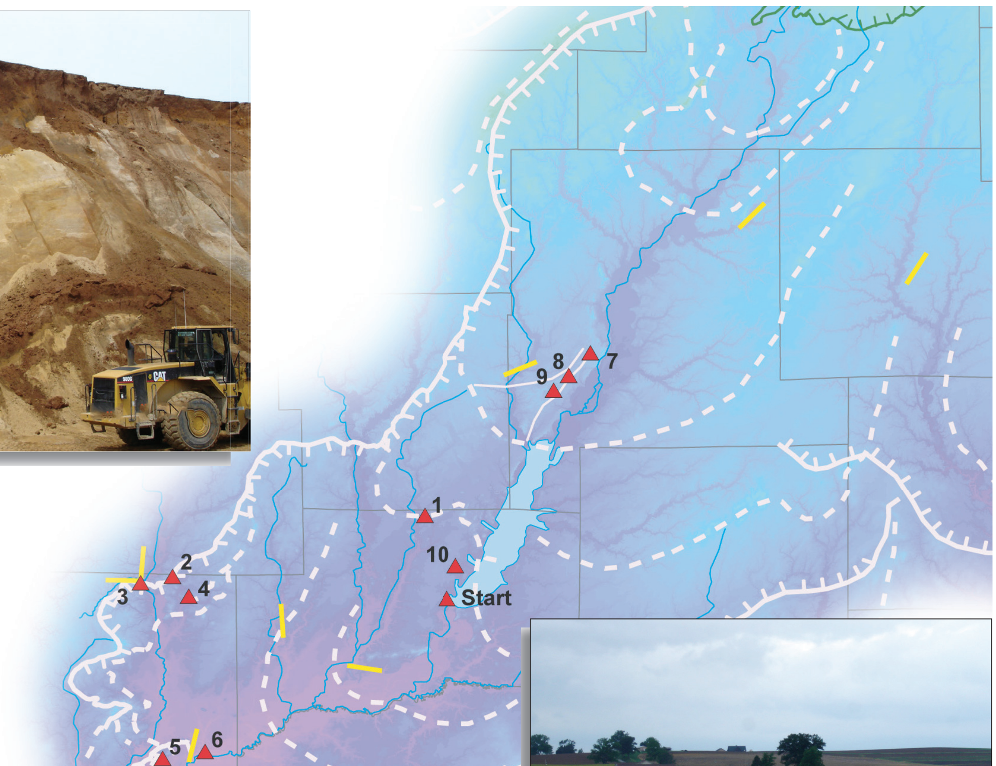
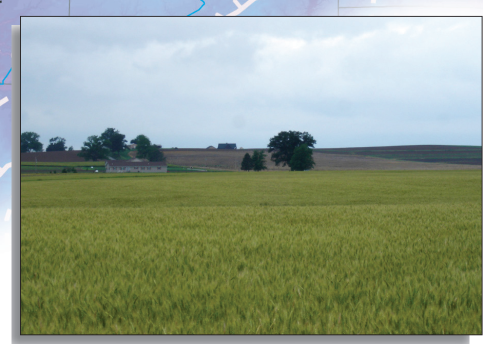
Ridges, Mounds, and Valleys: Glacial–Interglacial History of the Kaskaskia Basin, Southwestern Illinois

55th Midwest Friends of the Pleistocene 2011 Field Conference

Edited by **David A. Grimley and Andrew C. Phillips**





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Midwest Friends of the Pleistocene

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Cover photographs: (*top left*) *High-angle reverse faults in Hagarstown sand on the southwest wall of the Keyesport Sand and Gravel pit* (*Stop 1*). (*bottom right*) *Illinois episode glacial moraine* (*Ralls Ridge*) *a few miles east of Red Bud, Illinois.*

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Edited by **David A. Grimley[1] and Andrew C. Phillips[1]**

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FIELD TRIP THEMES

- Origin of constructional ridges on the Illinois Episode till plain
- Possible existence of a sublobe or ice stream in the Kaskaskia Basin
- Illinois Episode glacial-sedimentary processes and lithostratigraphy
- History and environment of paleo-slackwater lakes in the Kaskaskia Basin
- Climatic records during the middle to late Pleistocene and Holocene
- Archeological history
- Sodium-affected surface soils
- Integration of geophysical studies with research and mapping

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Meetings of the Midwest Friends of the Pleistocene*

1	1950 Eastern Wisconsin	Sheldon Judson
2	1951 Southeastern Minnesota	H.E. Wright Jr., R.V. Ruhe, and L. Gould
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	L. 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 southeastern Missouri	S.P. Esling and M.D. Blum
43	1996 Eastern North Dakota and northwestern Minnesota	K.I. Harris and others
44	1998 North-central Wisconsin	J.W. Attig and others
45	1999 North-central Indiana and south-central Michigan	S.E. Brown, T.G. Fisher, and others
46	2000 Southeast Nebraska and northeastern Kansas	R.D. Mandel and E.A. Bettis III
47	2001 Northwestern Ontario and northeastern Minnesota	B.A.M. Phillips and others
48	2002 East-central Upper Michigan	W.L. Loope and J.B. Anderton
49	2003 Southwestern 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
55	2011 Southwestern Illinois	D.A. Grimley and A.C. Phillips

*Field conferences were not held in 1968, 1970, 1974, 1977, 1997, 2009, or 2010.

A Brief History of the Kaskaskia Basin

The glaciers came and the glaciers went; the rivers rose and lakes formed. The wind blew and the dust covered all. Life rejoiced when the ice was no more.

The snails and the trees thought they were so free. The basins were filled and the rivers had glee.

But once again the ice advanced, covering all with relentless power. Lakes and rivers were filled to the brim, only to drain when the game was over. Hills and plains were left to see, by mammoths and mastodons that still ran free.

Ostracodes and mollusks were left to ponder, when will the ice come from over yonder?

And once again, the glaciers advanced, but this time they had no chance. The ice could only view the Kask without much effect to its inner core. Yet the powerful Miss ran its course and bumped its neighbor to its place. The wind blew and the dust covered all. Life rejoiced when the ice was no more.

Addendum: Native peoples entered the basin, trading materials from here to there. How did they live and where did they go? The evidence is now gone, under 20 feet of H_2O.

An Anonymous Kaskaskian

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FIGURES

PART I: Quaternary Geologic Overview of the Lower Kaskaskia Basin

INTRODUCTION

The intention of this field trip is to provide a tour of Quaternary deposits and history of the Kaskaskia Basin region of southwestern Illinois. To our knowledge, a Friends of the Pleistocene or Geological Society of America field trip related to the Quaternary or Pleistocene history has never been conducted in the Kaskaskia region. This area was glaciated during both the pre-Illinoian and Illinoian Stages and includes the controversial Ridged-Drift area (Leverett 1899; Jacobs and Lineback 1969), which contains constructional hills of various forms that protrude from the dissected Illinoian till plain. During this trip, we offer our current hypothesis, which envisions a Kaskaskia Basin Sublobe, at times possibly fed by an ice stream, within an overall glacial advance from the Lake Michigan basin during oxygen isotope stage (OIS) 6. We believe that such a sublobe may explain the presence and general distribution of morainal ridges, esker forms, and other ice-contact landforms in southwestern Illinois.

This guidebook includes a traditional road log, a general introduction with our latest thoughts on the regional Quaternary history, and a description for each stop. The field trip stops include sand and gravel pits, natural exposures, archeological sites, soil pits, and core samples of ridge deposits. The principal theme of the trip is the origin of the Ridged Drift and the former existence of a possible sublobe; however, a secondary theme is the occurrence of a slackwater lake, with fluctuating water levels, in the lower reaches of the Kaskaskia Basin during each major glaciation. The following general discussion of the Quaternary history in southwestern Illinois is based in part on detailed 1:24,000-scale mapping and research that has been conducted primarily in the western portion of the field trip area (Madison, Monroe, and St. Clair Counties) over the past decade.

QUATERNARY HISTORY AND STRATIGRAPHY, LOWER KASKASKIA BASIN, SOUTHWESTERN ILLINOIS[1]

The Quaternary and Pleistocene recently have been redefined to include time back to about 2.6 million years ago (Mya; Head et al. 2008); however, little is known about Early Pleistocene deposits (Gelasian and Calabrian Stages) in the region. The Kaskaskia Basin (Figures Q0 and Q1) was overrun by continental glaciers twice during the Middle Pleistocene (Ionian Stage; Head et al. 2008): first, during a pre-Illinois Episode glaciation [sometime between ~700 and 420 thousand years ago (ka)], and second, during the Illinois Episode glaciation (~180 to 130 ka; Figures Q2, Q3, and Q4). Deposits of both glaciations were documented in southern Illinois by MacClintock (1929) and have been repeatedly confirmed by observations in outcrops and cores (McKay 1979; Phillips 2004b; Grimley and Phillips 2011). During both glacial episodes, glaciers generally advanced from the northeast, with Illinois Episode glaciers emanating from the Lake Michigan basin (Willman and Frye 1970; McKay 1979; Grimley et al. 2001) and pre-Illinois Episode glaciers probably emanating from the eastern Great Lakes basins (Michigan, Huron, Erie, and Saginaw). During both episodes, glacial meltwater in southwestern Illinois appears to have been focused toward an ancestral valley in the Kaskaskia Lowland (Figures Q0 and Q5), as documented by sandy outwash in the subsurface (Phillips and Aper 2006; Grimley and Phillips 2011). The ancestral location of the Kaskaskia Valley somewhat coincides with the current valley and provided a southwestern passage of meltwater toward the Mississippi River valley or into ancient slackwater lakes. During the last glaciation, the Wisconsin Episode (~60 to 12 ka), glacial ice reached northeastern Illinois, northern Iowa, Wisconsin, and Minnesota, but did not reach southern Illinois (Hansel and Johnson 1996). Yet the effects of this glaciation were left throughout southwestern Illinois as torrents of glacial meltwater deposited sand and gravel outwash in the Mississippi River valley (Grimley et al. 2007) and, to a lesser extent (and with a finer texture), in the Kaskaskia River valley (Grimley and Webb 2010; Grimley and Phillips 2011). Finer silty sediment in the outwash valleys, especially the Mississippi Valley, was deflated by wind and incrementally draped over the uplands, providing a record of thick loess deposits (McKay 1979; Wang et al. 2003; Grimley and McKay 2004). Reflecting this source area, the loess thickness on uneroded uplands decreases exponentially away from the Mississippi Valley (McKay 1979; Hansel and Johnson 1996; Grimley and Phillips 2011). Concurrent with loess and outwash deposition, widespread deposition occurred in a large slackwater lake in the lower Kaskaskia Valley (Shaw 1921; Phillips 2008; Grimley and Webb 2009). This lake, known as glacial Lake Kaskaskia (Willman and Frye 1970), is expressed geomorphically today by multiple terrace levels. Postglacial deposits occur in modern valleys, with deposits of fine sand, silt, and clay [30 ft (9.1 m) thick] in the formerly active meander belts of the Kaskaskia

[1]Lithostratigraphic names are shown in bold at first use.

Valley. Channelization, straightening, and confinement by levees of these rivers and of some smaller creeks have altered the natural sedimentary regimes and processes in the past century and a half.

Surficial deposits in the middle to lower Kaskaskia drainage region of southwestern Illinois (Figures Q0, Q1, and Q2) vary widely from thick glacial and postglacial alluvium [combined >100 ft (>30.5 m)] in the broad Kaskaskia Valley, to thin glacial sediment [<25 ft (<7.6 m)] on bedrock topographic highs, to areas of thick loess [>25 ft (>7.6 m) thick], to ice-contact deposits [up to 150 ft (45.7 m) thick] in a train of ridges, in some areas subparallel to the Silver Creek and Kaskaskia River valleys (Grimley and Phillips 2011).

Exposures of bedrock, predominantly Pennsylvanian rocks, occur where creeks have incised into bedrock topographic highs. The trend of preglacial ridges and valleys on the bedrock topographic surface follows the strike of Paleozoic rocks on the west side of the Illinois Basin (Figure Q3). For instance, north–south- and northwest–southeast-trending buried bedrock valleys occur in eastern St. Clair County (Figure Q3) and likely developed adjacent to preglacial cuestas with shales in lowlands and sandstone or limestones forming the highlands. The lower reaches of the Kaskaskia River flow into an area of more competent Mississippian bedrock (Kolb 2010), causing the valley to be more constricted in this area. Highlands underlain by Mississippian rocks in far southwestern Illinois, containing karstic limestone, also helped to impede the flow of glacial ice and prevented significant advances into eastern Missouri from northeastern-source glaciers during the middle Pleistocene. The limit of glaciation during both the Illinois and pre-Illinois Episodes appears to have been in the Mississippian highlands of southwestern Illinois (Grimley and Shofner 2008; Kolb 2010).

Early to Early-Middle Pleistocene Preglacial History (~2.6 Mya to 500 ka)

The early Pleistocene history, before invasion of the first glaciers, has a limited geologic record that includes weathering and alluvial deposition. Deposits are mainly limited to bedrock residuum, colluvium, and alluvium, not unlike present-day surficial deposits in unglaciated regions of Missouri and Kentucky that are far from loess sources. A hillier terrain during this preglacial time probably resulted in more active gravity-slope processes, including creep and slumping along the margins of valleys. Alluvial deposits from preglacial Quaternary time tend to be relatively sandy or clayey in texture, and they lack the predominantly loessal influence that caused a preponderance of silty-textured upland, lacustrine, and alluvial deposits of middle to late Quaternary age.

Canteen Member [Preglacial Alluvium and Colluvium]

Fine-grained alluvial and colluvial deposits, typically faintly stratified, have been informally classified as the **Canteen member,** a basal unit of the **Banner Formation** (Figure Q4; Phillips 2004a). This unit has now been documented by numerous subsurface observations and is preferentially found in buried bedrock valleys tributary to the ancestral Kaskaskia and Mississippi Valleys (Phillips and Grimley 2004; Grimley and Phillips 2011). A preglacial interpretation for deposition of the Canteen member is substantiated by the lack of glacial erratics, low magnetic susceptibility, low or negligible carbonate content, and high kaolinite/chlorite content. All the above reflect a domination of local bedrock influences, predominantly Pennsylvanian shale. In many areas, it may have a slight greenish gray cast, not unlike the color of some local shales, or it may reflect poorly drained soil conditions. Beds of sand occur locally near the unit base. In some areas of southwestern Illinois, weak to moderate soil structure indicative of a paleosol has been observed in the upper portions of the Canteen member. Most of the Canteen member was probably deposited during the early-middle Pleistocene based on amino acid geochronology of one core (Oches et al. 2002; Grimley and Oches 2015). The lower foot or so of the Canteen member, above Pennsylvanian bedrock, may include residuum or subangular sandy gravel. In a few areas, a few inches to a few feet thick of rounded chert gravel in deep bedrock valleys at the basal contact with bedrock may be correlative with the **Grover Gravel,** a Pliocene or early Quaternary fluvial deposit (Willman and Frye 1970). Thick residuum, formed *in situ* from weathered limestone or shale, has been classified as the **Oak formation** (Figure Q4; Nelson et al. 1991) and is probably Pliocene to early Pleistocene in age. The Oak formation may underlie the Canteen member or may be included within the Canteen member if the residuum is thin and similar in lithology.

Pre-Yarmouth (or Pre-Illinois) Episode Glaciation, Middle Pleistocene (~650 to 430 ka)

Overall, evidence from Quaternary mapping and glacial studies in southwestern Illinois indicates that probably only one glacial advance physically reached the middle to lower Kaskaskia Basin during pre-Yarmouth (or pre-Illinois) Episode time (Grimley and Phillips 2011). However, multiple glacial advances cannot be ruled out,

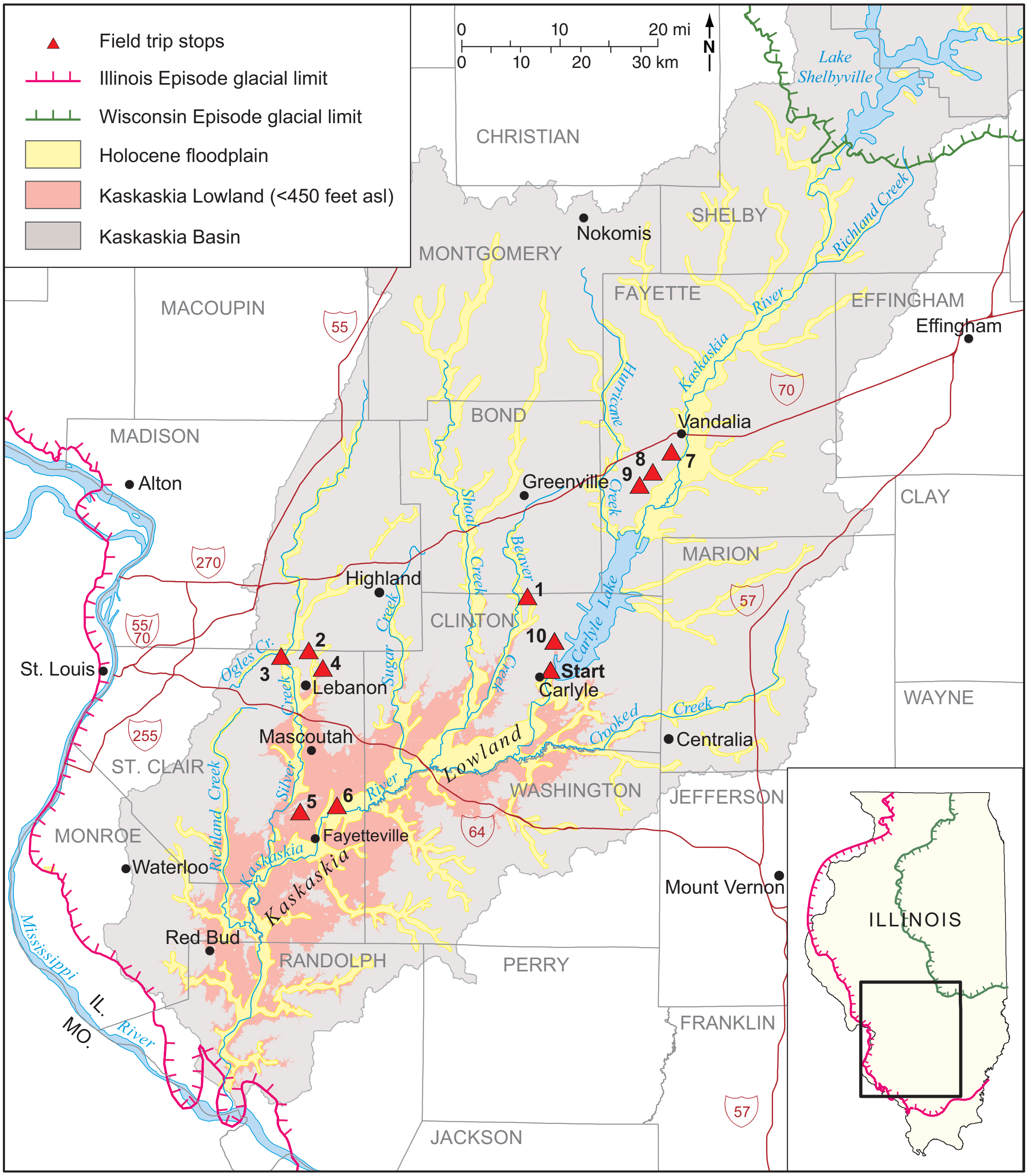


Figure Q0 Overall location map of the Kaskaskia Basin and Kaskaskia Lowland in southwestern Illinois. asl, above sea level. The same area appears in Figures Q1–Q3.

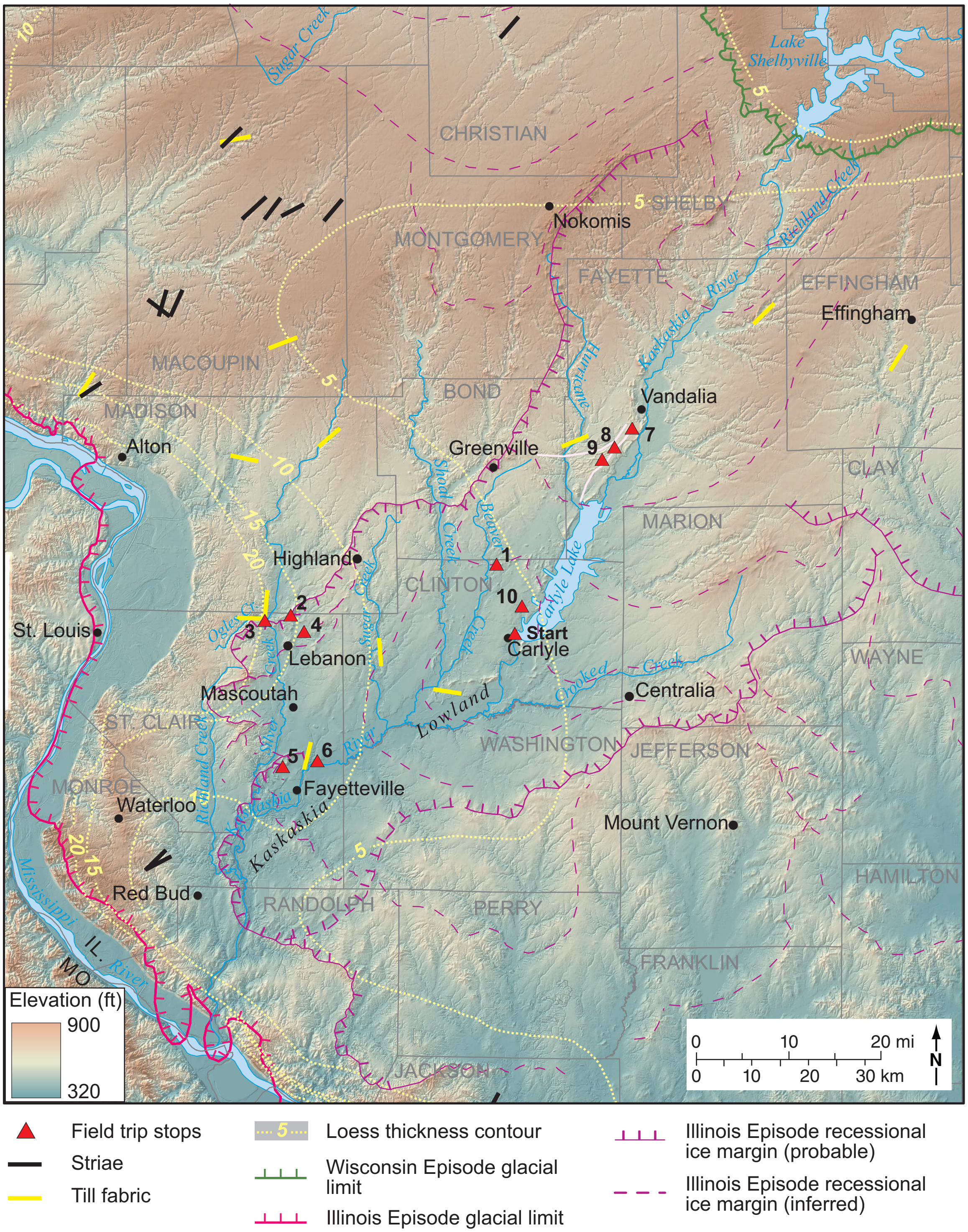


Figure Q1 Inferred ice margin positions in the Kaskaskia Basin region of southwestern Illinois. Striation directions are from Leighton and Brophy (1961), E.D. McKay (personal communication), and others. Till fabric directions are from Lineback (1971) and Webb (2009).

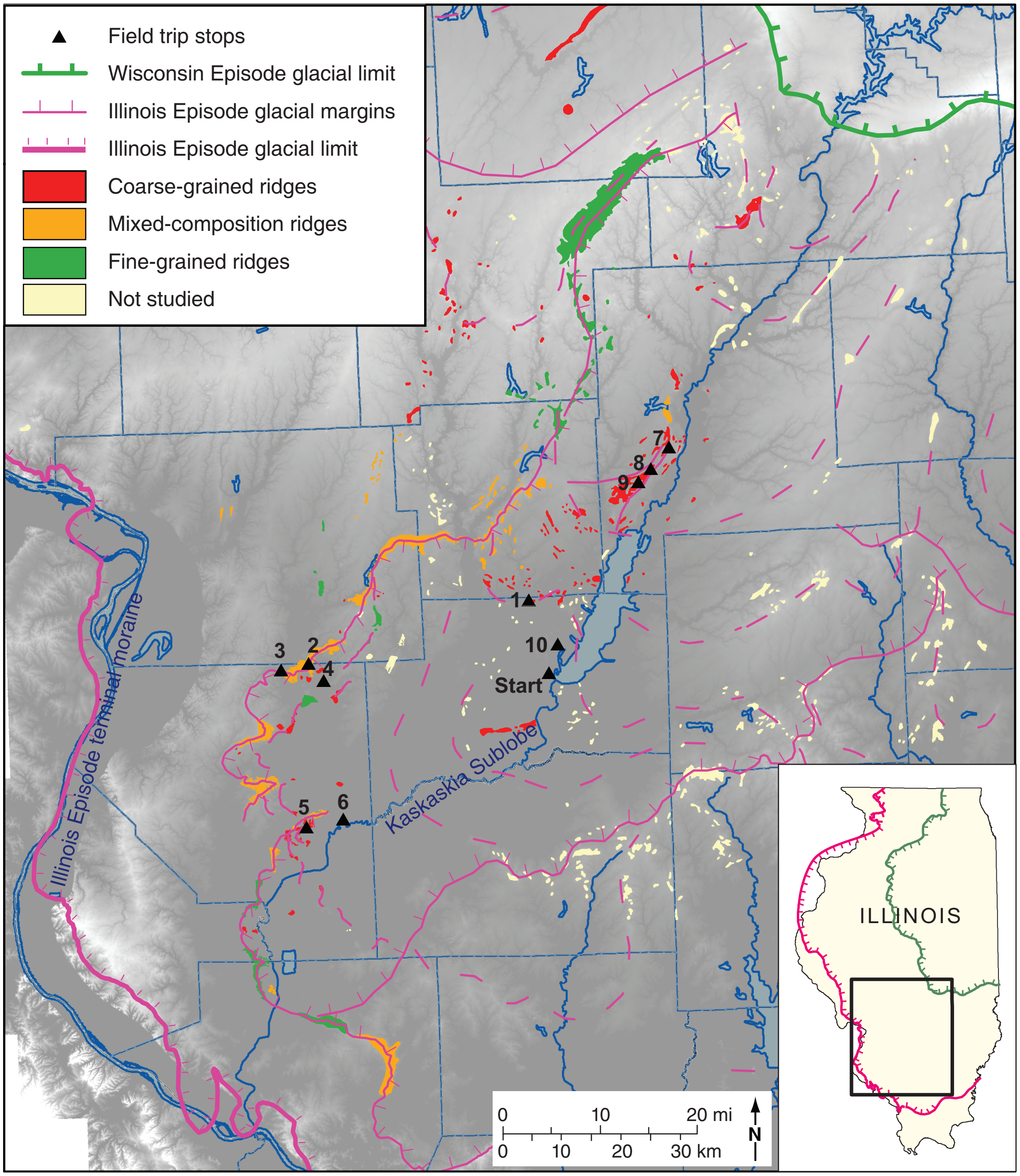


Figure Q2 Generalized lithologic composition of constructional glacial ridges in southwestern Illinois. Modified from Webb (2009), and based on various studies by Leighton and Brophy (1961), Jacobs and Lineback (1969), Stiff (1996), Grimley and Phillips (2011), and others. The background map is a grayscale digital elevation map, with lighter tones representing higher elevation and darker tones representing lower elevation.

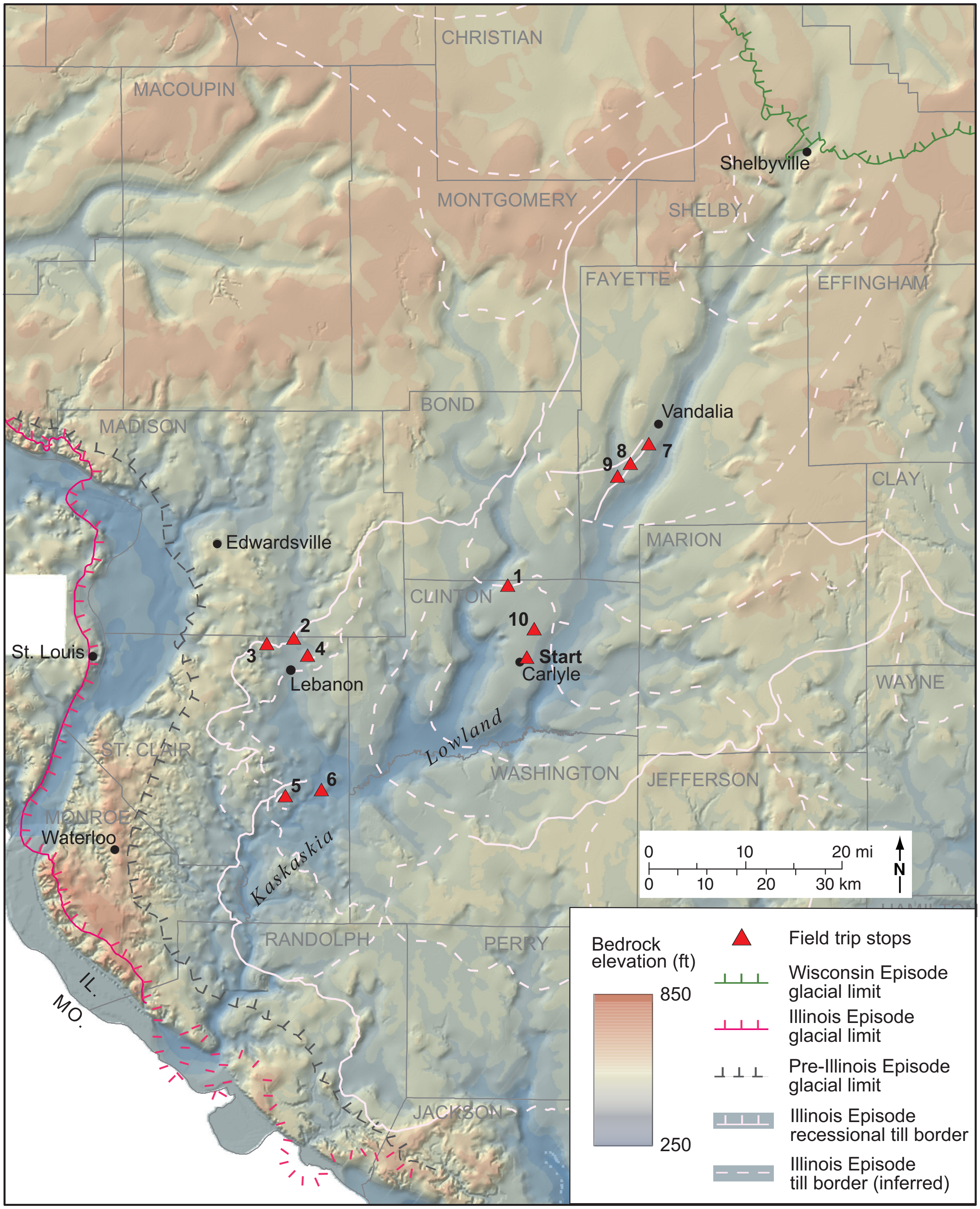


Figure Q3 Bedrock topography map of southwestern Illinois. Based on Grimley and Phillips (2011) and unpublished data in Madison, St. Clair, and Monroe Counties. Based on Herzog et al. (1994) for all other areas.

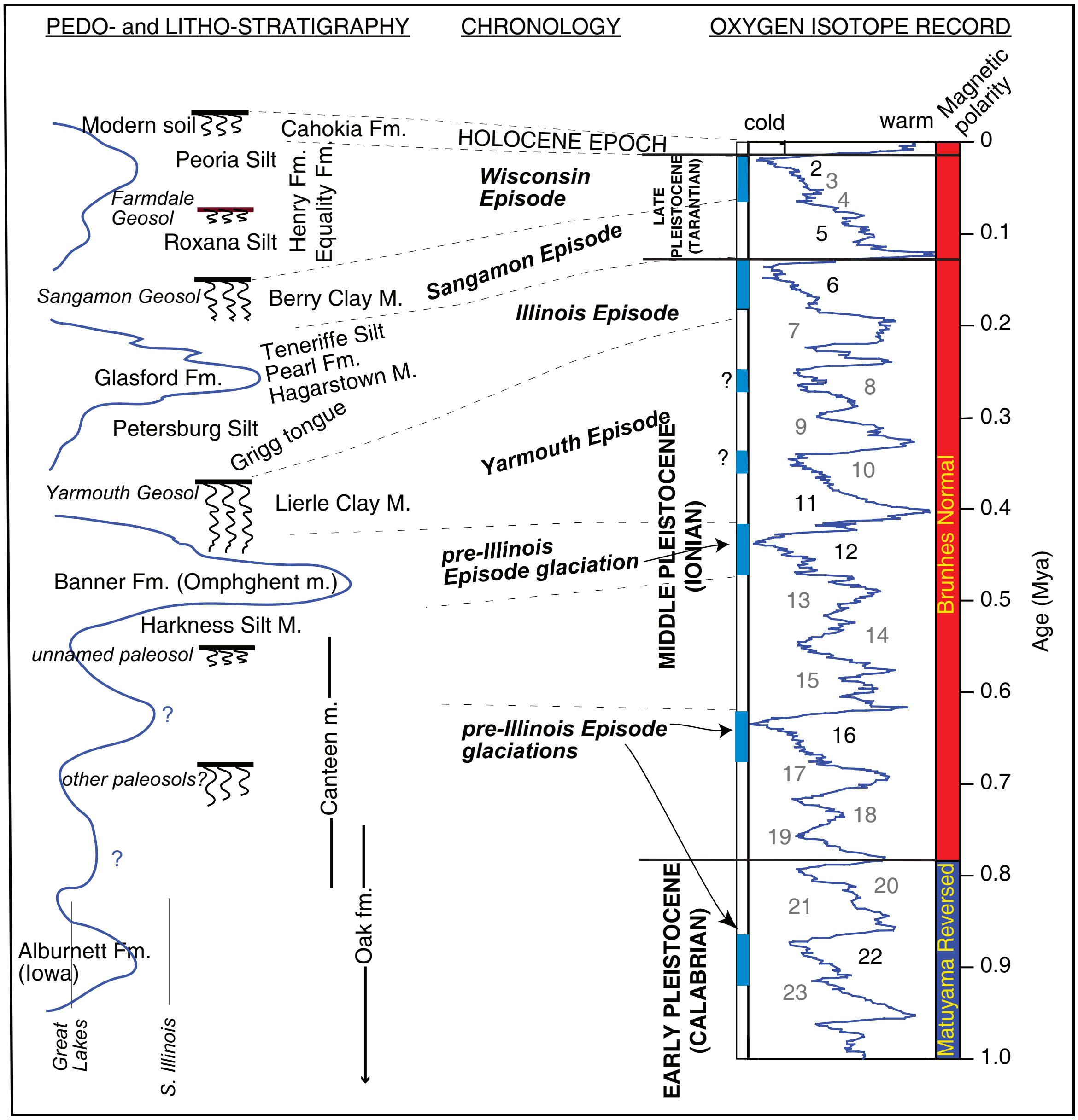


Figure Q4 Stratigraphic framework and age of Quaternary deposits in the Kaskaskia Basin region. Glaciations that probably reached the Great Lakes basins are highlighted in a greenish hue. Correlations modified from Curry et al. (2011). Oxygen isotope curve from Zachos et al. (2001). Mya, million years ago.

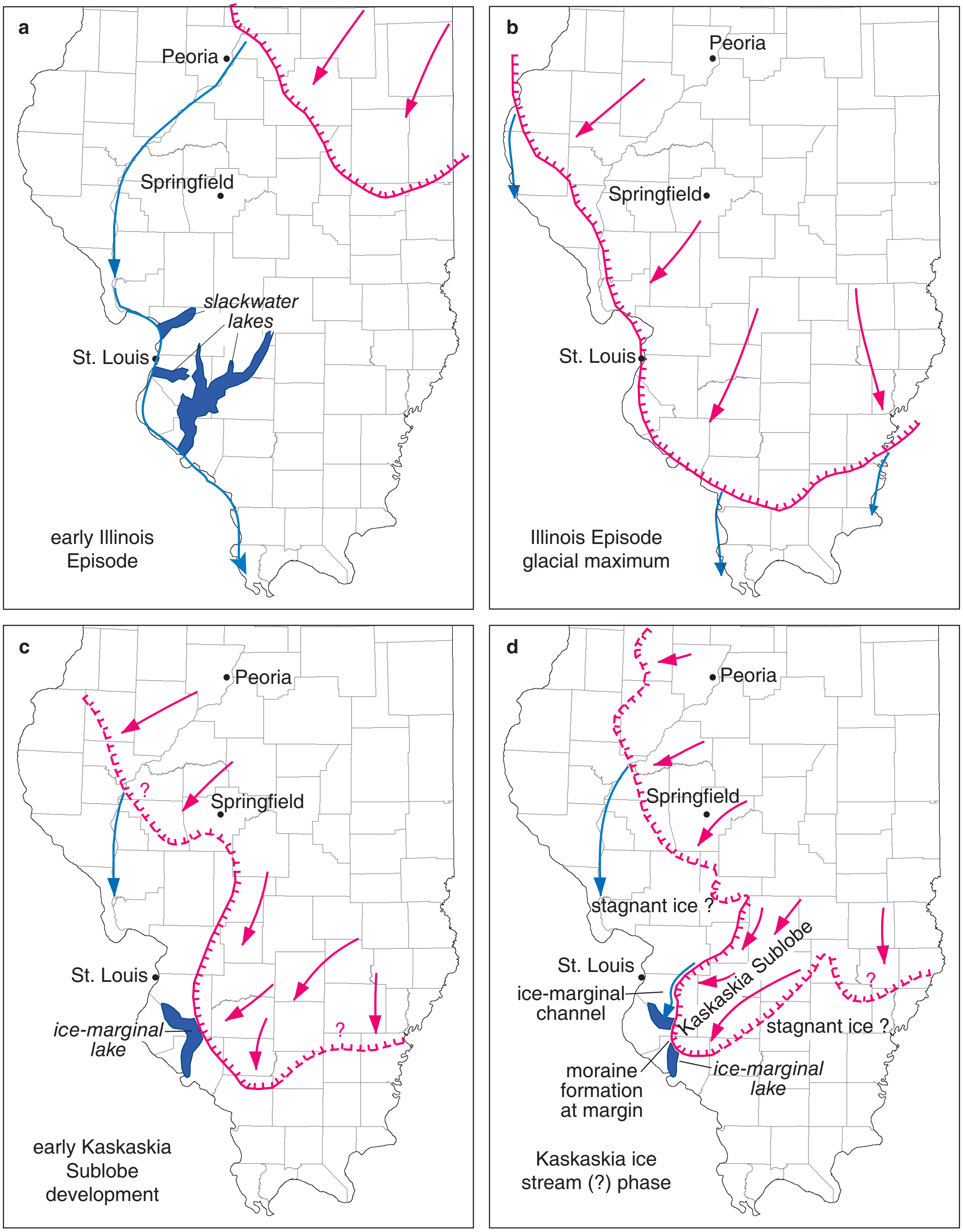
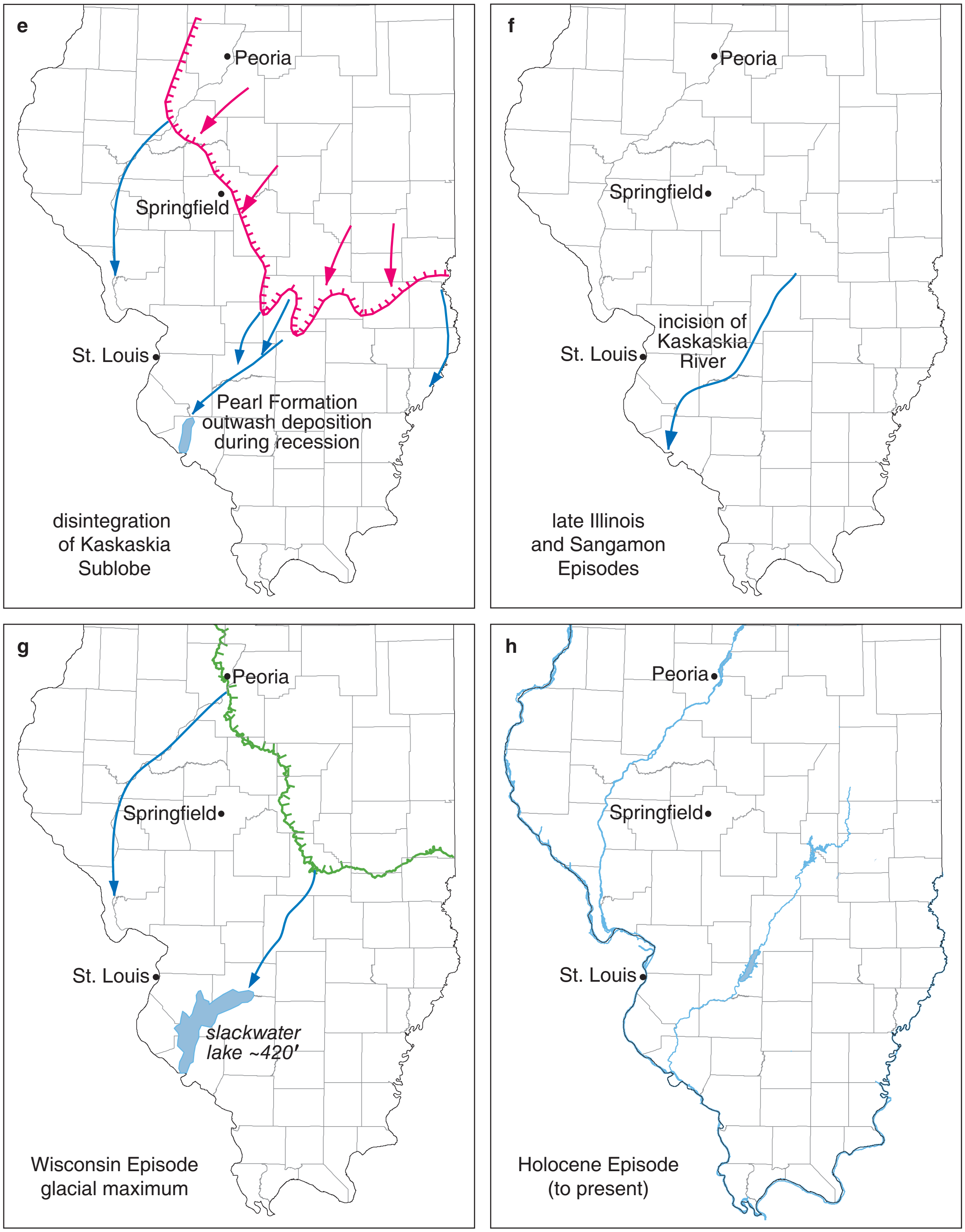


Figure Q5 (a–h) Cartoon diagrams of ice advances and geologic events in southern Illinois (with focus on the Kaskaskia Basin) during the middle to late Quaternary. Dark pink arrows indicate glacial ice flow during the Illinois Episode, and blue arrows indicate glacial meltwater flow.



Scale ~1:4,000,000

Figure Q5 (*Continued*).

and some lithological variations do occur within pre-Illinois Episode till. General correlations with the marine oxygen isotope record suggest that a regional glacial advance may have been synchronous with either OIS 12 or OIS 16 (Figure Q4), both periods of significantly large global ice volume and extreme cold (Shackleton 1987; Hansel and McKay 2010). Amino acid geochronology using pre-Illinois Episode fossil gastropods also suggests an age between about 650 and 400 ka (Grimley and Oches 2015). An age of about 450 ka (OIS 12) may be more likely based on compositional weathering indices for the duration of Yarmouth Geosol development in comparison with the duration of Sangamon Geosol development (Grimley et al. 2003).

Banner Formation [Till Deposits]

The limit of pre-Illinois Episode till observations in southwestern Illinois appears to be close to, and just within, the border of the Illinois Episode glaciation (MacClintock 1929; Willman and Frye 1970; Grimley and Shofner 2008; Kolb 2010), which locally parallels the strike of the Mississippian bedrock escarpment that rims the Illinois Basin. Within the interior of the Kaskaskia Basin, Banner till deposits are generally sporadic or absent from bedrock topographic highlands because of glacial erosion during the Illinois Episode. Banner till deposits were also significantly scoured by fluvial or glacial events in the main meltwater sluiceways, such as the Kaskaskia Valley. Correlations are, of course, difficult where key paleosol marker horizons have been removed by erosion.

Pre-Illinois Episode till (and related debris flow diamicton) in the western part of the St. Louis Metropolitan East region has been informally classified as the **Omphghent member** (pronounced Om-jent) of the **Banner Formation** (Figure Q4; McKay 1979, 1986). The Omphghent member only rarely occurs near surface (McKay 1979; Grimley and Phillips 2010), being found mainly in preglacial tributary bedrock valleys or lowlands, where it was protected from erosion during later geologic events. In its type area of central Madison County, the Omphghent member is a calcareous, grayish brown to yellowish brown (oxidized), pebbly silty clay loam to silty clay diamicton. The unit is interpreted primarily as till or debris flows, and can include small lenses of glaciofluvial sand and gravel. The diamicton includes usually <5% (by volume) pebbles of sandstone, chert, shale, siltstone, and rare crystalline erratics. The upper part of the till (or all of it if thin) is oxidized by Yarmouth Geosol interglacial soil development and may include the strongly developed soil solum; however, the solum is more commonly contained within the overlying Lierle Clay Member (see the Yarmouth Episode section). Till of the Omphghent member is thinner and more clayey in the southern and western areas of the Kaskaskia Basin (Grimley and Webb 2010) compared with farther north (Phillips and Grimley 2004) because of the greater local incorporation of soft Pennsylvanian shales, proglacial pre-Illinois Episode lake sediment, and bedrock residuum.

Where defined in central Madison County, the Omphghent till has a higher clay content, less illite, less dolomite, and more shale fragments than does the overlying Glasford Formation till, reflecting more input from the local substrate. Outside central Madison County, the lithological distinction between the pre-Illinois and Illinois Episode tills in southwestern Illinois is less apparent, but is possible to discern through detailed analyses and correlations with key localities. Near the glacial termini, local shale-rich deposits, residuum, and loess were incorporated into glacier beds during both glaciations (Grimley and Webb 2010), so the Banner and Glasford tills may have a similar field appearance in some areas (i.e., Monroe, western St. Clair, and Randolph Counties). In the up-ice direction to the east, the pre-Illinoian till is less clayey than that to the southwest and can have a texture similar to the loamy or silty Glasford Formation (Jacobs and Lineback 1969); hence, it is often referred to as Banner till or Banner diamicton (rather than Omphghent member). The most useful analytical parameter for distinguishing tills from the two glaciations (in the absence of paleosols and stratigraphic control) seems to be carbonate content or calcite/dolomite ratios (Jacobs and Lineback 1969; McKay 1979; Phillips and Grimley 2010), with the Banner Formation typically having less dolomite and less total carbonate. Because of a possibly different distal source area (perhaps Lake Huron or the Saginaw region), other geochemical or mineralogical signatures or ratios might exist that could further help with unit delineation.

Banner Formation [Lake Sediments, Loess, and Outwash]

Pre-Illinois Episode till deposits in the Kaskaskia Basin may be intercalated with, underlain, or overlain by lake sediments, fine-grained alluvium, sand and gravel outwash, or loess deposits. The most common unit found below Banner till in the lower Kaskaskia region is the massive to laminated, sometimes fossiliferous, silt to silty clay **Harkness Silt Member** of the Banner Formation (Figure Q4; Willman and Frye 1970). This unit may include lake sediments, fine-grained alluvium, and loess deposits, with the thickest occurrences consisting of laminated lake deposits. The lakes were likely of slackwater origin, dammed in the Kaskaskia Valley by high sediment accumulations in the Mississippi Valley.

Pre-Illinois Episode sand and gravel in the Kaskaskia Valley have not been given a formal member name, so are referred to as Banner sand or Banner outwash. Both the Harkness Silt and Banner sand are subsurface units, and both occur predominantly in preglacial bedrock valleys, where they are preserved below pre-Illinois or Illinois Episode till deposits.

The thickest known succession of the Harkness Silt Member was observed in a core below Omphghent diamicton in a buried bedrock valley a few miles west of the modern Kaskaskia Valley in northern Randolph County (Grimley and Webb 2010). At this site, 55 ft (16.8 m) of calcareous, laminated, grayish brown to slightly pinkish silty clay to silty clay loam fills the preglacial valley. A slackwater lacustrine environment is interpreted based on the presence of several gastropod and bivalve (*Pisidium* sp.) shells (~2 to 10 mm in size), ostracodes, aquatic vegetation, and small fragments of conifer wood and needles (probably *Picea* sp.) that likely washed into a generally shallowing lake. Aquatic gastropods include *Fossaria* sp., *Gyraulus* sp., *Valvata tricarinata*, and *Probythinella lacustris*. Distinctive ostracodes in the basal, more pinkish brown portion of the Harkness Silt may indicate a relatively deep and low-turbidity lake, perhaps when loess deposition was minimal. The slight pinkish color may reflect a time when pre-Illinois Episode ice began to advance across the Lake Superior region, an area that could have provided an abundant source of pinkish-brown sediment.

Some pre-Illinois Episode sand deposits do occur in the lower Kaskaskia region (Grimley 2010), although delineation of these deposits from younger sand units is sometimes difficult when using descriptions of water-well logs. On uplands, pre-Illinois Episode loess deposits that occur above the till are generally included within the Yarmouth Geosol weathering horizons. Proglacial loess that occurs below Banner till is included within the Harkness Silt unit but is a few feet in thickness or less. However, much of the silt-rich sediment within Harkness Silt lacustrine deposits is likely derived from loess on adjacent uplands. A pre-Illinoian loess deposit with a pre-Yarmouth weathering profile, informally classified as the Burdick loess by McKay (1979) in Madison County, Illinois, is perhaps one glacial stage older than the Omphghent till. Although this unit has not been found as yet in the Kaskaskia Basin, possibly correlative loesses have been found in unglaciated southern Illinois (Wang et al. 2009).

Yarmouth Episode (Interglacial), Middle Pleistocene (~430 to 190 ka)

The Yarmouth Episode, as used here, includes the time of development of the Yarmouth Geosol, a widely recognized interglacial paleosol. The presence of the Yarmouth Geosol, preserved within the uppermost pre-Illinois Episode deposits, allows these units to be delineated from younger Illinois Episode deposits (Leverett 1898; Willman and Frye 1970). In addition to being a time of weathering and soil development processes, the Yarmouth Episode was a time for deposition of accretionary and alluvial deposits (**Lierle Clay Member, Banner Formation;** Figure Q4) in lowlands, and was a time of overall landscape dissection and erosion.

The wetland or poorly drained facies of the Yarmouth Geosol is typically preserved within the Lierle Clay Member (described below) and is found above Banner till deposits, although partially eroded in many places (Grimley and Phillips 2010). The well-drained to moderately drained facies of the Yarmouth Geosol, formed on highlands and prominent uplands, were mainly scoured by glacial erosion during the succeeding Illinois Episode glaciation. Thus, these facies of the geosol are rarely observed within glaciated portions of the Kaskaskia Basin. However, oxidized, well-drained, reddish brown Yarmouth Geosol profiles are found in unglaciated areas of southern Illinois (Grimley et al. 2003; Wang et al. 2009). Mineralogical and magnetic properties, representative of the duration of soil development, suggest that Yarmouth Geosol alteration and soil development were about triple those for the Sangamon Geosol (Grimley et al. 2003). Considering the degree of interglacial soil alteration and thickness, and the relatively deep dissection of pre-Illinois Episode till plains (Willman and Frye 1970), the time of the Yarmouth Episode may best correlate to OIS 11 through 7 (~425 to 180 ka), compared with OIS 5 (~125 to 75 ka) for the Sangamon Episode (Hansel and McKay 2010).

Stream dissection and erosion occurred for a considerable time during the possibly 200 ka or greater length of the Yarmouth Episode. This likely resulted in the scouring of valleys and lowering of stream base levels during the late Yarmouth Episode or possibly early Illinois Episode (Willman and Frye 1970). After this period of dissection, Illinois Episode deposits served to preferentially fill in these low areas with lake sediments, outwash, redeposited loess, till, and debris flows (Grimley and Phillips 2011).

Lierle Clay Member, Banner Formation [Accretionary Deposits]

Only one unit deposited during the Yarmouth Episode is formally recognized as a lithostratigraphic unit. The Lierle Clay Member of the Banner Formation is an accretionary deposit that occurs in paleo-lowlands or depressions and may be up to 15 ft (4.6 m) thick. This unit is clay-rich, leached of carbonates, and high in expandable clay minerals (Willman and Frye 1970). Pedogenic alteration, including clay skins, iron staining, soil structure, and mottled colors (greenish gray), along with some faint laminations within this unit, record the interglacial soil development of the Yarmouth Geosol in a lowland or wetland environment. Iron and manganese concretions are abundant in places, and <5% small pebbles of angular chert are typical.

Illinois Episode, Late Middle Pleistocene (~190 to 130 ka)

During the maximum glaciation of the Illinois Episode, glaciers advanced southwest to present-day downtown St. Louis (Goodfield 1965) and to near the confluence of the Kaskaskia Valley with the Mississippi Valley (Kolb 2010). Illinois Episode glaciers thus completely covered pre-Illinois Episode deposits in the Kaskaskia Basin (Figures Q3 and Q5), leaving behind a record of glacial diamicton, glaciofluvial sediments, and other ice-marginal sediments. On some bedrock highlands, the already thin and weathered pre-Illinois Episode deposits were removed by direct glacial or glaciofluvial erosion. In other areas, typically lowlands or tributary valleys, a full sequence of pre-Illinois Episode deposits containing a strongly developed Yarmouth paleosol was preserved and buried by the younger sequences. As the Illinois Episode glacial margin receded from its maximum extent and ice stagnated and thinned, a series of recessional morainic margins developed in the Kaskaskia Basin that are envisioned to be part of a regional sublobe (Webb 2009; Grimley and Webb 2010; Grimley and Phillips 2011), referred to here as the Kaskaskia Sublobe (Figure Q5d). Thinning glacial ice would have allowed the local bedrock topography to have an increasing influence on glacial flow, resulting in small lobate forms up to a few miles across that protruded from the overall margin. In this model, glacial sedimentation was enhanced in reentrants between the sublobes, where convergent ice flow would have concentrated direct glacial, glaciocolluvial (debris flows), and glaciofluvial sedimentation (subglacial and supraglacial stream outflow). Many prominent glacial ridges today (e.g., Shiloh Ridge, Turkey Hill in St. Clair County) that formed in these convex ice-marginal areas are composed of the **mixed facies of the Hagarstown Member, Pearl Formation** (Figure Q4; Grimley and Webb 2010; Grimley and Phillips 2011). Divergent glacial ice in the concave areas of the ice margin was more sediment starved, was more typically morainal in character, and tended to be composed mainly of fine-grained material with fewer sand bodies. These deposits are predominantly fine grained, with many sheared inclusions of Yarmouth paleosol and older materials, and are classified as a morainal facies of the **Glasford Formation.** Other types of ridges, more abundant within the interior of the Kaskaskia Basin, contain sandy or locally gravelly glaciofluvial deposits (**sandy facies of the Hagarstown Member, Pearl Formation**). Landforms containing such deposits include ice-walled channels (tunnel eskers or subaerial) and kames, and tend to be most abundant on the northwest side of the present-day Kaskaskia Valley. The occurrence of the three lithologic end members (Glasford morainal, Hagarstown mixed facies, and Hagarstown sandy facies) within constructional Illinois Episode glacial ridges in southwestern Illinois, and their related origin, will be a primary focus of this field trip.

