4	l	45	1.78	7.4	0.12	0.22	1570
	60	73	14.6	52.1	34.0	13.9	sl	71	1.72	7.4	0.09	0.20	1571
2Btb1	73	82	18.7	49.5	32.1	18.3	l	71	1.63	7.3	0.13	0.39	1572
2Btb2	82	102	10.2	46.1	29.3	24.6	l	67	1.71	7.4	0.18	0.76	1573
2Btb3	102	115	11.7	49.8	28.7	21.5	l	71	1.68	7.4	0.21	0.53	1574
	115	135	12.8	62.9	21.4	15.7	sl	102	1.74	7.3	0.15	0.50	1575
2Btb4	135	154	3.2	65.4	16.4	18.2	sl	111	1.73	7.4	0.14	0.40	1576
2Btb5	154	169	3.0	41.7	33.5	24.8	l	54	1.82	7.5	0.17	0.40	1577
	169	187	6.7	40.2	33.1	26.8	l	54	1.72	7.4	0.17	0.57	1578
2BCtb	187	209	1.5	19.9	58.5	21.6	sil	20	1.64	7.5	0.16	0.39	1579
2CBtb	209	220	41.3	59.2	28.9	11.9	sl	104		7.4	6.61		1580
2C1	220	245	26.8	58.7	32.8	8.5	sl	96	1.87	7.6	7.29	0.09	1581
2C2	245	285	31.1	54.2	35.5	10.3	sl	78	1.83	7.9	7.64	0.07	1582
Profile 18A													
Ab1	0	14	0.1	15.7	69.8	14.5	sil	20	1.59	8.3	0.61	0.68	1730
Ab2	14	24	0.1	15.7	72.6	11.7	sil	20	1.72	8.2	0.41	0.42	1731
ABgb	24	39	0.1	26.2	62.5	11.3	sil	29	1.85	7.7	0.32	0.31	1732
Bgb	39	49	0.1	36.3	53.6	10.1	sil	37	1.97	7.6	0.25	0.18	1733
2AEb	49	55	1.5	44.4	46.6	9.0	l	50		7.7	0.13	0.26	1734
	55	63	2.3	48.1	43.5	8.3	l	54	1.95	7.7	0.14	0.20	1735
	63	73	3.5	51.8	37.9	10.3	l	63	1.90	7.6	0.15		1736
2EAb or 2Eb	73	88	24.2	55.0	34.4	10.7	sl	70	1.66	7.8	0.10	0.35	1737
2BEtb	88	98	6.8	39.9	34.4	25.7	l	53		7.8	0.17		1738
2BEtb	98	118	4.7	50.1	18.0	31.9	scl	123			0.24		1756
2Btb2	147	185	1.2	36.9	30.0	33.1	cl	58			0.22		1757
Profile 20													
Lacustrine	+10	0	0.0	3.0	83.6	13.4	sil	15			3.12		1740
Ab1	0	12	0.0	10.7	82.0	7.3	si	17	1.46		3.73		1741
Ab2	12	21	0.1	14.9	71.6	13.5	sil	21			0.80	1.08	1742
Bgb1	21	36	0.1	20.4	68.9	10.6	sil	25	1.89		0.40	0.25	1743
Bgb2	36	54	0.2	34.9	53.9	11.2	sil	39	1.83		0.26	0.21	1744
Bgb3	54	68	0.1	42.8	44.7	12.5	l	50	1.89		0.24	0.24	1745
AEb	68	88	2.2	47.0	40.7	12.3	l	58	1.83		0.15	0.17	1746
2BEtb	88	101	6.2	45.5	32.4	22.1	l	68	1.73		0.22	0.41	1747
2Btb1	101	133	6.9	40.8	22.7	36.5	cl	91	1.69		0.29	0.85	1748
2Btb2	133	158	4.6	41.1	27.2	31.8	cl	71	1.74		0.19	0.61	1749
2Btb3	158		3.0	45.6	25.5	28.9	scl	74	1.69		0.18	0.70	1750
		212	4.0	37.5	27.8	34.7	cl	66			0.19		1751
2BCtb	212	249	15.0	49.0	32.5	18.5	l	74	1.96		3.67	0.17	1752
2CBtb1	249	265	32.5	49.9	32.5	17.7	l	73	1.74		5.35	0.36	1753
3CBtb2	265	289	1.1	23.4	54.2	22.4	sil	26			3.51		1754
4C	289	289+	45.1	57.5	30.8	11.7	sl	108			7.22		1755
Profile 23													
Lacustrine	+50	0	0.0	5.5	46.9	47.6	sic	8	1.61	7.6	4.79	0.53	1533
Ab	0	20	0.0	8.7	87.2	4.0	si	17	1.35	7.1	5.89	2.43	1534
Bgb1	20	63	0.0	13.7	72.8	13.5	sil	20	1.63	7.2	0.65	0.90	1535
Bgb2	63	72	0.0	20.6	67.3	12.1	sil	22	1.80	7.2	0.48	0.39	1536
Bgb & 2Ab	72	86	0.3	44.4	46.3	9.3	l	44	1.89	7.2	0.23	0.29	1537
2AEb	86	96	3.5	49.3	40.8	9.9	l	54	1.91	7.3	0.14	0.14	1538
2EAb	96	105	4.1	48.2	39.6	12.2	l	58	1.86	7.3	0.14	0.17	1539
2BEtb	105	112	6.3	44.5	30.1	25.4	l	64	1.71	6.9	0.22	0.50	1540
2Btb1	112	132	2.9	36.1	30.6	33.3	cl	50	1.60	6.5	0.29	1.11	1541
2Btb2	132	163	2.9	53.6	23.4	22.9	scl	86	1.75	6.2	0.18	0.52	1542
2Btgb	163	197	3.2	46.2	22.9	30.9	cl	82	1.68	6.2	0.23	0.83	1543
2BCtb	197	231	10.7	31.6	49.6	18.8	l	32	1.92	7.7	4.93	0.29	1544
2C1	231	256	22.1	49.4	38.4	12.2	l	67	1.93	7.8	6.69	0.12	1545
3C2	256	286	6.0	40.7	49.5	9.8	l	40		7.8	5.92		1546

sl = sandy loam, sil = silt loam, l = loam, scl = sandy clay loam, cl = clay cloam increases in the topographic low at Profile 23. Bulk density of the whole soil diamicton is >1.83 gm/cm^3 in three samples, only slightly higher than in some of the overlying Bt horizon samples. Other physical and chemical characteristics were either not measured or were sufficiently invariable along the hillslope to provide insight into genesis of the F-S Geosol complex. Organic carbon is very low, and total carbon is as much as 7.6%, reflecting both primary carbonates not leached from the sediment during Sangamon soil formation and also secondary carbonate that was precipitated from groundwater flow following burial. Carbonate pendants were commonly observed on the underside of cobbles in the diamicton and especially underlying gravels, but we did not quantify the amount or attempt to distinguish a chronology of deposition. Clay mineralogy of calcareous Glasford Formation diamicton is dominated by illite (average of 68%), but also contains kaolinite plus chlorite (16%), vermiculite (10%), and expandable minerals (6%).

The Sangamon Geosol formed in the Glasford Formation diamicton, based on grain size characteristics, namely vertical continuity of siliceous coarse fragments and relatively high sand content in all Sangamon Geosol soil horizons (Table 1). Sand and silt content are especially variable with depth in profiles 17 and 23, the topographic high and low profile, respectively (Figure 4). In no horizons do the particle size characteristics resemble an eolian sediment such as loess or dune sand. Clay content is enriched considerably, with pedogenic alteration effects described below.

The Robein Member of the Roxana Silt was identified on the basis of no more than trace amounts of coarse fragments, high silt content (Figure 4) and soil horizonation characteristic of the Farmdale Geosol. While there was no clear macromorphological evidence of downslope movement and sorting of the silt, which is the defining characteristic of the Robein Member, thickness and grain size characteristics suggest the unit did experience hillslope mobilization following deposition (Figures 4 and 5). The decrease in relief from the Sangamon to Farmdale surface is a reflection of downslope movement of eolian silt on the landscape that filled the topo-

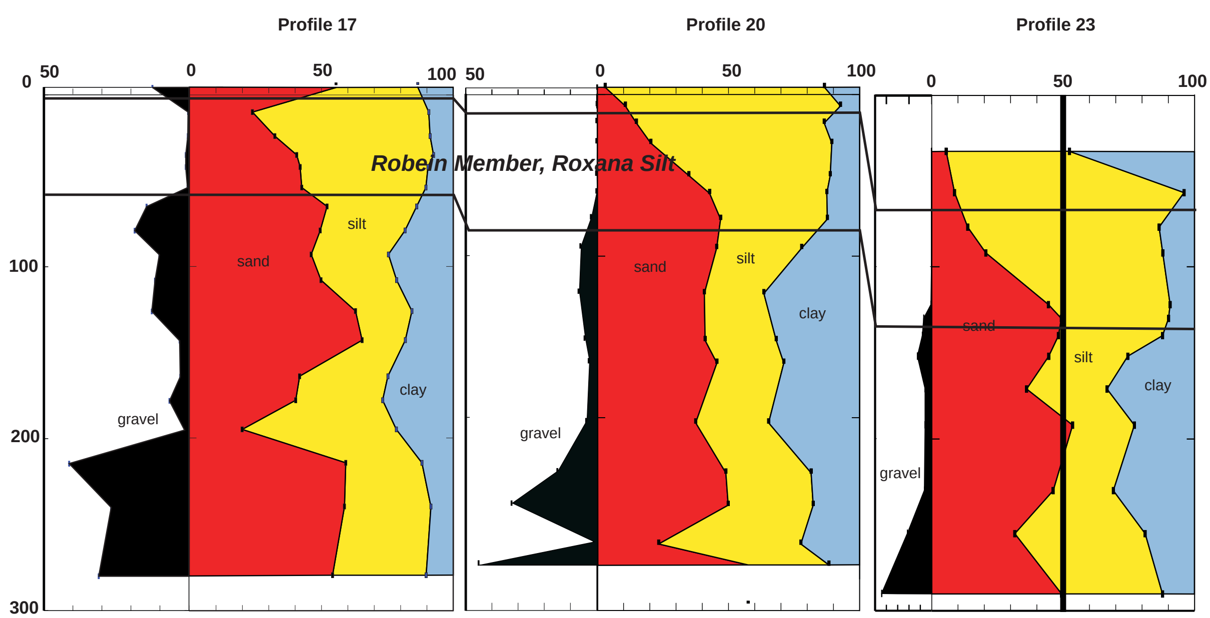


Figure 4 Relative abundance of gravel (total sample) and of the sand-silt-clay fraction of selected profiles shown in Figure 3. Grain-size classes include: gravel (> 2mm); sand (2mm – 0.0625 mm), silt (0.0625 – 0.002 mm) and clay (< 0.002 mm).

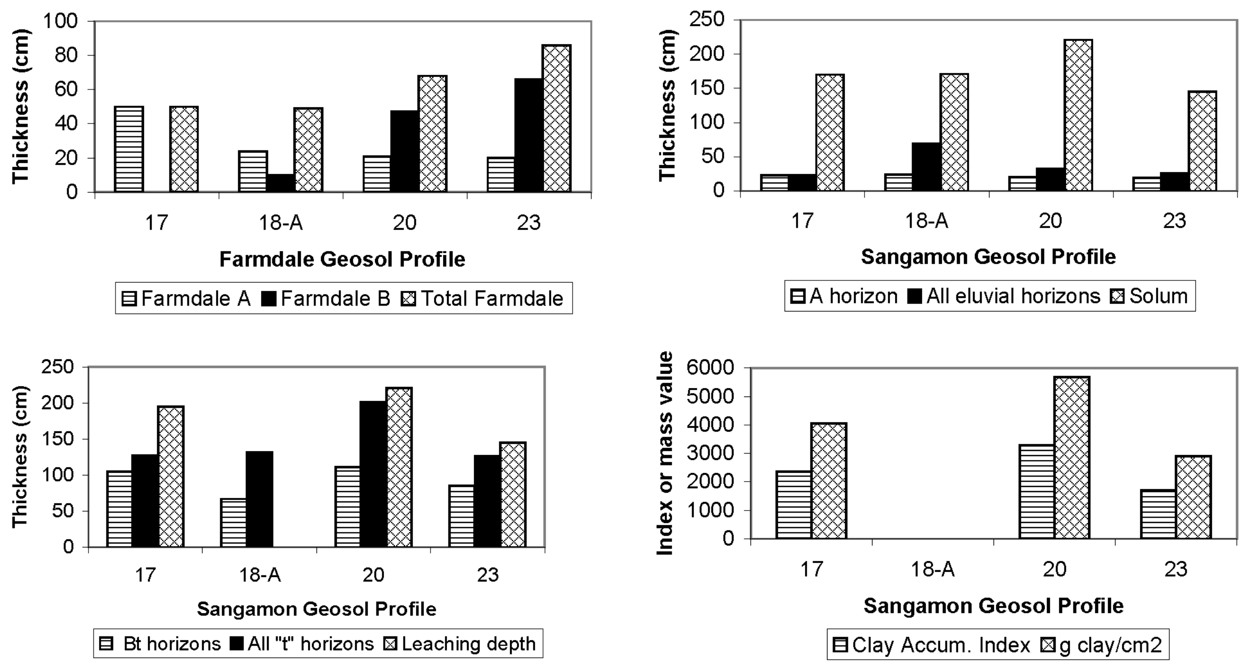


Figure 5 Morphological measure of the Farmdale and Sangamon Geosol profiles along the hillslope transect. (a) A and B horizon and total profile thicknesses of the Farmdale Geosol profiles; (b) thickness of former A horizons (2AEb), all eluvial horizons, and the solum of the Sangamon Geosol profiles; (c) thickness of Bt horizons (2Btb), all subsurface horizons described with clay coatings (e.g., 2BCtb), and the depth of leaching of the Sangamon Geosol profiles; (d) two measures of clay accumulation in Sangamon Geosol profiles relative to unleached Glasford Formation diamiction at the base of each profile (see Birkeland, 1999).

graphic low. While the clay-free geometric mean particle size fines somewhat down slope, only in the lowest profile is the mean particle size considerably finer, perhaps indicating relatively uniform contributions of particle from across the hillslope (Table 1).

An interesting mineralogical feature of the Robein Silt is the occurrence of hydroxy interlayed vermiculite (HIV), which is not as dominant in the Robein Silt in other regions of Illinois. HIV (or HI smectite) is identified on the basis of a very distinctive peak at about 6.1° 2Θ that does not collapse after a heat treatment (Figures 6 and 7); Barnhisel and Bertsch, 1989). The significance of HIV in the clay of the Robein is that it may provide clues to the provenance of dust during MIS 3, discussed below.

Conformably overlying the brown, organic-rich silt of the Robein Member is pinkish-brown, fissile, rhythmically bedded, laminated silt of the Equality Formation. The unit is calcareous, and locally contains fossils of the lacustrine ostracode *Cytherissa lacustris* and the pillclam *Pisidium conventus*. In many places, the lower contact is marked by a thin, continuous layer of very fine sand. Many laboratory parameters serve to distinguish the lake sediment from the buried soil such as the former's low organic carbon content, higher bulk density, high illite content, and high contents of clay-sized calcite and dolomite. The overlying diamicton of the Tiskilwa Formation is identified by its hardness, high sand and gravel content, high density, and lack of stratification. The Equality and Tiskilwa have very similar mineral types and proportions in the clay-size fraction.

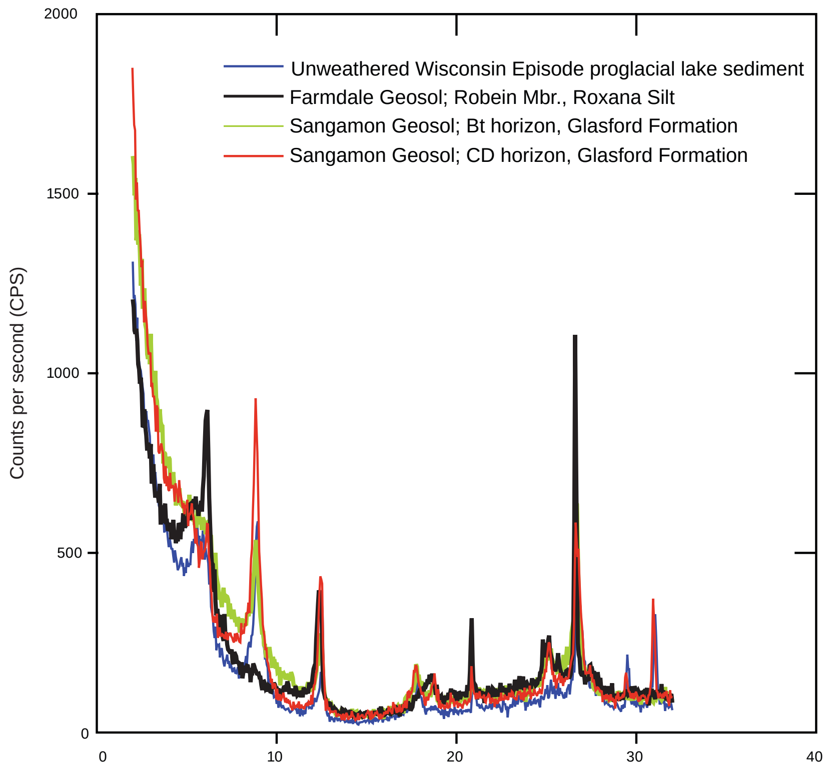


Figure 6 Representative X-ray diffractograms of key units. The scanned samples are oriented aggregate slides and were solvated in ethylene glycol.

Macro and micromorphological characteristics of the pedostratigraphic units

Sangamon Geosol. Morphological and physical characteristics of the Sangamon Geosol indicate that subtle hillslope characteristics do influence morphological expression and measures of soil development over interglacial timescales, but that depth variation in parent material uniformity appears to exert greater control over measures of soil thickness and clay accumulation. Sangamon Geosol solum thickness is least (145 cm) in the low, highest in the middle (221 cm) and slightly less in the upper (170 cm) hillslope positions (Figure 5). Horizonation is generally similar in all profiles along the transect. We recognized elu-

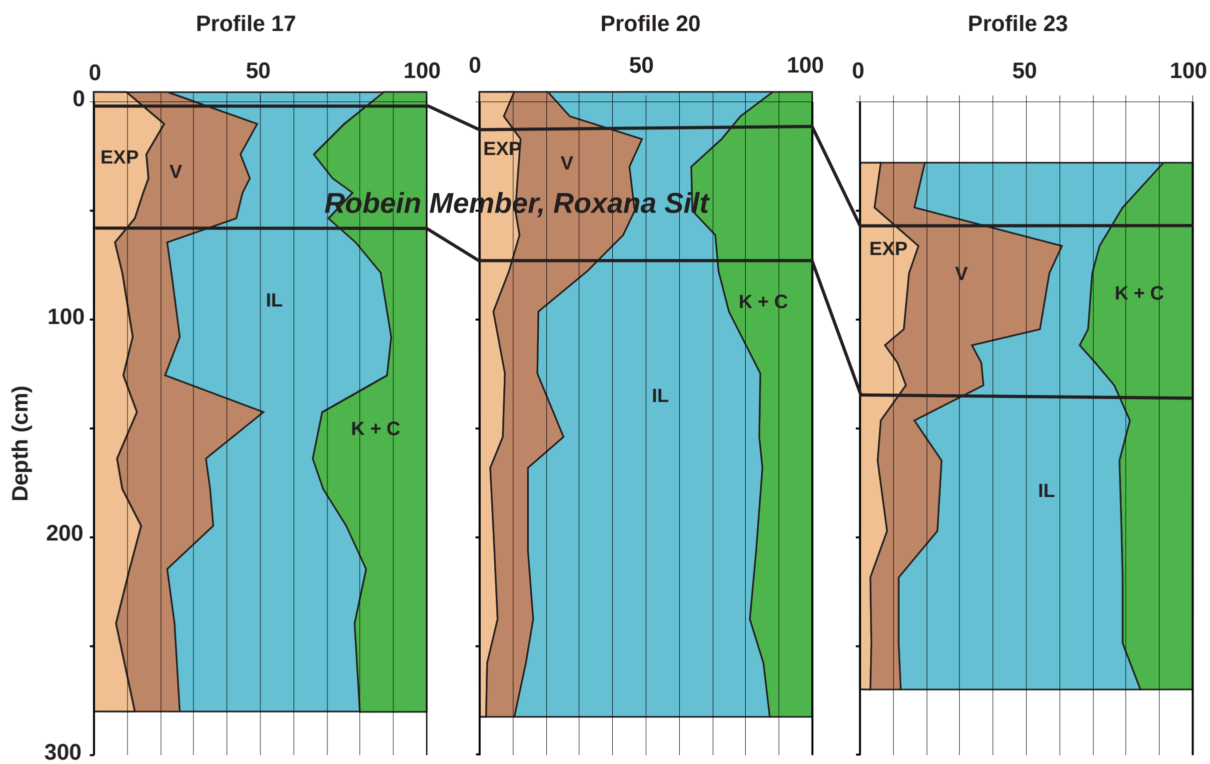
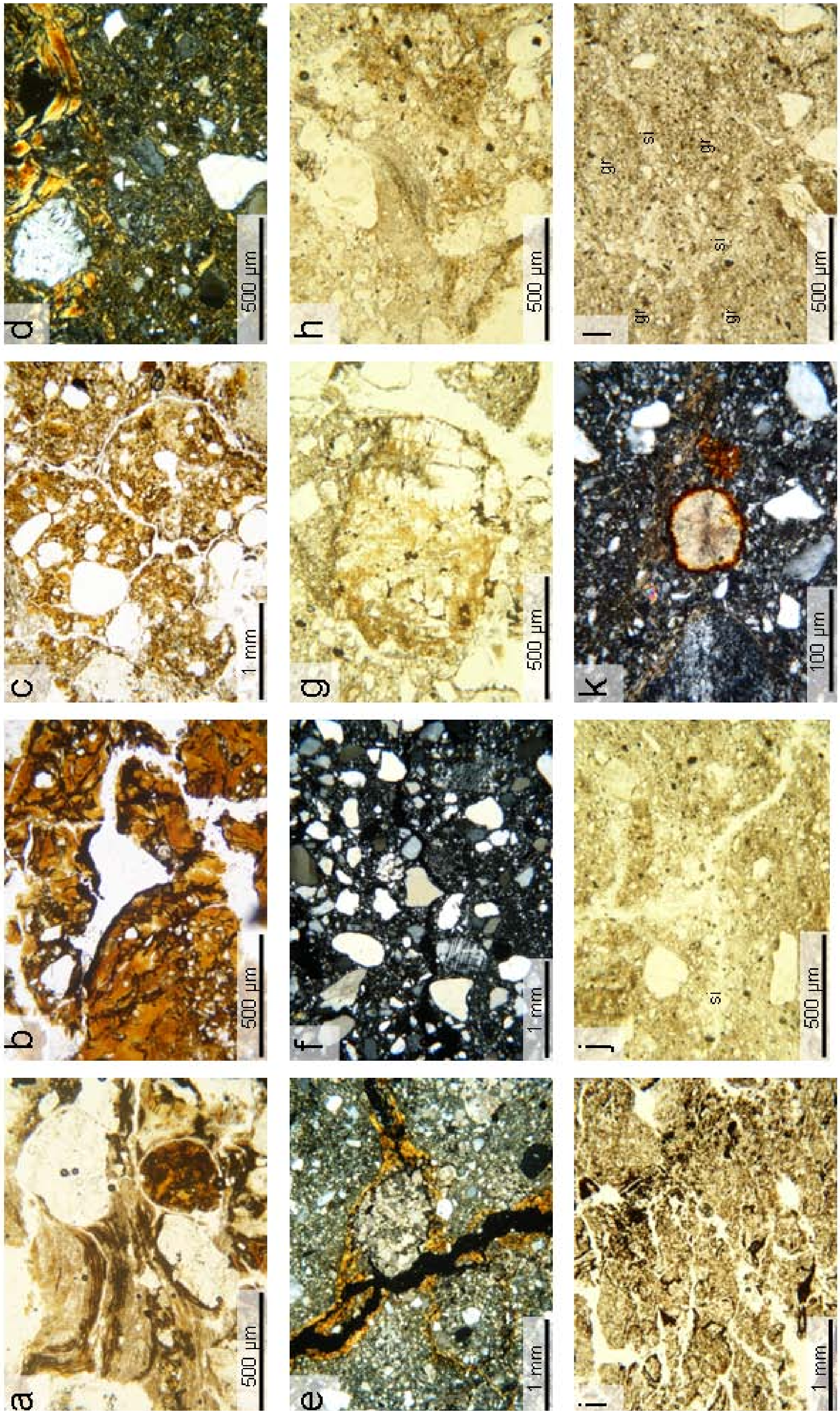


