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Geologic Controls on Plio-Pleistocene Drainage Evolution of the Kentucky River in Central Kentucky

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GEOLOGIC CONTROLS ON PLIO-PLEISTOCENE DRAINAGE EVOLUTION OF THE KENTUCKY RIVER IN CENTRAL KENTUCKY

ABSTRACT OF DISSERTATION

A dissertation submitted in partial fulfillment of the
Requirements for the degree of Doctor of Philosophy in the
College of Arts and Sciences
At the University of Kentucky

By
William M. Andrews Jr.

Lexington, Kentucky

Director: Dr. William A. Thomas, Professor of Geological Sciences

Lexington, Kentucky

2004

ABSTRACT OF DISSERTATION

GEOLOGIC CONTROLS ON PLIO-PLEISTOCENE DRAINAGE EVOLUTION
OF THE KENTUCKY RIVER IN CENTRAL KENTUCKY

The primary goal of this project is to develop a relative chronology of events in the geologic history of the Kentucky River, and to consider the geologic controls on those events. This study utilized published geologic and topographic data, as well as field observations and extensive compilation and comparison of digital data, to examine the fluvial record preserved in the Kentucky River valley in central Kentucky. Numerous fluvial features including abandoned paleovalleys, fluvial terraces and deposits, bedrock benches, and relict spillways between adjacent river valleys were identified during the course of the study.

The morphology of the modern valley coincides with bedrock lithology and can be used to describe the distribution and preservation of modern and ancient fluvial deposits and features in the study area. Bedrock lithology is the dominant control on valley morphology and on the distribution and preservation of fluvial deposits and features in the study area. Some stream trends are inherited from the late Paleozoic drainage of the Alleghanian orogeny. More recent inheritance of valley morphology has resulted from the erosion of the river from one lithology down into another lithology with differing erosional susceptibility, thus superposing the meander patterns of the overlying valley style onto the underlying lithology.

One major drainage reorganization related to a pre-Illinoisan glacial advance disrupted the northward flow of the Old Kentucky River toward the Teays River system and led to organization of the early Ohio River. This greatly reduced the distance to base-level, and led to abrupt incision and a change in erosional style for the Kentucky River.

The successful projection of valley morphologies on the basis of bedrock stratigraphy, the history of erosion suggested by fission track data and the results of this study, as well as soil thickness and development, all argue against the existence of a mid- to late-Tertiary, low-relief, regional erosional surface. This study instead hypothesizes that the apparent accordance of ridge-top elevations in the study area is a reflection of a fluvially downwasted late Paleozoic depositional surface.

KEYWORDS: Quaternary geology, fluvial geomorphology, landscape evolution, Kentucky River Palisades, Teays River

William Morton Andrews Jr.

June 15, 2004

GEOLOGIC CONTROLS ON PLIO-PLEISTOCENE DRAINAGE EVOLUTION
OF THE KENTUCKY RIVER IN CENTRAL KENTUCKY

By

William Morton Andrews Jr.

William A. Thomas
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June 15, 2004

RULES FOR THE USE OF DISSERTATIONS

DISSERTATION

William Morton Andrews Jr.

The Graduate School

University of Kentucky

2004

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To Dr. John C. Ferm (1925-1999) and Dr. Nicholas Rast (1927-2001):

I never got to say this before: your words and teachings have served me well. Thank you.

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CHAPTER ONE: INTRODUCTION AND BACKGROUND

Overview

The primary goal of this project is to develop a relative chronology of events in the Pliocene and Pleistocene geologic history of the Kentucky River, and to consider the implications for broader understanding of regional geologic history, landscape evolution, and controls on this fluvial system. The Kentucky River is a significant tributary of the modern Ohio River, and was a major tributary of the ancient Teays River system, the northwest-flowing precursor of the modern Ohio River (Teller and Goldthwait, 1991). An understanding of the geologic history of the Kentucky River could help to answer questions about the formation of the Ohio and the demise of the Teays. The Kentucky River flows across a variety of sedimentary rocks as it drains central and eastern Kentucky, and provides an opportunity to consider effects of bedrock lithology on development of the river system. A clear understanding of the evolution of the Kentucky River can also provide clues to broader mechanisms of landscape evolution and landform development.

Study Area and Setting

This study will focus on the area of the modern main stem of the Kentucky River and associated tributaries and paleo-features, from approximately Irvine to Carrollton, Kentucky (Figure 1). The geographic extent of the study is constrained by the distribution and reliability of identification of ancient fluvial deposits in central Kentucky. Many of these deposits are identified through the presence of quartz pebbles presumably derived from Lower Pennsylvanian and younger conglomeratic sandstones that are exposed along the western outcrop belt of Pennsylvanian strata in eastern Kentucky and farther east. Similar quartz pebbles are not found in older Paleozoic strata exposed in central Kentucky.

Physiography

The Kentucky River basin crosses parts of the Cumberland Plateau and Bluegrass regions of Kentucky (Figure 2). The Cumberland Plateau consists of relatively steep, high-relief hills and ridges on Pennsylvanian shales and sandstones. Relief and average slopes increase to the southeast. The Cumberland Escarpment separates the Cumberland Plateau from the topographically lower Bluegrass region to the west. The Bluegrass can be separated into three

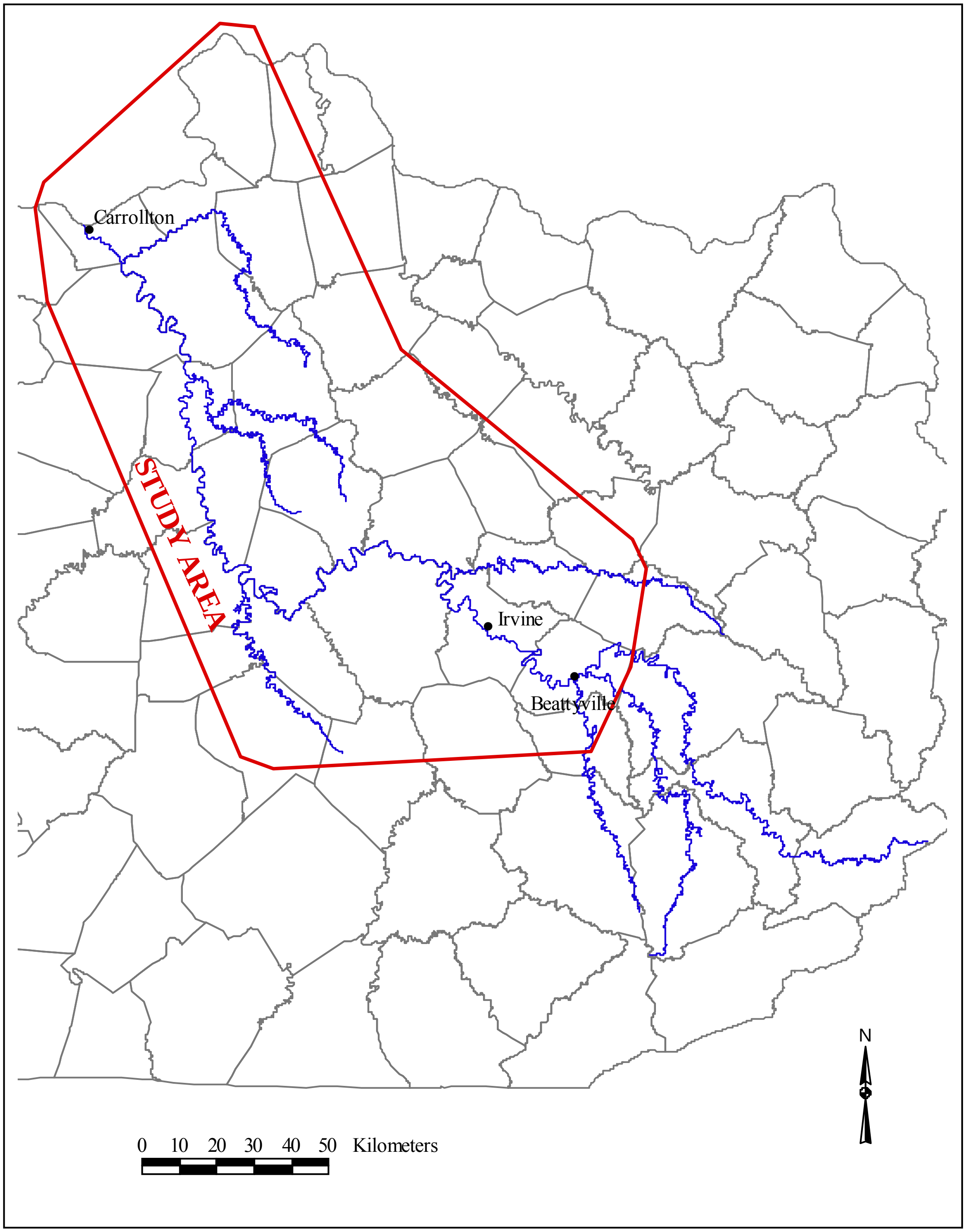


Figure 1. Map of the Kentucky River and key tributaries, showing the area of consideration in the current study.

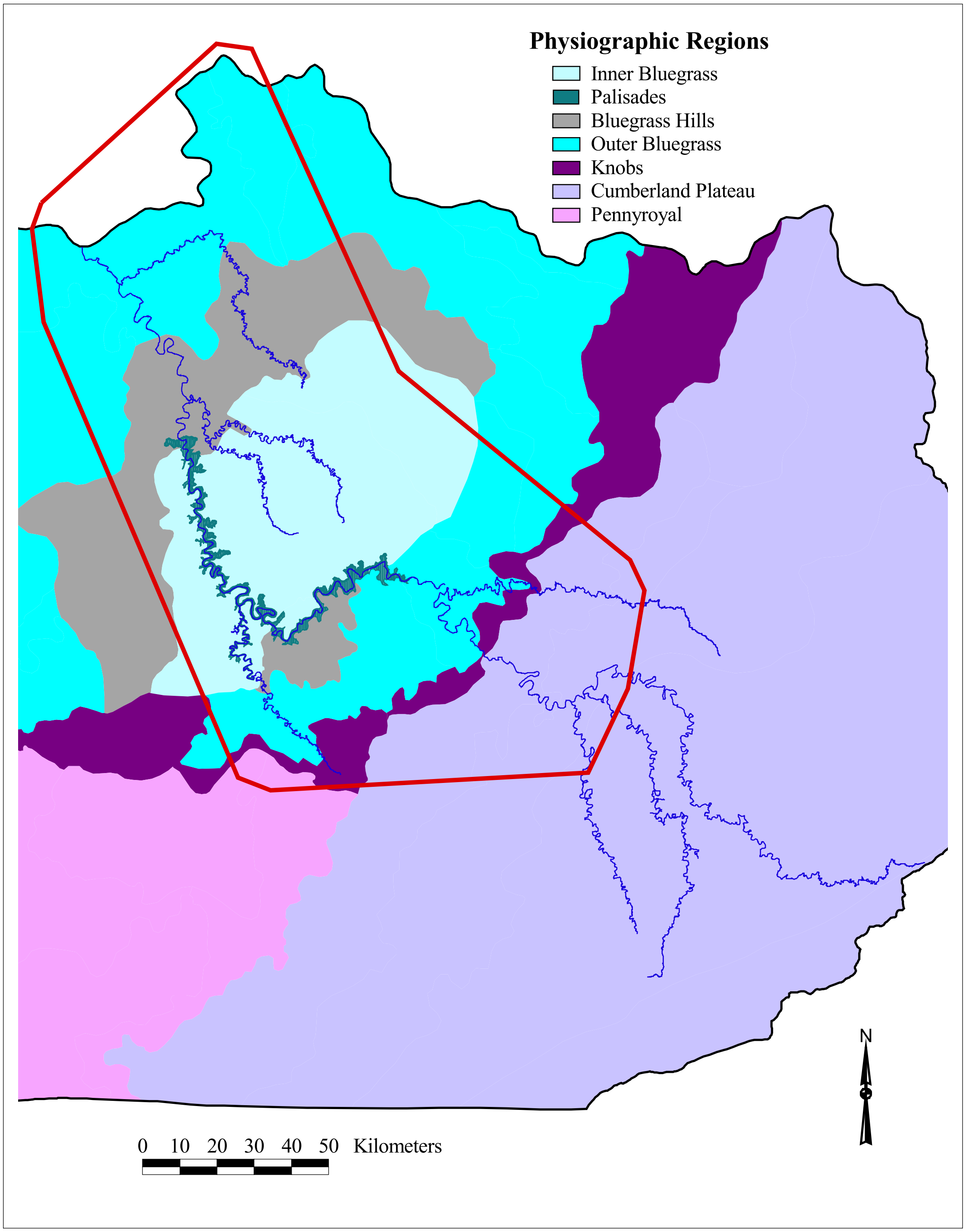


Figure 2. Physiography of the study area.

general subregions, on the basis of topographic characteristics (McFarlan, 1943). The Inner Bluegrass consists of gently rolling hills, with extensive karst in Upper Ordovician limestones. Where it crosses the Inner Bluegrass, the Kentucky River has eroded a steep gorge known as the Palisades (Figure 3a). The Bluegrass Hills region (Eden Shale Belt of McFarlan, 1943) is a belt of steep hills surrounding the Inner Bluegrass. The Bluegrass Hills are in areas dominated by the Clays Ferry and Kope Formations. The Outer Bluegrass topography is more gently rolling than the Bluegrass Hills, but less so than the Inner Bluegrass. The Outer Bluegrass region is on limestones and shales of Late Ordovician age. Around the circumference of the Bluegrass region in Kentucky is a belt of conical hills, known as the Knobs, on Devonian and Mississippian rocks (Figure 3b) (McFarlan, 1943). The Knobs are erosional outliers of the Pennyroyal to the south and the Cumberland Plateau to the east. The Pennyroyal, on Mississippian limestones, is a plateau intermediate in elevation between the Cumberland Plateau and the Bluegrass region.

Climate and Original Landcover

The modern climate of Kentucky can be classified as humid subtropical (Ulack and others, 1998). The average annual temperature is 54°F to 58°F (12.2°C to 14.5°C). The state receives an average of 45 to 50 inches (114 to 127 cm) of precipitation per year, including 10 to 12 inches (25 to 30 cm) of snow. Prior to European settlement, the vegetation of Kentucky was dominated by deciduous forests with local savannas (Woods and others, 2002). Pollen records indicate that during previous, colder Late Pleistocene climates, pine and spruce dominated the native vegetation (Wilkins and others, 1991).

Stratigraphy

From 1960 to 1978, the joint Kentucky Geological Survey (KGS) and U.S. Geological Survey (USGS) Geologic Mapping Program provided complete 1:24,000-scale geologic mapping for the entire state of Kentucky. The program provided published paper geologic maps for each 7.5-minute USGS quadrangle, as well as a comprehensive update of the stratigraphy and lithology. Cressman and Noger (1976), Cressman (1973), Weir and others (1984), McDowell (1983), Ettensohn and others (1984), Rice and Others (1979) and Rice (1984) summarized the stratigraphic data for the study area. McDowell (1986) summarized the stratigraphic results in a text report to accompany a 1:250,000-scale geologic map of Kentucky (McDowell and others, 1981) derived from the 1:24,000-scale mapping. A more recent program of the KGS (1996 to

A

B

Figure 3. Photographs of physiographic features in central Kentucky. A: View of the Palisades of the Kentucky River at Camp Nelson. B: The wooded hills on the horizon are the Knobs on the southern border of the Bluegrass in Boyle County.

present), funded by the USGS StateMap Program, is digitizing the original 1:24,000 geologic maps to enable use with Geographic Information System (GIS) software. The digitized geologic maps are available for the entire study area, and were used extensively in this study. Appendix 1 summarizes the geological quadrangles and Appendix 2 summarizes the digital datasets used in this study.

Strata exposed in the Kentucky River basin range from Late Ordovician to Middle Pennsylvanian in age. Figure 4 shows the distribution of the different rock units in the Kentucky River basin. Figure 5 shows the generalized stratigraphic column for the study area. Except where noted, the following descriptions are summarized from discussions in McDowell (1986).

High Bridge Group

The Upper Ordovician High Bridge Group is the oldest stratigraphic unit exposed in Kentucky, and contains three formations. From oldest to youngest, they are the Camp Nelson Limestone, the Oregon Formation, and the Tyrone Limestone. The Camp Nelson Limestone consists of thick-bedded to massive, fine-grained, dolomitic limestone (Figure 6a). The Camp Nelson is not completely exposed, but 320 feet (97.5 m) can be measured near Camp Nelson. Shale content is minimal, except for a zone within 20 feet (6 m) of the top of the formation. The Oregon Formation is characterized by thick-bedded dolostone interbedded with fine-grained limestone. The thickness varies across the area, reaching a maximum of 65 feet (20 m) near Boonesboro. The thickness variation is reciprocal with the overlying Tyrone Limestone. The Tyrone ranges from 55 to 155 feet (17 to 47 m) thick, and is characterized by thick-bedded, micritic limestone (Figure 6b). The unit contains a significant quantity of shale in the upper half, including at least two thick and several thinner bentonite layers. The bentonites are clay-rich and thus relatively easy to erode, and can assist in undermining the overlying limestone beds of the Tyrone.