Ice-blocked proglacial lakes, likely coalescing with preexisting slackwater lakes, developed and inundated tributary valleys to the lower Kaskaskia Valley, resulting in lacustrine sediment deposition before (**Petersburg Silt**) and after (**Teneriffe Silt**) Illinois Episode glacial advances. Ice-dammed or moraine-dammed lakes may have caused rising levels in the preexisting lakes, which may have been as extensive as the Wisconsin Episode version of glacial Lake Kaskaskia (Willman and Frye 1970). Consequently, slackwater lake sediments were extensively deposited in the lower Kaskaskia drainage basin as a result of Mississippi River aggradation (see section on glacial Lake Kaskaskia).

During and after the stagnation and final melting of Illinois Episode glaciers, proglacial outwash (**Mascoutah facies, Pearl Formation**) was deposited in the middle and lower Kaskaskia drainage basin and probably fed into the proglacial or slackwater lake system. Ice-marginal fans (Stop 2) and overflow channels with sand and gravel outwash contributed to the network of glaciofluvial deposits that fed into the Kaskaskia Basin, one of the prominent areas of discharge for glacial meltwater for the Illinois Episode ice sheet in Illinois. Ogles Creek and Richland Creek in St. Clair County appear to have developed as ice-marginal outwash channels on the west side of a prominent recessional morainic border (Grimley and Phillips 2011). As the Kaskaskia Sublobe receded further from a temporary margin (Figure Q5), other ice-marginal or proglacial meltwater streams drained through south- and southwest-flowing tributary valleys, such as those of Silver Creek, Sugar Creek, Shoal Creek, Beaver Creek, and Hurricane Creek, all feeding into the Kaskaskia Lowland and depositing Pearl Formation outwash (Figures Q0, Q1, D1, and D2).

Petersburg Silt [Sub-Glasford Lake Deposits and Loess]

Fine-grained, massive to stratified, and sometimes fossiliferous sediments preserved below Illinois Episode till deposits are classified as the Petersburg Silt (Willman and Frye 1970). This deposit of silt to silty clay with some fine sand beds is genetically and predominantly lake sediment, but also includes fine-grained alluvium and loess. Stratigraphically, the Petersburg Silt occurs below the Glasford Formation and above pre-Illinois Episode deposits (Willman and Frye 1970). However, the vast majority of this unit in the Kaskaskia region was deposited in proglacial lakes that formed as a result of ice-blockage or slackwater conditions. The Petersburg Silt is thus generally found in small lowlands and valleys on the pre-Illinois Episode paleolandscape. In the lower valley region, this unit is mainly interpreted as slackwater lake sediment resulting from the impoundment of the Kaskaskia Valley and its tributaries in response to Mississippi River valley aggradation during the advance of the Illinois Episode ice sheet (Grimley and Phillips 2011). The lake would have been present before burial by glacial ice and deposition of Glasford till and ice-marginal sediment. Fossil gastropods in the Petersburg Silt, indicative of relatively shallow lacustrine conditions and fluctuating water levels, have been noted across the lower Kaskaskia Basin (Geiger 2008; Grimley and Webb 2009). Numerous species of terrestrial and aquatic gastropods have been observed (Geiger 2008), such as those at the Ogles Creek Section (Stop 3).

Glasford Formation [Till and Ice-Marginal Sediment]

In southwestern Illinois, glacial diamicton, including till and debris flow deposits, are classified as the Glasford Formation (Willman and Frye 1970; Grimley and Phillips 2011). Although members of the Glasford Formation (Smithboro and Vandalia Members) have been differentiated in the Vandalia area by Jacobs and Lineback (1969), these units have not proved to be lithologically traceable to the west as the tills become gradually siltier and less illitic. The Fort Russell till member was differentiated informally in Madison and St. Clair Counties (McKay 1979), but this unit appears to be gradational with till units to the east, perhaps in facies, so cannot be clearly differentiated. We have thus chosen to use only the term Glasford Formation or Glasford till, rather than member names, until additional data are acquired and more research has been conducted in the middle Kaskaskia Basin (Bond, Clinton, and Fayette Counties). Our model also now envisions a series of recessional ice margins that all originate generally from a Lake Michigan Lobe source. Thus, the lithological variations in Illinois Episode tills of the Kaskaskia Basin are probably local in origin, reflecting the composition of overridden Pennsylvanian bedrock, pre-Illinois Episode deposits, or proglacial Illinois Episode water-laid deposits in central and southern Illinois.

Mixed and Sandy Facies, Hagarstown Member, Pearl Formation [Ice-Contact Deposits]

The Hagarstown Member, originally defined by Jacobs and Lineback (1969) and Willman and Frye (1970) within the Glasford Formation, includes poorly to well-sorted gravel, and sand interbedded with gravelly diamicton. The depositional environment of the unit suggested this was an ice-walled channel deposit or was of other glaciofluvial origin. This use of the Hagarstown Member appears to have evolved such that some Illinois Episode glacial ridges in the Kaskaskia Basin that were not known to be coarse grained became mapped by Lineback (1979) as Hagarstown Member, probably by geomorphic inference. Later, Killey and Lineback (1983) reclassified the Hagarstown Member within the Pearl Formation, rather than the Glasford Formation so that it is more analogous to the Wasco facies of the Henry Formation, which are mainly Wisconsin Episode ice-contact deposits. However, in the course of detailed mapping, it became clear that the constructional glacial ridges varied in lithologic character (Figure Q2). Deposits in ridges with mixed or alternating layers of coarse sand and gravel, fine sand, diamicton, silt, or silty clay were therefore classified informally as the Hagarstown Member mixed facies. If the material was predominantly diamicton (>75%), then the material was classified as the Glasford Formation. If the material was predominantly sand or coarser (>75%), then it was classified as the Hagarstown Member sandy facies. The sandy facies may include coarse-grained ice-contact sediments in areas of eskers, ice-walled channels, kames, or ice-marginal fans.

Mascoutah Facies, Pearl Formation [Outwash]

Of considerable significance in the Kaskaskia Valley are thick Pearl Formation sand deposits [up to 55 ft (16.8 m) thick], with some gravel, that occur in loess-covered terraces or in lowlands below younger Cahokia, Henry, and Equality Formation sediments (Grimley and Phillips 2011; units described below). Pearl Formation outwash, as defined by Willman and Frye (1970), is predominantly Illinois Episode in age, but basal portions may include undifferentiated pre-Illinois Episode fluvial deposits. The Mascoutah facies was informally defined by Grimley and Webb (2010) to more clearly distinguish Pearl Formation proglacial outwash deposits from more lithologically variable ice-contact facies sandy deposits (Hagarstown Member, Pearl Formation). Deposits classified as the Mascoutah facies consist of mostly fine to coarse sand and are typically horizontally stratified (Grimley 2010; Grimley and Webb 2010). The unit contains Sangamon Geosol weathering in generally fine-grained upper portions, if not eroded.

The Mascoutah facies can be traced lithologically in the subsurface for several miles from eastern St. Clair to northern Randolph Counties (Phillips 2008; Grimley and Webb 2009; Grimley 2010) and likely extends beneath much of the middle and lower Kaskaskia Valley. It also occurs in feeder tributary valleys (particularly those with south- or southwest-draining orientations), such as the Silver Creek and Shoal Creek valleys.

Grigg Tongue, Pearl Formation [Sub-Glasford Outwash]

The **Grigg tongue** was recently defined as a basal tongue of sand and gravel that occurs below the Glasford Formation in some areas of the lower Kaskaskia Basin, typically within a few miles of the present-day valley (Grimley 2010; Grimley and Webb 2010). The presence of this unit is based primarily on water-well and engineering boring logs, but data convincingly indicate its occurrence in the lower Kaskaskia Valley, and many private water wells tap into this aquifer material locally (Grimley and Webb 2010). The origin of this unit is likely proglacial sand and gravel deposited in front of an advancing Illinois Episode ice sheet. Sedimentologically and lithologically, it appears similar in character to the Mascoutah facies of the Pearl Formation but lies stratigraphically below the Glasford Formation. Because this unit conceptually connects to the Mascoutah facies beyond the limit of glaciation, it has been defined as a tongue of the Pearl Formation (analogous to the Ashmore tongue of the Henry Formation for Wisconsin Episode deposits; Hansel and Johnson 1996).

Teneriffe Silt [Late-Glacial Lake Sediment and Loess]

The Teneriffe Silt is a massive to stratified silt deposit that may include some clayey or sandy beds; it overlies other Illinois Episode deposits and contains the Sangamon Geosol in its top (Willman and Frye 1970). This is typically a lake deposit, but may also include late Illinois Episode loess and beds of fine-grained outwash. Although it was originally defined as only within the limit of Illinois Episode glacial deposits, it has since been extended to Illinois Episode lake deposits beyond the glacial border (Lineback 1979; Heinrich 1982; Grimley et al. 2009). The Teneriffe Silt has been observed and mapped as the surficial deposit (below Wisconsin Episode loess) in several isolated areas in southwestern Illinois. Areas of deposition may include proglacial lakes where meltwater was trapped between the receding ice and a recessional moraine, such as in northern Randolph County (Grimley and Webb 2010; Ostendorf 2010) and St. Clair County (Phillips 2004b). Other areas of Teneriffe Silt include proglacial lakes formed between glacial ice and areas of high topography in the Mississippian bedrock uplands of Monroe and southern St. Clair Counties (Phillips 2010; Grimley and Phillips 2011) or locally in ice-marginal water-laid deposits (Stop 2, this guidebook).

Sangamon Episode, Late Pleistocene (~130 to 60 ka)

The Sangamon Episode is the time represented by the development of the Sangamon Geosol into Illinois Episode deposits (Figure Q4). The original Sangamonian Stage was defined by this weathering zone (Leverett 1899; Follmer et al. 1979). In the Kaskaskia Basin of southwestern Illinois, the Sangamon Geosol is ubiquitous in upland areas (Jacobs and Lineback 1969; Grimley et al. 2001). Although sometimes partially truncated or eroded, the Sangamon profile is typically well preserved in upper portions of Illinois Episode deposits and separated from the modern soil profile by a 5- to 30-ft-thick (1.5- to 9-m-thick) protective blanket of Wisconsin Episode loess deposits.

The Sangamon Episode (mainly correlative with OIS 5) was a time of generally warm interglacial climate, probably similar to today’s climate on average or perhaps slightly warmer and more humid at times (Ruhe et al. 1974; Eyles and Clark 1988; Curry and Baker 2000). Climatic fluctuations within the time of the Sangamon Episode are known globally from the deep-sea record (Martinson et al. 1987) and have been recorded by variations in pollen and ostracode taxa assemblages in south-central Illinois (Curry and Baker 2000; Teed 2000; Curry et al. 2010b). The best known record of Sangamon Episode deposits in the U.S. Midwest comes from lake deposits in kettle basins in south-central Illinois. One of these sites, Pittsburg Basin (Stop 9), is visited on this field trip and has been extensively studied for paleoecological records (Grüger 1972b; Curry 1995; Teed 2000). On the basis of such studies, as well as the Alfisol expression of Sangamon Geosol profiles in southern Illinois and Indiana (Ruhe et al. 1974; Follmer 1983; Grimley et al. 2003), woodlands are envisioned to have been extensive during the last interglacial. The occurrence of species of more southern or eastern affinity (such as sweetgum and American beech, respectively), in addition to the typical oak-hickory biota, is evidence that the forest diversity was slightly greater during the Sangamon Episode than during the Holocene (Teed 2000).

Berry Clay Member, Glasford Formation (or Pearl Formation) [Accretionary Deposit]

The **Berry Clay Member** (Figure Q4) is a deposit of clay, silt, and sparse pebbles that was deposited as accretionary sediment in depressions and lowlands during the Sangamon Episode (Willman and Frye 1970). This deposit also includes syndepositional weathering of the Sangamon Geosol so that the sediment package is simultaneously a lithologic unit and an accretionary soil. In most areas, the Berry Clay Member contains characteristics of a wetland soil with iron mottling and gleying. It has been called an accretion-gley (Willman and Frye 1970); however, in some areas the unit has a more oxidized character, where drainage conditions were later improved with gradual stream dissection of the surrounding terrain. Sediment sources for the Berry Clay Member include redeposited late Illinois Episode loess and the uppermost till from surrounding areas on the till plain. The Berry Clay Member is typically found in isolated, closed depressions on till plains but may include broad areas as much as 1.0 mi (1.6 km) or so in diameter. The use of the Berry Clay Member in the Kaskaskia Valley has been extended to being an upper member of the Pearl Formation in areas where clay loam or sandy clay loam deposits containing the Sangamon Geosol occur above the Pearl Formation (Grimley 2010). In these areas, the sediment source for the Berry Clay Member is, again, mainly eroded loess that was redeposited into lowland areas. On shallow lake plains and floodplains, material similar to the Berry Clay Member is difficult to distinguish from the Teneriffe Silt, so these units have been mapped as a complex (Grimley and Phillips 2011). The sediment deposited in the Pittsburg Basin (Stop 9), and in similar kettle basins, during the Sangamon Episode could be classified as a Berry Clay Member-Teneriffe Silt complex.

Wisconsin Episode, Late Pleistocene (~60 to 12 ka)

During the last period of continental glaciation, the Wisconsin Episode, glacial ice did not reach the area, but glacial meltwater from ice margins in the upper Midwest deposited silt, sand, and gravel (outwash of the **Henry Formation**) in the Mississippi Valley and, to a lesser extent, in the Kaskaskia Valley. Silty waterlaid deposits in the Mississippi Valley were repeatedly entrained by prevailing westerly winds into intense dust clouds and, with subsequent settling of silt particles, deposited as a cover of loess (**Peoria and Roxana Silts;** Figure Q4). The loess is up to 90 ft (27 m) thick on the Mississippi River valley bluffs but thins southeastward to about 7 ft (2 m) thick on uplands adjacent to the Kaskaskia Valley (Figure Q1). Concurrent with loess deposition on uplands and outwash deposition in major valleys, a large slackwater lake called glacial Lake Kaskaskia formed in the Kaskaskia drainage basin up to about 425 ft (130 m) above sea level (asl), probably the result of high aggradation of outwash sediment in the Mississippi Valley. Radiocarbon dating of fossil gastropod shells and conifer wood in lake deposits (**Equality Formation;** Figure Q4), preserved in terraces and valley fill sequences, indicates the lake was at its maximum extent between about 25,000 and 15,000 calendar years ago (Grimley and Webb 2009, 2010).

Henry Formation [Outwash]

Fine to medium sand deposits found immediately below the Cahokia and Equality Formations in the Kaskaskia Valley are, in places, interpreted as the Henry Formation, such as at the Highbanks Road Section (Stop 6). The distinction between older and younger alluvial units can be subtle and difficult to differentiate in mapping (Grimley 2010). Overall, the Henry Formation (in the Kaskaskia Valley area) tends to be a bit coarser than the Cahokia Formation sand but has less coarse sand and gravel than the Pearl Formation, probably because of its closer proximity to the ice margin. However, overlap in the grain size of these units may be considerable. Sand in the Henry Formation can be noncalcareous or calcareous and may be intercalated with, overlain by, or underlain by calcareous silt loam beds of the Equality Formation.

Peoria and Roxana Silts [Loess]

Wisconsin Episode loess deposits in the Kaskaskia Basin (Figure Q1) range from about 30 ft (9 m) thick on the western edge of the basin (Grimley and Phillips 2011) to less than 5 ft (1.5 m) in the northeastern part of the basin (Fehrenbacher et al. 1986). The loess deposits consist of two units, the Peoria Silt and the Roxana Silt (Hansel and Johnson 1996). The Peoria Silt, ubiquitous across central U.S. uplands (Follmer 1996), is a tan to grayish brown silt loam containing modern soil development in its top. Less altered zones of the Peoria Silt, several feet or more below the surface, may be calcareous (mainly dolomitic) and locally contain fossil terrestrial gastropods. The Roxana Silt is a more pinkish brown- to grayish brown-colored silt loam. It directly underlies the Peoria Silt and is thinner, with a thickness about 20 to 60% that of the Peoria Silt. The Peoria and Roxana Silts are differentiated in part by the occurrence of the Farmdale Geosol (a weakly developed interstadial soil) in the top of the Roxana Silt, although this paleosol may be difficult to recognize where the total loess thickness is less than 10 ft (3 m). In the Kaskaskia Basin, the Roxana Silt is nearly always leached of carbonates. Because of their eolian origin, the Peoria and Roxana Silts drape older Illinois Episode units across the landscape with a relatively uniform thickness locally (except where eroded). The Sangamon Geosol normally underlies the Roxana Silt and separates these Wisconsin Episode loesses from Illinois Episode deposits. Peoria Silt deposition was concurrent with portions of the Equality Formation lake deposits (see below) and locally interfingers with, as well as partially overlies, such deposits on terraces. Intact loess deposits are not found on the younger Holocene surfaces, so the presence of a loess cover can help delineate ages of geomorphic surfaces.

Equality Formation [Lake Sediment]

The Equality Formation includes generally fine-grained sediment (silt and clay), typically faintly laminated, that was deposited as lake sediment (Willman and Frye 1970; Hansel and Johnson 1996). The environment of deposition in the Kaskaskia Basin was primarily as a slackwater deposit as the Mississippi River aggraded significantly in its valley during the last glacial maximum (LGM; see additional details in the Geologic History of Glacial Lake Kaskaskia section). Other tributary valleys of the Mississippi Valley, such as the Big Muddy Valley (Trent and Esling 1995), were similarly back-flooded and infilled. In calcareous, unoxidized zones (at depth) of the Equality Formation in the Kaskaskia Basin, it is common to find preserved organic debris, fossil conifer wood, and mollusk and ostracode shells that are representative of fluctuating lake levels in a shallow, freshwater lake (Geiger 2008; Grimley and Webb 2009). A site with thick deposits of fossiliferous Equality Formation will be seen on Stop 6 of this field trip (Highbanks Road Section).

Holocene (12 ka to the Present)

Near-surface Holocene deposits (from the last 11,700 years; Walker et al. 2009) are up to 30 ft (9 m) thick in the Kaskaskia Valley and include various alluvial facies of the Cahokia Formation, from fine sandy point bar and channel deposits to fine-grained overbank deposits and silty clay abandoned meander fills. Many of the tributary streams are dominated by silty sediment consisting of mainly redeposited loess. An early- to mid-Holocene terrace [~395 ft (~120 m) asl] is present in the lower Kaskaskia Valley and nearby tributaries. This terrace contains probable early or middle Holocene postglacial deposits (high-level clayey facies), which may have resulted from slackwater conditions or extensive overbank flooding. The lower Kaskaskia River was significantly straightened and channelized from Fayetteville southward by the removal of several natural river meanders for a navigation project in 1974. In advance of the project, archeological studies were conducted, and numerous sites with projectile points, tools, and fire-cracked rocks were noted in the floodplain, terraces, and adjacent lands (Conrad 1966). From human activity during the past 150 years, the landscape of the Kaskaskia Basin contains significant areas of anthropogenic fill in urban areas, landfills, dams, artificial levees, former strip mines for coal, limestone quarries, aggregate mines, and interstate interchanges.

Cahokia Formation [Alluvium]

Clayey and sandy facies of the Cahokia Formation, at times interstratified, have been differentiated in the large valley meander belts of the postglacial Kaskaskia Valley (Grimley and Phillips 2011). Near-surface deposits consist of interstratified fine to medium sand with silt loam, silty clay loam, and silty clay. Sandy deposits [up to 25 ft (8 m) thick] in channels and point bars of the Kaskaskia River are mapped as **sandy facies of the Cahokia Formation.** These deposits are typically fine to medium sand, moderately well sorted, and noncalcareous. They range in age from recent to possibly several thousand years old (mid to early Holocene) at higher elevations and in the subsurface. The **clayey facies of the Cahokia Formation** includes silt loam, silty clay loam, and silty clay in overbank deposits, swale fills, and abandoned meander fills on the modern floodplain. This unit is also present as a cap on low terraces in the lower Kaskaskia Valley (Phillips 2008; Grimley and Webb 2010).

ORIGIN OF GLACIAL RIDGES (RIDGED DRIFT) IN SOUTHWESTERN ILLINOIS

Historical Studies and Observations

Glacial ridges in southwestern Illinois, historically known as the Ridged Drift (Leverett 1899), have now been studied for more than a century by various researchers. The conical to elongate hills are relatively abundant and prominent in southwestern Illinois, especially immediately west of the Kaskaskia River valley, where they can be up to a few miles long, a mile wide, and 30 to 150 ft (9 to 46 m) in relief above the Illinois Episode till plain (Figures Q0, Q2, and D1). Several bands of prominent ridges and hills, varying from continuous to fragmentary and elongate to varying degrees, are found around the margins and within the interior of the Kaskaskia Basin (Figures Q1 and Q2). In many areas, scattered or isolated knolls (low conical hills) are found in the vicinity of the more elongate ridges. Some ridge belts appear to parallel river systems, some appear to align in arcuate or lobate patterns on a regional scale, and some appear to protrude oddly and singly from the surrounding till plain.

Through the history of investigation and mapping of the ridges, they have been interpreted as predominantly morainal (Leverett 1899; MacClintock 1929; Willman et al. 1963; Phillips 2008); as eskers, esker-like, or ice-walled channels (Ball 1940; Jacobs and Lineback 1969; Burris et al. 1981); as crevasse fills (Leighton 1959; Leighton and Brophy 1961); or as some combination of the above (Stiff 1996; Grimley and Webb 2010). The interpretations of researchers seem to reflect regions or aspects of the Kaskaskia Basin that were studied most intensely (e.g., central axis of basin, marginal ridges of basin). Leverett (1899) hypothesized that the prominent northeast- to southwest-trending ridge system on the western side of the Kaskaskia Basin marks the western border of a lobe persisting in southern Illinois after ice had retreated from western Illinois. MacClintock (1929) mainly supported this hypothesis, noting the morainic topography in the region with outwash extending westward from the ridge deposits. In subsequent years, however, Ball (1940) concluded the ridge system likely had a glaciofluvial origin, based on several observations of sorted sediment. Jacobs and Lineback (1969) similarly favored a glaciofluvial origin, based on observations in sand and gravel pits in the Vandalia area, and they envisioned ice-walled channels with subsequent collapse causing an influx of supraglacial debris. This model could explain the sedimentological complexity of ice-contact deposits in the Vandalia area ridges, which appear to include ice-walled channels or eskers (either subglacial or subaerial). The largely coarse-grained deposits in this area (field trip Stops 7 and 8) were termed the Hagarstown Member (Jacobs and Lineback 1969; Willman and Frye 1970; Killey and Lineback 1983). The glaciofluvial or eskerine origin of another ridge, in the vicinity of Taylorville, was supported by the geophysical data of Burris et al. (1981).

In contrast, Willman et al. (1963), in partial agreement with Leverett (1899) and MacClintock (1929), hypothesized that the ridges constitute an interlobate morainic complex, based on slightly different mineral compositions in tills east and west of a prominent ridge belt. Still often cited today, Leighton (1959) and Leighton and Brophy (1961) strongly argued that the ridge system has characteristics of a crevasse fill system with related moulin kames. They concluded that ridge orientations are parallel to ice flow (based on available striation data) and that a local trellis drainage pattern was genetically related to crevasses. More recently, a study of ridge geomorphology and a GIS analysis of water-well log descriptions in a portion of the ridge system northwest of Vandalia (Stiff 1996) highlighted the varying subsurface composition of ridge deposits, ranging from predominantly fine grained in the Nokomis area to predominantly coarse grained in the Vandalia area (Figure Q2). Recent 1:24,000-scale surficial geologic mapping in southwestern Illinois has further documented the spatial variability in texture, composition, and geomorphology of ridges in the southwestern part of the Kaskaskia Basin (Phillips 2008; Grimley and Webb 2009; Grimley 2010).

Ice-Directional Indicators

The direction of glacial ice flow in the Kaskaskia Basin during the Illinois Episode is recorded at several sites with till fabric measurements and rare instances of striations (Figure Q1). Striations are the most reliable ice-directional indicators but are rarely observed because of limited bedrock exposure. Striations were observed by Leighton (1959) mainly in limestone quarries, with a few additional observations by Grimley et al. (2001), E.D. McKay (personal communication, 2011), and others. Most striations have been observed in bedrock around the margins of the basin, where limestone bedrock highlands were scoured. Some degree of caution is needed because some striation directions may possibly reflect pre-Illinois Episode glaciation. Till fabric data from Glasford till in southwestern Illinois consist of data obtained by Lineback (1971), who noted mainly southwestern flow directions in the upper Kaskaskia Basin, and by Webb (2009), who noted more variable directions in the lower part of the basin (Figure Q1).

Current Working Hypothesis—Kaskaskia Sublobe or Ice Stream

Our current hypothesis is that a glacial sublobe, here referred to as the Kaskaskia Sublobe, episodically stabilized at successive positions in the Kaskaskia Basin during overall ice-margin retreat during the waning phase of Illinois Episode glaciation (late OIS 6). Consequently, a series of constructional ridges (morainic, eskerine, kamic, etc.) were formed in association with various ice-marginal positions as the sublobe progressively but episodically retreated (Figures Q1 and Q5). Active and stagnant ice conditions varied both spatially and temporally within the sublobe. Meltwater stream outlets, overflow channels, and ice-marginal lakes formed during melting phases and added to the complexity of deposits in the Kaskaskia Lowland. Minor readvances of glacial ice sometimes overrode and obscured earlier deposits and landforms.

During the glacial maximum of OIS 6 (Illinois Episode), a regional lobe emanated from the Lake Michigan basin and overwhelmed the Kaskaskia Basin as it flowed southwestward to its terminus near present-day St. Louis, Missouri, and Valmeyer, Illinois (Figure Q5b). Later, as the ice downwasted and thinned, an area of active ice may have become restricted to the Kaskaskia Basin lowlands (Figure Q5c,d), perhaps with stagnant melting ice along the sides of the basin. If stagnant or slower moving ice was adjacent to the Kaskaskia Sublobe, some of the lateral marginal moraines might be ice stream shear-margin moraines (Stokes and Clark 2002). When several observations are taken into consideration (listed below), a case can be made that the Kaskaskia Sublobe was, in fact, an ice stream during a portion of its history (Webb 2009). The geographic dimensions of the Kaskaskia Basin and the presence of a preexisting substrate of soft shale, fine-grained till, and lake deposits would have been favorable for the development of an ice stream (Stokes and Clark 2001). A scenario can be envisioned in which a resurgent ice stream flowed into the basin, with thinner, more stagnant ice decaying on the adjacent bedrock-cored highlands (Figures Q3 and Q5d). The complexity of deposits on the northwestern margin of the basin may have similarities to sedimentary processes in the Kettle Moraine region of southeastern Wisconsin during the last glaciation (Carlson et al. 2005). The occurrence and geometry of some Illinois Episode recessional moraines in the Kaskaskia Basin also appear similar in form to some of the last glaciation in central Illinois (Johnson and Hansel 1999), such as the Cerro Gordo Moraine. Although differences are found in glacial processes and landforms between the Illinois and Wisconsin Episode glaciations in Illinois, they may be less than once thought.

The presence of a sublobe or ice stream in the Kaskaskia Basin can explain a number of glaciological or geomorphic features:

1. *recessional morainal ridges* formed progressively as ice downwasted and receded to the northeast;
2. *push moraines* formed at the toe of the sublobe, suggesting active ice conditions (Grimley and Webb 2010);
3. *ice-contact sediment in eskers and kames* is more prevalent in the interior of the sublobe, evidence of stagnation (Jacobs and Lineback 1969);
4. *ice-marginal drainage* and overflow channels formed (Phillips 2004b, 2008);
5. *proglacial lakes* (depositing Teneriffe Silt) were temporarily formed by ice blockages (Grimley and Webb 2010; Ostendorf 2010);
6. *curved stream valleys* appear to follow former ice margin positions (Figure Q0); and
7. *large-scale flutelike features* appear on statewide digital elevation models (Luman et al. 2003) in the middle-upper Kaskaskia Basin.

More details on this hypothesis are discussed in Stops 1, 2, 5, 7, and 8.

GEOLOGIC HISTORY OF GLACIAL LAKE KASKASKIA

The lower reaches of the Kaskaskia Basin, here referred to as the Kaskaskia Lowland [<450 ft (<137 m) elevation southwest of Carlyle Lake; Figure Q0], contain a cyclical Quaternary record of fluvial, lacustrine, and glacial deposits from the past approximately 500 ka, with all deposits overlying Paleozoic bedrock. During this time, the Kaskaskia Lowland (including the modern valley and its ancestors) experienced a succession of cut-and-fill sequences, with overall aggradation during glacial episodes and predominantly incision and removal of deposits during the interglacial times (Figure Q5), although with some exceptions. During continental glaciations that advanced south of the Great Lakes (pre-Illinois, Illinois, and Wisconsin Episodes), widespread back-flooding occurred in the Kaskaskia Lowland in response to high aggradation in the Mississippi Valley [as much as 70 ft (21 m) above the current river level]. The timing of formation of the large slackwater lakes was concurrent with extensive outwash and loess deposition along the Mississippi Valley. Back-flooding and the formation of slackwater lakes in the Kaskaskia Lowland were particularly extensive owing to the very low gradient of the valley. Today's Kaskaskia River valley drops only about 60 ft (18 m) in elevation from areas near Carlyle [~415 ft (127 m) asl] to its confluence with the Mississippi River [~355 ft (108 m) asl], about 65 mi (105 km) of straight valley distance downstream. Thus, the average gradient of the modern Kaskaskia Valley is <1 ft/mi (not considering meanders). During interglacials (Yarmouth and Sangamon Episodes, Holocene), the Kaskaskia River and its tributaries generally incised the valley fills in response to periods of downcutting of the Mississippi River (Curry and Grimley 2006). Although the term glacial Lake Kaskaskia was originally defined as a slackwater lake that formed during the peak of the Wisconsin Episode glaciation (Willman and Frye 1970), we have here extended the use of this name to earlier Illinois and pre-Illinois Episode occurrences of slackwater lakes in the Kaskaskia Lowland.

The former flooding of the Kaskaskia Lowland by glacial Lake Kaskaskia during the Wisconsin Episode has been well known for decades (Shaw 1921; Willman and Frye 1970). The timing of this most recent occurrence of glacial Lake Kaskaskia was coincident with many other slackwater lakes in tributary valleys to the Mississippi, Illinois, Ohio, and Wabash River valleys during the LGM (Willman and Frye 1970; Frye et al. 1972; Lineback 1979; Moore et al. 2007; Grimley et al. 2009). Rising base levels in the Mississippi Valley began in the mid-Wisconsin Episode at least as early as approximately 45 ka (Curry and Grimley 2006), when glacial ice was probably in the Lake Superior region, and perhaps as early as 55 ka, the estimated basal age for Roxana Silt deposition in the region (McKay 1979; Leigh 1994). During the initial phases of glacial Lake Kaskaskia, a finger-like slackwater lake was probably restricted to areas in Randolph County and perhaps southern St. Clair County. Increasingly, higher base levels during the LGM caused further blockage and led to the maximum extent of glacial Lake Kaskaskia probably extending as far northeast as the Carlyle area (Figure Q5g). The peak extent of the lake was probably between about 25,000 and 15,000 calendar years ago, based on radiocarbon dating of fossil gastropod shells and conifer wood in the Equality Formation lake deposits (Grimley and Webb 2009, 2010). This timing is approximately synchronous with Peoria Silt deposition (McKay 1979, 1986; Wang et al. 2003) and overlaps the timing of high terrace levels and peak flooding in the St. Louis region at approximately 23 to 19 cal ka (Hajic et al. 1991) in the Ohio-Wabash system (Frye et al. 1972) and the lower Mississippi River valley (Grimley et al. 2009). Typical elevations for the terrace that contains glacial Lake Kaskaskia deposits range from about 410 to 430 ft (125 to 131 m) asl in St. Clair County (Phillips 2008; Grimley 2010). Multiple terrace levels do exist (Grimley 2010; Soil Survey Staff 2015), but tracing them regionally is difficult.

Early versions of glacial Lake Kaskaskia were formed during the Illinois and pre-Illinois Episodes. Although less is known about the earlier lakes, the Wisconsin Episode version of the lake can be used as an analogue to help understand the older lakes. Paleontological records have also been very helpful in providing independent evidence for depositional environments. The Illinois Episode version of glacial Lake Kaskaskia was probably initially of slackwater origin, based on paleontological and sedimentological records from Petersburg Silt deposits at the Prairie du Pont Section (Grimley et al. 2001), Ogles Creek Section (Stop 3), NAW-2 core (Phillips 2008), and other localities. Well-preserved mollusk and plant macrofossil records at these localities give an indication of fluctuating levels in a generally shallow, alkaline lake [probably 0 to 10 ft (0 to 3 m) deep in many areas]. Paleoenvironments varied from shoreline to near-shore and deeper lacustrine. During periods of receding lake levels, fluvial or terrestrial conditions (with loess deposition) may have prevailed.

One important distinction in the character of the Illinois Episode glacial Lake Kaskaskia is that the approaching glacier eventually affected conditions in the lake (burying the Petersburg Silt with debris flow and till deposits). In many areas, the influence of loess and glacial silt is apparent in upper Petersburg Silt deposits and is inferred to be a result of increasing amounts of surrounding loess washing into the lake or from increased sedimentation from an approaching glacier. Small pebbles that could be dropstones are rarely found, in contrast to some pebble- to cobble-sized stones in Petersburg Silt deposits west of the Kaskaskia Basin (Grimley et al. 2001; Grimley and McKay 2004). A shoaling of the lake is apparent, based on the molluscan record in upper Petersburg Silt (e.g., more *Fossaria* sp.; less *V. tricarinata*), before the lake was covered by glacial ice and till deposition. As much as 100 ft (30 m) of Petersburg Silt, much of slackwater origin, was documented in a core in the New Athens West Quadrangle (Phillips 2008).

The pre-Illinois Episode version of glacial Lake Kaskaskia appears to have existed for a considerable time as well, based on a core in the Red Bud Quadrangle. In this core, the presence of 55 ft (17 m) of fine-grained, fossiliferous Harkness Silt implies that a large slackwater lake existed during the pre-Illinois Episode (Grimley and Webb 2010) as it did during the later glaciations. Initially, a relatively deep, low-turbidity lake is suggested by fossil mollusks in the basal Harkness Silt. During deposition of the basal zone, the pre-Illinois Episode ice had likely not yet advanced significantly into the Mississippi River drainage basin. The molluscan fauna of the upper Harkness Silt is consistent with a shallow freshwater lake with fluctuating water levels. The mapped extent of Harkness Silt (mainly lake deposits) in St. Clair County was shown in Grimley and Phillips (2011).

GEOPHYSICAL SURVEYS: TWO-DIMENSIONAL RESISTIVITY IMAGING

Timothy H. Larson

The Illinois State Geological Survey (ISGS) has been conducting electrical resistivity surveys in central Illinois for more than 70 years (e.g., Hubbert 1934; Bays 1946; Buhle and Brueckmann 1964; Heigold et al. 1985). The method exploits the strong contrast in electrical conductivity between clay (high conductivity/low resistivity) and sand (low conductivity/high resistivity), with especially high resistivity in dry sands. Historically, most of these surveys were specifically designed to assist in locating shallow groundwater supplies. Typical acquisition strategies concentrated on covering large areas that had relatively sparse information on subsurface geology. The goal was to locate areas of anomalously high resistivity that might correspond to shallow aquifers. Taking advantage of advances in modern electronics over the past five field seasons (≈2007–2011), we used a different approach to acquire detailed resistivity profiles as a tool for mapping Illinois Episode ridges in southwest Illinois. The method, known as two-dimensional resistivity imaging, uses a computer-controlled acquisition system to obtain nearly continuous resistivity measurements in two dimensions and to produce a simulated cross section beneath the resistivity profile. The goal of the profiling is to image the sediments across or along the ridges to determine the bulk composition of the various ridges (coarse grained or fine grained), the lateral variations in composition, and the approximate geometries of sediment bodies. In addition, the method can be used to map deposits of fine-grained lacustrine and alluvial sediments between ridges. The resistivity profiling method is not as reliable for mapping the buried bedrock surface in the Kaskaskia Basin. The lithology of Pennsylvanian- and Mississippian-aged bedrock is characterized by extremes in resistivity—shale and coal have low resistivity, whereas sandstone and limestone have high resistivity. Depending on the lithology of the bedrock surface at any particular location, it may or may not have a resistivity contrast with the overlying sediments. Furthermore, resolution of the resistivity profiling method decreases with increasing depth. Although the bedrock surface is interpreted in several of the images presented in this report, the location of this surface is only approximate. Electrical resistivity surveys were conducted in areas around Stops 2, 5, and 7.

PART II: Field Trip Itinerary, Findings, and Discussion: Road Logs and Field Stops

DAY 1 ROAD LOG (May 21, 2011)

Leave Mariner's Village Motel (8 a.m.—load buses at 7:45 a.m.)

- Turn right on William Road. [0.5 mi]
- Turn right on IL-127. [7.8 mi]
- Turn left on Keyesport Road. [1.0 mi]
- Turn left on the gravel road into the area of the Keyesport Sand and Gravel pit.

STOP 1: Keyesport Sand and Gravel Pit (see Figure D1)

- Turn right and head east on Keyesport Road. [1.0 mi]
- Turn left on IL-127 North. [2.0 mi]
- Turn left on IL-143 West. [12.3 mi]
- Turn left on Picatte Street. [0.3 mi]

REST STOP: Pierron City Park

- Go west on Edward Street. [0.1 mi]
- Turn right onto Bauman Road. [0.3 mi]
- Turn left on IL-143 West and enter Madison County. [1.8 mi]
- At the stop sign, turn left on IL-143 West/US-40 West. [3.1 mi]
- After the traffic circle, continue straight onto US-40 West. [7.4 mi]
- Exit right on the ramp to IL-4. [0.3 mi]
- Turn left on IL-4 South. [4.4 mi]
- Turn right on Meriwether Road to the circle at the end in Archview Estates.

STOP 2: Terrapin Ridge

- Turn right on IL-4 South. [1.4 mi]
- Turn right on Widicus Road. [0.6 mi]
- Continue on Old Lebanon–Troy Road. [0.4 mi]
- Stay right and continue on Old Lebanon–Troy Road. [0.7 mi]
- Take a left to stay on Old Lebanon–Troy Road. [3.0 mi]
- Turn left onto Scott Troy Road. [0.7 mi]
- Take the first left onto Weil Road. [1.1 mi]
- Continue to the end of the road at the small gate.

STOP 3: Ogles Creek Section

- Turn around and go west on Weil Road. [1.1 mi]
- Turn right onto Scott Troy Road. [0.7 mi]
- Turn right onto Old Lebanon–Troy Road. [3.0 mi]
- Turn right to stay on Old Lebanon–Troy Road. [0.7 mi]
- Bear left on Old Lebanon–Troy Road. [0.4 mi]
- Take the first right onto Widicus Road. [1.1 mi]

LUNCH STOP: Homer Recreation Park

- Turn right and head south on Widicus Road. [0.4 mi]
- Take the second left onto Acorn Way. [0.4 mi]
- Turn left on IL-4 North. [0.5 mi]
- Turn right onto Midgley Neiss Road. [1.5 mi]
- Turn left onto Emerald Mound Grange Road. [0.3 mi]
- Turn left into the long driveway.

STOP 4: Emerald Mound

- Head south on Emerald Mound Grange Road. [1.9 mi]
- Turn right onto US-50 West [1.5 mi] and enter the historic town of Lebanon.
- Turn left onto IL-4 South [9.4 mi] and pass through the town of Mascoutah.
- Turn right onto Grodeon Road. [1.2 mi]
- Turn left onto Brickyard Road. [0.3 mi]
- Turn right into the park.

REST STOP: Silver Creek Nature Preserve

- Head south on Brickyard Road. [3.0 mi]
- Turn right (west) onto Pleasant Ridge School Road. [0.5 mi]
- Follow the turn to the south on Pleasant Ridge School Road. [0.5 mi]
- Park buses on the gravelly area at the bend in the road.

STOP 5: Pleasant Ridge Area

- Head west on Pleasant Ridge School Road. [1.4 mi]
- Turn left onto Karch Road. [2.2 mi]
- Turn left onto IL-4 North. [3.7 mi]
- Turn right onto Town Hall Road. [2.3 mi]
- Turn right onto Highbanks Road. [2.0 mi]
- Turn right onto Slm Road. [0.2 mi]

STOP 6: Highbanks Road Section

- Head southeast on Slm Road. [0.2 mi]
- Turn left onto Highbanks Road. [4.0 mi]
- Turn right onto IL-177 East. [3.5 mi]
- Turn left onto IL-160 North. [9.1 mi]
- Turn right to merge onto US-50 East. [17.2 mi]
- Turn left onto IL-127 North. [0.2 mi]
- Take the first right onto William Road. [0.5 mi]

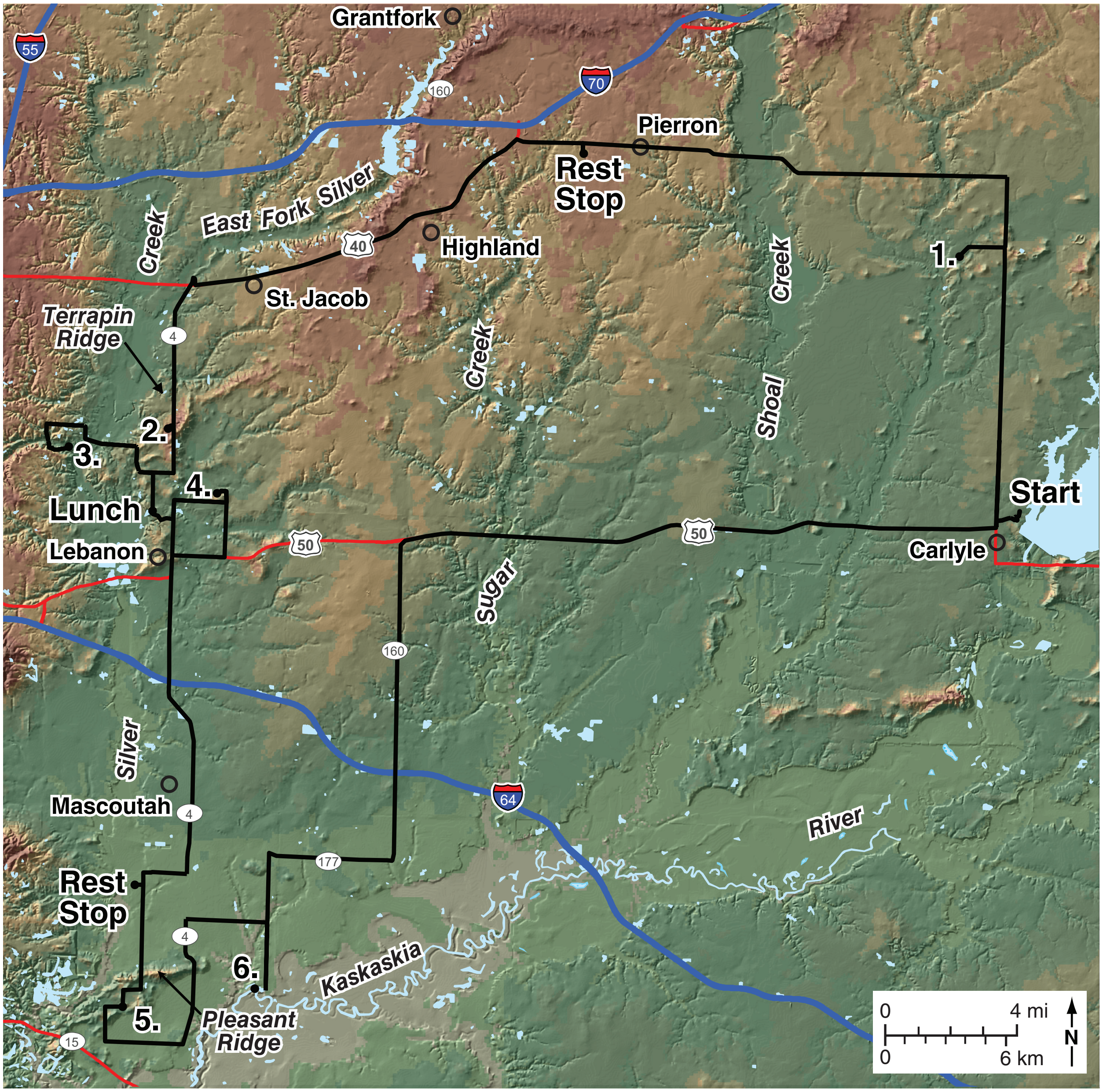


Figure D1 Location map of planned trip route for Day 1 of the field trip. Colors are from the statewide digital elevation map (30 m resolution).

STOP 1: Keyesport Sand and Gravel Pit

David A. Grimley, Andrew C. Phillips, and Nathan D. Webb

Please use caution at this site—do not walk or excavate beneath the undercut or steep slopes; sand is soft and caves in easily.

Overview

The Keyesport Sand and Gravel pit is located in an area that contains (or formerly contained) numerous small, conical or irregularly shaped hills dispersed among more streamlined or elongate hills (Figures D1 and 1.1). Mining has been continuing on a southwestern track for the past several years to follow a coarser tract of sand and gravel near the border of Bond and Clinton Counties (James Koerkenmeier, Plant Manager, Keyesport Sand and Gravel LLC, personal communication, 2011). Keyesport Sand and Gravel LLC, a division of Shakespeare Oil Company, has been mining in the vicinity since 1978, beginning with a hill (now gone) at the northwest corner of Illinois Route 127 and Keyesport Road. The current pit, where mining has been ongoing in Clinton County (on the south side of Keyesport Road) for the past several years, extends about 2,500 ft (750 m) in a southwesterly direction (Figure 1.1), with a width of about 820 ft (250 m) and a maximum depth of about 100 ft (30 m). Earlier mining is now evidenced by a northeast- to southwest-trending man-made lake on the north side of Keyesport Road in Bond County. This lake extends more than 2,500 ft (750 m) northeast of the northern extent of the current pit. In the areas mined, it is important to note that the sand and gravel deposits, in places up to 100 ft (30 m) thick, have been present in areas that were previously hills as well as in areas in direct line between the hills. Thus, coarse-grained deposits are not exclusive to the hills and ridges. Several other smaller kamic hills have been mined by various operations in areas to the east and northeast of the Keyesport pit, in the direction of Carlyle Lake (formerly the Kaskaskia Valley before the lake was created by the U.S. Army Corps of Engineers in the 1960s).

Many other pits in the area (e.g., Stops 7 and 8) utilize dredge operations because most deposits extend below the water table. Extensive exposures are viewable at the Keyesport pit because the groundwater is being pumped continuously to allow for dry mining. In areas where recent slumping has not affected the exposures, a panoramic view of the geological strata is afforded, at least on three sides (southeast wall, southwest wall, and northwest wall). About 90% of the sand and gravel deposits at Keyesport are used for making concrete. Of the remaining 10%, some fine sand is used for mortar sand, and some of the gravel is used for decorative landscaping (James Koerkenmeier, personal communication, 2011).

The overall sediment package that is being mined varies from well-sorted, bedded to cross-bedded, fine- to medium-grained sand, to coarse sand and gravel channel deposits (Figures 1.2 and 1.3). Over the various areas of exposure of Illinois Episode deposits, one can find planar-bedded fine sand, coarse gravelly sand channel deposits, diamicton inclusions, and silty clay lake sediments interbedded with sand layers. The thickness of extractable sand and gravel varies from about 40 to 100 ft thick (12 to 30 m thick), depending on the current and paleo-landscapes (which typically overlie an erosion surface on Glasford till). The sand was clearly deposited during the Illinois Episode, based on the presence of a strong interglacial paleosol (Sangamon Geosol) with its solum developed in the upper 6 to 8 ft (2 to 2.5 m) of the sand. The sand and gravel are thus classified as the Pearl Formation, predominantly within the Hagarstown Member (sandy facies). The Sangamon Geosol is a reddish brown to gray (depending on landscape position) silty clay loam to clay loam (becoming sandier toward the base) with a moderate blocky soil structure. Leaching and clay infiltration extend several feet below the paleosol surface, and an irregular leaching front can be seen at the top of the pit walls. Thus, clean sand suitable for extraction occurs at a minimum of about 15 ft (5 m) below land surface (Figures 1.2 and 1.3). The Pearl Formation and Sangamon Geosol are blanketed by about 5 ft (1.5 m) of loess (Peoria and Roxana Silts), consisting of friable tan to brown silt loam and containing modern soil development. However, the upper 10 ft (3 m) of deposits (loess and upper Sangamon Geosol) is stripped back from the main highwall and is not visible from inside the pit.

The exposures at the Keyesport pit were visited several times by the authors between 2004 and 2011 (Webb et al. 2012). A few general comments can be made, based on observations during successive visits, regarding the three primary walls of the active pit: the southeast, southwest, and northwest walls.

Southwest Wall (active mining wall, sand and gravel, high-angle reverse faults)

The southwest wall at the Keyesport pit is the active mining wall, with progressive excavation ongoing in a southwesterly direction. The coarsest and thickest deposits of sand and gravel are found on the southwest wall, with grain sizes ranging mostly from medium to

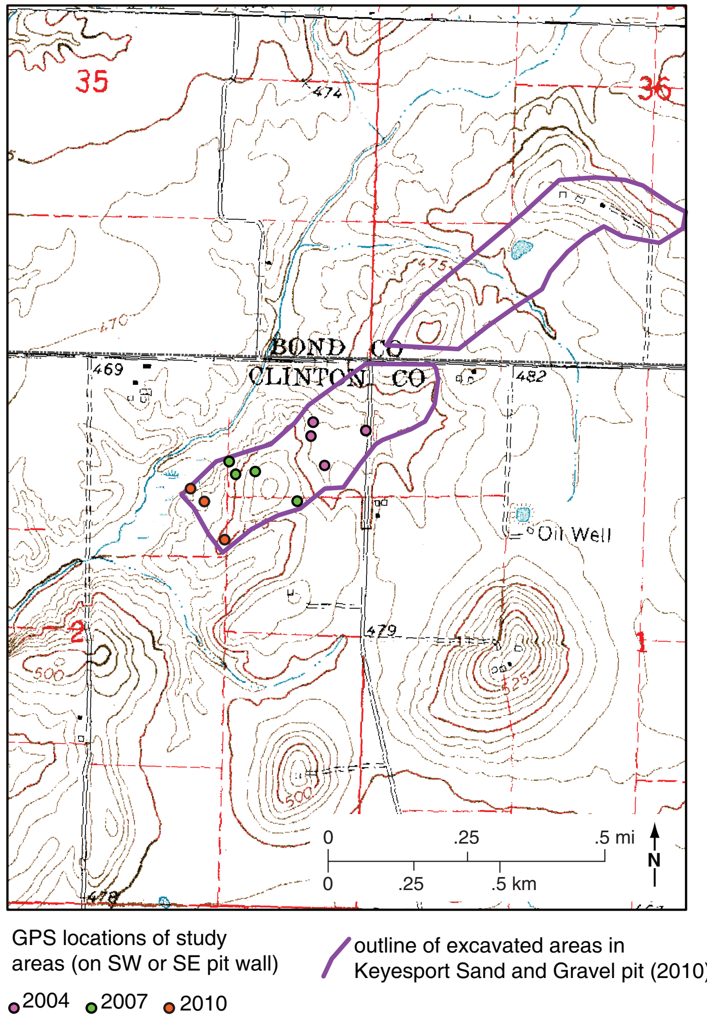


Figure 1.1 Location map for the Keyesport Sand and Gravel pit, Bond-Clinton Counties, Illinois (Stop 1), on the Stolletown U.S. Geological Survey 7.5-minute quadrangle. The GPS locations of some observations during the past several years are shown (2004–2010). Mining has progressed to the southwest following the coarser sand deposits. Topographic map courtesy of the U.S. Geological Survey.

coarse sand, with up to 15 to 20% gravel in some areas. Prominent trough cross-bed sets, up to a few inches thick and dipping generally to the southwest, were found along the axis of the ridge, near the center of the southwest highwall (Figure 1.3a). The total thickness of sand deposits was as much as 100 ft (30 m) thick in previous years, when the mining proceeded through the crest of the glacial ridge. The sand and gravel have been observed to overlie a dense, subglacial, pebbly loam diamicton (Glasford Formation till) at the base of the central part of the pit. In a few areas, a thin deposit of coarse sand with gravel was observed as a basal lag or coarse zone immediately above the Glasford Formation. The elevation of the Glasford till surface rises, in some places quite sharply, toward the northwest; thus, the thickness of sand decreases toward this direction.

High-angle reverse faulting was noted in the southwest highwall on numerous occasions from 2004 through 2011 (Figure 1.3b), particularly in sand deposits beneath high areas of the ridge. In 2008, near the top of a well-exposed highwall, beds of laminated silt were observed within the sand that had been cut and offset by a high-angle reverse fault. This type of faulting is similar to that noted by McDonald and Shilts (1975) in ice-proximal glaciofluvial environments.