Figure 7 Relative abundance of clay minerals of profiles shown in Figure 4. Clay mineral groups include: EXP (expandable clay minerals that swell to 17Å under ethylene glycol-solvation), V (non-expandable vermiculite), IL (illite), and K + C (kaolinite plus chlorite).

vial upper solum characteristics, designated as an AE horizon on the basis of upper profile location, low fines content, pale low chroma colors, and weak soil structure that is in many places platy and has segregated sand or silt coatings. Thickness of the AE horizon ranged from 23 cm in the upper profile to 10 cm in the low profile. In the low profile, we additionally described a 9 cm thick EA horizon, which when combined with the AE horizon thickness indicates that Sangamon upper solum horizons are relatively similar all along the transect (Figure 5). The apparent differences in horizonation (AE vs. EA) may relate to post-burial modification in the different hillslope positions through accretion of early Wisconsin Episode loess and overprinting by Farmdale Geosol pedogenesis. In the middle and lower slope positions we described a BE horizon, where faces of subangular blocks have some sand coatings and gave the appearance of being "degraded" through clay dispersion and eluviation; clay coatings occur on some ped faces in these BE horizons.

Micromorphology of the upper solum horizons provides evidence of weatherable mineral depletion, clay loss, and fabric reorganization (Figure 8). In thin section, the AE horizons all appear mineralogically mature, being dominated by quartz with only minor amounts of weatherable grains that are visibly being transformed to clays, or the accumulation of opaque Fe oxide segregations or coatings. Microstructure evident in thin section mostly consists of channels and vughs with occasional granules, but in the upper hillslope position platy structure is also evident. Zones of segregated silt that may have originated along the faces of peds often appear isolated in the interior of massive fabric areas. We suspect the collapse following burial, especially under glacial ice, contributed to the degradation of aggregation. The b-fabric of these horizons is variable. In some places, it is nearly absent, reflecting the lack of fine grained (plasma) and dominance of skeletal materials. In other profiles, small areas of speckled or striated b-fabric appear to be detached and embedded from underlying horizons, features characteristic of E horizons (Bullock and Thompson, 1985). The lack of grain coatings and void infills of birefrigent

Figure 8 (Facing page) Photomicrographs of vertical thin sections with up on top of image, PPL = plane polarized light, XPL = cross polarized light. Images (a-e) are of Sangamon Geosol B horizon characteristics. (a) Compound infillings with alternating impure and limpid clay, plus rounded embedded fragment of a clay coating (upper profile, 2BEtb, PPL); (b) illuvial argillans in channel, note most clay is iron stained, but is overlain by an opaque coating of Fe/Mn that also fills cracks in the argillan (upper profile, 2Btb1, PPL); (c) subangular blocky microstructure showing abundance of illuvial clay that occurs as fragmented clay coatings, matrix infillings, and grain coats, while clay coatings in channels are largely absent (middle profile, 2Btb1, PPL); (d) well-oriented illuvial clay lining channels and matrix dominated by mosaic stippled b-fabric (low profile, 2Btb1, XPL); (e) blocky microstructure with accommodating planar voids lined with illuvial clay coatings in horizon still containing primary carbonate grains (two large grains labeled c) marks the advancing front of pedogenesis into the lower solum horizon (low profile, 2BCtb, XPL). Images (f-h) are of Sangamon Geosol upper solum characteristics. (f) 2AEb horizon of upper profile showing remarkably clay-free porphyric fabric reflecting eluvial loss of clay (XPL); (g) feldspar grain weathering to clay surrounded by clay-free matrix and void indicating the clay production potential of the upper sola horizons was probably limited to mineral weathering sources (middle profile, 2AEb, PPL); (h) sorted silty bands and occasional clay remaining in fabric of the upper solum of the lower profile (2AEb, PPL). Images (i-k) are of Farmdale Geosol characteristics. (i) Moderately separated granular microstructure with some tendency toward platy structure probably as a result of compression; note humus staining is common throughout and opaque areas are Fe/Mn oxides that may have precipitated around organic materials (middle profile, Ab1, PPL); (j) thin silt coatings only a few grains thick, and lack of illuvial clay coatings (middle profile, ABgb2 PPL); (k) secondary calcite accumulation surrounded by Fe oxide, indicating calcite precipitation prior to Fe; note also stipple speckled and granostriated b-fabric but no illuvial clay pedofeatures (middle profile, Bgb3, XPL); (l) apparently massive microstructure that appears to consist of humus-stained granules (gr) separated by unstained silt (si) (lower profile, Bgb2, PPL).


a
500 µm
b
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1 mm
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material indicates long term dispersion and translocation of clay from these horizons to subjacent horizons. We are confident that these horizons represent Sangamon Geosol eluvial upper sola horizons, and that these profiles were not truncated prior to deposition of Robein Silt.

All of the Sangamon Geosol profiles are visually dominated by horizons with argillic horizon characteristics such as blocky soil structure that is coated with readily identifiable clay coatings. Soil colors of upper Bt horizons along the transect vary systematically along the hillslope, ranging from 5YR in the highest profile to 7.5 YR to 10 YR hues in the lowest position. In all instances, the "brightest" color development is expressed in the upper part of the Bt horizons. In most horizons, color values and chromas are $\geq$4 and indicate oxidation and retention of iron oxyhydroxides in the solum. In most upper Bt horizons color values are $\geq$4 and chromas are $\geq$4. Only in the lower Bt and subjacent horizons of the lower profile do colors display chromas $\leq$2.

All Bt horizons of Sangamon Geosol profiles along the transect display easily recognized evidence of clay illuviation. In upper Bt horizons of every profile, structure is moderate medium and fine subangular blocky and covered with clay coatings. In the upper profile the coatings are pervasive and control the color of the horizons, whereas in the lower profile the coatings are moderately thick, patchy, and have darker colors than the matrix. In lower Bt horizons, clay coatings are moderately thick to thick and continuously cover the faces of moderate medium and coarse subangular blocky structure. Clay coatings extend into horizons described as BCt and CBt that become progressively more till-like in terms of density and fabric arrangement. The coatings in these horizons are often confined to ped faces of coarse structural units, but also commonly coat clasts and sand grains. The color of the clay coatings in the lowermost horizons is distinctly redder (7.5 YR hue) than the matrix (often 2.5Y hue).

In thin section, the single most obvious pedogenic effect in B horizons of the Sangamon Geosol is accumulation of illuvial clay, although as with the upper sola horizons, the Bt horizons also appear mineralogically mature in the sand and coarse silt fractions. The large proportion of clay occurs as grain coatings, as infills in packing voids, and as illuvial coatings in pores and along ped faces. Microstructure is most typically blocky and vughy, with b-fabric that is striking in the amount of clay that is oriented around individual grains, coating voids, or filling voids. Clay bodies often show well developed extinction patterns, indicating preferred orientation of the clay and an illuvial origin (Stoops, 2003). The colors of illuvial clay coatings generally reflect the drainage position along the hillslope, being distinctly reddish in the upper hillslope position, more yellowish in the middle position, and then a mix of red and yellow in the lowermost hillslope position. While fragmented bits of clay coatings occur in the matrix of most Bt horizons, a large amount of clay occurs as infilled voids, which probably reflect void space that was created by the dissolution of carbonate grains subsequently filled with clay. Deeper in the profiles, the distribution of illuvial clay changes from being pervasively distributed throughout the fabric to being more concentrated as illuvial clay coatings along the walls of voids, with the interior of aggregates showing a weakly expressed stipple speckled b-fabric. These features reflect the less weathered nature of the lowest horizons, where carbonate dissolution and clay production and migration had not yet occurred prior to burial. In addition, redoximorphic features such as opaque Fe/Mn coatings that overlie illuvial argillans and fill cracks in the argillans, indicate that Fe/Mn accumulated late during pedogenesis or perhaps following burial by Wisconsin Episode glacial sediment.

Farmdale Geosol. Changes in thickness and morphology of the Farmdale Geosol along the hillslope transect are more marked than the Sangamon Geosol, despite the overall greater simplicity of the morphological expression of the Farmdale Geosol profiles. We attribute the thickness differences to slope-redistribution of loess and the morphological differences to changes in landscape drainage.

In the upper landscape position, the solum is 50 cm thick, color is oxidized (10YR 4/4), soil structure is dominated by moderate fine plates that part into fine subangular blocks. Much of the platy structure is coated with gray segregated silt coatings and high chroma iron stains (7.5 YR 5/6). In this landscape position we did not describe a genetic B horizon in the Farmdale Geosol. The uniformity of color, texture, and structure led us to the interpretation that the entire 50 cm thickness was an A horizon aggraded on the Sangamon landscape. Given the uniformly thinner Farmdale Geosol A horizons in the other profiles along the hillslope, it is distinctly possible that this soil possessed B horizon characteristics prior to burial but that compaction and post-burial degradation led to loss of those characteristics. Alternatively, this soil may never have been differentiated with a B horizon because the topographically higher landscape position was likely not as wet as the topographically lower profiles; our recognition of subsoil horizon characteristics in the lower profiles appears to largely reflect differences in drainage and how wetness impacted soil color and soil organic carbon storage.

In the middle landscape position, the solum is 68 cm thick and is clearly differentiated into A and B horizons (Figure 5). The A horizons are 21 cm thick, colors range from dark gray (10YR 4/1) to black (10YR 2/1), and soil structure is thick platy that parts to moderate fine granular. The thick plates are interpreted as resulting from compression of the granular structured A horizon. The three B horizons are designated Bg horizons on the basis of low chroma colors. The horizons variably have subangular blocky or granular structure and gray, segregated silt coatings are common. Only the lowest B horizon (Bgb3) had the field appearance of gleying rather than dark colors from organic matter accumulation.

In the low landscape position, the solum is 86 cm thick, and while differentiated into multiple horizons have cumulic morphology. The uppermost horizon is black (2.5Y 2.5/1) and appears massive with some platiness, which supported by soil organic carbon values >2% and low bulk density (Table 1), is interpreted as a sapric organic horizon (muck). Wood and charcoal fragments are common and correspond to the source of radiocarbon ages presented above. The underlying horizons are very dark gray (10YR 3.5/1) and consist throughout of very friable moderate fine granular structure that is somewhat compressed. Redoximorphic features such as common fine gray (2.5Y 5/1) depletion zones and few fine yellowish brown (10YR 5/6) iron accumulation zones are typical.

In thin section the Farmdale Geosol A horizon is readily identifiable in all three hillslope positions on the basis of one or more of the following: charcoal, humus staining, or compressed granular structure (Figure 8). The strength of expression of these features increased downslope. Although A horizon characteristics such as color and aggregation are readily identifiable in thin section, the silt coatings typically observed during morphological description and sampling were not pervasive in thin section and proved to be only be a few grains thick. This is attributed to the silt coatings losing their field moisture more readily than the organic-rich and clayier ped interiors. The compressed granular structure may simply reflect compaction under the Wisconsin Episode glacier and sediments, but when considered with the silt coatings, may reflect the occurrence of segregated ice in the profile prior to burial (Harris, 1985).

Particle size and clay mineralogical characteristics of the pedostratigraphic units

Particle Size Trends. In conjunction with morphology and micromorphology, profile distribution of soil particle size separates all indicate that the Sangamon Geosol profiles are texture contrast profiles with an important part of horizon differentiation being the translocation of clay from eluvial upper sola horizons to subsurface B horizons. Particle size characteristics in each profile are variable with depth, and still strongly reflect sedimentation patterns of the glacial sediment

(Table 1). Sand and silt values often change by 20 or 30 percent between samples, and the clay-free geometric mean particle size often changes by more than 50 µm over short intervals. We attribute these characteristics to pre-pedogencic, geogenic processes discussed above.

Coarse fragments support this interpretation; they are common in the Sangamon Geosol and decrease upward in the Farmdale Geosol. Beneath the former Sangamon surface, coarse fragment content is variable with occasional concentrated layers or lenses that also correspond with increases in sand content that strongly suggest they are the result of sedimentation rather than pedogenesis.

Clay content in each profile is greater in subsurface horizons than overlying upper sola horizons consistent with texture-contrast profiles, but depth distributions are not uniform nor do they consistently decay from a horizon with peak clay content (Table 1). The Bt horizons with peaks in clay content also have less sand, which may reflect finer grained packages of stratified sediment that either contained more clay initially, acted to trap more illuvial clay, or provided a weathering environment that was conducive to *in situ* weathering and clay production.

Recognizing the variable sedimentation pattern is critical to our understanding of formation of the Sangamon Geosol because particle size discontinuities influence soil water and air movement through the profile, which ultimately affects soil depth, leaching, clay translocation and other morphological and weathering features. Along the hillslope transect, the middle profile shows the most uniform particle size trends, and it is also the deepest profile. We believe the greater depth of this profile is directly a result of the greater uniformity of particle size distribution with depth that enabled more carbonate leaching and solum extension than adjacent profiles where particle size, hence pore size, discontinuities impeded water percolation. The greater thickness of genetic soil horizons in the midslope profile also produces the highest measures of profile clay accumulation (Figure 5), since, in addition to clay content and bulk density, these measures are thickness dependent.

Clay Mineralogy. Down profile trends in clay mineral composition in the Sangamon Geosol profiles are consistent with our interpretation of the soils forming into stratified diamiction. We have divided the profiles into an upper zone that has characteristics of a weathered clay mineral assemblage, and a lower zone with a secondary peak that appears to mimic weathering trends. In the upper sola of each profile of the Sangamon Geosol, there are increases in kaolinite, vermiculite, and expandable minerals at the expense of illite, a typical mineral weathering sequence observed in the Sangamon Geosol (Willman et al.,1963). In all three profiles, this trend is evident for 30-40 cm beneath the former Sangamon landsurface. With depth, however, vermiculite and kaolinite again increase relative to illite, which is not an expected weathering pattern in midcontinent soils and paleosols. This trend is most pronounced in the upper profile, and we interpret this trend to be the result of sediment stratification during deposition.

Throughout the Sangamon Geosol the percentage of illite remains above 25%, often >60%, and in diffractograms the illite peak remains sharp throughout the solum. We do not observe clear evidence of mixed-layered kaolinite and expandable minerals (K/E) in these profiles unlike those developed in loess in southern Illinois (Hughes et al., 1993; Grimley et al., 2003). We attribute this to either less intensive weathering due our sites more northerly location or differences in parent material. Compared to most other sites in which Sangamon Geosol profiles have been characterized, our profiles have parent material that is clay-poor and carbonate rich; many physical characteristics described above may be attributed to carbonate dissolution.

The mineralogy of the Robein Member (Roxana Silt) is unusual in that several samples have little or no illite, and abundant HIV vermiculite (Figure 6). The latter mineral has been identified, usually as a minor component, of most glacial drift and loess deposits in Illinois. The absence of illite is especially noteworthy because sharp illite peaks are observed in the underlying weathered diamicton. Illite also occurs in nearly all Ordovician-through-Pennsylvanian-age rocks in Illinois. This suggests that the outwash source of this loess was derived from areas where upper Paleozoic rocks are uncommon. We suspect that the provenance of the HIV vermiculite is weathered mica from the granitic and other crystalline rocks of central Wisconsin. In this scenario, the Rock River valley was the outwash conduit that brought materials with this unusual clay mineral composition to near the LaFarge site.

Questions for discussion

Why is there no evidence of eolian silt (loess) associated with Illinois Episode glaciation?

This region of Illinois contains ≥1 m of late Wisconsin Episode Peoria Silt that originated as a loess deposit mantling the deglaciated MIS 2 landscape. Plus, Forman and Pierson (2002) offered IRSL evidence from several Midwestern loess sections that they interpreted as evidence of two episodes of loess sedimentation during the period MIS 5 through MIS 4. The Feltes site is a distal sedimentation site at ~150 km from a presumed Mississippi Valley source area, yet we see no compelling evidence in soil morphology or particle size characteristics to indicate any recognizable additions of loess to the surface of the Sangamon Geosol prior to the Robein Silt. That is, we see no evidence of MIS 6, 5, or 4 loess overlying Glasford Formation glaciogenic sediments in this distal setting. We suggest that environmental conditions that controlled dust sources and/or distribution at the end of the Illinois and Wisconsin Episodes glaciations were different.

What mechanisms explain the texture-contrast morphology of the Sangamon Geosol?

Multiple causes of texture-contrast soil profiles with coarser textured upper sola and clayey B horizons have been described and occur in a variety of environments (Phillips, 2004). For the Sangamon Geosol, we document morphological and micromorphological evidence of eluviation-illuviation (lessivage), probably along with subsoil clay production by weathering. Macro-, meso-, and micromorphological evidence all indicate that upper sola horizons are depleted in clay and that soil horizons are enriched in clay that is well oriented and likely illuvial.

In contrast, Paton et al. (1995) presented a model of the origin of texture contrast soils where the biological actions of soil mixing and surface mounding followed by downslope rainwash depletes the upper solum of fines, leaving the characteristic coarser texture. Johnson and Balek (1991) proposed the Sangamon Geosol texture contrast originated through such a model, illustrating their idea through the use of hypothetical chronographs. Our data do not support this model, however. Evidence such as the uniformity of thickness and particle size characteristics of the upper solum horizons in all topographic positions, along with sandy upper sola textures in the topographic low do not support the contention that long-term soil mounding by fauna and subsequent erosional winnowing by rainwash led to accumulation of fines in the low position, as described by Paton et al. (1995). We acknowledge that our catena does not provide a full understanding of the geometry and flow paths of surface water in the paleolandscape, but given the short 9% slope along the transect, we believe if this process had been significant to the origin of these texture contrast profiles that some sedimentary record would be preserved by variation in sandy topsoil thickness and fines accumulated in the low topographic position.

STOP 7: DeKalb mounds

Archives of deglacial history and postglacial environments

B. Brandon Curry, Michael E. Konen, Timothy H. Larson

Abstract

The "type" DeKalb mounds of northeastern Illinois, USA, archive a rich flora and fauna dating from about 20,700 to 18,700 cal yr BP (17,500 to 15,250 C-14 yr BP). The loess-mantled mounds are formed of (1) a basal lag of sand and gravel, (2) laminated, rhythmically bedded, fossiliferous silt loam and very fine sand and, (3) weathered sand and gravel or diamicton. Generally from 2 to 7 m thick, the unoxidized fossiliferous sediment contains well-preserved fossils of ostracodes, pillclams, tundra plants, and chironomids.

In Illinois, DeKalb mounds are found on ground moraine and moraines of the last glaciation as far south as Royal, Illinois (40.17ºN). DeKalb mounds are especially abundant on the Gilman, Woodstock, and Tinley moraines, and on the southern arm of the Marseilles Morainic System. South of about 40.15ºN, the ice-walled lake deposits tend to be < 5 m thick, and the plant fossils are pyritized and unsuitable for radiocarbon dating. Other fossils, such as beetle elytra and ostracode valves, persist. The ice-walled lake plains near Royal were formed shortly after the Putnam Phase (Wisconsin Episode), and are the oldest such features in Illinois, but they have yet to yield a meaningful chronology. The youngest ice-walled lake plains so far explored in Illinois are located on the Deerfield Moraine in Lake County. From the base of a lacustrine succession, abundant *Dryas integrifolia* twigs and leaves yielded an age of about 16,270 cal yr BP (13,650 ± 140 C-14 yr BP; UCIAMS-46829), ages that are consistent with the oldest radiocarbon ages associated with the Glenwood Spit formed by Glacial Lake Chicago at about 16,310 cal yr BP (13,890 ± 120 C-14 yr BP; ISGS-1649).

The sedimentology and architecture of the mound successions, as well as modern analogs of Lake Michigan ostracode fauna, indicate the mounds were once ice-walled lakes as much as 8 to 15 m deep. Ice-walled lakes similar in depth and form to the DeKalb mounds occur today on the slowly melting Kara ice sheet on the Siberian Taymyr Peninsula. Paleoecological and stable isotopic evidence suggests that the moisture source for the ice-walled lakes was primarily precipitation, and less from meltwater. The ostracode fauna of the "type" DeKalb mounds is dominated by *Cytherissa lacustris* and *Limnocythere friabilis*. *C. lacustris* valves from the mounds yield $\delta^{18}O$ values of -2.3‰ to +1.5‰, which are similar to values obtained from modern ostracode valves from nearby hard-water lakes which receive about 75% of their precipitation from moisture evaporated from the Gulf of Mexico.