Lexington Limestone

The Upper Ordovician Lexington Limestone is a complex mosaic of rock units dominated by thin-bedded limestone. The formation ranges from 190 to 320 feet (58 to 98 m) thick across central Kentucky. Various members in the formation are primarily distinguished on the basis of shale content and bedding characteristics. The lower part of the Lexington Limestone is dominated by the Grier Member, which contains thin- to medium-bedded limestones with

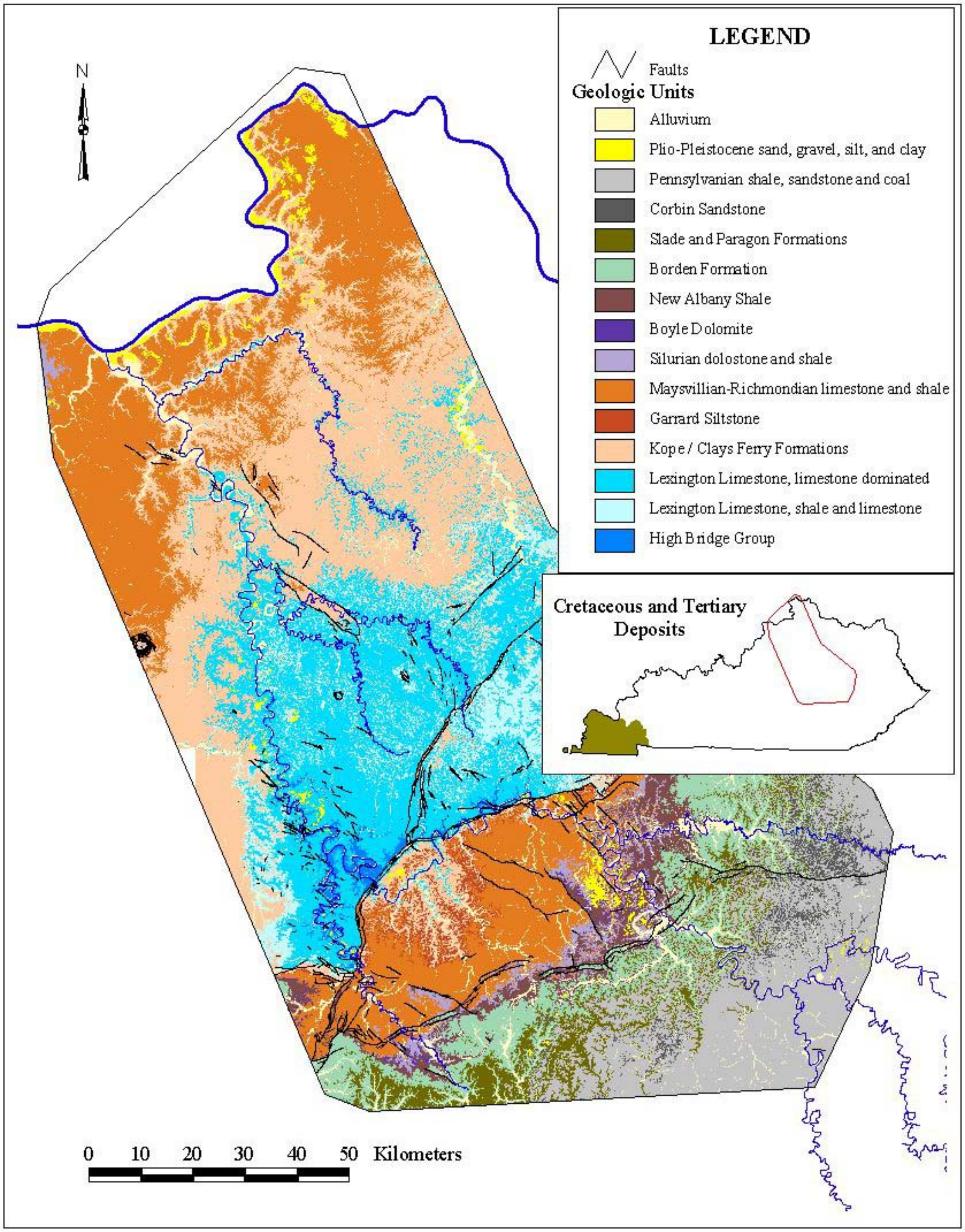


Figure 4. Geologic map of the study area, compiled from Kentucky Geological Survey digital geologic map data base.

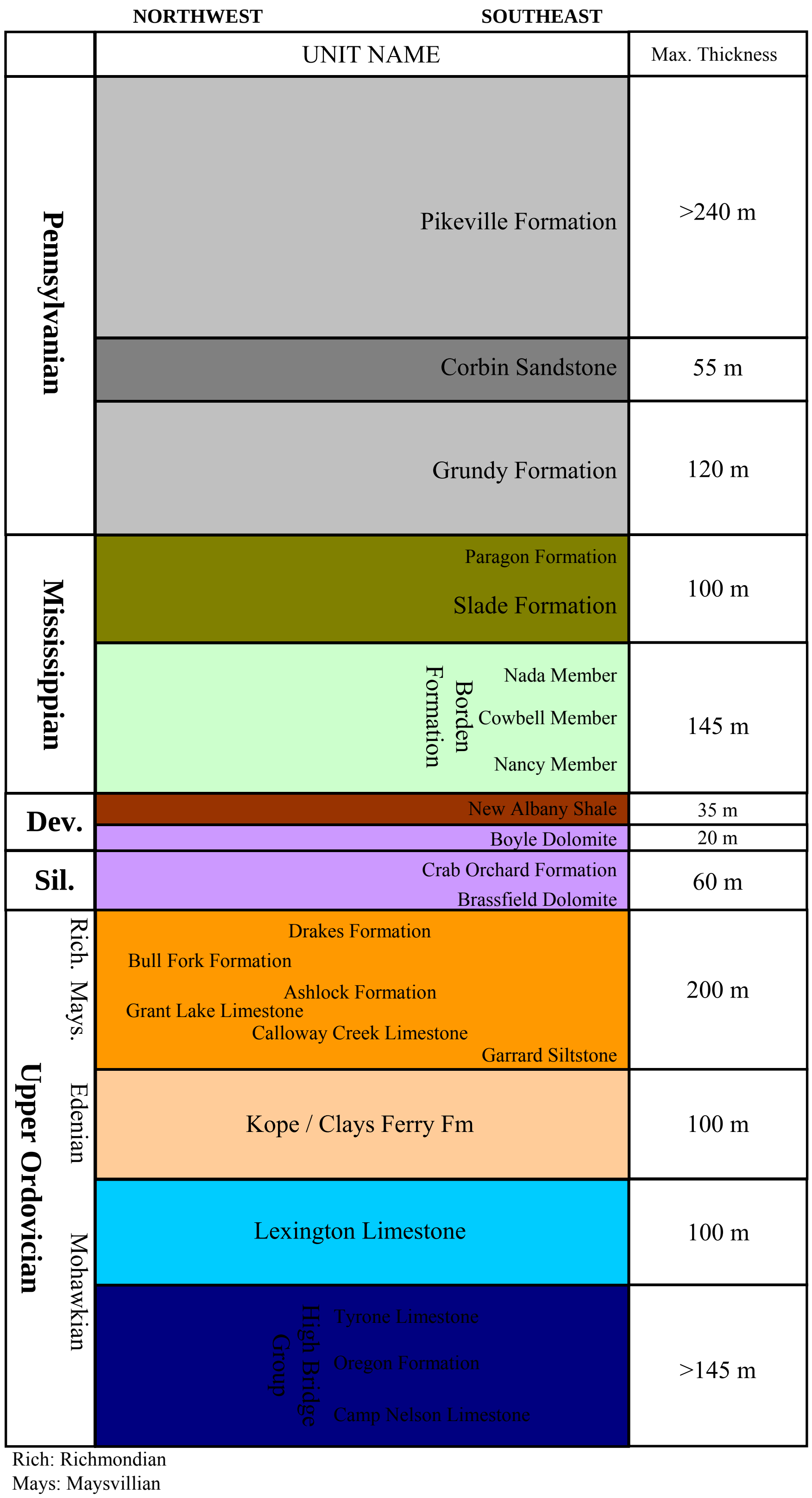


Figure 5. Stratigraphic column of bedrock units in the Kentucky River valley.

A

B

Figure 6. Photographs of High Bridge Group rocks in central Kentucky. A: Thick-bedded to massive limestone in the Camp Nelson Limestone. B: Medium- to thick-bedded limestone interbedded with thin shale partings in the Tyrone Limestone. Wooden stake is 1.2 m long.

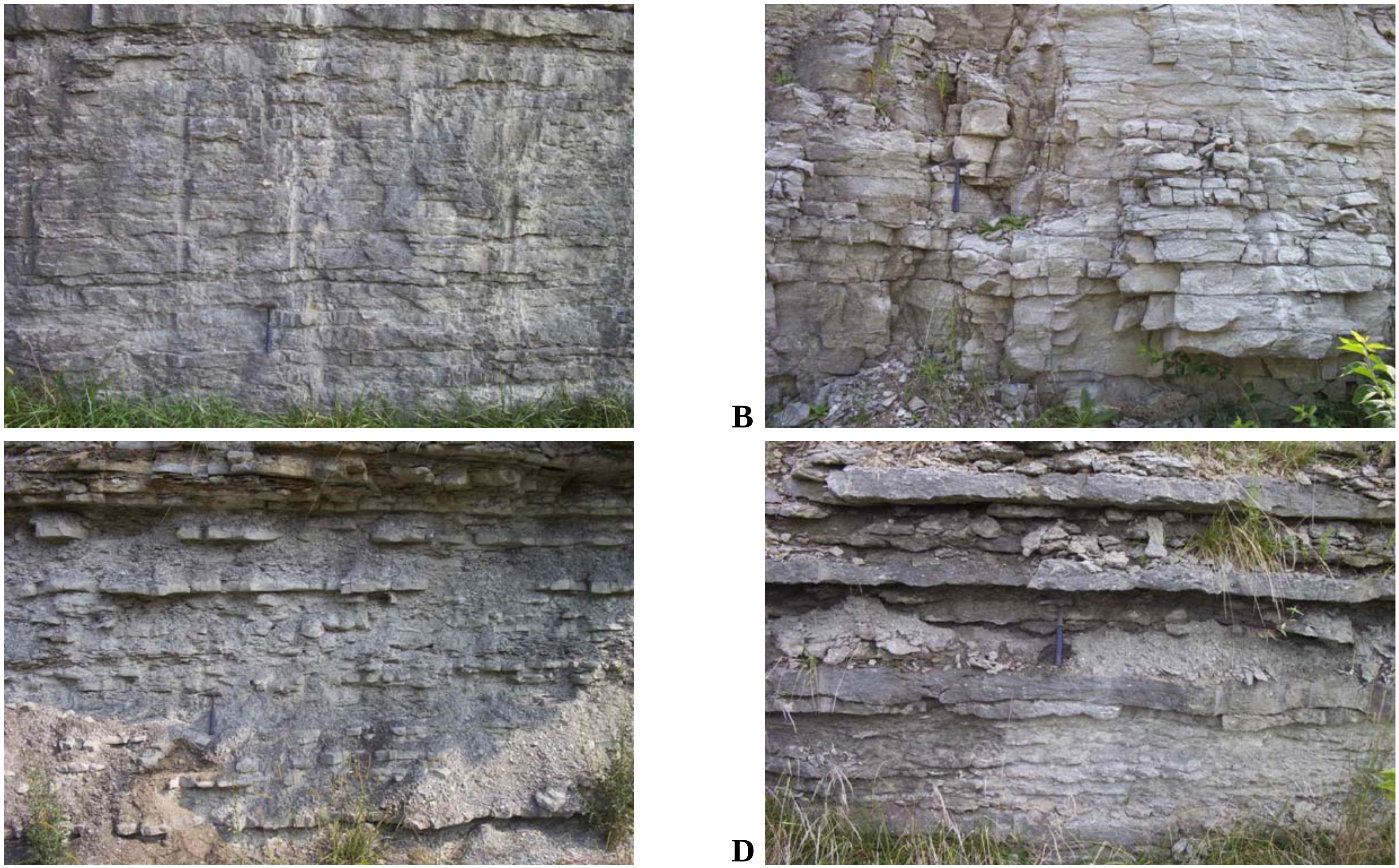
common thin shale partings (Figure 7a); it ranges from 100 to 180 feet (30 to 55 m) thick. The Tanglewood Member includes multiple zones of cross-bedded calcarenite in the upper part of the Lexington Limestone (Figure 7b). The Logana Member, near the bottom part of the Lexington, and the Brannon and Millersburg Members (Figure 7c,d) in the upper part of the Lexington contain significant amounts of shale (>30%) interbedded with thin-bedded limestone. The upper parts of the Lexington (the Brannon Member and above) intertongue laterally with the overlying Upper Ordovician Clays Ferry Formation (Kope Formation). The uppermost parts of the Lexington Limestone are likely Upper Ordovician (Edenian).

Younger Upper Ordovician Units

The Upper Ordovician strata of central Kentucky consist primarily of thin-bedded, interbedded limestone, calcareous shale, and siltstone. The units are distinguished on the basis of the dominance of a particular lithology compared to adjacent strata. Thickness of the combined Upper Ordovician section is approximately 300 m; the Kope / Clays Ferry Formation has a maximum thickness of 100 m. The Clays Ferry, Kope, and Bull Fork Formations are dominated by interbedded shale (commonly >50%) and limestone (Figure 8a). The Garrard Siltstone consists of thin- to thick-bedded siltstone interbedded with shale and limestone. The Calloway Creek Limestone contains approximately 70% thin-bedded, fossiliferous limestone interbedded with shale. The Grant Lake Limestone contains 70 to 90% thin-bedded limestone interbedded with shale. The Ashlock Formation is a complex unit of intertonguing limestone and shale. The Ashlock grades northward into the Grant Lake Limestone. The Drakes Formation contains a lower unit, the Rowland Member, dominated by thin- to thick-bedded silty dolostones and limestone. The overlying Preachersville Member in the southeastern Bluegrass contains interbedded shale and limestone and dolostone. In the western Bluegrass, the Rowland is overlain by the shaly Bardstown Member and the massive, dolomitic Saluda Member.

Silurian and Devonian Units

Silurian strata in the study area consist primarily of interbedded dolostone and shale. The basal unit, the Lower Silurian Brassfield Dolomite, consists of a lower massive dolostone overlain by thin-bedded dolostone with thin shale interbeds. The overlying Middle Silurian Crab Orchard Formation (or Group, see discussion in McDowell, 1983, 1986) consists of alternate units of clay shale and thin- to thick-bedded dolostone (Figure 8b). The Middle Devonian Boyle



A B C D

Figure 7. Photographs of Lexington Limestone rocks in central Kentucky. A: Thin-bedded limestone interbedded with shale in the Grier Member. B: Thin- to medium-bedded limestone in the Tanglewood Limestone Member. C: Interbedded shale and thin-bedded limestone in the Brannon Member. D: Interbedded thin-bedded limestone and shale in the Millersburg Member.

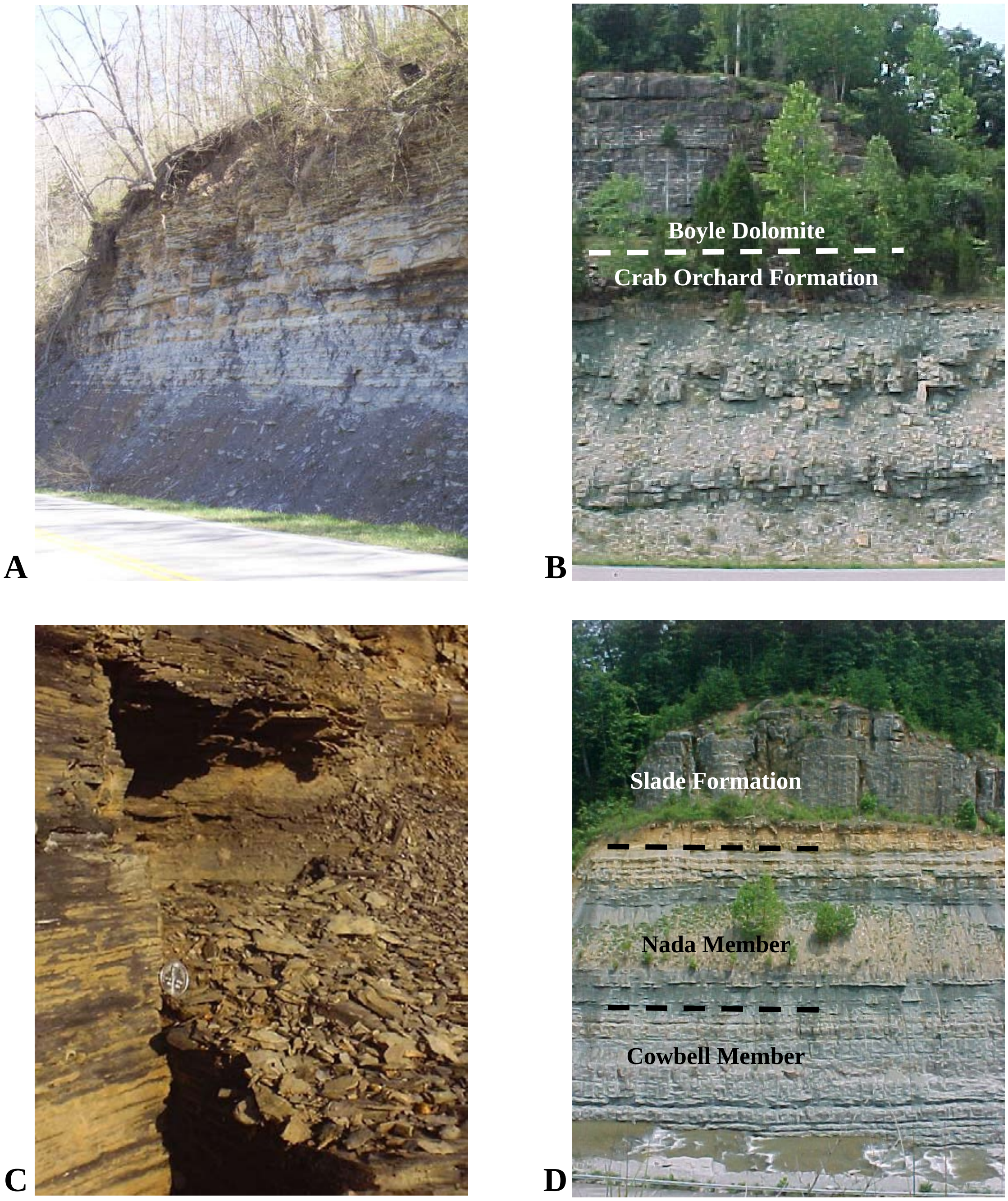


Figure 8. Photographs of Upper Ordovician, Silurian, Devonian, and Mississippian rocks in east-central Kentucky. A: Interbedded shale and thin-bedded limestone in the Upper Ordovician Clays Ferry Formation. B: Interbedded shale and dolostone of the Silurian Crab Orchard Formation overlain by massive dolostone of the Middle Devonian Boyle Dolomite. C: Black shale of the New Albany Shale (photo by Warren Anderson). D: Siltstone and shale in the Mississippian Borden Formation, overlain by limestone and dolostone in the Slade Formation.