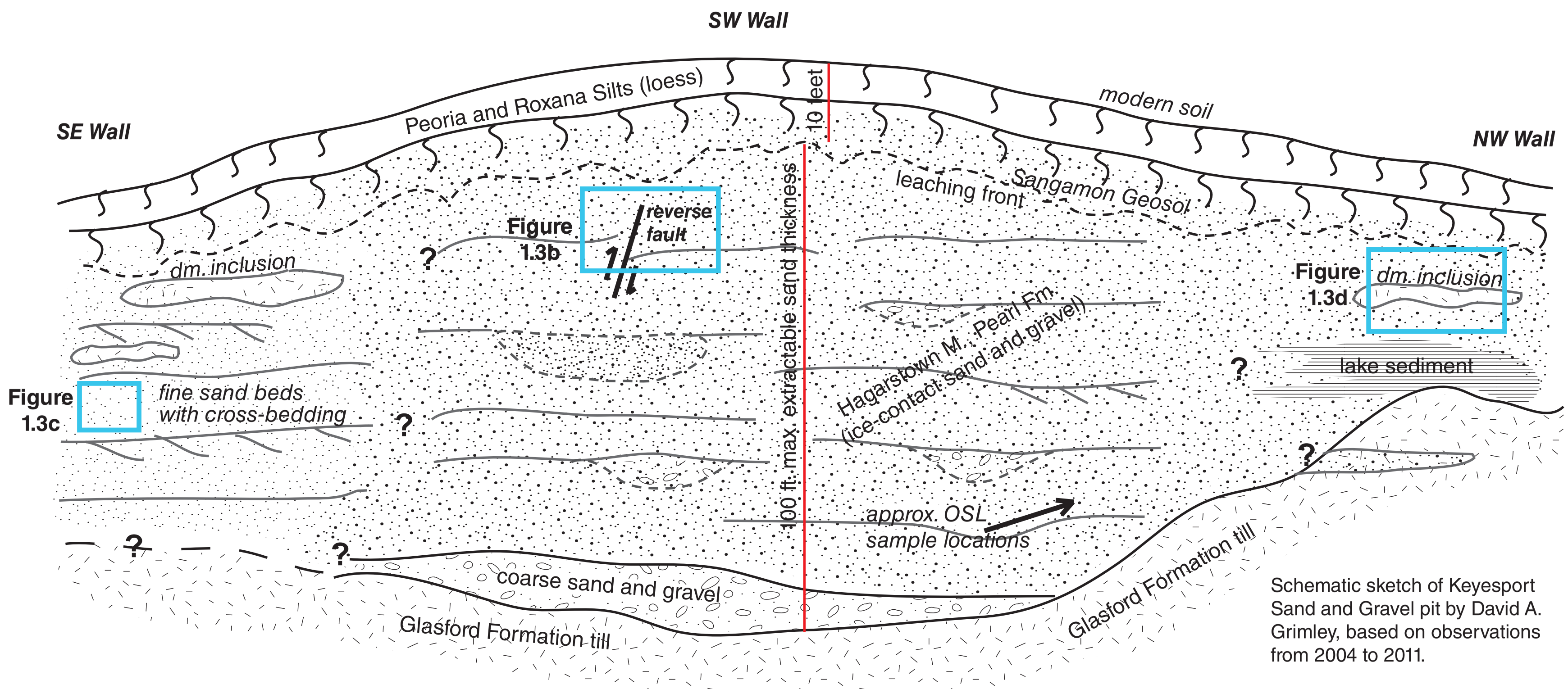


Figure 1.2 Sketch of Quaternary deposits exposed at the Keyesport Sand and Gravel pit. dm, diamicton; OSL, optically stimulated luminescence. Modified from Webb, N.D., D.A. Grimley, A.C. Phillips, and B.W. Fouke, Origin of glacial ridges (OIS 6) in the Kaskaskia Sublobe, southwestern Illinois, USA: Quaternary Research, v. 78, p. 341–352. © 2012, with permission from Elsevier.

Southeast Wall (finer grained sands)

Deposits of the Pearl Formation along the southeastern wall tend to be planar bedded and consist primarily of fine- to medium-grained sand (Figure 1.3c). These deposits, in places, are overlain by fine-grained, stratified silt beds and, together, are altered by a somewhat poorly drained, more gray-colored Sangamon Geosol, covered by about 5 ft (1.5 m) of Wisconsin Episode loess. Because the sand has tended to be finer toward the southeast wall, mining has progressed southwesterly rather than widening to the southeast. Many areas of the southeast wall, formerly well exposed, are now covered, slumped, or unsafe to access along steep slopes, so we will not view these deposits at close range on the field trip. However, the overall character of these deposits can be seen from a distance of several hundred feet as we walk down into the pit. This wall is parallel to the overall meltwater flow direction, so beds dipping generally to the southwest are sometimes visible in cross section.

Other notable features on the southeast wall exposure are a few small detached bodies of diamicton, one of which was noted to be about 3 ft (1 m) thick and 20 ft (6 m) long in 2008. During visits to the pit in previous years, diamicton bodies [3 to 6 ft (1 to 2 m) thick and more than 40 ft (12 m) long] have been observed within the sand on either the southeast or west-northwest walls (Figure 1.3d)—it is possible that one or more might be seen on the field trip. The diamicton occurs within otherwise uniform sand. Because of slumping of the sands, some diamicton bodies could not be traced laterally over a great distance. One diamicton inclusion that was observed contained a number of prominent, irregularly shaped sand lenses up to a couple inches thick. The basal contact of the diamicton with the underlying sand was irregular, showing the sand pushing up into the diamicton body, and a gravel lag was found along the contact in some areas. Sand below the diamicton body exhibited contorted bedding. Some diamicton inclusions, such as the one described, may be debris flows. Other diamicton inclusions may have been incorporated into high-velocity glacial meltwater flows as the main channel, prevented from lateral migration by ice walls incised into the underlying till. Breaking along fracture plane discontinuities, subglacial till bodies may have been carried a short distance by glacial meltwaters.

Northwest Wall (sand against wall of till, sporadic lake sediments)

The northwestern wall has the most variable and complex sequence of deposits in the exposure. In 2004, on our first visit to the site, the operator described a wall of clay on this side of the pit, and our observations confirmed it as a wall of dense Glasford till. In other words, the sand abruptly ended on the northwest side of the mined area. James Koerkenmeier, plant manager, related that the pumping of water from this pit did not affect the water-well level for a house a short distance to the northwest, but did affect those in other directions from the pit. The abrupt contact of sand and gravel with hard, pebbly loam diamicton (Glasford Formation) is visible in Figure 1.3e. In a few areas, the sand appeared to have been injected into fractures in the till, implying high pressure or lateral filling by high-velocity glacial meltwater flows. Below the Glasford Formation, a few feet of calcareous fine sand or silt deposits (Petersburg Silt) with wood fragments was locally exposed; these are interpreted as proglacial lake or deltaic deposits.

In late 2010 and early 2011, as mining proceeded to the southwest, an area of rhythmically bedded lake deposits was observed immediately above thin Glasford Formation till deposits and below about 30 to 40 ft (9 to 12 m) of Pearl Formation sand. Lateral areas have >60 ft (>18 m) of sand and gravel. The lake deposits, adjacent to sand and gravel deposits laterally, are too fine to be economically useful and have been left in a protruding peninsula, with the surrounding deposits already mined away. The lake deposits consist of prominently laminated silt, silty clay loam, and very fine sand and are as much as 5 to 10 ft (1.5 to 3 m) thick. In some places, the lake deposits (a tongue of the Teneriffe Silt) are intercalated with medium to coarse sand beds (Pearl Formation), representing alternating lacustrine and fluvial conditions. The lacustrine environment was likely ice-proximal with a rapid sedimentation rate, as evidenced by the occasional small pebbles, low clay content, prominent strata, calcareousness, and lack of fossil shells or wood. The lake sediments could have been localized in a small area blocked by ice, before further melting of the glacier caused drainage of the lake and connection to the full-fledged meltwater outburst and drainage southwest within the Kaskaskia Basin. It is possible, however, that the lake deposits are remnants of a much larger regional lake within the Kaskaskia Sublobe area that temporarily formed by morainal ice blockage in the lower Kaskaskia Valley and drained upon the breaching of such moraine(s). The elevation of the lake deposits is approximately 430 to 440 ft (130 to 135 m), similar to the elevation of deposits of Teneriffe Silt many miles to the southwest (Phillips 2004b; Grimley and Webb 2009, 2010) and the Petersburg Silt in cores and outcrops, such as at Stop 2 and Stop 3, respectively. It is thus intriguing to consider the possibility of a large-scale moraine-dammed lake, or a series of lakes, temporarily existing in the lower Kaskaskia Lowland.

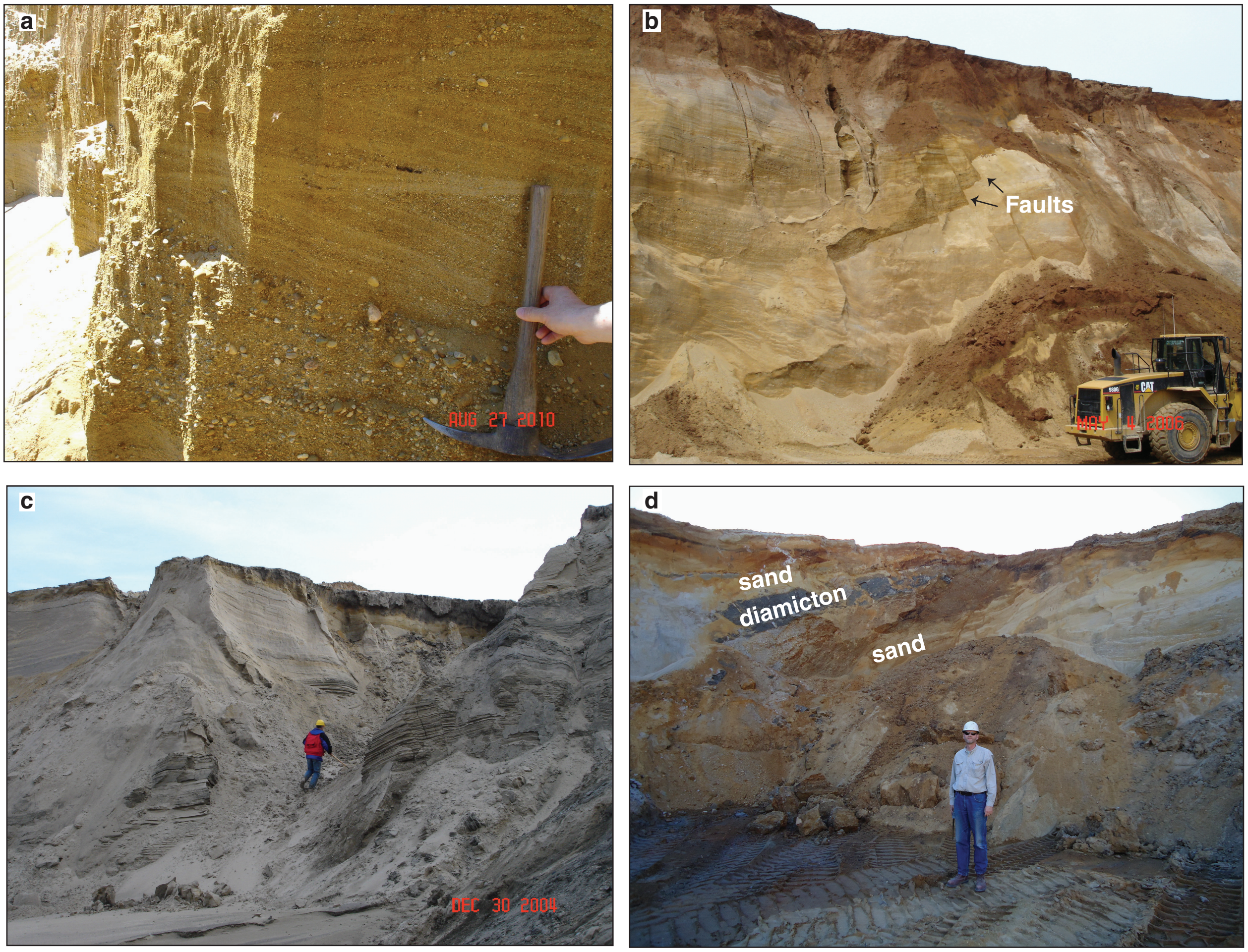


Figure 1.3 Photographs of various exposures at the Keyesport Sand and Gravel pit. (a) Cross-bedded Hagarstown Member sand and gravel on the southwest wall (2010). (b) High-angle reverse faults in Hagarstown sand on the southwest wall (2006). The Sangamon Geosol solum is visible at the top (loess has been stripped). (c) Planar-bedded fine sand (Hagarstown Member) on the southeast wall of the pit (2004). The Sangamon Geosol is visible near the top (loess stripped). (d) Diamicton inclusion (gray) within sand in the west corner of the pit (2010; also viewed on the southeast wall). (e) Wall of Glasford Formation till below and adjacent to Pearl Formation sand on the northwest wall of the pit (2004).

Also observed in 2011 near the lake deposits, but below the Glasford Formation, was a few feet of noncalcareous, stratified, brownish gray silty clay loam, interpreted to represent preglacial or interglacial alluvium. Several large [1 to 3 ft (0.3 to 1 m) wide], angular blocks of fossiliferous limestone were also found scattered about in the area within about 50 ft (15 m) laterally from the exposure of preglacial silt. Although not in place, these limestone blocks appear to be fragments of local Pennsylvanian bedrock, likely incorporated from nearby local bedrock into the Glasford till deposits. The depth to intact bedrock is not known because in-place bedrock is not exposed at the pit.

Luminescence Ages

Two sand samples for optically stimulated luminescence (OSL) age dating were taken from the southwestern wall on October 31, 2007. Sample Keyesport-1 was taken about 55 ft (17 m) below original ground surface and about 15 ft (5 m) above the bottom of the pit. Sample Keyesport-2, some tens of feet lateral to the first sample, was obtained about 45 ft (14 m) below ground surface and about 25 ft (8 m) above the base of the pit. These samples yielded OSL ages of 153,700 ± 19,400 yr [University of Nebraska–Lincoln (UNL)-1873] and 147,100 ± 18,700 yr (UNL-1872), respectively, confirming a correlation of the Illinois Episode deposit with OIS 6 (Table 1.1). The water content of these samples was estimated to have been about 23%, based on complete saturation for most of its history. Natural moisture contents were not used because the sand is now being pumped dry for mining; however, iron staining and coloration indicate a natural water table that was many feet above the sampled location.

Summary/Regional Interpretations

At the Keyesport Sand and Gravel pit, the sediment package of well-sorted fluvial sands interspersed with faulting and diamicton bodies is interpreted as an ice-walled channel deposit, essentially an open-air esker system. Yet initial deposition could have been within a subglacial tunnel followed by ice-ceiling collapse and transition to an open-air channel, which is a typical pathway (Warren and Ashley 1994). The high-angle reverse faults are interpreted to reflect the removal of a supporting lateral ice wall common in ice-contact glaciofluvial sediments (McDonald and Shilts 1975). Constraining of the main channel (more than several hundred feet wide) by ice walls probably led to incision of the channel into the underlying Glasford Formation till deposits and pressurized some sand deposits laterally into fractures in the subglacial till walls. Consequently, some large inclusions of diamicton bodies were eroded out of the lateral till walls or basal till floor and carried into the meltwater channel deposits. The diamicton inclusions were not carried far and were sedimented along with the sand and gravel strata. Other inclusions may represent glacial debris flows.

Optically stimulated luminescence ages of the sand deposits constrain the deposition of the Hagarstown Member, Pearl Formation, between about 170 and 130 ka (Table 1.1), confirming an OIS 6 age for these deposits and for the Illinois Episode glaciation in general (McKay et al. 2008; Hansel and McKay 2010; Curry et al. 2011). The Keyesport Sand and Gravel pit exposures and nearby assortment of landforms point to localized stagnant ice conditions during the melting and deterioration of the Kaskaskia Sublobe and deposition of the Pearl Formation. However, deposition of the underlying Glasford Formation may have occurred during a preceding active ice phase. Although a sublobe per se was not suggested by previous researchers, stagnant ice conditions are a recurring theme in the ridged-drift literature for southwestern Illinois (Leighton 1959; Jacobs and Lineback 1969; Stiff 1996). Overall, the timing of the ice-walled channel deposits here at Keyesport likely represents the deterioration of the Kaskaskia Sublobe, coinciding with the opening of meltwater drainage to the southwest down the Kaskaskia Valley. The Kaskaskia Basin was probably temporarily ponded, perhaps multiple times, upon moraine formation in the lower basin—but the recessional moraines were probably rather quickly overtopped or eroded through by glacial meltwaters from the melting lobe or ice stream. Proglacial lakes, trapped between glacial ice and the moraines, might have been localized or might have been similar in size and duration to Wisconsin Episode proglacial lakes found in central and northeastern Illinois (Lineback 1979; Hansel et al. 2001).

Table 1.1 Equivalent dose, dose rate data, and optically stimulated luminescence (OSL) ages for the two Keyesport Sand and Gravel pit samples[1,2]

Field no.	UNL lab no.	Latitude (N)	Longitude (W)	Depth (m)	U (ppm)	Th (ppm)	K_2O (wt %)	*In situ* H_2O (%)[3]	Dose rate (Gy/ka)	D_e (Gy) ± 1 std. err.	Aliquots (n)[4]	Optical age ± 1σ
Keysport 1	UNL-1872	38.73792	89.39483	16.7	0.5	1.5	1.0	23.0	0.85 ± 0.07	131.08 ± 10.79	24/40	153,700 ± 19,400
Keysport 2	UNL-1873	38.73825	89.39507	13.7	0.6	1.7	1.2	23.0	0.97 ± 0.08	142.17 ± 12.25	21/32	147,100 ± 18,700

[1]Modified from Webb, N.D., D.A. Grimley, A.C. Phillips, and B.W. Fouke, Origin of glacial ridges (OIS 6) in the Kaskaskia Sublobe, southwestern Illinois, USA: Quaternary Research, v. 78, p. 341–352. © 2012, with permission from Elsevier.
[2]The OSL measurements utilized multiple grains of quartz (90 to 150 μm) with the single aliquot method. Analyses were made at the University of Nebraska–Lincoln (UNL) laboratory.
[3]Assumes 30% error in estimate.
[4]Accepted disks/all disks.

STOP 2: Terrapin Ridge: Cores and Geophysics

Andrew C. Phillips and Timothy H. Larson

Terrapin Ridge, containing the Archview Estates subdivision (where we plan to stop for the view), is an impressive hill about 3 mi (5 km) north of the town of Lebanon. The sharp-crested and steep-sided, northeast- to southwest-trending ridge rises up to 100 ft (30 m) above the surrounding plain and is traversed by Illinois Route 4, a major transportation corridor in the region. The Terrapin Ridge area is geomorphically and geologically complex, but fortuitously, we have several high-quality continuous cores from mapping projects, various unpublished data, and several miles of geophysical transects across the area from the past decade (Phillips 2004b; Grimley and Phillips 2011).

As a cultural note, the historic town of Lebanon [3 mi (2 km) south] is perched on another smaller glacial ridge and is home to the oldest private college in Illinois, McKendree University, established in 1828. The town boasts that Charles Dickens lunched here on his way from St. Louis to visit Emerald Mound (Stop 4). In addition, Olympic miler Craig Virgin grew up on Terrapin Ridge.

Geomorphology

Terrapin Ridge occurs along a recessional ice margin just east of where the ridge system crosses the Silver Creek valley (Figures D1, 2.1, and 2.2). The ridge appears to be amalgamated from three or more lithologic and geomorphic elements. The peak of the ridge, at our field trip stop, appears to define the head of a west-facing fan superimposed upon the overall ridge form. The southern end of the ridge has significantly lower relief and is separated by a swale, probably composing a separate element. The longer, linear Becker Ridge intersects the northern tip of Terrapin Ridge at a high angle (Figure 2.1). A shallow basin to the east includes possible ice-walled lake plain and deltaic forms. Small oriented ridges and isolated mounds dot the surrounding landscape, especially to the west.

The Silver Creek watershed is the second largest tributary to the Kaskaskia River. East Fork Silver Creek, interpreted as an ice-marginal stream valley, joins Silver Creek immediately upstream of the area (Figure 2.1) and flows adjacent to the ridge system from its headwaters near Grantfork to west of St. Jacob (Figure D1). Immediately west of Terrapin Ridge, the Silver Creek valley is relatively wide except at the Ogles Creek confluence, where the valley narrows abruptly as it cuts through the ridge system. Shale bedrock crops out in the bottom of Ogles Creek, so bedrock control on narrowing of the valley is possible. The area of restricted valley and high bedrock may also have controlled the location of archeological sites (see discussion at the end of Stop 4). No evidence exists for continuation of a large ice-marginal stream west of Ogles Creek. It is thus possible that such a stream did not exist until the ice had retreated from the highlands to the west and terminated along the east side of the valley.

Low mounds and ridges protrude from the plain west of Terrapin Ridge. Although some appear elongated or oriented, most are irregularly shaped and distributed. On the east side of Terrapin Ridge is a small enclosed basin. Several landforms are compelling, but their origin is not known. Flat-topped landforms with upper elevations just above 490 ft (150 m) asl suggest the basin held a proglacial, ice-marginal lake after the ice margin retreated a short distance eastward from Terrapin Ridge. One rounded, flat-topped mound approximately 500 ft (150 m) across and 10 ft (3 m) high and northeast of Terrapin Ridge (Figure 2.1) resembles ice-walled lake plains recently investigated in northern Illinois (Curry et al. 2010a). Deltaic forms extend from the south flank of Becker Ridge. As deltas or fans, they may have been fed by small streams flowing from decaying ice.

Geologic Succession

Sedimentary and stratigraphic interpretations were made from eight high-resolution borings to bedrock and five shallow percussion probe or hand-auger borings that were acquired along ridge crests and in the surrounding plains (Figure 2.1). These are supplemented by several lithologic logs provided by geotechnical borings at bridges and water-well records. In addition, 4 mi (6 km) of shallow seismic shear wave and 2 mi (3 km) of resistivity profiling reveal some of the sediment architecture. An auger hole to a depth of 27 ft (8 m) by Alan Jacobs (unpublished data, ISGS archives) encountered 10 ft (3 m) of sand below 16 ft (5 m) of loess (Figure 2.3). A borehole by wireline (Virgin Core, Figure 2.3) completed to bedrock just below the hillcrest and only 700 ft (200 m) to the east corroborated Jacobs' findings, but also encountered slickensides within the Glasford Formation and repetition of parts of the Illinois Episode sequence, consistent with glacial deformation. These features may corroborate imbricated sheets of till interpreted from seismic profile data (Figure 2.4). At the base of the Virgin Core are pre-Illinois Episode till, outwash, glacial lake, and preglacial sediment.

The correlation diagram (Figure 2.3) shows the subsurface cross section along the crest of the ridge complex and onto the western flank. Shale and sandstone bedrock rise slightly from the south, from the trough of the ancestral Silver Creek valley. In the

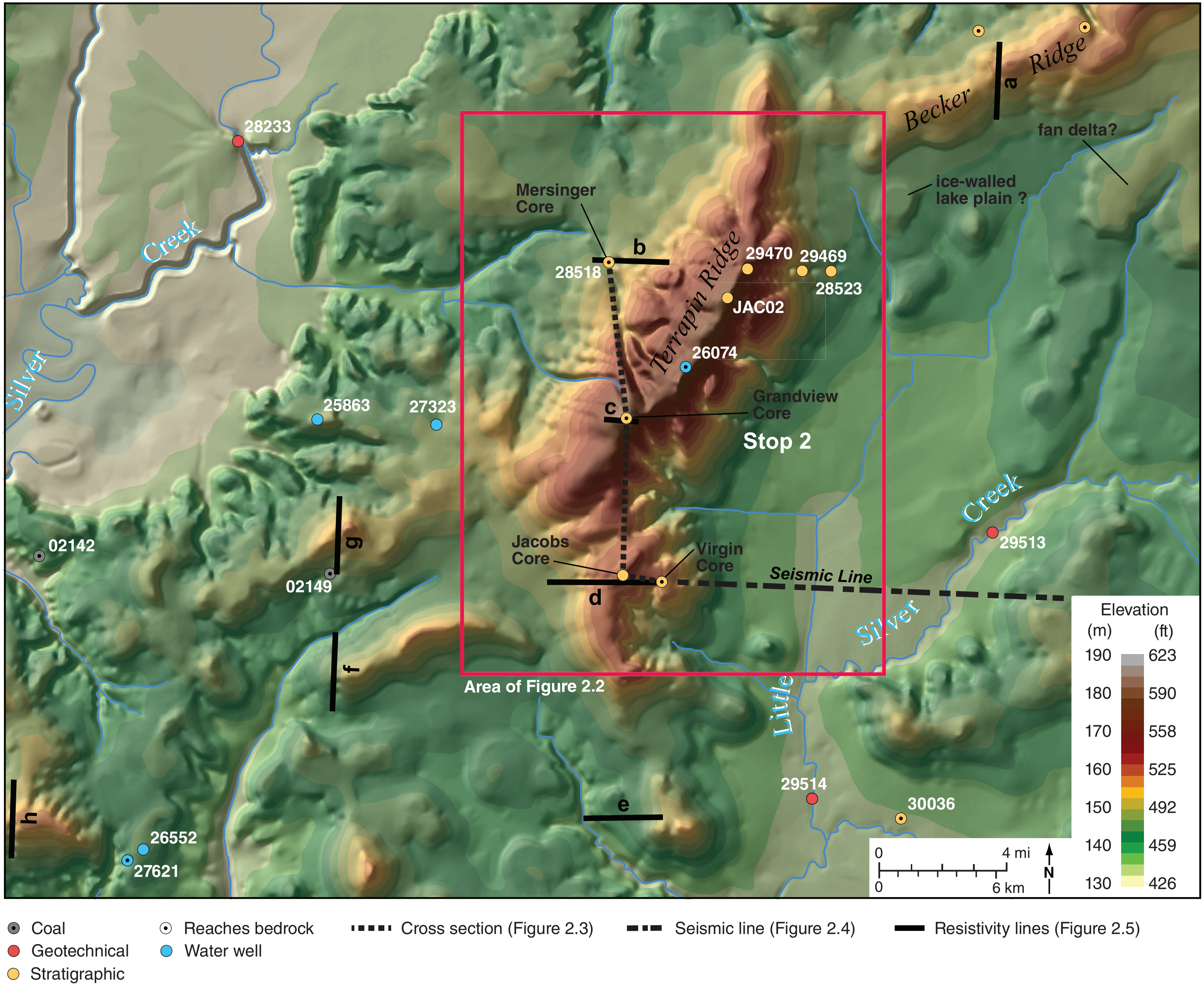


Figure 2.1 Geomorphic map of Terrapin Ridge and vicinity (St. Clair County, Illinois) based on a colorized digital elevation map. The locations of a detailed topographic map (Figure 2.2), a geologic cross section (Figure 2.3), and geophysical transects (Figures 2.4 and 2.5) are shown in context with various important subsurface boring data.

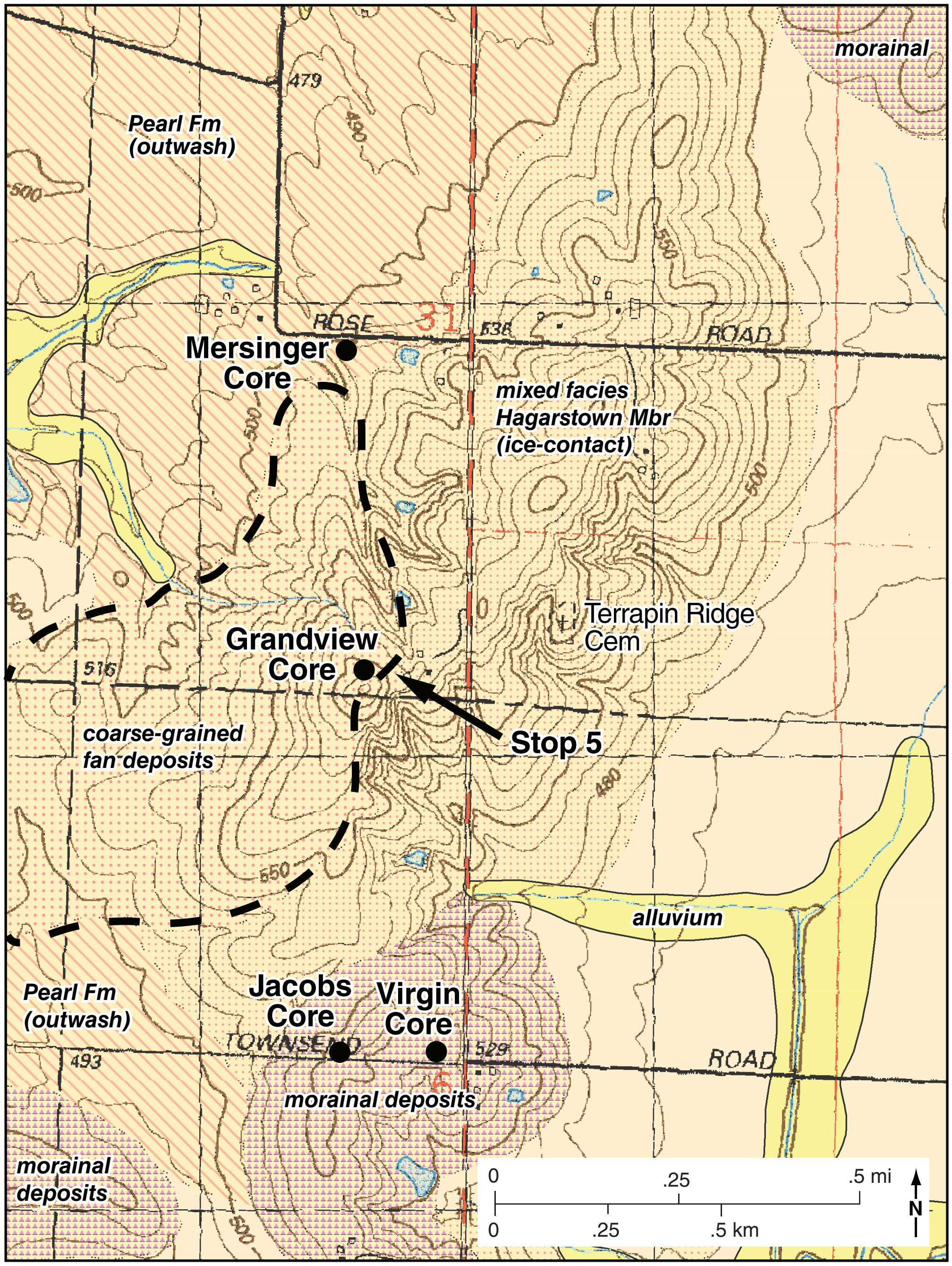


Figure 2.2 Location map for the Grandview/Terrapin Ridge area (Stop 2), with a surficial geologic map (Grimley and Phillips 2011) overlain on a portion of the St. Jacob 7.5-minute quadrangle. Most areas are blanketed by 10 ft or more (≥3 m) of Wisconsin Episode loess except the yellow areas with Holocene alluvium (Cahokia Formation). Hachured areas (near Mersinger Core) are underlain by Pearl Formation outwash facies; areas with dark pink patterns (near Virgin Core) are underlain mainly by diamicton (morainal); and stippled areas over most of Terrapin Ridge are underlain by mixed or sandy facies of the Hagarstown Member, Pearl Formation. Topographic map courtesy of the U.S. Geological Survey.

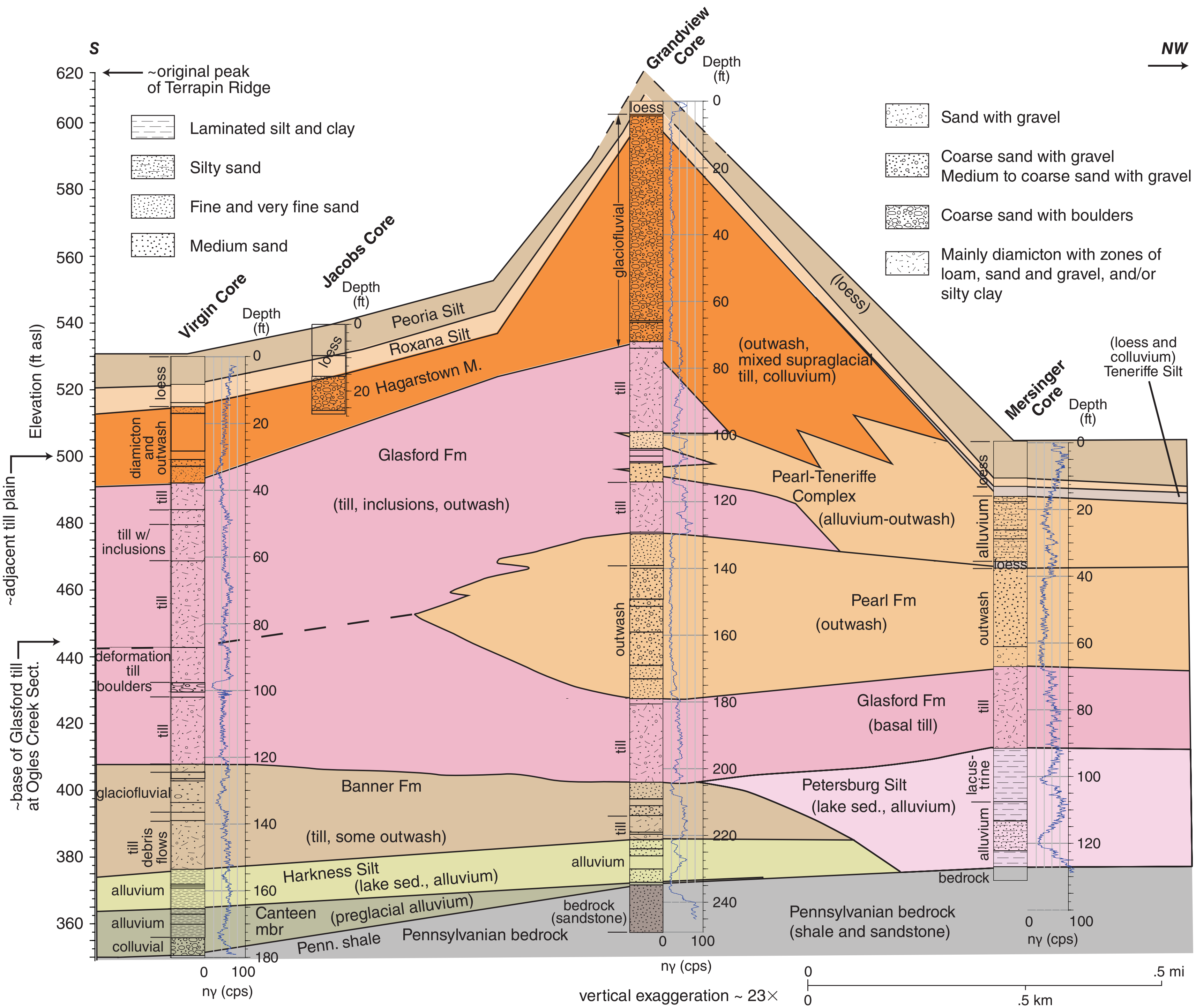


Figure 2.3 Correlation diagram (cross section) of Terrapin Ridge boreholes, based on downhole geophysics and core sediment descriptions. Location of the section shown in Figure 2.1. asl, above sea level; cps, counts per second.

Table 2.1 Fossil gastropods from Illinois Episode lake deposits in the Mersinger Core

Unit, depth	Genus	Species	No.	Comments
Petersburg Silt, 106 ft (32.3 m)	*Pomatiopsis*	sp.	15	Many other immature shells
Petersburg Silt, 106 ft (32.3 m)	*Carychium*	sp.	5	>2 mm
Petersburg Silt, 106 ft (32.3 m)	*Succinea*	sp.	1	Immature

trough, pre-Illinois Episode, preglacial (Canteen member), proglacial (Harkness Silt), and glacial (Omphghent member) sediments are preserved (Virgin and Grandview Cores, Figure 2.3). These deposits were eroded out west of the ridge (Mersinger Core, Figure 2.3) by ancestral Silver Creek, possibly during an ice advance, or else by interglacial fluvial erosion.

Slackwater lake sediment and associated silty alluvium (Petersburg Silt) were deposited during the Illinois Episode ice advance (Mersinger Core, Figure 2.3). Slackwater conditions in Silver Creek valley were possibly caused by ice or sediment blockage of the ancestral Kaskaskia River. The Petersburg Silt unit in Mersinger can be correlated across the valley to the exposure we will see during Stop 3 (Figure D1). Fossil gastropods in one layer within the Petersburg Silt at a depth of 106 ft (32 m) include predominantly *Pomatiopsis* sp. (up to 6 mm long), the much smaller *Carychium* sp. (~2 mm; terrestrial species), and one individual of *Succinea* sp. that is immature (Table 2.1). *Pomatiopsis* is an amphibious genera (with lungs) that typically lives along riverbanks or in moist areas near streams (Burch 1989). *Carychium* is found in moist or very damp areas, such as under forest debris near water (Baker 1939). *Carychium* is a terrestrial genera not common to frequently flooded areas (Burch and Jung 1988), although it could have washed into the floodplain. *Succinea* is also terrestrial, but some species live in wet and marshy places near the edge of streams (i.e., *Succinea retusa*; Baker 1939). The overall paleoenvironmental interpretation based on this assemblage is consistent with a floodplain, the edge of a floodplain, or possibly a marshy shoreline of a lake against a hillside. All three genera here are also found in the Petersburg Silt at the Ogles Creek Section (Stop 3), which allows for a reasonable biostratigraphic correlation.

Above the Petersburg Silt, a deposit of Glasford till marks the advance of the Illinois Episode glacier from the northeast to its fullest extent (near St. Louis). This interface in the Virgin Core (Figure 2.3) is marked by a striated limestone boulder and sharp drop in gamma log intensity, interpreted as a boulder pavement. The erosional character of the till-lake sediment interval to the west, including a basal mixing zone (deformable bed?), is exposed at Ogles Creek (Stop 3).

After its maximum advance, the glacier retreated to immediately east of Terrapin Ridge, although it was likely forming other ridges along the Kaskaskia Sublobe margin at the time (Figure Q5d). Outwash, possibly transported down the then ice-marginal East Fork Silver Creek, was deposited within the Silver Creek valley, including at the western base of Terrapin Ridge. At Ogles Creek (Stop 3), the outwash, with an upper elevation at approximately 450 ft (137 m) asl, can be seen to cut out the till, and it lies directly on slackwater lake sediment.

Ice next readvanced and began to construct the end moraine that constitutes the ridge by shearing and redepositing older sediment as well as by direct till deposition. These processes are evident from mixed borehole lithologies in the Virgin and Grandview core and inclined seismic reflectors on the up-ice side of the ridge (Figure 2.4).

The highest portion of Terrapin Ridge, the site of Stop 2, was formed by construction of an alluvial fan from a supraglacial stream. The topographic expression of the fan is clear in Figure 2.2. The approximately 70 ft (21 m) of rounded, coarse gravel encountered in the Grandview core (Figure 2.3) is the thickest and coarsest glaciofluvial deposit we have found in our investigations of these ridges. High-resistivity (sandy) landforms determined from electrical earth resistivity surveys (Figure 2.5) on the valley plain could be correlated to this fan. The fan interfingered with or overrode relatively fine outwash and alluvium within the valley. This is illustrated in the Mersinger core by interbedded fine sand, laminated silt, and laminated silty clay from a depth of 13 to 38 ft (4 to 12 m). The fine sediment may have originated from the fan source or from smaller streams along the glacier front, as is evident from the high land surface on the west side relative to the east side of Terrapin Ridge.

Rounded, flat-topped terraces and terrace lobes at an elevation of approximately 490 ft (149 m) are compelling to interpret as ice-walled lake or fluvial sediment that was deposited in a lake dammed between Terrapin

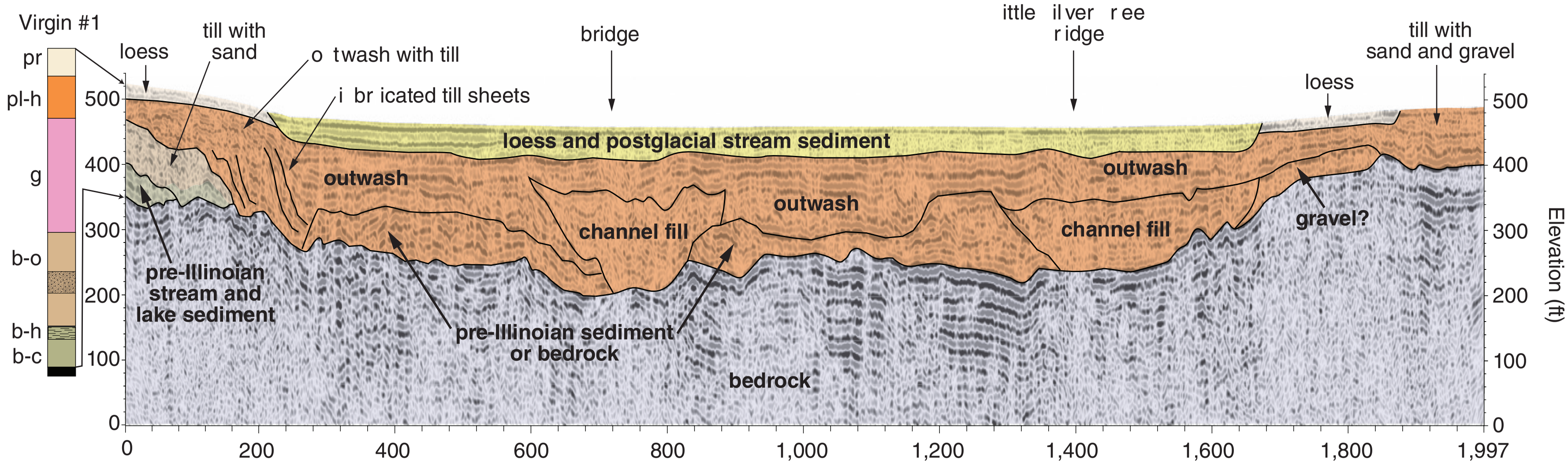


Figure 2.4 Seismic profile from Virgin Core and eastward. Location of the profile shown in Figure 2.1. pr, Peoria and Roxana Silts; pl-h, Hagarstown Member, Pearl Formation; g, Glasford Formation; b-o, Omphghent member, Banner Formation; b-h, Harkness Member, Banner Formation; b-c, Canteen member, Banner Formation. Modified from Phillips (2004b).

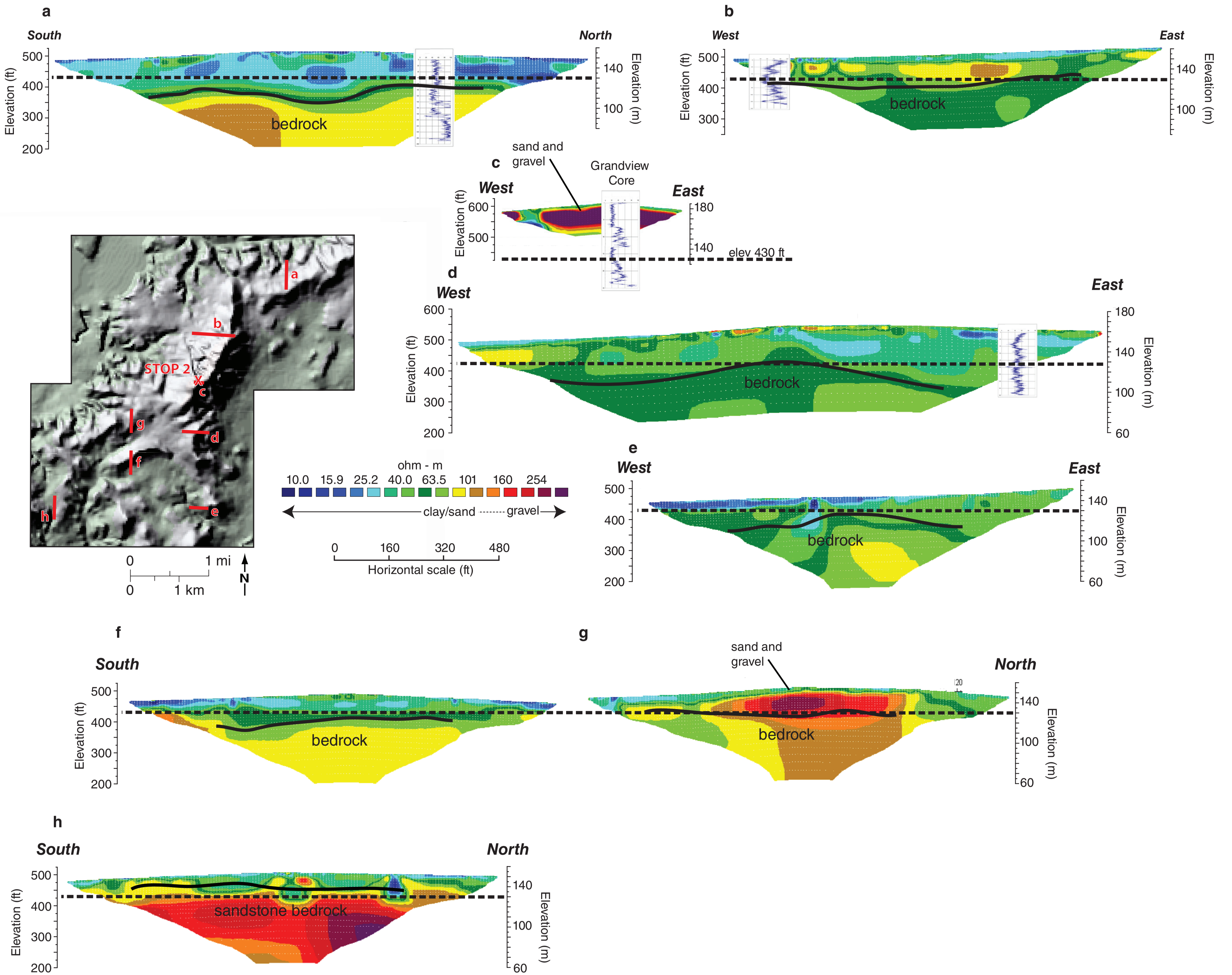


Figure 2.5 Resistivity profiles in the Terrapin Ridge area (locations in Figure 2.1). The dashed line at 430 ft (131 m) is the approximate elevation of nearby Silver Creek. All profiles are shown at the same horizontal, vertical, and resistivity scales. The solid line represents the interpreted contact between unconsolidated Quaternary sediments and bedrock.

Ridge and a recessional ice front or moraine to the east after final retreat of the ice from Terrapin Ridge. Stratigraphic boring #30036, acquired in the Little Silver Creek valley and the interpreted intermorainal basin east of Terrapin Ridge, revealed water-laid sediment throughout the sequence (Figure 2.1) rather than the expected Illinois Episode till beneath last glacial loess. The top of the Illinoian package includes 50 ft (15 m) of bedded loam, silt, and clay, interpreted as glaciofluvial and glaciolacustrine sediment intercalated with debris flow deposits. This could represent the period of buried ice melting and landscape reconfiguration after retreat of the ice front. Underlying the glaciofluvial sediment was 10 ft (3 m) of till (Glasford Formation) over 15 ft (5 m) of fossiliferous lacustrine silt (Petersburg Silt). At the base of the sedimentary sequence, a few feet of soil developed in preglacial alluvium (Canteen member) lay on top of a fluvial conglomerate composed of weakly cemented, imbricated, discoid cobbles (bedrock, probably Pennsylvanian).

Electrical Resistivity Imaging

Approximately 10,500 ft (3,200 m) of continuous resistivity data was acquired on eight profiles in the Terrapin Ridge area (Figure 2.5). These profiles were acquired with a dipole–dipole electrode configuration using an ABEM Terrameter SAS 4000 and LUND acquisition system (Sundbyberg, Sweden). Profiles were processed with RES2DINV, a computer algorithm (Loke and Barker 1996). Four short profiles were aligned transverse to the direction of the ridge at approximately 0.5-mi (0.3-km) intervals along the length of the ridge. One profile (profile c) was acquired at the field trip stop site, just south of the subdivision.

Of the four Terrapin Ridge profiles (b, c, d, and e), only one (profile c) crossing the crest of the ridge at Stop 2 encountered high-resistivity sediments. Profile c also corresponds to the location of the Grandview Core. The gamma log from the Grandview Core (Figures 2.3 and 2.5) correlates the high-resistivity sediments to coarse sand and gravel. The profile here was very short [less than 656 ft (200 m)] and was imaged to a depth of only about 100 ft (30 m), but the resistivity character and presence of sand and gravel are similar to those within a large ridge south of Pleasant Ridge at Stop 5 (Kessler borehole). A similar deposit of high-resistivity material was encountered in one other resistivity profile west-southwest of Terrapin Ridge (profile g, Figure 2.5). Ridge-forming deposits imaged in the other profiles (such as profile d) are relatively low-resistivity sediments, interpreted as likely diamicton or lake sediments based on test drilling (Figure 2.3). In general, the resistivity images suggest that the sand and gravel deposit encountered at the crest of Terrapin Ridge has limited areal extent. Very low resistivity deposits were encountered at the base of the ridges in several of the profiles. These deposits probably correlate to alluvial or lacustrine sediments. Very high resistivity deposits encountered in the western and eastern profiles (such as profile h) at elevations below 400 ft (120 m) probably reflect sandstone or limestone bedrock. High-resistivity bedrock was not imaged beneath Terrapin Ridge. A slight increase in resistivity at some locations may indicate the bedrock.

Topics for Discussion

What different processes or events combined to produce the varied ridge compositions and forms?

- How did the ridges underneath the town of Lebanon (Figure D1) form, and what was their timing relative to the ridges farther north?
- Why is the breach of the ridge system by Silver Creek so narrow?
- Are those really sheared till sheets in the geophysical profiles?

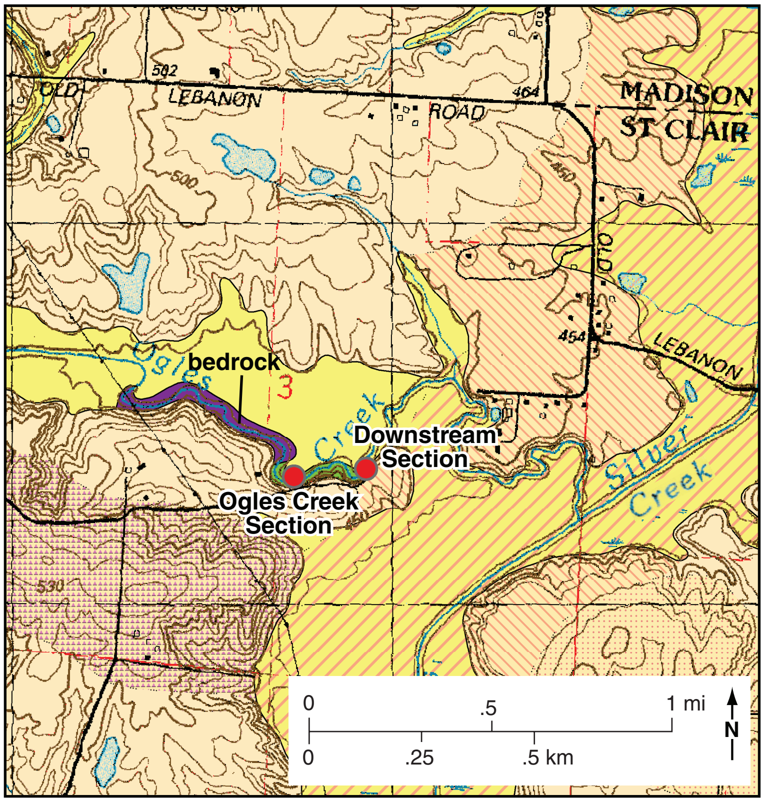


Figure 3.1 Location map for the Ogles Creek Section, St. Clair County, Illinois (Stop 3). Surficial geologic map (Grimley and Phillips 2011) is overlain on a portion of the St. Jacob 7.5-minute quadrangle. Topographic map base courtesy of the U.S. Geological Survey.

STOP 3: Ogles Creek Section

David A. Grimley, Andrew C. Phillips, Samantha W. Kaplan, and Elizabeth C. Geiger

Please use caution at this site—do not walk beside or excavate beneath the undercut or steep slopes; watch for slump blocks.

Introduction

This section is located near O'Fallon, Illinois (St. Clair County), along a cutbank on the south side of Ogles Creek in the NE¼SW¼ Section 3, T 2 N, R 7 W of the St. Jacob 7.5-minute U.S. Geological Survey Quadrangle (Figure 3.1). The Ogles Creek Section was rediscovered in 2004 by Andrew Phillips, David Grimley, and others when exploring the area for exposures of Quaternary deposits as part of the St. Jacob Quadrangle surficial mapping project (Phillips 2004b). Unknown to the mappers at the time, about 8 ft (2 m) of "preglacial silt" (Petersburg Silt in current terminology) had previously been noted by Udden and Shaw (1915) at approximately this location. The quality of the exposure from 2004 through 2010 was ideal for modern study of the geologic units, with fresh exposures cut by an actively eroding stream (Figure 3.2). This outcrop is a key location for understanding the paleoclimate, paleoenvironments, and glacial processes in southern Illinois during the peak of the Illinois Episode glaciation (probably late OIS 6).