Exogenic fossils of tundra plants preserved in mound successions provide material for C-14-age control on early deglacial history and postglacial landform development. The span of time when deglacial topographic inversion occurred is revealed by the C-14 age of terrestrial plant remains preserved in ice-walled lake sediment compared to those C-14 ages of organics from basal kettle successions that accumulated after melting of most surficial ice. In northeastern Illinois, this topographic inversion occurred between about 18,600 and 16,650 cal yr BP (15,150 and 14,000 C-14 yr BP.).

INTRODUCTION

Numerous circular to semicircular flat-topped hills from 30 to 10,000 m across occur throughout DeKalb and Kane counties in northeastern Illinois, USA. The hills are subtle features, rising from about 1.5 to 9 m above adjacent glacial deposits. Flemal at al. (1973) named these features the DeKalb mounds. They attributed their genesis to sediment infilling of melting pingos (protrusions in areas of continuous permafrost caused by deep, upwelling groundwater) although they did not dismiss that the mounds could be ice-contact ridges, a hypothesis championed by Ianacelli (2003). Some of the key characteristics of the DeKalb mounds (Figure 1) include: 1) circular to semicircular appearance from aerial photography, with light-toned rims and darker interiors (evocative of the moniker “glacial doughnuts”); 2) narrow moats; 3) “parasitic” or “satellite” doughnuts; 4) rim-breaching channels; and 5) interiors of laminated silt and fine sand. Mapping DeKalb mounds is facilitated by interpretation of shaded relief maps of 2-m DEM data from LiDAR or detailed photogrammetry (Curry, 2006). Our new contribution to understanding these landforms is the discovery of micro- and macrofossils, including abundant ostracode valves, rare to common tundra plant leaves and stems (Figure 2), rare chironomid head capsules, and beetle elytra. Ostracode autecology and valve chemistry indicate moisture from the Gulf of Mexico provided the bulk of water in the ice-walled lakes.

METHODS

Color infra-red and black-and-white aerial photography were examined in the type area of the DeKalb mounds (Figure 3) to assess their distribution, shape, and size. In order to conduct future statistical analyses of these attributes, the mounds were mapped using ArcGIS. Study cores were sampled with PowerProbe and CME-750 drill rigs, as well as a Giddings rig. Many cores were sampled on the Hampshire, Illinois, 7.5-minute Quadrangle (Curry, 2008), in particular at a mound located 2 km west of Burlington, Illinois (Stop 7). Additional cores were sampled from mounds located on or adjacent to the Woodstock, Tinley, Deerfield, and Champaign moraines, and the Marseilles Morainic System (Figure 3).

Core H-22 was selected for detailed study (Figure 1). Plant macrofossils, ostracodes, and other fossils were picked and identified from prepared core segments about 4 cm long. Moist samples were disaggregated by pretreating in boiling water with a pinch of baking soda, cooled to room temperature, and wet-sieved using a shower spray on a Tyler #100 sieve (150 µm openings). Plant macrofossils, if used for C-14 analysis, were kept refrigerated in vials containing tap water and a drop of 10% hydrochloric acid. Ostracode valves were picked from the dried residue and identified using the NANODe website (Forester et al., 2006). Four to five valves of *Cytherissa lacustris* were selected for chemical analyses, and cleaned by soaking in household 2% hydrogen peroxide until particulate matter, if present, became detached from the valves. The valves were then bathed in methanol for 1-2 minutes. Air-dried shells were analyzed for C-13 and O-18 using a Finnigan-MAT 252 mass spectrometer with a Kiel II device. The phosphoric acid residue from each sample was subsequently analysed for dissolved Mg^{2+}, Ca^{2+}, and Sr^{2+} using an ICP-MS (USEPA, 1994). The mineralogy of the <2µm fraction was determined by X-ray diffraction of ethylene glycol-solvated, oriented aggregates (Hughes et al., 1994). The clay slides were also analyzed for color parameters (L*, a*, and b*) defined by the Commission Internationale de l'Eclairge (CIE, 1978). The a* parameter measures redness (positive values) to green (negative values). Particle-size analysis was conducted using a modified pipette procedure (Soil Survey Staff, 1996).

High-resolution resistivity profiling was used to determine the geometry of sedimentary units with contrasting particle-size characteristics. Two east-west profiles were acquired at Stop 7 (Figure 1). Both lines were acquired with a multi-electrode resistivity system using a dipole-

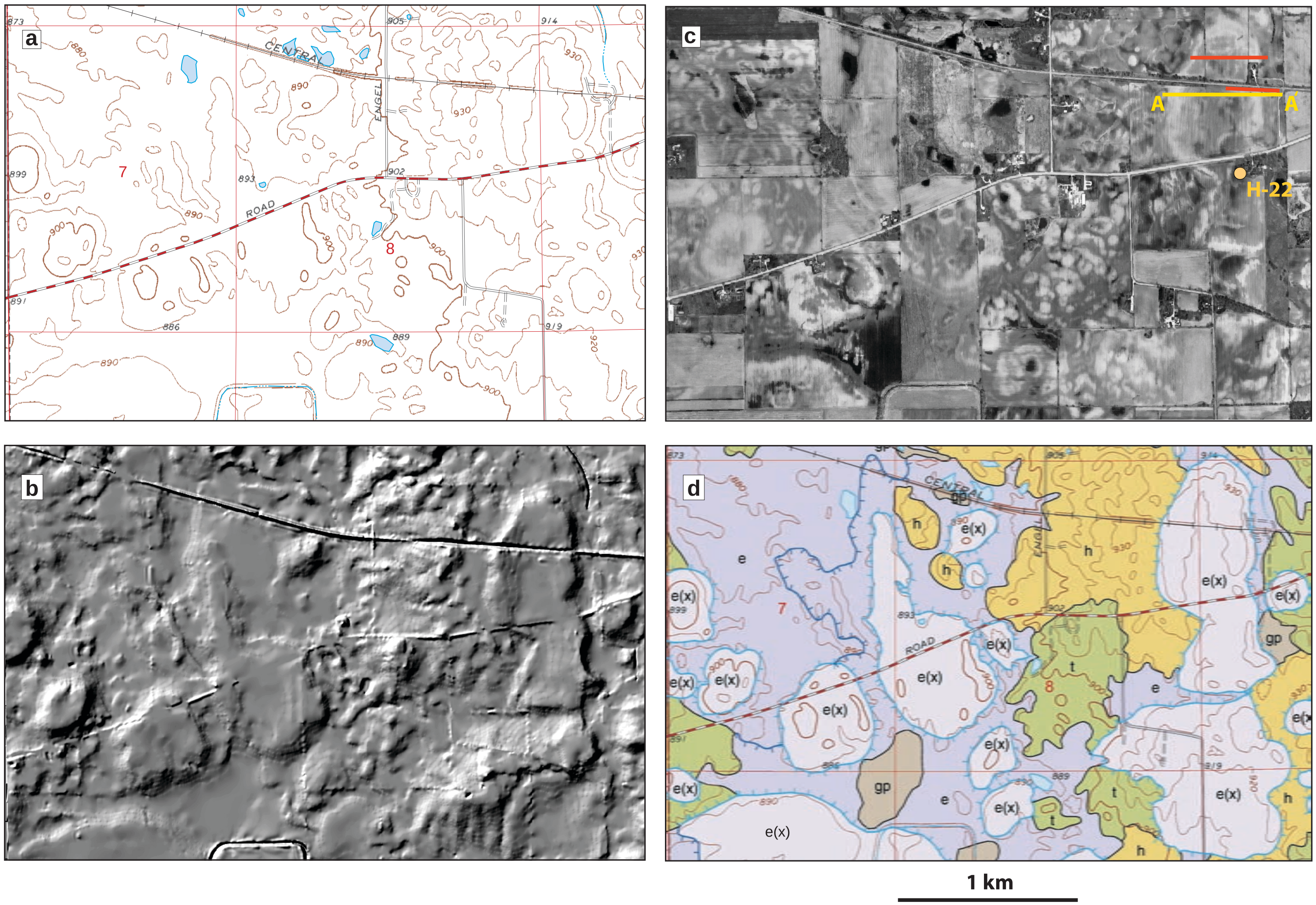


Figure 1 Located on the Hampshire, Illinois, 7.5-minute Quadrangle, these four maps are of the same area along Plank Road, immediately west of Burlington, Illinois. DeKalb mounds are mapped as unit e(x) on map d. The maps show (a) topography with ten-foot contour intervals, (b) U.S. Geological Survey digital orthophoto quadrangle (DOQ) imagery, (c) shaded relief made from a DEM of 2-foot contours, and (d) the surficial geology map. On the latter map, gp is Grayslake Peat; e, Equality Formation; e(x) Equality Formation complex; h, Henry Formation sand and gravel and t is clay loam diamicton of the Tiskilwa Formation (from Curry, 2008). Map c also shows (1) the line of section for the resistivity profiles (red lines), the geological cross section A-A' (yellow; Figure 7), and location of boring H-22.

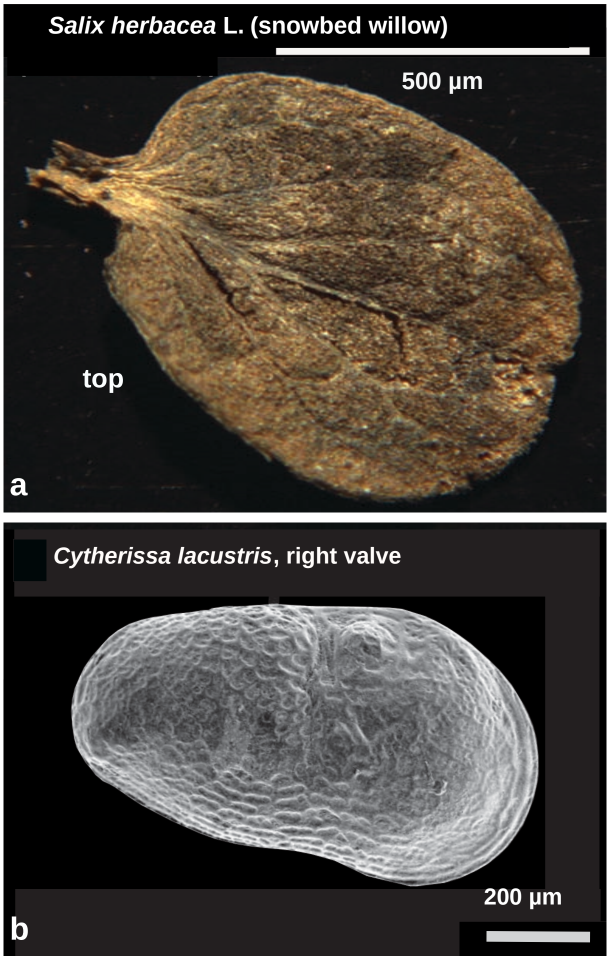


Figure 2 Examples of fossils recovered from laminated silt loam forming the DeKalb mounds (a) leaf of *Salix herbacea* (snowbed willow) and (b) *Cytherissa lacustris,* an ostracode.

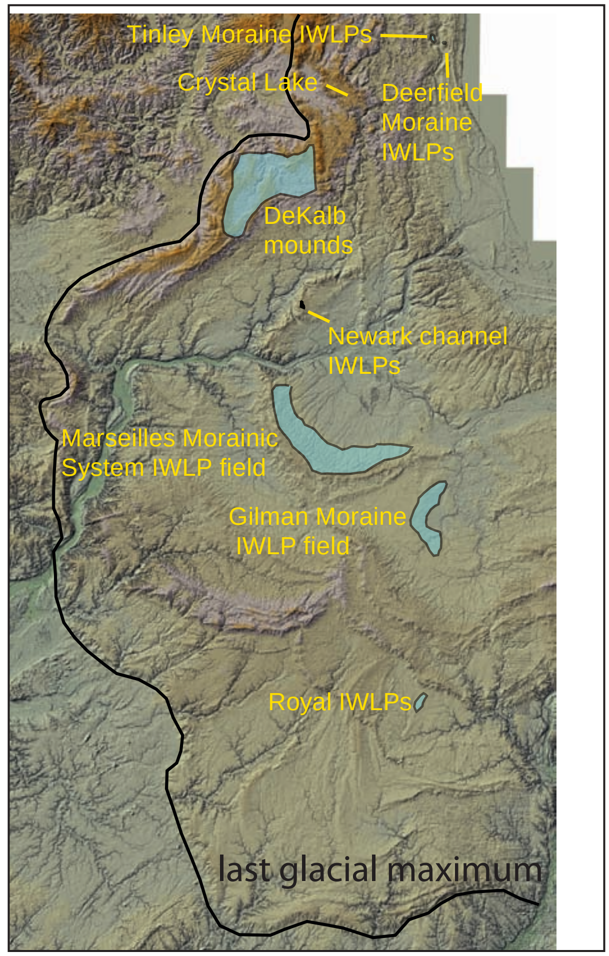


Figure 3 Location of areas with high concentrations of “glacial doughnuts”. The location of Crystal Lake, McHenry County, is also shown.

dipole array and 2-m electrode separation. Approximate depth of penetration for this array was 10 m and the resolution about 1 m laterally and vertically. Raw data were processed into 2-D topographically corrected resistivity models using a least-squares inversion technique (Loke and Barker, 1996).

RESULTS

Thousands of DeKalb mounds occur throughout northeastern and central Illinois in areas well beyond the “type” area described by Flemal et al. (1973; Figure 3). On aerial photography, mound shapes are round, multilobate (Figure 1), elliptical (Figure 4), and irregular (Figure 5). In the southern half of the DeKalb, Illinois, 7.5-minute Quadrangle, 311 features interpreted to be DeKalb mounds have been identified on aerial imagery, and all four shapes are present (Table 1). On the Hampshire, Illinois, 7.5-minute Quadrangle, 95 DeKalb mounds were mapped by Curry (2008).

Occurring above diamicton of the last glaciation, the DeKalb mounds are composed of five layers, including (1) basal sand and gravel, (2) fossiliferous, uniform silt and very fine sand, (3) fossiliferous, rhythmically bedded, (varved?), laminated silt loam and very fine sand, (4) weathered

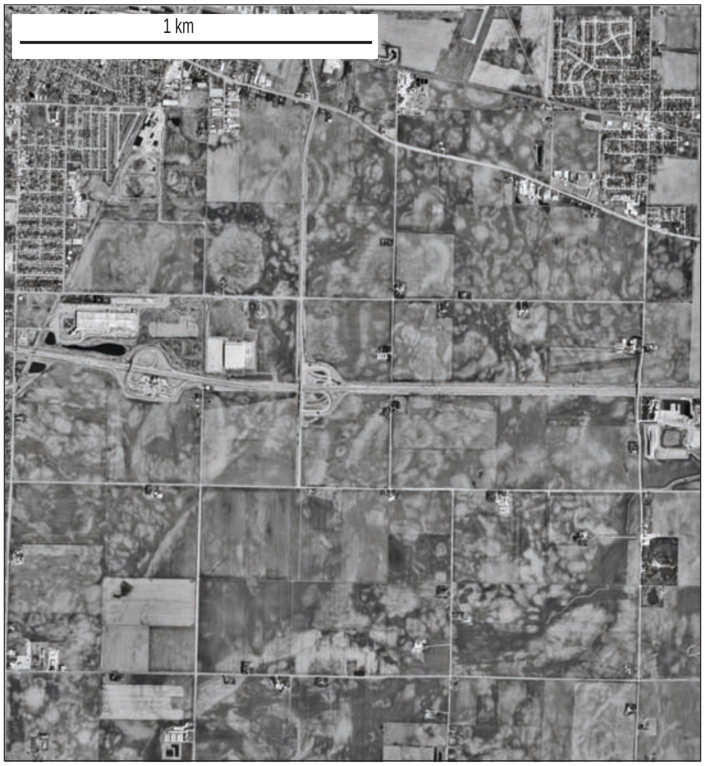


Figure 4 Typical distribution and shape of DeKalb mounds in the SW ¼ of the Sycamore, Illinois, 7.5-minute Quadrangle. Note the lighter colored rims and the variety of mound sizes and shapes. The large elliptical mound in the southwest corner of this image is typical of many of the larger mounds with their long axes oriented northeast-southwest.

Figure 5 Typical distribution and morphology of irregularly-shaped mounds in the NW ¼ of the DeKalb Quadrangle. Note the lighter toned rims and the variety of mound sizes and shapes.

Boring H-22: 2675'NL, 475'WL, Section 9, T. 41 N., R. 6 E.; Elevation: 925 ft (from 2 ft DEM). API: 12-089-35696
Depths are in cm.
0-18; Silt loam, very dark brown (10YR 3/1); friable [Ap horizon]
18-35; Silt loam; as above, med crumb struct., friable, earthworms, leached [A hor]
35-51; Silt loam; as above, but not as friable, and somewhat finer textured [A3 hor]
51-79; Silt loam, thick, continuous, dark gray (2.5Y 4/1) argillans, very firm, varicolored (deoxidized), sandier with depth [Bt horizon in Peoria Silt]

79-92; Transition zone [IIC1 in Henry Formation]
92-114; Diamicton, sandy loam, leached, yellowish brown (10YR 5/4)
114-122; Sand and gravel, loose, oxidized, slowly calcareous, light yellowish brown (10YR 6/4) [IIC2 horizon in Henry Fm]

122-139; Silt, coarse, laminated, calcareous, light brown (7.5YR 6/4); Equality Formation
139-152; Core loss
152-164; Sand, very fine, stratified, with thin beds of uniform silt; sand is yellowish brown (10YR 7/6), silt is light brown (7.5YR 6/4)
164-207; Silt; uniform to laminated, light brown (7.5YR 6/4); mottles (5%), gray 5Y 6/1) 207-210; Krotovina
210-227; Ibid.
227-263; Silt loam, brown (8YR 5/3), rhythmically laminated with very thin beds to laminae of very fine sand; silt loam layers are 5 to 10 mm thick; very fine sand beds are 1 to 2 mm thick and are light yellowish brown (10YR 6/4), oxidized
263-275; Transition zone to less oxidized sediment; silt is progressively grayer; very fine sand is brighter (yellowish brown (10YR 5/8), oxidized
275-291; Silt loam, grayish brown (10YR 5/2) with common black (N 2.5/0) smudges [Equality Formation, cont.]
291-305; Core loss
305-430; As above, with brightly oxidized layers at 318-320 cm and 338-340 cm
430-457; Core loss
457-563; As above [Equality Formation, cont.]
563-610; Core loss
610-650; As above, but formed folds due to sediment resistance to sliding up tube during sampling; fold amplitudes as least 10 cm [Equality Formation, base]

650-662; Gravel, cobbles, with sandy matrix, brown (10YR 5/2) [unnamed tongue, Henry Formation]
662-692; As above, but light olive brown (2.5Y 5/3); abrupt, planar lower boundary [Henry Formation, cont.]

692-765; Diamicton; clay loam matrix, brown (8YR 5/2), hard, calcareous, uniform [Tiskilwa Formation]
End of boring

SUMMARY:
0-79 cm; Modern soil in loess (Peoria Silt)
79-122 cm; Sandy loam diamicton
122-650 cm; Silt and very fine sand; laminated ice-proximal lacustrine sediment; oxidized and fossil-poor from 122-275 cm; fossiliferous from 275-650 cm)
650-692 cm; Sand and gravel outwash
(79-692 cm: Equality Formation Complex)
692-765 cm; Diamicton, uniform; clay loam matrix; till (Tiskilwa Formation)

Figure 6 Representative core segments from site H-22 sampled with a PowerProbe. Core diameters are about 5 cm.

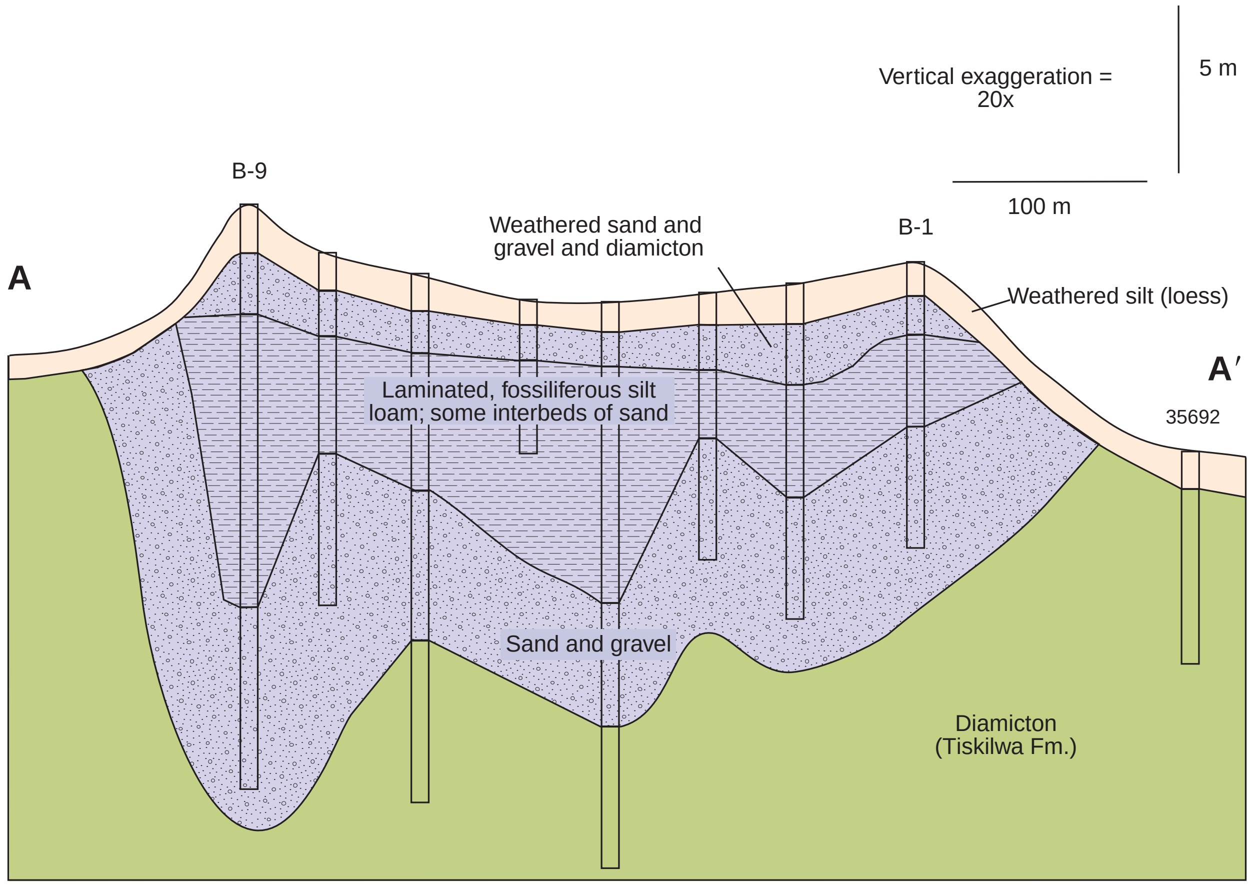


Figure 7 Cross section across part of an ice-walled lake deposit. The line of section is shown in Figure 1c.

sandy loam diamicton or sand and gravel, and (5) weathered silty loess (Figures 6 and 7). The high resolution resistivity profiles distinguished between the diamicton, silt-rich units (units 2 and 3), and sandy unit 4, but was unable to detect unit 1 (the basal sand and gravel), and the mantle of loess (Figure 8). Sediment cores of the other ice-walled lake plains reveal nearly identical sediment successions, although they are floored by younger diamictons, including the Batestown, Yorkville, Haeger, and Wadsworth units of Hansel and Johnson (1996).