Dolomite uncomformably overlies Middle Silurian to Upper Ordovician strata across central Kentucky. The Boyle is a thick-bedded to massive dolostone, ranging from 0 to 60 feet (0 to 18 m) thick across the area (Figure 8b). The Upper Devonian New Albany Shale consists of laminated to thin-bedded carbonaceous shale. This black, organic-rich shale typically weathers into intact plates or slabs (Figure 8c).

Borden Formation

In the study area, the Lower Mississippian Borden Formation is divided into three members. The Nancy Member is dominated by green-gray clay shale with minor amounts of silty shale and siltstone. The Cowbell Member contains medium- to thick-bedded siltstone with thin shale interbeds (Figure 8d). The Nada Member contains clay shale with sparse thin limestone interbeds (Figure 8d). The overlying Renfro Member, a thick-bedded dolostone, was mapped with the Nada Member during the KGS-USGS Geologic Mapping Program (McDowell, 1986), but has since been re-assigned to the Slade Formation and associated with overlying carbonate rocks (Ettensohn and others, 1984). Upper parts of the Borden Formation locally contain numerous geodes.

Slade and Paragon Formations

The Upper Mississippian Slade Formation is dominated by thick-bedded limestone with minor dolostone (Figure 8d), and is locally truncated or eroded beneath a Lower Pennsylvanian unconformity (Rice, 1986). The St Louis Member of the Slade contains the distinctive coral fossil *Lithostrotion sp*. The Slade Formation is the caprock on the Cumberland Escarpment in the main valley of the Kentucky River. The Upper Mississippian Paragon Formation is thin, and is present only locally in the Kentucky River basin. It contains red and green clay shale and limestone.

Pennsylvanian Units

Lower Pennsylvanian shale, sandstone and coal overlie the Mississippian strata and are at the surface throughout the headwaters region of the Kentucky River (the Three Forks of the Kentucky River). The Corbin Sandstone is a thick conglomeratic sandstone in the Lower Pennsylvanian rocks. The Corbin contains abundant quartz pebbles (Figure 9a), and is inferred to be a source of quartz gravel found in fluvial deposits downstream. The Corbin, however, is only

A

B

Figure 9. Photographs of Corbin Sandstone in eastern Kentucky. A: Slabs of conglomerate from the Corbin Sandstone, showing abundant quartz gravel. B: Prominent cliffs in Corbin Sandstone near Slade, Kentucky.

a few tens of feet thick in the main valley of the Kentucky River; much of the original thickness and extent in this area have been eroded away. In the Red River valley, however, the Corbin Sandstone is more than 200 feet (60 m) thick, and is a prominent cliff former (Figure 9b). Along much of the Cumberland Escarpment, the Corbin Sandstone or similar thick quartzose sandstones form the caprock of the Cumberland Escarpment, above the Slade Formation. The Grundy and Pikeville Formations are composed of shale, sandstone, siltstone, and coal, and are typical of the strata underlying most of the Cumberland Plateau in the headwaters region of the Kentucky River.

Cretaceous and Early Tertiary Units

Although Mesozoic through mid-Tertiary rocks have not been identified in the Kentucky River basin, significant Late Cretaceous through Early Oligocene deposits are found in the Jackson Purchase of far western Kentucky (Figure 3, inset). The Jackson Purchase region represents the northernmost extension of the Gulf Coastal Plain. The unconsolidated deposits of gravel, sand, silt, and clay were deposited in a complex facies mosaic of fluvial, deltaic, lacustrine, lagoonal, and marine environments (Figure 10) (Olive, 1980). These deposits are significant to the current study because some of the units are interpreted as marine in origin, and thus suggest that sea-level (global base level) was in or near far-western Kentucky from the Late Cretaceous until the Late Eocene. The close proximity of sea level would preclude deep fluvial incision below that level, and thus discourage extensive fluvial erosion and denudation in adjacent areas.

Late Tertiary and Quaternary

The youngest sedimentary deposits identified in Kentucky are of Late Tertiary to Quaternary in age. The KGS-USGS mapping program and earlier workers identified fluvial, glacial, glacio-fluvial, lacustrine, and eolian sediments mantling bedrock and older unconsolidated sediments across the state. Older fluvial sediments are inferred to be Miocene(?) to Pleistocene in age, and are represented by gravel, sand, silt, and clay in terrace and paleochannel deposits at various distances above the elevations of modern streams. Younger fluvial deposits include gravel, sand, silt, and clay in terraces and alluvium in valley bottom settings along most streams in the study area. Glacial deposits include till and drift deposits found in northern Kentucky. Older till, inferred to be pre-Illinoisan in age, has been mapped

A

B

Figure 10. Photographs of Cretaceous and Tertiary sediments in the Jackson Purchase region. A: Gravel deposit in the Cretaceous Tuscaloosa Formation. B: sand, clay, and lignite in the Eocene Claiborne Formation, overlain by brownish-orange late Tertiary fluvial gravel (photo by Brandon Nuttall).

along uplands adjacent to the Ohio River from Oldham County to Pendleton County. Illinoisan till has been identified in and near the Ohio River valley in the same area. Wisconsin till has not been identified in Kentucky. Glacio-fluvial deposits include pre-Illinoisan outwash preserved in paleovalleys in Boone, Gallatin, and Carroll Counties, as well as Illinoisan and Wisconsin outwash, which comprise most of the valley-fill sediments in the Ohio River valley. Lacustrine deposits are mapped primarily in Ohio-River tributaries that were impounded by rapid alluviation and high flow volumes of Illinoisan and Wisconsin glacial outwash in the main Ohio River valley. Eolian deposits include sand dunes and loess derived from outwash deposits in the Ohio, Wabash, and Mississippi River valleys. Sand dunes are restricted mainly to the southern valley margins of the Ohio River valley. Loess of various thicknesses blankets much of the uplands south of the Ohio River. Further discussion of the Late Tertiary and Quaternary deposits of the study area is provided in Chapter 2.

Structural Geology

The study area straddles the Cincinnati arch, a broad, north-northeast trending, regional anticline separating the Appalachian basin on the east from the Illinois basin on the west (McDowell, 1986). The Jessamine dome is a relatively high area along the crest of the Cincinnati arch, and lower Upper Ordovician rocks are exposed at the surface. Progressively younger rocks are preserved away from the dome in all directions. The Appalachian basin, which includes eastern Kentucky, is a composite foreland basin, activated several times during the Paleozoic by loading during mountain-building events in the Appalachian orogen to the east (Colton, 1970; Milici and de Witt, 1988; Rast, 1989). Pennsylvanian rocks are exposed at the surface in the Appalachian basin in Kentucky. The Illinois basin is a continental interior basin, also with Pennsylvanian rocks at the surface. Strata in the Appalachian basin in eastern Kentucky dip to the southeast, whereas those in the Illinois basin in north-central Kentucky dip westward. Pine Mountain, in far southeastern Kentucky, is the northwesternmost exposed thrust fault of the final Appalachian orogeny, the Alleghenian.

The Lexington fault system trends north-northeast along the crest of the Cincinnati arch (Figure 11). Near Camp Nelson, the fault system juxtaposes younger Upper Ordovician rocks against High Bridge strata. The Kentucky River fault system branches northeast from the Lexington fault system; this fault system also brings High Bridge rocks to the level of the Upper

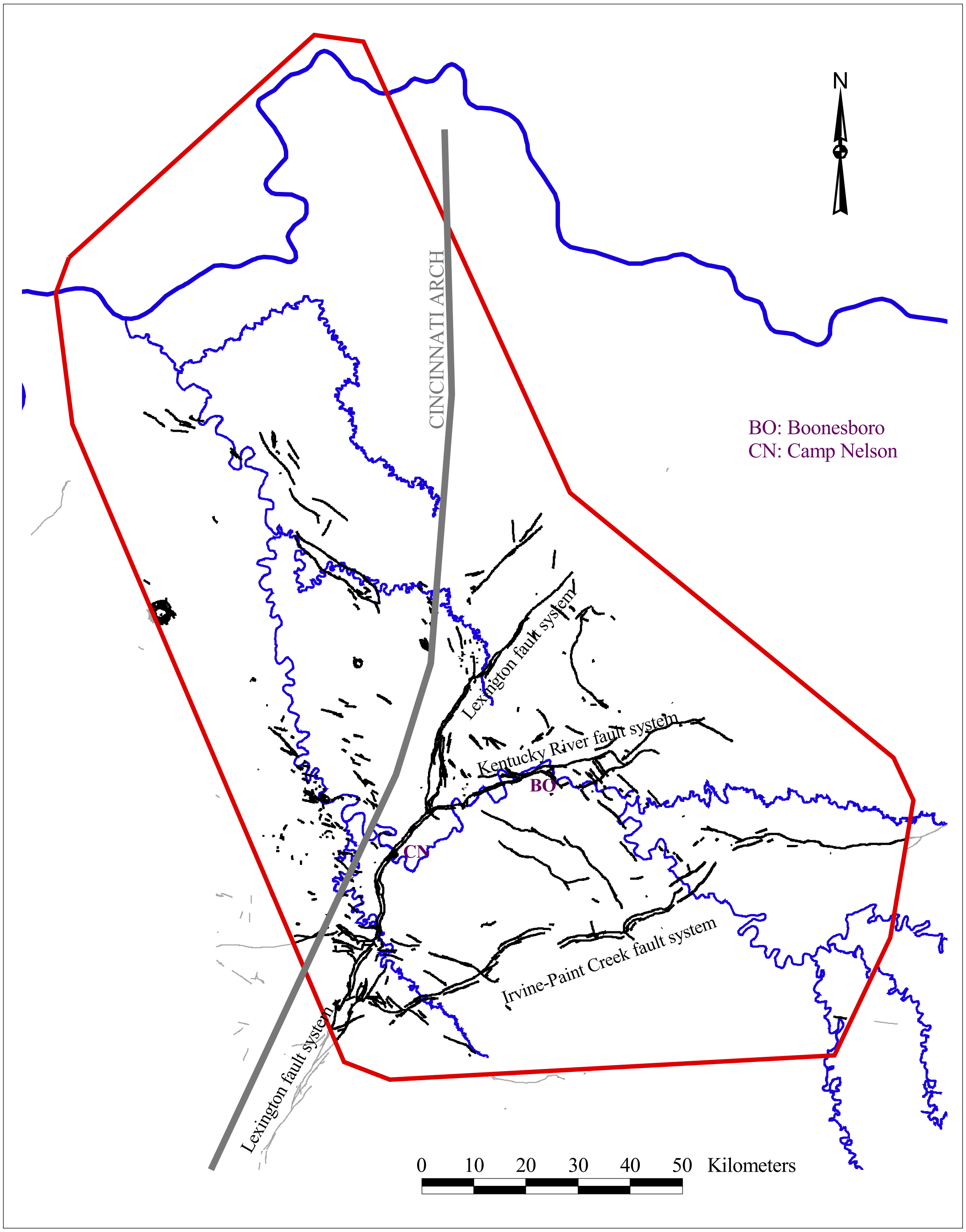


Figure 11. Major structural features in central Kentucky.

Ordovician Clays Ferry Formation. The Kentucky River generally parallels the Kentucky River fault system between Boonesboro and Camp Nelson. The Irvine Paint Creek fault system has approximately 200 ft (60 m) of down-to-the-south offset, and roughly parallels the Kentucky River fault system (Figure 11). Numerous smaller faults are mapped through the basin, many with a general northwest trend.

Structural and Stratigraphic Profile

The stratigraphy and structure along the longitudinal profile of the Kentucky River valley are shown in Figure 12. The elevations of stratigraphic tops were collected from geologic quadrangle maps along the entire length of the valley at 1 to 3 kilometer intervals. In general, the rock units dip away from the axis of the Cincinnati arch, the top of which is shown on Figure 12 approximately 180 km from the mouth of the Kentucky River. Fault offsets accommodate a significant portion of the dip, especially along the eastern flank of the arch. The reversed dips, toward the axis of the arch, are an artifact of the valley meanders. Locally, fault offsets introduce significant lithologic contrasts along the valley.

Background

Fluvial Processes

The mechanical process of fluvial incision is a key consideration in discussing the evolution of a fluvial system through time. Much of the modern Kentucky River is flowing over a thin veneer of unconsolidated fluvial deposits. However, to incise the existing valley, the river must have eroded through the native bedrock, rather than through unconsolidated materials. In rivers eroding into bedrock (“bedrock streams”), the key processes that have been identified are quarrying and abrasion; cavitation and dissolution also play important roles. Both bedrock-erosion processes require the stream flow to have access to the bedrock surface, and thus any overlying sediment must be removed before bedrock erosion occurs. Although multiple variables are involved in the erosion processes, they all track with flow velocity, so it can be considered that the velocity of water flow near the rock surface primarily controls both processes (Hancock and others, 1998).

Of the two bedrock-erosion processes, quarrying is the most effective mechanism where bedding planes or joints are spaced closely enough together and optimally oriented (Miller,

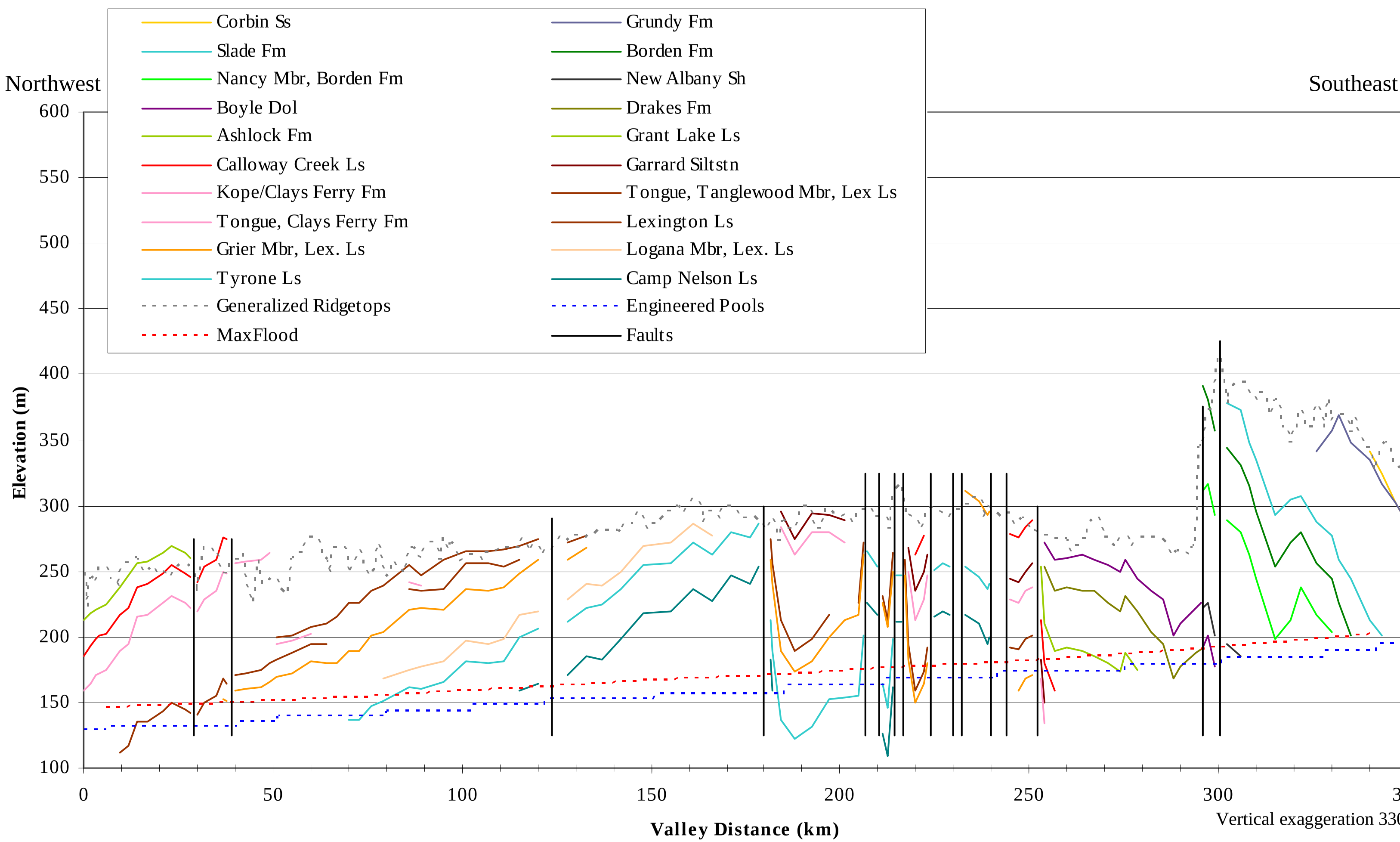


Figure 12. Stratigraphic and structural profile along the Kentucky River valley. Lines represent tops of stratigraphic units.

1991). Before blocks can be moved, they must be geomorphically "prepared" for removal by enlargement, weathering, or weakening of bounding surfaces of the block. When a threshold velocity of flow is surpassed, quarrying of properly prepared blocks can progress (Figure 13). Flows greatly exceeding the threshold will be able to quarry significantly larger blocks. Areas of constricted channel width and knickpoints will be areas of increased water-surface slope, and thus focused locations of bedrock quarrying (Hancock and others, 1998). Bedrock with very thick bedding and/or widely spaced joints produces blocks too large for streams to remove by quarrying, and limits the effectiveness of the process.

Abrasion results from the impact of entrained sediment grains upon the bedrock surface, dislodging a fraction of the impacted bedrock. The process, by definition, requires the presence of a significant sediment load in the stream (Hancock and others, 1998). Streams with minimal sediment load are not able to effectively abrade underlying bedrock. Because of channel-margin friction, flow velocities immediately adjacent to the bedrock surface are typically quite low, so enough momentum must be acquired in the main flow, and the sediment separated from the main flow, for grains to impact the bedrock surface with enough kinetic energy to dislodge parts of the bedrock (Hancock and others, 1998). As such, only very high flows are effective agents of abrasion. Abrasion is likely a punctuated process; significant erosion occurs only during maximum flows.