Geologic Sequence

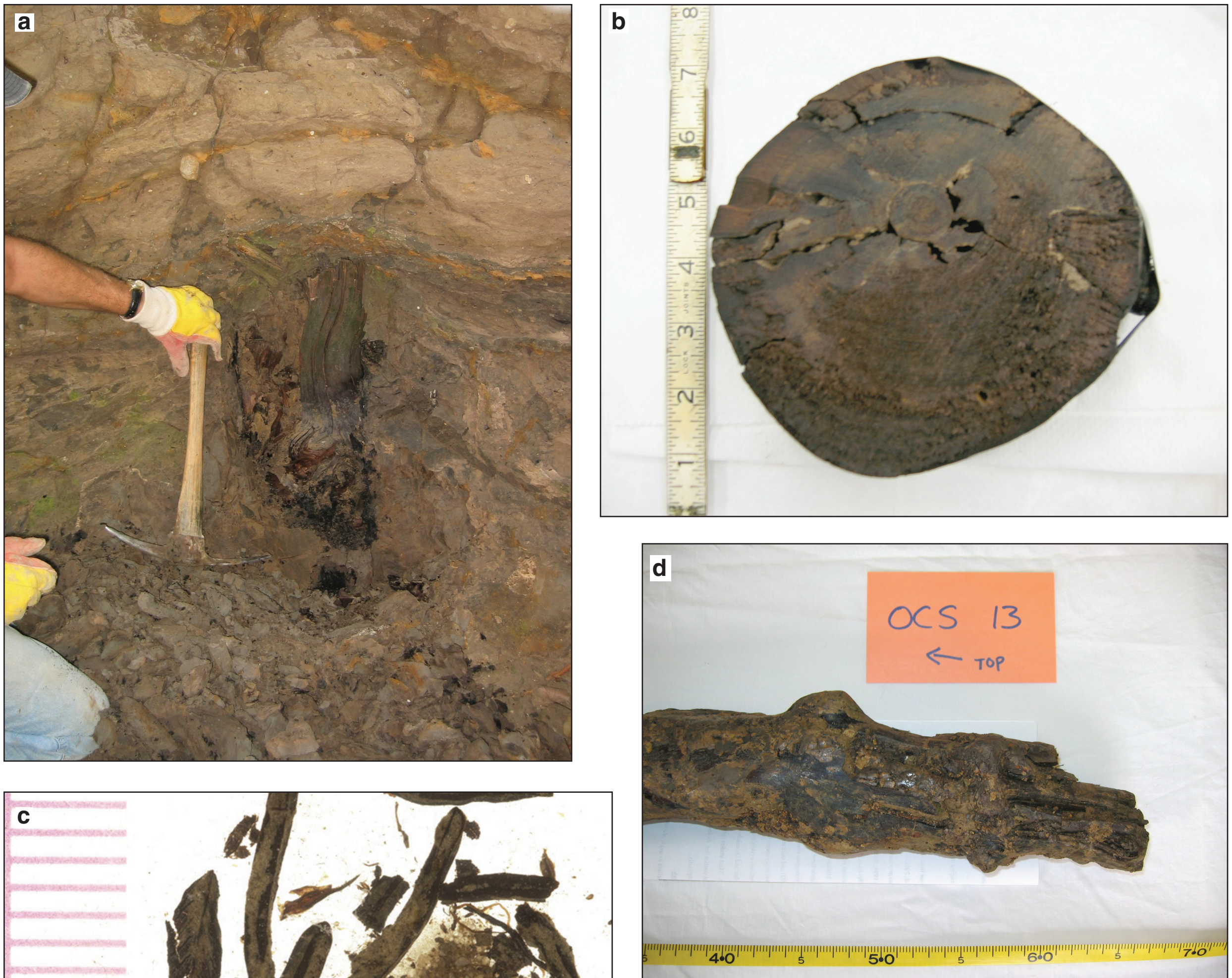
The Ogles Creek Section reveals 17 ft (5 m) of loess (Peoria and Roxana Silts) overlying 22 ft (7 m) of diamicton (Glasford Formation, mainly till) overlying about 6 ft (2 m) of alluvial and lacustrine silts (Petersburg Silt) to creek level (Figure 3.2). Shale bedrock is not exposed at the main section, but it crops out nearby in downstream and upstream directions (Phillips 2004b). Locally, the bedrock topographic surface rises westward from the Silver Creek bedrock valley (Grimley and Phillips 2011). The Glasford Formation, in its upper 5 ft (1.5 m), contains a strongly developed, reddish brown interglacial paleosol (Sangamon Geosol solum) that is clay loam in texture and consists of altered till. Below the paleosol solum is 8 ft (2 m) of olive brown, pebbly loam diamicton (oxidized C horizon) and then 9 ft (3 m) of gray, pebbly loam diamicton (unoxidized D horizon), both interpreted as subglacial till. The oxidized zone contains abundant iron stains on the till fractures, with oxidized zones around gray

Figure 3.2 Ogles Creek Section during the 2004 rediscovery. The hole in the bank is likely a sand and gravel channel near the base of the Glasford Formation. The Petersburg Silt is fossiliferous, with *Picea* stumps and minute gastropods in the lowest 5 ft (1.5 m) of the exposure (not visible in the photo). The logjam in the creek (with modern trees) is from recent flooding.

diamicton interiors near the CD horizon transition. The unoxidized zone consists mainly of massive pebbly loam diamicton but includes isolated sand and gravel lenses up to several feet wide and 3 ft (1 m) thick (possibly R-channels). The basal zone of till also contains some sheared, perhaps formerly frozen, inclusions of underlying lake sediment and fossil *Picea* wood [one horizontal log is 4.5 ft (1.4 m) long and 4 in. (10 cm) in diameter].

The Petersburg Silt, exposed below the Glasford Formation, consists of weakly laminated to massive fossiliferous silt loam. The silt is interpreted as a combination of lake sediment, shoreline deposits, and alluvium that consists mainly of redeposited loess. Within the unit are numerous fossil *Picea* logs [up to 6 in. (20 cm) in diameter], branches, and needles (Figure 3.3) as well as aquatic and terrestrial gastropods (Table 3.1), small pill clams, and sparse ostracodes. Two lithofacies are distinguished within the Petersburg Silt: (1) a relatively intact lower zone of dark grayish brown silt loam with a weak [2-in.-thick (5-cm-thick) A horizon] paleosol in its top, and (2) an upper zone of intermixed silt, diamicton, sheared paleosol stringers. Fossil *Picea* logs were found in both facies (Figure 3.4). The upper sheared facies varies from 0 to 3 ft (0 to 1 m) thick in the zone between the lower *in situ* facies and the overlying Glasford Formation (Figure 3.4). The upper facies appears to have been significantly modified and distorted by the weight and movement of the Illinois Episode glacier as it traversed the area and deposited the Glasford till, and it pinches out laterally. The lower facies served as the soft substrate over which the glacial ice traversed and advanced to the terminal Illinois Episode margin (Figure Q5b).

About 500 ft (152 m) downstream (east) from the main Ogles Creek Section is a smaller 25-ft-high (8-m-high) exposure on the east bank where the river makes a short bend to the north. This section shows Wisconsin Episode loess over a Sangamon paleosol developed in Pearl Formation sand and gravel that in turn overlies several feet of *in situ* Petersburg Silt (which floors the creek locally). The Pearl/Petersburg contact slopes toward Silver Creek valley. Here, it is inferred that the Glasford till was eroded before or during deposition of the Pearl Formation.



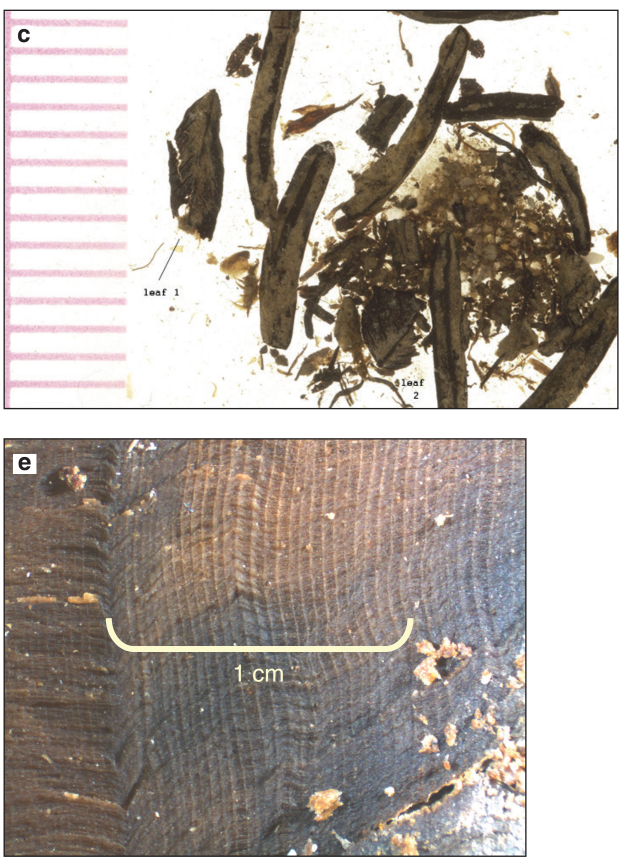


Figure 3.3 *Picea* logs and needles from Petersburg Silt, Ogles Creek Section. (a) *In situ Picea* log in Petersburg Silt, which appears to have been sheared off by the glacial advance that deposited the Glasford Formation. (b) Cross section of a 6-in. (15-cm)-diameter *Picea* log. (c) *Picea* needles washed out from Petersburg Silt. (d) *Picea* log OCS-13. (e) Narrow rings shown in a *Picea* log cross section. Panels b, d, and e from Kaplan and Grimley (2006).

Table 3.1 Molluscan fauna from the Petersburg Silt (Illinois Episode), Ogles Creek (OC) Section

Genus	Species	OC-1	OC-2	OC-3	Total no.	Environment
Carychium	*exile canadense*	5	105	1	111	Terrestrial—wet fens and woodlands
Catinella	*avara*	4	8	0	12	Terrestrial
Columella	*alticola*	0	1	0	1	Terrestrial—cold
Fossaria	cf. *obrussa*	0	4	1	5	Shallow aquatic
Fossaria	*dalli*	8	16	1	25	Shallow aquatic
Gastrocopta	*tappaniana*	0	9	0	9	Terrestrial—wet fens and woodlands
Gastrocopta	*pentodon*	1	0	0	1	Terrestrial
Hawaiia	*minuscula*	2	0	0	2	Terrestrial
Helicodiscus	*singleyanus*	1	0	0	1	Terrestrial
Nesovitrea	*electrina*	7	3	0	10	Terrestrial
Punctum	*minutissimum*	3	3	0	6	Terrestrial
Strobilops	*labyrinthicus*	0	1	0	1	Terrestrial
Succinea	cf. *grosvenori*	0	0	1	1	Terrestrial
Succinea	cf. *ovalis*	0	3	0	3	Terrestrial
Triodopsis	*algonquinensis*	0	1	0	1	Terrestrial
Vallonia	*gracilicosta*	1	0	0	1	Terrestrial—cold
Vertigo	*elatior*	3	44	0	47	Terrestrial—boreal
Vertigo	sp. 1	0	21	0	21	Terrestrial
Total		35	219	4	258	

Below is a description of a stream bank 500 ft (152 m) downstream of the Ogles Creek Section:

0–15 ft [0–4.6 m]:
Peoria and Roxana Silts; tan to reddish brown silt; not studied [loess]

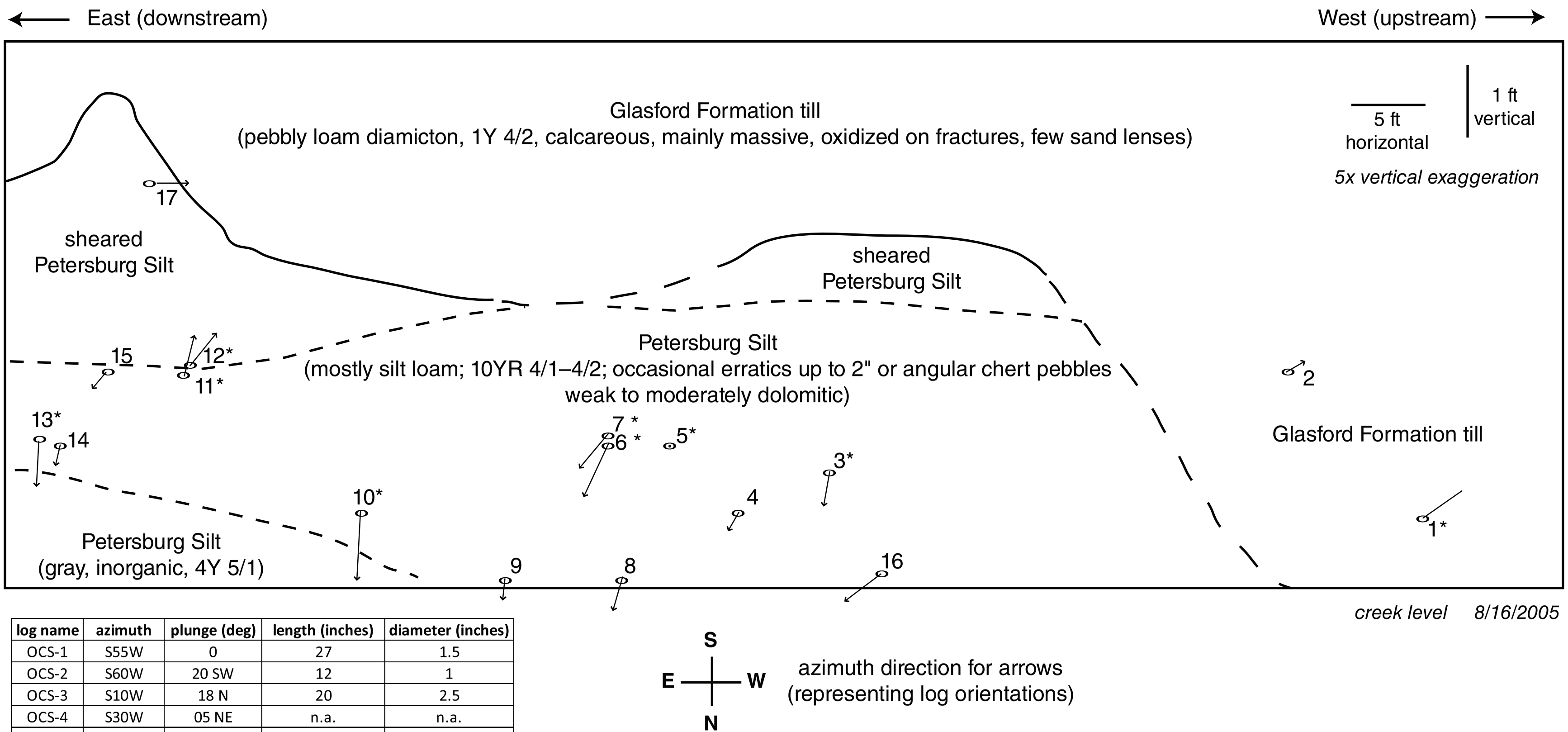
15–18 ft [4.6–5.5 m]:
Sangamon Geosol developed in Pearl Formation; pebbly silty clay loam to silty clay diamicton, deeply weathered to orange brown, clasts ~60% angular, 40% rounded, pebbles to few cobbles; sharp, undulating lower contact [fluvial]

18–25 ft [5.5–7.6 m]:
Petersburg Silt; very dense orange brown (mottled) silt to silt loam, massive, dry, leached, Liesegang banding, Sangamon Geosol B horizon well developed in upper part, forms creek bed [loess and lacustrine silt]

Plant Mesofossils (*Picea* Logs), Macrofossils, and Pollen in Petersburg Silt

Most noteworthy in the Petersburg Silt (at the main section) is the presence of abundant fossil *Picea* (spruce) logs, which occur in both vertical and horizontal positions (Figure 3.3). On the first visit to the site in 2004, one log was observed in its original growth position (*in situ*). In 2005, Samantha Kaplan and David Grimley made a detailed survey and sampling study of the numerous fossil logs (Figure 3.4; Kaplan and Grimley 2006). More than 20 logs were recovered from within the Petersburg Silt (mostly lower facies), with some logs up to 41 in. (1 m) in length and several inches in diameter. The logs in the *in situ* Petersburg Silt were generally oriented north–south, with a slight dip to the north (Figure 3.4), perhaps reflective of a fan deposit along the margin of a floodplain against a hillside. Three additional logs were noted within the overlying till (Glasford Formation), having been incorporated into the glacial ice during advance. The tree rings in the *Picea* fossils are very narrow (Figure 3.3e), especially in later stages of growth, with some having 200 to 400 rings in cross sections. The minute growth rings suggest harsh growing conditions immediately before glacial burial.

Smaller plant mesofossils in sieved fractions (identified by Catherine Yansa, Michigan State University) included *Picea glauca*, *Picea mariana*, *Larix laricina* (?), *Carex* spp., Bryophytae (numerous), Pteridophyta (cf. *Pteridium*), and Ericaceae (cf. *Chamaedaphne*). This assemblage is indicative of shoreline conditions. *Picea glauca* and *P. mariana* were distinguished based



log name	azimuth	plunge (deg)	length (inches)	diameter (inches)
OCS-1	S55W	0	27	1.5
OCS-2	S60W	20 SW	12	1
OCS-3	S10W	18 N	20	2.5
OCS-4	S30W	05 NE	n.a.	n.a.
OCS-5	vertical	90	6+	2
OCS-6	S25W	19 NE	34+	3
OCS-7	S40W	15 NE	26+	2.75
OCS-8	S16W	18 N	19+	3.5
OCS-9	S06W	25 N	12+	3
OCS-10	S03W	18 N	41	2.5
OCS-11	S14W	03 S	23	3.5
OCS-12	S39W	14 S	23.5	6
OCS-13	S03W	16 N	26+	3.5
OCS-14	S13W	14 N	n.a.	1.5
OCS-15	S40W	14 N	n.a.	1.5
OCS-16	S53W	05 N	26+	2
OCS-17	east	29 W	17+	2.5

Figure 3.4 Sketch of the Ogles Creek Section in August 2005 with exposed fossil *Picea* (spruce) log orientations in the Petersburg Silt and lower Glasford Formation (studied and sampled by David Grimley and Samantha Kaplan). Fossil logs or branches that were measured for azimuth and plunge are shown, with the arrowhead indicating the plunge direction. Line lengths are proportional to the length of exposed wood where known (many lengths are minimums because the wood extended into the bank). Log 5 was in a vertical position (no azimuth). An asterisk indicates sampling by Samantha Kaplan for a dendrochronology study. The OCS number is labeled next to each log on the sketch. Length and diameter were measured in the field, prior to drying.

on needle size. The pollen record sampled in intact Petersburg Silt was analyzed by Samantha Kaplan in 2- to 4-in. (5- to 10-cm) vertical increments (Figure 3.5). Spruce (*Picea*), sedges, and mosses dominate both the pollen assemblage and the macrofossil component. *Pinus*, although present in lower portions of the silt, is absent from the profile immediately below the till, whereas *Picea* is still abundant.

Molluscan Fauna in Petersburg Silt

In addition to plant macrofossils, the Petersburg Silt contains small aquatic and terrestrial gastropods and some pill clams. They are all <0.4 in. (<1 cm) in size, the larger of which are visible in the field upon close inspection. Geiger (2008) collected three bulk samples [OC-1 (lowest), OC-2, and OC-3 (highest)] from the exposure for examination of the molluscan fauna (Table 3.1). The most abundant species were *Carychium exile canadense*, *Fossaria dalli*, and *Vertigo elatior*. The modern distribution of *V. elatior* is from the southern shores of Hudson Bay to the Great Lakes region (Nekola and Coles 2010), and they live in such areas as fens, conifer swamps, and other wet areas. *Carychium exile* is another terrestrial snail that lives in similar wetland-type environments as *V. elatior*. The aquatic snail *F. dalli* occurs in the northern United States and southern Canada and lives in wet marshy areas, shallow lakes, and ponds (Clarke 1981; Burch 1989). The presence of one individual of *Columella alticola*, whose modern distribution is northward in Manitoba, Ontario, and Quebec, hints at the possibility of a more northern climate (Nekola and Coles 2010).

Ostracodes in Petersburg Silt

Although uncommon in sediment washings, a few species of ostracodes were noted in some samples of the Petersburg Silt. Species (identified by B. Brandon Curry) include *Candona inopinata*, *Candona candida*, *Cyclocypris sharpei*, *Cypridopsis vidua*, and *Candona* juveniles. These fauna suggest that a shallow freshwater lake existed, at least temporarily. The overlapping distributions of these species point to a modern analogue in the northern Great Lakes region (B. Brandon Curry, personal communication, 2011). On the basis of the ostracode occurrences, along with other paleoecological records, the site probably fluctuated between a floodplain, alluvial fan, and shallow lake environment.

Overall Interpretations

According to regional mapping (Grimley and Phillips 2011), the Petersburg Silt at the Ogles Creek Section probably represents the shoreline (littoral zone) of a large slackwater lake in the lower Kaskaskia Basin. As the Mississippi River base elevation rose in response to the influx of glacial sediment, the Kaskaskia Valley and its tributaries were dammed and back-flooded up to elevations of about 440 ft (134 m) asl (approximately the top of the Petersburg Silt here) in this drainage basin. The lake levels fluctuated seasonally, and terrestrial material (loess, wood, terrestrial gastropods, and other organics) periodically washed into the ancient lake from the surrounding uplands. Advancing glaciers during the Illinois Episode buried and overrode the sediments, in part explaining why the upper contact of the Petersburg Silt has a few feet of relief and its upper facies is sheared and mixed. Because this site is only about 20 mi (32 km) within the terminal margin of Illinois Episode glaciation (Figure Q1), the fossils are perhaps reflective of the glacial climate and environment during the full glacial conditions of OIS 6.

Altogether, the pollen, plant macrofossils, plant mesofossils, and mollusks are consistent with a boreal condition, with the nearest modern analogue perhaps in western Ontario (north of Lake Superior), based on the plant and molluscan fauna. The climate and vegetation envisioned in the immediate vicinity of the Illinois Episode ice front appear to have been fairly similar to full glacial Wisconsin Episode conditions in southern Illinois (Curry and Delorme 2003). Future work may involve finding more specific modern analogues for the assemblages in the hopes of better quantifying climatic conditions in the central United States during the glacial maximum of OIS 6. Further study of the tree rings might provide insights into short-term climate change, seasonality, and periodicities during the few hundreds of years immediately before glacial advance.

Topics for Discussion

- What clues can the sheared and deformed layer in the upper Petersburg Silt tell us about glacial dynamics during the Illinois Episode?
- Is this site representative of the glacial climate and vegetation in southern Illinois during the peak glaciation of OIS 6?

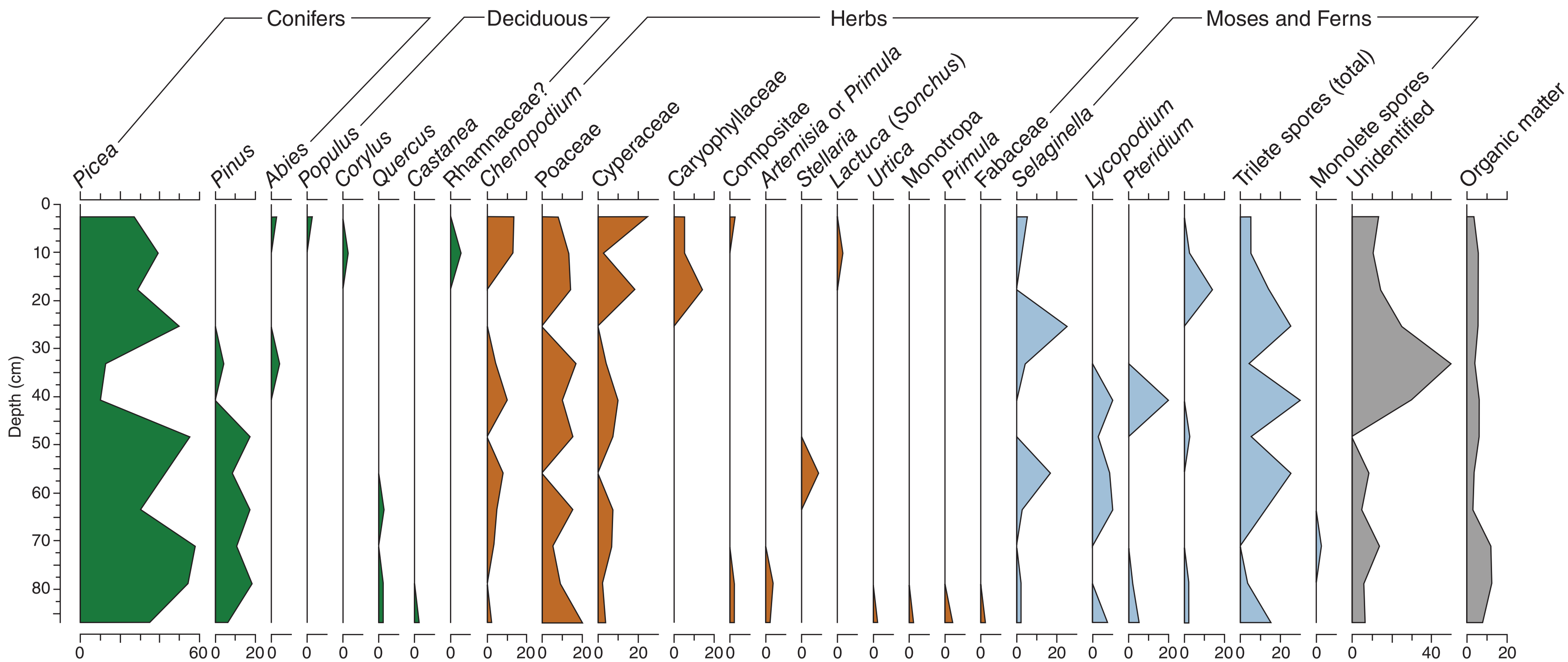


Figure 3.5 Pollen diagram of material from *in situ* Petersburg Silt (Illinois Episode) at the Ogles Creek Section (analyst, Samantha Kaplan). The zero depth indicates the top of *in situ* Petersburg Silt (below sheared Petersburg).

STOP 4: Emerald Mound: Archaeology and History

Brad H. Koldehoff

Background and Geologic Setting

The Illinois Episode till plain in southwestern Illinois is dotted with many isolated knobs and broad hills (Phillips 2004b). The Emerald archeological site (Figure D1) is located atop one of these knobs (Koldehoff et al. 1993). This ancient Native American settlement contains several earthworks, the largest of which is Emerald Mound. The Emerald site was a mound and town complex constructed by people of the Mississippian culture. Named for the middle Mississippi Valley, the Mississippian culture developed and disappeared (A.D. 1050–1350) long before Columbus sailed for the New World in 1492 (Pauketat 2004). Earthen monuments, such as those at the Emerald site and the world-famous Cahokia site [located 15 mi (24 km) to the west in the Mississippi Valley], are the most visible vestiges of this ancient civilization. Researchers believe that the ancient inhabitants of the Emerald site selected this elevated glacial knob for settlement because of its commanding view and its proximity to an ancient overland trail that linked the Mississippi Valley with the Wabash Valley (Koldehoff et al. 1993; Pauketat 1998, 2004). From a glacial geology perspective, the site occurs within an irregular, possibly intermorainic, plain (Figure D1). Emerald Mound is composed of remolded loess deposits (Peoria Silt) that were probably mined locally, based on a 1960s borrow pit cut into its eastern side.

Archeological and Recent History

Emerald Mound is a flat-topped pyramidal platform, built from thousands of basket loads of silty clay loam sediment that was topped with a temple or elite residence constructed from wooden posts covered with mats, thatch, or both. Earthen pyramids, such as Emerald Mound and Monks Mound at Cahokia, are characteristic of Mississippian culture. Recent archaeological excavations conducted by researchers from the University of Illinois at Urbana-Champaign in the modern agricultural fields that surround Emerald Mound have documented habitation features—house floors and cooking or storage pits filled with domestic refuse, such as fragments of pottery vessels, stone tools, animal bones, and charcoal. These durable residues of daily life, along with the mounds, are all that remain of the Mississippian culture. Because they left behind no written records, in the 19th century their remains were attributed to the mysterious Mound Builders. Today, however, it is well established that Mississippian culture is an ancient Native American agriculture-based civilization with far-flung trade networks centered at the massive Cahokia site, which is in large part preserved as the Cahokia Mounds State Historic Site. Located across the Mississippi River from St. Louis, the Cahokia site once held more than 100 earthen monuments. It is the earliest and largest center of Mississippian culture and one of the largest and most complex sites in North America. As such, it is one of only a handful of archaeological sites in the United States that is listed on UNESCO's World Heritage List.

When first reported, Emerald Mound's flat-topped pyramid (with a lower terrace) stood about 50 ft (15 m) tall and measured 400 by 250 ft (120 by 80 m; Snyder 1877, 1909). This structure was the centerpiece of the Emerald site, a Mississippian settlement with four smaller mounds and associated village areas that encompass several hectares of the surrounding area (Figure 4.1). The two conical, flat-topped mounds on the east side of the site (Mounds 1 and 2) were removed in the 1960s. Such mounds are rarely found outside the Cahokia site. The two mounds on the west side (Mounds 3 and 4) were oval in shape when first reported but may originally have been flat-topped pyramids. Their original size and shape have been greatly altered by much plowing and erosion. Immediately north of these two mounds are three possible mounds that have also been significantly lowered by plowing and erosion. The broad, level area between these five mounds and Emerald Mound on the northeast may have been the location of a community plaza. No confirmed borrow areas have been identified, but the northern slope of the site, which borders a small stream and spring that supplied the settlement with fresh water, is suspected to be the location where silty clay loam sediments (Peoria Silt) were excavated to construct the mounds. Although controlled excavations into the mounds have been limited, evidence of typical Mississippian basket-load construction has been identified, and Emerald Mound was built in at least four major stages (Timothy Pauketat, personal communication, 2010).

Obscured by trees and brush, Emerald Mound derives its name from the green prairie grasses that covered the mound when the area was first settled by European Americans. The Emerald site is the easternmost mound center directly affiliated with the sprawling Mississippian mound center of Cahokia (Koldehoff et al. 1993; Pauketat 1998, 2004). Its location in the uplands away from a major stream valley and adjacent to rolling tall-grass prairie is atypical for Mississippian mound centers, which are typically situated in or along river valleys. Researchers have traditionally considered Mississippian economies to have been primarily based

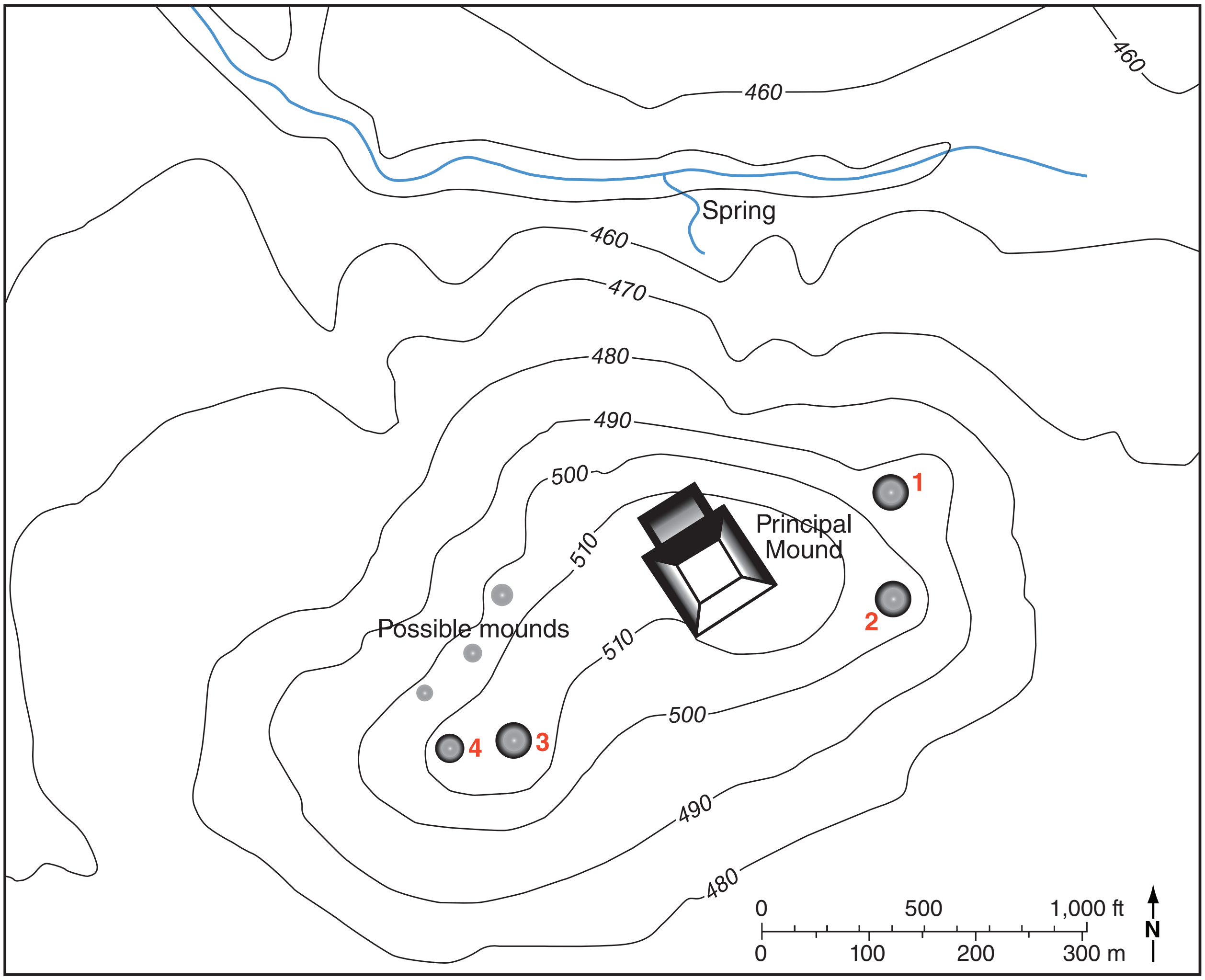


Figure 4.1 Emerald site map (from Koldehoff et al. 1993). Used by permission of the Illinois Archaeology Survey.

on exploitation of the fertile and renewable resources found in alluvial valleys, such as fish, waterfowl, and soils suitable for hoe agriculture. Common crops were maize, squash, tobacco, sunflowers, and other native cultigens.

Investigations at the Emerald site and at other upland settlements have demonstrated that upland populations, perhaps numbering in the thousands, made a living by exploiting local soils and resources (Koldehoff et al. 1993; Pauketat 1998, 2004). Although only limited excavations have been conducted at the Emerald site, they have nonetheless routinely yielded evidence of hoe agriculture, as have excavations at other upland settlements. Thus, it has been suggested that upland prairies, in addition to alluvial soils found along upland streams, may have been farmed by local Mississippian populations (Pauketat 1998). Stone hoe-blades, chipped from imported Mill Creek chert (from Alexander County, Illinois), and the flakes produced from resharpening their worn and soil-polished bits are especially common at the Emerald site. Although these tools were likely used in mound construction, their ubiquity at the site is characteristic of agricultural pursuits. For example, several caches (hoards) of hoe blades have been discovered. The earliest reported cache is also the largest and was found by accident. The following information is reported by John F. Snyder (1909, p. 76), pioneer archaeologist in Illinois:

> In 1840 Mr. Baldwin, then proprietor of the premises, built a dwelling house that encroached several feet upon the large square mound near its eastern corner. In excavating the cellar and foundations . . . he unearthed, from about a foot beneath the mound edge, sixteen large flint spades, from ten to eighteen inches in length, smoothly polished at their broad ends by continued use. . . .

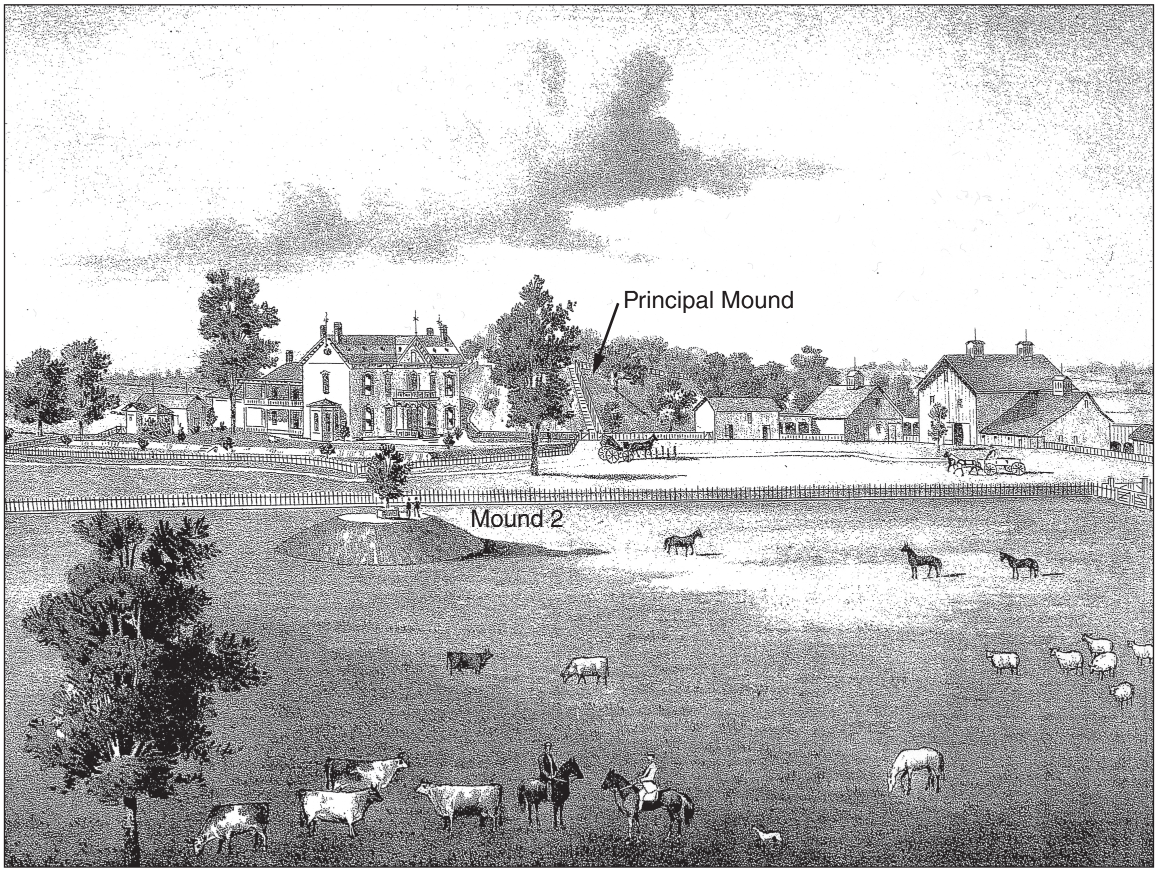


Figure 4.2 Drawing of Mr. Henry Seiter's "Mound Farm" (from Brink, McDonough, and Company 1881). Mound 2 is in the foreground.

The Baldwin house mentioned by Snyder is depicted in the 1881 history of St. Clair County as part of Mr. Henry Seiter's "Mound Farm" (Brink, McDonough, and Company 1881). Emerald Mound is in the background, and Mound 2 is in the foreground (Figure 4.2). An aerial photograph taken in February of 1965, some 80 years later, shows Mr. Seiter's house still standing on the east side of Emerald Mound (Figure 4.3). Also visible is a T-shaped depression atop Emerald Mound. This exploratory trench was excavated in 1964 by Robert Hall, then Illinois State Museum archaeologist. Hall's test discovered a post pattern (foundation) of a large building (probably a temple or elite residence) that once stood atop the mound (Koldehoff et al. 1993).

Hall's work at the site, as well as investigations conducted several years earlier by other State Museum archaeologists, were undertaken because John Hoak, the owner of the Mound Farm in the 1960s, considered the mounds to be a nuisance. Consequently, he began removing them—selling their ancient basket loads of sediment by the truckload for fill dirt. After Mounds 1 and 2 were removed, he began cutting into Emerald Mound itself just north of the old Henry Seiter house (Figure 4.3). It was at this time, with much local support, that the State of Illinois purchased Emerald Mound to ensure its protection. The rest of the site remains in private ownership, but this and other sites in the state are accorded some protection by state and federal laws.

Despite erosion and soil borrowing, the Emerald site is one of the best preserved Mississippian mound and town complexes in Illinois. It is the largest upland mound center in the region and is the easternmost mound center within the sphere of greater Cahokia. Its location on a high upland knob is atypical for a major Mississippian site and likely stems in part from its position along an ancient east–west transportation corridor (Koldehoff et al. 1993). Today, U.S. Highway 50, which runs through the nearby town of Lebanon, essentially retraces the route of the historic St. Louis-to-Vincennes wagon road, which in turn likely followed a footpath blazed by ancient Native Americans. This overland corridor is the shortest route between the Wabash Valley

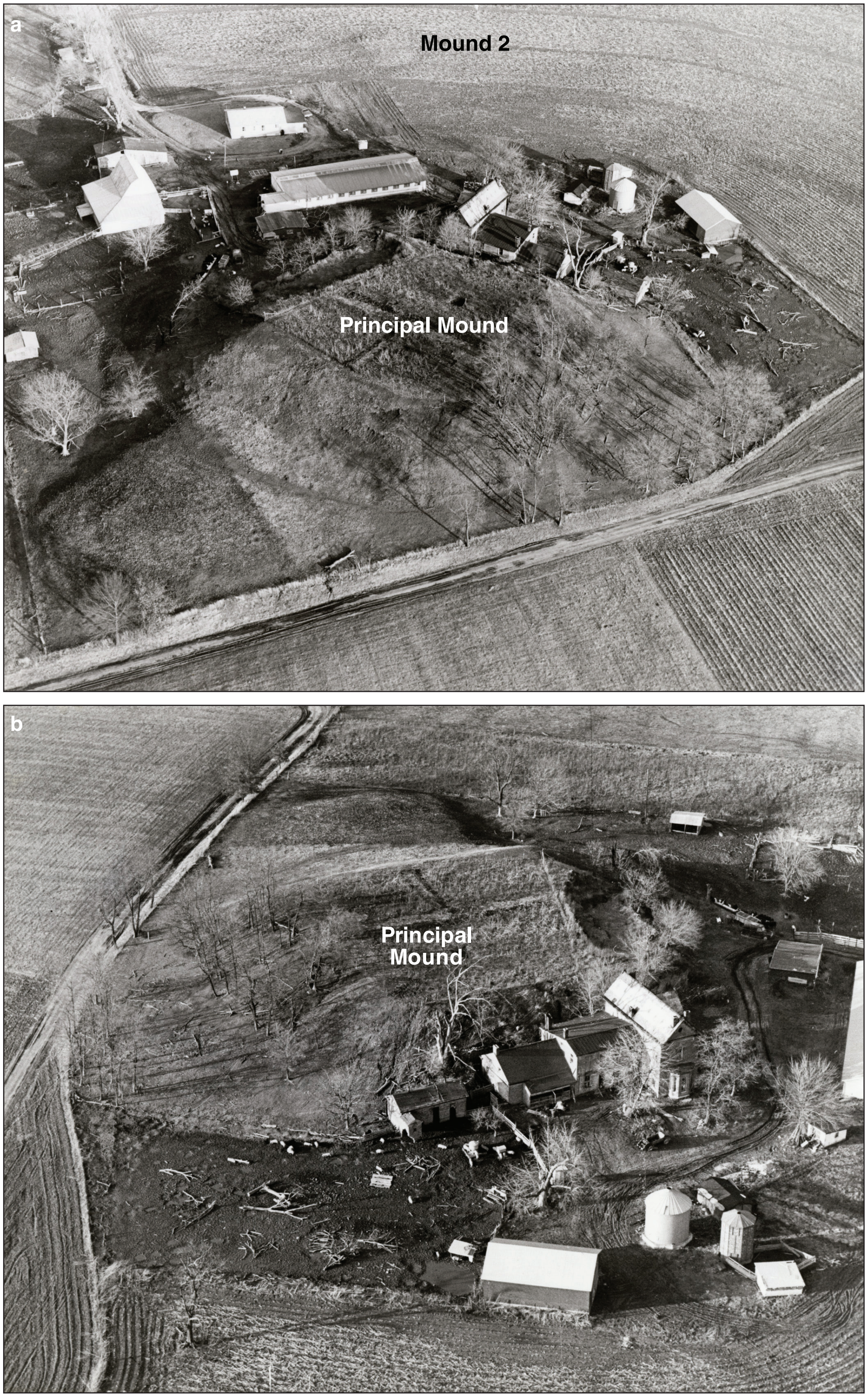


Figure 4.3 February 1965 aerial photographs of Emerald Mound (from Koldehoff et al. 1993). (a) View to the east, with Principal Mound in the middle and Mound 2 visible as a small knob in the upper photo. (b) View to the north. Photographs courtesy of Richard Norrish. Used by permission of the Illinois Archaeology Survey.

and the Mississippi Valley. The prominent location of Emerald Mound likely provided a commanding view of the ancient footpath, as well as of nearby settlements.

Nearby Archeological Sites

The Copper site is a nearby, and much smaller, Mississippian mound center. Unlike the Emerald site, which is located about 4 mi (6 km) away on a high knob, the Copper site is located on a ridge along the east bank of Silver Creek at a spot where the otherwise broad and swampy floodplain of the creek constricts, thus affording a good location for a stream crossing. The idea that this general locale was where an ancient trail crossed Silver Creek is supported by the location of other nearby archaeological sites. For example, located across Silver Creek, near the mouth of Ogles Creek, is the Bostrom site, a late Pleistocene Clovis culture campsite. This site was investigated before the expansion of a residential development that ultimately destroyed the site (Tankersley et al. 1993; Koldehoff and Walthall 2004). Its position high above the creek valley would have been an ideal location for monitoring the movement of herd animals, especially if they were crossing the creek at this narrow spot in the valley (Figure 3.1).

STOP 5: Pleasant Ridge Area: Cores and Geophysics

Timothy H. Larson and David A. Grimley

Introduction

The Pleasant Ridge area, in St. Clair County 3 to 5 mi (5 to 8 km) south of Mascoutah, Illinois, is part of a series of glacially constructed Illinois Episode hills and ridges in southwest Illinois (Figures D1 and 5.1). Pleasant Ridge is a relatively narrow, steep, and somewhat elongate hill just north of a complex of several broader ridges that are about 80 to 90 ft (24 to 27 m) above the surrounding lake plains (that were inundated by slack-water lakes during the Wisconsin Episode). Some of the ridges in the vicinity of Pleasant Ridge are composed predominantly of fine-grained diamicton and appear to be marginal moraines, whereas others are composed predominantly of sand and gravel, interpreted as glaciofluvial. Distinguishing the types of ridge deposits is important geologically in interpreting their origin and economically because the coarse-grained deposits are valuable as construction aggregate. The ridge-forming deposits are capped with 10 to 13 ft (3 to 4 m) of silty loess and underlain by thin pre-Illinois Episode deposits or Paleozoic bedrock.

The research presented at this stop integrates extensive geophysical studies (electrical resistivity) with borehole studies taken as part of a 1:24,000-scale surficial geologic mapping project (Grimley 2010). Both shallow and deep (to bedrock) stratigraphic test holes and lithologic descriptions of water-well records were used to help calibrate the geophysical data.

Geomorphology, Geologic History, and Sediment Cores

The Pleasant Ridge area consists of curvilinear hills and knolls, mostly having an east–northeast to west–southwest orientation. Lithologically variable deposits in the ridges are classified on the surficial map (Figure 5.1) as either the mixed (fine–coarse) or sandy facies of the Hagarstown Member, Pearl Formation. However, the subsurface geology is more complex than can be shown on the 1:24,000-scale surficial map or cross sections.

Deposits within the main portion of Pleasant Ridge (Sections 19 and 20, T 1 S, R 6 W; center of Section 24, T 1 S, R 7 W) include interbedded sand, loam, and variably textured diamicton, as well as ice-thrusted inclusions of pre-Illinois Episode paleosol fragments (Yarmouth Geosol) or other older materials. The hills south of the main ridge include up to 110 ft (34 m) of various grades of well-sorted to poorly sorted sand, including gravelly zones and zones of very fine sand. These deposits, shown in the cross section in Figure 5.1, are classified as the sandy facies, Hagarstown Member. Sandy areas (with loess cover) are shown with large stipples in Figure 5.1 and tend to occur in hills on the southeast side of the main ridge system, based on test holes, water-well records, and geophysical studies. In the circular-shaped lowland between the ridges, clayey lacustrine sediments (silty clay loam mainly) were found in a shallow boring to 37 ft (11 m; MSC-11: county #30361). This shallow boring encountered mainly last glacial lake sediment (Equality Formation), with a radiocarbon age on peaty/organic material at a depth of 34.5 ft (10.5 m) determined as 40,500 ± 1,200 ^{14}C yr BP (radiocarbon years before present; ISGS-6001). The gray, silty material in the few feet above the peat (30.5 ft depth; 9.3 m depth) was calcareous, had some thin beds of very fine sand, and contained the ostracodes *Candona ohioensis* (juvenile), *Cypretta brevispina*, and *Pelocypris tuberculatum*.

The origin of Pleasant Ridge may be related to the presence of a localized high area on the buried bedrock surface underlying the ridge system (Grimley 2010). The bedrock high or obstacle may have restricted the flow of glacial ice, leading to a natural position of a glacial margin for the envisioned Kaskaskia Sublobe (Figures Q1 and Q3). A stationary or locally fluctuating glacial margin in the area for a period of time could explain the complex association of deposits (lacustrine, fluvial, subglacial) in the Culli boring (Figure 5.2) and, consequently, the formation of a moraine and related deposits. The large sand-rich hills on the southern side of Pleasant Ridge may be related to subglacial channels or to ice-marginal channels that developed on stagnant ice between the Pleasant Ridge moraine (on the bedrock high) and the active, but receding, glacial margin. The Kaskaskia Sublobe glacier need not be generalized as active or stagnant ice overall; these conditions would have varied both spatially and temporally, depending on the regional topographic configuration and glacial thickness profile.

Electrical Resistivity Imaging

A two-dimensional resistivity survey was conducted from July 10 to 14, 2006, on Pleasant Ridge to assist with mapping of Quaternary deposits and with economic assessment of sand and gravel resources. The resistivity survey includes parts of Sections 19 and 30 of T 1 S, R 6 W and parts of Sections 24, 25, and 26 of T 1 S, R 7 W, St. Clair County. Approximately 23,000 ft (7,000 m) of continuous resistivity data was acquired on five connecting profiles (Figures 5.1 and 5.3) using the Wenner electrode configuration with an ABEM Terrameter SAS 3000 and LUND acquisition system

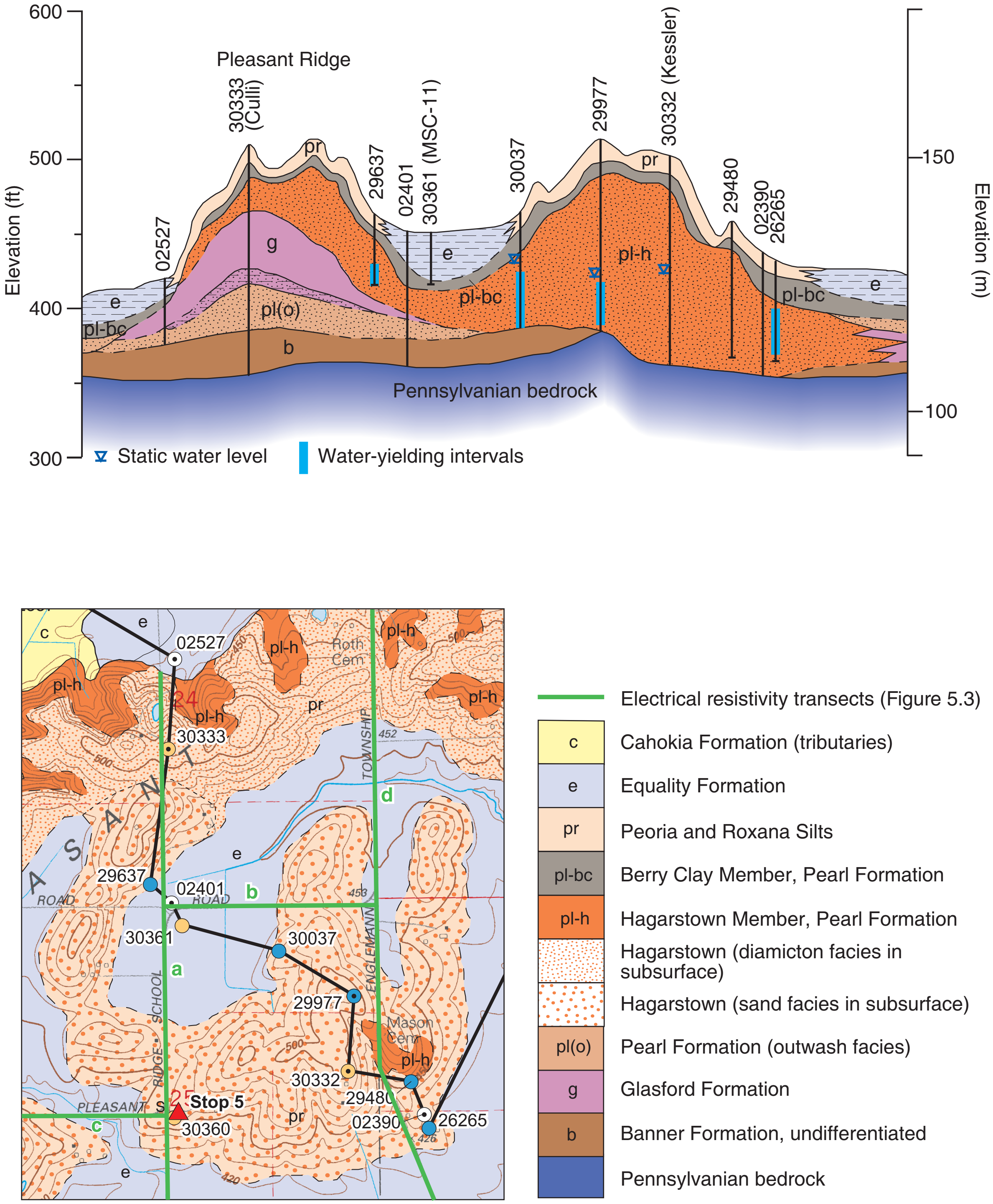


Figure 5.1 Location map (1:24,000) of the Pleasant Ridge area, St. Clair County, Illinois (Stop 5), a portion of the Mascoutah 7.5-minute topographic quadrangle. The surficial geologic map and cross sections are from Grimley (2010). Portions of the electrical resistivity transects are shown. Topographic map courtesy of the U.S. Geological Survey.

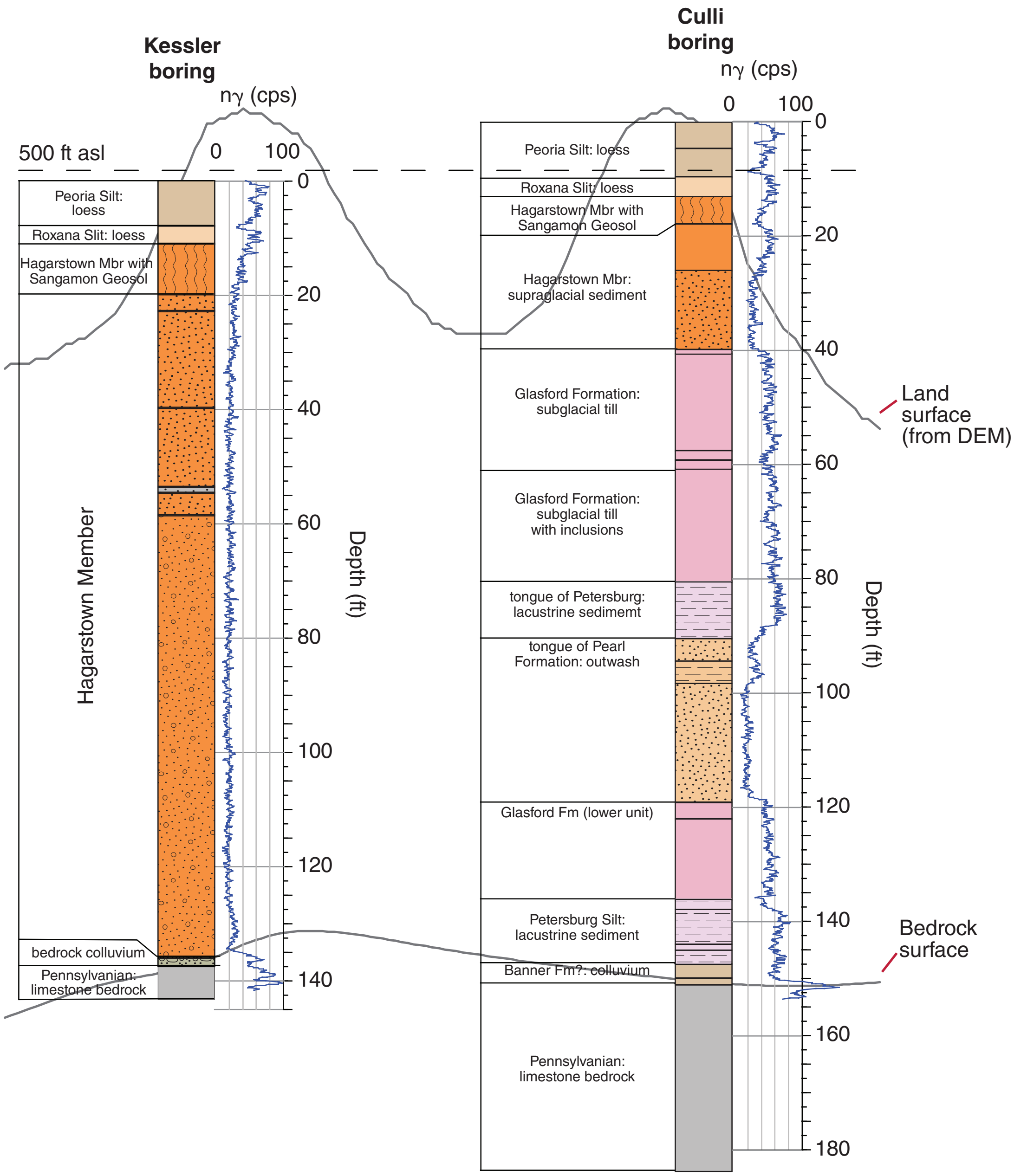


Figure 5.2 Stratigraphic column for the Culli and Kessler borings. The Culli boring is on Pleasant Ridge proper (morainal), whereas the sand- and gravel-rich Kessler boring is on a large hill just south of Pleasant Ridge. asl, above sea level; nγ, natural gamma; cps, counts per second; DEM, digital elevation model.