Unit 2 contains only fossils of wood fragments. The fine cellular structure and presence of resin indicates that the wood fragments are from boreal trees. A radiocarbon age of 24,950 ± 150 yr BP indicates that most, if not all, of the wood fragments are reworked from the Robein Member (Roxana Silt).

The fossiliferous laminated silt of unit 3 also contains fragments of boreal trees, but they are less common than valves of ostracodes, head capsules of chironomids, aquatic plant seeds,

Table 1 DeKalb mound sizes classified by shape. (L = long axis; W = short axis)

	Round		Elliptical		Irregular	
	W (m)	L (m)	W (m)	L/W	W (m)	L (m)
Average	84	122	74	1.7	138	276
Median	66	104	66	1.6	114	208
Maximum	312	824	578	7.0	1003	1354
Minimum	38	38	28	1.1	47	66
Number	54		170		87	

Figure 8 High-resolution resistivity profiles across the DeKalb mound visited at Stop 7. The lower profile approximately follows the line of section for the cross section in Figure 7, which has been stretched to approximately match the line of section of the resistivity profile. The 'hottest" colors – magenta and red – are interpreted to be deposits of sand and gravel. Most domains with warm colors (orange, and yellow) are interpreted to be clay loam diamicton of the Tiskilwa Formation. Cool colors of green and blue are deposits of silt loam, silty clay loam, and silty clay. Some structures in the shallow sediment are attributed to drainage tiles.

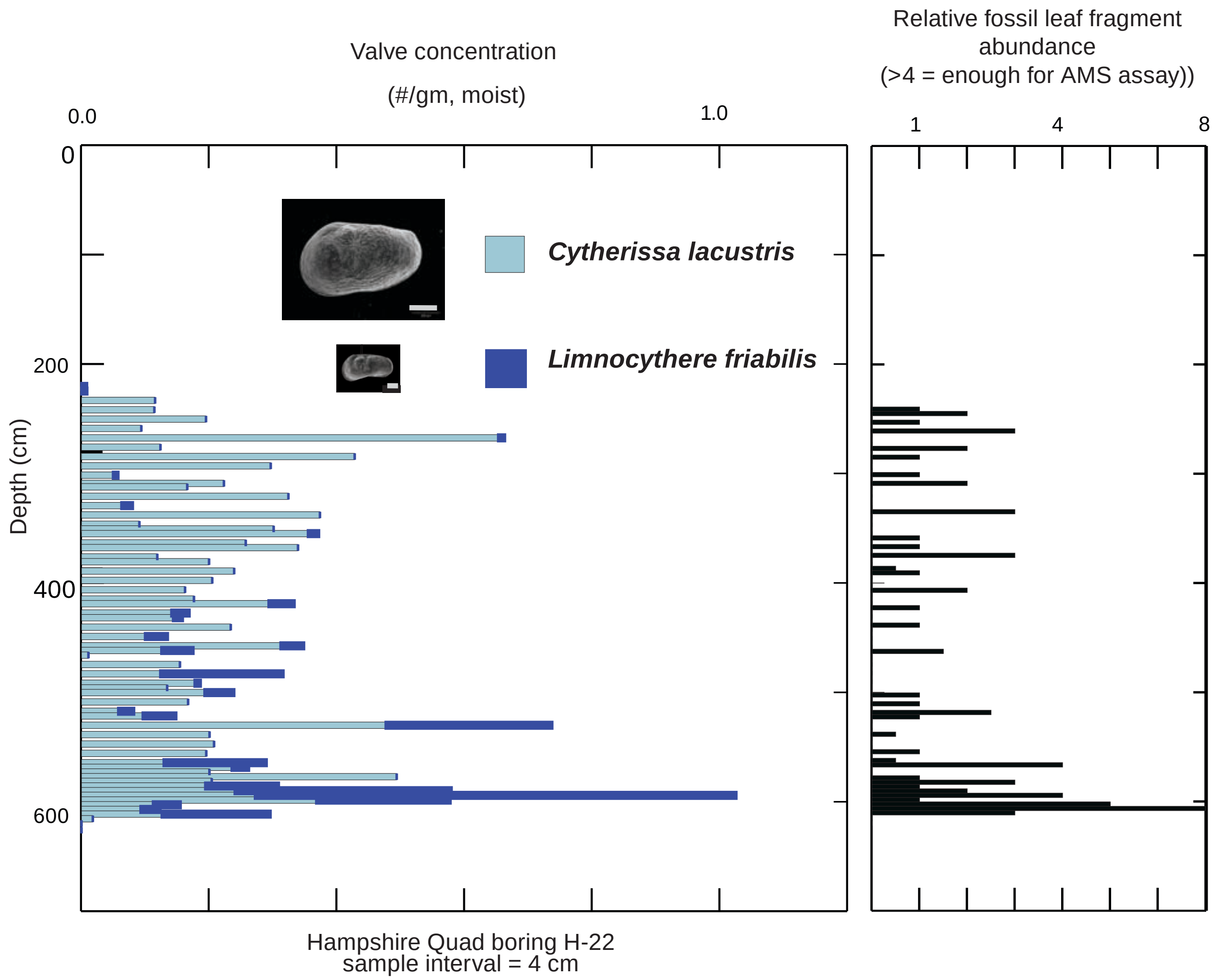


Figure 9 Profiles of ostracode abundance (left) and plant fragments (right) from 4-cm long subsamples of core H-22.

resting cocoons of flatworms (Tuberellia), shells of pillclams (*Pisidium* sp.), tundra plants, and beetle elytra. The older mounds (e.g., ones located in the "type" area, Woodstock Moraine, and on the Marseilles Morainic System) contain tundra plant fossils of Arctic bilberry (*Vaccinium uliginosum* ssp. *alpinium*) and snowbed willow (*Salix herbacia*), whereas plant fossils from younger mounds on Lake Border moraines are dominated by stems and leaves of Arctic dryad (*Dryas integrifolia*). The fossil plant remains tend to be concentrated at the base and at the top of laminated unit 3 (Figure 9). As of yet, needles and cones of boreal trees have not been identified from DeKalb mound sediments in Illinois. Sediment deformation and core loss incurred during sampling precludes precise counting of the laminae which may be varves.

The ostracode assemblage identified from core H-22 includes *Cytherissa lacustris*, *Limnocythere friabilis*, and uncommon *Heterocypris* sp. Valve concentration ranges from nil to 1.4 valves per gram dry sediment, with abundant *L. friabilis* at the base, and near-constant abundance of *C. lacustris* (Figure 9). Additional species have been identified from other mounds (Table 2). Especially common are valves of *Fabaeformiscandona rawsoni* and *F. caudata.*

In core H-22, XRD clay mineral analyses of unweathered laminated silt yielded values of about 14% expandable clay minerals, 75% illite, 12% kaolinite, and 8% chlorite, nearly identical to the

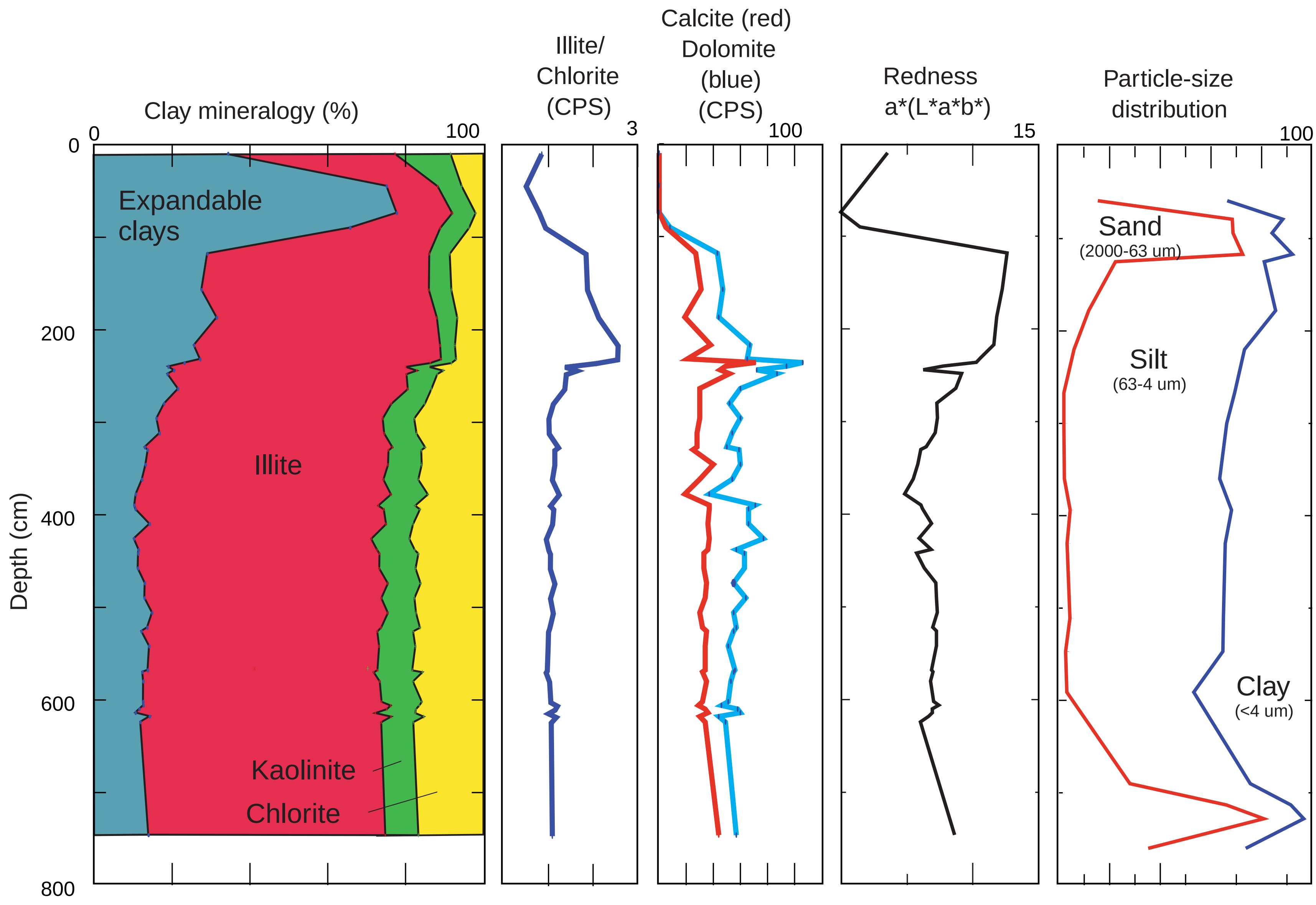


Figure 10 Semi-quantitative clay mineralogy, and calcite and dolomite abundance (in CPS, counts per second) of the < 0.002 mm fraction calculated from X-ray diffractograms of oriented, aggregate slides (ethylene glycol-solvated) from core H-22. Also shown are profiles of the illite/chorlite ratios, redness (a*) (L*a*b*), and particle-size distribution determinations.

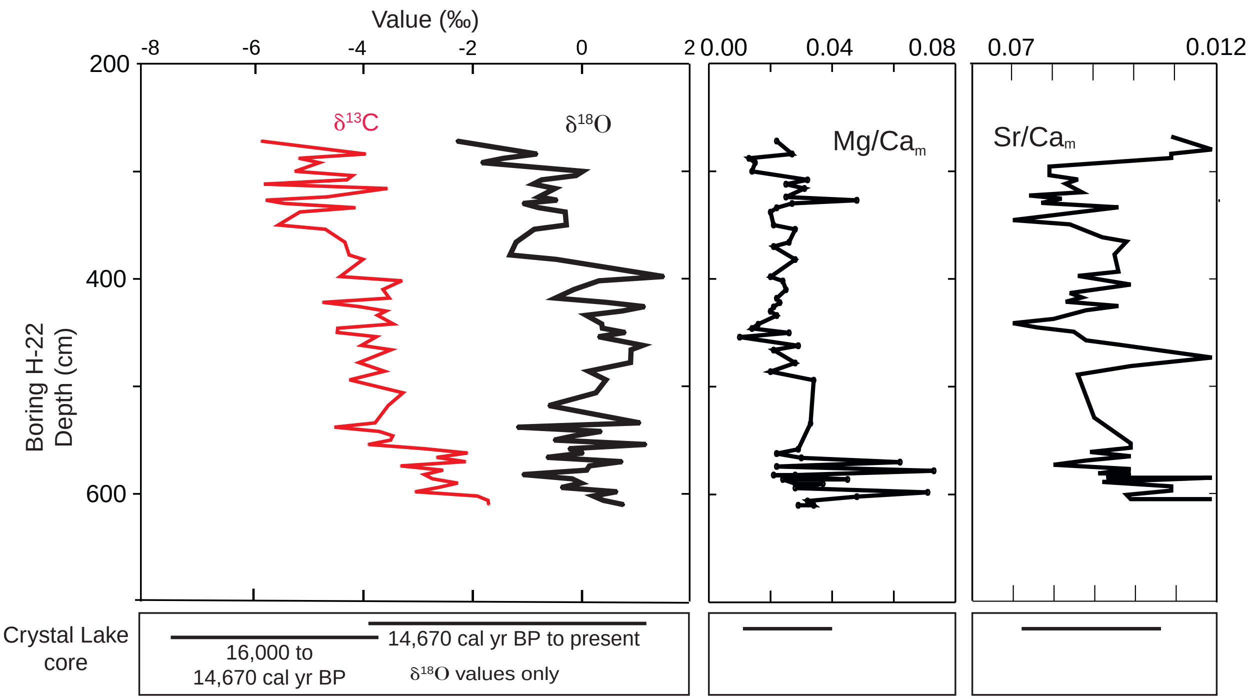


Figure 11 Profiles of biogeochemical profiles of valves of *Cytherissa lacustris*, including $\delta^{13}C$, $\delta^{18}O$, $Mg^{2+}/Ca^{2+}{}_m$, and $Sr^{2+}/Ca^{2+}{}_m$ from boring H-22. The range of values from a sediment core from Crystal Lake (Figure 3) is shown except for $\delta^{13}C$. The $\delta^{18}O$ values of *Candona ohioensis* have adjusted +0.65 ‰ to account for the greater vital effect compared to *Cythersisa lacustris* (von Grafenstein et al., 1994, 1999)

mineralogy of the subjacent glacial diamicton, and dissimilar to the mantle of smectite-rich loess (Figure 10). Loss of chlorite, high illite/chlorite ratios, loss of carbonate minerals, and greater sample redness are all indications that the upper 1.5 m or so of the laminated fines are weathered. These sediments contain abundant sesquioxide concretions, and contain no fossils.

The biogeochemical profiles of *Cytherissa lacustris* from boring H-22 shows ranges of -5.9‰ to -1.7‰ for $\delta^{13}C$,–2.3‰ to +1.5 ‰ for $\delta^{18}O$, 0.0071 to 0.012 for $Sr^{2+}/Ca^{2+}{}_m$, and 0.008 to 0.073 for $Mg^{2+}/Ca^{2+}{}_m$. The isotopic profiles reveal upward trends towards "heavier" values, whereas the trace element ratios reveal no trends (Figure 11).

DISCUSSION

The combined evidence of cross sections and modern analogs of the ostracode assemblages indicates that the DeKalb mounds are deposits of ice-walled lakes. Cross-sections across the largest mounds (not shown) suggest ice thickness was at least 8 m; the ostracode evidence, presented below, suggests depths of at least 15 m. These estimates do not support the theory of mound genesis through degradation of pingos as promoted by Flemal et al. (1973). Water depths in modern degrading pingos are usually < 4 m; the greatest water depth estimated for deposits attributed to Pleistocene pingos in western Europe is about 12 m (DeGans, 1988, and references therein). The sediment successions in the European paleo-pingos are composed of primarily biogenic sediment (gyttja, marl, and peat), whereas the De Kalb Mounds contain little biogenic sediment. Likely modern analogs for the De Kalb mounds are the ice-walled lakes of the Siberian Taymyr Peninsula formed in slowly melting remnants of the Kara ice sheet (Figure 12; ≈ 75 º N, 100º E; Alexanderson et al., 2002). Here, the tundra-plant laden soil cover is thin

(< 2 m), and susceptible to slumping (see guidebook cover). Bathymetric maps reveal that some lakes are flat-bottomed, and have steep slopes that drop abruptly to more than 14 m deep (Figure 12). Other more-often photographed modern ice-walled lakes are located on stagnant arms of active mountain glaciers with supraglacial debris tens of meters thick, such as the Bering Glacier, Alaska. Although thick supraglacial drift was not present during formation of the DeKalb mounds, such conditions have been interpreted in areas peppered with high-relief ice-walled lake plains on the North American prairies (i.e., Pirazek, 1970; Ham and Attig, 1996).

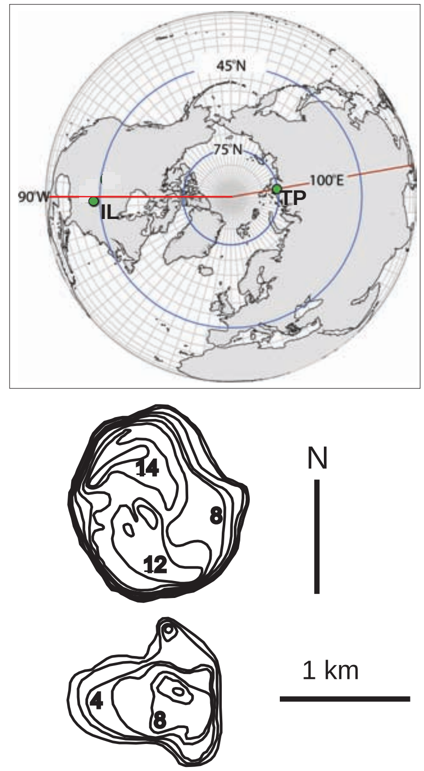


Figure 12 Bathymetric maps of two ice-walled lakes from the Taymyr Peninsula, Siberia. Contour interval = 2 m. On the inset map, TP = Taymyr Peninsula and IL = Illinois.

The assemblage of *Cytherissa lacustris* and *Limnocythere friabilis* is known from Lake Michigan in water >15 m deep (Buckley, 1975). *Cytherissa lacustris* is also common in Canadian lakes containing water with low total dissolved solids (TDS; 196-365 mg/L) that are > 3 m deep (Delorme, 1989) although they have been identified in shallow water of such deep lakes (Forester, personal communication). Importantly, DeKalb mounds from outside the "type" area have yielded other assemblages, including Cytherissa lacustris, *Fabaeformiscandona rawsoni*, *F. caudata*, and *Limnocythere varia* (Table 2) that collectively are also known from the profundal zone of Lake Michigan (Buckley, 1975). Only *Candona subtriangulata* is missing from the profundal assemblage, perhaps because the salinity of the ice-walled lakes was too great for this species to tolerate (Forester et al. (1994).

Much source water of the ice-walled lakes was likely derived from precipitation. Today, water evaporated from the Gulf of Mexico accounts for about 75% of the precipitation received in the Upper Midwest (Simpkins, 1995). The $\delta^{18}O$ values of lakes in northeastern Illinois typically range from about -5 to +2‰; shallow groundwaters typically range from -9 to -5‰. The other likely water source, glacier meltwater, had $\delta^{18}O$ values estimated to be from -24‰ to -17‰ (Remenda et al., 1994; Dettman et al., 1995). The $\delta^{18}O$ values of adult *Cytherissa lacustris* in core H-22 range from -2.3 to+1.5‰; given this species vital effect of about +1.5‰ (von Grafenstein et al., 1999), the values of the host water ranged from +0.0 to -3.8‰. This suggests that the water in the ice-walled lakes was derived from precipitation rather than meltwater. Moreover, the range of ostracodal $\delta^{18}O$ values in boring H-22 fall within the range of $\delta^{18}O$ values yielded by fossil valves of the ostracode *Candona ohioensis* from Crystal Lake, Illinois (location shown in Figure 3). The age range of these analyses is from 14,670 cal yr BP to the present. Interestingly, the $\delta^{18}O$ values from Crystal Lake dating from 16,000 to 14,670 cal yr BP are much "lighter" from the younger samples. This earlier period was likely when the water was colder and possibly influenced more by groundwater or possibly meltwater.

Radiocarbon ages associated with ice-walled lakes allow a more robust assessment of deglacial history than previously possible. Bottom ages from the laminated, fossiliferous silt loam facies indicate when glacial conditions shifted from active to stagnating (about 17,500 C-14 yr BP for the DeKalb mounds). Top ages provide minimum age estimates of the period of stagnation. The span of time between the latter ages and those ages from the base of kettle sediment succesions encompass the period of final melting. In the Hampshire area, this period was between about 15,150 and 14,000 C-14 yr BP (Curry, 2008). Ages from several localities on the Woodstock Moraine indicate the period lasted from about 14,900 to 13,900 C-14 yr BP. This span of time represents an important topographic reversal during deglaciation. At the onset, ice-walled lakes were the lowest parts of the glacial landscape; later, kettles formed in the low spots of the deglaciated landscape that included terraces formed of lacustrine deposits that had accumulated in the ice-walled lakes. A concomitant response to the inversion of topography is a reversal in the direction of shallow groundwater flow.