In addition to quarrying and abrasion, slaking and dissolution may also contribute to bedrock erosion. Discussions of the quarrying and abrasion processes focus upon resistant lithologies such as limestone and sandstone. Shale, however, comprises a significant component of the stratigraphy in the study area, and must be considered. Three different distinct shale types can be generalized from the stratigraphy of central Kentucky: clay shale, calcareous shale, and black (organic-rich) shale. Clay shales typically have a very low slake durability, thus implying they mechanically erode very easily in the presence of water. Although clay has a high cohesion, aggregates of clay minerals can be excavated in a way similar to that of larger and less cohesive silt and sand grains. Calcareous shales, some with carbonate contents as much as 70% (Fisher, 1968) but with enough clay to behave somewhat like clay shales, are susceptible to erosion by combination of dissolution and mechanical/slaking erosion. The black shales in Kentucky have high organic contents, and weather very differently from the clay and calcareous shales. Some

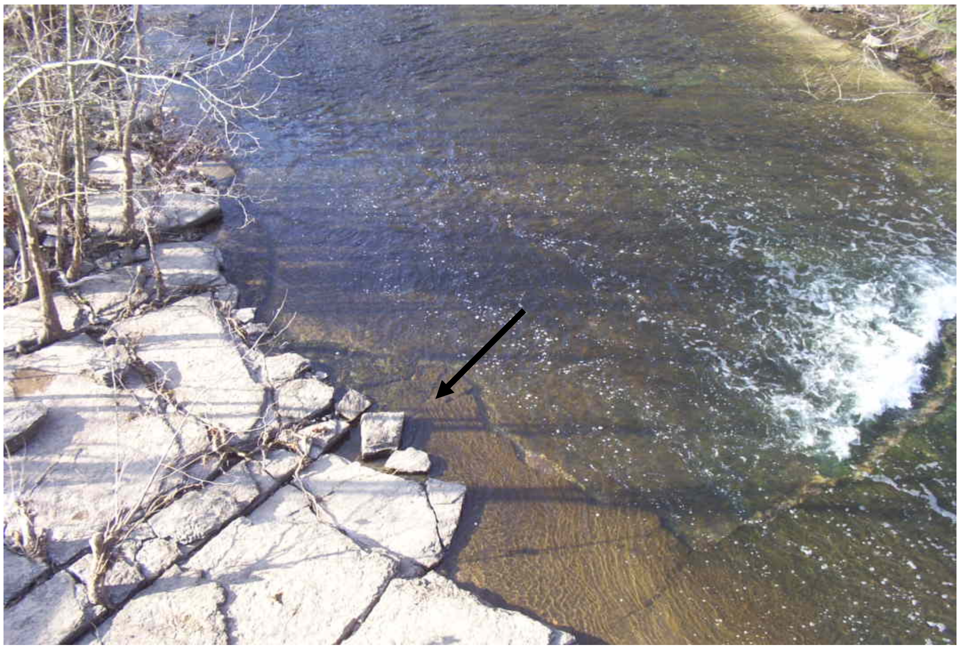


Figure 13. Photograph of fluvial quarrying of bedrock along Hickman Creek in eastern Jessamine County (Highway 1541 bridge). The stream is flowing over a knickpoint in the Middle Ordovician Oregon Formation at this site. Block indicated by arrow is approximately 1 m wide.

parts of the black shales tend to erode by quarrying into chips, plates, and slabs, and have somewhat higher slake durability than clay shales.

Interbeds of shale with more resistant rock greatly facilitate preparation of blocks for quarrying as the intervening shale is more readily eroded (by slaking) than the resistant beds. The rapid excavation and removal of shale leaves the intervening resistant layers more exposed and susceptible to quarrying in less time than in relatively "pure" resistant rocks. Thus, characteristics of the bedrock—especially bedding thickness, joint spacing, and shale content—in conjunction with the velocity of the stream flow, play key roles in the erosion of bedrock by streams.

Average velocity in a stream varies with stream slope and discharge, among other variables. Even in relatively low-slope areas, high discharges result in relatively high velocities (Knighton, 1998). Drainage area is a reliable proxy for relative discharge estimations in areas of similar hydrologic response (Knighton, 1998). Velocity is not, however, uniformly distributed within a channel profile, and is locally susceptible to numerous variables. Friction with the stream bed slows velocities near the water-bed interface. In relatively straight stretches of streams where channels are relatively symmetrical, the highest velocities are concentrated in the center of the channel, and close to the surface (Knighton, 1998). In asymmetric channels, such as those in the bends of meandering streams, flow velocities are highest toward the outside of the bend, because the momentum of the flow carries it closer to the outside bank (Knighton, 1998). The sharp velocity gradients along the outer banks of meandering streams cause greater turbulence and this results in lateral erosion being a major component of the erosional activity of the stream.

Base level is identified as the elevation to which a stream is ultimately flowing, whether a lake, ocean, or other larger body of water. Different scales of base level can be identified, from a global base level (mean sea level) to localized base levels (lakes, larger streams, ponds, etc). A stream only immediately responds to the next local base level downstream.

Ideally, a fluvial system will evolve over time to produce an equilibrium longitudinal profile that is concave upward with an exponentially decreasing slope from headwaters to mouth. This ideal profile is sometimes referred to as the "graded" profile of the stream (Gilbert, 1877; Mackin, 1948). The slope of this curve is inversely proportional to the discharge of the stream. In areas with minimal variation in climate or precipitation delivery, drainage area can serve as a general proxy for potential discharge. Average stream slope, ideally, would thus decrease with progressive increase in contributing drainage area. Change in base level, change in discharge, or adjustments to the course or drainage area of the stream (through the process of stream capture) will result in an adjustment of the longitudinal profile, either through erosion (degradation) or sedimentation (aggradation) in the stream valley.

Sharp deflections from the ideal longitudinal profile, where distinct convexities exist in the longitudinal profile, are called knickpoints. Knickpoints primarily form by one of four different mechanisms: upstream propagation of adjustments to base-level, differential erosion of bedrock in the stream bed, localized neotectonic offsets, and sites where coarser bedload are introduced into the stream (Seidl and others, 1994). A knickpoint may develop where lithologies of contrasting erodibility are exposed in the stream channel. Resistant lithologies will erode more slowly than softer lithologies, producing a convex deflection in the stream profile as the channel downstream erodes more swiftly into softer bedrock. Knickpoints may develop where an abrupt geomorphic disruption (e.g. introduction of a large volume of sediment from a tributary) to the stream equilibrium produces a localized steepening of the stream. Introduction of coarse bedload into a stream channel from a tributary may create a knickpoint as the stream adjusts to the coarse deposit in the channel. The knickpoint will migrate upstream, and the gradient below the knickpoint will gradually become shallower, until the knickpoint merges into a new equilibrium profile somewhere upstream. Neotectonic offset of the stream bed can also cause development of a knickpoint by uplifting one side of a fault, or by juxtaposing contrasting lithologies in the stream bed.

Colluvial Processes

Fluvial erosion is a relatively localized process, and works in conjunction with colluvial processes to modify the landscape. As fluvial erosion deepens a valley and steepens the adjacent slopes, colluvial activity works to reduce the angle of the oversteepened slopes (Easterbrook, 1999). Bedrock lithology determines the style of colluvial activity (Ritter and others, 1995). Bedrock units with thick resistant beds usually degrade by rock falls enabled by sapping of underlying soft shaly strata. Slopes on shale-dominated stratigraphy degrade by slumping and creep processes. Slope angles decrease gradually until a stable angle of repose is reached, whereupon slope-wash processes become dominant. Slope wash gains effectiveness with increasing contribution area and slope gradient (Knighton, 1998), and thus, in the long term, will tend to not greatly reduce slopes but more likely cause them to retreat parallel or steepen, reactivating the colluvial processes. The toe of a colluvial slope can gradually become armored with colluvial debris, and thus limit further mass wasting activity (Ritter and others, 1995), unless a mechanism such as fluvial erosion can "clean" the slope toe and remove the colluvial debris from the system. Nevertheless, significant slope retreat would typically require fluvial erosion of the adjacent valley to maintain the retreat process.

Karst

The exposure of numerous limestone-dominated stratigraphic units in the study area has led to significant karst development in the soluble lithologies. Karst is significant for this study, because karst processes operate differently from fluvial or colluvial processes to erode or modify landscapes. Fluvial erosion may be an agent of very rapid denudation; local fluvial incision rates as great as 600 m/m.y. have been documented in the eastern United States (Stanford and others, 2002). However, fluvial erosion is dependent on stream flow, is commonly most effective only under high-energy conditions, and is thus episodic and spatially concentrated. Karst dissolution may operate almost continuously in low- to high-energy situations in humid climates. Landscape denudation through karst dissolution is largely controlled by climate, and more specifically by precipitation rates (Jennings, 1985; White, 1988). Soil cover and vegetation help to accelerate karst dissolution processes (Trudgill, 1977). Karst denudation rates have been estimated to range from 35 to 40 m/m.y. (Jennings, 1985; White, 1988). Although karst dissolution is a slower method of landscape denudation in the short term, the more continuous nature of karst dissolution may have significant impact in landscape modification (Simms, 2004). In tectonically stable areas, karst dissolution may progressively bevel a landscape to a low-relief corrosion plain (Smart and others, 1986). Karst conduit development also can divert surface flow underground, and thus reduce the surface flow available for fluvial erosion. Underground conduit systems may also capture surface drainage and reroute surface flow under surface drainage divides.

Karst in central Kentucky is concentrated in areas where limestone-dominated stratigraphic units are exposed at the surface, especially the High Bridge Group and Lexington Limestone (Figure 14). Thrailkill (1982) provided a review of karst distribution and processes in the central Kentucky area. Sinkhole distribution in the area north of Lexington displays a distinctly linear pattern, and is likely controlled by bedrock joints (Thrailkill, 1982). Sinkholes follow more meandering paths south of this area, and may reflect geomorphically controlled development in old valley systems. Across much of the Bluegrass region, karst conduit systems are shallow and nearly horizontal, draining to and graded to larger streams in the area. Smaller karst systems are stratigraphically controlled, perched above less soluble lithologies. Closer to the modern Kentucky River valley, the karst systems have a more vertical aspect, with steep pits and drops, compared to more horizontally oriented systems away from the deeply entrenched valley (Thrailkill, 1982).

Landscape Evolution

Two fundamentally different landscape-evolution models were classically applied to the study of landscapes in the eastern U.S. One followed the model of Davis (1902, 1909) in hypothesizing a cyclic pattern of landscape evolution. This model proposed that a landscape begins as an uplifted surface of relatively low relief, and that as it matures, it will be progressively dissected. Once the landscape is thoroughly dissected, slopes will progressively flatten and retreat from stream valley axes. Eventually a regional, low-relief, erosional landscape is graded to sea-level or a regional base-level. This surface would bevel the landscape regardless of underlying lithology. Davis (1902) termed this regional erosional surface a peneplain. If base-level changes, the surface is "reactivated" or "rejuvenated," and incision begins to dissect the peneplain, restarting the cycle. The implication would be a record of geomorphic history in the planation surfaces and subsequent rejuvenations; each peneplain would record the end of a long period of stability. Phillips (2002), among others, noted that peneplain development requires a prolonged interval of tectonic, climatic, and base-level stability.

Although the term "peneplain" technically refers to any nearly flat surface, the term should primarily be associated with the genetic implication of the Davis model. As other terms exist to describe low-relief surfaces (pediments, stripped surfaces, etc) without resorting to the term "peneplain," this term as used in this report will refer only to the regional erosional surface that is the mature end stage of the Davis model, following the assertion of Thornbury (1954):

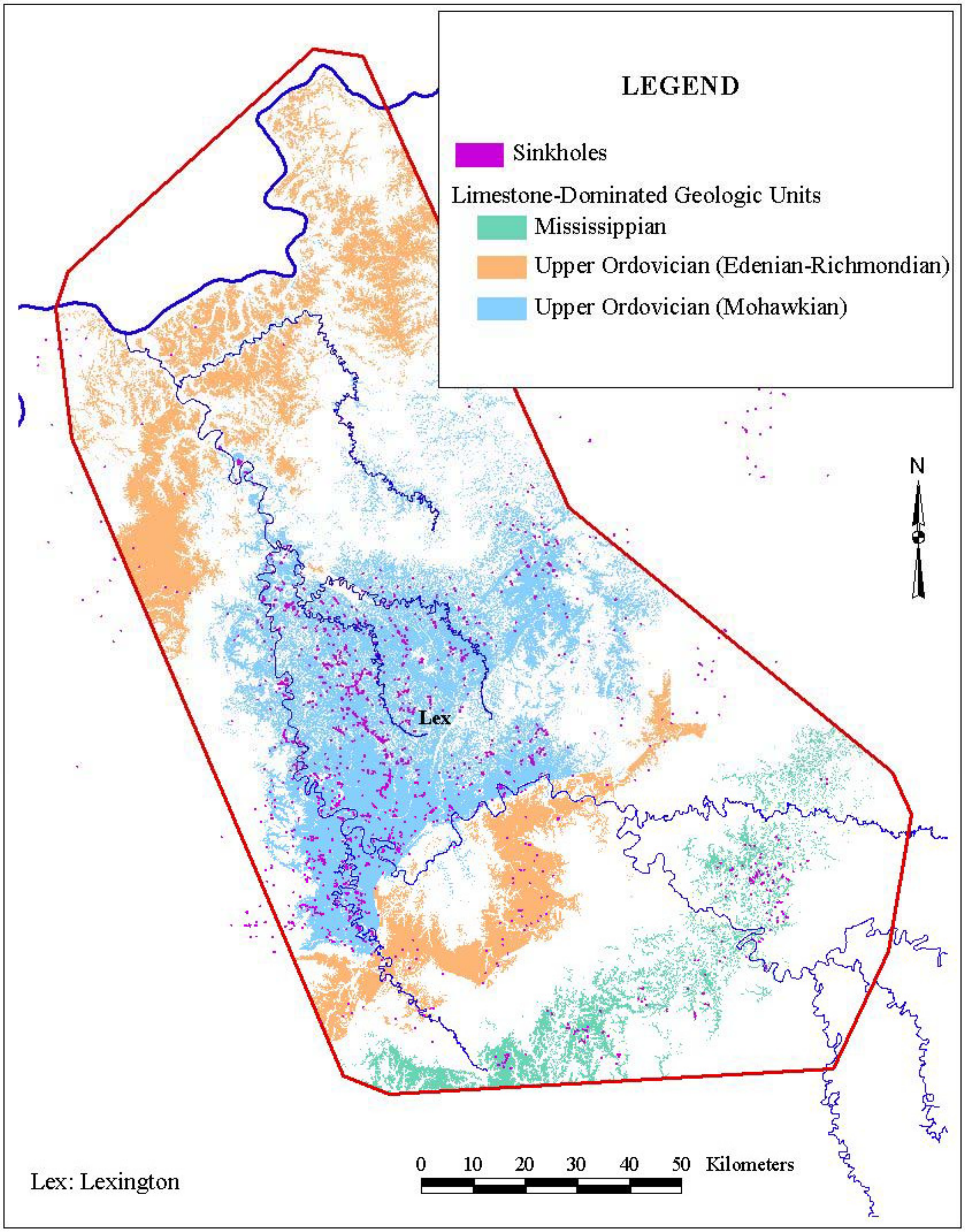


Figure 14. Distribution of sinkholes in central Kentucky, showing distribution of limestone-dominated bedrock units in the study area.

> The term peneplain should be restricted to those gently undulating landscapes which develop under a base level control toward the end of a humid fluvial cycle in part through lateral planation by streams but more through mass-wasting and sheet wash on interstream areas than by stream erosion. Used in this sense, it has a definite meaning and implication in regard to the geomorphic history of the region that retains such a condition (p. 187)

Another model of landscape evolution is that developed by Hack (1960, 1965, 1966). Hack identified lithologic control on erosional processes as the primary source of differing elevations and apparent planation surfaces in the eastern and southeastern United States. Hack proposed that landscapes downwaste at similar rates through time, with multiple fluvial and colluvial processes in delicate balance, unless the equilibrium of the system is somehow disturbed. The system rapidly adjusts, or attempts to adjust, to the disturbance to regain an equilibrium state. Distinctive landforms develop on particular rocks or structures, and persist through time as the entire landscape is downwasted somewhat steadily through time. The implication of the Hack model would be that planation surfaces are related to the erosional resistance of underlying strata, and do not represent chronological features.

Most previous considerations of landscape evolution in central Kentucky have relied upon a Davisian cyclic-erosion model to explain development of landforms and landscape features. Fenneman (1939), Jillson (1930, 1945a, 1963), Thornbury (1965), Straw (1968), and Warwick (1985) have all invoked peneplain theory to describe the physiography and landscape evolution of central Kentucky. An apparent accordance of ridge-top elevations gently sloping to the northwest has been identified as the Lexington peneplain. The Lexington peneplain is held as a classic example of a peneplain surface produced by extended erosion, and is inferred to have been “reactivated” in the mid-Tertiary (Jillson 1930, 1943a, 1950, 1963) by regional uplift. Uplift is inferred on the basis of intrenchment of major streams and tributaries into the regional peneplain. The source or mechanism of this uplift has not been conclusively identified.

Previous Work

The Kentucky River is a prominent geologic and geographic feature in central Kentucky. The scenery of the river, and the Palisades in particular, has attracted numerous artistic, photographic, and historical/cultural studies, as well as aroused scientific speculation as to the origin of the steep gorge cut through an otherwise gently rolling landscape. The Kentucky River is a major tributary of the modern Ohio River, and has been identified as the largest tributary of the Plio-Pleistocene Teays River system.

Kentucky River Deposits and Geomorphology

Although he cites others who had noted the presence of quartz pebbles, water-worn geodes, and Mississippian corals in the uplands on the Bluegrass, Miller (1895) was the first to suggest that these pebbles and cobbles were deposited by a through-flowing trunk stream, rather than let down as a residual deposit. He noted the restricted occurrence within relatively close proximity to the modern river, and typically below the elevation of the highest ridges but well above the modern river, as evidence that they were more likely related to older versions of the river system than to regional landscape lowering. A residual origin, he argued, would have led to their distribution across the entire area, rather than only close to the river valley. Campbell (1898) was the first to formally name high-level fluvial deposits between Richmond and Irvine as the Irvine Formation, but only provided limited description of details of these deposits. Foerste (1906) provided detailed descriptions of the Irvine Formation in the same area of east-central Kentucky. Miller (1919), Leverett (1929), McFarlan (1943), and Straw (1968) all correlated the Irvine Formation with other upland gravel deposits along the Ohio River below the mouth of the Kentucky and with the Lafayette gravel of the Mississippi Embayment. Leverett (1929) identified a sharp bend in the Salt River near Lawrenceburg, Kentucky, southwest of the Kentucky River basin, as a capture of an early tributary of the Kentucky by the Salt River system.