(Appendix A). The profiles were processed with RES-2DINV, a computer algorithm (Loke and Barker 1996). Two north–south resistivity profiles are each more than a mile long and extend from the lake plain on the south across the southern ridges, interridge sediments, and Pleasant Ridge. The long profiles illustrate the contrasting sediment characteristics within these ridges. Line A begins in a field and continues north along the west side of the road across the low between the ridges and then up and over Pleasant Ridge.

The resistivity profiles (Figure 5.3) suggest that Pleasant Ridge (the long northern ridge) is composed of low-resistivity, fine-grained sediment (diamicton). This fine-grained sediment also underlies surficial lacustrine sediment between the ridges. Because of their near-surface occurrence and extremely low resistivity values, the lacustrine sediments (mostly Equality Formation, possibly underlain by Berry Clay/Teneriffe Silt) can be accurately imaged in two dimensions with the resistivity profile data. This is particularly clear in lines d, c,

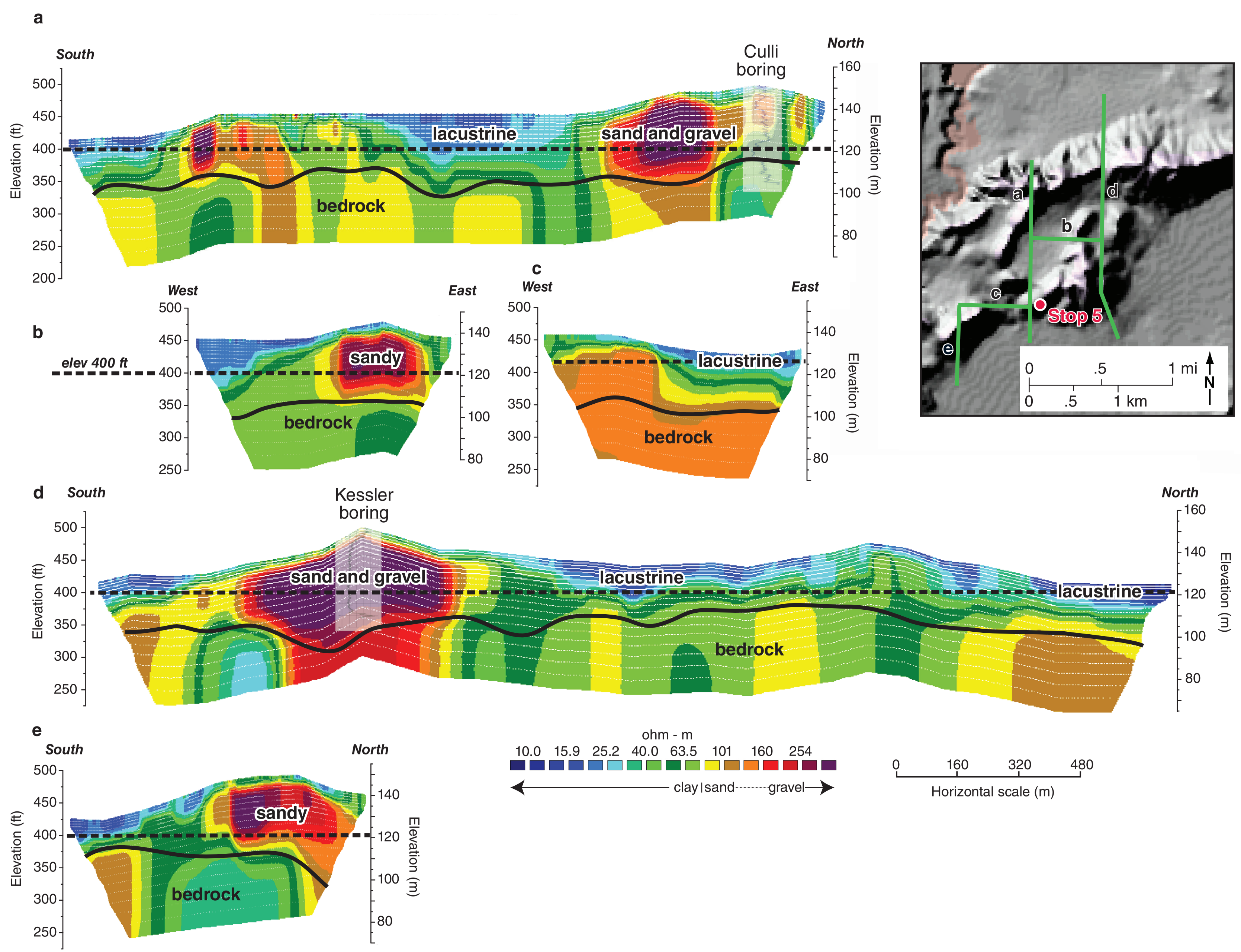


Figure 5.3 Resistivity profiles in the Pleasant Ridge area. Zones of very high, moderately high, and very low resistivity are highlighted in red, orange, and blue respectively. Distances and elevations are in meters. Bedrock surface is indicated by the solid black line.

and a. The low-resistivity lacustrine sediments onlap onto the high-resistivity coarse-grained (Hagarstown Member) sediments of the ridge. In contrast to the main east–west hill of Pleasant Ridge, the associated ridges to the south (more circular or smaller in size) are generally cored by high-resistivity coarse-grained sediments (sand and gravel). This interpretation is confirmed by the Kessler borehole (county #30332; Figures 5.1 and 5.2), in which sand and gravel were encountered, with some zones containing significant secondary calcite cement. Diamicton is of variable and generally low resistivity, but is not specifically highlighted in Figure 5.3. These deposits form the core of Pleasant Ridge and underlie the very low resistivity sediments in other areas. The approximate location of the bedrock surface is indicated on each of the images. The bedrock position is generally determined by a vertical change in resistivity. The depth is confirmed at the Kessler borehole and at a water well on line d. Four geophysical zones were delineated based on the electrical resistivity data.

Geophysical Zone 1—Very High Resistivity in Ridge-Core (Red)

This geophysical zone is caused by a combination of coarse sand and gravel along with dry conditions (low water table) on the ridge crests. Materials in this zone have the highest resistivity (Figure 5.3) and contain the most consistently coarse-textured sediment encountered in the survey. Locally, zone 1 areas may be affected by secondary carbonate zones. This unit is found on the large hill south of Pleasant Ridge (southern line d and eastern line b) and is confirmed by thick sand and gravel in the Kessler borehole (Figure 5.2). This resistivity unit is also found on line e and on the northern end of line 1 (but south of the northernmost area). Although northern line a appears to be on Pleasant Ridge moraine, a separate feature is actually adjoining Pleasant Ridge just south of the Culli boring. Lithostratigraphically, these areas include thick deposits of the sandy facies, Hagarstown Member, with about a 10- to 13-ft (3- to 4-m) blanket of loess.

Geophysical Zone 2—Moderately High Resistivity in Ridge-Core (Orange)

This zone is found on the southern part of the Koltz Road Line and the eastern part of the Pleasant Ridge School Road East Line. These materials have somewhat lower resistivity compared with zone 1 (Figure 5.3) and, in general, do not attain the same thickness or elevation. In part according to calibrations to test borings and well records (Grimley 2010), zone 2 correlates with sand and gravel that is finer grained or has a higher water content than in zone 1. Zone 2 is found on line e, line c, and line b and on southern line a (where we have our bus stop). Lithostratigraphically, these areas include material of the sandy facies, Hagarstown Member, with about a 10- to 13-ft (3- to 4-m) blanket of loess.

Geophysical Zone 3—Moderate-Resistivity Ridges (Green)

This zone includes areas on Pleasant Ridge proper (northern lines a and d)—that is, the morainic areas that include the Culli boring and geomorphically similar areas to the east and west. Resistivity values are moderate, intermediate between areas dominated by sand and gravel (zones 1 and 2) and areas dominated by fine-grained lake sediment (zone 4). Zone 3 deposits include a mixture of some sand and gravel, some buried lake sediments, and significant thicknesses of diamicton with inclusions of older materials. Lithostratigraphically, zone 3 includes material of the Glasford Formation, the Petersburg Silt, mixed facies of the Hagarstown Member, and tongues of the Pearl Formation. These areas have a 10- to 13-ft (3- to 4-m) blanket of loess.

Geophysical Zone 4—Low-Resistivity Lowland (Blue)

This zone is found below flat intraridge areas on eastern Pleasant Ridge School Road, the center of Koltz Road, the west part of the Pleasant Ridge School Road East Line, and the north part of the Brickyard Road Line on lines a, b, c, and d between the ridges (Figures 5.1 and 5.3). The maximum thickness of this zone is about 65 ft (20 m) in the center of the Koltz Road Line near its junction with the Pleasant School Road East Line. This geophysical zone probably reflects the presence of fine-grained lacustrine sediments, as substantiated by deposits in borehole MSC-11 (county #30361; Figure 5.1). The areas of lowest resistivity within this zone occur at a lower surface elevation, on the south ends of the south–north profiles. The fine-grained material ranges from 30 to 65 ft thick (9 to 20 m thick), with the base of these zones at an elevation similar to other areas in zone 4. Lithostratigraphically, these areas include the Equality Formation and the Berry Clay Member, and perhaps some areas of fine-grained Teneriffe Silt or Glasford Formation.

Bedrock

Bedrock in the area varies from high-resistivity carbonate to low-resistivity shale. In many areas, it is difficult to distinguish the bedrock interface from other materials because the resistivity contrast is low. Where bedrock can be distinguished, it generally occurs at elevations below 360 ft (110 m). This resistivity method is not appropriate for detailed mapping of the bedrock surface in this area, although a best interpretation is shown in Figure 5.3.

STOP 6: Highbanks Road Section

David A. Grimley and Elizabeth C. Geiger

Please use caution at this site—slopes may be very slippery.

Introduction

The Highbanks Road Section is on the Mascoutah 7.5-minute quadrangle in the SW, SW, Section 22, T 1 S, R 6 W in St. Clair County, Illinois (Figure 6.1), and was discovered during surficial geology mapping (Grimley 2010). The section contains an extensive exposure, several hundred feet long and up to 35 ft (11 m) high, of mostly fine-grained slackwater lake deposits (Wisconsin Episode) in a terrace along a cutbank of the Kaskaskia River (Figure 6.1). An active meander has resulted in 100 to 200 ft (30 to 60 m) of northward erosion into the terrace deposits during the last 20 years, based on orthophotographic comparisons. From local anecdotes, the erosion has led to slumping of the soft banks and the destruction of a small road and several cabins over recent decades. Stratified silty and clayey lacustrine deposits are exposed in the bank, along with deltaic or fluvial sands. Some zones in the slackwater succession are fossiliferous, thus providing a paleoenvironmental record of glacial Lake Kaskaskia during the Wisconsin Episode. Although the site reveals mainly deposits from the last glaciation, Holocene alluvium (Cahokia Formation) also occurs as a cut-and-fill deposit into the slackwater terrace deposits at the western or downstream portion of the section. Other well-exposed Holocene alluvial sections can be viewed by canoeing downstream along the Kaskaskia River between here and Fayetteville.

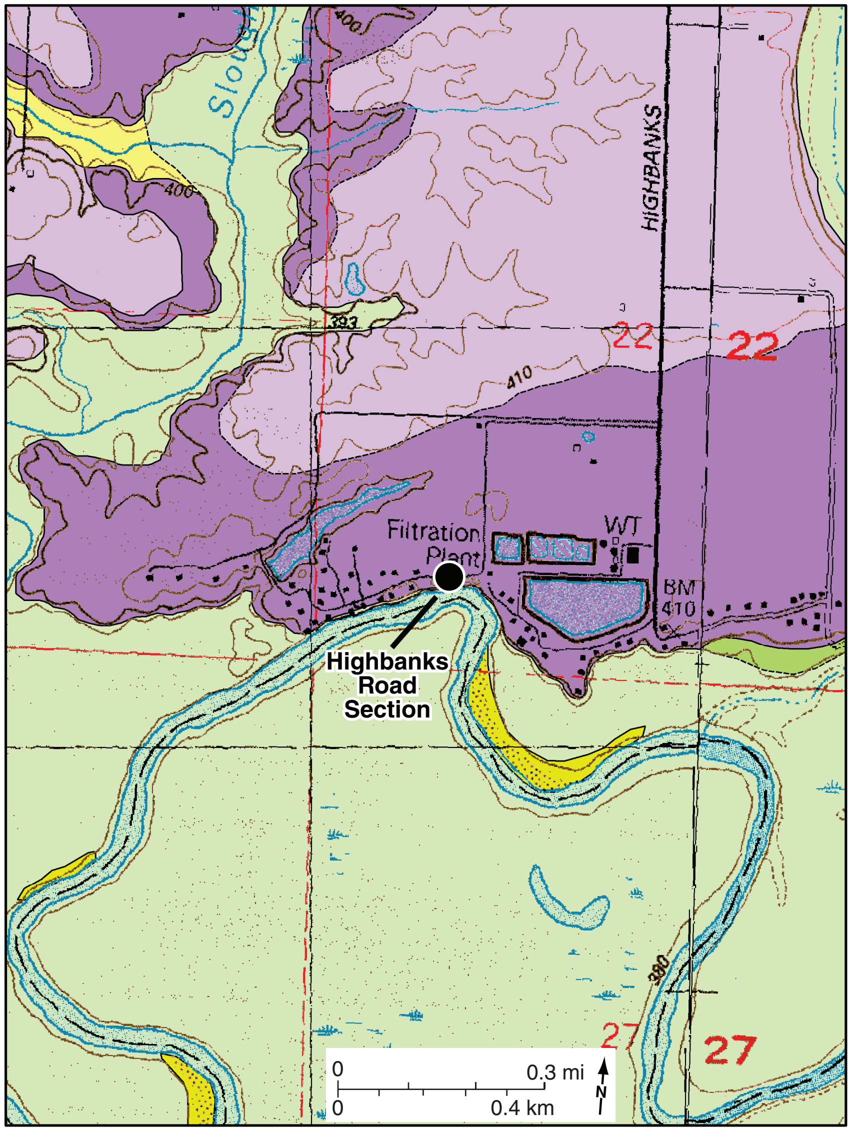
Geologic Strata

The geologic strata exposed at the Highbanks Road Section are divided into five units (Figures 6.2 and 6.3). At the base of the section, about 3 ft (1 m) of well-sorted sand (unit A) is exposed at low stages of the Kaskaskia River. The base of this unit was not exposed. Once interpreted as the Pearl Formation (Grimley 2010), unit A has been reinterpreted as the Henry Formation (Wisconsin Episode) based on younger than anticipated radiocarbon ages in the overlying units in a seemingly conformable sequence. Sandy deposits in the subsurface of the lower Kaskaskia Valley are as much as 50 ft (15 m) thick or more (Larson 1996; Grimley 2010; Grimley and Phillips 2011), with basal portions probably of Illinois Episode age. It is anticipated that a considerable thickness of sand occurs below the exposure of unit A.

Units B through F are mainly lacustrine sediments (Equality Formation). Unit B is interbedded sand and silt loam that varies from 0.5 to 1.5 ft (0.2 to 0.5 m) thick. Above that is 6 ft (2 m) of silty clay loam in unit C. The next younger stratum is a fossiliferous silt deposit about 6 ft (2 m) thick, unit D, containing reddish brown iron nodules. Unit E contains abundant molluscan fossils and thus was the primary focus of paleoecological study. Unit E is about 10 ft (3 m) thick and is subdivided into three units: E1 (lowest) is a calcareous silty clay loam with a 1.0-in. (2.5-cm) layer of secondary carbonate; E2 is a laminated silt that contains interbedded very fine sand (probably fluvial or deltaic); and E3 contains a more clayey zone within mostly laminated silt. Unit F, the uppermost unit at the section, is a 12-ft-thick (4-m-thick) dark brown, silty clay loam to silty clay that is laminated, with common secondary carbonate concretions (especially near the basal contact) and a hackly structure. The clayey, smectitic material is highly fissured from shrinkage when dry. The modern soil solum is contained within the uppermost unit F. The uppermost 1.0 ft (0.3 m) of the soil is silty and could be interpreted as eolian (loess), but because of weathering, an interpretation of water-laid silt is difficult to rule out.

Below is a description of the Highbanks Road Section:

> According to field descriptions by David Grimley on November 8, 2005, and Elizabeth Geiger in May 2006; large cutbank (several hundred feet or more wide) along Kaskaskia River; near intake for water treatment plant; river level ~372′ [113 m] from topographic map; base of section at river level.
>
> **API** 121633027500
> **ELEVATION** 405 asl—section top (approx.)
> **FIELD ID** MSC-3f
> **LATITUDE** 38.425236 N
> **LONGITUDE** 89.756133 W
>
> **0 to 1 ft [0 to 0.3 m]**
> Peoria Silt?, heavy silt loam
>
> **1 to 10 ft [0.3 to 3.0 m]**
> Unit F of Equality Formation, silty clay loam, brown, thickly bedded, contains multiple alluvial deposition events and paleosols
>
> **10 to 18 ft [3.0 to 5.5 m]**
> Unit E of Equality Formation (E1, E2, and E3); 1.5 to 4.0 tsf [tons/ft^2], contains fossil mollusks (mainly aquatic) including *Fossaria*, *Pomatiopsis*, and small fingernail clams



Cahokia Formation (silty to sandy alluvium)—floodplain level
Cahokia Formation (clayey alluvium)—floodplain level
Cahokia Formation (clayey alluvium) (high Holocene terrace)
Equality Formation (fine-grained lacustrine)—low terrace
Equality Formation (fine-grained lacustrine)—high terrace

Figure 6.1 Location map of the Highbanks Road Section, St. Clair County, Illinois (Stop 6). The surficial geologic map (Grimley and Phillips 2011) is overlain on a portion of the Mascoutah 7.5-minute quadrangle. Areas shaded light and dark purple are the last glacial terraces containing Equality Formation. Areas shaded yellow (sandy) and green (clayey) contain late Holocene alluvium.

Figure 6.2 Photograph of the Highbanks Road Section taken in 2010, looking to the northwest. The light-colored bed is the silty and fossiliferous unit E of the Equality Formation.

E3 (upper): silt loam, light olive brown (2.5Y 5/4); well sorted, calcareous, laminated; fossiliferous w/ gastropods concentrated in a bed near top of unit

E2 (middle): silt, light brownish gray (10YR 6/2); well sorted, dolomitic to calcareous, contains zones of very fine sand, cross-bedding in some areas (dipping to east)

E1 (lower foot): silty clay loam, with a 1-in. layer of secondary calcium carbonate; yellow (2.5Y 7/6), well sorted, fossiliferous

18 to 23 ft [5.5 to 7.0 m]
Unit D of Equality Formation, silty clay loam to silty clay; light yellowish brown 2.5Y 6/3 to 10YR 5/4; weakly dolomitic; contains few fossil mollusks, some iron nodules (7.5YR 6/8)

23 to 30 ft [7.0 to 9.1 m]
Units B and C of Equality Formation, silty clay loam, grayish brown (2.5Y 5/2) to dark gray (5Y 4/1), leached to slightly calcareous; some secondary carbonate and iron nodules; basal 0.5 ft [0.2 m] is a noncalcareous, yellowish brown (10YR 5/4), silt loam

30 to 33 ft [9.1 to 10 m]
Unit A: Henry Formation, medium sand, well sorted, leached, 10YR 7/6 (yellow), base of unit at Kaskaskia River level (11/8/2005).

Paleontology/Paleoecology

Identification by Geiger (2008) of the molluscan fauna from units D and E at the Highbanks Section has been useful for paleoecological interpretations and paleoclimate estimations during the time of glacial Lake Kaskaskia (see the Introduction section of this volume for lake history). Furthermore, gastropods in key zones were utilized for isotopic analyses and radiocarbon age determinations.

The study by Geiger (2008) focused on the gastropod fauna, although small bivalves (*Pisidium* sp. and *Sphaerium* sp.) were also present. Four large grab samples (one-gallon bags) were collected from zones with visibly abundant gastropod shells in unit E1 (HBC-1,

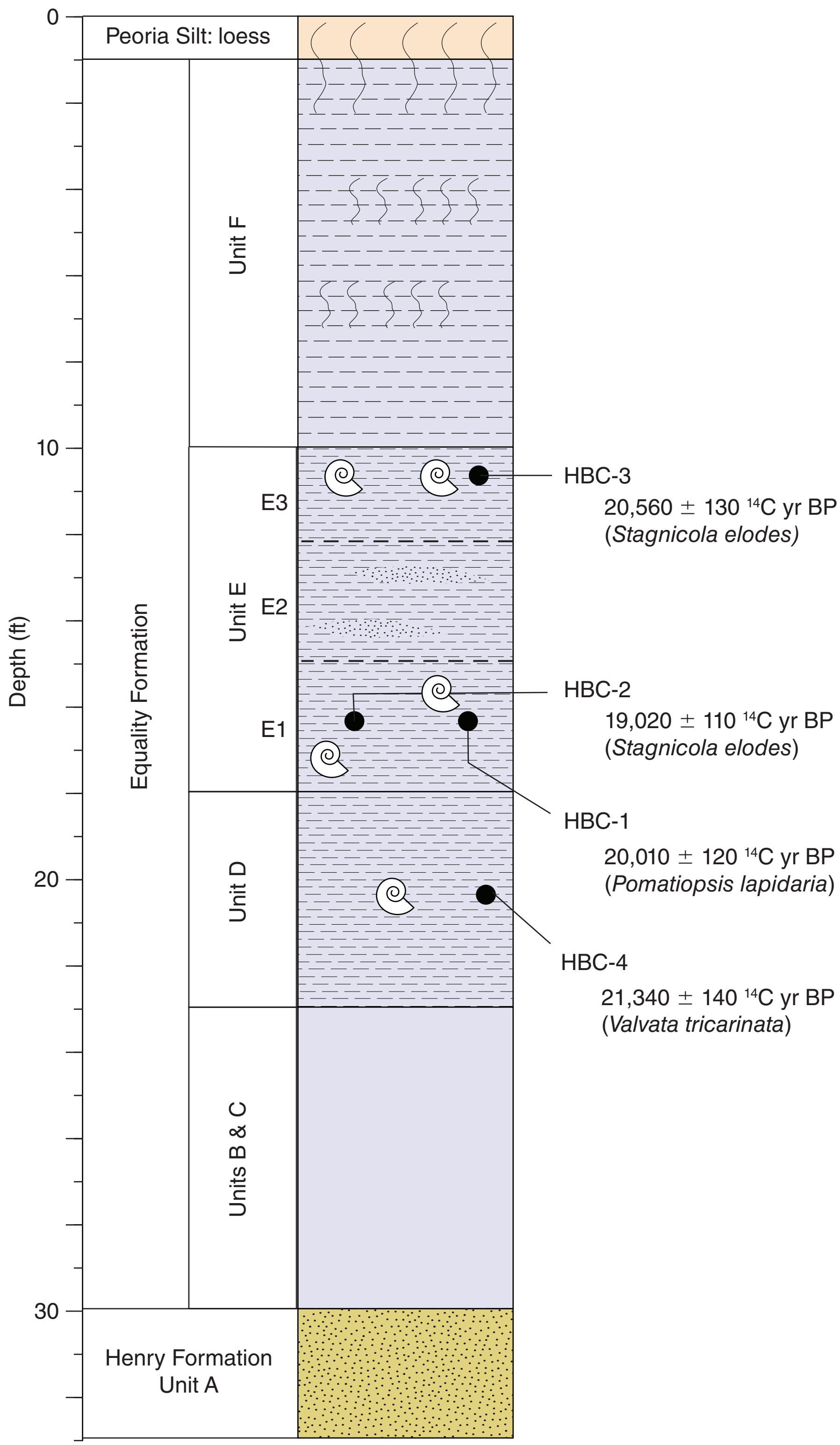


Figure 6.3 Stratigraphic column and location of mollusk collection sites at the Highbanks Road Section. One accelerator mass spectrometry radiocarbon age was determined on gastropod shells from each of the four grab sample locations (HBC-1, HBC-2, HBC-3, HBC-4).

HBC-2) and unit E3 (HBC-3) in May 2006, and in unit D (HBC-4) in November 2006. After samples were soaked in water for 48 hours in the laboratory, they were wet-sieved and the collected shells were identified and catalogued. Subsequently, accelerator mass spectrometry (AMS) radiocarbon dates were obtained on some of the larger sized gastropods in fall 2010 to provide age constraints on glacial Lake Kaskaskia.

Most mollusk shells were <1 cm (<0.4 in.), except *Stagnicola elodes*, which was up to 1.5 cm (0.6 in.) in length. The results of the molluscan study [somewhat revised from Geiger (2008) after further study] are reported in Table 6.1. However, extremely minute (~1 to 2 mm) juvenile individuals of various species are not reported. For instance, hundreds of minute juveniles of *Fossaria* sp. and *Stagnicola* sp. were found in sample HBC-2, probably indicating a period of rapid deposition and burial before maturity.

Unit D, the oldest unit sampled (HBC-4), has a molluscan assemblage dominated by *Valvata tricarinata*, *Fossaria* sp., and *Gyraulus parvus*, with few *Physella gyrina*. This assemblage suggests perennial lake conditions (genera *Valvata* and *Gyraulus* have gills rather than lungs) with high alkalinity (Cvancara 1983). The presence of a few individuals of the terrestrial *Catinella avara*, usually found along vegetated lakeshores, in swampy places, or under logs (Burch and Jung 1988), suggests minor wash-in from a nearby shore. Overall, the assemblage is consistent with a large slackwater lake in the Kaskaskia Basin that contained aquatic vegetation as a food source and habitat. The lake existed during the last glaciation, based on a radiocarbon age of 21,340 ± 140 ^{14}C yr BP (ISGS-A1652) on *V. tricarinata* shells, with δ^{13}C = –8.0.

In unit E1, one sample (HBC-2) has primarily freshwater aquatic species, whereas the other sample (HBC-1) has a mixture of terrestrial and shallow aquatic species. Sample HBC-2 is dominated by *Fossaria* sp., *G. parvus*, *Helisoma anceps*, *P. gyrina*, and *S. elodes*. A few terrestrial species, such as *V. elatior*, were also found. Overall, the implied environment is a shallow lake with some species washing in from nearby muddy lakeshores (i.e., *Oxyloma retusum*). The lake level likely fluctuated, with some aquatic species having lungs (*H. anceps* and *S. elodes*) and thus being able to withstand temporary lake conditions. The presence of *Planorbula armigera* indicates a eutrophic lake (Cvancara 1983). Climatically, the occurrences of the northern species *Fossaria galbana* (common name: Boreal Fossaria) and *V. elatior* point to a modern analogue in the northern Great Lakes region, western Ontario, or Manitoba (Burch 1989; Nekola and Coles 2010). *Stagnicola elodes* shells in unit E1 were radiocarbon dated at 19,020 ± 110 ^{14}C yr BP (ISGS-A1650), with δ^{13}C of –9.4.

The presence of an individual of the terrestrial species *C. alticola* in sample HBC-1 (unit E1) further substantiates cold, moist conditions during the LGM. *Columella alticola* has a present distribution in the northern portions of Quebec, Ontario, and Manitoba (Nekola and Coles 2010), as well as in high elevations of the Rocky Mountains (Hubricht 1985). However, the presence of one individual should not be overinterpreted. Other indicator species include *V. elatior* and *Vertigo modesta*, which have considerable abundance in HBC-1 and have distributions today in Ontario, Manitoba, and the northern Great Lakes area, with *V. modesta* restricted mainly to Canada (Nekola and Coles 2010). The additional presence of *C. exile* suggests a moist area or spring at the base of a slope or edge of a floodplain. Last, the occurrences of *Fossaria* sp. (shallow lacustrine), *Pomatiopsis lapidaria* (amphibious), and *S. elodes* (shallow lacustrine) imply a fluctuating slackwater lake that seasonally or periodically flooded the mudflats along the shoreline. *Pomatiopsis* shells were radiocarbon dated at 20,010 ± 120 ^{14}C yr BP (ISGS-A1649), with δ^{13}C of –6.1. This age is 1,000 years older than the shells dated from sample HBC-2, which was collected 10 ft (3 m) laterally in the same unit E1, so the actual subunit age could be somewhere between 20,000 and 19,000 ^{14}C yr BP.

Unit E3 is the uppermost fossiliferous unit of the Equality Formation. The molluscan assemblage in this unit (sample HBC-3) records a return to shallow lacustrine or quieter water conditions following the fluvial-deltaic conditions represented by unit E2 (not fossiliferous). Terrestrial and amphibious gastropods are absent in E3, but the fossil concentration is very low in this sample. The occurrences of *F. dalli*, *P. gyrina*, and *S. elodes* all suggest shallow lacustrine conditions in a mesotrophic or eutrophic lake (Cvancara 1983). One *S. elodes* shell was radiocarbon dated at 20,560 ± 130 ^{14}C yr BP (ISGS-A1651), with δ^{13}C of –9.1. This age is slightly older than ages for unit E1, suggesting minor hardwater effects, rapid deposition, or slight reworking of shell material.

Isotopic Data

Carbon and oxygen isotopes were analyzed for several individuals of the genera *Carychium*, *Vertigo*, and *Gastrocopta* (all terrestrial species) from the last glacial Equality Formation at the Highbanks Road Section. The samples analyzed were from HBC-1 (unit E1) because it contained the greatest abundance of terrestrial gastropods. For comparison, modern shells of the same genus,

Table 6.1 Molluscan fauna from four grab samples (one-gallon bags) from Equality Formation units at the Highbanks Road Section[1]

Genus	Species	Habitat	HBC-3, unit E3	HBC-1, unit E1	HBC-2, unit E1	HBC-4, unit D
Gastropods (total number discludes many immatures)			8	225	405	103
Carychium	*exile canadense*	Terrestrial	0	31 (4 iso.)	1	0
Catinella	*avara*	Terrestrial	0	15	0	6
Cochlicopa	*lubrica*	Terrestrial	0	3	0	0
Columella	*alticola*	Terrestrial (cold)	0	1	0	0
Columella	*simplex*	Terrestrial	0	4	0	0
Discus	*whitneyi*	Terrestrial	0	9	0	0
Euconulus	*fulvus*	Terrestrial	0	2	0	0
Fossaria	*dalli*	Shallow lake	4	30	0	0
Fossaria	*galbana*	Shallow, cold lakes	0	0	31	0
Fossaria	*obrussa*	Shallow lake	0	20	1	1
Fossaria	*parva*	Amphibious	0	0	26	23
Gastrocopta	*tappaniana*	Terrestrial	0	4 (2 iso.)	2	0
Gyraulus	*hornensis*	Perennial lake	0	0	6	2
Gyraulus	*parvus*	Perennial lake	0	0	200	30
Helisoma	*anceps*	Shallow lake	1	1	22	0
Nesovitrea	*electrina*	Terrestrial	0	6	0	0
Oxyloma	*retusum*	Lakeshore	0	0	17	0
Physella	*gyrina*	Shallow lake	1	10	24	6
Planorbula	*armigera*	Eutrophic lakes	1	5	19	0
Pomatiopsis	*lapidaria*	Amphibious	0	34 (^{14}C)	0	0
Punctum	*minutissimum*	Terrestrial	0	1	0	0
Stagnicola	*caperata*	Shallow lake	0	0	12	7
Stagnicola	*elodes*	Lake	1 (^{14}C)	4	39 (^{14}C)	0
Vallonia	sp.	Terrestrial	0	5	0	0
Valvata	*tricarinata*	Perennial lake	0	0	0	28 (^{14}C)
Vertigo	*elatior*	Terrestrial	0	15 (3 iso.)	5	0
Vertigo	*hubrichti*	Terrestrial	0	20 (3 iso.)	0	0
Vertigo	*modesta*	Terrestrial (cold)	0	5	0	0
Bivalves						
Sphaerium	*lacustre* (~5–8 mm)	Eutrophic lakes	0	0	>60	0
Pisidium	*casertanum* (4–5 mm)	Lake	0	0	>20	0
Pisidium	sp. (<3 mm)	Lake	4	6	0	0
Ostracodes *candona*						
Fabaeformiscandona	*caudata*?	Lake	0	0	>1	0
Fabaeformiscandona	*candida*?	Lake	0	0	>1	0
Unknown ostracodes			7	0	0	0

[1]Identifications by E. Geiger and D. Grimley. Samples utilized for isotopic (iso.) or radiocarbon (^{14}C) analyses are indicated.

collected in the early 20th century from near Du Bois, Illinois [Washington County, about 31 mi (50 km) to the southeast], were also analyzed to determine whether significant climatic or vegetation shifts might be recorded by the isotopic data. The modern shells are from the Frank C. Baker Collection (Illinois Natural History Survey; Kevin Cummings, curator) and were sampled by A.A. Hinkley in the vicinity of the town of Du Bois in the early 20th century, probably from a woodland in the Little Muddy River watershed. *Carychium exile* and *Gastrocopta tappaniana* were found at both the fossil and modern sites; however, the modern *Vertigo* species, *Vertigo milium*, differed from the HBC-1 *Vertigo* spp. that are more typical of northern climates (Nekola and Coles 2010). Individual shells, both Pleistocene and modern, were carefully cleaned of sediment by gently crushing and spraying with distilled H_2O on a filter funnel with a vacuum connection before analyses.

Overall, the mean $\delta^{18}O$ values measured were 0.6 ± 1.6‰ (n = 13) for the last glacial shells (HBC-1) and –1.9 ± 1.0‰ (n = 16) for the modern shells (Frank C. Baker Collection). The $\delta^{13}C$ values had more contrast, with a mean of –7.5 ± 1.4‰ (n = 13) for the last glacial shells compared with –11.6 ± 2.0‰ (n = 16) for modern shells (Table 6.2). The internal consistency within species and within age groupings seems reasonably significant. It is therefore clear that the isotopic composition of equivalent-genera last glacial shells is consistently heavier compared with modern shells, particularly in the carbon isotope. Similar differences have been found when comparing isotopic compositions of last glacial and modern terrestrial shells in northern Spain (Yanes et al. 2012). The $\delta^{18}O$ of terrestrial gastropod shells can be affected by temperature, relative humidity, and the $\delta^{18}O$ of the water (typically rainwater) ingested by the snail. One reasonable interpretation suggested by Yanes et al. (2012) for the Spanish record is that the heavier $\delta^{18}O$ values in glacial-age shells are a result of the lower relative humidity, and that this overwhelmed the effects of cooler temperatures. It should be noted that isotopic values will reflect the time period of shell growth only during the warmer months, which is a shorter time frame in colder climates. The carbon isotope data are generally reflective of the gastropod diets and thus the type of vegetation present during the seasons of shell growth (Colonese et al. 2010). A $\delta^{13}C$-depleted diet is suggested during glacial times compared with the modern samples, which were likely collected from a woodland area, perhaps indicating a higher proportion of C_4 to C_3 plant ingestion than during the late Holocene. So perhaps conditions during the LGM were more of an open woodland/wetland rather than a dense floodplain forest, which is more typical of this area today and before settlement. Some detrital carbonate ingestion or higher stress on C_3 plants during glacial times could also have caused somewhat heavier carbon isotopic compositions (Yanes et al. 2012). Within the Wisconsin Episode gastropods from HBC-1, the *Vertigo* samples appeared to have less variation than the *Carychium* samples.

Regional Interpretations and Conclusions

The Highbanks Road Section records the episodic deposition of the Henry and Equality Formations in fluvial and extensive lacustrine environments, respectively. Depositional environments in units B through F varied with fluctuating lake levels and included shallow lacustrine facies, near-shore facies, lakeshore facies, and brief periods of fluvial deposition (unit E2). The deposition of units D and E, documented by radiocarbon dating between about 21.5 and 19.0 ^{14}C ka BP, is coincident with the early part of the global LGM (Mix et al. 2001). This timing of Equality Formation deposition probably records a period of maximum onlap of glacial Lake Kaskaskia in this area, located about 30 mi (50 km) northeast of the confluence of the Kaskaskia and Mississippi Rivers. The spread in ages of approximately 2,500 years over about 13 ft (4 m) of section thickness suggests an average deposition rate of 0.6 in. (1.6 cm) per decade for units D and E. On the basis of the radiocarbon ages, the overlying, nonfossiliferous unit F is likely the latest Wisconsin Episode, but possibly may be early Holocene if found to be younger than 11,700 calendar years (Walker et al. 2009), equivalent to about 10,000 ^{14}C years.

The gastropod shell radiocarbon ages are between 20.5 and 19.0 ^{14}C ka BP for unit E (including subunits E1, E2, and E3). This time of the LGM is also synchronous with the advance of Wisconsin Episode glaciers in east-central Illinois to their southern terminus (Shelbyville Moraine), based on *Picea* wood and organic silt ages ranging from 20.6 to 19.3 ^{14}C ka BP below till at the Charleston Quarry (Hansel et al. 1999). This correspondence in age may suggest input of Lake Michigan Lobe sediment to unit E from the upper Kaskaskia Valley, which would have drained a relatively small portion of the southwesternmost Lake Michigan Lobe [~500 mi^2 (1,295 km^2)] during this geologically brief time period. Units E2 and E3 have a distinctly lower clay content than do units below (Appendix B) and include beds of very fine sand. This may suggest more rapid deposition rates from a distal source of outwash that was transported southwesterly into Glacial Lake Kaskaskia. The

Table 6.2 Carbon and oxygen isotopic data from gastropod shell carbonate[1]

Sample no.	Sample ID	ISGS no.	$\delta^{13}C_{VPDB}$	$\delta^{18}O_{VSMOW}$	$\delta^{18}O_{PDB}$[2]
HBC-1	*Carychium exile* (A)	KL008074	–6.76	33.31	2.33
HBC-1	*C. exile* (B)	KL008075	–5.37	27.54	–3.27
HBC-1	*C. exile* (C)	KL008076	–5.53	29.67	–1.20
HBC-1	*C. exile* (D)	KL008077	–6.67	29.76	–1.12
HBC-1	*Vertigo elatior* (A)	KL008078	–6.59	32.32	1.37
HBC-1	*V. elatior* (B)	KL008079	–9.39	31.71	0.78
HBC-1	*V. elatior* (C)	KL008080	–8.55	32.31	1.36
HBC-1	*Vertigo hubrichti* (A)	KL008090	–7.57	32.35	1.40
HBC-1	*V. hubrichti* (B)	KL008082	–9.40	31.08	0.17
HBC-1	*V. hubrichti* (B) Dup.	KL008087	–9.17	33.63	2.64
HBC-1	*V. hubrichti* (C)	KL008083	–8.14	31.89	0.95
HBC-1	*Gastrocopta tappaniana* (A)	KL008085	–7.53	32.95	1.98
HBC-1	*G. tappaniana* (B)	KL008086	–6.92	31.46	0.54
Baker[3]	*C. exile* (A)	10168	–11.38	29.88	–1.00
Baker	*C. exile* (B)	10142	–10.90	29.92	–0.96
Baker	*C. exile* (C)	10143	–6.41	28.37	–2.46
Baker	*C. exile* (D)	10144	–11.55	29.14	–1.71
Baker	*C. exile* (E)	10145	–11.84	28.95	–1.90
Baker	*C. exile* (F)	10146	–11.67	29.81	–1.07
Baker	*Vertigo milium* (A)	10153	–14.83	27.79	–3.03
Baker	*V. milium* (B)	10171	–13.45	25.83	–4.92
Baker	*V. milium* (C)	10155	–12.05	28.87	–1.98
Baker	*V. milium* (D)	10156	–13.38	28.66	–2.18
Baker	*V. milium* (E)	10173	–13.41	28.85	–2.00
Baker	*G. tappaniana* (A)	10147	–12.02	29.68	–1.19
Baker	*G. tappaniana* (B)	10169	–11.91	30.15	–0.73
Baker	*G. tappaniana* (C)	10149	–11.81	29.54	–1.32
Baker	*G. tappaniana* (D)	10150	–10.41	28.62	–2.22
Baker	*G. tappaniana* (E)	10152	–8.45	29.27	–1.59

[1]Shari Fanta, analyst, Isotope Geochemistry Section, Illinois State Geological Survey (January 2011, HBC-1 Pleistocene samples; August 2012, modern samples).

[2]Conversion equation used: $\delta^{18}O_{VPDB} = 0.97002 \times \delta^{18}O_{VSMOW} = 29.98$ (Clark and Fritz 1997; Coplen et al. 1983). Isotopic values were calibrated based on NBS18 and NBS19 standards. VPDB = Vienna Pee Dee Belemnite; VSMOW = Vienna Standard Mean Ocean Water. Isotopic composition is reported relative to the VPDB (Carbon) and VSMOW (Oxygen) standards

[3]All Baker entries indicate sample ID numbers from the Frank C. Baker Collection, Illinois Natural History Survey: *C. exile* (Z-23830); *V. milium* (Z-25101); *G. tappaniana* (Z-25060).

rapid sediment accumulation during this time resulted in quick burial and excellent preservation of the molluscan shells. Alternatively, more rapid deposition could have been controlled by faster loess deposition and subsequent transport into valleys as a result of more open vegetation on landscapes during this time period. However, the distal glacial source is preferred as the dominant source for unit E. Deposition of the more clayey units (units B, C, D, and F) perhaps represents times before and after the peak glacial advance (time of unit E), when distal outwash inputs were minimal or nonexistent. During these periods, the Lake Michigan Lobe would not have been far enough south to provide glacial meltwaters to the Kaskaskia Valley; the Lake Michigan meltwaters would have been trapped in proglacial lakes or discharged elsewhere, such as to the Wabash or Illinois valleys.

On the basis of the modern distribution of gastropod fossils (Clarke 1981; Nekola and Coles 2010), the climatic conditions during deposition of units D and E were perhaps similar to those found today in western Ontario about 49 to 50° latitude (north of Lake Superior). This region is far enough north to have the cold-climate species *F. galbana* (Burch 1989), *V. modesta* (Nekola and Coles 2010), and *C. alticola* (Nekola and Coles 2010) but south enough to be within the modern distribution of species such as *G. tappaniana* (Nekola and Coles 2010) and *P. lapidaria* (Clarke 1981). The regions in Manitoba and Ontario between the Great Lakes and Hudson Bay are also good analogues for the boreal forest vegetation that is known to have been present in this area during the last glaciation, based on numerous *Picea* fossil wood finds and other paleobotanical records (Curry and Delorme 2003).

DAY 2 ROAD LOG (May 22, 2011)

Leave Mariner's Village Motel (8 a.m.)

- Turn right on William Road. [0.5 mi]
- Turn right on IL-127. [16.9 mi]
- Turn right and merge onto Interstate 70 toward Effingham. [6.8 mi]
- Take the exit ramp at Exit 52 (Mulberry Grove exit). [0.3 mi]
- Turn left on Mulberry Grove Road/County Road 1900 East. [0.4 mi]
- Turn right on US-40 East. [0.9 mi]
- Bear right onto County Road 700 East (sign to Hagarstown). [7.9 mi]
    - Proceed under Interstate 70. [The now-reclaimed borrow pit wall of the former Mulberry Grove Section (Jacobs and Lineback 1969) is to your immediate right.]
    - Pass through the town of Hagarstown after a few miles.
- Turn right on County Road 700 East at VFW Post 3862. [0.4 mi]
- Turn left and go down into the sand and gravel pit.

STOP 7: Vandalia Sand and Gravel Pit

- Turn left and go south on County Road 700 East. [0.9 mi]
- Turn left on County Road 1375 North. [0.1 mi]
- Turn right on 715 East/County Road 10/Fayette County Road 1. [1.5 mi]
- Turn right on County Road 1275 North, which becomes County Road 1300 North. [0.8 mi]
- Turn left on the gravel road to the office, then south on the dirt road (weather permitting) to the site. [0.7 mi]

STOP 8: Central Illinois Materials Sand and Gravel Pit

- Take the dirt road south to the paved road. [0.1 mi]
- Turn right onto County Road 1225 North, which becomes County Road 1150 North. [1.8 mi]
- Pull off on the grassy upland area on the right side of the road (an abandoned, or periodically used, small sand pit is found here on this ridge); it overlooks circular basins to the south and north.

STOP 9: Pittsburg Basin

- Go southwest on County Road 1150 North to County Road 1100 North. [0.6 mi]
- Turn left on County Road 300 East. [2.0 mi]
- Turn right at County Road 10/County Road 900 North and pass through Pittsburg; the road winds around and crosses a creek. [4.0 mi]
- Turn left on Mulberry Grove Road (1900 E), which becomes Mulberry Street. [8.3 mi]
- Mulberry Street turns right and becomes West Clinton Street. [0.1 mi]
- Turn left on Main Street. [0.1 mi]

REST AREA/SNACK BREAK—Keys Restaurant and Lounge (nice view of the lake from the upstairs deck)

- Turn around and go left (west) on West Clinton Street/Keyesport Road. [1.0 mi]
- Turn left onto North Emerald Road, and turn west after a couple miles. [3.5 mi]
- At the stop sign, turn left on Hopewell Road. [1.0 mi]
- Bear right on Marydale Road. [0.5 mi]
- Take the first left onto Fisher Road. [2.0 mi]
- Turn right on Hazlet Park Road. [0.3 mi]
- Turn right into the gravel driveway.

STOP 10: Sodium-Affected Soils in South-Central Illinois

- Go west on Hazlet Park Road. [0.6 mi]
- Continue on to Lake Road. [2.8 mi]
- Turn right onto Resort Drive [0.1 mi], which takes you back to the Mariner's Village Motel.

LUNCH

Enjoy your lunch in the adjacent park or take it with you for the road! Have a safe trip home.

DAY 2 BACKGROUND: Illinois Episode Glacial Sediments and Landforms in the Vandalia Area

Geomorphology

A striking array of ridges arc southwestward along the west side of, and subparallel to, the Kaskaskia Valley near Vandalia (Figures D2, 7.1, and 7.2). From an arbitrary head at the peak of Thrill Hill immediately north of the Vandalia Geologic Area to the banks of Hurricane Creek, the train of ridges is about 11 mi (17 km) long. The train can be divided into three parts: the Thrill Hill area, the Hickory Ridge area, and the anastomosing fan. Thrill Hill appears to be an amalgamation of several ridge forms, generally peaked and steep sided. The Vandalia Geologic Area (Figure 7.2) was designated by the ISGS and the Illinois Department of Natural Resources as representative of the "Ridged Drift" for the purposes of landscape preservation in the state (Department of Landscape Architecture 1978). The impressive view is now enjoyed by subdivision residents. South of Vandalia, the system transitions to Hickory Ridge, which is a flat-topped, narrow ridge with steep flanks. With summits of approximately 650 ft (200 m) asl elevation, the ridge stands about 125 ft (38 m) in relief over the surrounding plain. The southwestern third of the system is an anastomosing, fan-shaped network standing 25 to 35 ft (8 to 11 m) in relief over the surrounding plain (Figure 7.2). This network terminates gradually to about the edge of the Hurricane Creek valley. Within and around the fan-shaped network of thread-like ridges are several flat-bottomed and mainly circular basins (e.g., Stop 9). West of the main ridge belt, isolated hills or knobs and small subparallel ridges are dispersed across the landscape. In overall shape, the train of ridges appears to compose a continuous feature but is interrupted by several saddles. Sand and gravel are exploited from several pits along the train, some of which we will visit (Stops 7 and 8). Another train of ridges, here referred to as the Kaskaskia Bluff ridges (Figure 7.2), rises from the area immediately west of the Kaskaskia Valley wall and extends northward, eventually amalgamating with the Thrill Hill complex. They are peaked rather than flat topped, with higher summits in some areas than Hickory Ridge.

Landform–Sediment Assemblages

The sediment thickness of the plain west of the Vandalia ridge system is 40 to 80 ft (12 to 24 m), based on limited data, and it thickens toward the Kaskaskia Valley as the buried bedrock surface deepens. The bedrock lithology is predominantly shale. Pre-Illinois Episode till, distinguished from overlying deposits by the occurrence of the Yarmouth Geosol, was described from an outcrop on the western valley wall at the Vandalia Bridge Section (Jacobs and Lineback 1969). However, the pre-Illinois Episode till has not been readily identifiable within the area of Hickory Ridge, probably because of being eroded. Two till members (Smithboro, Vandalia) of the Illinois Episode Glasford Formation have previously been described across the map area (Jacobs and Lineback 1969). These till units have a total thickness of about 30 ft (9 m) where described and are loam to clay loam diamicton that are distinguished mainly by clay mineralogy and texture. Overlying the Glasford till, sand with gravel deposits typically 15 ft (5 m) thick are ubiquitous. These outwash deposits are classified as the Pearl Formation. The upper portion is typically strongly weathered by the Sangamon Geosol. The weathered sand, locally known as "red dog," whose uses include baseball infields, locally has accumulated sufficient alluvial clay to be described as pedogenic diamicton (typically clay loam texture in a Sangamon Bt horizon).

In some geotechnical borings along Interstate 70 west of Vandalia (Figure D2), 3 to 10 ft (1 to 3 m) of silty clay, possibly organic, was described, presumably within or below the Sangamon Geosol. These lacustrine to alluvial deposits occur at least locally across the plain west of the Vandalia ridge system (Figure 7.2). Jacobs and Lineback (1969) mapped lacustrine sediments in a flat-bottomed basin between Hickory Ridge in the Vandalia ridge system and the Kaskaskia Bluff Ridges (Figure 7.2). This basin is in a reasonable geomorphic position to be an ice-marginal lake. Furthermore, the upper reaches of Raccoon Creek, which cuts the saddle between Hickory Ridge and the ridges in the ice-contact fan, are geomorphically compelling as a spillway for the drainage of that lake. Larson concluded, in his electrical resistivity study (Teed 2000, Zhu and Baker 1995; see Stop 7), that below a surficial cover of fine-grained deposits, the basin is underlain by 10 to 130 ft (3 to 40 m) of sand, likely fluvial in origin.

Kettle lake deposits also dot the landscape, particularly entwined in the anatomizing threads of the southwestern train. When geologists investigated the area in the early 20th century (e.g., P. MacClintock, ISGS field notes, 1927), several of the basins were still closed and undrained. Most basins have since been drained for agriculture. Because of the excellent preservation in sub-water-table unoxidized conditions, lacustrine deposits in the kettle basins contain unusual and important records of the fauna and flora from the Wisconsin, Sangamon, and late Illinois Episodes (Zhu and Baker 1995; Teed 2000; see Stops 8 and 9).

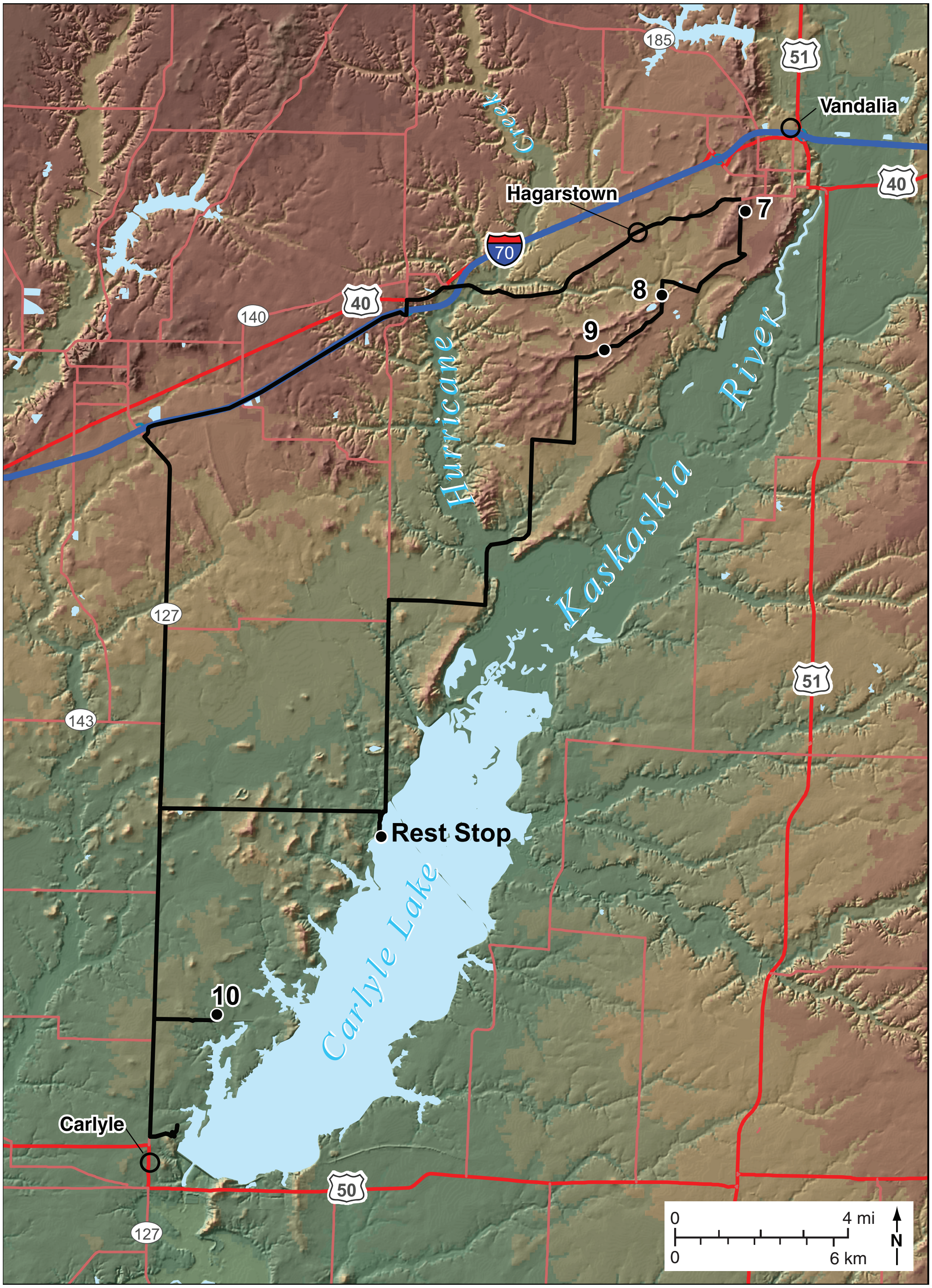


Figure D2 Location map of the planned trip route for Day 2 of the field trip. Colors are from a 30-m statewide digital elevation map (Luman et al. 2003). Higher elevations are in orange-red [to >600 ft (182.9 m)] and lower elevations are in green [to <425 ft (<129.5 m) southwest of Carlyle Lake].