Our core studies, modern analog analyses of fossil ostracodes, and high-resolution resistivity profiling indicate that DeKalb mound sediments were deposited in ice-walled lakes. The best evidence of this is the laminated sediment with deep-water ostracode assemblages that extend the edge of the mounds (Figures 7 and 8). The lake bottom was irregular with meters of relief such that lake sediment thickness varies from nil to more than 10 m (Figure 7). Below the loess mantle, mounds are capped on the sides and on the top by a continuous 1 -2 m-thick layer of sand and gravel and sandy loam, matrix-supported diamicton. It is important to note that the laminated, silty, fossiliferous sediment in the mounds contains little sediment coarser than coarse sand, and yet it separates deposits of sand and gravel that contain little silt.

The architecture and sediment texture discussed above suggests the following set of processes during the ontogeny of a DeKalb mound. These scenarios are consistent with the formation of “perched” ice-walled lake deposits as envisioned by Clayton (1967) and others (e.g., Syverson, 2007). The first step was to form the accommodation space (“hole”) occupied by the ice-walled lake as suggested by Ham and Attig (1997). Formation of the “hole” likely developed from a surficial “seed” lake; once the lake formed, its expansion was enhanced through the latent heat released when water freezes. The second step was for a period of stabilization when the lake was seasonally open to allow terrestrial plants to be washed in to the lake. In order to keep coarse material from the ice walls from falling into the lake, conditions were likely colder than the initial period of formation. The source of the silty material probably was not from debris in the ice-walls of the lake, but rather from suspended sediment that was perhaps discharged into the lake by way of subterranean springs. The provenance of the suspended sediment was likely the debris-rich ice that was thermally and mechanically eroded by water in englacial conduits with access to seasonal atmospheric precipitation. This accounts for the similar clay mineralogy of the lake sediment and underlying diamicton, as well as the isotopically "heavy" $\delta^{18}O$ values of the lake water as indicated by analyses of ostracode valves.

The third and final step in the ontogeny of the DeKalb mounds was final collapse. The special conditions that allowed the ice-walled lakes to slowly fill with fossiliferous lake sediment abruptly changed such that the once-stable ice walls melted back. Saturated, coarse debris eventually spread across the shallow frozen lake. During this final stage, permafrost may have formed in the upper 1-2 m of the lake sediment which may account for some of the remarkable features observed on aerial photography, such as doughnut-in-doughnut features (Flemal et al., 1973; Ianacelli, 2003).

In summary, the ecology and age of terrestrial and aquatic fossils of the De Kalb mounds indicate they formed in the karstified, stagnating Lake Michigan lobe during the last deglaciation. Endogenic fossils of DeKalb mounds in Illinois include *Cytherissa lacustris, Limnocythere friabilis*, *L. varia*, *Fabaeformiscandona rawsonia*, and *F. caudata*, an ostracode assemblage known only from profundal Lake Michigan (Buckley, 1975). Exogenic fossils of tundra plants preserved in mound successions provide some temporal control on the onset and duration of glacier stagnation.

Table 2 Location and ostracode fauna from ice-walled lake plains located in Illinois. Thickness refers to the thickness of the complex of sediments comprising the mounds (i.e., the Equality Formation complex shown on Figure 7), including the mantle of lloess.

Moraine	7.5-minute map	N Lat	W Long	diameter (km)	thickness (cm)	CYTH	LFRI	HET	FRAW	FCAU	LVAR	CYCA
Deerfield	Wadsworth	42.43	-87.90	0.872	640	C	A	NP	NP	C	NP	NP
Tinley	Wadsworth	42.44	-87.97	0.642	698A	A	NP	NP	C	NP	NP	
West Chicago	Streamood	42.02	-88.20	0.8741	716A	A	R	NP	C	R	R	
Burlington	Hampshire	42.05	-88.57	1.17251	624	A	A	R	NP	NP	R	NP
Marseilles (N)	Newark	41.50	-88.54	0.6059	853	A	A	NP	C	C	R	NP
Marseilles (S)	Dwight	41.08	-88.49	0.3151	411	NP	A	NP	A	NP	NP	NP
Gilman	Gilman	40.82	-87.99	0.5166	411	C	C	NP	A	NP	NP	NP
Champaign	Royal	40.17	-87.98	0.5496	381	NP	A	R	C	NP	NP	C

Qualitative abundance per ca. 200 gm sample (moist): A = abundant (hundreds of valves); C = common (tens); R = rare (single valves); NP = not present

CYTH = *Cytherissa lacustris,* LFRI = *Limnocythere friabilis,* HETI = *Heterocypris incongruens*(?), FRAW = *Fabaeformiscandona rawsoni,* FCAU = F. *caudata, LVAR = L. varia; CYCA = Cyclocypris ampla.*

Stop 8: Spring Lake Sand and Gravel Pit

Glaciotectonic deformation of Wisconsin Episode sediment: shallow décollements, sheath folding, and corroborating strain indicators in overlying diamicton

I. C. Higuera-Diaz, J. Stravers, and D. Kulczycki

Introduction

Sediments underlying moving ice sheets undergo deformation processes due to the shear stresses caused by local ice pushing, dragging, and differential rates of movement and friction between the ice, bed, and substrate. Such stress is amplified at glacier margins where mass balance of ice flux necessitates compressive flow. However, there is still discussion about the relationship between directional strain indicators and paleo-iceflow directions. Clast macrofabric analysis has been used to evaluate deformation attitudes in subglacial sediment (Bennett et al., 1999; Hooyer and Iverson, 2000; Boulton et al., 2001; Benn, 2002; Iverson and Hooyer, 2002). Macrofabric analysis of deformed subglacial diamictons have shown correlation between prolate clast orientation and glacier-induced shear stress direction and intensity (Phillips and Auton, 2008; Van der Wateren et al., 2000). Complementary to macrofabric studies, micromorphological and microstructural analysis can provide insights into complex deformation histories in subglacial sediments (Phillips and Auton, 2000; Menzies, 2000; Thomason and Iverson, 2006; Hooyer et al., 2007). Evidence of deformation of unlithified subglacial diamicton has been described at sites worldwide and across the geologic record (Banham and Ranson, 1965; Hart, 1997; Thomas, 1984; Owen, 1988; Owen and Derbyshire, 1988; Kluiving et al., 1991; Matoshko, 1995; Van der Wateren, 1999; Pedersen, 2000; Van der Wateren et al., 2000; Bennett et al., 2000; Lian et al., 2003; Möller, 2006).

A complete understanding of the relationships between paleo-iceflow direction and bed or substrate deformation requires three-dimensional mapping of folds, joints, and faults, as well as measurement of sediment fabrics. The sediments exposed in man-made exposures are ideal for making statistically significant and verifiable analyses of the geometry and kinematics of sediment deformation, as well as determining the directions of paleo-iceflow that were likely the stresses responsible for subglacial deformation. Geophysical data, such as the seismic reflection data of McBride et al. (2007), may reveal interesting glaciotectonic features deeper in the sediment succession.

The study site is located at the Spring Lake sand and gravel pit, Spring Grove, Illinois (42.43ºN, -88.24ºW; Figure 1). The quarry is located on the edge of the Ringwood Upland, a geomorphic landsystem element in McHenry County (Curry et al., 1997). The upland is bounded on the east by an ice-contact moraine (the Fox Lake Moraine of Willman and Frye, 1970), and on the other sides by anastomosing surficial channels with underfit drainage, possibly tunnel valleys, that were formed during the earliest phase of the Wisconsin glaciation (Curry et al., 1997).

Previous work

The succession of deformed sediments revealed at the Spring Lake pit was first mapped by Stravers et al., (2001). Although much has been written about the lithostratigraphy and environments of deposition of the region's glacial deposits (Willman and Frye, 1970; Hansel and Johnson, 1996; Curry et al., 1997) there are few published analyses of the glaciotectonic structures in this region. The seismic reflection data of McBride et al. (2007), located about 25 km east of the Spring Lake pit, indicates a north to south-trending décollement which occurred along the drift/

bedrock interface. Moreover, they interpreted low-angle listric faulting of glacial drift of the Wadsworth Formation, the unit forming the Valparaiso Morainic System, Tinley Moraine, and Lake Border moraines to the east. Layers of folded sand and gravel deposits with diapiric intrusions of silty clay diamicton (the Yorkville Member of the Lemont Formation) have been observed at the Meyers sand and gravel pit immediately west of the municipality of McHenry. This site is located on the southern margin of the same upland element as the Spring Lake pit (Figure 1).

In this investigation, we determine the directional stress fields that produced structural elements observed in a sequence of folded and faulted proglacial and subglacial sediment of the last glaciation, and compare these directional data with pebble fabrics in the overlying diamicton. The stratigraphic units and deformation involved in our analysis are older than those described by McBride et al. (2007). Ostensibly, the direction and intensity of paleostresses imparted on the now-deformed sediment were controlled primarily by paleotopography, local ice flow velocity, and the conditions at the base of the glacier — whether the bed was frozen onto the substrate or was sliding along a film or layer of water. The latter factor likely changed over time as the ice margin first crossed the area and eventually reached its terminal position at the Woodstock Moraine. Our analyses show (1) a correlation between ice-flow direction and the orientation of the shear stresses causing the deformation, and (2) that deformation occurred in a system with a strong coupling along the ice-sediment interface and ductile deformation in the underlying outwash sediments.

Methods

Except for the mantle of Peoria Silt, and diamicton of the upper part of the Haeger Member, all sedimentary units exposed in the highwalls of the Spring Lake quarry are deformed. Subunits of sorted sediment of the Beverly Tongue (Henry Formation) and diamicton of the Haeger Member (Lemont Formation) are described based on lateral continuity and inferred mechanical behavior. These units were mapped in detail throughout the quarry along with a detailed census of structural features and attitudes. We recorded approximately 300 structural data from bedding surfaces, joints, faults, axial and foliation planes. At each pebble fabric station located in Figure 2, we measured the orientation of 25 prolate clasts > 2 cm long within a horizontal circle 1 m wide. Although some studies use larger numbers of clasts in fabric analyses (Hart and Smith, 1997; Benn, 1995; Huddart and Hambrey, 1996), our eigenvector analyses are statistically significant by using 25 measurements as prescribed by Lawson (1979). The strength and character of our eigenvector plots are similar to those of the Haeger Member by Hansel and Johnson (1986) who also used 25 measurements.

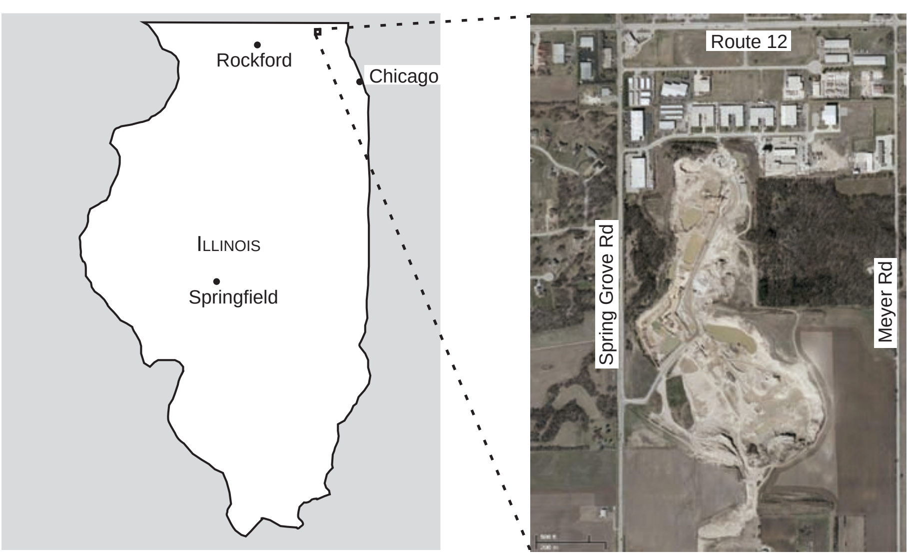


Figure 1 Location of the project area in northeastern Illinois. Detailed satellite image of the Spring Lake Quarry taken from Google Maps.

Stratigraphy and Mechanical Stratigraphy

The stratigraphic succession below the floor of the quarry is not known, but a nearby boring (MC-13; Curry, 1995), also located on the Ringwood Upland, reveals about 95 m of glacial drift including from bottom to top: a.) pre-Wisconsin Episode sand and gravel about 10 m thick; b) undifferentiated sorted sediment and reddish-brown diamicton about 10 m thick; c) laminated silt, silt loam, and fine to medium sand more than 30 m thick that are part of the Equality Formation; d.) sand and gravel about 10 m thick that form the Beverly Tongue of the Henry Formation; e) sandy loam diamicton with interbeds of sand and gravel about 10 m thick that are part of the Haeger Member (Lemont Formation) and, f.) silt loam and silty clay loam about 1 m thick (Peoria Silt). To the west of the Spring Lake quarry, diamicton of the Tiskilwa Formation becomes an important component in the stratigraphic succession (Wickham et al., 1988; Curry et al., 1997). Tiskilwa diamicton, if it occurs at depth at the quarry, may correlate with the second unit from bottom to top in the MC-13 boring. Only the four upper units identified in the MC-13 boring are found and were mapped in the high walls of the quarry.

The units shown in Figure 2a were defined by their inferred contrasting mechanical competence, and are lithotectonic units (Wood and Bergin, 1970; Fischer and Jackson, 1999). Our geometrical and kinematical analyses are based on these units. The Equality Formation is the oldest unit exposed in the mapped area, and is composed of laminated gray to brownish-red clays and silty clays. Due to the limited exposures of this unit, it was not possible to characterize its mechanical behavior from field data. However, our structural analyses suggest this is a mechanically weak unit that serves as a basal detachment for deformation.

Overlying the Equality Formation is the Beverly Tongue of the Henry Formation which we subdivide into: a) the lower gravel (HBlg), a matrix-supported gravel with clast sizes ranging from granule to pebble (Figure 3d). Matrix composition is mostly sandy with traces of silt and clay. Mechanically this is a strong unit that formed large folds tens of meters across without developing obvious structural micro- or mesofabrics; b) the gravel and sand (HBgs) package is the thickest subunit of the Beverly Tongue (Figures 3a, f). It consists of meter-scale fining-upwards successions of pebbles, gravel, and medium to fine-grained sand. Mechanically, the interlayering of thin, strong gravel beds with thicker, weak sand beds resulted in the unit as a whole behaving as a weak unit. Meter scale disharmonic folding is common with short throw (< 1 m) reverse faulting in the gravel beds; c) the sand unit (HBs) is stratified medium to fine sand. Mechanically, this unit is the weakest among the Beverly Tongue. Disharmonic folding at small to large scales, foliation, and jointing are common features of this unit. Eye-shaped folds exhibit sheath folding (Alsop et al., 2007) within this unit (Figures 3b, c); d) the upper gravel (HBug) is clast-supported gravel with particle sizes ranging from granule to cobble. Although the grain-to-grain contacts of large clasts would suggest a strong unit, it behaved as a weak unit.

The sand and gravel of the Beverly Tongue is unconformably overlain by two subunits of the Haeger Member. Both subunits are composed of sandy loam diamicton, but the upper unit (LHud) contains lense-shaped inclusions of silt-coated diamicton and attenuated lenses of sand. Deformation of the lower diamicton subunit (LHld) is coupled with the underlying Beverly tongue. The upper subunit is not involved with deformation, and is considered to be rigid.

Structural Domains

The mapped area can be divided in four structural domains with characteristic deformation patterns (Figure 2b). The northernmost domain involves the two upper lithotectonic units of the Beverly Tongue and the two units of the Haeger Member. This domain is the highest structural level revealed in the quarry. Folds have meter-scale wavelengths and amplitudes, with axial

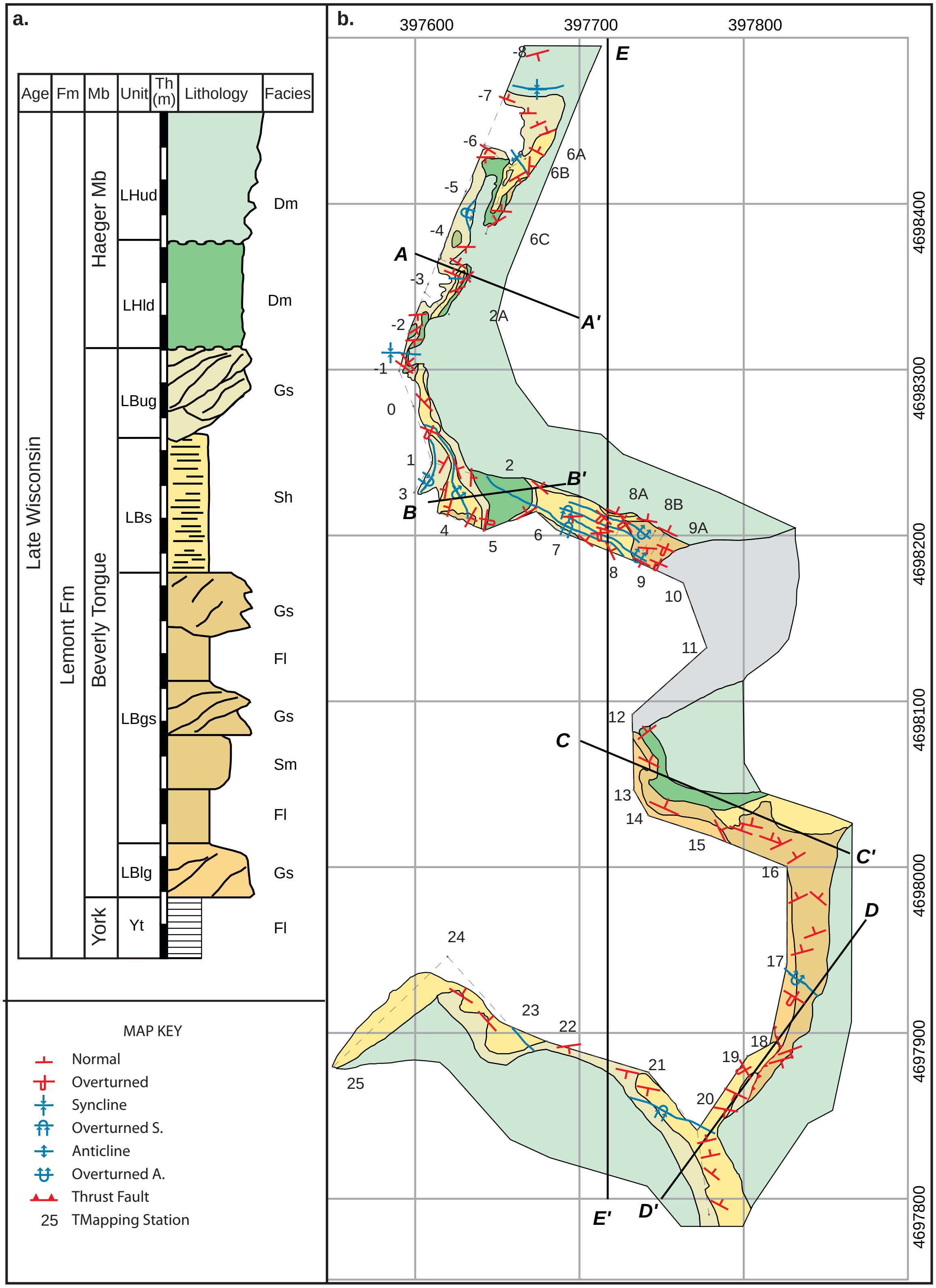


Figure 2a Stratigraphic units exposed at the Spring Lake pit (after Hansel and Johnson, 1996). Subunits are defined in part by their likely mechanical behavior during folding. **Figure 2b.** Detailed geological map of pit walls showing the location of representative field-collected data, sampling stations, and structural cross-sections location.