Willard Rouse Jillson conducted the most intensive field study focused on the deposits of the ancient Kentucky River. In a series of pamphlets (Jillson 1943b, 1944a, 1944b, 1945b, 1946a, 1946b, 1946c, 1946d, 1947, 1948a, 1948b), Jillson characterized the nature of the deposits, including silt and sand and distinctive gravel composed of well-rounded quartz pebbles, water-worn geodes, and subrounded chert pebbles. He also noted slabs of sandstone in the deposits at numerous locations. He termed these "typical Irvine gravels" and associated them with the formally named deposits identified farther upstream by Campbell (1898). On the basis of only rudimentary topographic mapping (the best available, but still crude by modern standards) and intensive field work, Jillson's pamphlets also delineated and named abandoned paleochannels of the Kentucky River from Camp Nelson to Carrollton (Figure 15). Maps in the pamphlets note the location of more than 370 separate observations of fluvial sand and gravel and related valley features. Mappers in the joint KGS-USGS geologic-mapping program expanded upon the work of Jillson, Campbell, and Foerste, and delineated specific deposits of high-level fluvial sediment across central Kentucky.

Jillson (1945a, 1963) also hypothesized an even older course of the Kentucky River traversing the Inner Bluegrass region from Winchester, through Georgetown, to Monterey, on the basis of quartz and granite pebbles found in selected sites across the region. Jillson speculated that the granite cobbles indicated that the Kentucky River originally had its headwaters in the Blue Ridge Mountains. He suggested that structural activity related to the Kentucky River fault system caused abandonment of this proposed ancient course of the river. Although a Blue Ridge paleo-origin of the Kentucky River is beyond the geographic scope of the current study area, the hypothesis can be constrained by the observations of this study.

A series of Masters theses completed at Eastern Kentucky University, summarized by Vanarsdale (1986) and Vanarsdale and Sergeant (1992), used mapped fluvial deposits in the Kentucky River basin to investigate potential neotectonics in central Kentucky. The theses selected a series of fluvial deposits that overlie mapped faults identified during the KGS-USGS mapping. Trenching and coring were used to look for potential offset and disturbance in the deposits that might indicate structures active since deposition of the fluvial material. The only disturbances identified were in deposits well above the course of the modern river. Although the disturbances might have been related to karst dissolution in underlying limestone bedrock, Vanarsdale (1986) cited the recurrence of deformation in multiple trenches and other site-specific evidence to argue for a tectonic origin, supporting Jillson's hypothesis of structural activity along these structures.

Tight (1903) named the Teays River system for an abandoned high-level valley system he recognized in West Virginia and southeastern Ohio and inferred it to be "pre-glacial,"

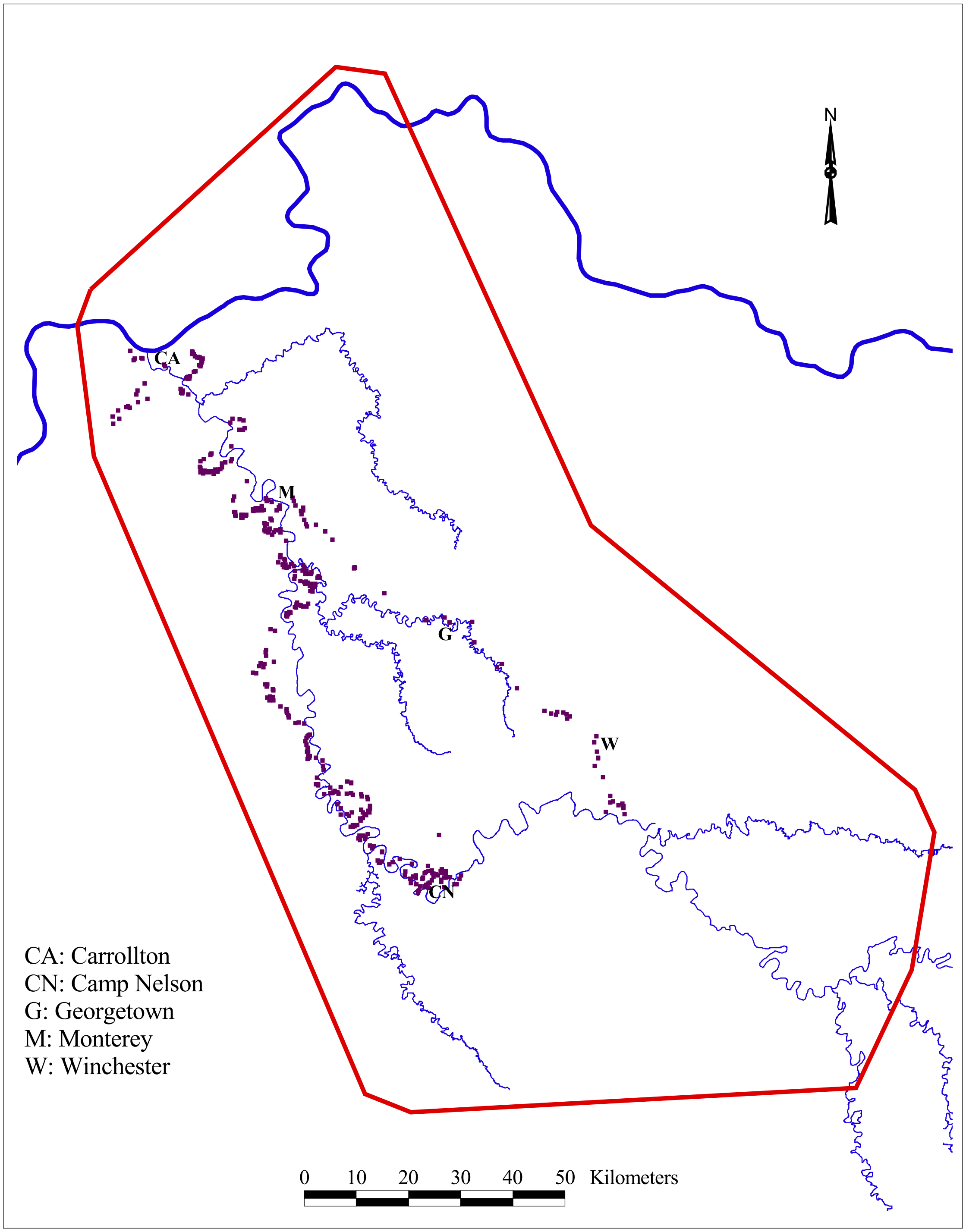


Figure 15. Distribution of fluvial deposits and associated features noted in the publications of Willard Rouse Jillson (references in text).

hypothesizing that subsequent Pleistocene glaciations obstructed the Teays, facilitating formation of something similar to the modern Ohio. Subsequent work has traced the course of the Teays River system from West Virginia, across Ohio, central Indiana, and through Illinois (Figure 16) (summarized in Melhorn and Kempton, 1991).

On the basis of detailed topographic mapping, extensive mapping of bedrock terraces and straths, delineation of bedrock valleys buried under Quaternary glacial deposits, and mapping of glacial, fluvial, and lacustrine deposits in upland paleovalleys, the Ohio River was likely formed by destruction of the Teays River system by pre-Illinoisan glaciation (Teller, 1973; Ray, 1974; Melhorn and Kempton, 1991, and references therein). The earliest Ohio River probably had its headwaters along the Madison divide, developed on resistant Silurian lithologies near Madison, Indiana (Figure 16). Another divide near Manchester, Ohio, separated Licking River-related drainage from New/Kanawa River drainage (Figure 16). Flooding of the Teays Valley as a result of prolonged blockage by pre-Illinoisan ice probably led to overflow across cols at these divides and development of new drainage pathways through the subsequently eroded cols. This newly expanded Ohio River flowed in a loop north of the modern site of Cincinnati, but otherwise followed the modern course of the river. A subsequent glaciation (Illinoisan?) caused the establishment of a new channel south of Cincinnati, notable for barbed tributaries in the vicinity of Anderson Ferry. The Wisconsin glaciation was marked by aggradation of the Ohio River valley by a thick valley-train deposit of glacio-fluvial outwash.

The paleodrainage of the Kentucky River has been a matter of varied opinions and study for several decades. Some early workers disagreed over whether the ancient Kentucky River turned southwest at Carrollton and followed an early version of the Ohio River (Leverett, 1902; Fenneman 1914, 1916, 1938; Stout and others, 1943), or turned northeast and flowed into the ancient Teays River system (Fowke, 1898, 1900, 1925, 1933; Tight, 1903; Malott, 1922; Wayne, 1952). The early Ohio River hypothesis proposed a paleodrainage divide somewhere upstream along the old Ohio from the mouth of the Kentucky, at Cincinnati or beyond. The Teays hypothesis suggested a paleodivide at Madison, Indiana.

Swadley (1971) delineated and characterized a paleocourse of the Old Kentucky River from Carrollton, Kentucky, to near Lawrenceburg, Indiana (Figure 16), on the basis of data from his geologic mapping in the area. He based a northward to southward reversal of flow on the

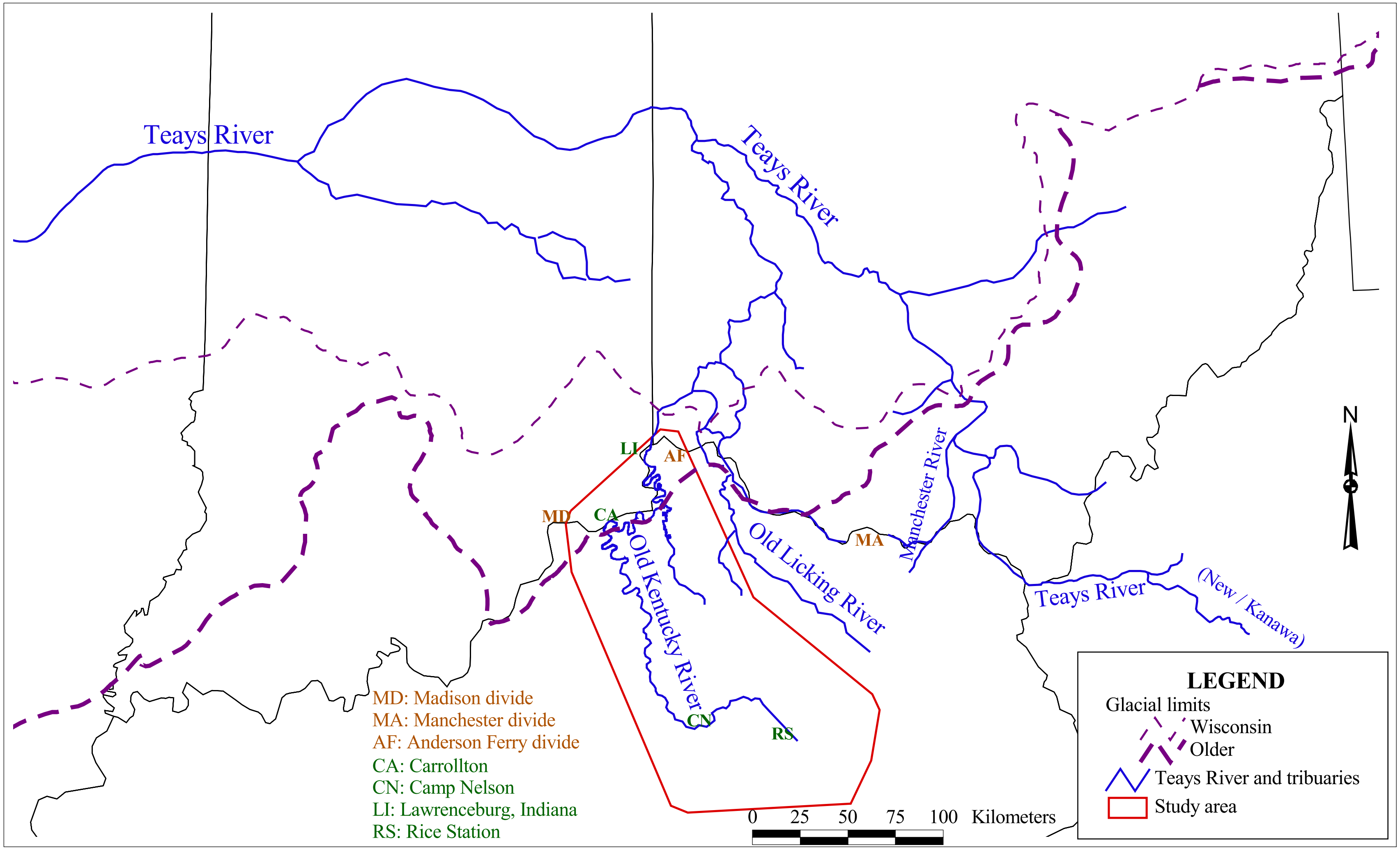


Figure 16. Relationship of the Old Kentucky River to the ancient Teays River system.

juxtaposition of lacustrine and glacial deposits over the high-level fluvial material. No paleochannels or high-level fluvial deposits were identified along the Ohio River valley between Carrollton and Madison during the mapping program. Also on the basis of mapping from the KGS-USGS program, Luft (1980, 1986) examined high-level deposits in the Licking River valley and discussed their implication for geologic evolution of the Licking River and its major tributaries. On the basis of this mapping and other mapping work around Cincinnati ("AF" on Figure 16) (references in Luft, 1980), both the ancient Licking and the ancient Kentucky were traced to the Teays system (Figure 16). The major drainage divide between the Teays River and early Ohio River basins was at Madison, Indiana.

Warwick (1985) tested a hypothesis of river response to regional base level lowering in the Kentucky River valley. In a study that extended from Carrollton to Camp Nelson (Figure 16), Warwick used valley and knickpoint geometries to develop a model of how the Kentucky River responded to base-level lowering upon abandonment of the ancient Teays River and establishment of the early Ohio River. Although Warwick noted numerous instances of lithologic/stratigraphic control on knickpoints and valley morphology, his conclusions focus on geomorphic/temporal control on knickpoint development in streams tributary to the Kentucky River. Warwick (1985) inferred migration of a master knickpoint up the Kentucky River valley.

The only quantitative geochronologic evidence from Pleistocene and older Kentucky River deposits comes from recent applications of cosmogenic-radionuclide geochronology. A quantitative technique for estimating the exposure or burial ages of sediments relies upon the production of selected radionuclides from the bombardment of the Earth by cosmic radiation. Granger and Smith (2000) produced a cosmogenic beryllium-production model age of approximately 1.5 to 1.74 Ma for Irvine Formation deposits near Rice Station in Estill County, Kentucky (Figure 16). Dating of deposits near Carrollton (Figure 16) produced an age estimate of 1.3 to 1.45 Ma (Granger, personal communication, 2001).

Regional Glacial Geology

Classic studies identified four Pleistocene glaciations in the midwestern United States. These glaciations were named for states in which "type" deposits were recognized, from youngest to oldest: Wisconsin, Illinoisan, Kansan, and Nebraskan (eg. Ray, 1974). Subsequent geochronological work in the midwestern states, however, has identified as many as 12 different till deposits and noted miscorrelations among many glacial deposits in the region. As such, only two "named" glaciations are currently recognized in the region: the Wisconsin and the Illinoisan. Glacial deposits older than the Illinoisan are termed "pre-Illinoisan" and identified sequentially by letter (Richmond and Fullerton, 1986). As many as seven different distinct tills have been identified in the buried Teays Bedrock Valley in Indiana (Bleuer, 1991) and four pre-Illinoisan tills have been identified south of the Wisconsin glacial limit in Decatur County, Indiana (Teller, 1972). Wisconsin ice did not reach as far south as Kentucky, and only glacial outwash deposits from the Wisconsin are found in the Ohio River valley. Illinoisan deposits include drift and/or till deposits along the Ohio River from just above Cincinnati to possibly as far south as Louisville (Ray, 1974). Two pre-Illinoisan glacial deposits have been inferred for northeastern Kentucky (Leighton and Ray, 1965; Ray, 1966, 1974), although published USGS geologic quadrangle maps do not differentiate the pre-Illinoisan deposits. Johnson (1986) and Fullerton (1986) summarized the stratigraphic relationships of glacial, fluvial, lacustrine, and eolian deposits and related pedologic features in Illinois, Indiana, Ohio, and northern Kentucky.

The pre-Illinoisan glaciations disrupted the Teays River system, and possibly caused the subsequent organization of the Ohio River system (Tight, 1903; Gray, 1991; Teller and Goldthwait, 1991). A glacial advance into the Teays valley would have caused impoundment of the river valley, until a low point on the downstream divide ("col") was overtopped, allowing the valley to drain. If the blockage persisted, and the col acted as a spillway for extended periods of time, then significant bedrock erosion through the col could lead to rerouting of the drainage system through the newly developed channel. Ettensohn (1974), Teller and Last (1981), and Luft (1980, 1986) identified extensive, high-level lacustrine deposits in the Licking River valley and attributed them to impoundment of the Licking River valley by a pre-Illinoisan glaciation. As such, spillways should have existed to drain the Licking River basin (and likely all upstream drainage from the Teays/early Ohio River system) into the Kentucky River basin to the southwest. No spillways of this sort have been documented, except for ice-proximal courses identified by Teller (1973). Swadley (1971) and Ettensohn (1974) have identified lacustrine deposits in the Old Kentucky River paleocourse between Carrollton and Cincinnati and inferred glacial impoundment as the cause of abandonment of this segment of the river valley. Teller and Goldthwait (1991) have suggested that the lack of thick lacustrine deposits upstream in the Kentucky River may indicate that the Kentucky was diverted to the southwest prior to the arrival of the earliest glacial advance in the region. If the main stem of the Kentucky (upstream from Carrollton) were impounded, lacustrine deposits should exist in the valley or paleovalleys, or a low-relief col very close to the elevation of the ancient Kentucky River would have allowed drainage to occur without significant slackwater impoundment of the river valley.