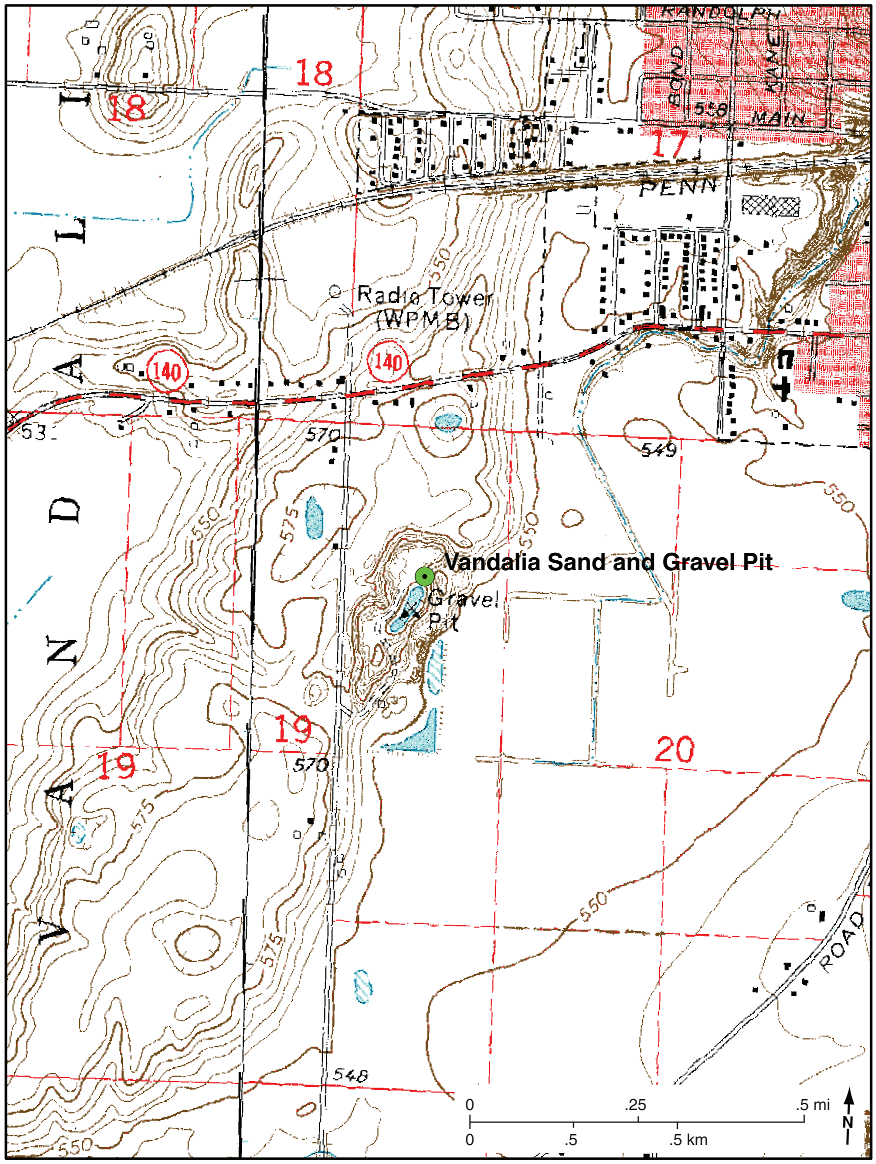


Figure 7.1 Location map for the Vandalia Sand and Gravel pit, Fayette County, Illinois (Stop 7). The topographic base map is from portions of the Hagarstown and Vandalia 7.5-minute quadrangles. Topographic map courtesy of the U.S. Geological Survey.

Ridge deposits across southwestern Illinois vary from being dominated by sand and gravel to diamicton. Aspects of this regional variability were recognized by early researchers, including Leverett (1899) and MacClintock (1929). Jacobs and Lineback (1969) described the Vandalia area ridges as composed of well-sorted gravel or sand that grades into gravelly till between the ridges. Recent investigations have shown that not all the ridges are dominated by sorted sand and gravel, and that sorted sand and gravel occur between ridges on the surrounding plain (T. Larson, Stop 7; Stiff 1996; Grimley and Phillips 2011).

Jacobs and Lineback (1969) distinguished four sediment–landform associations within the "Hagarstown beds" in their Vandalia area study: (1) gravelly till–drift plains, (2) poorly sorted gravel in drift plains near ridges, (3) well-sorted gravel in elongate ridges, and (4) sand in drift plains near Mulberry Grove. These units have graded facies relationships, and their distribution was interpreted as representing a drainage system on top of stagnant glacial ice (Jacobs and Lineback 1969). Deposition of sediments in ice-walled and englacial channels eventually led to the presence of elongate

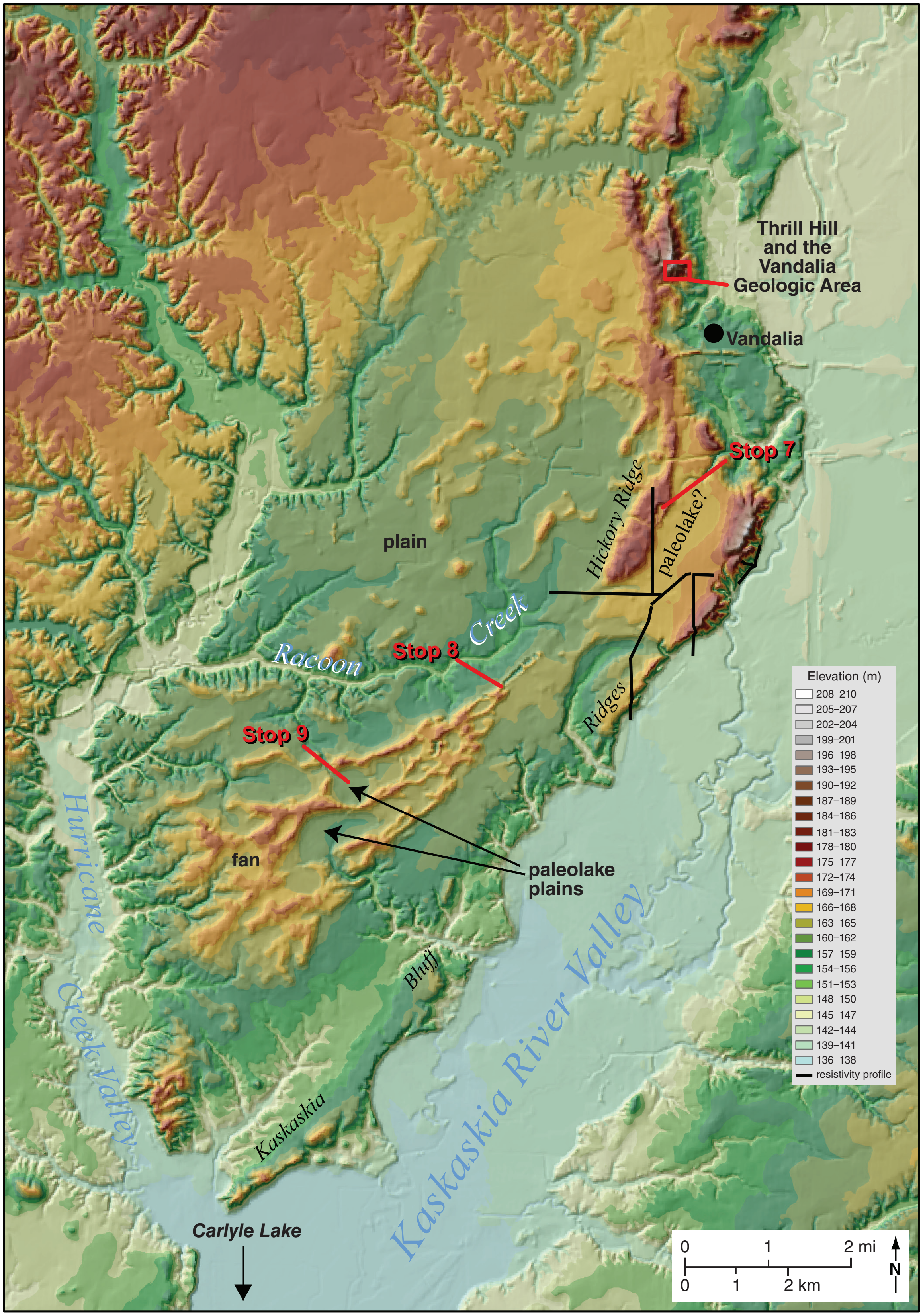


Figure 7.2 Digital elevation map of the Vandalia ridge system, including the esker and terminal fan (Stops 7, 8, and 9 are labeled). Electrical resistivity lines from Figure 7.5 are shown in black in the Hickory Ridge area.

ridges and mounds, following postdepositional melting and topographic inversion. The presence of elongate ridges, along with intact cross-bedding, was interpreted by Jacobs and Lineback (1969) to indicate widespread glacier stagnation. Mean cross-bed dip vectors indicate drainage toward the southwest in parallel with the Kaskaskia bedrock valley. All uplands are blanketed by a surficial cover of 3 to 9 ft (1 to 3 m) of Wisconsin Episode loess, containing the modern soil profile. This loess cover is thick enough to preserve, nearly intact, the Sangamon Geosol developed into the upper part of the Illinois Episode glacial ridge deposits. The last glacial loess in this area provides just enough separation of the Sangamon Geosol and modern soil sola so that little overlap is found in primary horizonation and they are thus clearly distinguished in most areas.

STOP 7: Vandalia Sand and Gravel Pit

Andrew C. Phillips, David A. Grimley, Timothy H. Larson, and Hong Wang

Historical Exposure

The Vandalia Sand and Gravel pit (Vandalia Sand and Gravel, Inc.) is the location of Jacobs and Lineback's (1969) Hickory Ridge Section. The area of their study probably sits just south of where the scale and office are located (currently an open area with slightly vegetated highwalls). The deposits we can view today in the eastern part of the pit (Figure 7.1) appear to be rather similar in character, but perhaps slightly sandier and less gravelly (Figure 7.3).

The Hickory Ridge Section was identified as the type locality for the Hagarstown Member of the Glasford Formation (Jacobs and Lineback 1969; Willman and Frye 1970, p. 58), later reclassified within the Pearl Formation (Killey and Lineback 1983). Both Jacobs and Lineback (1969) and Willman and Frye (1970) noted the gravel pit exposure as being located at SW, NW, Section 30, T 6 N, R 1 E. However, the inset map by Jacobs and Lineback (1969, p. 4) shows the site 1 mi (1.6 km) northeast in Section 20, where the Vandalia gravel pit is currently located and has been operating for several decades. No potential outcroppings are found in the northwest quarter of Section 30, so we feel confident this location was a typographic error and should have been Section 20.

According to the owner, Michael Themig, and as generally confirmed by aerial photography, excavation of sand began in the 1930s as a dryland operation in the southeastern end of the Vandalia Sand and Gravel pit (Figure 7.1). The operators expanded northward in 1970, just after Jacobs and Lineback (1969) described the type section, and they may well have described the northern face that is visible in aerial photographs from 1973. Much of the current demand for the sand is for concrete production.

Below is a reformatted description of the Hagarstown Member type section (modified from Jacobs and Lineback 1969):

Peoria and Roxana Silts (Wisconsin Episode)
0 to 3 ft [0 to 0.9 m]
Silt and loess

Hagarstown beds (Illinois Episode)
3 to 13 ft [0.9 to 4.0 m]
Gravel, silty and clayey, brownish red, poorly bedded and cross-bedded, noncalcareous; B-zone of Sangamon Soil in upper 6.6 ft [2.0 m] of unit

13 to 33 ft [4.0 to 10 m]
Gravel and sand, well sorted, very light brown, cross-bedded, largely uncemented, calcareous; intermixed in thick lenses, base of section

Current Exposure (2010–2011)

In 2011, the pit was excavated into the east flank of a wide [maximum width of ~0.5 mi (0.8 km)], elongate [~1.9 mi (3.1 km)], flat-topped ridge segment that rises about 50 ft (15 m) above the plain to the west, but only about 30 ft (9 m) above the basin to the east (Figures 7.1 and 7.2). The ridge top includes several flat-bottomed depressions that are consistent with a kettle origin. Soils mapped on top of the ridge indicate about 5 ft (1.5 m) of loess over paleosol developed in sand (Soil Survey Staff 2011).

Here we will have an opportunity to examine the edge of a large sand-dominated ridge (Figure 7.3). We will have access to the pit wall on the southern end of the dredge area. The overlying loess is sometimes visible below this cover [~30 ft (~9 m)] of spoil or fill from previous mining. The lower part of the Sangamon Geosol, developed in sand, and the relatively unweathered (but oxidized) sand below are more accessible (Figure 7.4a). We will find intercalated gravel and sand beds that generally fine upward within multiple sequences or pulses of deposition (Figures 7.4b,c). We will look for sedimentological clues to interpret flow dynamics and discuss englacial versus subglacial origins. One difference between the present section and the historical Hickory Ridge Section (Jacobs and Lineback 1969) is that the latter section was described as predominantly gravel, implying less sand than we observe today. The Hickory Ridge Section was possibly located more near the center of the ridge and thus in the area of higher flow velocities. Dredging below the water table began in the northern part of the pit in about 1990. The dredge typically excavates from depths of 35 to 40 ft (11 to 12 m) below the water level. The dredge reached as deep as 53 ft (16 m) in the northeast corner. "Blue clay," which could be till, lake sediment, or possibly weathered shale (see T. Larson; Figure 7.5), occurs below this level.

A brief description was made at the section viewable in fall 2010. About 30 ft (9 m) of mine spoil caps the natural sequence of deposits described below:

Original ground surface elevation (*top of described section*) *approximately 575 ft* [*175.3 m*]:

0–6 ft [0–1.8 m]:
Peoria and Roxana Silts [loess], silt loam, yellow brown to brown, contains modern soil solum in upper part; partially truncated by anthropogenic activity.

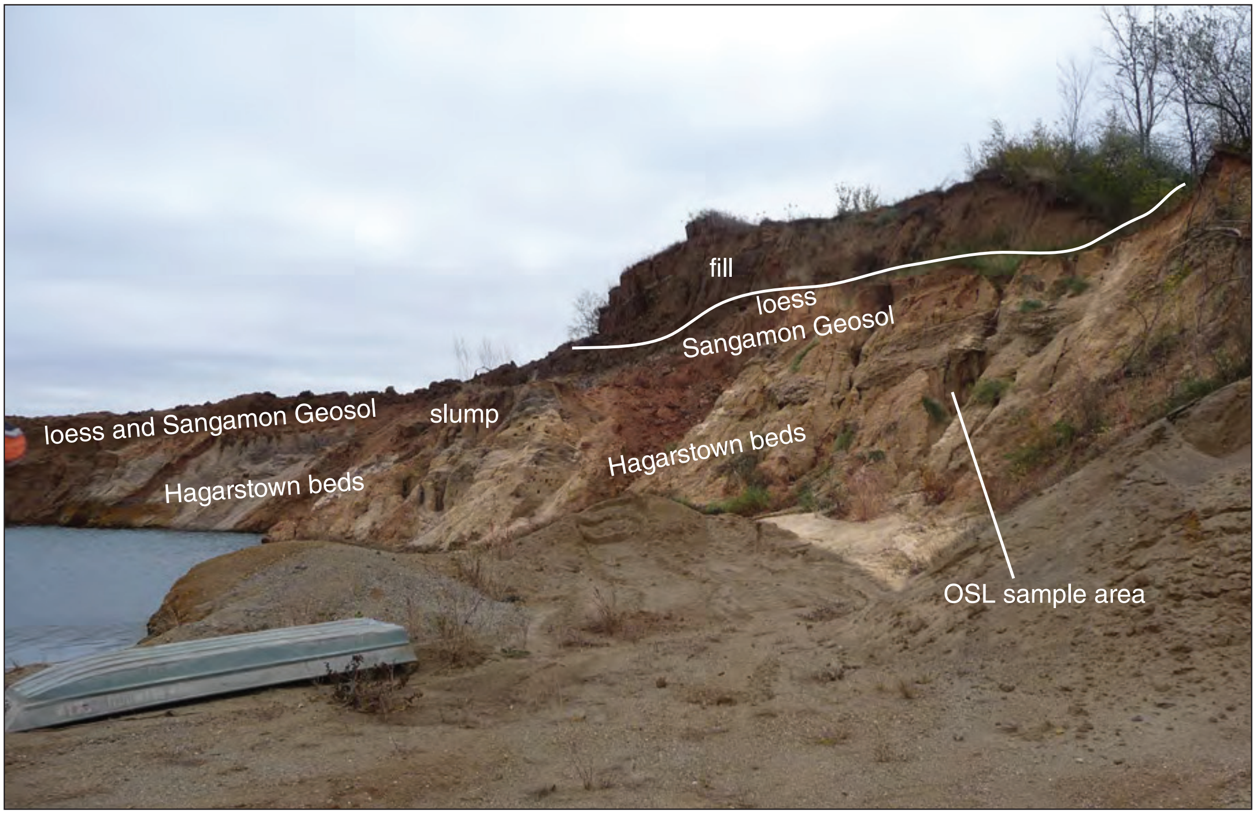


Figure 7.3 View of the active wall at the Vandalia Sand and Gravel pit. Note the reddish brown Sangamon Geosol in the upper portion of interbedded sand and gravelly sand of the Hagarstown Member, Pearl Formation. Beds are 3.3 to 6.6 ft (1 to 2 m) thick, and several can be traced across the outcrop. Some of the bed dip may be depositional, but slumping after undermining also occurs. The several-foot-long rescue punt provides a scale. OSL, optically stimulated luminescence.

6–12 ft [1.8–3.7 m]:
Sangamon Geosol developed in Hagarstown Member, Pearl Formation [ice-contact]; gravelly clay loam diamicton [pedogenic], reddish brown; Bt and BC horizons of Sangamon Geosol, included weathered beds of fine and medium sand.

12–35 ft [3.7–11 m] (base at lake level):
Hagarstown Member, Pearl Formation [ice-contact]; alternating beds of sandy gravel and fine to medium sand; the gravelly beds are 1 to 5 ft [0.3 to 1.5 m] thick; the fine–medium sand beds are 2 to 8 ft [0.6 to 2.4 m] thick, well-sorted, yellow-brown (10YR 5/4–6/4), low-angle to planar cross-bedded [Figure 7.4], and include a few thin layers that contain small blocky coal fragments (<5 cm); sampled for OSL at approximately 20 ft (6.1 m; VSG-2) and 29 ft depth (8.8 m; VSG-1) in areas of well-sorted fine or medium sand [see photos of sand in Figure 7.4d]; unit measured base is at the lake level.

OSL sample (*VSG-2*) *at ~20 ft* [*~6.1 m*] *depth from well-sorted medium sand, light yellowish brown* (*10YR 6.4*)*; about 5 in.* [*13 cm*] *above a fine sand bed and 18 in.* [*46 cm*] *below a sandy gravel bed.* [*Easting = 316333 m; Northing = 4313657 m; UTM16, NAD83*]

OSL sample (*VSG-1*) *at ~29 ft* [*8.8 m*] *depth in fine sand to medium sand, cross-bedded, well sorted, yellow brown* (*10YR 5.5/4*)*, a layer of blocky coal fragments up to 5 cm* [*2 in.*] *occurs within adjacent sand beds* (*not in layer sampled*)*. Laterally about 75 ft* [*23 m*] *east of VSG-2.* [*Easting = 316357 m; Northing = 4313659 m; UTM16, NAD83*]

35–75 ft [11–23 m] (all below lake):
Hagarstown Member, Pearl Formation [ice-contact]; sand and gravel mined up to 40 ft [12 m] below lake level from dredging (according to operator Mike); material excavated is mostly sand with about 15% gravel; hit “blue clay” at the bottom, which could be either till [Glasford Formation], lake sediment [Petersburg Silt], or weathered shale.

Figure 7.4 (a) Sand and gravel deposits of the Hagarstown Member, Pearl Formation. Of note at the top of the gravel is a protruding cobble, possibly emplaced by rafting or sliding and buried by the low-angle, thin-bedded fine to medium sand. (b) Poorly sorted gravel with rounded clasts grades upward to gravelly sand. The base of the gravel is conformable with the underlying fine sand; this contact is traceable for ~32.8 ft (~10 m) across the outcrop. Crude internal bedding within the gravel, highlighted by small coal clasts near the middle-top of the photo. (c) Dipping sequence of climbing ripple drift grading upward to low-angle to planar cross-bedding in fine sand. (d) Location of optically stimulated luminescence sample VSG-2 collected from fine sand near the section base.

Luminescence Ages

Optically stimulated luminescence ages were measured from Hagarstown Member sand samples VSG-1 and VSG-2 at 20-ft (6-m) and 29-ft (9-m) depths, respectively, below the original ground surface. The ages are 177 ± 15 ka (ISGS-71) for VSG-1 and 163 ± 15 ka (ISGS-72) for VSG-2, although these samples are likely very close in age. The OSL ages (Table 7.1) for Illinois Episode deposits are not precise, and OSL ages beyond 100 ka can be problematic (Wintle 2008; Lai 2010), but they generally confirm that deposition and ridge formation occurred during OIS 6. From global records of ice volume within OIS 6, which indicate more extensive ice volumes within the latter half of OIS 6 (Martinson et al. 1987), we favor the younger age range for the Illinois Episode glacial deposits (i.e., closer to 150 ka than to 180 ka). Optically stimulated luminescence ages were also measured from similar glacial ridge-forming sandy deposits (Hagarstown Member) at the Munie sand and gravel pit, about 30 mi (48 km) to the southwest in the Grantfork Quadrangle within Madison County (Grimley and Phillips 2005). The OSL ages from the ice-contact sand deposits at the Munie pit were determined as 140 ± 11 ka (ISGS-73) and 125 ± 10 ka (ISGS-74) for samples MGP-1 and MGP-2, respectively, sampled 7 in. (20 cm) apart laterally. The more reasonable MGP-1 mean age is within late OIS

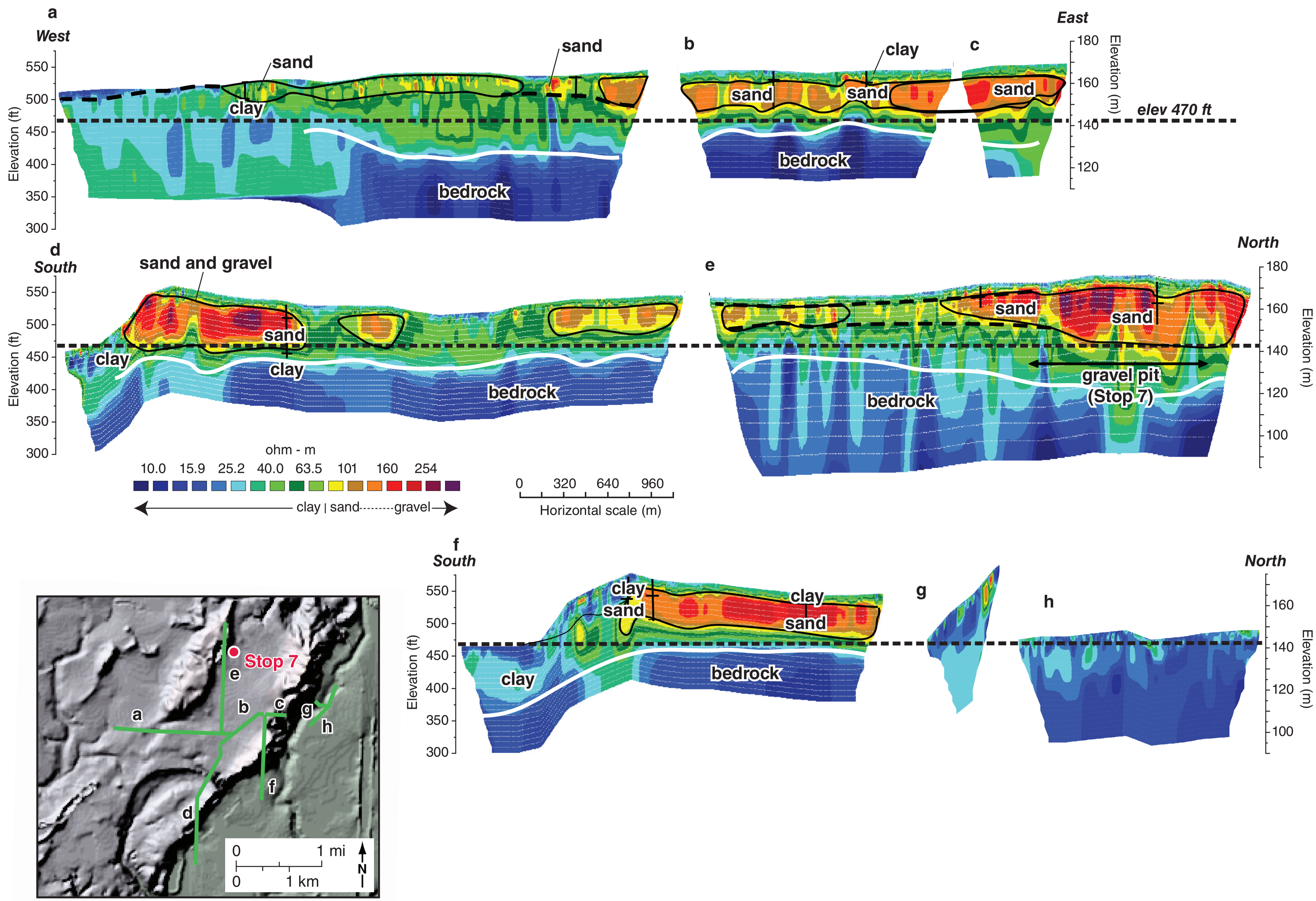


Figure 7.5 Resistivity profiles south of Vandalia, Illinois. High-resistivity areas (generally coarse grained) have more reddish colors. The color legend differs from those used in Figures 2.5 and 5.2. The dashed line at 470 ft (142 m) is approximately the elevation of the Kaskaskia River. Vertical lines on profiles show the locations of water-well borings with material descriptions that were used to confirm the electrical resistivity interpretations. Distances and elevations are in meters. Bedrock surface is indicated by the solid white line.

Table 7.1 Calculations of optically stimulated luminescence (OSL) ages at the Vandalia Sand and Gravel pit (VSG) and Munie Sand and Gravel pit (MGP)[1]

ISGS lab no.	Sample no.	Depth (m)	U (Bq/kg)	Th (Bq/kg)	^{40}K (Bq/kg)	Dose rate (Gy/ka)	H_2O (%)	Equivalent dose (Gy)	Aliquots (no.)	OSL age (ka ± 1σ)
ISGS 071	VSG-1	8.80	8.82 ± 0.4	1.14 ± 0.1	237 ± 9.5	0.998 ± 0.02	3.0	176.6 ± 11.2	21	176.99 ± 14.66
ISGS 072	VSG-2	6.10	8.50 ± 0.3	0.30 ± 0.01	203 ± 8.1	0.886 ± 0.02	4.2	141.4 ± 10.7	22	159.58 ± 14.58
ISGS 073	MGP-1	6.10	0.31 ± 0.01	2.27 ± 0.1	279 ± 11.1	0.991 ± 0.03	5.0	139.4 ± 8.6	22	140.61 ± 9.80
ISGS 074	MGP-2	6.10	9.40 ± 0.4	3.61 ± 0.1	293 ± 11.7	1.149 ± 0.03	12.0	144.2 ± 8.7	22	125.47 ± 9.80

[1]Bq/kg = concentration of radioactive material. Final OSL age = dose rate/equivalent dose. Water contents are from the field samples, so are approximate for paleo-water content values.

6, and both ages are within late OIS 6 when the error bars are considered. Conceptually, by recessional morainic relationships, the ridge dated at the Munie pit is interpreted as slightly older (but probably no more than a few centuries or millennia older) than the Vandalia area ridges. It is thus perhaps reasonable to average the sets of ages together to provide an approximate age for the entire Kaskaskia Basin ridge system. The mean and standard deviation of the four ages (177, 165, 140, and 125 ka) is 152 ± 24 ka. The mean and standard deviation for six OSL ages on ice-contact sand deposits (Hagarstown Member) in the region, including 154 and 147 ka from UNL analyses at the Keyesport site (Webb et al. 2012), field trip Stop 1, is 151 ± 18 ka. This leads to a reasonable age estimate for the Illinois Episode glaciation of between 133 and 169 ka—clearly within the range of peak global ice volumes during the penultimate glaciation. The ages in the Kaskaskia Basin are also similar to those reported previously (McKay and Berg 2008; McKay et al. 2008; Berg et al. 2012) for Illinois Episode deposits in the middle Illinois River valley region.

Electrical Resistivity Imaging

Approximately 34,800 ft (10,600 m), or more than 6.5 mi (10 km), of continuous resistivity data were acquired on eight profiles using the dipole–dipole electrode configuration with an ABEM Terrameter 3000 and LUND imaging system (one profile with the ABEM Terrameter 4000 system), and subsequently processed with RES-2DINV software (Loke and Barker 1996). The profiles link together to form two long north–south profiles and one long west–east profile across plains and ridges south of Vandalia, one short profile along the Kaskaskia River floodplain, and one short profile ascending the bluff (Figures 7.2 and 7.5). One of the north–south profiles (Figure 7.5e) continues north along the large ridge, which has been extensively mined at the Vandalia Sand and Gravel pit (Figure 7.2). Additional geologic information along these profiles is provided from logs of water wells. A heavy dashed line across all the profiles at an elevation of 470 ft (142 m) marks the approximate level of the Kaskaskia River.

In general, the resistivity profiles are characterized by zones of elevated resistivity highlighted in yellow, orange, and red. Several areas with very high resistivity values are 30 to 50 ft (9 to 15 m) thick, so we have used a different resistivity color scale for this figure compared with those in Stops 2 and 5. Similar very high resistivity values were recorded in the ridges south of Pleasant Ridge (Stop 5), but they were more limited in extent. As expected, high-resistivity deposits dominate profile (e) near the sand and gravel pit. However, similar deposits also extend beneath the plain to the southeast in profiles (b), (c), (d), and (e). The very high resistivity values could be caused by thick, coarse sand and gravel, unsaturated sand and gravel, cemented sand and gravel, or a combination of these.

The very high resistivity deposits do not extend to the southwest in profile (a). Materials in a small rise crossed in profile (a) could be the southern continuation of the ridge crossed in profile (e). These materials have elevated resistivity (highlighted in yellow), but not to the same magnitude as beneath the main ridge. However, these moderately high resistivity values are similar to the resistivity values obtained beneath Grandview Farm north of Lebanon (Stop 2) and beneath several of the ridges south of Mascoutah (Stop 5).

A series of ridges south of Vandalia intersect the west bluff of the Kaskaskia Valley. Resistivity profiles (c), (f), and (d) were designed to investigate the deposits within these ridges. As expected, an area of very high resistivity was found beneath the southern end of profile (d). This area of very high resistivity is beneath the ridge and continues south to the face of the bluff. Preliminary interpretation of this zone of very high resistivity is a deposit of unsaturated sand and gravel. A similar area of very high resistivity beneath profile (f) extends to the north, but not beneath the ridge at the river bluff. Profile (g), although very short, is similar to profile (f). The very low resistivity deposits encountered beneath the Kaskaskia River floodplain on profiles (f), (g), (h), and, to a lesser degree, profile (f) suggest that the sand and gravel deposits are restricted to higher stratigraphic levels, with fine-grained deposits on the edge of the floodplain.

STOP 8: Central Illinois Materials Sand and Gravel Pit: Catfish Pond Paleoecology

David A. Grimley, Rebecca Teed, and Andrew C. Phillips

Overview

The Central Illinois Materials sand and gravel pit (Central Illinois Materials, Inc.), only a few kilometers southwest of Stop 7, occurs at the apex of a large ice-walled fan system (Figures 7.2 and 8.1) that discharged glacial meltwater and sediment to the southwest. The fan system is interpreted as ice-walled because of the presence of an anastomosing distributary system of raised sand ridges separated internally by large [~0.5 mi (0.8 km)] circular to oval lake plains (Figure 7.2). One such lake plain, Catfish Pond, is discussed as part of this stop, and another one, Pittsburg Basin, is the highlight of Stop 9. The ice-walled fan is interpreted to have been genetically connected with the esker or ice-walled channel system that formed Hickory Ridge (Stop 7). This meltwater system would have drained the Kaskaskia Sublobe from the Vandalia area downgradient to the southwest.

The areas of sand-rich sediment emanate from the surrounding flats and lake plains as intertwining fingers to the southwest. The mining of sand and gravel deposits clearly follows these somewhat elevated, narrow, winding hills [~300 to 1,000 ft wide (100 to 300 m wide)]. The coarsest sediment has presumably been mined from the fan apex, but a systematic study or detailed mapping has not yet been conducted. A dozen water supply test holes were drilled in these hills by the village of Mulberry Grove in 1969, but apparently, insufficient groundwater was found because the city presently obtains its water supply from Hurricane Creek valley aquifers. Only a somewhat cursory description (see below) is provided in this guidebook because it was not in our area of detailed mapping. For this field trip, we will likely stop where fresh excavations are found at the end of a narrow and elongate area of dredge mining (below what was once a raised ridge) toward the southwest end of the pit area. A brief description was made during a visit by David Grimley and Steven Brown on November 8, 2010:

0–5 ft [0–1.5 m] depth:
loess with modern soil development (Peoria and Roxana Silts)

5–10 ft [1.5–3.0 m]:
Sangamon Geosol in gravelly deposits (Hagarstown M.)

0–14 ft [3.0–4.3 m]:
gravel, with numerous igneous clasts, chert, sandstone, etc. (Hagarstown M.)

14–30 ft [4.3–9.1 m]:
fine to medium sand, stratified; limited areas of coarse sand and gravel (Hagarstown M.); base of observation at water level in lake

30–55 ft [9.1–17 m]:
plant manager says an additional 20 to 30 ft [6 to 9 m] of sand with gravel is found below the lake level (probably similar to above unit)

>approximately 55 ft [17 m]:
plant manager says that shale or clay shale was encountered [possibly bedrock or could be hard, clayey, shale-rich till]

Of practical concern, subhorizontal zones of iron-cemented sand and gravel are found just above the lake level in some parts of the pit. The hardness of these zones can make extraction difficult according to the owner, Charles Barenfanger. The iron oxide in the cementing agent is mainly goethite, based on ISGS X-ray diffraction analyses. It is suspected that the zones of iron cementation reflect mobilization of iron oxides from the upper beds and reprecipitation at or near the water table level during the Sangamon Episode and at later times, perhaps continuing up through the Holocene.

Origin of the Ridge Fan: Observations

The flat, pitted surficial expression and the sand- and gravel-rich sediments within Hickory Ridge are consistent with its interpretation as an ice-walled channel. The transition from the ice-walled channel to the fan network was possibly a transition from pipe flow to open channel flow during earlier eskerine phases, although the flat top of Hickory Ridge indicates open channel flow during at least the waning stages. The depositional environment of the fan network itself is less certain. Although stratified coarse sediments attest to the glaciofluvial origin of the fan network, the system now stands above the surrounding plain, so it must have been supported laterally by ice. Gravel pits and test holes indicate that the sand and gravel also extend 20 to 40 ft (6 to 12 m) below the surrounding plain, suggesting channelized flow. The occurrence of contemporaneous buried ice blocks is evident from kettle lakes entwined with the ridge threads. A mode of origin is suggested below.

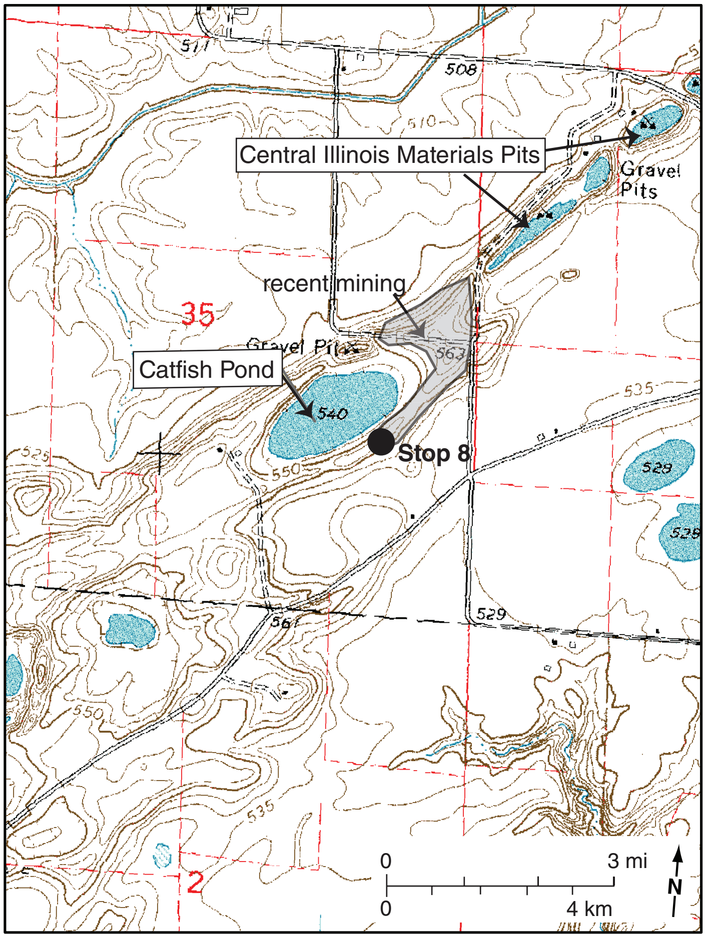


Figure 8.1 Location map for the Central Illinois Materials sand and gravel pit, Fayette County, Illinois (Stop 8). Topographic map (Hagarstown 7.5-minute quadrangle) courtesy of the U.S. Geological Survey.

Origin of the Ridge Fan: Hypothesis

A subglacial drainage system broke to the glacier surface south of what is today Raccoon Creek (Figure 7.2). A distributary fan then developed as meltwater flowed through a stagnant, decaying ice margin. The anastomosing ridge pattern developed as either multiple contemporaneous channel threads or as isolated threads that were stacked during postdepositional melting. Some or all of the channels eroded through the ice into the substrate. Blocks of ice were buried or isolated between the channel threads. These blocks melted long after the ice margin retreat, leaving basins that were later filled with redeposited loess and colluvial sediment. A modern analog of this scenario was described in Iceland by Evans and Twigg (2002). This scenario appeals to the anastomosing shape, the irregular elevations of ridge crests, and the weak correlative deltaic facies evidence.

Diatom and Pollen Record from Catfish Pond

Rebecca Teed

Immediately adjacent to current mining in the Central Illinois Materials sand and gravel pit, and still separated by an earth berm, is Catfish Pond, one of the few natural water bodies of this type in an area that was not drained for agriculture. As part of a thesis study (Teed 1999), Rebecca Teed, Herb Wright Jr., and B. Brandon Curry sampled a core in Catfish Pond to help decipher the late Wisconsin Episode and Holocene records (Figure 8.2), which are largely missing or are of poor quality in the Pittsburg Basin record (Stop 9). Previously, little was known of the Holocene pollen record in southern Illinois. Below are some key details and findings of the study:

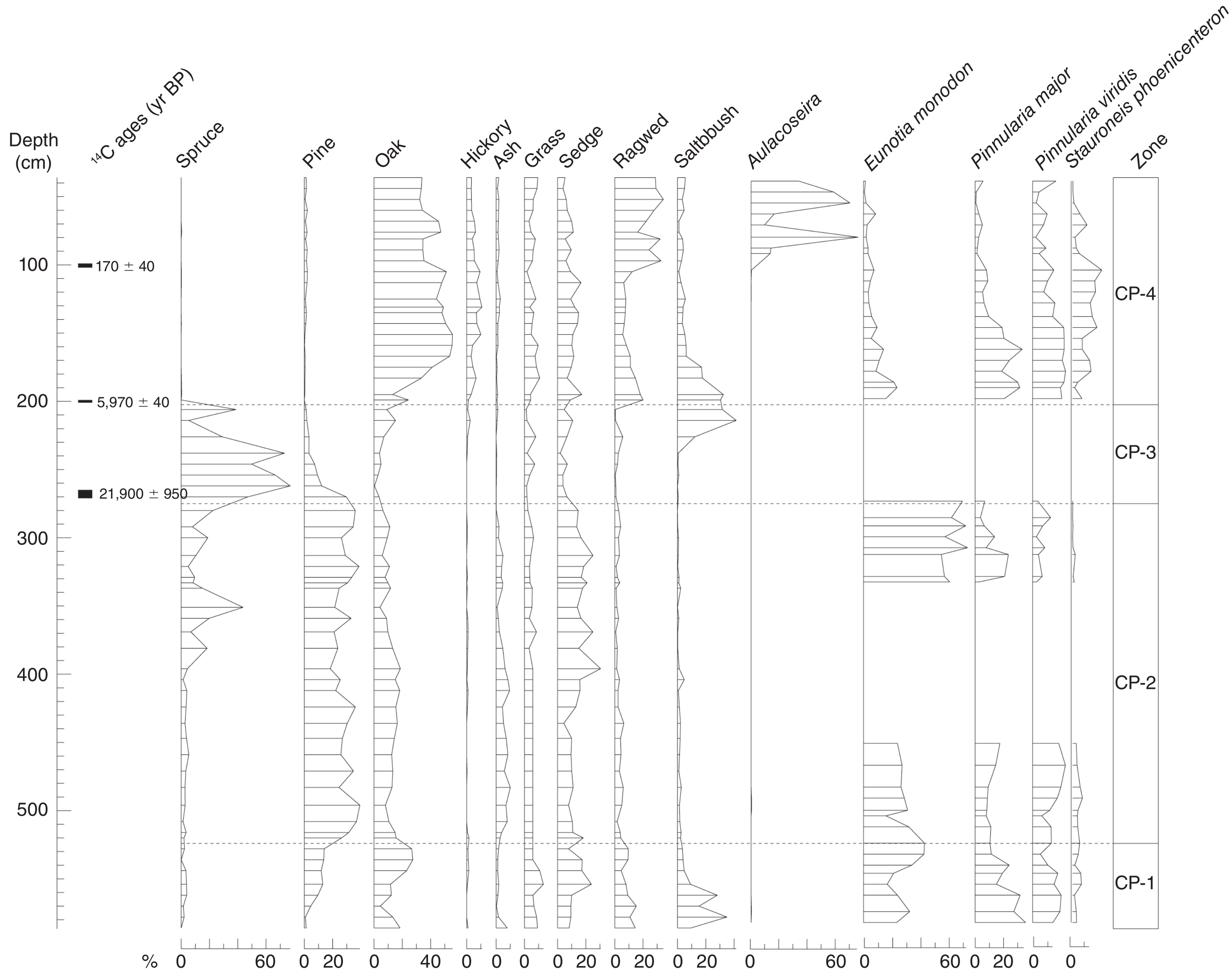


Figure 8.2 Catfish Pond (CP) pollen (modified from Teed 1999) and diatom percentages. The five Latin-named species on the right-hand side are diatoms. Zones barren of diatom counts in CP-2 and CP-3 had a few diatoms that were unidentifiable because of partial dissolution or poor preservation. Used by permission of R.A. Teed.

- *Methods:* The pollen [19 ft (5.7 m) of data] and diatom records from a core taken from Catfish Pond were analyzed in 1996. Three AMS radiocarbon dates were obtained on charcoal, wood fragments, and seeds (Teed 1999).

- Diatom assemblages at the deepest levels contained mostly *Eunotia monodon* and large (+80 µm) *Pinnularia* spp., including *P. major* and *P. viridis*. These levels were also rich in pine, oak, ash, and sedge pollen.

- Sediments dated to late Wisconsin Episode, just over 22,000 ^{14}C years ago, also contained abundant spruce pollen (Figure 8.2), indicating a cooler climate. *Eunotia monodon* accounted for more than 60% of the diatom valves counted, up from about 20% in older levels.

- The pollen assemblages indicated that between 6,000 and 200 yr BP, the landscape was a mosaic of oak–hickory forest and prairie (Figure 8.2). *Eunotia monodon* decreased to less than 10% of the diatom valves counted and the diatom flora became more diverse, including substantial amounts of *Stauroneis phoenicenteron* and *Neidium affine*.

  In the last 200 years, the sedimentation rate and ragweed pollen percentages have increased, indicating homesteading and associated intense agriculture in south-central Illinois. The diatom assemblages have become even more diverse, dominated in the upper levels by *Aulacoseira italica*.

Throughout the record, there was a close correlation between *E. monodon*, as a percentage of the diatom valves counted, and coniferous tree pollen, as a percentage of the terrestrial pollen types ($r^2 = 0.7924$, $p < .0001$). Perhaps both the diatom flora and tree taxa were responding synchronously to climate change, or perhaps there is a more direct link. For instance, coniferous tree litter tends to decompose slowly and to acidify the soil of the catchment area. This would in turn decrease pH and increase the dissolved organic carbon content of the lake, which might favor *E. monodon* over its competitors.

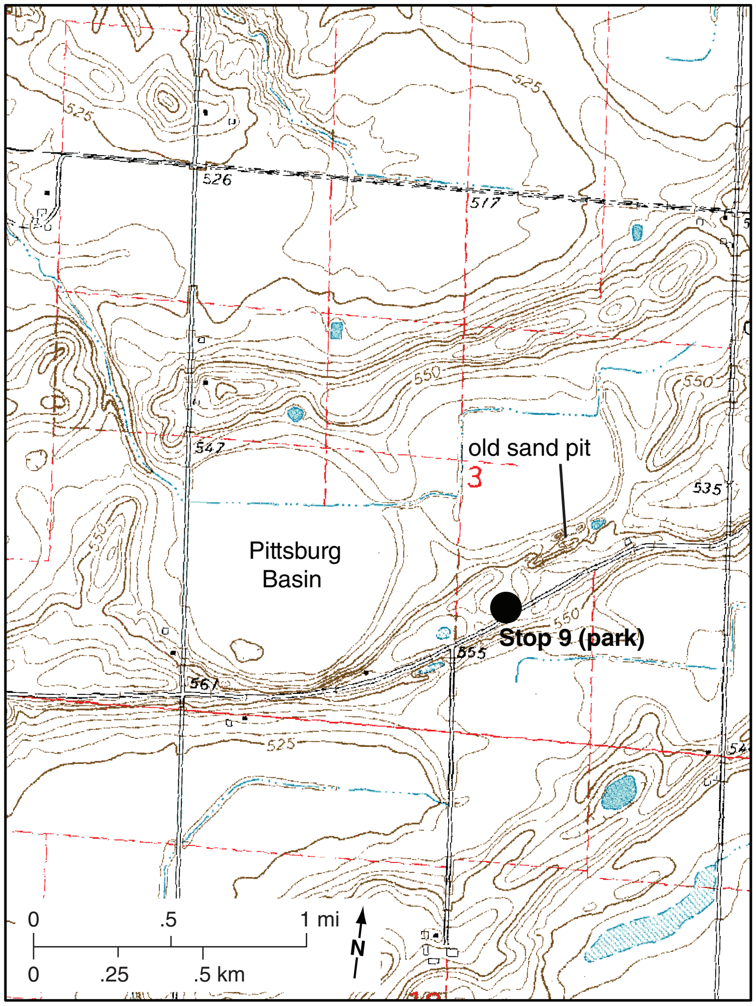


Figure 9.1 Location map for Pittsburg Basin, Fayette County, Illinois (Stop 9). Approximate locations of borings taken for the paleoenvironmental study shown in Figure 9.2. Topographic map (Hagarstown 7.5-minute quadrangle) courtesy of the U.S. Geological Survey

STOP 9: Pittsburg Basin: Paleoenvironmental History from Fossil Pollen and Ostracode Records in South-Central Illinois

B. Brandon Curry, Hong Wang, Jeffrey A. Dorale, Bonnie A.B. Blackwell, and Craig C. Lundstrom

Pittsburg Basin Record

Regionally, Pittsburg Basin (Figure 9.1) is one of five kettle basins on the Illinois Episode till plain that have yielded fossiliferous lacustrine deposits dating from the late Illinois, Sangamon, and Wisconsin Episodes. The other sites are Hopwood Farm, Raymond Basin, Catfish Pond, and Bald Knob Basin (Figure 9.2). All four basins are present within or abut constructional hills formed by Illinois Episode glacier ablation or, more commonly, by ice-walled meltwater channels.

In 1816, Pittsburg Basin was “a beautiful lake surrounded by a handsome bluff; except a small outlet at the NW end where there is an outlet leading NWterly” (Illinois Archives Land Records 1949). A lake is shown on the 1857 plat map (Illinois Archives Land Records 1949). In 1911, the Pond Lily Drainage District began to drain the lake, but a small pond was shown in the southwest part of the basin in 1915 (Curry 1995). In the 1920s, Pittsburg Basin was completely drained for agricultural purposes by blasting the overflow channel, dredging sediment in the channel, and installing field tiles.

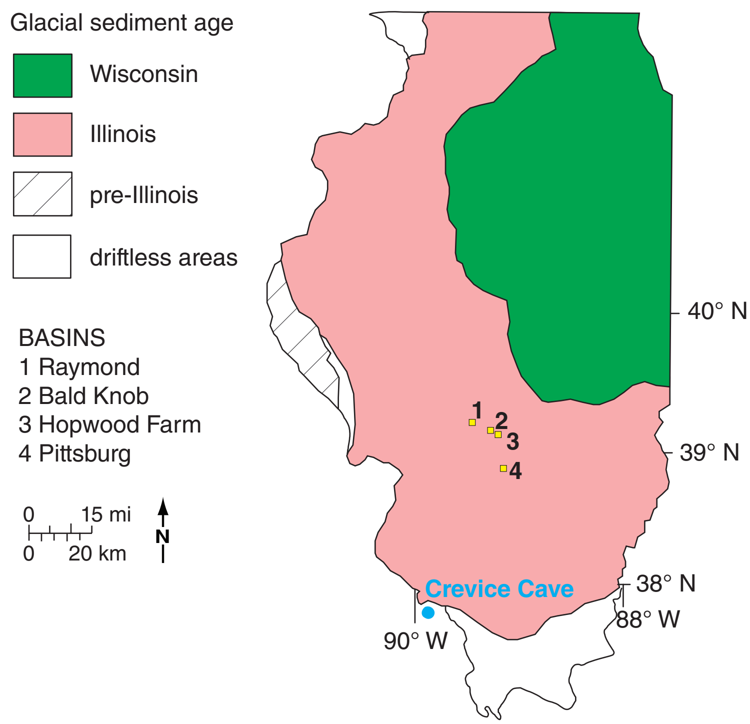


Figure 9.2 Location of Pittsburg Basin and other coring sites that have yielded fossil records in south-central Illinois spanning from the late Illinois to the early Wisconsin Episodes. Modified from Curry and Baker (2000).

Present-day Pittsburg Basin, about 0.5 mi (0.8 km) in diameter, is the westernmost and largest of a series of smaller basins outlined by sandy ridge deposits (Figures 7.2 and 9.1). The sandy ridges around the rim of Pittsburg Basin (where we will stand on the field trip) are about 35 to 50 ft (11 to 15 m) above the basin floor. This general area, about 2 mi (3 km) northwest of the modern Kaskaskia River valley, was clearly an area where glacial meltwater was focused from the melting Kaskaskia Sublobe (Webb et al. 2012). The basins and narrow ridges are seemingly arranged within a large subglacial or ice-marginal fan delta (Figure 7.2) that formed in a reentrant area of the receding ice sheet. The kettle basins, being nearly circular, most likely originated from the melting of *in situ* remnant ice blocks that were left isolated among meltwater streamflow channels as the glacier ablated in the Kaskaskia Basin (see Hypothesis for Fan Origin described in Stop 8).

Over the past few decades, the paleoenvironmental (Grüger 1972a,b; Curry 1995; Teed 1999, 2000) and enviromagnetic (Geiss and Banerjee 1997) records at Pittsburg Basin have been studied from various sediment cores across the feature. Important reference cores for studying paleoecology were all collected near the center of the basin (Curry 1995; Geiss and Banerjee 1997; Teed 2000). The basin floor was found to have a uniform elevation, based on several cores taken in 1994 (Teed 2000). One representative core from near the center of Pittsburg Basin yielded the following stratigraphy (Curry 1995):

0–1.03 m [0–3.38 ft]:
varicolored silty clay, crudely laminated in places; leached; biotubated; no fossils. [Equality Formation]

1.03–1.95 m [3.38–6.40 ft]:
finely laminated dark silty clay intercalated with gleyed (gray) silty clay loam; strong to medium platy structure imparted by laminae. [Equality Formation]

1.95–6.14 m [6.40–20.14 ft]:
black silty clay and dark; gleyed; sapric peat with gastropod shell fragments. [Equality Formation/ Berry Clay-Teneriffe Silt]

6.14–6.84 m [20.14–22.44 ft]:
olive gray to black silty clay; few gastropod shells; some zones with abundant ostracode valves. [Berry Clay-Teneriffe Silt]

Table 9.1 Ratio of ancient lake surface area to watershed area (km^2)[1]

Basin	Lake area	Watershed area	Ratio
Hopwood Farm	0.1	4.2	0.03
Bald Knob	1.4	5.5	0.4
Raymond	5.8	4.7	1.2
Pittsburg	0.6	0.4	1.5

[1]From Curry (1994).

6.84–6.97 m [22.44–22.87 ft]:
black and gray laminated silty clay; leached. [Berry Clay-Teneriffe Silt]

6.97–7.00 m [22.87–22.97 ft]:
black rock fragment.