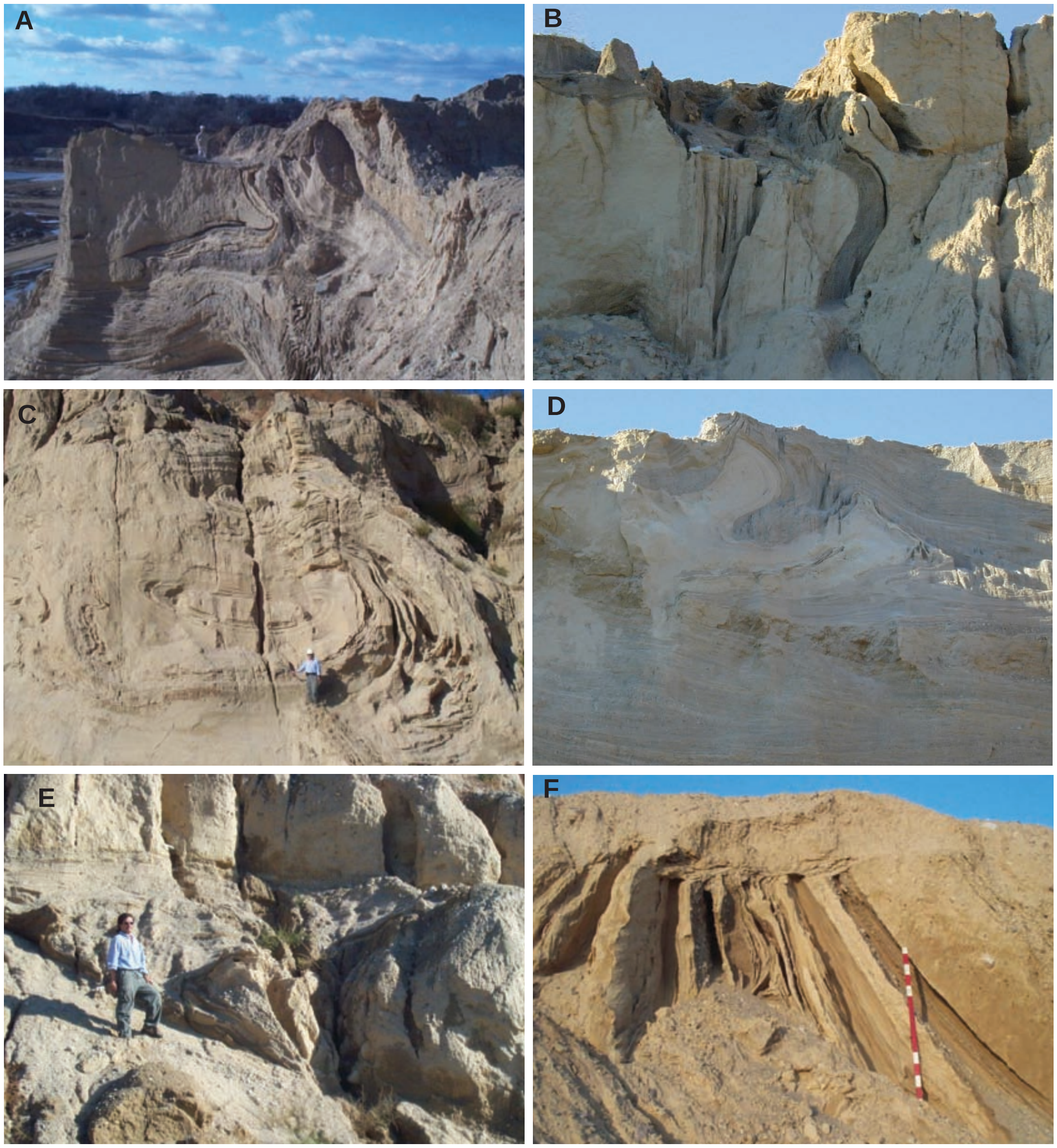


Figure 3 Characteristic deformation of unconsolidated sand, gravel, and diamicton. Station locations are shown on Figure 2b. **a.)** Overturned anticline involving the upper gravel of the Beverly Tongue (HBug) and the lower diamicton of the Haeger Member (LHld). The anticline is beheaded by the contact with the upper diamicton of the Haeger Member (LHud). In the upper half of the picture note the overturned and nearly vertical lower diamicton (LHld) cut by the horizontal upper diamicton (LHud). Picture taken from the east towards station D 12**.** **b.)** Subvertical anticline in the upper gravel of the Beverly Tongue decapitated by the upper diamicton of the Haeger member (LHud). The lower diamicton (LHld) was removed from the crest of this anticline. Picture taken from the east towards station D 6. **c.)** Section cutting through the noses of a succession of sheath folds in the middle sandy member of the Beverly Tongue (HLBs). Picture taken between stations D 7 and 8. **d.)** Thrust faults and folds above a detachment in the gravel and sandy members of the Beverly Tongue (HBgs). Both the thrust and the fold axial planes dip towards the north in this picture. Picture taken between stations D 16 and 18; photo courtesy of B. Curry). **e.)** Cross-section of sheath folds involving the lower diamicton of the Haeger Member (LHld), and the upper gravel of the Beverly Tongue (LBug). Picture taken near station D 2. **f.)** Drag features caused by the overriding the of the upper diamicton of the Haeger Member (LBud) on the vertical layers of the sandy unit of the Beverly Tongue (HBs) and the lower diamicton of the Haeger Member (LBld). Picture taken near station D 13.

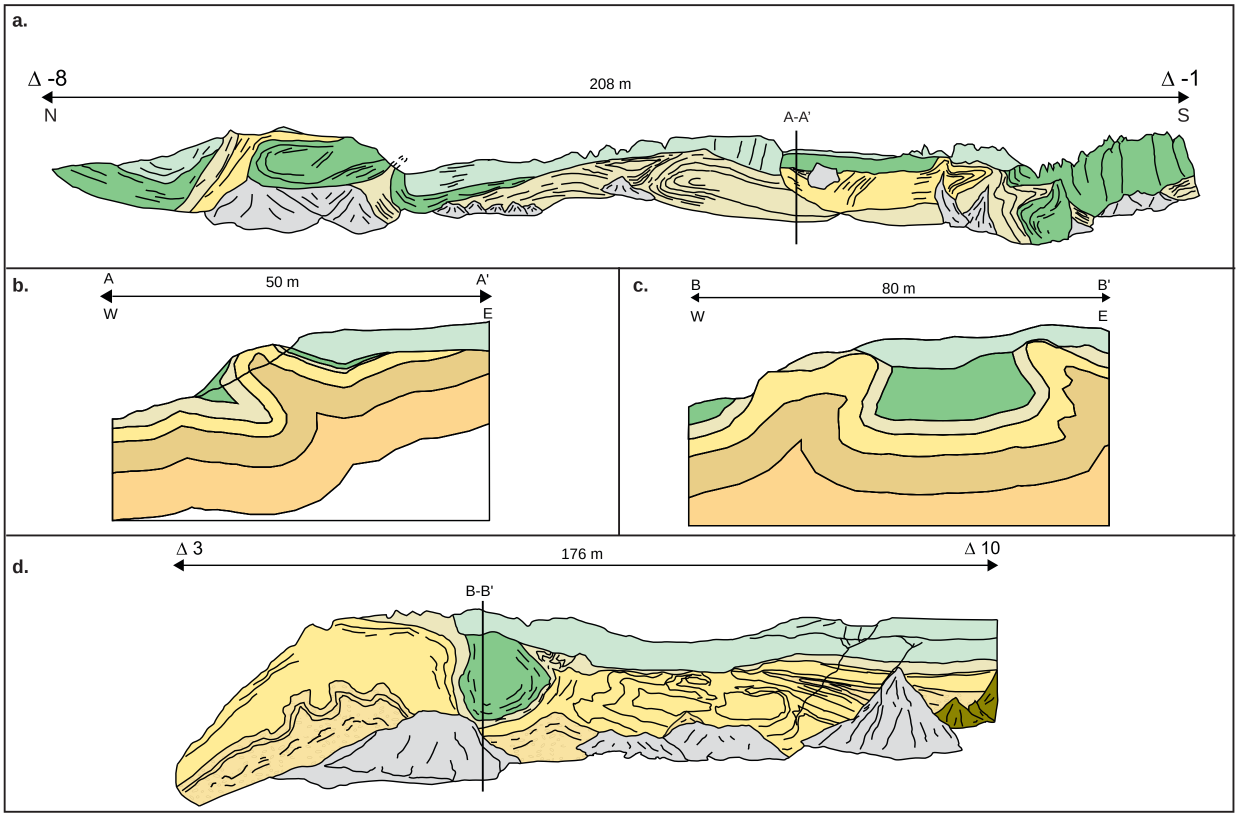


Figure 4 Detailed wall profiles and structural cross sections of the northern half of the Spring Lake quarry. For wall profiles and structural cross sections locations refer to Figure 2b, and for color codes of the units refer to Figure 2a. Gray units are modern slumps. **a.)** Meter-scale wavelength folding of the Beverly Tongue, and the lower diamicton of the Haeger Member. The upper diamicton dip is nearly horizontal and lies on top of the deformed sequence. **b.)** Cross-section through a meter-scale fold with an axial plane that dips to the east. **c.)** An isoclinal fold involving the lower diamicton of the Haeger Member, and the Beverly Tongue. **d.)** Meter-scale wavelength folding and sheath folds in the Beverly Tongue, and the lower diamicton of the Haeger Member. This wall is cut along the strike of the sheath folds and perpendicular to the main fold axis.

planes verging (angle respect to a vertical plane) towards the west (Figure 4b), and fold axes trending north to south. The lower diamicton (LHld) is involved with folds parallel to the Beverly Tongue (Figure 4a) whereas the lower contact of the upper diamicton is a nearly horizontal thrust fault that cross-cuts several features such as anticlines (Figures 3a,b,d,f).

The north-central domain reveals deformation of most lithotectonic units. The domain involves two structural levels: an upper level that is a continuation of domain A, and an intermediate structural level with subhorizontal folding (Figures 4c, d). The amplitude and wavelength of folds in this level ranges from meters to tens of meters with horizontal axial planes. Axial planes of non-horizontal folds in this level trend east-west and verge southwards.

The south-central and southernmost domains (C and D in Figure 2) involve all lithotectonic units, and reveal a third, deeper structural level (Figure 5). The shallow and intermediate structural levels occur in these areas as well. Unique characteristics of the southern domains are the lack of small scale folding in the rigid units, and major south-verging thrust faulting in the deepest levels of the Beverly Tongue.

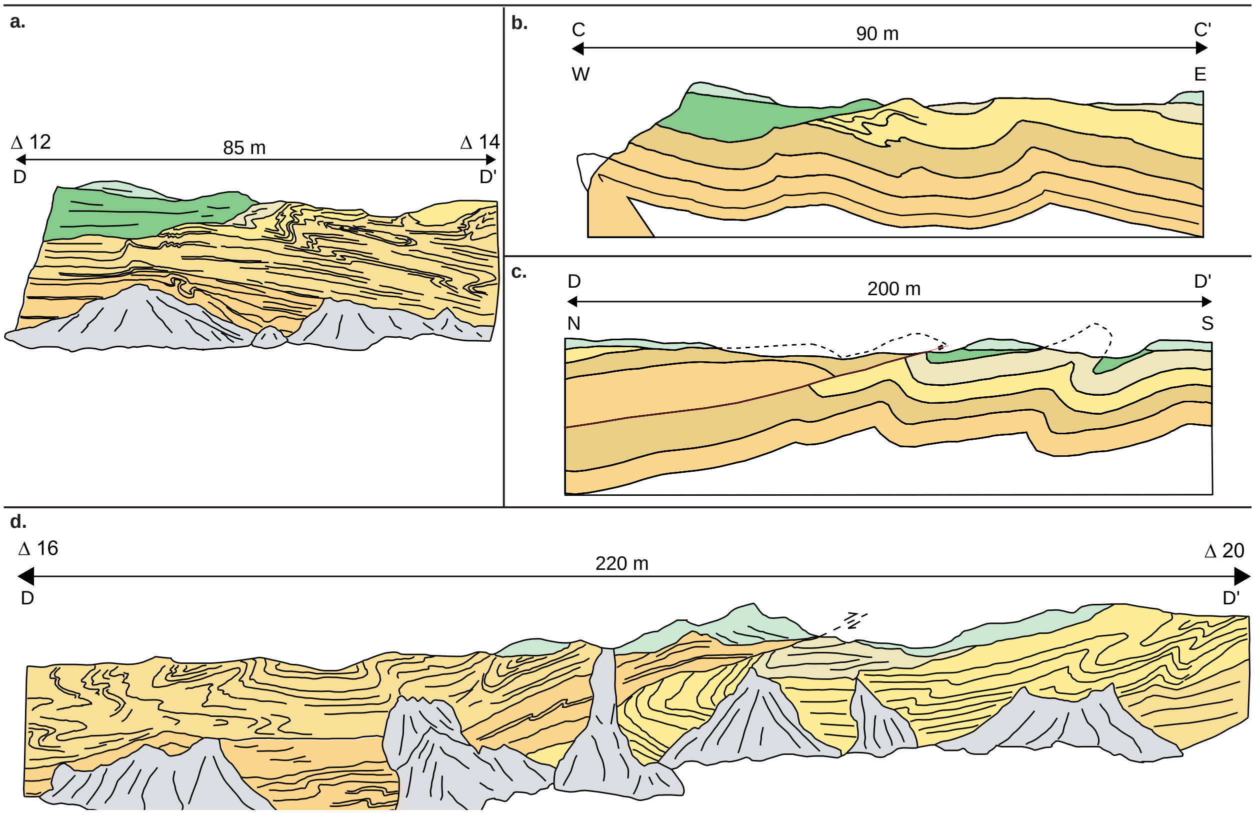


Figure 5 Detailed wall profiles and structural cross sections for the southern half of the Spring Lake quarry. For wall profiles and structural cross sections locations refer to Figure 2b, and for color codes of the units refer to Figure 2a. **a** and **b)** Disharmonic short-wavelength folds of the Beverly Tongue. **c** and **d)** South-dipping thrust fault with associated disharmonic folding in the hanging wall, and folds with wavelengths of ≈ 10 m or more in the footwall.

Structural Analysis

To determine the orientation of the paleostresses responsible for the deformation described of the subglacial sediments, we used direct inversion analysis of fold geometry, and orientation of axial planes, fault planes and joint sets. Our analyses show that each structural domain has elements with similar orientations. Fold axes orientation determined from a ϖ diagram trend NEE to SWW (Figure 6a). Consistent with these data, a wind rose analysis of the axial planes of 24 individual folds show a mean orientation of 230° (Figure 6b). Five of six joint planes indicate a main stress of 221° (Figure 6c). This value is statistically insignificant, but is given because the results are in step with the fold data analysis. An analysis of 24 faults planes indicate that the maximum principal stress was oriented 262° with a dip of 52° (Figure 6d).

Clast fabrics

Fabric data were measured at five stations each for the lower diamicton unit, and the upper diamicton unit of the Haeger Member (see Figure 2a for station locations). At each station we measured plunge and bearing of the long axis of 25 prolate clasts. Across the mapped area, fabrics from LHud stations show a consistent NW-SE trend (Figure 7a). The fabrics shown in Figure 7b are of folded layers of LHld; Figure 7c shows these fabrics rotated such that bedding

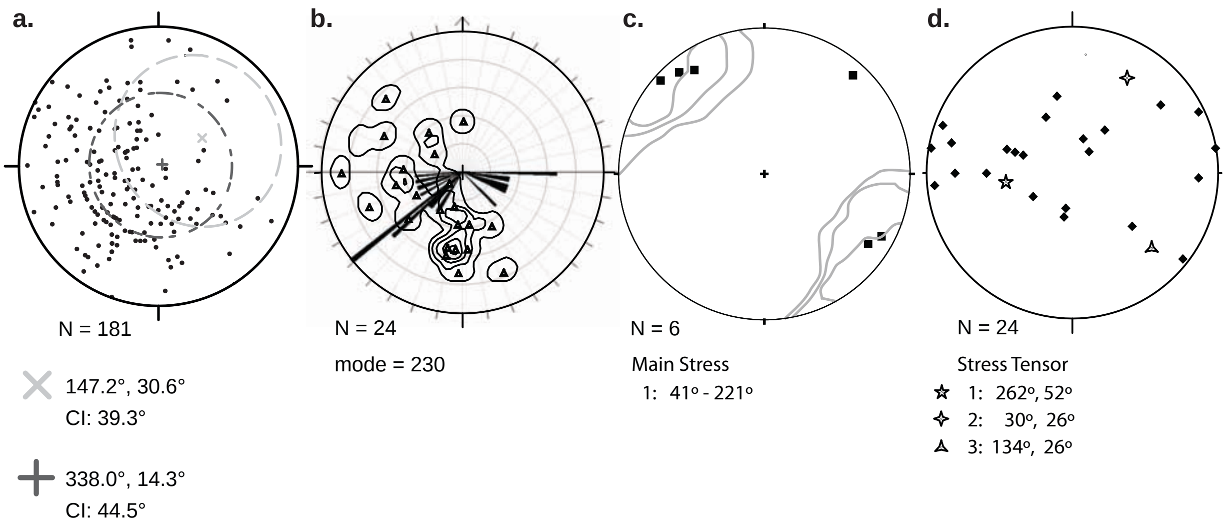


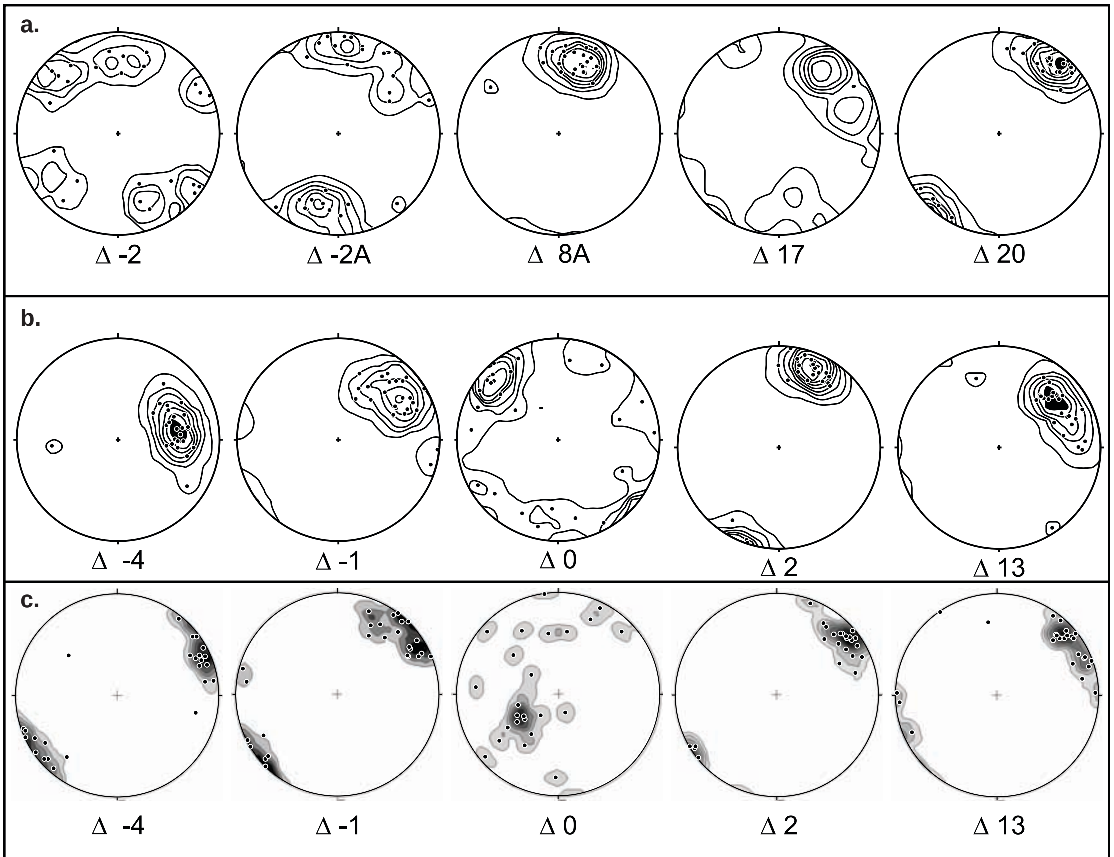
Figure 6 Structural analyses of deformation. **a.)** Equal area lower hemisphere stereographic projections of poles to bedding for the SLQ. P- and B diagram indicate the folding is conical. **b.)** Wind rose for the azimuths of the axial planes of metric-scale wavelength folds with a mode orientation of 230˚. **c.)** Equal area lower hemisphere stereographic projections of poles to joints. **d.)** Equal area lower hemisphere stereographic projections of poles to faults. Stress tensors responsible for the mean fault plane orientation were determined by multiple iterations of direct inversion techniques.

is horizontal. The transformation produces stronger fabrics, and a slight to no change in fabric orientation. We examined the statistical robustness of the fabric dataset by using Bingham axial distribution analyses to determine the eigenvalues and eigenvectors for every station (Figure 7d). These values are used to determine if our sediment fabrics are consistent with other evidence of paleoice- flow direction. Moreover, pebble fabric data are routinely used to interpret the depositional setting of subglacial sediments (Benn, 1994, 1995; Bennet et al., 1999). Other insights into kinematics are possible; for example, the ratio between E1 and E3 eigenvectors provides a measure of fabric isotropy, and the E2:E3 ratio indicates fabric elongation. The pebble fabric data from Spring Grove quarry show that at most stations, the lower and upper diamicton units have high E1 values indicative of stong preferred prolate pebble orientations (Figure 7e).

Discussion

Paleoice flow and paleo stress directions

Our collective field data indicate that biaxial stresses imparted simple shear near the base of the glacier resulting in folding and faulting of proglacial sand and gravel, and beds of loam diamicton. Orientations of axial planes, joints and faults indicate that maximum principal stresses were oriented along two main trends: 221˚ to 230˚, and 262˚ (Figure 6). The direction of ice-flow determined from fabrics in the lower diamicton shows two distinctive families trending along 36º-216º and 90º-270º and two slightly different families for the upper diamicton units trending along 49º-229º and 63º-243º. Since the upper diamicton is not involved in folding, we suggest that its deposition occurred in a later event not related to the deformation of the lower diamicton unit and subjacent sorted sediment. The correlation between main stresses and paleoice flow directions for the deformed sediments indicates that deformation occurred during deposition of the lower diamicton. Hence, we attribute folding of Beverly Tongue sediment was due to simple shear applied by the ice depositing the Lower Haeger Member. The younger ice that deposited



d.

STATION	E1	E2	E3
UPPER DIAMICTON (LHud)			
-2	0.5499	0.2844	0.1657
-2a	0.8450	0.0968	0.0582
8a	0.8808	0.0963	0.0228
17	0.8359	0.1157	0.0484
20	0.9213	0.0548	0.0239
LOWER DIAMICTON (LHld)			
-4	0.8380	0.0977	0.0633
-1	0.8457	0.1242	0.0300
0	0.5510	0.3587	0.0903
2	0.5536	0.3025	0.1439
13	0.8100	0.1722	0.0178

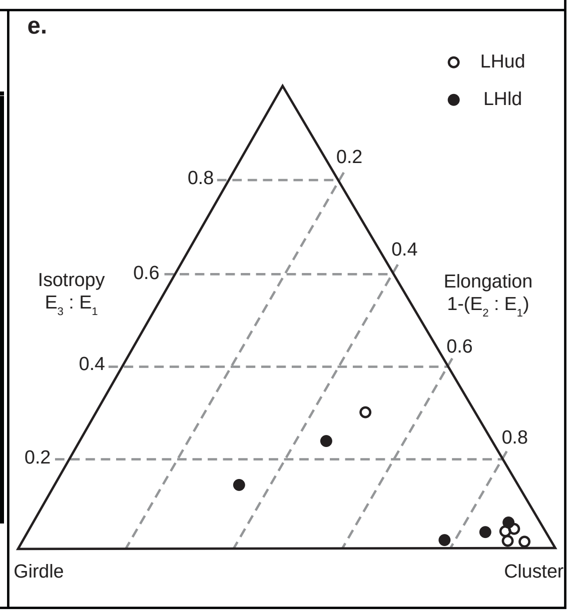


Figure 7. Equal area lower hemisphere stereographic projections of poles and contouring of principal axes of prolate casts in the upper diamicton **(a)** and the lower diamicton of the Haeger Member **(b)**. The diagrams in the bottom row **(c)** show the results of rotating the clasts of the lower Haeger Member (7b) to the horizontal, thus eliminating the effects of folding, and showing the pre-folding direction if ice-flow. **d.)** Table summarizing the eigenvalues of clasts fabrics for every station in the lower and upper diamicton. **e.)** Triangular eigenvalue plot after Benn (1994) and Bennett et al. (1999) emphasizing the importance of clast-fabric shape and showing that our fabric data is non-isotropic and strongly clustered.