Methods

Much of the previous research in the Kentucky River basin preceded the development of desktop computing. Recent advances in Geographic Information System (GIS) software and technology, and the resultant release and development of GIS-compatible data sets have provided a new tool available for spatial examination of new and previously published data for the study area. The first step for this project was assembling previously collected information gathered by other workers. This included information on the bedrock geology, fluvial deposits, soils, elevation, streams, and glacial features, all in digital format (Appendix 2). These data sets are very large (see Appendix 2), and required significant computing power to allow simultaneous comparison of the different data sets for the basin. Until recently, this would not have been possible without specialized computer facilities. The Kentucky Geological Survey graciously provided computer and software support for this project.

KGS personnel have digitized all of the 1:24,000 geologic maps for the study area (Appendix 1). These maps include not only bedrock geology, but also unconsolidated materials such as alluvium, lacustrine deposits, and glacial deposits. Additional locations of fluvial deposits were collected from Jillson’s series of pamphlets on abandoned paleochannels of the Old Kentucky River (Jillson 1943b, 1944a, 1944b, 1945b, 1946a, 1946b, 1946c, 1946d, 1947, 1948a, 1948b, 1963), from Joseph Ray (Kentucky Division of Water, personal communication, 2001), and from field work during the course of this study. Figure 17 shows the locations of 174 sites visited during the course of this study. Selected county soil surveys, published by the U.S. Department of Agriculture, have been digitized and included in the USDA Soil Survey Geographic (SSURGO) database. Soil mapping was completed at higher resolution than geologic mapping and, in some places, identified fluvial parent materials in soils where geologic maps overlooked them (Figure 18). Only the soil data for Scott County in central Kentucky were not available in digital format for the study area at the time of analysis. The USGS has developed high-resolution (10-meter cell size) digital elevation data for Kentucky and scanned topographic

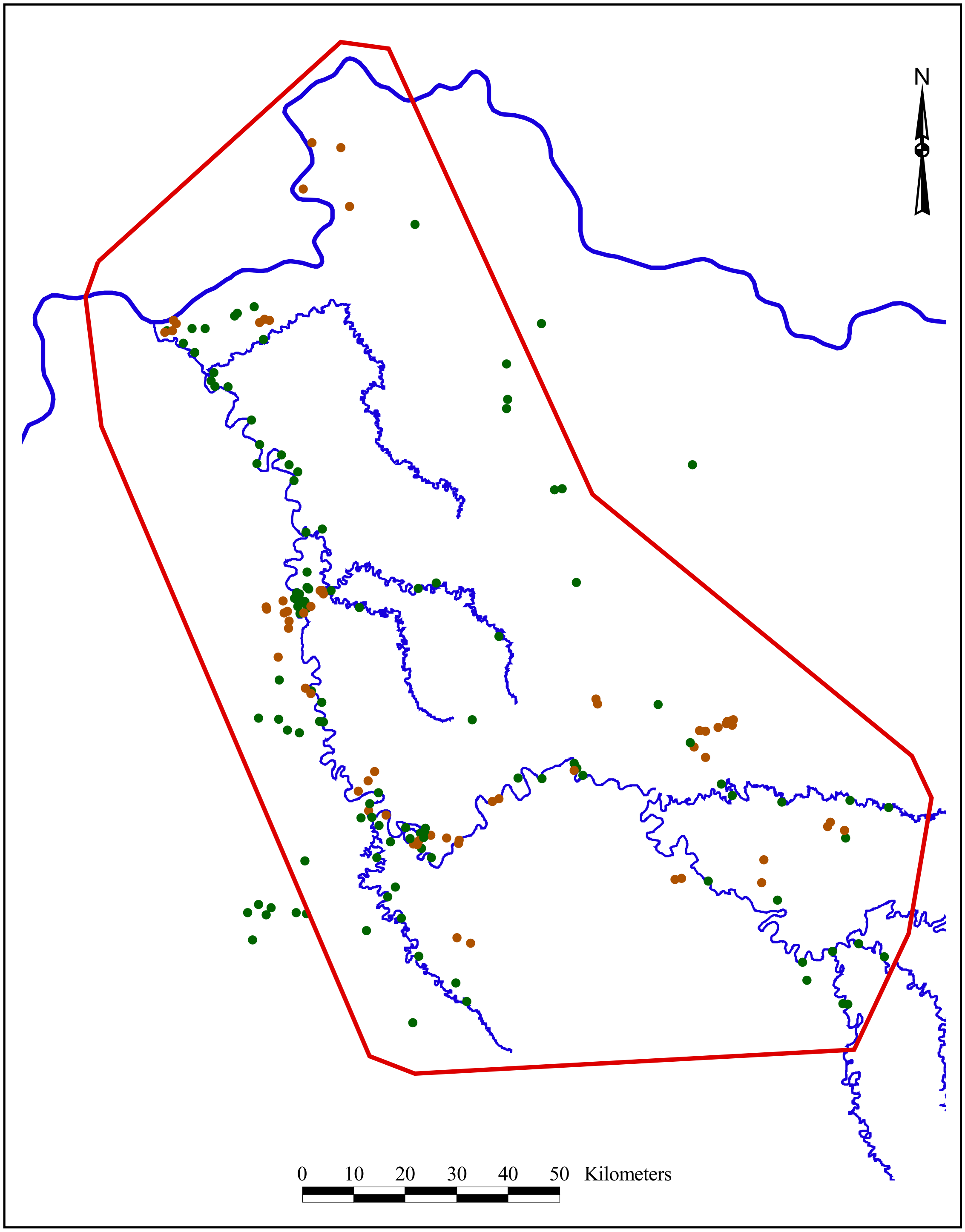


Figure 17. Location of sites visited during the course of the current study. Green dots represent sites where landscape features such as stream valleys, paleovalleys, and terraces were noted. Orange dots represent sites where sedimentary deposits were observed in addition to the landscape features.

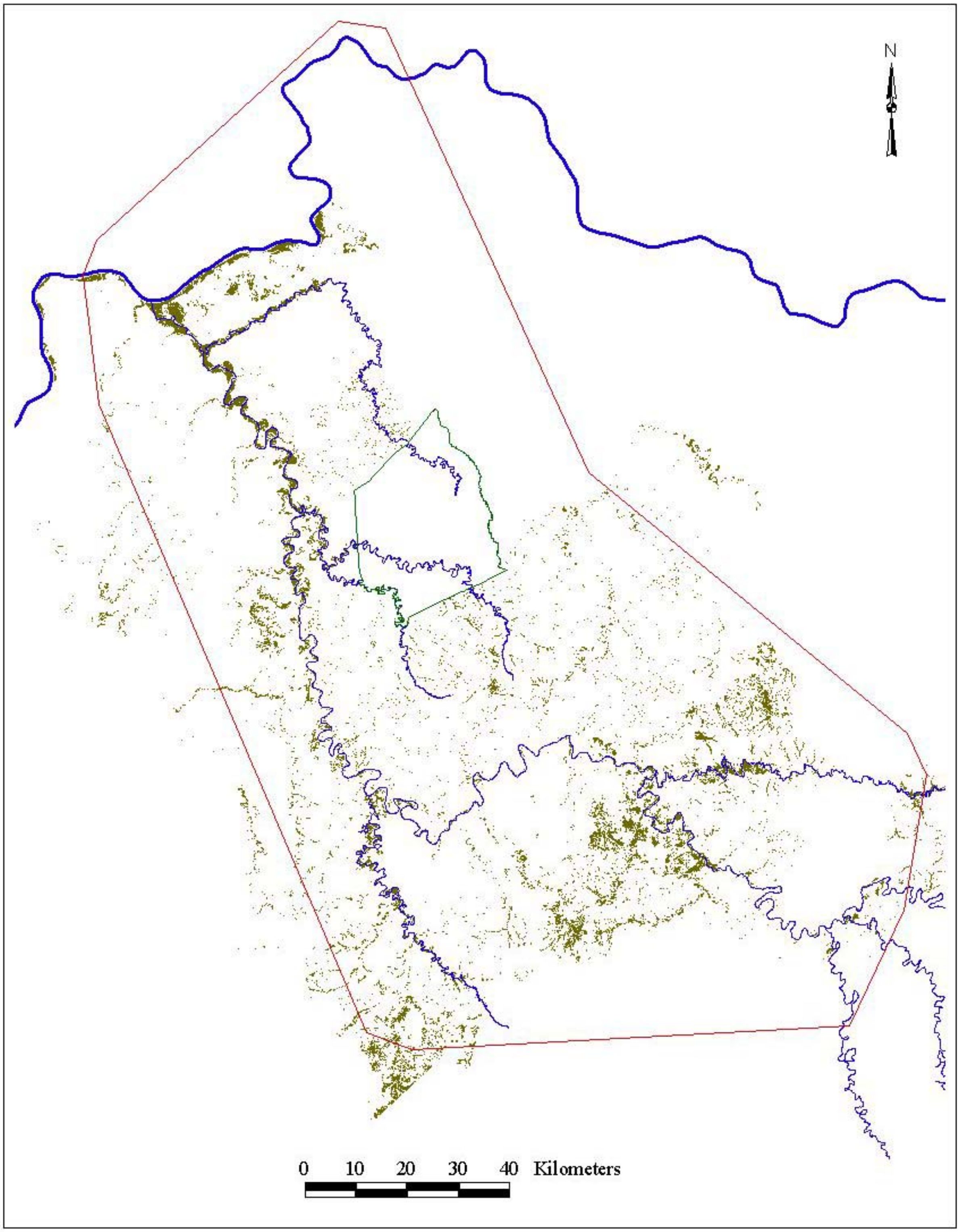


Figure 18. Distribution of soils derived from terrace sediments and associated deposits along the Kentucky River and major tributaries. Active floodplains are excluded. Digital data from the USDA SSURGO database. Scott County shown by green outline.

maps for the study area, and both are available through the Kentucky Office of Geographic Information Systems (OGIS). Another USGS data set, the National Hydrologic Dataset (NHD) provides attributed centerlines of mapped streams in Kentucky.

The elevation, geology, and soils data were examined to identify the locations of fluvial deposits and potential paleovalleys and high-level abandoned meanders. In only a few places were the combined soil and geologic data sets insufficient to confirm potential fluvial deposits or paleovalleys, and field work was conducted to search for diagnostic sediment such as geodes or quartz pebbles (Figure 17). Numerous bedrock benches were identified from the digital elevation data and scanned topographic maps. The bedrock benches are too numerous to efficiently field check even a majority of the features, and they are delineated and identified separately from the confirmed fluvial deposits on cross sections in this study.

Compilation of regional geochronologic measurements can serve to constrain the temporal framework of the relative chronology. Geophysical and hydrologic modeling, although imprecise, can also serve to constrain the possible processes and responses expected from lithospheric flexure and lake impoundment resulting from glacial loading. Published erosion and denudation rates can constrain the process rates in this study, in the absence of quantitative measurements and data. These calculations and constraints are discussed in more detail in Chapter 3.

Ideally, comparison of elevations or discreet longitudinal profiles or alignments of the ancient river deposits will allow for development of a detailed relative chronology of downcutting and stream captures in the high-level deposits to be a straightforward, albeit tedious, process. Available geologic, geomorphic, and climate data can be used to infer controls on each capture or downcutting event. Possible controls on fluvial events in the Kentucky River basin are discussed in Chapter 4. Each system adjustment should also have resulting impacts on the rest of the system, as well, so ideally the sequence of events and their inter-relationships can be identified.

Traditionally, distances along the Kentucky River have been measured in U.S. statute miles (5,280 ft, 1609.3 m), surveyed by the U.S. Army Corps of Engineers (USACE). Distances given in this study are derived from measurements along the lines provided in the U.S. Geological Survey National Hydrologic Dataset (NHD) data for the Kentucky River. The digitized centerlines in the NHD dataset do not conform to the USACE-surveyed channel of the river, and thus do not duplicate the USACE values. Also, the NHD dataset is measured from the centerline of the Ohio River, thus adding approximately 0.4 km to the overall measured distance. Taking the 0.4-km difference into account, the NHD-derived distances are typically 2-3% greater than the USACE-surveyed values. The USACE survey followed the navigable channel of the Kentucky, whereas the NHD data follows the river centerline. Likewise, distances measured along the estimated centerline of the river valley do not duplicate values derived from the NHD river-centerline values, because of meandering of the river within the bedrock valley. Table 1 compares distance values from the different sources. This study will primarily use valley distances calculated from the estimated centerline of the valley ("valley-km"), with values for selected landmarks summarized in Appendix 3.

Table 1. Comparison of different measures of distance along the Kentucky River

Feature	**USACE** *miles*	**USACE** *km*	**NHD** *channel km*	**This Study** *valley km*
Mouth of Eagle Creek	11.0	17.7	18.6	15.3
Lock and Dam No. 4 (Frankfort)	65.0	104.6	108.2	80.2
Mouth of Dix River	118.2	190.2	195.3	152.7
Mouth of Red River	190.8	307.0	314.8	261.5
Confluence of North and South Forks	254.8	410.0	418.0	350.3

CHAPTER TWO: OBSERVATIONS AND INTERPRETATIONS

This goal of this project is to interpret the geologic controls on the evolution of the Kentucky River. This requires a description and interpretation of preserved fluvial deposits and the geomorphology of the river valley. Map perspectives of valley geomorphology include drainage area, drainage density, meanders, and stream trends. Longitudinal perspectives include stream profiles and knickpoints of the master stream and tributaries, as well as flood-stage profiles of the main stream. Quantitative considerations of valley morphology include valley width and sinuosity. Qualitative descriptions of valley morphology include the identification of meanders and paleovalleys. Previously published observations and interpretations in adjacent stream valleys provide context for the observations of this study. Anthropogenic factors such as engineering and land-use changes obscure the natural geologic processes, and areas susceptible to these human-induced controls are excluded from this study.

Engineering and Land-use Impacts

Land-use changes and engineering modifications of the Kentucky River valley in the last 200 years have had significant impacts on erosion and sedimentation in the basin. Archaeological studies at local sites, such as Fort Boonesboro (O'Malley, 1989) have identified significant landscape alterations caused by erosion and sedimentation attributed to engineering improvements and land-use change. Similar changes have been documented in the Salt River basin during planning for construction of Taylorsville Lake in the 1980s (Collins and Norville, 1980). The land-use and engineering impacts have altered the surface of the alluvial valley fill, but not significantly changed buried deposits or terraces along the valley shoulders, both of which pre-date the land-use impact. Analysis of these surficial modifications would require stratigraphic and geochronological studies beyond the scope of the current effort. This study, therefore, will focus on evolution of the Kentucky River basin prior to human land-use impacts.

Native peoples inhabited or hunted the Kentucky River basin from approximately 12,000 years ago until displaced by European settlers in the late 18th century. The earliest inferred land-use impacts in the basin are the localized burning of selected central Kentucky areas by Late Prehistoric native peoples to encourage growth of grassland and savannas, in order to attract buffalo for sustenance (Kingsolver, 1992). These burning events could have led to changes in sediment supply and run-off to the adjacent streams. The first non-native pioneers settled in the basin in 1775, and began to intensively clear forests for agriculture, construction materials, and defense against native attacks (Harrison and Klotter, 1997).

Timber, iron ore, and coal have been major economic natural resources within the basin, especially within the basins of the Three Forks. Surface disturbance and land-clearing associated with these industries led to major soil-erosion problems and significant siltation in the basin (Verhoeff, 1911; Johnson and Parrish, 1999). Timber has always been a significant resource for eastern Kentucky throughout the 19th and 20th centuries. Timber-cutting operations for much of this time caused significant soil erosion, and sediment loading of the Kentucky River. Iron was produced at charcoal-fired furnaces in the basin throughout much of the 19th century. Forests for miles around each furnace were cleared to produce the needed charcoal to fire the furnaces (Fig, undated). Coal was produced from mines at Beattyville as early as 1790, and shipped to markets down stream on rafts and boats. In the early 20th century, the penetration of railroads into eastern Kentucky allowed major expansion of the coal industry. With the advent of surface mining in eastern Kentucky in the 1950s, more sediment was introduced to the river system.

To facilitate the transportation of these resources to downstream markets, a series of 14 locks and dams was constructed on the Kentucky River (Table 2, Figure 19). The first dam was completed near Carrollton in 1840, but because of financial and political delays, the final dam was not completed at Beattyville until 1917 (Johnson and Parrish, 1999). These dams caused modifications to the flow and sedimentation patterns of the river (Figure 20), and thus contribute to the modifications mentioned above. However, the construction records of the dams provide information on the pre-settlement valley-fill deposits and, thus, serve as a source of geotechnical data for this study.