7.00–7.20 m [22.97–23.62 ft]:
gray sand. [Pearl Formation]

A similar core, also taken near the center of Pittsburg Basin, was analyzed for pollen by Teed (2000); the results are shown in Figure 9.3. The dominance of *Picea* in pollen zone PB-1 suggests a cold and moist climate during the late Illinois Episode. An abrupt shift to deciduous tree and nonarboreal pollen marks the interglacial Sangamon Episode, represented by pollen zones PB-2, PB-3, and PB-4 (Figure 9.3). Teed (2000) interpreted all three zones as correlating to OIS 5e, but it is also possible that the zones correlate to other or all substages of OIS 5 (Curry and Baker 2000). Zone PB-5 is characterized by abundant nonarboreal pollen, and zone PB-6 is characterized by *Picea*, *Pinus*, and nonarboreal pollen. Grüger (1972b) found that the relative percentage of tree pollen to nonarboreal pollen was strongly influenced by the size of the basin, with cores from smaller basins yielding higher percentages of tree pollen. Radiocarbon ages on *Scirpus* seeds (47,900 ± 320 ^{14}C yr BP; AA21053) at the base of PB-6 verify that this pollen zone was likely deposited during the Wisconsin Episode (Teed 2000). Zone PB-7 contains pollen from deciduous forest and prairie taxa, and correlates to the Holocene Epoch (Teed 2000).

Paleoenvironmental Records from Late Pleistocene Basins in South-Central Illinois

From core records in Raymond, Bald Knob, and Pittsburg Basins (RBKP) in south-central Illinois, regional pollen zones have been developed (RPZ-1 through RPZ-7) with statistically unique pollen spectra. A fourth basin, Hopwood Farm, has a different pollen signature (Figure 9.4) that may reflect more rapid sedimentation resulting from a higher watershed-to-lake size ratio (Curry 1994; Table 9.1). In central Illinois and throughout the Midwest, the regional pollen zones contain distinctive proportions of three basic pollen types: boreal trees, deciduous trees, and herbaceous taxa (e.g., sagebrush and grass).

The paleoclimate and paleohydrology of the past 150,000 years (150 ka) have been estimated from pollen and ostracode records from the kettle basins, including Pittsburg Basin (Table 9.2). Of all the studied basins, Raymond Basin [about 35 mi (56.3 km) northwest] has the longest and most complete fossil record for paleoenvironmental and paleoclimate study, containing a record from the late Illinois Episode to the early or middle Wisconsin Episode (Zhu and Baker 1995). Zone RPZ-4 (or R4), representative of the middle Sangamon Episode, is of special interest to researchers because it contains abundant *Liquidambar* pollen (Zhu and Baker 1995), a deciduous tree most abundant in southern deciduous forests today. In the RBKP core samples, the subtropical ostracode *Heterocypris punctata* occurs in low percentages (its northernmost occurrence today is near Galveston, Texas; Curry 1995; Curry and Baker 2000). Future correlation of the RBKP and Hopwood Farm Basin records will be especially important because the fossil *Geochelone crassiscutata* (an extinct giant tortoise) was found in an outcrop at Hopwood Farm (King and Saunders 1986). This large reptile has similar-sized relatives that live today on the Galapagos and Seychelles Islands, and they cannot burrow to escape winter cold (King and Saunders 1986). Although *G. crassiscutata* and *H. punctata* likely indicate a limited extent of subfreezing temperatures during the Sangamon Episode, these species were not encountered at the same site (Curry and Baker 2000). Paleosalinity reconstructions and open hydrological conditions suggest that lake salinity was likely too low at the Hopwood Farm site to support *H. punctata.*

Another notable feature in the fossil records is the abrupt species shift in pollen and ostracodes across the boundary between RPZ-4 and RPZ-5. The lower part of RPZ-5 in Raymond Basin contains ostracodes that today live in springs and streams (e.g., *Limnocythere reticulata* and *Pelocypris tuberculatum*) and a mixture of *Picea* and *Liquidambar* pollen that suggests low lake levels and sediment mixing under a variable continental climate (Curry and Baker 2000). The lack of evidence for weathering or desiccation in sediments, with abrupt changes in the ostracode and pollen records, suggests a climatic shift across the RPZ-4/RPZ-5 boundary. However, the age of this climatic shift has been difficult to elucidate.

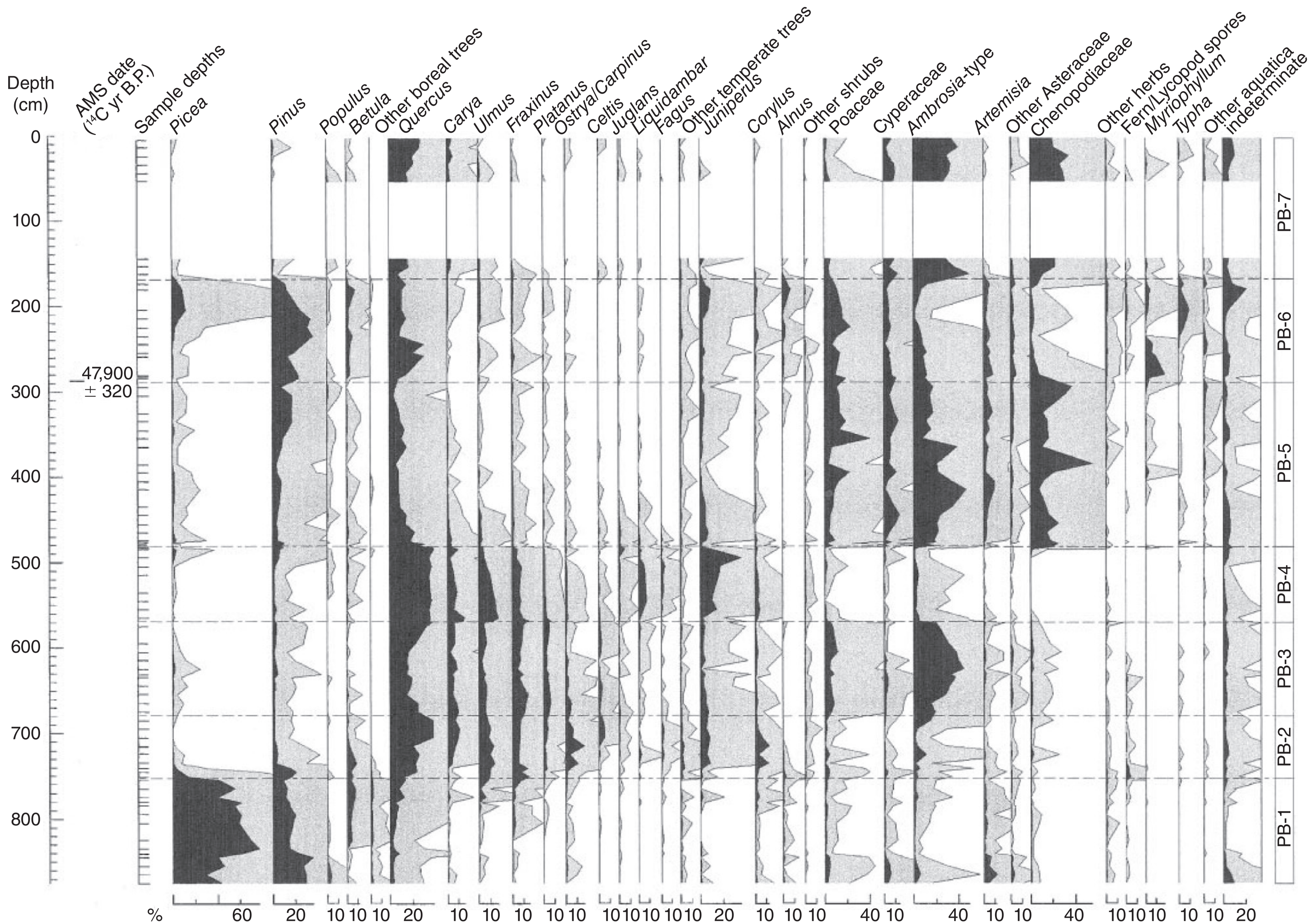


Figure 9.3 Pollen percentages from analyses of Pittsburg Basin (PB) core (from Teed 2000); location of borings in Figure 9.1. From Teed, R.A., A 130,000-year-long pollen record from Pittsburg Basin, Illinois: Quaternary Research, v. 54, p. 264–274. © 2000, with permission from Elsevier.

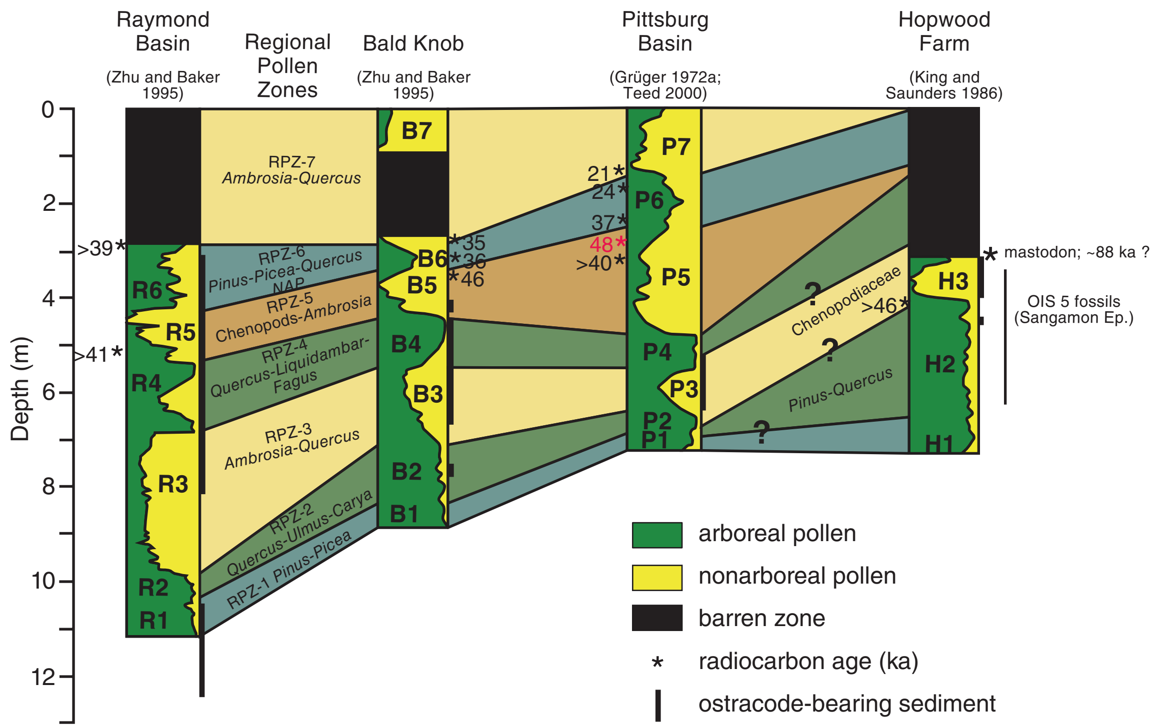


Figure 9.4 Correlation of regional pollen zones (RPZs) across sites in south-central Illinois. The stratigraphic position of the mastodon skeleton discussed in the text is shown in the Hopwood Farm section. Modified from Curry (1995) and Curry and Baker (2000).

Table 9.2 List of paleoecological and chronological studies

Basin and publication[1]	Pollen	Ostracodes	Plant macros	Verts	Chronology
Hopwood Farm					
King and Saunders (1986)	x			x	
Blackwell et al. (1990)*					x
Bald Knob					
Zhu and Baker (1995)	x				
Raymond					
Zhu and Baker (1995)	x				
Curry et al. (1997)		x			
Curry and Baker (2000)	x	x			
Curry et al. (2002)*	x				x
Dorale (2004)*					x
Curry and Dorale (2008)*					x
Curry et al. (2010b)	x	x			x
Curry and Wang (2010)*					x
Pittsburg					
Grüger (1972a,b)	x		x		
Teed (2000)	x		x		
Ito and Curry (1998)*		x			

[1]An asterisk (*) denotes an abstract.

Table 9.3 Radiocarbon ages from Pittsburg Basin, Bald Knob Basin, Raymond Basin, and Hopwood Farm[1]

Lab. no.	Core no., depth (cm)	Age (yr BP)	$\delta^{13}C$ (%)	Reference[2]
ISGS-11	PB-1972, 275–280	38,100 ± 1,000	ND[3]	1
ISGS-13	PB-1972, 317–322	>40,000	ND	1
ISGS-47	PB-1972, ca. 200	21,370 ± 810	ND	2
ISGS-53	PB-1972, ca. 220	24,200 ± 1,900	ND	2
ISGS-65	PB-1972, 255–260	24,200 ± 800	ND	3
ISGS-67	PB-1972, 263–265	34,000 ± 1,200	ND	3
ISGS-71	PB-1972, 295–297	37,200 ± 900	ND	3
ISGS-738	PB-1979, 263–265	32,590 ± 930	–24.8	4
ISGS-748	PB-1979, 269–271	39,800 ± 1,200	–25.5	4
ISGS-746	PB-1979, 275–277	>42,000	–27.4	4
ISGS-750	PB-1979, 281–283	40,030 ± 990	–26.8	4
ISGS-742	PB-1979, 289–291	41,110 ± 810	–28.3	4
ISGS-2291	BK-1, 320–326	35,480 ± 800	–27.6	5
ISGS-2292	BK-1, 393–399	35,860 ± 850	–26.89	5
ISGS-2290	BK-1, 427–430	46,200 ± 2,100	–26.4	5
ISGS-2294	BK-4, 198–209	26,090 ± 710	–26.9	5
ISGS-2293	BK-4, 287–305	>35,300	–28.9	5
Beta-53295	RB-2, 271–280	>39,410	–6.5	6
ISGS-2147	RB-2, 491-503	>41,000	–20.5	5
ISGS-2217	HF-4, outcrop	>45,900	–26.7	5
AA-472091	PB-4, 275	47,900 ± 320	ND	7

[1]PB, Pittsburg Basin; BK, Bald Knob Basin; RB, Raymond Basin; HF, Hopwood Farm.
[2]Reference: (1) Jacobs (1970); Kim (1970); Grüger (1972a,b); (2) Jacobs (1970); Coleman (1972); Grüger (1972a,b); (3) Jacobs (1970); Coleman (1972); (4) Liu et al. (1986); (5) C.-L. Liu (personal communication, 1991, 1992); (6) M. Tamers (personal communication, 1992); (7) Teed (2000).
[3]ND, not determined.

Chronology

Several problems have been encountered regarding numerical ages from Pittsburg Basin and the other three kettle basin sites. Over the past 40 years of study, four themes of age determination have been utilized at these sites: (1) bulk ^{14}C ages of detritus gyttja, (2) U-series and electron spin resonance (ESR) dating of mastodon teeth at Hopwood Farm, (3) pollen record correlations to the well-dated Crevice Cave speleothem record in Missouri (Dorale et al. 1998), and (4) OSL dating of sediments.

Radiocarbon Dating (Pittsburg Basin and Others)

Radiocarbon ages of bulk organic-rich gyttja in core samples from Pittsburg Basin provided the first chronological data (Grüger 1972a,b), including seven ages ranging from 24,200 ± 800 to >42,000 ^{14}C yr BP (Table 9.3). Most ages are in chronological order with respect to depth (Table 9.3, Figure 9.4). Teed (2000) reexamined the pollen record and chronology at Pittsburg Basin, but only one horizon at about 9 ft (3 m) depth (Figure 9.3) yielded enough material (bulrush seeds) for an AMS radiocarbon age of 47,900 ± 320 ^{14}C yr BP (AA21053; Teed 2000). Additional ages from the other basins are presented in Table 9.3.

ESR and U-Series Dating (Hopwood Farm)

At the Hopwood Farm Section, a nearly complete mastodon skeleton (*Mammut americanum*) was found above pollen-bearing sediment (asterisk in Figure 9.4), covered by 10 ft (3 m) of smectite-rich clay and about 2.3 ft (0.7 m) of Wisconsin Episode loess. Electron spin resonance analysis of a mastodon molar fragment returned ages ranging from 73 ± 9 ka to 141 ± 17 ka, depending on the uranium (U) uptake model used in the age derivation (Blackwell et al. 1990). The possible age showed large variation because ESR ages are highly dependent on U uptake history. More recently, the ESR age estimate on this fossil has been recalibrated to 88 ± 8 ka (Bonnie Blackwell, Williams College, unpublished data, 2011) using a newly acquired U-series dentine age of about 26 ka to represent late-stage U uptake in the sediment and fossil.

The enamel and dentine from the mastodon were dated through mass spectrometric analysis (by Craig Lundstrom, Department of Geology, University of Illinois at Urbana-Champaign) of U and two of its short-lived radiogenic daughters, ^{230}Th (thorium) and ^{231}Pa (protactinium). The combination of U-Th and U-Pa dating provides two independent checks on the chronology

(Cheng et al. 1998) as well as information on the extent the system has remained closed. The basic assumption is that U (a fluid mobile element) is taken up by the tooth during the life of the mastodon, but that Th and Pa (fluid immobile elements) are more stable. Thus, the amounts of ^{230}Th and ^{231}Pa relative to U provide two different numerical ages for the time that has elapsed since the mastodon was alive.

For the dentine, which is U rich and unlikely to reflect a closed system, the ^{230}Th and ^{231}Pa ages are 31 and 26 ka, respectively. For the enamel, the ^{230}Th and ^{231}Pa ages are not concordant with ages of 7 and 19 ka, respectively. These data lie to the left of "Concordia" (which shows values that would give concordant dates), indicating the enamel system was not closed and is most easily explained as late uptake of U. However, by drawing a line corresponding to the removal of the late added U, one can determine a possible date when the sample was last concordant as about 36 ka (unpublished data). The 36-ka age is the best estimate from the U-Th and U-Pa analysis according to Lundstrom. If Th or Pa was fluid mobile during diagenesis, then the estimated age by this technique might be inaccurate.

In the opinion of Blackwell et al. (1990), the original U and Pa were leached from the system, and the reported ages primarily reflect resetting of the U-Pa "clock" by younger U. Within the context of our understanding of sediment age (e.g., Curry and Baker 2000), the revised ESR age estimate by Blackwell et al. (1990) is a better fit with our previous models and with the field stratigraphy (having Wisconsin Episode loess cover), and is more in line with new (unpublished) amino acid ratio determinations on fossil gastropods from near the same layer. In sum, a best estimate of the mastodon is about 90 ka (based on ESR ages). The *Geochelone* fossils, which provide evidence for nonfreezing winters at some point in time, are likely older that about 90 ka and younger than the late Illinois Episode (~135 ka), so they would fall within the Sangamon Episode or within marine OIS 5.

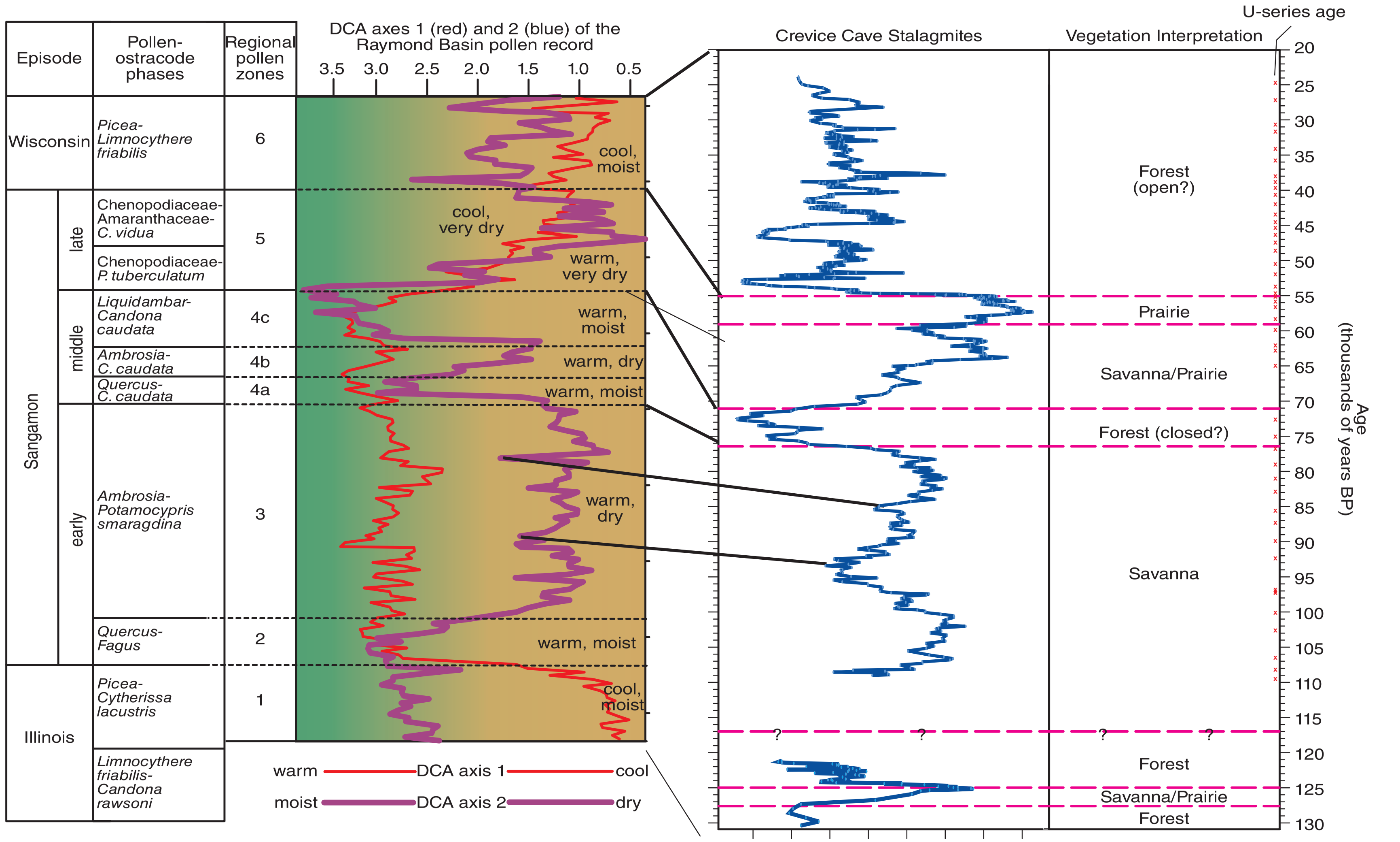
Correlation of Pollen and Speleothem Records

An alternative chronology for the regional pollen zones from RBKP was obtained by wiggle-matching the pollen record of Zhu and Baker (1995) with the δ^{13}C profiles of well-dated stalagmites collected at Crevice Cave, Missouri (Dorale et al. 1998; Curry et al. 2000). The pollen record is represented by the stratigraphically constrained detrended correspondence analysis (DCA) axis 2 (Figure 9.5). The DCA analysis is a multivariate statistical technique used to find the main factors or gradients in ecological data, such as changes in relative pollen abundance stratigraphically. The DCA axis 1 (Figure 9.5) is used as a proxy for temperature (Zhu and Baker 1995). In the DCA axis 2, larger values are associated with arboreal (tree) pollen and smaller values with nonarboreal (nontree) pollen, and it is assumed that arboreal vegetation requires more effective precipitation than does nonarboreal vegetation. The principal assumption for the correlation of the pollen-based DCA axis 2 and the δ^{13}C records is that they are proxies for paleoprecipitation. In the stable isotope record, higher δ^{13}C values are associated with vegetation that photosynthesizes by using the C_4 pathway, rather than C_3. Dorale et al. (1998) argued that the C_4 vegetation in the region is primarily prairie plants, whereas the C_3 vegetation is forest.

With a wiggle-matching program, written by Robert Henson (University of North Carolina at Greensboro), an r^2 value of 0.74 was determined by applying a linear regression of the pollen (DCA axis 2) and speleothem records. Curry et al. (2002) showed through Monte Carlo analysis that the correlation could not be random. The results of this method, utilizing correlations to dated speleothems, provided age range estimates of 77 to 70 ka and 70 to 55 ka for RPZ-4 and RPZ-5, respectively (Dorale 2004; Curry and Dorale 2008).

Luminescence Dating (Raymond Basin)

Encouraged by the infrared-stimulated luminescence results of Sanda Balescu at Pittsburg Basin, ranging from 69 ± 7 ka at the top of RPZ-4 to 126 ± 16 ka at the top of RPZ-2 (Teed 1999), a recent attempt was made to validate the speleothem-tuned chronology at Raymond Basin with OSL dating. The lacustrine sediments sampled were from a replicate of the original core, RB-2 (Figure 9.6), used for ostracode, pollen, particle-size, and clay mineral analyses (Curry 1995; Zhu and Baker 1995; Curry et al. 1997; Curry and Baker 2000). Initial samples for chronological data were sent to the Luminescence Dating Research Laboratory at the University of Illinois at Chicago (UIC). A sample of lake sediment just above Glasford Formation till (Vandalia Member) at a depth of 1,250 cm (RPZ-1) yielded ages of 64 ± 5 ka and 67 ± 5 ka. Another sample at a depth of 508 cm (RPZ-5) yielded an age of 54 ± 4 ka. The layer of silty sediment at 300 cm (RPZ-6), interpreted as redeposited Peoria Silt, yielded an age of 14 ± 1 ka (Table 9.4). The basal ages were much younger than expected, based on the widely held assumption that Glasford Formation sediments were deposited during OIS 6 (e.g., Curry and Follmer 1992; Curry and Baker 2000; Hansel and McKay 2010). To test the initial results, additional OSL ages were determined for ice-contact glaciofluvial sediments (Hagarstown Member, Pearl Formation) exposed



Stratigraphic plot of DCA axes 1 and 2 values, their general paleoenvironmental interpretations, and biostratigraphic correlations (Zhu and Baker 1995; Curry and Baker 2000)

U-series $\delta^{13}C$ record from Crevice Cave, Missouri, and vegetation (intepreted independently of the pollen record; Dorale et al. 1998; Dorale and Edwards 1999).

Figure 9.5 (Left) Detrended correspondence analysis (DCA) axis 1 (red) and DCA axis 2 (thick purple) from the Raymond Basin pollen record (Zhu and Baker 1995). The pollen-ostracode phase column is modified from Curry and Baker (2000). (Right) Carbon isotope data are from Crevice Cave, Missouri, speleothems (Dorale et al. 1998).

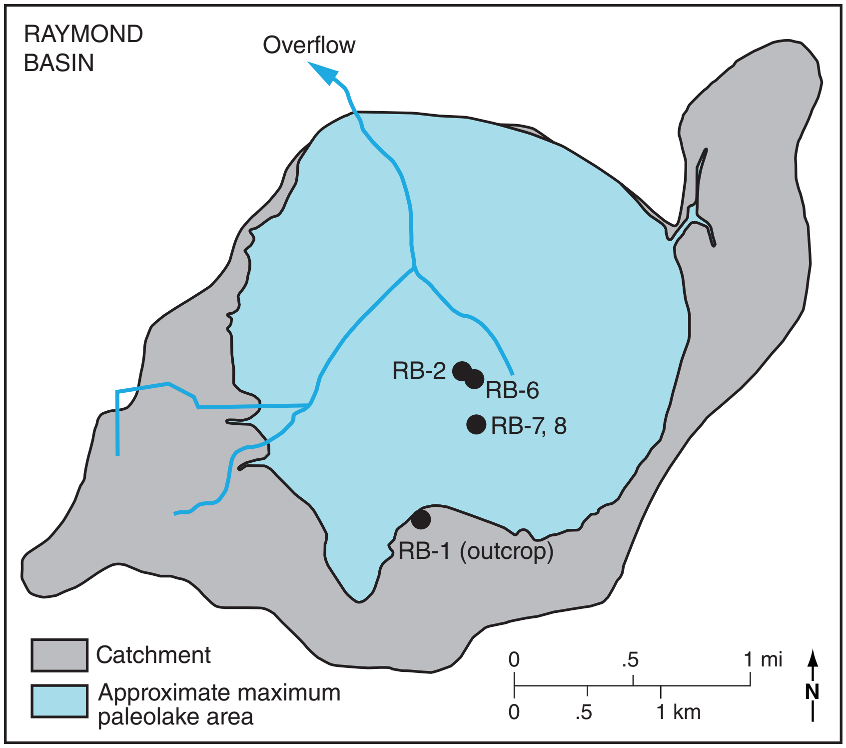


Figure 9.6 Location of cores studied in Raymond Basin (RB) and the adjacent exposure in a small sand and gravel pit. Modified from Curry and Baker (2000).

in a sand pit in a constructional Illinois Episode ridge adjacent to Raymond Basin (see Figure 9.6). Duplicate samples were submitted to two laboratories: the UIC laboratory and the OSL Laboratory at the ISGS. The UIC laboratory results (Table 9.4) were consistent with their earlier results. Optically stimulated luminescence ages from the ISGS laboratory were 138 ± 10 ka and 122 ± 11 ka, nearly twice as old. The age difference may be explained in part by the use of different methodologies. The ISGS laboratory used the single-aliquot regenerative (SAR) protocol on 90- to 250-µm quartz, and the UIC laboratory used the multiple-aliquot regenerative protocol on fine-fraction (4- to 11-µm) polyminerals, including quartz. Additionally, each laboratory calculated the dose rate differently. The UIC laboratory determined the *in situ* radiation dose from inductively coupled plasma mass spectrometry sediment analysis, with the assumption being that most radiation is derived from ^{40}K and ^{238}U. The ISGS laboratory determined the dose rate from gamma-ray spectrometry measurements of subsamples in contact with the primary OSL sample. It should also be noted that, in general, luminescence ages for sediments >100 ka can be problematic (Wintle 2008; Lai 2010).

Encouraged by the older results from the ISGS laboratory, B. Brandon Curry and Hong Wang sampled replicate cores RB-7 and RB-8 located between core RB-2 and the highest ancient shoreline at Raymond Basin (Figure 9.6). Key stratigraphic contacts were marked by layers of sand in cores RB-7 and RB-8. The sand layer between the basal, barren silty clays and organic-rich silts (RPZ-1) yielded an OSL age of 130 ± 8 ka, whereas the sand layer between the organic silts of RPZ-1 and gray clays of RPZ-2 yielded an age of 124 ± 6 ka. These ages are consistent with results of the pollen-speleothem wiggle-matching exercise, albeit by virtue of the ages being older than the oldest correlated Crevice Cave age of 109 ± 1 ka. However, the date from the sand layer that caps the fossiliferous succession was 107 ± 7 ka compared with an estimate of about 50 ka based on the pollen-speleothem calibrated ages (Figure 9.7). Two additional samples, dated by the ISGS, from the lower part of the silty lithofacies at depths of 230 and 270 cm have ages of 41 ± 3 ka and 53 ± 3 ka, respectively (Figure 9.7). These ages are not unreasonable, considering the regional records of the Roxana Silt (mainly loess) deposited between about 55 and 28 ka (Curry and Follmer 1992; Leigh 1994).

Table 9.4 Optically stimulated luminescence (OSL) ages

Lab no.[1]	Depth (cm)	U (Bq/kg)	Th (Bq/kg)	^{40}K (Bq/kg)	Dose rate (Gy/ka)	H_2O	Equivalent dose (Gy)	Aliquots (no.)	OSL age ± 1σ (ka)
Cores RB-7,8									
ISGS lab no.									
ISGS 075	230	29.7 ± 1.2	35.7 ± 1.4	512 ± 20.5	2.531 ± 0.1	15.5	103 ± 5.2	18	41 ± 3
ISGS 076	270	27.8 ± 1.1	31.7 ± 1.3	491 ± 19.7	2.353 ± 0.1	17.5	125 ± 5.8	18	53 ± 3
ISGS 059	350	12.7 ± 0.5	8.16 ± 0.3	241 ± 9.7	1.021 ± 0.0	23.5	109 ± 5.4	28	107 ± 7
ISGS 063	580	22.5 ± 0.9	19.3 ± 0.8	234 ± 9.4	1.297 ± 0.0	24.5	161 ± 4.3	18	124 ± 6
ISGS 058	700	1.3 ± 0.1	0.70 ± 0.03	215 ± 8.6	0.680 ± 0.0	19.5	88 ± 3.1	25	130 ± 8
ISGS 057	800	17.3 ± 0.7	20.3 ± 0.8	574 ± 22.9	2.100 ± 0.1	21.5	150 ± 11.7	14	71 ± 7
Core RB-6									
UIC lab no.		U (ppm)	Th (ppm)	K_2O					
2290BL	300	3.3 ± 0.1	10.5 ± 0.1	2.34 ± 0.02	3.76 ± 0.2	30 ± 5	54 ± 1.0	30	14 ± 1
2291BL	509	2.4 ± 0.1	5.4 ± 0.1	1.56 ± 0.02	2.34 ± 0.1	30 ± 5	127 ± 3.7	30	54 ± 4
2289BL	1,250	2.9 ± 0.1	10.1 ± 0.1	3.91 ± 0.04	4.50 ± 0.2	35 ± 5	302 ± 3.0	30	67 ± 5
2289BLq	1,250	2.9 ± 0.1	10.1 ± 0.1	3.91 ± 0.04	4.51 ± 0.2	35 ± 5	290 ± 1.6	30	64 ± 5
Outcrop		U (Bq/kg)	Th (Bq/kg)	^{40}K (Bq/kg)					
ISGS 021	400	12.1 ± 0.5	6.4 ± 0.3	317 ± 12.7	1.24 ± 0.0	20.0	151 ± 11	22	122 ± 11
ISGS 022	400	8.9 ± 0.5	5.8 ± 0.2	362 ± 14.5	1.23 ± 0.0	21.5	177 ± 10	22	138 ± 10
		U (ppm)	Th (ppm)	K_2O					
UICRB 1	400	0.8 ± 0.1	2.1 ± 0.1	1.22 ± 0.01	1.35 ± 0.1	15 ± 5	85 ± 9	30	63 ± 8
UICRB 2	400	0.6 ± 0.1	1.6 ± 0.1	1.02 ± 0.01	1.16 ± 0.1	15 ± 5	55 ± 7	30	47 ± 8

[1]RB, Raymond Basin; UIC, University of Illinois at Chicago.

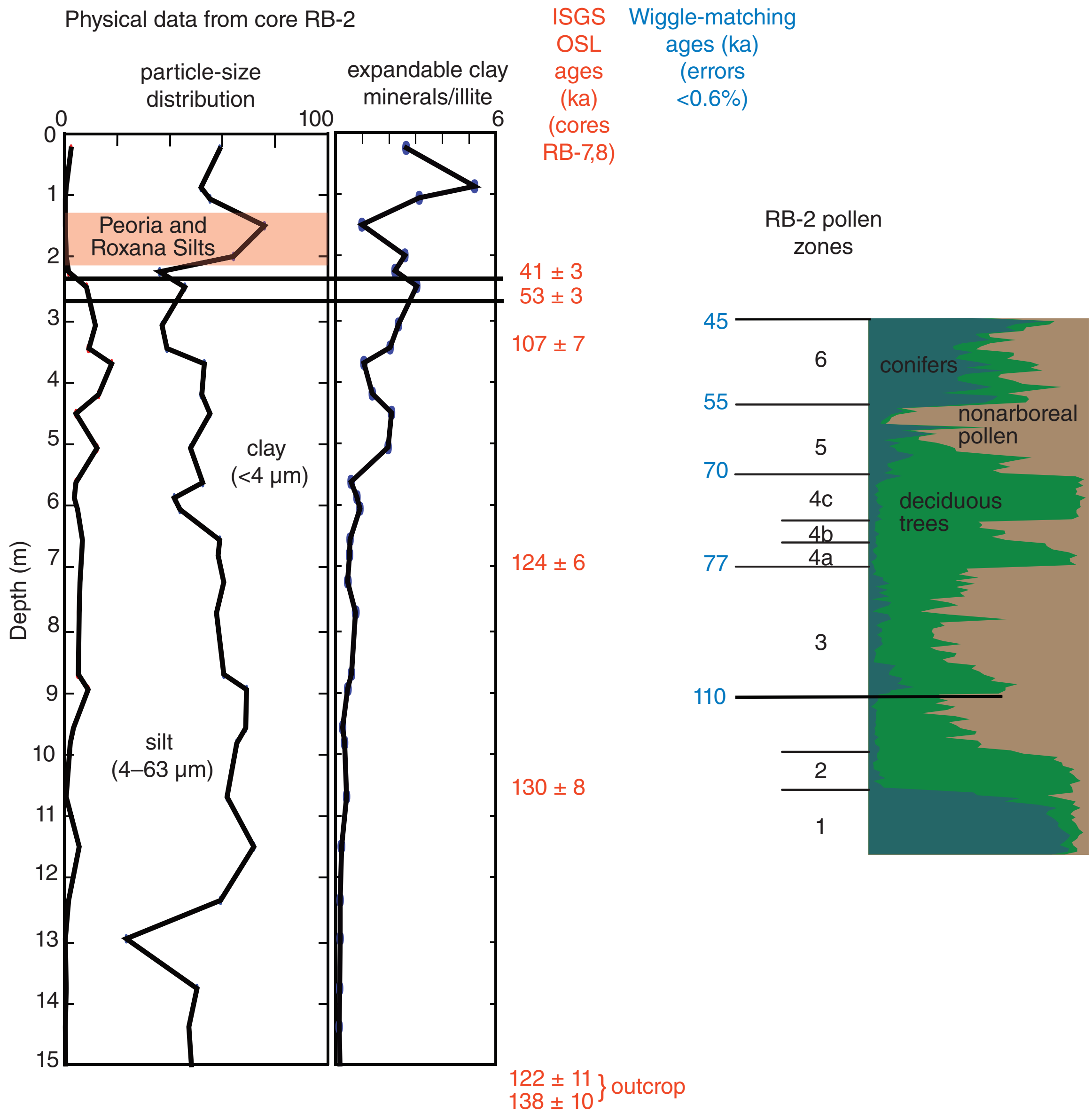


Figure 9.7 Comparison of core Raymond Basin (RB)-2 (location in Figure 9.6) compositional data, pollen zones, and estimated chronology. Two geochronological records are shown: (1) ISGS-analyzed optically stimulated luminescence (OSL) ages (in red) and (2) speleothem calibrated ages (in blue) of key biozone contacts. Modified from Curry (1995).

Summary—Where to Go from Here?

It is clear, based on the preceding discussion of the results to date, that considerable conflict exists in the data acquired from the various geochronological methods, and even from the same method performed at different laboratories. We hope that the age discrepancy will be resolved in the coming years as new approaches are taken and methods are improved to date older (pre-55 ka) materials. One method being used to help resolve the controversy at Raymond Basin is to determine meteoric beryllium (^{10}Be) profiles (cf. Graly et al. 2011), focusing on the interval above the fossiliferous succession. If the OSL ages from the ISGS laboratory are accurate, then the thin [~3-ft (~1-m)] interval above the fossiliferous succession and below the redeposited Peoria Silt should be concentrated in ^{10}Be relative to the sediment above and especially below. If the speleothem-tuned chronology is accurate, then the ^{10}Be profile will be relatively constant. In either case, the results of the Monte Carlo simulation indicate that the wiggle-matching is not random and that a fundamental pattern may exist at a smaller harmonic. We plan to refine our OSL results by determining the saturation of outer atomic shells after incremental heat treatments (cf. Stevens et al. 2011). As for other analyses, amino acid racemization analyses of the aquatic gastropods at Hopwood Farm are also being studied. Previously determined values of amino acid ratios from gastropods at this site were variable and inconclusive (McCoy et al. 1995). However, the advancement of this technique with multiple amino acid assays (Kaufman and Manley 1998; Oches and McCoy 2001) and comparisons with recently acquired amino acid data in the region (Curry et al. 1997; Grimley et al. 2001, 2009, 2010; Grimley and Oches 2015) may help resolve the age controversy for the mastodon skeleton and Hopwood Farm succession. Preliminary results with amino acid dating at Hopwood Farm (B. Curry, D. Grimley, D. Kaufman, and L. Fay, 2013–2014) confirm an OIS 5 age for the gyttja and marl layers (Sangamon Episode).

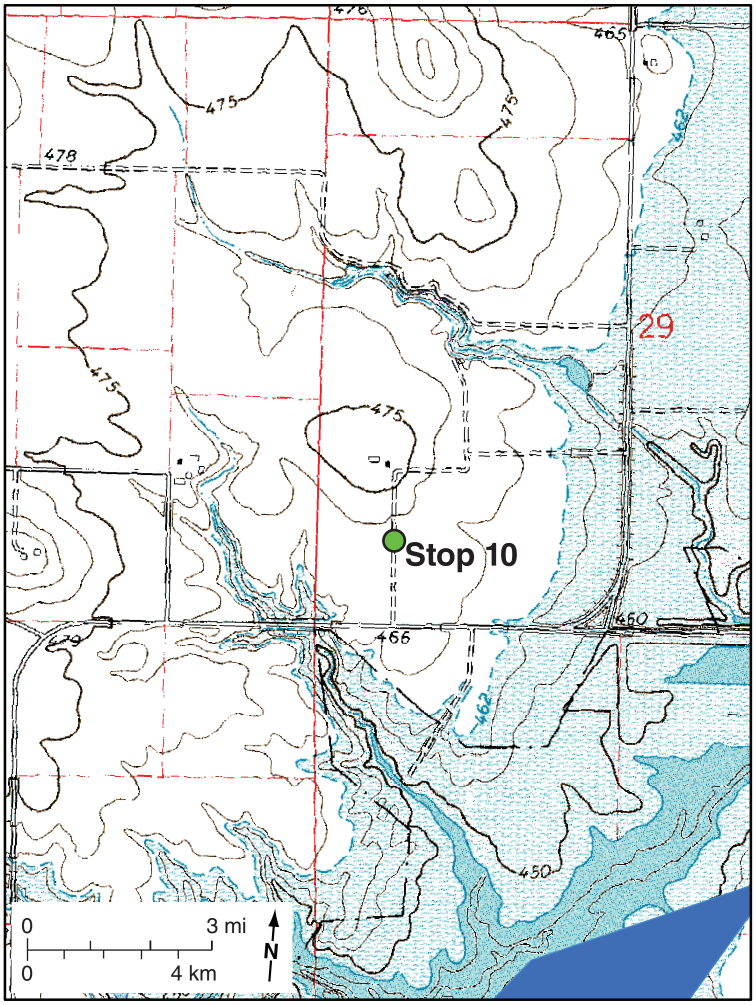


Figure 10.1 Location map for the field trip stop examining sodium-affected soils in Clinton County, Illinois (Stop 10). Topographic map (Keyesport 7.5-minute quadrangle) courtesy of the U.S. Geological Survey.

STOP 10: Sodium-Affected Soils in South-Central Illinois: Relationships with Relict Patterned Ground

Samuel J. Indorante, Michael Konen, and Erik A. Gerhard

Sodium-affected soils (SAS) are those that have been adversely affected by sodium salts, exchangeable sodium, or both (Indorante 2002; Indorante et al. 2011). They usually occur in arid, semiarid, and subhumid climates where rainfall is insufficient to leach soluble salts from soils where internal drainage is restricted. To a much lesser extent, SAS occur in humid regions with a mean annual precipitation of >39 in. (>100 cm) because of factors that restrict leaching of soluble salts from the soil (Indorante 2002). The main factors for SAS soil development are (1) sources of sodium, (2) high water tables, (3) clayey, dense subsoils, (4) impermeable underlying geologic strata, and (5) seasonal periods of high evapotranspiration.

High-pH SAS occupy approximately 946,414 acres (383,000 hectares) in south-central Illinois and are particularly common in the Kaskaskia Basin. In general, SAS soils in Illinois occur in areas where the Wisconsin Episode loess is 3 to 7 ft (1 to 2 m) thick over the leached, less permeable Sangamon Geosol, such as in the case of this field trip stop (Figure 10.1). Sodium feldspars in the parent loess are the primary source of the sodium (Wilding et al. 1963). The SAS occur in seemingly unpredictable patterns among normal acidic soils and are typically mapped in complexes with

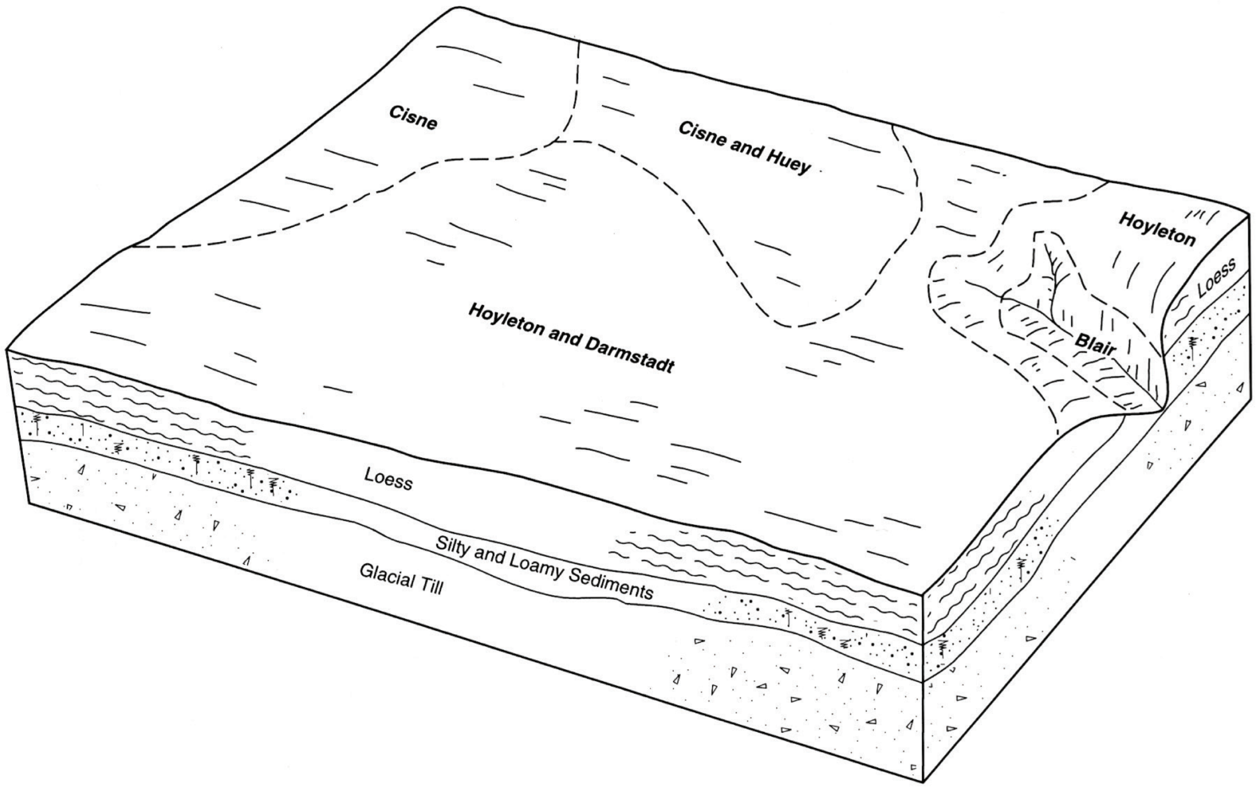


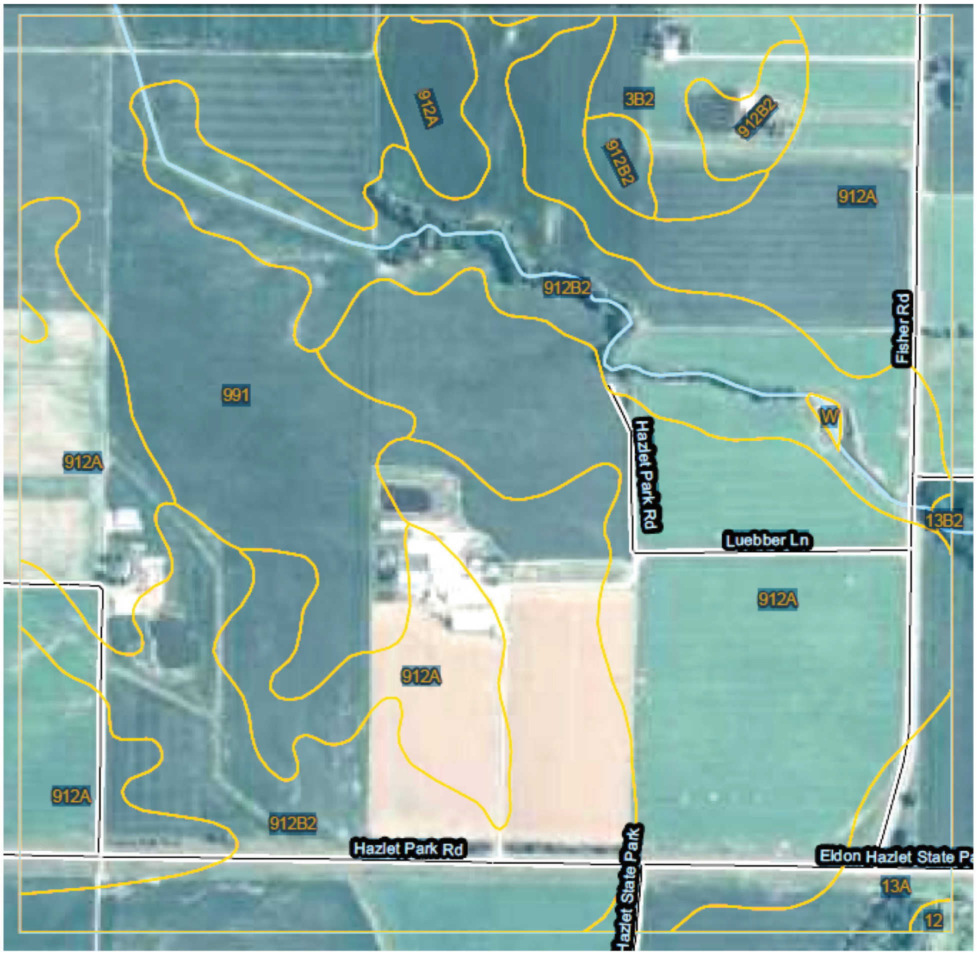
Figure 10.2 Soil landscape block diagram of the stop area. Typical pattern of soils and parent material in the Hoylton-Darmstadt-Cisne-Huey association (Hamilton 2002). Figure courtesy of the U.S. Department of Agriculture.

non-SAS (Figures 10.2 and 10.3). Most areas of SAS/non-SAS soil complexes are presently farmed. Before settlement, these areas had a mixture of prairie and forest vegetation.

The impact of SAS on Illinois agriculture is significant. The non-SAS Cisne soil (fine, smectitic, mesic Vertic Albaqualf) and the SAS Huey soil (fine-silty, mixed mesic Typic Natraqualfs) are commonly mapped in complex in south-central Illinois. Under the same climate and management factors, and under nonirrigated conditions, the Huey soils with more than 15% exchangeable sodium or with sodium adsorption ratios >10 in the subsoil had an average of 18% yield reduction for soybeans and an average of 38% yield reduction for corn when compared with the Cisne soils (Table 10.1). The yield reduction is caused primarily by the poor physical makeup of the dispersed soil resulting from the high sodium concentration.

The intricate pattern of SAS and non-SAS on level to nearly level uplands indicates differential redistribution of sodium derived from primary minerals in the loess. Previous research suggests that differential water movement and variations in evapotranspiration, both associated with current and historical soil landscape settings, were the mechanisms responsible for redistribution of sodium in solution (Wilding et al. 1963; Frazee et al. 1967).

Recent updating and digitization of Illinois soil surveys have revealed large areas of polygonal patterned ground associated with SAS (Figure 10.4), although not all SAS is associated with polygonal patterned ground. The patterned ground is interpreted to have formed as a result of Wisconsin Episode permafrost formation and degradation (Johnson 1990), particularly during the LGM (~19 to 25 ka calendar years). Most polygons are 30 to 260 ft (10 to 80 m) in diameter with 13- to 20-ft-wide (4- to 6-m-wide) borders. Polygon interiors are darker colored and slightly lower in elevation, and polygon borders are lighter colored and slightly higher in elevation. Although multiple pathways may be responsible for SAS, our working model is that permafrost-related processes led to a unique microtopography that affected local hydrology and resulted in SAS development in south-central Illinois (Indorante et al. 2011).



Clinton County, Illinois (IL027)

Map Unit Symbol	Map Unit Name	Acres in AOI	Percentage of AOI
3B2	Hoyleton silt loam, 2% to 5% slopes, eroded	12.7	2.7
12	Wynoose silt loam	0.6	0.1
13A	Bluford silt loam, 0% to 2% slopes	12.5	2.6
13B2	Bluford silt loam, 2% to 5% slopes, eroded	0.4	0.1
912A	Hoyleton-Darmstadt complex, 0% to 2% slopes, eroded	218.0	45.5
912B2	Hoyleton-Darmstadt complex, 2% to 5% slopes, eroded	153.1	31.9
991	Cisne-Huey complex	81.4	17.0
W	Water	0.5	0.1
Totals for Area of Interest (AOI)		479.3	100.0

Figure 10.3 Soil map and legend of the stop area (from Hamilton 2002; Soil Survey Staff 2015). The site is located in Clinton County, Illinois (S½, Section. 29, T 3 N, R 2 W) between Carlyle and Keyesport. Figure and data courtesy of the U.S. Department of Agriculture.

Table 10.1 Chemical properties of selected soils in Clinton County, Illinois[1]

Map symbol and soil name	Depth (in.)	Cation exchange capacity (mEq/100 g)	Effective cation exchange (mEq/100 g)	Soil reaction (pH)	$CaCO_3$ (%)	Salinity (mmhos/cm)	Na absorption ratio
912A: Hoyleton (non-SAS)	0–9	14–22	—	4.5–7.3	0	0	0
	9–15	9.0–17	—	4.5–6.5	0	0	0
	15–36	—	21–28	4.5–6.0	0	0	0
	36–60	9.0–21	—	5.1–7.3	0	0	0–5
Darmstadt (SAS)	0–14	7.0–20	—	5.1–7.3	0	0.0–2.0	0–5
	14–20	16–23	—	4.5–7.8	0	0.0–2.0	10–20
	20–40	16–23	—	6.6–9.0	0–5	0.0–2.0	10–25
	40–60	9.0–20	—	7.4–9.0	0–5	0.0–2.0	5–20
912B2: Hoyleton (non-SAS)	0–7	14–22	—	4.5–7.3	0	0	0
	7–12	9.0–17	—	4.5–6.5	0	0	0
	12–32	—	21–28	4.5–6.0	0	0	0
	32–60	9.0–21	—	5.1–7.3	0	0	0–5
Darmstadt (SAS)	0–6	7.0–20	—	5.1–7.3	0	0.0–2.0	0–5
	6–14	16–23	—	4.5–7.8	0	0.0–2.0	10–20
	14–27	16–23	—	6.6–9.0	0–5	0.0–2.0	10–25
	27–60	9.0–20	—	7.4–9.0	0–5	0.0–2.0	5–20
Cisne (non-SAS)	0–8	11–22	—	4.5–7.8	0	0	0
	8–15	—	9.0–17	4.5–6.0	0	0	0
	15–51	—	21–28	4.5–6.0	0	0	0
	51–60	13–19	—	5.6–7.3	0	0	0–3
Huey (SAS)	0–7	11–22	—	5.1–7.8	0	0.0–2.0	0–20
	7–12	6.0–10	—	5.1–7.8	0	0.0–2.0	0–20
	12–15	12–22	—	5.6–8.4	0–15	0.0–2.0	0–20
	15–51	15–21	—	7.4–9.0	0–25	0.0–2.0	15–40
	51–60	11–21	—	6.6–8.4	0–30	0.0–2.0	10–40

[1]From Soil Survey Staff (2015). Table courtesy of the U.S. Department of Agriculture.