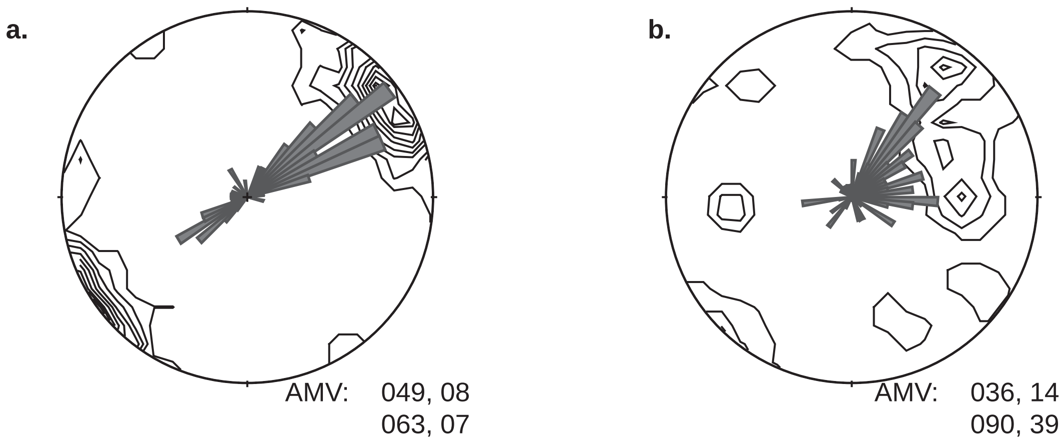


Figure 8 Summary of fabric data shown in Figure 7. **Figure 8a** shows the combined data for the fabrics shown in Figure 7c (the lower diamicton unit), and **Figure 8b** shows the combined data shown in Figure 7a (the upper diamicton unit). Each diagram shows the superposition of the equal area lower hemisphere projection contouring of poles with the wind rose diagram of the azimuths of clasts.

the upper diamicton was not as strongly coupled to its substrate and did not induce further folding. Instead, the debris-rich ice eroded some of the anticline crests formed during the previous folding. These results are in step with recent work in glaciotectonics indicate that subglacial deformation is caused by simple shear at the base of the glacier of the base of the ice sheet the overriding the sediments (Alsop et al., 2007).

In a simple simple shear regime clasts should be oriented parallel to the shear surface giving strong clustered fabrics (Hicock et al., 1996). The orientation and strength of fabric will also be affected by bedrock topography and the thickness of the deforming layer. In deformation tills formed by the deformation of unconsolidated sediment beneath a moving glacier, the style of fabric will depend on the style of deformation (Bennet, 1999). In deforming subglacial sediment, clast fabric resembles the cumulative strain ellipsoid of the deforming matrix through rotation (Benn, 1995). In places with strongly defined boundaries the ellipsoid is closely constrained causing the fabric to be non-isotropic and strongly clustered. Alternatively, fabric strength in deforming subglacial glacier beds may be attributed to the thickness of the deforming bed and strain magnitude (Hart, 1994). A thin layer should provide strong fabrics in the direction of deformation whereas a thick layer should provide weak fabric strengths.

Folding geometry and kinematics

Strong coupling along the interface between the basal debris-rich ice was needed for the subglacial sediments to be dragged and folded. This does not necessarily imply that the bed was completely frozen onto the sorted sediment; significant shear can be applied at the base of a glacier with a continuous film of water if the bed roughness is great enough, which has not been observed or measured here. Another important variable is the ice flow velocity, with higher velocities imparting higher shear stresses, all other variables being equal. Although the folded sediment is impressive here, it is important to realize that the actual amount of strain (bed shortening) is much less than the strain that deformed the diamicton during its deposition at the base of the glacier. In addition, as the sediment deformed, void spaces likely enlarged as particles rotated against one another, leading to lower pore water pressure, and greater bulk strength of the deforming body. The increase in strength may have been a primary cause for the cessation of folding process.

In the weaker units, several characteristics of the folded sand units are consistent with sheath folds (Figure 9). Many folds of sandy beds at the Spring Lake quarry are conical (Figure 6a) and appear as bull's-eyes in some quarry walls (Figures 3c and 4d). Another feature attributed to sheath folds are folds with sub-horizontal axial planes (Figures 4d and 5d). Sheath folds are flattened ellipsoidal features with stacked interior folds that close along both the base and the top of the fold. Both the outer and interior folds show double vergence similar to boudinage (Alsop and Holdsworth, 2004; Alsop et al., 2007; Searle and Alsop, 2007). The importance of the sheath folds is that it indicates that the weaker beds in the deforming sequence became realigned such that they no longer show geometries indicative of a fluvial sedimentary regime, but had been transformed to a structural fabric consistent in concept with the aligned pebbles in the overlying diamicton.

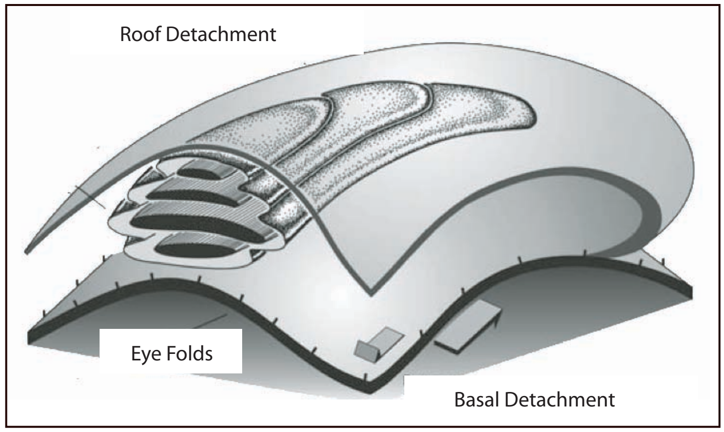


Figure 9 Three-dimensional model of a sequence of sheath folds showing the characteristic eye folding and wrapping of the eye folds by basal and roof detachments. Modified from Searle and Alsop, (2007).

In the Spring Lake quarry folding, fracturing and internal strain seems to be confined to a deformation sheet bounded by basal and roof detachments, respectively (Figure 10). Neither, the basal, or the roof detachments were observed in the field. However, by using established area-balance restoration techniques (Epard and Groshong, 1993) it was possible to determine the position of the depth-to-detachment for the basal décollement. The roof detachment was most likely near the ice sediment interface, and since we have observed deep erosion by the upper diamicton (Figure 4a) we suggest that it may have been completely eroded away.

We constructed a balanced and restored cross-section to estimate the minimum shortening that sediment layers underwent in this area. We assume parallel folding for all layers and that for the short extent of the quarry the layer thickness remain constant. Our restoration technique uses

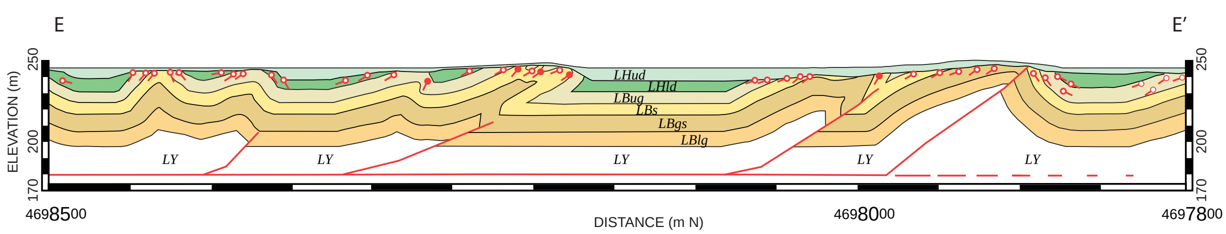


Figure 10 Balanced and restored cross-section of the sediment sequence at the Spring Lake quarry. The line of section is shown in Figure 2b. Only the southernmost thrust fault was field-mapped. Locations of the other thrusts faults are interpretative and based on geometric and fault kinematics criterion. The basal detachment is located in the Equality Formation and the depth to-detachment was estimated using area-balancing methods. The roof detachment of the sheath-fold complex is eroded away, and corresponds to the surface between the early Haeger ice, and the lower diamicton. Restoration using a constant line-length method shows a maximum shortening of 22 %. The section pinpoint was located along the vertical axial plane of the northernmost syncline. The inclined southern end of the cross-section indicates an increasing shortening from top to bottom due to the shear stress and the difference of the coupling between the basal and roof detachment.

a method that measures the line-length of the deformed layers and transfers those lengths to a template where all layers are horizontal. Hence, layer shortening is estimated. This technique accounts for deformation due to folding and faulting, but does not include internal deformation such as layer flattening or stretching due to sediment compaction, reaccommodation, or internal shear strain. Our shortening estimates indicate that maximum shortening was about 22%, with somewhat lower values occurring lower in the field section. There is less deformation in the conglomerates.

The geometry of a basal detachment and multiple thrust faults rooting to it that we interpreted for our structural cross section is similar to the interpretation of shallow seismic reflection lines by McBride et al., (1997). Although deformation occurred was interpreted to have occur in different (younger) stratigraphic units at greater depths that in this study, collectively, our results indicate that glaciotectonic deformation associated deposits of the Lake Michigan lobe should be expected.

Acknowledgments

We would like to thank Jeff Thurlwell owner of the Spring Lake Quarry for allowing us access to the quarry. We gratefully acknowledge reviews by B. Curry and J. Thomason whose suggestions greatly improved and made this paper readable. This research is part and was supported by of the ISGS-USGS EDMAP.

I am also indebted to Paradigm Geophysical, Inc. for the free use and support of the Geosec2D™ software. Structural analyses of fabric data were done using Stereoplot© a software written by R. Allmendinger, and Daisy© a software written by F. Salvini.

Stop 9: Thelen Sand and Gravel Pits

Sedimentology of kame terrace deposits at the Thelen sand and gravel pits, northwestern Lake County, Illinois

Christopher Stohr, Timothy Kemmis, Andrew Stumpf, Jason Thomason, Brandon Curry

Introduction

Deposits forming the Fox Lake Moraine and a kame terrace west of the Fox River valley near Spring Grove, Illinois, are important aggregate resources in northwestern Lake County. Investigation of excavations in two large sand and gravel surface-mines at Illinois Highway 173 and Wilmot Road (Figure 1) focused on the sedimentary characteristics and the environment of deposition of surficial outwash sediments attributed to a proglacial delta. Using a reflectorless total station (RTS), bedding characteristics were described with a lithofacies code, and bedding plane altitudes were mapped along a 3.5 kilometer-long profile. Use of the RTS allowed geologists to describe from a safe distance the ca. 27m-high walls exposed in the pits.

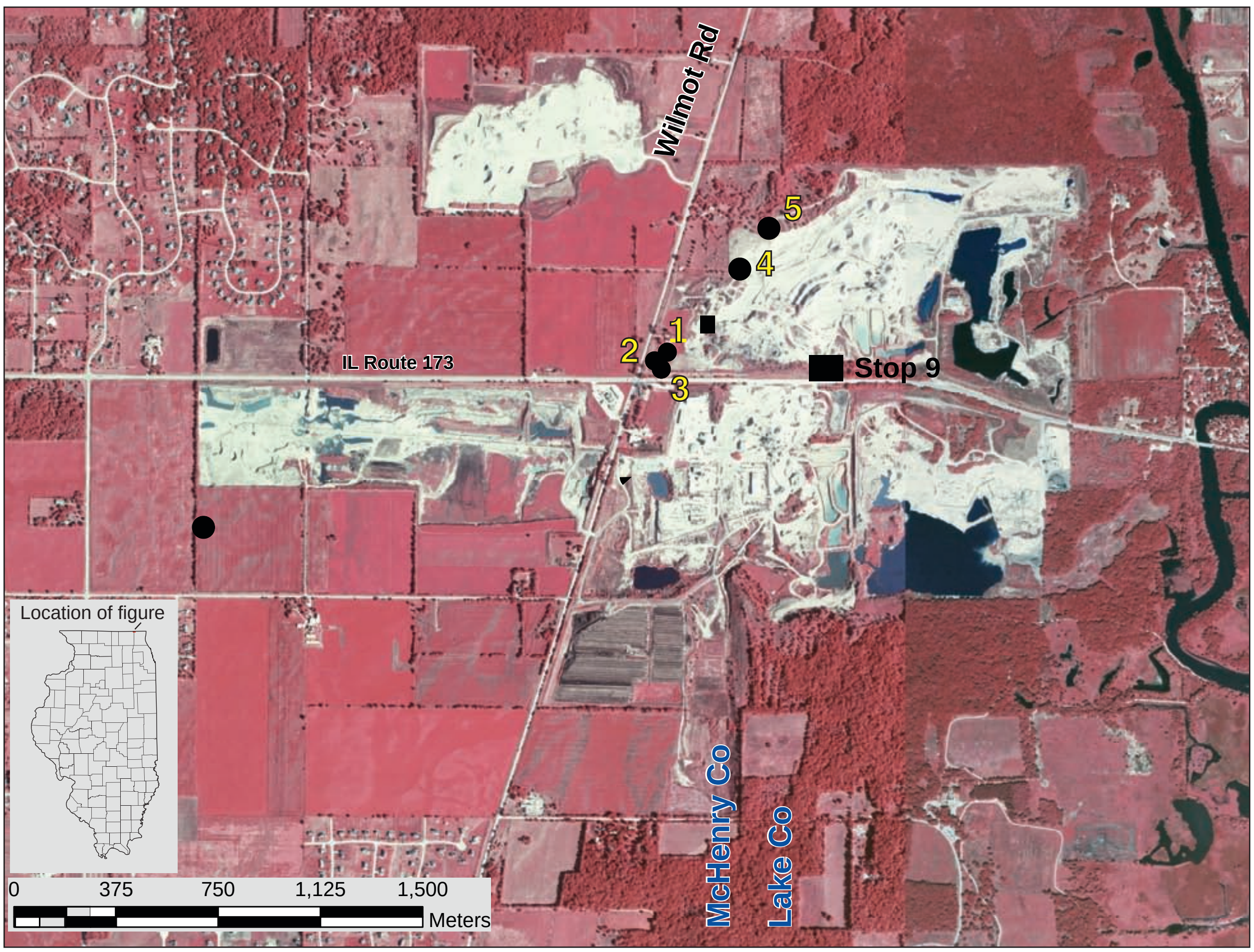


Figure 1 National Agriculture Imagery Program (2004) color-infrared orthophotograph showing the location of profiles surveyed at the Thelen pits, and the approximate location of Stop 3. The image is located in northwestern Lake and northeastern McHenry counties, Illinois.

Figure 2 Equipment used for surveying.

Quadrangle-scale mapping indicates that the deltaic sediment is younger than the folded beds of sand and gravel and diamicton observed at Stop 8. The deltaic sediments were deposited during an early (earliest?) part of the Milwaukee Phase as the Harvard sublobe paused at the position of the Fox Lake Moraine prior to melting back as far north as Milwaukee, Wisconsin. Later, the Lake Michigan lobe readvanced as the Joliet sublobe, forming the Valparaiso Morainic System immediately east of the Fox River during the Crown Point Phase.

Methods

A reflectorless total station (RTS) (Figure 2) permitted detailed descriptions of six profiles as much as 27 m high at a safe distance from the working face of a sand pit. Bed thicknesses and unit contact elevations were surveyed to a precision less than 3 cm. The georeferenced data can be used in 3-D maps and models.

The RTS sends a pulsed laser to a target (Kemmis et al., 2005) but unlike conventional instruments, a reflective prism is not needed at the surveyed point. Ground control for the RTS surveys was made using high-precision, dual-frequency Leica 399 GPS receivers processed with US National Geodetic Survey's Online Positioning User Service. Two vertical control monuments at the Thelen study site were established with 3 cm and 9 cm precision; horizontal control precision was 0.02m N and 0.09m E, and 0.02m E and 0.04m N.

Each unit was described using a lithofacies code (Table 1) based on estimation of grain size, sorting, and sedimentary structures and bedforms related to flow-regime as observed visually and through the RTS telescope. The lithofacies code permitted rapid description of the successions. The altitude of every bedding plane in the six profiles was surveyed. Each profile was measured, described, and discussed by our team in about two hours. Binoculars were helpful for team interpretation and discussion. The recording of absolute X, Y, and Z coordinates for

Table 1 Lithofacies code for fluvial and glaciofluvial deposits (modified from Kemmis et al. 1988)

GROSS PARTICLE SIZE – first symbols	
BG	boulder gravel
CG	cobble gravel
PG	pebble gravelcm clast-supported
ms	matrix-supported
cm	clast-to-matrix supported
S	sand
F	fines
BEDDING STRUCTURES6 - second symbols, in parentheses	
(m)	massive
(pl)	planar-bedded; crudely horizontal, may be slightly undulatory
(h)	horizontally laminated; may be slightly undulatory
(r)	ripple-drift cross-laminated (various types)
(t)	trough cross-bedded; size (scale) and single or multiple sets noted on log
(w)	wedge cross-bedded; size (scale) and single or multiple sets noted on log
(p)	planar cross-bedded; size (scale) and single or multiple sets noted on log
(c)	cross-bedded deposits with complex upper and lower contacts; generally large-scale, solitary sets: lower contacts commonly undulatory over irregular channel floor; upper contacts commonly undulatory, truncated by overlying bedding structures
(la)	lateral-accretion deposits
(ccf)	channel cut-and-fill; massive or simple structures mimicking the scoured channel cross-section
(ccfc)	channel cut-and-fill structure with complex facies changes within the fill (see Ramos and Sopena, 1983)
(ccft)	channel cut-and-fill structure with transverse fill (see Ramos and Sopena, 1983)
(ccfms)	channel cut-and-fill structure with multi-storey fill (see Ramos and Sopena, 1983)
(lag)	lag at base of channel or cross-bed set
(g)	normally graded
(ig)	inversely graded
(n-i)	normal to inversely graded
(i-n)	inversely to normally graded
(l)	low angle (<10o) crossbeds
(e)	erosional scours with intraclasts
(s)	broad shallow scours
(sc)	laminated to massive fines
(-t)	various bedding structures comprising deltaic topset beds
(-f)	various bedding structures comprising deltaic foreset beds
(-b)	various bedding structures comprising deltaic bottomset beds

each bed or unit contact (Stohr, 2007) is a geospatial registration that permits comparison of elevations with correlated contacts at other sites to calculate the slope of boundary surfaces, and geometry of large-scale bedforms, such as channels. Establishment of the three-dimensional characteristics of major features, such as kame-deltas, is a valuable contribution to multisource profiles and 3-D models.

Geology of the Spring Grove Kame Terrace

Regional Geology and Geomorphology

The Thelen site is located on the eastern edge of a dissected plain that dips south towards Nippersink Creek (Figure 3). The western edge of the terrace abuts against a semicircular scarp about 15 m high that steps up to the Ringwood Upland of Curry et al. (1997). On the east, the plain abuts against the Fox Lake Moraine and thence to the lowlands of the Chain 'O Lakes Lowland, a drop in elevation of more than 40 m in some places. This relief reflects minimum ice thickness during formation of the moraine. In McHenry County, the upland is formed of sandy loam diamicton of the Haeger Member (Lemont Formation) and proglacial outwash (Beverly Tongue, Henry Formation) that was visited previously at Stop 2. The succession observed at the Thelen site is inset into deposits of the Ringwood Upland. The upland succession was formed during the Woodstock glacial phase; the terrace deposits, described below, were formed during the earliest part of the Milwaukee lake phase.

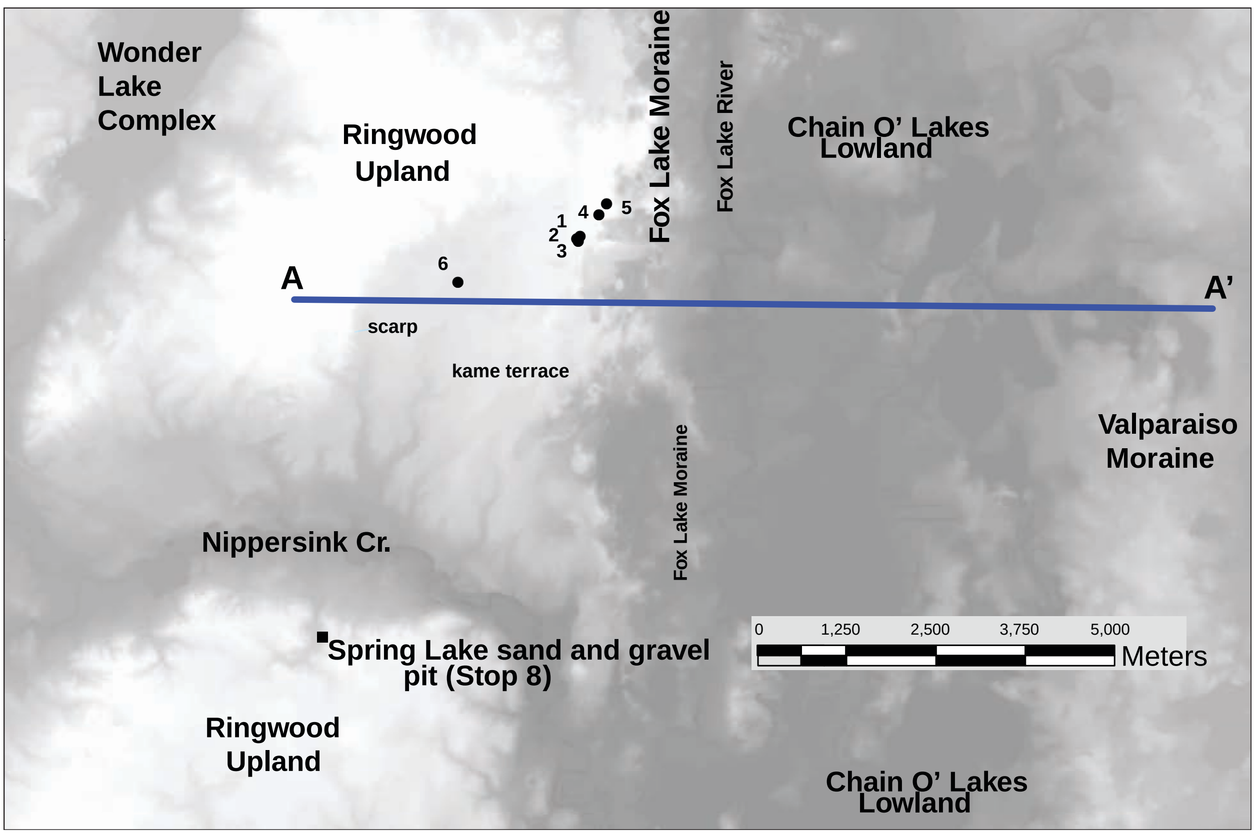


Figure 3 Digital elevation model of 30-meter USGS elevation data showing the Spring Grove kame terrace west of the Chain O Lakes lowland, Fox River, and the Valparaiso Moraine. The scarp of the terrace is shown inset into part of the Ringwood Upland. The Wonder Lake Complex (including the valley of Nippersink Creek) is a series of anastomosing channels that may have been exploited to drain early Glacial Lake Milwaukee. The line of section, A-A', is for Figure 4. The location of surveyed sections (1-6) is shown along with the location of figure 5 (x).