Kentucky River Geomorphology

Drainage Area

The Kentucky River basin is contained entirely within the Commonwealth of Kentucky and drains a total area of 18,042 km^2 (Figure 21) (Bower and Jackson, 1981). The mouth of the river is at Carrollton where it joins the Ohio River at mile 545.8 (measured from Pittsburgh, Pennsylvania). The main stem of the Kentucky River has its head 350 valley-km from its mouth at Beattyville at the confluence of the North, Middle, and South Forks of the Kentucky,

Table 2. Locks and dams constructed on the Kentucky River

Feature	Valley-km	Upper pool (m)	Construction	Completed
Ohio River at Carrollton		128.6		
Lock and Dam No 1	6.0	131.1	Ky	1841
Lock and Dam No 2	40.6	135.3	Ky	1840
Lock and Dam No 3	51.5	139.3	Ky	1840
Lock and Dam No 4	80.2	143.4	Ky	1840
Lock and Dam No 5	102.6	147.9	Ky	1841
Lock and Dam No 6	121.9	152.2	USACE	1891
Lock and Dam No 7	150.9	156.9	USACE	1897
Lock and Dam No 8	185.5	162.5	USACE	1900
Lock and Dam No 9	211.9	167.8	USACE	1903
Lock and Dam No 10	241.5	173.0	USACE	1905
Lock and Dam No 11	275.3	178.5	USACE	1906
Lock and Dam No 12	300.9	183.7	USACE	1910
Lock and Dam No 13	327.9	189.2	USACE	1914
Lock and Dam No 14	341.8	194.3	USACE	1917

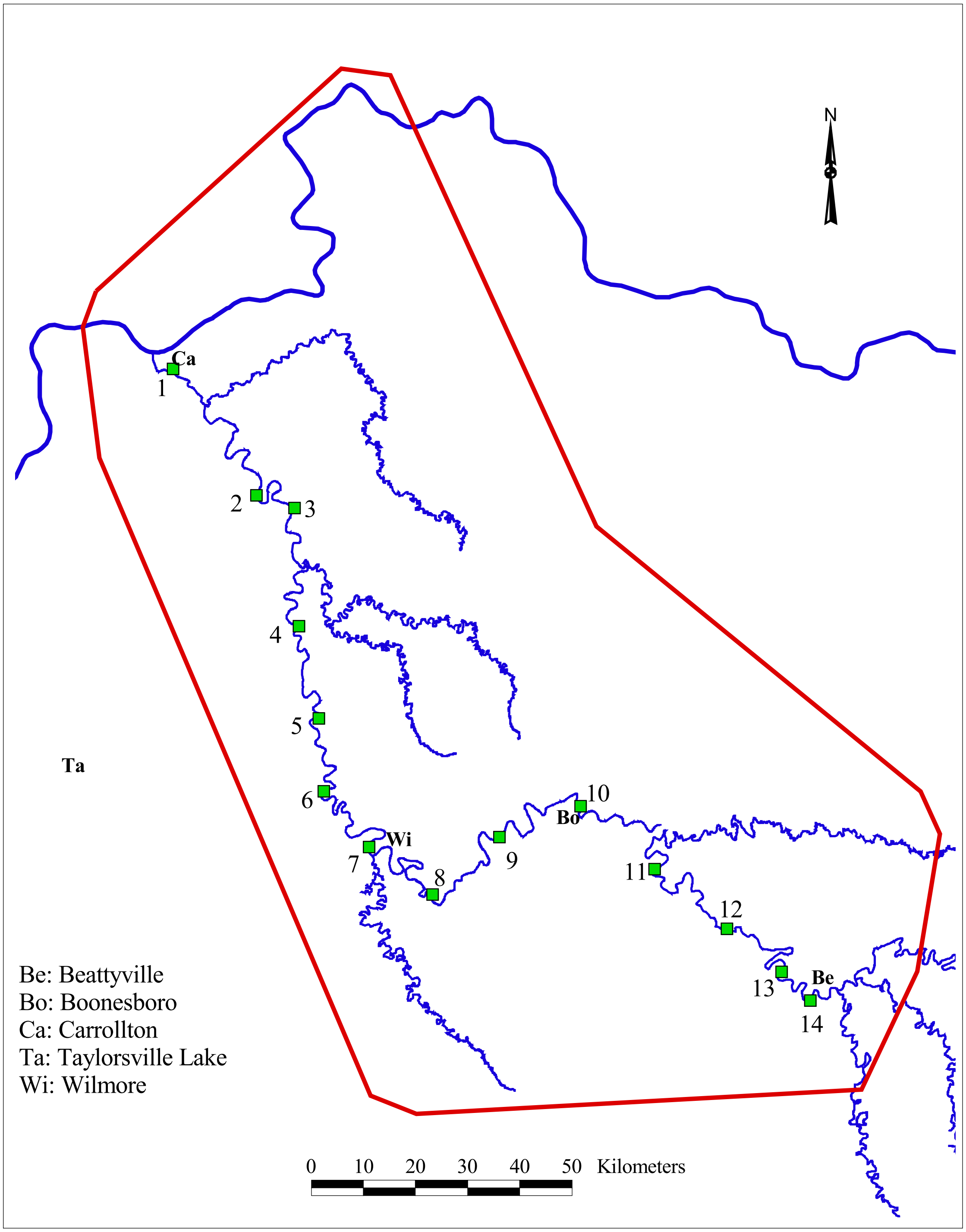
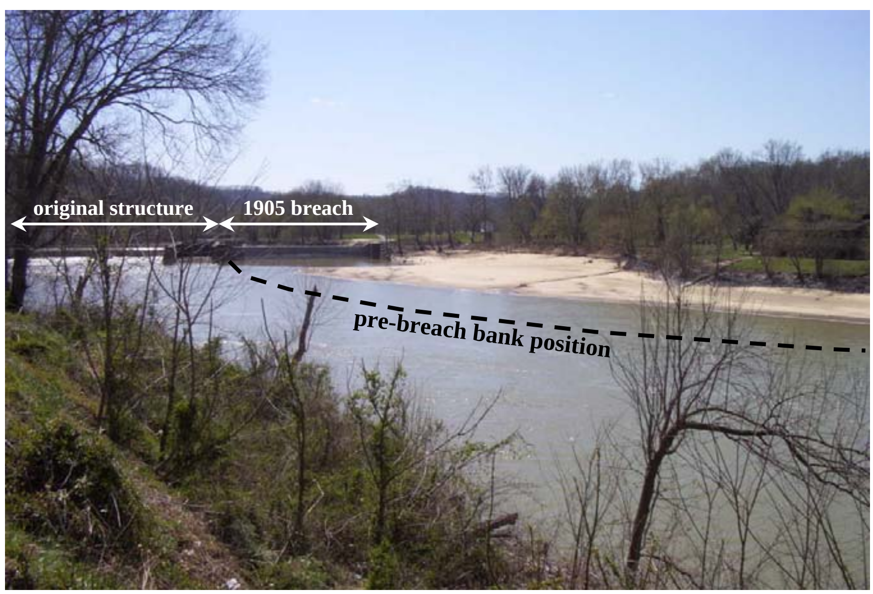


Figure 19. Locations of the 14 locks and dams built along the Kentucky River between 1840 and 1917.

A



B

Figure 20. Photographs of locks and dams along the Kentucky River. A: View of Lock and Dam No. 7 near Wilmore. B: View of Lock and Dam No. 10 near Boonesboro. In 1905 a flood breached Lock and Dam No. 10, eroding a new 73-m (240-ft) wide channel on the landward side of the lock (Johnson and Parrish, 1999). An extension was later built to block the breach.

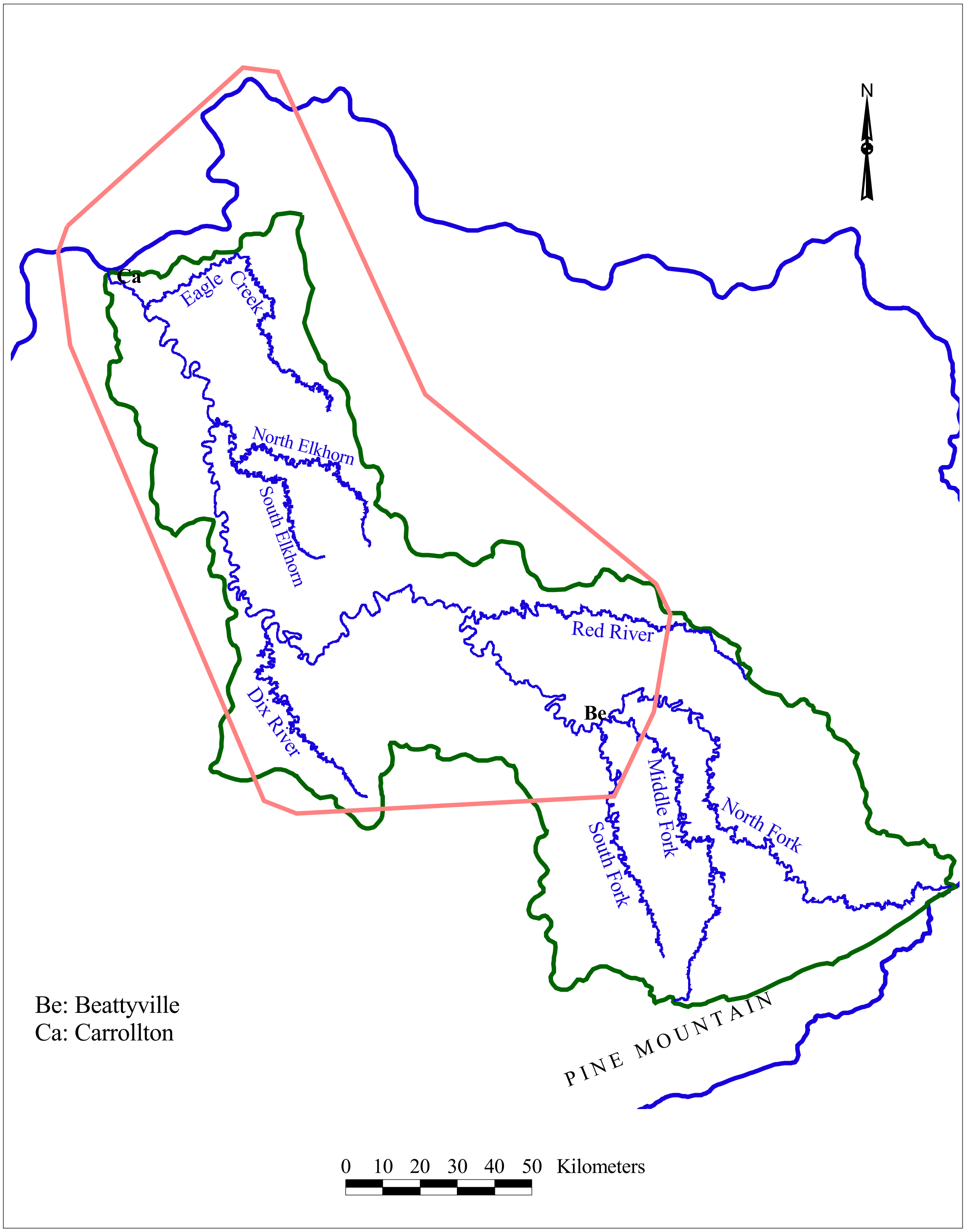


Figure 21. Drainage basin and major tributaries of the Kentucky River.

collectively referred to as the Three Forks. The Three Forks each have their headwaters on Pine Mountain, in the southeastern part of Kentucky; together they contribute 6,814 km^2 to the drainage area of the basin. Major tributaries of the Kentucky River include Eagle Creek, Elkhorn Creek (including North Elkhorn Creek and South Elkhorn Creek), Dix River, and Red River; each contributes more than 1000 km^2 to the drainage area of the basin. Table 3 summarizes the drainage areas of these and smaller tributaries.

Drainage Density

Drainage density is one parameter commonly used to characterize drainage basins. It is usually expressed as the length of streams in the basin divided by the total drainage area.

$$D_d = l_s/A$$

The USGS NHD dataset (Figure 22) contains a total length of streams (l_s) of 25,864 km within the Kentucky River basin, and the drainage area of the basin (A) is 18,042 km^2. Using the equation above, the average drainage density for the entire basin is 1.434 km/km^2.

Careful consideration of the streams illustrated in Figure 22 shows that the drainage density in the Kentucky River basin is not uniform throughout the basin. To quantify the variability of drainage density across the basin, the drainage density was calculated for the area of each 7.5-minute quadrangle within the Kentucky River basin. The length of streams within each quadrangle was calculated using ArcView software and the area of each quadrangle that is within the basin was used as the value for A. Figure 23 shows that the drainage density varies systematically across the basin. Areas underlain by the Upper Ordovician Lexington Limestone have low drainage densities. Karst development in these areas has led to underground drainage, resulting in fewer surface streams. Areas in the southeastern part of the basin are underlain by shale-dominated Pennsylvanian coal-bearing strata, and have higher relief than the rest of the basin. High-density areas in the northwest end of the basin are on shaly Upper Ordovician strata, as is the higher density area in the south-central part of the basin. The high-density area in the northeastern part of the basin is on shaly strata of the Lower Mississippian Borden Formation, and the shaly rocks of the Lower and Middle Pennsylvanian rocks.

Table 3. Drainage areas of the Kentucky River and selected tributaries

Stream	Drainage area (km^2)*	Valley-km
Eagle Creek	**1,344**	15.3
Big Twin Creek	105	24.0
Drennon Creek	253	28.7
Sixmile Creek	210	40.8
Cedar Creek	167	51.2
Elkhorn Creek	**1,295**	65.0
Benson Creek	277	81.2
Clear Creek	169	119.7
Dix River	**1,145**	152.7
Jessamine Creek	104	167.0
Hickman Creek	262	179.2
Sugar Creek	108	189.8
Paint Lick Creek	282	194.5
Silver Creek	326	200.6
Boone Creek	114	232.2
Otter Creek	169	242.8
Muddy Creek	176	253.9
Red River	**1,261**	261.5
Station Camp Creek	562	297.9
Millers Creek	193	306.7
Sturgeon Creek	287	341.5
North Fork Kentucky River	3,416	
Middle Fork Kentucky River	1,448	
South Fork Kentucky River	1,937	
Three Forks, above confluence	**6,814**	350.3
Kentucky River (total)	**18,042**	

From Bower and Jackson (1981)

**: tributaries with drainage area less than 100 km^2 not included*

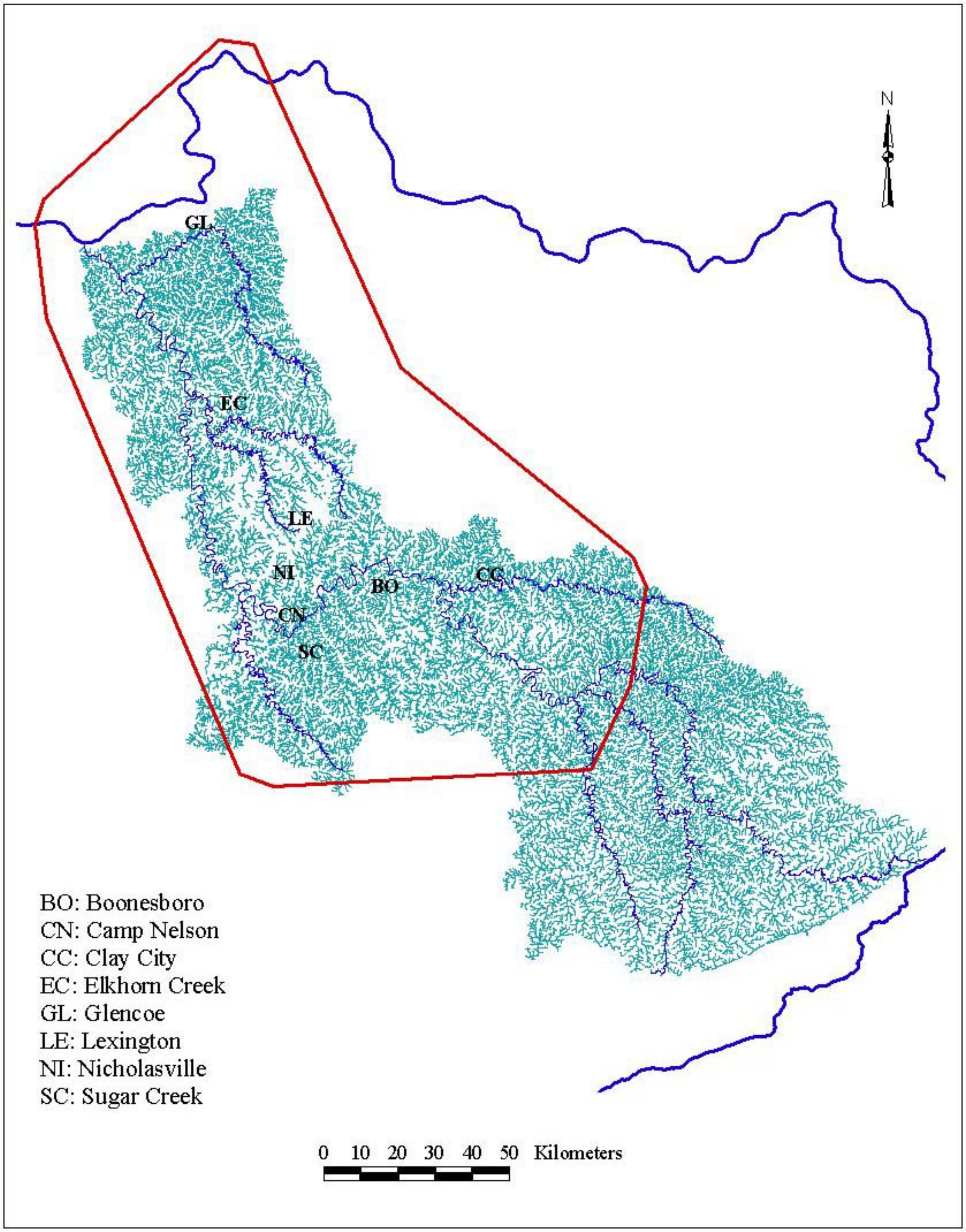


Figure 22. Streams in the National Hydrologic Dataset for the Kentucky River basin.