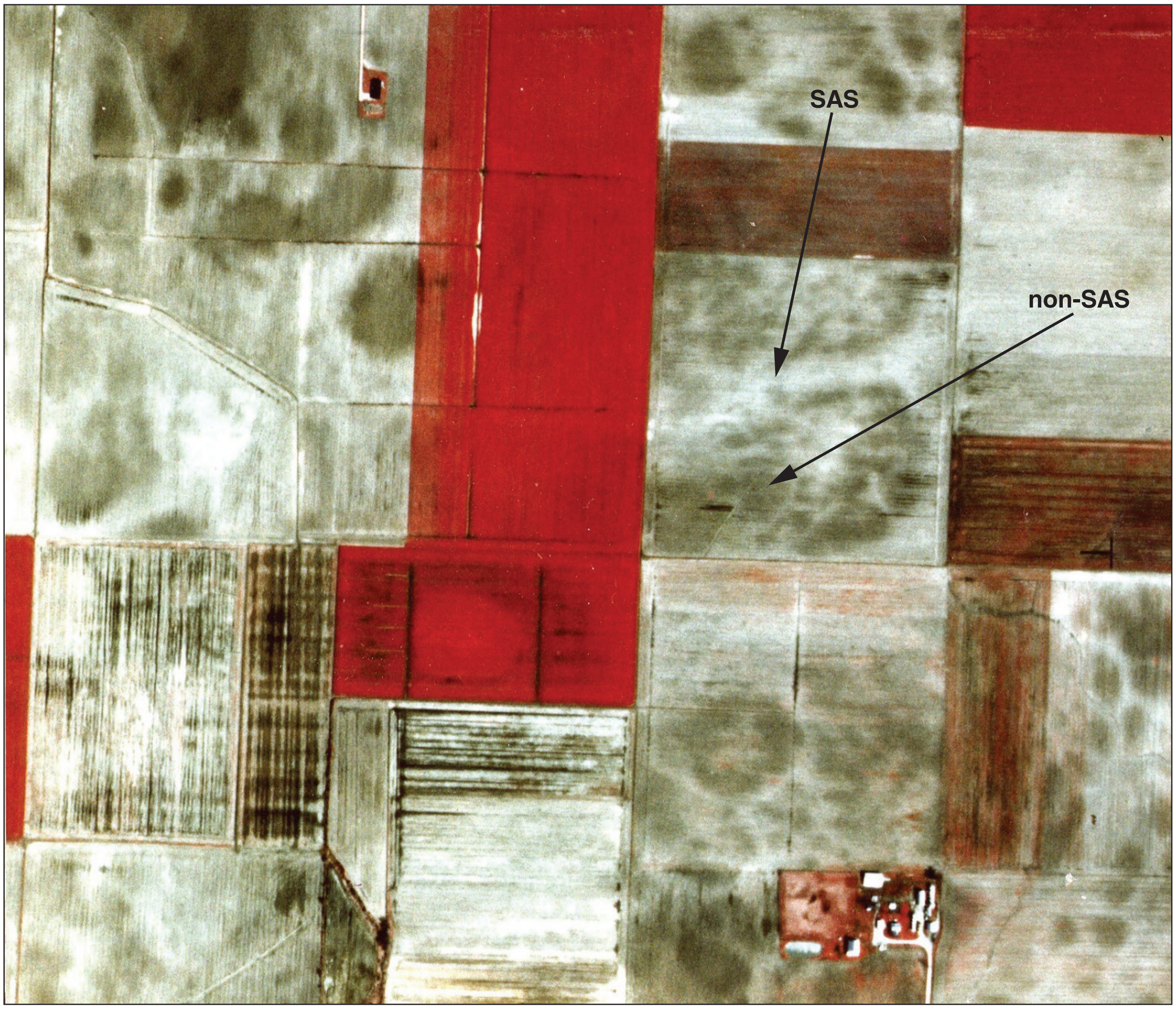


Figure 10.4 Example of the sodium-affected soils (SAS)/non-SAS complex in Washington County, Illinois (Sabata 1998; Soil Survey Staff 2015). The patterned ground is inferred to be related to relict ice-wedge polygons and several small relict thermokarst lakes (thaw lake basins; Indorante et al. 1995, 2011). Used by permission of the Soil Science Society of America.

PART III: Supplemental Data

APPENDIX A: Geophysical and Laboratory Methods

Electrical Resistivity

Timothy H. Larson

Methodology

Resistivity data were acquired using two different electrode configurations. Data from the Mascoutah area were obtained using the Wenner electrode configuration, in which four stakes (electrodes) are equally spaced along a line with two potential electrodes inside two current electrodes. Each measurement involves a hemispheric volume of material surrounding the potential electrodes. Deeper measurements were made by increasing the spacing between electrodes. Data from the Lebanon and Vandalia areas were acquired using the dipole–dipole electrode configuration, in which four electrodes (stakes) are placed in a line, with two electrodes forming a current dipole next to two electrodes forming a potential dipole. The depth of penetration can be controlled by varying the distance between the two electrodes in each dipole (dipole length) and the spacing between the dipoles (dipole separation). Although raw data from the two configurations look different, processing software accounts for these differences and produces comparable images.

For the high-resolution electrical earth resistivity images constructed for this project, up to 60 stainless steel stakes were pushed into the ground at intervals of 16 ft (5 m) along each resistivity profile. The stakes were connected through four 328-ft-long (100-m-long) multicore cables to a computer-controlled resistivity meter (ABEM Terrameter SAS 1000 or 4000) and switching system (LUND imaging system). A control program sequentially switched more than 100 combinations of electrodes, operated the instrument, and stored the data. After a set of readings were completed, the resistivity equipment was moved in 328-ft (100-m) increments and more data were acquired.

With this method, profiles of continuous resistivity measurements were obtained at 15-ft (5-m) spaces and a depth of more than 200 ft (60 m). For this project, individual resistivity profiles ranged in length from about 0.2 to 1.2 mi (0.4 to 2.0 km). A two-dimensional resistivity image was produced from the electrical data with a finite element inversion program (RES2DINV; Loke and Barker 1996) that calculates an approximation of the true resistivity of the earth materials. Finally, topographic information was added to the model along the line of profile.

Color Coding in Figures

Because of the large variation in resistivity values within this study area, resistivity values are represented using two different logarithmic scales. Data from the Mascoutah and Lebanon areas are plotted using a resistivity range of 10 to 320 ohm-m. Typically, geologic materials having resistivity values less than 50 ohm-m are silt- or clay-rich in texture. Lower resistivity corresponds to higher clay content. The materials with lower resistivity are coded green and blue on the figures. Materials identified by dark green, yellow, orange, or red colors have resistivity greater than 50 ohm-m and are generally composed of coarser sediments, such as sands and fine gravel. In some instances, these materials may also contain beds or lenses of sand and clay. Very coarse sand and gravel in this area have resistivity values of >226 ohm-m and are identified by the orange and red shading on the figures.

In the Vandalia area, materials were encountered that have very high resistivity values. For this area only, a different color scale was used to better represent the wider range in resistivity values, ranging from 10 ohm-m to greater than 1,810 ohm-m. With this scale, the unconsolidated sediments with resistivity values of less than 56 ohm-m are silty or clay-rich. Smaller (or lower) resistivity corresponds to more clay content. These materials are coded blue. Materials depicted in shades of green, between 56 and 226 ohm-m, generally are composed of increasingly coarse sands. In some instances, these materials may also include intervening layers of sand and clay. Very coarse sand with gravel in this area, having resistivity values that exceed 226 ohm-m, are shaded in yellow and brown. These very high resistivity values may be caused by dry (unsaturated) conditions or zones of calcite cement. These zones are depicted by dark brown to black shades. Small areas with similarly very high resistivity were encountered in several of the ridges south of Mascoutah, but because these were localized, we did not use the wide-range color scale here.

OSL Dating

For OSL dating at the ISGS laboratory (ages reported at Stop 7 for the Vandalia and Munie sand and gravel pits, and at Stop 9, Raymond Basin) directed by Hong Wang, samples were sieved to extract 90- to 125-µm (or up to 150-µm) size fractions and were treated with 2 M HCl and bleach. Quartz grains were separated using a density of 2.70 g/mL^3 of lithium metatungstate liquid, followed by 40% HF and 2 M HCl etching for 80 and 50 minutes, respectively. The SAR protocol used here corrected the natural and five regenerative OSL signals by a corresponding OSL signal produced by a small but constant test dose, known as sensitivity-corrected OSL (Murray and Wintle 2000; Wintle and Murray 2006). The paleodose in quartz grains was obtained from the interpolation of sensitivity-corrected natural OSL signals on the sensitivity-corrected regenerative dose–response curve on a six-cycle sequence with a preheated temperature of 260 °C and a cut-heat temperature of 220 °C. All paleodoses were measured with a Risø Model TL/OSL-DA-20 reader with a $^{90}Sr/^{90}Y$ built-in source, blue LEDs centered at 470 nm, Hoya 2×U340 and Schott BG-39 filters, and an EMI 9635 QA photomultiplier tube. The dose rate was obtained by using an Ortec GEM-40190P gamma spectrometer on 100-g pulverized samples that were preheated at 450 °C and tightly held in Marinelli beakers sealed with layers of Parafilm along the edges.

Optically stimulated luminescence dating at the UNL laboratory (Keyesport samples, Stop 2) used the methodology of Miao et al. (2007). This analysis used the SAR protocol. Multiple aliquots were prepared from a single sample, each of which produced an individual value for the equivalent dose (De). After discarding aliquots with unsatisfactory behavior during luminescence measurements, the remaining aliquot De values were combined to determine the final age. Samples were collected in aluminum tubes pushed into the exposure. Dose rate values were calculated from the bulk sediment concentrations of K, Rb, U, and Th as measured by inductively coupled plasma mass spectrometry and inductively coupled plasma atomic emission spectroscopy. Samples were opened in the laboratory under dim amber light and wet-sieved to separate the 90- to 150-µm grains. The grains were treated by flotation in 2.7 g/cm^3 of sodium polytungstate to remove heavy minerals, and with hydrofluoric and fluorosilicic acid treatments to etch the quartz and remove feldspars. The isolated quartz grains were then applied to aluminum disks with silicon spray and a 5-mm mask. The quartz aliquots were analyzed with a Risø Model TL/OSL-DA-20 reader. The range in final age estimates incorporated errors from the De values and analytical errors in elemental concentrations, and has an assumed relative variation of ±30% in water content. The water content for the Keyesport sand samples was estimated based on water content for saturated sand of similar grain sizes in Illinois Department of Transportation bridge borings in the region.

Other Laboratory Methods (Compositional and Grain Size Analyses)

Data for *carbonate analyses* were obtained by utilizing either the methodology of Dreimanis (1962) for the Chittick apparatus or an alternative high-precision method devised by Hong Wang (ISGS) for the percentage of carbonate based on the dissolution and capture of purified carbon dioxide. Both methods used the <74-µm fraction. For the Chittick analyses, calcite and dolomite percentages were calculated in addition to total carbonate; however, the percentage of calcite is not very reliable (has low precision). *Clay mineralogy* procedures and percentage calculations were based on the methodology of Hughes et al. (1994) and Moore and Reynolds (1997) using the glass slide method developed by H.D. Glass. *Grain size analyses* (Tables B2 and B3) used standard hydrometer analyses and were performed at the ISGS Sediment Laboratory.

APPENDIX B: Supplemental Data

Table B1 Core descriptions

Core name	Top (ft)	Bottom (ft)	Description
Kessler	0.0	8.0	Peoria Silt, silt loam to heavy silt loam, mainly 10YR 5/5, modern soil in upper 4.0 ft (1.2 m).
Kessler	8.0	11.0	Roxana Silt, heavy silt loam, 8YR 4/4, leached.
Kessler	11.0	20.0	Sangamon Geosol in Hagarstown Member, loam to clay loam, 7.5YR 3.5/4, leached, upper 5.0 ft (1.5 m) is solum of Sangamon Geosol.
Kessler	20.0	23.0	Hagarstown Member, medium to coarse sand, reddish brown (7.5YR 4/6), moderately well sorted, some fines.
Kessler	23.0	40.0	Hagarstown Member, medium to coarse sand, some gravel and some fines, 10YR 4/6, leached, moderately sorted.
Kessler	40.0	54.0	Hagarstown Member, medium to coarse sand, common gravelly zones [mostly <1.0 in. (<2.5 cm)], subangular, some cemented zones, calcareous.
Kessler	54.0	55.0	Limestone cobble; drilled through.
Kessler	55.0	59.0	Hagarstown Member, fine to medium sand, well sorted, calcareous.
Kessler	59.0	135.0	Hagarstown Member, zones of medium to coarse sand and gravel [lots of core loss in places], moderately sorted, subangular to angular, noncohesive, <10% fines, calcareous.
Kessler	135.0	137.0	No sample.
Kessler	137.0	138.5	Bedrock colluvium, pebbly diamicton, 10YR 5/6.
Kessler	138.5	142.5	Pennsylvanian bedrock, limestone and siltstone, gray, layer of soft shale at the base, has fossils including crinoids.
Culli	0.0	4.5	Peoria Silt, heavy silt loam to silty clay loam, 10YR 5/6, leached, modern AE and Bt horizons [AE in upper 0.5 ft (15 cm)].
Culli	4.5	9.5	Peoria Silt, silty loam, 10YR 5/4, leached, contains clay lamellae ~2 mm thick every 1 to 2 cm.
Culli	9.5	13.0	Roxana Silt, heavy silt loam, 9YR 4/4, leached, weak crumb soil structure.
Culli	13.0	18.0	Hagarstown Member (Sangamon Geosol A and Bt), sandy clay loam diamicton, 7.5YR 4/6, clay skins, weak mottling, more gravelly in lower portion.
Culli	18.0	20.0	No sample.
Culli	20.0	26.0	Hagarstown Member, sandy loam diamicton, 10YR 5/6, leached, some Mn stains and minor clay accumulations, rotten weathered pebbles, upper C horizon [supraglacial drift].
Culli	26.0	40.0	Hagarstown Member, predominantly loamy sand and medium sand, some coarse sand, fine sand and minor clay, 10YR 5/4–4.5/4, moderately sorted, leached, significant core loss (but gamma log confirms mainly sand).
Culli	40.0	40.8	Glasford Formation, pebbly loam diamicton, more dense than above, 2.5Y 5/4, calcareous, C horizon, some iron stains.
Culli	40.8	58.0	Glasford Formation, pebbly loam diamicton (upper) to silt loam diamicton (lower), 5Y 4/1.5, dense, hard, uniform [subglacial till], variety of pebbles include coal, shale, carbonate, and fine-grained sandstone, calcareous.
Culli	58.0	59.5	Glasford (complex); intermixed leached greenish (10Y 4/1) silty clay with brown (2.5Y 4/2) calcareous diamicton.
Culli	59.5	61.0	Very fine sand, 5Y 5/2, faint stratification, strongly calcareous, well sorted.

Continued on next page.

Table B1 (*continued*)

Core name	Top (ft)	Bottom (ft)	Description
Culli	61.0	81.0	Glasford? (complex); mostly silty clay loam diamicton (2.5Y 4/1.5 and calcareous) with significant inclusions of greenish gray silty clay (10Y 4/1 to 10GY 4/1—leached Lierle Clay with clay skins and soil structure) as well as inclusions of brown silty clay to clay lake sediment (2.5Y 4/2), such as at 65 to 66.7 ft (20 m) [this unit is a mixed zone of subglacial diamicton with clasts of the underlying Lierle Clay/Yarmouth Geosol Btg horizon and lake sediment that was incorporated into the glacier]; the Lierle Clay inclusions are sometimes more than 1.0 ft (0.3 m) thick in the core and in other areas are small, wispy contortions that are in contrast to intervening diamicton.
Culli	81.0	91.0	Tongue of Petersburg Silt; clay to silty clay (lake sediment), 2.5Y 3/2, faintly laminated, knife cuts it like soap or chocolate!
Culli	91.0	95.0	Very fine sand to fine sand, moderately sorted, some fines, 2.5Y 5/2, calcareous, loose to weakly cohesive.
Culli	95.0	99.0	Coarse silt and very fine sand, few Mn stains, mostly oxidized with some mottling, 2.5Y 6/5, friable.
Culli	99.0	105.0	Mostly coarse sand, some very coarse sand, medium sand and fines, moderately sorted, subangular to subrounded, calcareous, 2.5Y 5/5, friable.
Culli	105.0	115.0	No sample, but coarse sand according to gamma log.
Culli	115.0	118.0	Coarse sand and very coarse sand, loose to weakly cohesive, subangular, moderately sorted, some fines, 2.5Y 5/4, calcareous.
Culli	118.0	120.0	No sample.
Culli	120.0	120.5	Silt loam, dense, faintly laminated(?), Liesegang banding, 4Y 5/4.
Culli	120.5	123.0	No sample (maybe silt or diamicton from gamma log).
Culli	123.0	137.0	Lower unit of Glasford Formation; pebbly loam diamicton (till), dense, some large [2-in. (5-cm)] fragments of carbonate and red shale, few conifer wood fragments (especially near base), few thin sandy or gravelly lenses [<0.3 ft (0.1 m) thick], pebbles include shale, carbonate, coal, 5Y 4/2.
Culli	137.0	138.6	Very fine sand and silt, crudely laminated, calcareous, some wood fragments (plant macrofossils?), 5Y 4.5/2.
Culli	138.6	139.0	Clay-rich diamicton (silty clay loam to silty clay), 2.5Y 4/2, calcareous, has some shear planes at 45° angle (mixed with underlying lake sediment).
Culli	139.0	145.0	Petersburg Silt; silty clay to clay, massive to laminated lake sediment, some layers slightly contorted, calcareous, 10YR 4/2 (slight pinkish hue).
Culli	145.0	146.0	Very pebbly silty clay loam(?) diamicton, lots of pebble lithologies, sandstone, granite, chert, other, 1Y 4/2.
Culli	146.0	148.5	Petersburg Silt; silty clay to clay, thickly laminated—reddish brown to gray, some contortions in layers, calcareous, 1Y 4/2.
Culli	148.5	151.0	Clay loam diamicton, clay, 10YR 4/3, local orange shale fragments, calcareous.
Culli	151.0	152.0	Clay loam diamicton and silty clay, mottles, 2.5Y 5/4 and 10YR 4/3 (diamicton) and 10Y 6/1 in silty clay (residual soil mixed with diamicton—colluvium?).
Culli	152.0	155.0	Limestone, argillaceous, fractures in upper part, fossiliferous, light gray.
Grandview	0.0	3.9	Soft brown (7.5YR 4/3) silt loam, laminated 0.5 to 1.5 ft (15.2 to 45.7 cm) but otherwise massive, few pebbles, nonreactive to very weakly dolomitic; loess, Roxana Silt.
Grandview	3.9	4.4	Strong brown very pebbly silt loam diamicton, leached but with local secondary calcite accumulation; contact between loess and glaciofluvial sediment, root (upper C) of Sangamon, Hagarstown Member.

Continued on next page.

Table B1 (*continued*)

Core name	Top (ft)	Bottom (ft)	Description
Grandview	4.4	65.7	Coarse gravel, rounded to bedded pebbly very coarse sand and medium sand, reactive, light yellow brown, n-gamma logs bedded [1.0–4.0 ft (0.3–1.2 m)] on left (complete recovery in augured 4.0- to 9.0-ft (1.2- to 2.7-m) interval but retained only subsamples; very poor recovery below); glaciofluvial, possibly outwash portal, Hagarstown Member.
Grandview	65.7	66.2	Yellow brown pebbly clay loam, laminated; lower flow facies, Hagarstown Member.
Grandview	66.2	72.0	No recovery, n-gamma logs left, likely coarse grained.
Grandview	72.0	74.0	No recovery, n-gamma logs right, likely fine grained (diamicton?), Glasford Formation.
Grandview	74.0	98.9	Bedded diamicton, more gray in upper part, more oxidized in lower 4.0 ft (1.2 m), including dark gray brown (2.5Y 4/2) loam, 2% clasts (round quartzite, siltstone); olive gray (5Y 5/2) clay; gray (2.5Y 5/1) loam, 3% clasts (quartz, siltstone), gray brown (2.5Y 5/2) loam, 3 to 5% clasts (siltstone, igneous quartz, red chert); silty and gravelly zones, other clasts include shale, brown chert, white chert, limonite nodule, disturbed zones; sheared sheets of till, 2 to 8.0 ft (0.6 to 2.4 m) thick, Glasford Formation.
Grandview	98.9	99.0	Gray brown (2.5Y 5/2) sandy loam diamicton, disturbed; Glasford Formation.
Grandview	99.0	104.0	No recovery, n-gamma transitional to right, defining higher block that is distinctive to 108.0 ft (32.9 m); inclusion of Banner Formation, Glasford Formation.
Grandview	104.0	104.5	Gray brown sandy loam diamicton, intercalating with gray brown loam diamicton, clear contact; sheared inclusions, Glasford Formation.
Grandview	104.5	106.2	Dark gray (10YR 4/1) oxidized to olive brown loam diamicton, 5 to 10% clasts (igneous, metaquartz, siltstone), gradual transition, >1.0 ft (>0.3 m) healed subvertical joint, secondary calcite, small intersecting horizontal joints; elevation is approximately elevation of surrounding plain; inclusion of upper C of Yarmouth in “Hillery,” Glasford Formation.
Grandview	106.2	107.9	Dark gray (10YR 4/1) loam diamicton, pink cast when dry, 5 to 10% clasts (igneous, metaquartz, siltstone), wood fragments, likely spruce; till, inclusion of Banner Formation (Hillery-like), Glasford Formation.
Grandview	107.9	108.3	Olive brown (2.5 & 4/3) very pebbly loam diamicton, 1- to 2-cm clasts (chert, sandstone, subrounded to subangular) at contact; color similar to ~98 ft (~30 m), but from gravel and n-gamma is transitional; oxidation may be from below; till, Glasford Formation.
Grandview	108.3	114.0	No recovery; n-gamma logs left—likely coarse-grained facies, tongue of Pearl Formation.
Grandview	114.0	114.5	Veneer of brown sand over silt, laminated to horizontal bedded, few granules, disturbed by coring, sharp contact; glaciofluvial sediment, tongue of Pearl Formation.
Grandview	114.5	119.8	Dark gray (2.5Y 4/1) to gray silty clay loam diamicton, bedded, with very silty to sandy beds, 7 to 10% clasts (siltstone, shale, coal, chert), mainly angular to subangular, reactive, contact missing, color distinctly grayer than surrounding units; morainic deposits, mainly till, Glasford Formation.
Grandview	119.8	123.0	No recovery, n-gamma similar to overlying unit, logging right, likely diamicton; Glasford Formation.
Grandview	123.0	124.0	Boulder, sandstone; Glasford Formation.
Grandview	124.0	129.0	0.5- to 1.0-ft-thick (0.2- to 0.3-m-thick) beds of olive gray pebbly loam, soft, dark gray (10YR 4/1), lean clay and dark brown clay (shear zone) at bottom, reactive; morainic till, Glasford Formation.

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Table B1 (*continued*)

Core name	Top (ft)	Bottom (ft)	Description
Grandview	129.0	129.5	Yellow brown (10YR 5/4) to light olive brown (2Y 5/4) silty clay loam diamicton, with siltstone, chert, metaquartz clasts, strong reaction, over strong brown clay (shear zone), n-gamma transitional from far right; glaciotectonized till, Glasford Formation.
Grandview	129.5	139.0	No recovery, n-gamma logging far left, likely sand and gravel, bedded; tongue of Pearl Formation.
Grandview	139.0	139.8	Brown fine sand over pebbly sand; tongue of Pearl Formation.
Grandview	139.8	149.1	No recovery, n-gamma near minimum, likely sand and gravel, bedded; tongue of Pearl Formation.
Grandview	149.1	151.3	Brown (oxidized) poorly sorted granular coarse sand grades down to gray medium to coarse sand, loamy sand, and sandy fine gravel, bedded, strong to very strong reaction; glaciofluvial sediment; tongue of Pearl Formation.
Grandview	151.3	159.0	No recovery, n-gamma similar to adjacent, likely sandy gravel, bedded; tongue of Pearl Formation.
Grandview	159.0	169.0	Recovered handful of pebbly sand or sandy gravel, n-gamma similar to overlying, but logging slightly to right; tongue of Pearl Formation.
Grandview	169.0	173.0	Gray very fine to fine sand and silty sand, few pebbles, disturbed by sampling, but weak bedding apparent [thickness?], fluidized zones, strong reaction; glaciofluvial sediment, tongue of Pearl Formation.
Grandview	173.0	179.0	No recovery, n-gamma similar to overlying, likely fine sand; tongue of Pearl Formation.
Grandview	179.0	180.5	Gray brown pebbly loam diamicton, subrounded to angular sandstone, rhyolite, milky quartz; till.
Grandview	180.5	204.0	No recovery, n-gamma logs mostly moderate to right, 2.0-ft-thick (0.6-m-thick) left blocks, possibly interbedded gravel and diamicton.
Grandview	204.0	209.0	Virtually no recovery except for a handful of granules, n-gamma strong to left; glaciofluvial lens.
Grandview	209.0	211.0	Gray brown (2.5Y 5/2) sandy loam diamicton, 10 to 50% gravel up to 2.5 cm, igneous and sedimentary lithologies, including sandstone, quartzite, chert, subangular to subrounded, strong reaction; till with glaciofluvial sediment.
Grandview	211.0	214.0	No recovery, no change in n-gamma.
Grandview	214.0	219.0	Recovered only two large gravel clasts and a smear of diamicton, n-gamma in same block as adjacent samples; till?
Grandview	219.0	219.6	Gray silty clay loam to silt loam diamicton, sharp contact; till.
Grandview	219.6	221.0	Light olive brown (2.5Y 5/3) silty clay loam diamicton, massive, 2 to 3% clasts (red, brown, and gray sandstone, rounded, brown coral), oxidized zones with secondary calcite, strong reaction, graded contact (reaction much stronger than below), n-gamma transitional from overlying block; till, but possibly inclusion of upper C horizon material.
Grandview	221.0	221.4	Gray (2.5Y 4/1) silty clay loam to silt loam diamicton, 2 to 3% clasts (granite, sedimentary), wood (flattened twigs, likely spruce), moderate reaction, contact missing; till.
Grandview	221.4	224.0	No recovery, n-gamma logs left, similar to below, likely sandy.
Grandview	224.0	226.0	Gray very fine sand, no pebbles, massive to weakly bedded, reactive; alluvium Petersburg or Harkness?

Continued on next page.

Table B1 (*continued*)

Core name	Top (ft)	Bottom (ft)	Description
Grandview	226.0	230.0	Dark gray brown (10YR 4/2) silt loam, massive, very dark gray (10YR 3/1) silty clay loam, laminated, browner silty clay loam and silt, laminated, variably leached to dolomitic to calcareous, sharp contacts, disturbed in upper foot, fine root or twig fragments (spruce?); lacustrine and alluvial sediment, Petersburg or Harkness?
Grandview	230.0	234.0	Gray brown loamy very fine to fine sand, horizontally bedded, contact missing, n-gamma in steady block.
Grandview	234.0	234.8	Olive gray (5Y 5/2) silty clay loam, weak bedding, irregular contact, no reaction; alluvium, Canteen member.
Grandview	234.8	249.0	Gray to olive gray fine sandstone, cross-bedded, weathered in upper part, missing 235.0 to 240.0 ft (71.6 to 73.2 m); bedrock, Pennsylvanian (undifferentiated)
Mersinger	0.0	4.5	1.0 ft (0.3 m) of roadbed recovered, lower part likely soft silt loam; loess with modern soil.
Mersinger	4.5	7.6	Soft, moist, dark yellow brown (10YR 3/4), mottled with reddish hue, silt loam, 0.5 tons/ft^2 (tsf), leached; loess, Peoria Silt.
Mersinger	7.6	10.5	Dark brown (7.5YR 4/3.5), distinct color change, silt loam, 0.75 tsf, leached; loess, Peoria Silt.
Mersinger	10.5	13.0	Brown (7.5YR 4/3.5), soft silt loam, massive; softens downward (2.25–0.75 tsf) and becomes more loamy with few fine subangular pebbles; gradational contact; cumulic soil below 12.4 ft (3.8 m); loess, Roxana Silt.
Mersinger	13.0	16.0	Stronger brown silt loam to loam, few round to subrounded pebbles, common root traces, blocky structure, increasing soil strength with depth, gradational contact; cumulic Sangamon Geosol in colluvium and loess, Teneriffe Silt.
Mersinger	16.0	17.6	Strong brown (7.5YR 5/6), mottled with orange hue, very fine sandy loam, 0.75 to 1.0 tsf grades to very soft strong brown (7.5YR 4/6) loam; leached; sharp contact. Sangamon Geosol in alluvium; Pearl-Teneriffe complex.
Mersinger	17.6	19.3	Dark yellow brown (10YR 4/6) loamy fine sand with granules, coarsens downward, weakly bedded, leached, sharp contact; upper C horizon, Pearl-Teneriffe complex.
Mersinger	19.3	22.2	Brown (10YR 4/6) pebbly loamy sand, very poorly sorted, bedded, graded; leached, contact missing; alluvium, Pearl-Teneriffe complex.
Mersinger	22.2	26.0	No sample, gamma log similar to above; alluvium, Pearl-Teneriffe complex.
Mersinger	26.0	28.8	Firm brown (10YR 6/4–5/6), mottled or banded, silt, massive to laminated; 0.25 to 0.75 tsf, increases to 2.25 × 28 ft (0.7 × 8.5 m); calcareous; sharp contact; alluvium, Pearl-Teneriffe complex.
Mersinger	28.8	29.5	Very fine sand, massive; alluvium, Pearl-Teneriffe complex.
Mersinger	29.5	33.9	Light yellowish brown (2.5Y 6/4) medium silt, laminated, clay laminae in lower 0.5 ft (0.2 m); 1.0 tsf; sharp color contact; alluvium, Pearl-Teneriffe complex.
Mersinger	33.9	35.5	Gray silty clay and clay, laminated, interbedded; 0.5 to 1.25 tsf; calcareous; sharp contact; alluvium, Pearl-Teneriffe complex.
Mersinger	35.5	37.5	Gray brown (2.5Y 5/2) silt, soft, bedded, sharp contact; loess, Pearl-Teneriffe complex.
Mersinger	37.5	38.4	Brown silt loam, laminated, dry, 3.25 tsf, grades down to fine sand, bedded; alluvium, Pearl-Teneriffe complex.
Mersinger	38.4	40.0	No sample, gamma log drops sharply at 40.0 ft (12.2 m).
Mersinger	40.0	61.0	Light yellow brown (10YR 6/5) to brown fine sand and loamy sand, loose to soft; contact missing (partial sample recovery); outwash, Pearl Formation.

Continued on next page.

Table B1 (*continued*)

Core name	Top (ft)	Bottom (ft)	Description
Mersinger	61.0	67.0	Clean gravel, clayey gravel, coarse sand, subangular to subrounded; bedded, loose; poor recovery, driller estimated lower limit; outwash, Pearl Formation.
Mersinger	67.0	67.9	Brown pebbly loam diamicton, weakly bedded, calcareous, sharp contact; debris flow, Glasford Formation.
Mersinger	67.9	72.2	Dark gray brown (2.5Y 4/2) pebbly loam diamicton, massive, very dense; coal fragment, cobbles >2.5 in. (>6.4 cm); >4.5 tsf; calcareous. Basal till; Glasford Formation.
Mersinger	72.2	77.5	Dark gray brown (2.5Y 4/2), slightly more olive than above, pebbly loam diamicton, bedded, gravel and silt beds less than 0.1 ft (3.1 cm) thick; calcareous, gradational contact; till, Glasford Formation.
Mersinger	77.5	80.0	Dark gray brown (2.5Y 4/2), pebbly loam diamicton, massive, no discernable change in effervescence, but lower total carbonate than above; contact missing; till, Glasford Formation.
Mersinger	80.0	86.0	No sample; gamma log is similar to above, but gravelly 84.0 to 86.0 ft (25.6 to 26.2 m); till, Glasford Formation.
Mersinger	86.0	91.6	Olive brown (2.5YR 4/3) silt loam diamicton, irregular fracture, possibly weakly bedded, small wood fragments; till with incorporated lacustrine sediment from below, deformable bed(?), Glasford Formation.
Mersinger	91.6	96.0	Olive gray (5Y 6/1) silt, laminated; very dense (>4.5 tsf), disseminated organics. Lacustrine sediment, possibly backwater lake of Kaskaskia, Petersburg Silt.
Mersinger	96.0	99.6	Olive gray silt, laminated, clay laminae, firm (1.5–3.75 tsf); disseminated organics with common wood fragments, including 0.25-ft (7.6-cm) spruce branch; increased deposition rate with onset of glaciation, Petersburg Silt.
Mersinger	99.6	106.0	No sample; gamma logs left in upper half, right in lower half; sandy zone? Petersburg Silt.
Mersinger	106.0	106.5	Dark gray brown silt loam, diffuse brown mottle, gastropods (*Pomatiopsis* sp., *Carychium* sp., *Succinea* sp.), calcareous, distinct contact; lacustrine sediment, Petersburg Silt.
Mersinger	106.5	107.5	Dark gray brown (2.5Y 4/2) silt loam to silty clay loam, massive to weakly bedded, weak paleosol at top, dolomitic; distinct contact; shallow lacustrine sediment, Petersburg Silt.
Mersinger	107.5	113.2	Gray brown (2.5Y 5/2) silt loam, olive hue, very stiff, grades to beds of silty clay loam and silt loam, laminated zones, sandy at ~109.5 ft (~33.4 m); weak paleosol, root traces through interval, leached to weakly dolomitic; gradational contact; top of ancestral floodplain, Petersburg Silt.
Mersinger	113.2	117.0	Dark gray brown (2.5Y 4/2) loam, coarsening with depth, root traces, >4.5 tsf; fractured, leached; weak paleosol in alluvium, Petersburg Silt.
Mersinger	117.0	118.0	Loam, laminated, organic, leached; alluvium, Petersburg Silt.
Mersinger	118.0	121.0	Recovered only a handful of fine sand; alluvium, Petersburg Silt.
Mersinger	121.0	122.2	Coarse sand [1 ft (0.3 m)], calcareous; alluvium, Petersburg Silt.
Mersinger	122.2	126.0	Silt loam to loam, laminated, over silty clay bed; calcareous; alluvium, Petersburg Silt.
Mersinger	126.0	127.0	Gray brown silty clay loam, laminated; alluvium, Petersburg Silt.
Mersinger	127.0	131.0	Recovered bit of shale at 127.0 ft (38.7 m). Bottom of hole.
Virgin	0.0	8.3	Dark yellow brown (10YR 4/4) silt loam contains A horizon of modern soil, lightly bleached and bedded below ~3.0 ft (~0.9 m) loess, Peoria Silt.

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Table B1 (*continued*)

Core name	Top (ft)	Bottom (ft)	Description
Virgin	8.3	13.7	Brown (7.5YR 4/4) silt loam, massive, leached, few silans. Loess, Roxana Silt.
Virgin	13.7	15.0	Strong brown (7.5YR 4/6) granular silty loam diamicton, massive, angular lithic facies, leached. Colluvial loess facies, Roxana Silt.
Virgin	15.0	17.0	Strong brown pebbly silty clay loam diamicton, massive, argillans >1 cm, Fe-Mn concentration >1 cm, angular clasts 1 to 10 mm, leached. B Paleosol in till, Sangamon/Hagarstown Member (diamicton).
Virgin	17.0	20.3	Soft, yellow brown (10YR 5/6), bleached-mottled, pebbly silty clay loam diamicton, to loam diamicton, crudely bedded, clasts to 25 mm, argillans, leached, strong horizontal fabric, possibly highlighted by pedogenic features, including Liesegang bands. Sangamon B horizon in supraglacial till, Sangamon/Hagarstown Member (diamicton).
Virgin	20.3	27.0	No sample.
Virgin	27.0	28.2	Soft, yellow brown (10YR 5/6), bleached-mottled, pebbly silty clay loam diamicton to loam diamicton, crudely bedded, clasts to 25 mm, argillans, leached, strong horizontal fabric, possibly highlighted by pedogenic features, including Liesegang bands. Sangamon B horizon in supraglacial till, Sangamon/Hagarstown Member (diamicton).
Virgin	28.2	30.8	Yellow brown loam diamicton, horizontally bedded, pebbly, strong subhorizontal fabric, calcareous, oxidized. Sangamon upper C, Hagarstown Member (diamicton).
Virgin	30.8	32.0	No sample—sand and gravel? Hagarstown Member.
Virgin	32.0	32.8	Yellow brown, gravel with fine sand matrix, graded, sharp contact, calcareous; outwash, Hagarstown Member.
Virgin	32.8	33.8	Yellow brown, pebbly loam diamicton, bedded, sand and gravel horizon <0.1 ft (<3 cm) thick, heterolithic clasts; outwash, Hagarstown Member.
Virgin	33.8	37.0	No sample—sand and gravel? Hagarstown Member.
Virgin	37.0	37.8	Yellow brown, coarse gravel over pebbly loam diamicton, horizontal fabric, platy structure in lower part, abrupt color contact, very stiff; outwash, Hagarstown Member.
Virgin	37.8	45.8	Gray brown (10.5YR 5/2), pebbly loam diamicton, massive, weak horizontal fabric (shearing?) highlighted by slight oxidation, calcareous, abrupt contact (erosion surface); till, Glasford Formation.
Virgin	45.8	50.3	Yellow brown (10YR 5/6) silty clay loam diamicton, soft, horizontal fabric (shearing?), argillans, Fe-Mn concentration <5 mm, clay-rich at bottom, leached, inclusion of Yarmouth B/Banner; till, Glasford Formation.
Virgin	50.3	61.1	Yellow brown, silt loam diamicton, massive, jointed with bleaching along joints, clasts mainly very coarse sand to fine gravel but up to coarse pebbles (4 cm), oxidized, calcareous, distinct color contact. Inclusion of Yarmouth upper C/Banner Formation; till, Glasford Formation.
Virgin	61.1	62.8	Dark gray, diamicton, massive, oxidized on joints, calcareous, gradational contact (mixture with lower unit). DC horizon, till, Glasford Formation.
Virgin	62.8	67.5	Dark gray brown, silt loam diamicton, heterogeneous, sand and thin granule layers. At 65.2 ft (19.9 m), sharp contact between brownish gray ("pink"?) and grayish brown. Unit possibly includes amalgamated till sheets, although no shear zones or slickensides are apparent; Glasford Formation.
Virgin	67.5	74.0	Dark gray brown to slightly olive (2.5Y 4/2) clay loam to silty clay loam diamicton, massive, fine shale clasts abundant, wood fragments, chert cobble at base, distinct color contact, subhorizontal fabric and bedding or shear planes; Glasford Formation.

Continued on next page.

Table B1 (*continued*)

Core name	Top (ft)	Bottom (ft)	Description
Virgin	74.0	87.1	[No sample 85.4–87.0 ft (26.0–26.5 m), 87.0–87.1 ft (26.5–26.6 m) cobble at top of run]. Brown (10YR 4/3) silt loam to loam diamicton, massive, similar color and texture to ~64 ft (~19.5 m; trace element data appear to support correlation), yellower in lower 2.0 ft (0.6 m), gradational, vague horizontal clast fabric, subangular to well-rounded sedimentary clasts, few wood fragments, quartzite cobble at base below gravel broken by drilling, sharp contact. C horizon till, Glasford Formation.
Virgin	87.1	97.7	Dark brown (10YR 3/3) loam diamicton, horizontal fabric, sedimentary clasts to very fine pebble size, fine wood fragments, possibly horizontal fabric, gravel and sandy loam beds ≤0.2 ft (≤6.1 cm) thick. Dark gray brown (10YR 4/2) silt bed at 93.7 ft (28.6 m), clay-rich. Yellow brown to gray with yellow brown mottle, horizontal fabric, shear zones at 90.0, 92.2, 94.2 ft (27.4, 28.1, 28.7 m) with sharp contacts. Angular to well-rounded chert and sedimentary lithic fragments. Distinctly higher natural gamma signature from overlying diamicton units. Till, mixing with underlying units, Glasford Formation.
Virgin	97.7	100.4	Limestone boulder.
Virgin	100.4	100.6	Disturbed silt and clay, originally stratified. Glasford Formation.
Virgin	100.6	102.0	No sample.
Virgin	102.0	119.2	Light olive brown (2.5Y 4/3) pebbly loam diamicton, massive; granite, sandstone, shale, chert clasts to fine pebble size overall larger than overlying units, subangular to well rounded; no wood; 0.1 to 0.3 ft (3.1 to 9.1 cm) thick sand, sandy gravel zones; graded clayey gravel to very fine sand bed in lowest foot, calcareous; till, Omphghent member.
Virgin	119.2	122.0	No sample, natural gamma log increasing from 120.0 ft (36.6 m).
Virgin	122.0	122.1	Light brown diamicton, massive, calcareous, sharp contact; Glasford Formation.
Virgin	122.1	124.5	Yellow brown (10YR 5/6) loam diamicton, massive, leached, Liesegang banding in upper foot, clay or Mn laminae at base. Truncated paleosol in till? Banner Formation.
Virgin	124.5	126.4	Light yellow brown (oxidized) to gray (lower part) loam diamicton, massive, calcareous, distinct but graded contact. Debris flow, Banner Formation.
Virgin	126.4	127.1	Yellow brown pebbly sand grades up to gray brown silty loam laminated, calcareous. Glaciofluvial sediment.
Virgin	127.1	132.0	No sample—sand and gravel?
Virgin	132.0	133.5	Olive brown fine sandy gravel with clay, loose, round to subangular clasts, calcareous. Glaciofluvial sediment.
Virgin	133.5	135.5	No sample.
Virgin	135.5	136.5	Reddish brown fine gravel, massive, rounded clasts, mainly resist lithologies, sharp contact. Glaciofluvial sediment.
Virgin	136.5	136.8	Reddish brown loam diamicton, massive; till, Banner Formation.
Virgin	136.8	137.2	Gray brown diamicton, disturbed, clay-rich; till, Banner Formation.
Virgin	137.2	137.5	Gray loam diamicton, massive, clasts to 3 cm; till, Banner Formation.
Virgin	137.5	139.0	No sample.
Virgin	139.0	153.5	Gray brown loam diamicton, horizontal bedded, sand lenses, rounded and angular shale fragments, coal, heterogeneous, random fabric of lithics, soft lithics unbroken imply short transport; grades to silty clay loam diamicton, massive, more uniform, with wood fragments at ~147.5 ft (~45.0 m), sharp contact; till, Banner Formation.

Continued on next page.

Table B1 (*continued*)

Core name	Top (ft)	Bottom (ft)	Description
Virgin	153.5	158.0	Dark gray to yellow brown silt loam, fractured in lower part, rich in disseminated organics, calcareous, 155.3 to 157.0 ft (47.3 to 47.9 m) missing, sharp contact. Alluvium, Harkness Silt.
Virgin	158.0	158.5	Gray brown, silty clay loam diamicton, massive, calcareous, sharp contact; Harkness Silt.
Virgin	158.5	159.2	Gray brown silt, horizontally bedded, calcareous at top but gradually leached with depth. Alluvium, Harkness Silt.
Virgin	159.2	165.3	Gray brown silt loam, horizontally bedded, dolomitic, gradational color contact. Alluvium, Harkness Silt.
Virgin	165.3	165.9	Olive brown (2.5Y 4/3), silt loam, massive to faintly bedded, vivianite, leached. Paleosol in alluvium, Canteen member A.
Virgin	165.9	167.0	No sample.
Virgin	167.0	170.0	Olive (5Y 4/3) silt loam, faintly laminated, vivianite, root traces in upper 0.5 ft (15.2 cm), leached, gradual color change. Alluvium.
Virgin	170.0	174.1	Olive brown very fine sandy loam, laminated grades to loam, horizontally bedded, grades to very fine sand, horizontally bedded to massive, leached, sharp contact. Alluvium.
Virgin	174.1	179.4	Yellow brown conglomerate, bedded, large shale fragments, iron-rich, poorly to well indurated. Alluvium or colluvium, reworked Tertiary or older units. Canteen member B.
Virgin	179.4	181.1	Gray shale, laminated. Bedrock.

Table B2 Particle size analysis of outcrop grab samples

Site name	Sample ID	Gravel (% >2 mm)	% within the <2-mm fraction: Sand (% 63–2,000 μm)	Silt (% 2–63 μm)	Clay (% <2 μm)	USDA texture
Ogles Creek Section	Top of Petersburg Silt	0	1	88	11	Silt
Ogles Creek Section	Bottom of Petersburg Silt	0	1	84	15	Silt loam
Highbanks Road Section	HBC-E3	0	1	86	13	Silt loam
Highbanks Road Section	HBC-E2	0	2	88	10	Silt
Highbanks Road Section	HBC-E1	70	9	60	31	Silty clay loam
Highbanks Road Section	HBC-E	0	2	67	31	Silty clay loam
Highbanks Road Section	HBC-D	1	3	56	41	Silty clay
Highbanks Road Section	HBC-C	0	13	59	28	Silty clay loam
Highbanks Road Section	HBC-B	0	25	56	19	Silt loam
Highbanks Road Section	HBC-A	0	98	0	2	Sand

Table B3 Particle size analysis of core samples

Core name	Depth (ft)	Gravel (% <2 mm)	% within the <2-mm fraction: Sand (% 63–2,000 μm)	Silt (% 4–63 μm)	Clay (% <4 μm)
Grandview	1	0	3	73	24
Grandview	75	7	35	39	26
Grandview	85	7	35	41	24
Grandview	107.1	5	39	41	20
Grandview	116	2	21	44	35
Grandview	126	22	42	33	25
Grandview	179	6	44	35	21
Grandview	210	9	40	38	22
Grandview	226.6	0	7	60	32
Grandview	234.1	0	17	63	20
Virgin	7	0	3	78	19
Virgin	148	3	19	51	30
Virgin	152	1	15	55	30
Virgin	158	2	14	54	33
Virgin	162	0	9	62	29
Virgin	168	0	23	63	14
Virgin	170	0	41	49	10

Table B4 Carbonate content data[1]

		% <74 μm			
Core name	Depth (ft)	Calcite (%)	Dolomite (%)	Total carbonate (%)	Method
Culli	40.0	NA	NA	13.2	H. Wang
Culli	42.0	NA	NA	16.4	H. Wang
Culli	44.0	NA	NA	15.8	H. Wang
Culli	46.0	NA	NA	14.3	H. Wang
Culli	48.0	NA	NA	10.5	H. Wang
Culli	50.0	NA	NA	7.8	H. Wang
Culli	52.0	NA	NA	7.3	H. Wang
Culli	58.0	NA	NA	1.7	H. Wang
Culli	60.0	NA	NA	11.8	H. Wang
Culli	62.0	NA	NA	0.6	H. Wang
Culli	64.0	NA	NA	5.0	H. Wang
Culli	66.0	NA	NA	24.5	H. Wang
Culli	68.0	NA	NA	13.9	H. Wang
Culli	70.0	NA	NA	13.3	H. Wang
Culli	72.0	NA	NA	5.2	H. Wang
Culli	74.0	NA	NA	5.5	H. Wang
Culli	76.0	NA	NA	17.6	H. Wang
Culli	78.0	NA	NA	5.3	H. Wang
Culli	80.0	NA	NA	10.8	H. Wang
Culli	82.0	NA	NA	16.0	H. Wang
Culli	84.0	NA	NA	7.1	H. Wang
Culli	86.0	NA	NA	7.8	H. Wang
Culli	88.0	NA	NA	14.7	H. Wang
Culli	90.0	NA	NA	6.2	H. Wang
Culli	92.0	NA	NA	23.6	H. Wang
Culli	95.0	NA	NA	29.6	H. Wang
Culli	98.0	NA	NA	24.9	H. Wang
Culli	100.0	NA	NA	23.9	H. Wang
Culli	102.0	NA	NA	28.6	H. Wang
Culli	104.0	NA	NA	29.2	H. Wang
Culli	115.0	NA	NA	19.5	H. Wang
Culli	116.0	NA	NA	19.5	H. Wang
Culli	117.0	NA	NA	19.5	H. Wang
Culli	118.0	NA	NA	19.5	H. Wang
Culli	119.0	NA	NA	17.9	H. Wang
Culli	120.0	NA	NA	17.9	H. Wang
Culli	121.0	NA	NA	17.9	H. Wang
Culli	122.0	NA	NA	17.9	H. Wang
Culli	123.0	NA	NA	17.9	H. Wang
Culli	124.0	NA	NA	15.2	H. Wang
Culli	126.0	NA	NA	13.6	H. Wang
Culli	128.0	NA	NA	13.5	H. Wang
Culli	130.0	NA	NA	9.0	H. Wang
Culli	132.0	NA	NA	16.0	H. Wang
Culli	134.0	NA	NA	15.5	H. Wang
Culli	136.0	NA	NA	13.1	H. Wang
Culli	138.0	NA	NA	16.4	H. Wang
Culli	140.0	NA	NA	24.4	H. Wang
Culli	142.0	NA	NA	24.0	H. Wang
Culli	144.0	NA	NA	22.7	H. Wang
Culli	146.0	NA	NA	13.2	H. Wang
Culli	148.0	NA	NA	6.4	H. Wang
Culli	150.0	NA	NA	5.8	H. Wang
Culli	152.0	NA	NA	3.4	H. Wang
Culli	154.0	NA	NA	78.9	H. Wang
Culli	155.0	NA	NA	85.6	H. Wang

Continued on next page.

Table B4 (*continued*)

Core name	Depth (ft)	% <74 μm Calcite (%)	Dolomite (%)	Total carbonate (%)	Method
Grandview	75.0	6.5	11.9	18.4	Chittick
Grandview	85.5	4.7	9.2	13.9	Chittick
Grandview	107.1	12.0	12.3	24.3	Chittick
Grandview	116.0	7.0	5.6	12.6	Chittick
Grandview	126.6	2.1	3.2	5.3	Chittick
Grandview	129.1	2.0	10.8	12.8	Chittick
Grandview	179.5	8.9	18.1	27.0	Chittick
Grandview	210.0	5.1	17.2	22.3	Chittick
Grandview	219.9	7.6	9.2	16.8	Chittick
Grandview	221.3	1.7	5.3	7.0	Chittick
Grandview	226.7	2.2	10.4	12.6	Chittick
Grandview	227.8	1.1	15.9	17.0	Chittick
Grandview	228.9	17.6	11.8	29.4	Chittick
Mersinger	19.0	NA	NA	0.0	H. Wang
Mersinger	21.0	NA	NA	0.0	H. Wang
Mersinger	27.0	NA	NA	27.8	H. Wang
Mersinger	29.0	NA	NA	27.2	H. Wang
Mersinger	31.0	NA	NA	27.9	H. Wang
Mersinger	33.0	NA	NA	28.1	H. Wang
Mersinger	35.0	NA	NA	32.1	H. Wang
Mersinger	37.0	NA	NA	30.1	H. Wang
Mersinger	39.0	NA	NA	27.7	H. Wang
Mersinger	41.0	NA	NA	28.2	H. Wang
Mersinger	47.0	NA	NA	26.5	H. Wang
Mersinger	49.0	NA	NA	28.7	H. Wang
Mersinger	51.0	NA	NA	25.6	H. Wang
Mersinger	53.0	NA	NA	31.6	H. Wang
Mersinger	55.0	NA	NA	33.0	H. Wang
Mersinger	57.0	NA	NA	33.2	H. Wang
Mersinger	59.0	NA	NA	35.8	H. Wang
Mersinger	65.0	NA	NA	25.0	H. Wang
Mersinger	71.0	NA	NA	27.4	H. Wang
Mersinger	73.0	NA	NA	24.2	H. Wang
Mersinger	77.0	NA	NA	22.2	H. Wang
Mersinger	79.0	NA	NA	14.4	H. Wang
Mersinger	87.0	NA	NA	11.4	H. Wang
Mersinger	89.0	NA	NA	17.6	H. Wang
Mersinger	91.0	NA	NA	7.4	H. Wang
Mersinger	93.0	NA	NA	16.0	H. Wang
Mersinger	97.0	NA	NA	16.1	H. Wang
Mersinger	108.0	NA	NA	1.9	H. Wang
Mersinger	110.0	NA	NA	1.0	H. Wang
Mersinger	112.0	NA	NA	0.3	H. Wang
Mersinger	114.0	NA	NA	1.7	H. Wang
Mersinger	116.0	NA	NA	2.6	H. Wang
Mersinger	122.0	NA	NA	0.3	H. Wang
Mersinger	126.0	NA	NA	16.0	H. Wang
Virgin	28.6	9.3	6.3	15.5	Chittick
Virgin	30.2	6.1	14.1	20.3	Chittick
Virgin	33.2	2.8	7.8	10.6	Chittick
Virgin	45.1	4.6	8.5	13.1	Chittick
Virgin	46.2	0.0	0.0	0.0	Chittick
Virgin	49.0	0.0	0.0	0.0	Chittick
Virgin	59.6	10.4	8.4	18.7	Chittick
Virgin	64.0	12.7	9.8	22.5	Chittick

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Table B4 (*continued*)

Core name	Depth (ft)	% <74 μm			Method
		Calcite (%)	Dolomite (%)	Total carbonate (%)	
Virgin	73.7	2.3	2.9	5.2	Chittick
Virgin	75.9	12.5	10.5	23.0	Chittick
Virgin	84.0	11.9	12.9	24.8	Chittick
Virgin	89.5	2.3	2.2	4.5	Chittick
Virgin	93.4	1.6	2.5	4.1	Chittick
Virgin	109.5	5.2	7.5	12.7	Chittick
Virgin	117.0	2.0	10.6	12.5	Chittick
Virgin	123.7	0.0	0.0	0.0	Chittick
Virgin	124.9	12.1	7.7	19.8	Chittick
Virgin	125.8	13.7	8.6	22.4	Chittick
Virgin	140.2	6.4	6.0	12.5	Chittick
Virgin	144.5	2.7	3.3	6.0	Chittick
Virgin	148.0	0.8	4.9	5.7	Chittick
Virgin	150.5	1.7	3.1	4.8	Chittick
Virgin	152.0	1.0	3.0	4.1	Chittick
Virgin	158.1	2.7	4.7	7.4	Chittick
Virgin	158.9	2.3	4.4	6.7	Chittick
Virgin	162.0	0.8	1.2	2.0	Chittick
Virgin	170.0	0.0	0.0	0.0	Chittick

[1]NA, not applicable; only total carbonate determined by capture of CO_2 during dissolution.

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