The erosion of the large semi-circular scarp marks a fundamental change in the behavior of the Harvard sublobe. Prior to scarp erosion, the sublobe had overridden and deformed proglacial outwash and diamicton (discussed at Stop 8) perhaps as ice advanced west and south to form the Woodstock Moraine. As the Harvard sublobe retreated into the Chain O' Lake lowland, deposition was primarily by meltwater. In the area of the Thelen site, meltwater eroded along the edge of the retreating ice to form the semicircular scarp and low area later filled by a proglacial lake correlated to Glacial Lake Milwaukee. The unconformity left by this period of erosion is irregular, with some erosional remnants of older, finer-grained material preserved immediately south of the Thelen pits (Jack Thelen, personal communication). Ostensibly, base level for this period of erosion was the channels of the Wonder Lake complex, an anastomosing network of subglacial channels inset into the Ringwood Upland that includes the valley associated with underfit Nippersink Creek (Figure 3; Curry et al., 1997). This suggests that the direction of flow in the channel of Nippersink Creek was the opposite of what it is today.

As meltwater formed the scarp, the retreating Harvard sublobe stalled to form the Fox Lake Moraine. A proglacial lake extended west from the morainic front to the scarp, leaving a depositional basin exploited by at least one proglacial delta, described in detail below. The Fox Lake Moraine is unusual for Illinois because it is formed in most places of stratified, folded, and faulted deposits of sand and gravel, as well as laminated and stratified silt and sand (Figure 5). Most other moraines in Illinois are formed primarily of diamicton (till) with less sand and gravel outwash than the Fox Lake Moraine.

Geologic Transects

Excavations at the Thelen pits reveal that the sand and gravel is composed of a deltaic sequence prograding west away from the former glacial margin marked by the Fox Lake Moraine (Figure 4). Sediments grade from cobbles and boulders to sand, but little sediment finer than fine sand is observed. The deltaic sediments are bounded on the distal (west) side by an upland similar to the kame terrace setting described by Fleisher (2003) rather than the valley ice tongue-type described by Embleton (1987) and Eyles (1983).

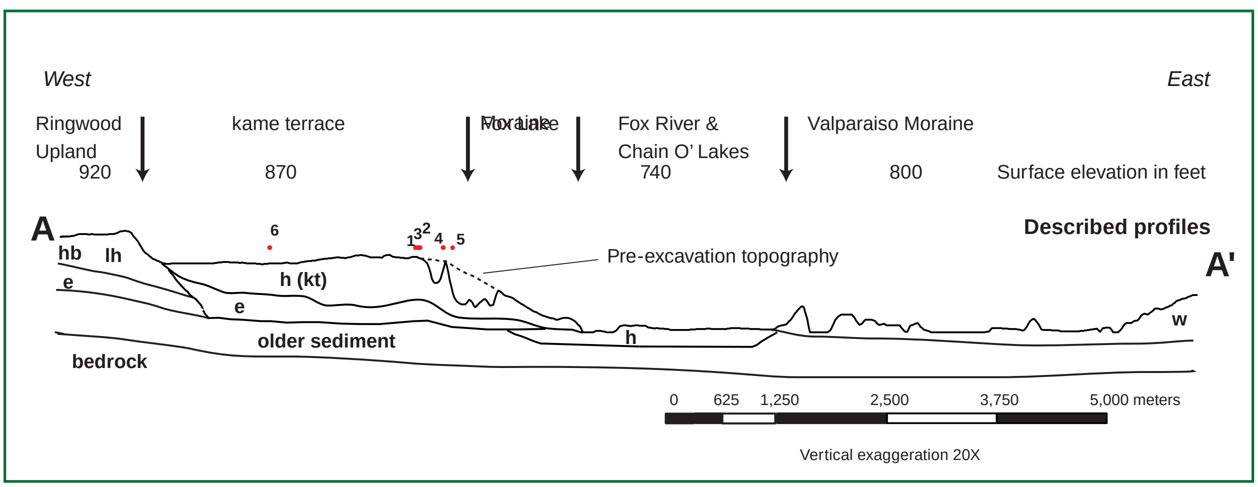


Figure 4 Schematic cross section A-A' showing changes in the geology crossing east from the Ringwood Upland, kame terrace, Fox Lake Moraine, Chain O' Lakes lowland, and Valparaiso Moraine. Units include (e) = Equality Formation (silt and clay), h = Henry Formation (sand and gravel), hb = Beverly Tongue, Henry Formation, lh = Haeger Member, Lemont Formation, h (kt) = kame terrace formed of Henry Formation), and w = Wadsworth Formation.

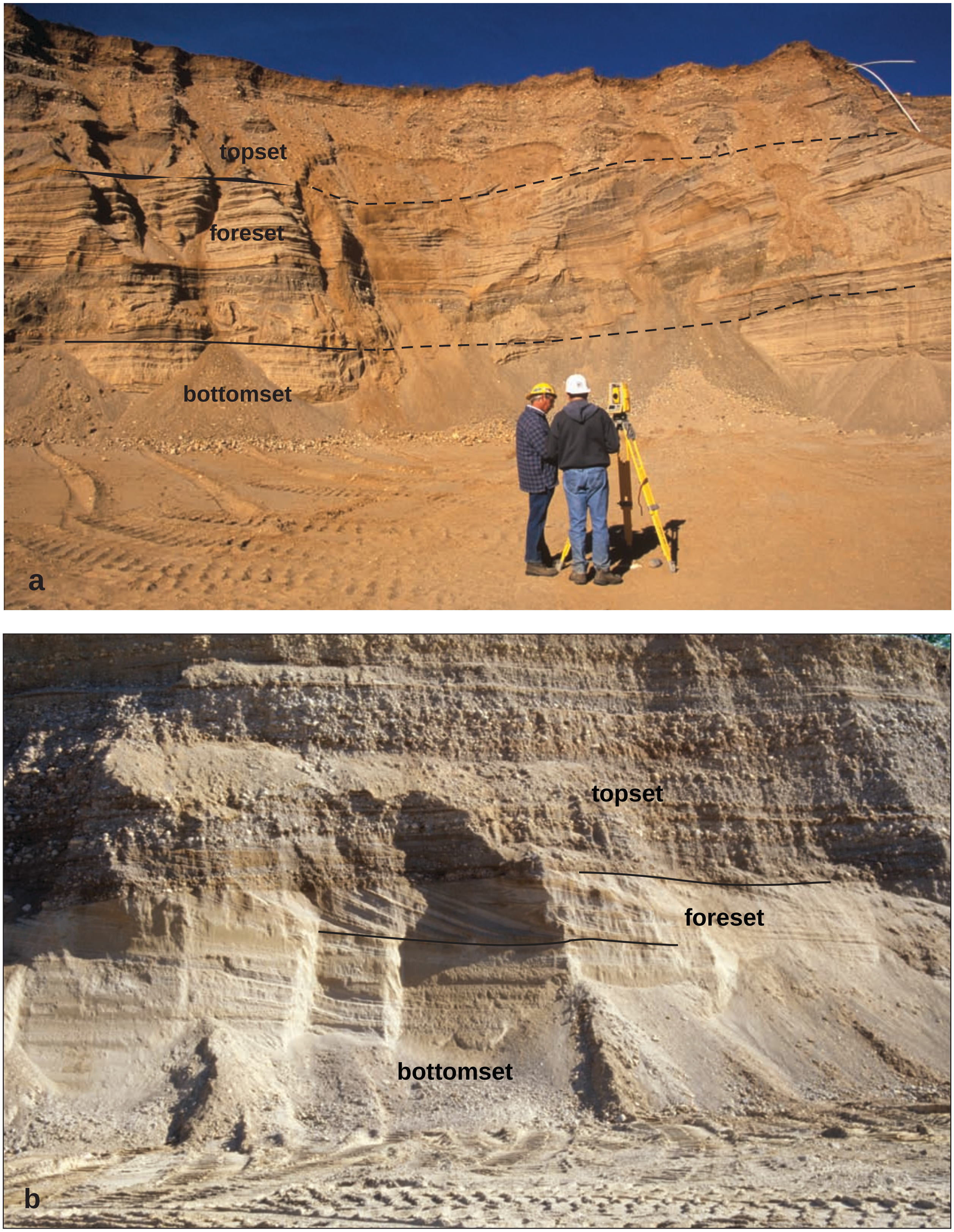


Figure 5 Folded and faulted deltaic sediments at eastern edge of kame terrace. The deformation is attributed to stresses at the glacial margin.

Topset sequences are composed of boulder and cobble gravels that fine westward, away from the glacial margin. Thickness of beds varies but generally decreases from up to 2.7 m (proximal, east) to about 1 m (distal, west). Fluctuating flows left planar-bedded, multistory and complex channel fills, trough cross-beds, and planar cross-bedding structures interbedded with ripple drift, cross laminated sand suggesting rapid deposition succeeded by sustained, low flows. At he bottom of the topset facies, the planar gravels layer, dips 1.2° west about parallel with the ground surface (1.8°) which forms the top of the deltaic deposits. The truncation of the foreset beds displaying contacts parallel with the ground surface is same as reported by Smith (1980) of ice-frontal deltas in New England.

Overall thickness of the foreset sequence abruptly decreases from 11 m at Profile 5 (proximal, Figure 5a) to the west (distal) only 0.7-3.2 m at Profiles 1 through 4, a distance of 100 to 400 meters (Figure 5b). This differs from the "ideal" delta described by Gilbert (1890). Foreset sequences are dipping, mostly planar or massive bedding structures with occasional ripple drift structure. Local, high angle faulting probably associated with loading rarely occur. Comparing Profile 5 with the other profiles shows how texture of the foreset beds coarsens eastward toward the glacier source. Steep dips of foresets grade into nearly horizontally bedded bottomsets which form the delta floor.

Bottomsets are composed of the finest-grained meltwater sediments in the deltaic sequence and progressively thicken and fine away from the glacier source (westward). Bottomset beds as observed in Profiles 5 to 2 contain large proportions of medium to fine and very fine sand which have little commercial value. The sand is typically horizontally laminated, massive and ripple-drift cross-laminated. Normally graded beds are attributed to local deposition from debris flows

Discussion

What features can you see on the shaded relief map in Figure 6 that were not obvious from the shaded relief map on Figure 3? The latter was created from 30-m elevation data, and Figure 6 was created from a variety of sources, including LiDAR (Lake County), 2-ft contour data (McHenry County), and 10-m DEM from the National Elevation Database. A portion of the Fox Lake and Richmond 7.5-minute Quadrangles are shown below for comparison. See if you can find the Spring Lake sand and gravel pit (Stop 8) rotational landslide scarps, and a small, but towering ice-walled lake plain. A more advanced exercise is for students to develop a geologic history based on cross-cutting surfaces observed on the kame terrace.

Acknowledgements

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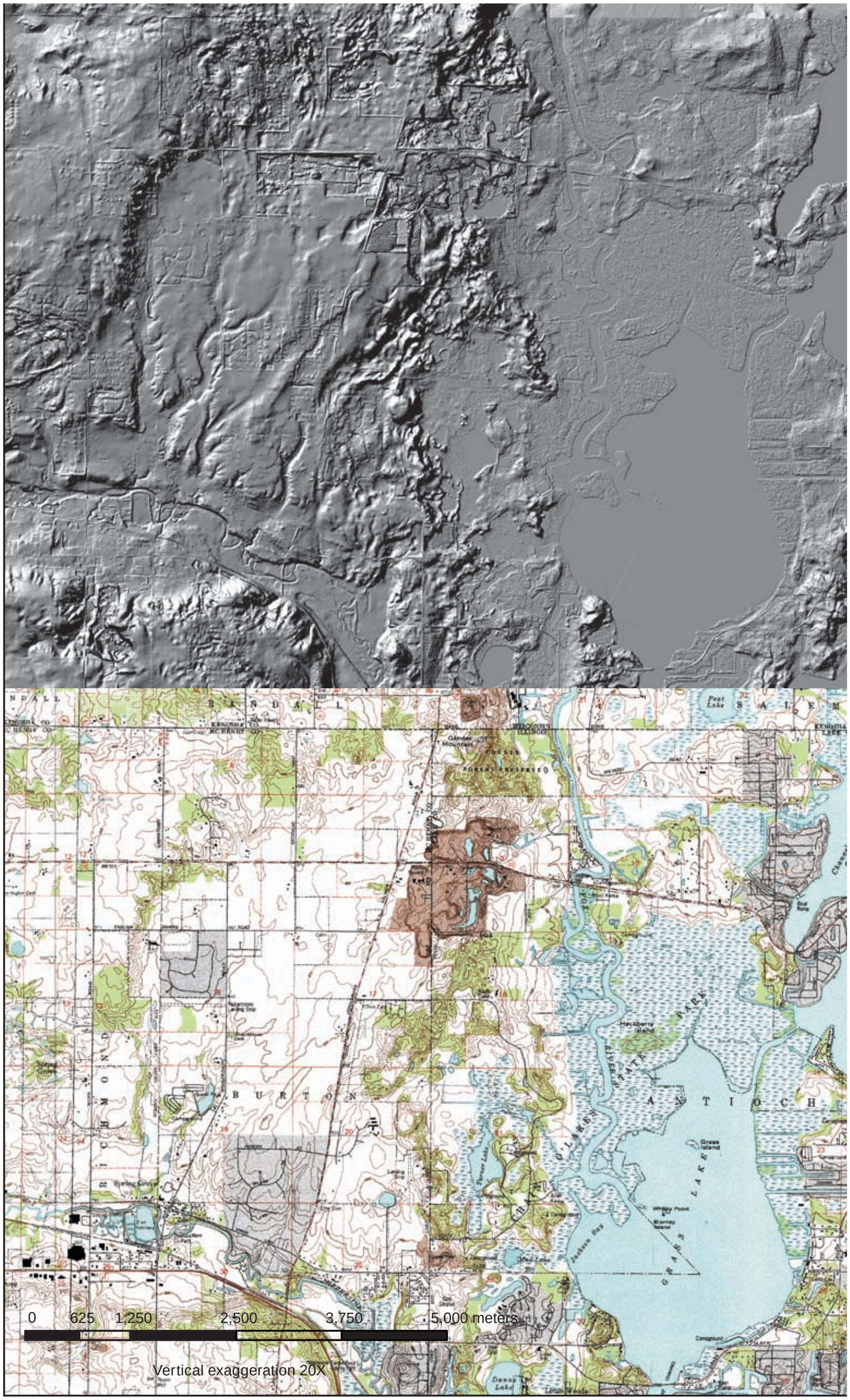

Figure 6 Shaded relief maps of portions of Lake and McHenry Counties, Illinois (top), and portions of the Richmond and Fox Lake, Illinois, 7.5-minute Quadrangles.

APPENDIX A. Description of Profile 1 at the Thelen Pit using the lithofacies code (Table 1).

THELEN NORTH PIT
PROFILE 1

Date: July 27, 2005

Described by: T.J. Kemmis, C. Stohr, A. Stumpf

Location: SW/SE/NE section 8, T. 46 N., R. 9 E., McHenry County, Illinois

Projected Coordinates: 401149.434 m East, 4703969.889 m North, 271.285 m orthometric - UTM16, NAD83

Geographic Coordinates: 42.4817728 North Latitude, 88.20270013 West Longitude, NAD83

Geographic Coordinates: 42.481753714 North Latitude, 88.202611897 West Longitude NAD27

Ground Surface Elevation: 890.041 ft MSL

Geologic Setting: The site is located on a kame terrace west of the Fox River Lowland that is incised into uplands which are underlain by Lemont Formation Haeger Member glacial deposits. The terrace is 1.5 to 2 miles wide (east-west) and just over 3 miles long (north-south), with an ice-contact face on its eastern margin that descends into the Fox River Lowland (Chain O'Lakes area). Ground surface elevations are about __ feet lower than the upland to the west and __ feet higher than the crest of the younger Valparaiso Morainic System east of the Fox River Lowland which is underlain by Wadsworth Formation glacial deposits.

Site Setting: The described section is located near the western end of the Thelen North Pit, close to the intersection between Illinois Route 173 and Wilmot Road, on the north face of the active pit.

Geologic Overview of the Described Section: The Thelen North Pit is located on the eastern edge of the kame terrace, but west of the ice-contact face. This section occurs on the north (south-facing) wall near the western edge of the North Pit.

Thick fine and very fine sand deltaic bottomset deposits occur at the base of the section. These grade into and are overlain by thin deltaic foresets of sand and pebbly sand. This section shows the lateral gradation of foresets and bottomsets away from the delta apex. The thickness of the bottomset sequence increased as the delta prograded westward to this location. Correspondingly, the foresets thin and become finer grained away from the delta apex because the basin became shallower as the bottomset sequence aggraded.

An angular unconformity separates the deltaic bottomset and foreset deposits from the overlying sequence of cobble and boulder gravels. The cobble and boulder gravels are laterally extensive, planar-bedded, matrix-supported (poorly sorted) cobble and boulder gravels interbedded with discontinuous broad, shallow channel fills composed of fine sand and pebble gravels that occur in a variety of bedding structures. Most of these channels are marked by thin fine sands that drape over broad, shallow scours (channels) on the surface of the underlying laterally extensive, planar-bedded cobble and boulder gravels. In thicker channels fills (up to about 0.5 m thick) various pebble gravels were often deposited. This sequence indicates high-magnitude meltwater events that deposited the laterally continuous, coarse cobble and boulder gravels. When stage waned, fine sands and pebble gravels were locally deposited in broad, shallow channels on the cobble gravel surfaces. The repetition of these deposits upward indicates significantly fluctuating flow as the sequence aggraded, with flow variations of this magnitude most likely on a seasonal scale.

Bed No.	Top Elevation (ft)	Top Depth (ft)	Thickness (ft)	Lithofacies	Description	USCS Classification
FILL (1.5 feet thick)						
1	890.0	0.0	1.5	Fill	Fill. Silt loam and gravelly loam fill; drainage tile at base.	CL; CL with grave l
1a	888.5	0.0	0.8	Fill	Base of soil A horizon developed in fill.	CL; CL with gravel
1b	887.7	0.8	0.7	Fill	Base of soil B horizon developed in fill.	CL; CL with gravel
UNIT Z – Planar-bedded cobble gravels with interbedded, discontinuous channel fills (23.3 feet thick)						
2	887.0	3.0	3.4	CGms(m)	Laterally extensive, matrix-supported, massive to faintly planar-bedded cobble gravel; matrix consists of fine to medium pebble gravel and coarse sand; beta soil horizon in upper 12 to 18 inches	GW with sand
3a	883.6	6.4	0.2	S(m)(ccf)	Massive coarse sand with very fine pebble gravel; this bed is part of a medium scale channel fill that is lenticular within the sequence and pinches out laterally.	SP to SP with gravel
3b	883.4	6.6	1.1	CGms(pl)(ccf)	Matrix-supported, planar-bedded cobble gravel; matrix is pebble gravel and coarse sand; cobbles constitute only about 10 percent of the unit; this bed is part of the same medium scale channel fill as the overlying S(m) bed and is lenticular within the sequence and pinches out laterally.	GW with sand
3c	882.3	7.7	2.8	S(t)(ccf)	Medium bedded, trough cross-bedded medium sand; this bed is part of the same medium scale channel fill that includes the overlying CGms(pl) and S(m) beds and is lenticular within the sequence and	SP

					pinches out laterally.	
4	879.6	10.5	2.9	CGms(pl)	Laterally extensive, matrix-supported, planar-bedded cobble gravel; matrix is primarily pebble gravel with lesser amount of coarse sand.	GW with sand
5a	876.7	13.4	0.6	S(t)	Small-scale trough cross-bedded medium sand.	SP
5b	876.1	14.0	2.5	PGms(t)(ccf)	Matrix-supported, trough cross-bedded to small-scale channel fill of coarse to fine pebble gravel; matrix is fine pebble gravel and coarse sand; some thin fine sand beds in lower part.	GW with sand
5c	873.5	16.5	3.1	CGms(t)(ccf)	Trough-cross bedded to small-scale channel fill of matrix-supported cobble gravel; matrix is medium to coarse pebble gravel and sand.	GW to GP
6	870.4	19.6	2.7	PGms(pl)	Lenticular bed of matrix-supported planar-bedded pebble gravel; matrix is coarse sand and very fine pebble gravel.	GW to GP
7a	867.6	22.3	0.1	S(m)(ccf)	Massive, reddish brown silty very fine sand draped over and infilling underlying cobble gravel that fills a broad, shallow channel.	SP-SM
7b	867.6	22.4	2.1	CGms(pl)(ig)(ccf)	Matrix-supported, planar-bedded cobble gravel filling broad, shallow channel; matrix is fine pebble gravel and sand with occasional medium gravel; inversely graded (coarsens upward).	GW with sand
8a	865.5	24.6	0.2	S(m)(ccf)	Massive, horizontal bed of reddish brown silty very fine sand draped over underlying cobble gravel that fills a broad, shallow channel.	SP-SM
8b	865.3	24.8	4.6	CGms(pl)(ig)(ccf)	Matrix-supported, planar-bedded cobble gravel filling broad, shallow channel; matrix is fine to coarse pebble gravel with sand; inversely graded (coarsens upward).	GW with sand
UNIT Y – Deltaic Foresets (2.8 feet thick)						
9	860.7	29.4	2.8	S(h-f)-S(pl-f)	Deltaic foresets of laminated to very thinly bedded medium and coarse sand; foresets dip to the west-southwest.	SP

UNIT X – Deltaic Bottomsets (3.9 feet exposed, extends below base of pit – at least 12.3 feet thick)

10	857.9	32.2	3.9	S(h-b)-S(r-b)	Deltaic bottomsets of laminated and ripple-drift cross-laminated fine and very fine sand.	SP
SLUMP						
	853.9	36.1	8.4	Slump	Top of scree at base of outcrop.	
	845.5	44.5		Pit floor	Bottom of outcrop.	