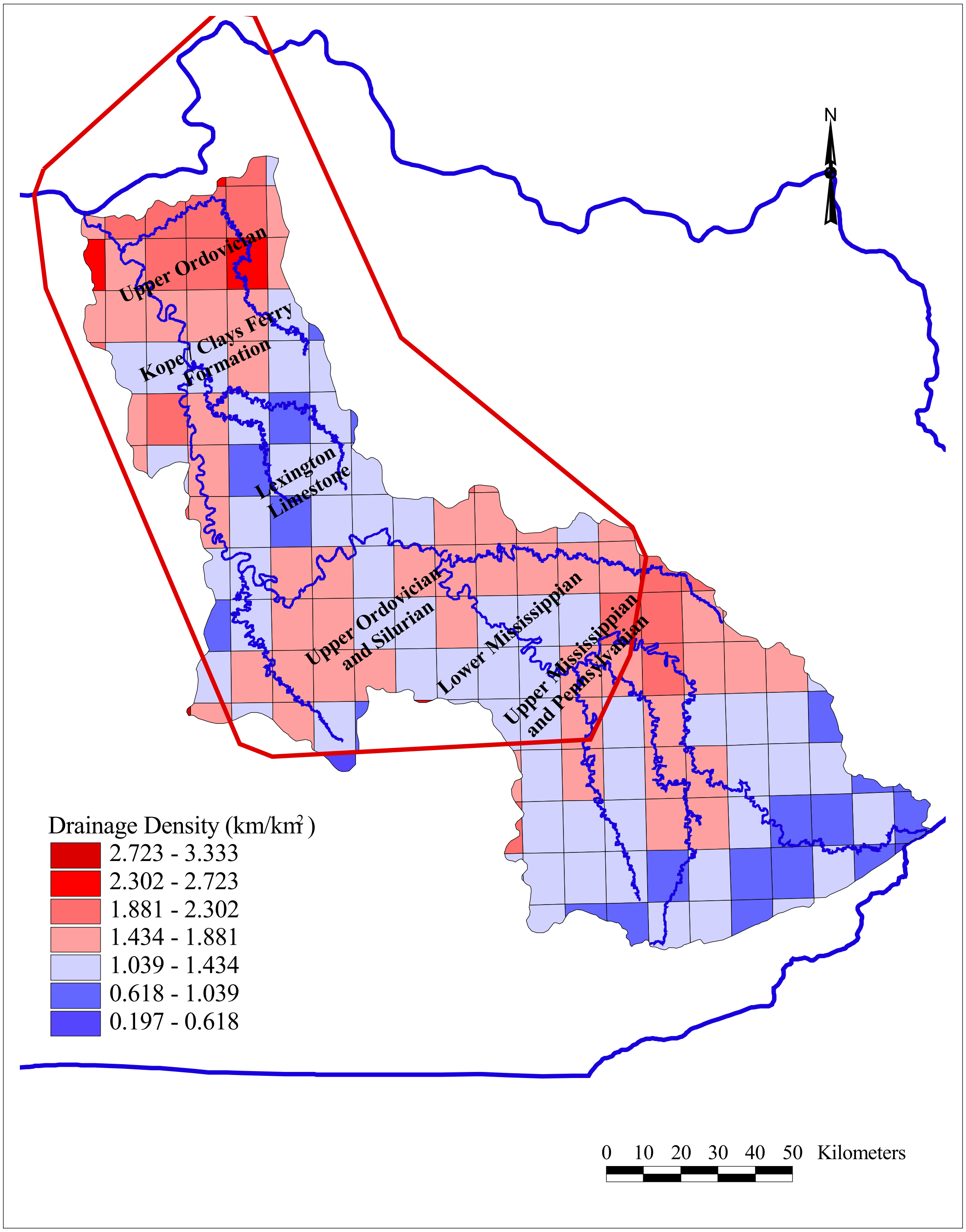


Figure 23. Drainage density for 7.5-minute quadrangles in the Kentucky River basin. The mean drainage density for the entire basin is approximately 1.434. General ages of bedrock shown by text; refer to Figure 4 for detailed bedrock geology.

Stream Trends

The Kentucky River and most of its major tributaries flow generally to the northwest throughout most of the drainage basin (Figure 22). A notable exception is the abrupt shift to the southwest of the main course of the Kentucky between Boonesboro and Camp Nelson. Another dramatic exception is the nearly 90-degree turn of Eagle Creek to the southwest near Glencoe. The Red River flows generally westward to join the Kentucky River west of Clay City. Tributary streams draining the Lexington-Nicholasville area flow south-southwestward to join the main stream. The only observed "barbed" tributaries are just upstream from Camp Nelson in the vicinity of Sugar Creek. Elkhorn Creek shows some sharp bends in the lower reaches.

Abandoned Meanders

Abandoned meanders are found only in selected stretches of the valley bottom (Figure 24). Most of the abandoned meanders are located between valley-kilometer 22 and 87 (at 22, 37, 53, 54, 60, 80, and 87), and are between 0.8 and 1.5 kilometers in diameter. A cluster of similar meanders is found along the lower reaches of Elkhorn Creek. A larger meander (> 2 km) is located at Carrollton. One very large meander (3.7 km in diameter) is located just west of Irvine. The Carrollton meander is in the Kope Formation. The Irvine meander is floored by Silurian strata, and has the New Albany black shale and Borden Formation in the valley walls. The other meanders are floored by the Lexington Limestone, and have the Kope Formation in the valley walls.

Longitudinal Profile

Figure 25 shows the longitudinal profile of the Kentucky River and major tributaries. Engineering and land-use impacts have caused significant modification of the thalweg of the stream, but even prior to settlement the stream was flowing on alluvial fill at least as far upstream as Beattyville.

Bedrock Valley

Depth-to-bedrock information along the Kentucky River valley is scarce and comes only from a few scattered water wells and from geotechnical information collected at the fourteen locks and dams built along the river. The data above Lock and Dam No. 4 at Frankfort show a relatively smooth profile that essentially parallels the thalweg and flood profiles, with an average

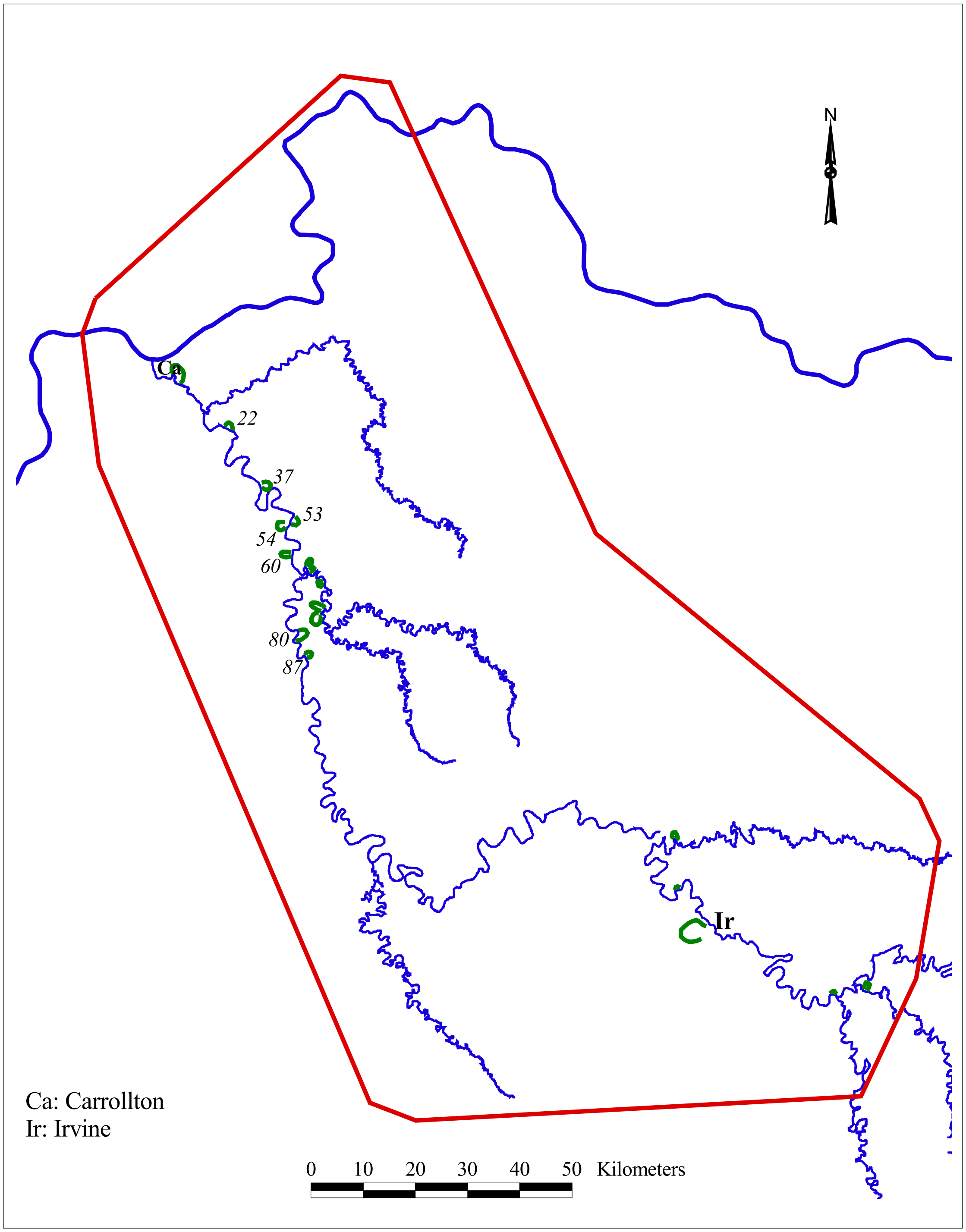


Figure 24. Distribution of abandoned meanders (shown in green) in the Kentucky River valley. Numbers denote distance (valley-km) from the mouth of the Kentucky River at Carrollton.

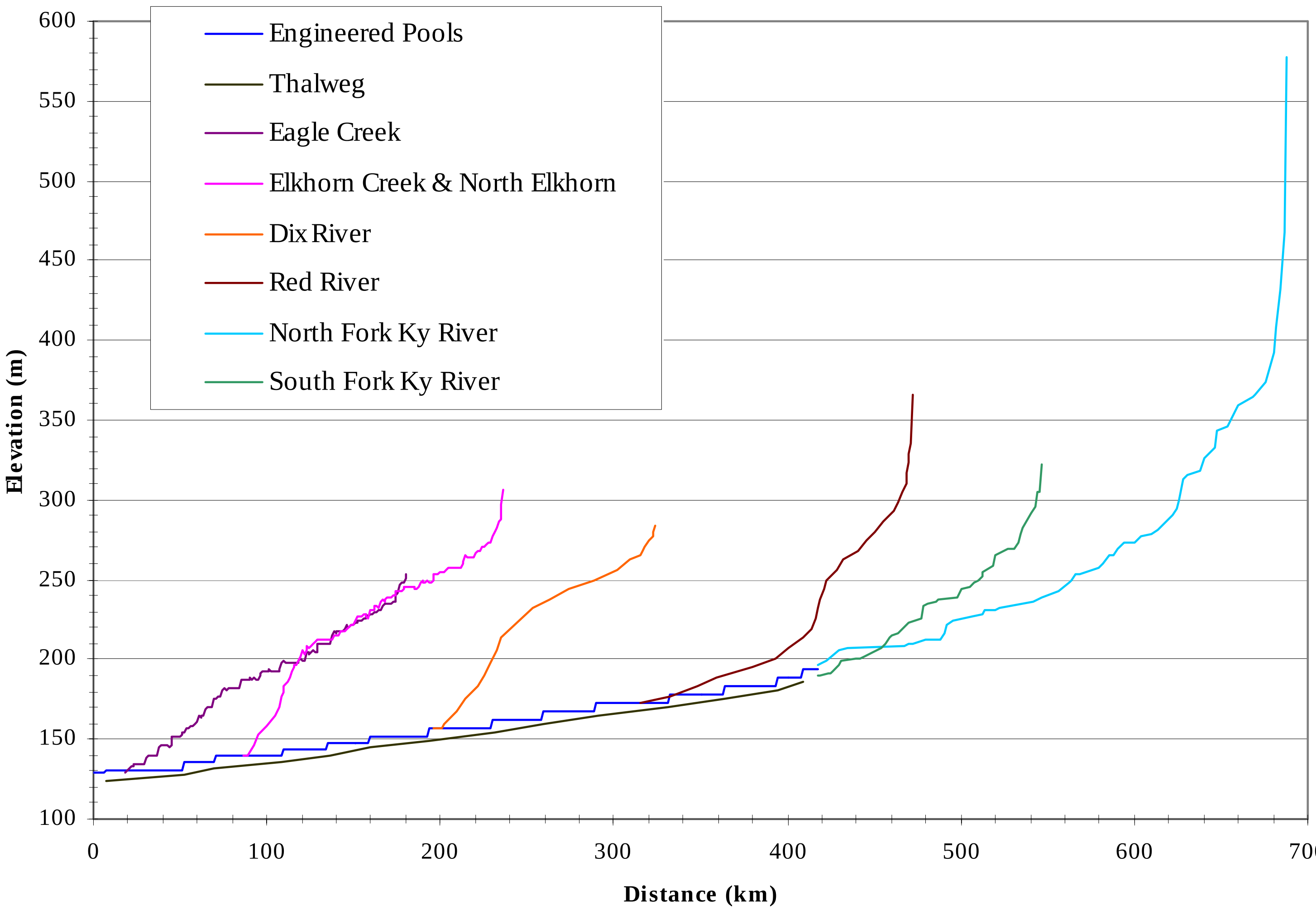


Figure 25. Longitudinal profiles of the Kentucky River and major tributaries.

gradient of 0.15 m/km (Figure 26). Downstream from Lock and Dam No. 4, the profile is apparently more irregular; knickpoints form a series of steps and benches. Although the bedrock profile is quite variable, the average gradient of the bedrock in this part of the river is 0.28 m/km, much steeper than that upstream. Detailed data at Lock and Dam No. 3 and No. 4 suggest that buried knickpoints exist just downstream from each dam; the dams were apparently built on bedrock ledges. Lock and Dam No. 3 is built on the Grier Member of the Lexington Limestone; lithologic control on the development of this knickpoint is possible. Lock and Dam No. 4 is built near the contact of the Lexington Limestone with the Tyrone Limestone. The knickpoints and irregularity downstream from Dam 4 produce a generally convex shape for this part of the profile (Figure 26), instead of the concave "equilibrium" profile that would be expected for a trunk stream. The average gradient of the entire bedrock valley, from Carrollton (valley-km 0) to Beattyville (valley-km 350), is 0.18 m/km.

Tributary Profiles

Most tributary streams in the Kentucky River basin are bedrock streams with visible knickpoints in some profiles (Figures 27, 28, 29). In general, knickpoints may form because of upstream propagation of adjustments to base-level, differential erosion of bedrock in the stream bed, localized neotectonic offsets, and sites where coarser bedload are introduced into the stream (Seidl and others, 1994). The locations of the knickpoints along tributaries of the Kentucky River do not correspond to mapped faults, or to the mouths of tributaries which might be introducing coarse bedload into the stream. The knickpoints along the major tributaries (drainage area > 1000 km^2) also evidently do not correspond to mapped stratigraphic contacts, or to mapped faults (Figure 30). Some stratigraphic units, however, are lithologically diverse. One exception to this is the prominent knickpoint in the Red River approximately 105 km above the mouth (Figure 27). This knickpoint is on the thick, massive, conglomeratic sandstones of the Lower Pennsylvanian Corbin Sandstone. The possibility of localized lithologic variation as a control on the formation of the other knickpoints is not eliminated. Eagle Creek, Elkhorn-North Elkhorn Creek, and Dix River each have a prominent knickpoint approximately 50 to 65 meters above the elevation of the confluence with the Kentucky River (Figure 27). Upstream from the knickpoints, each of these streams has a profile gradient of 55 to 60 m/km with steeper gradients downstream in each stream.

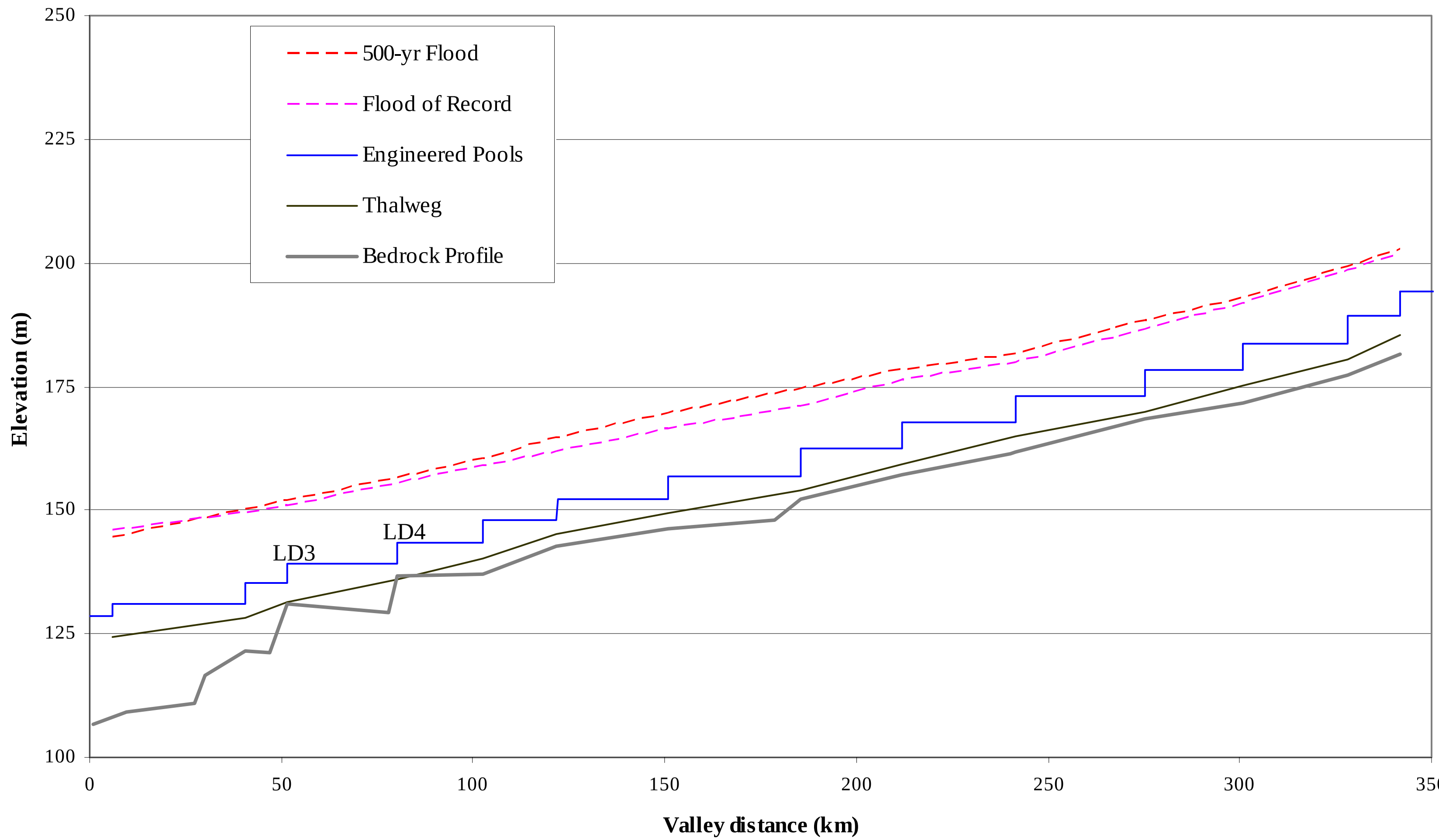


Figure 26. Longitudinal profile of the Kentucky River showing the profile of the buried bedrock valley.

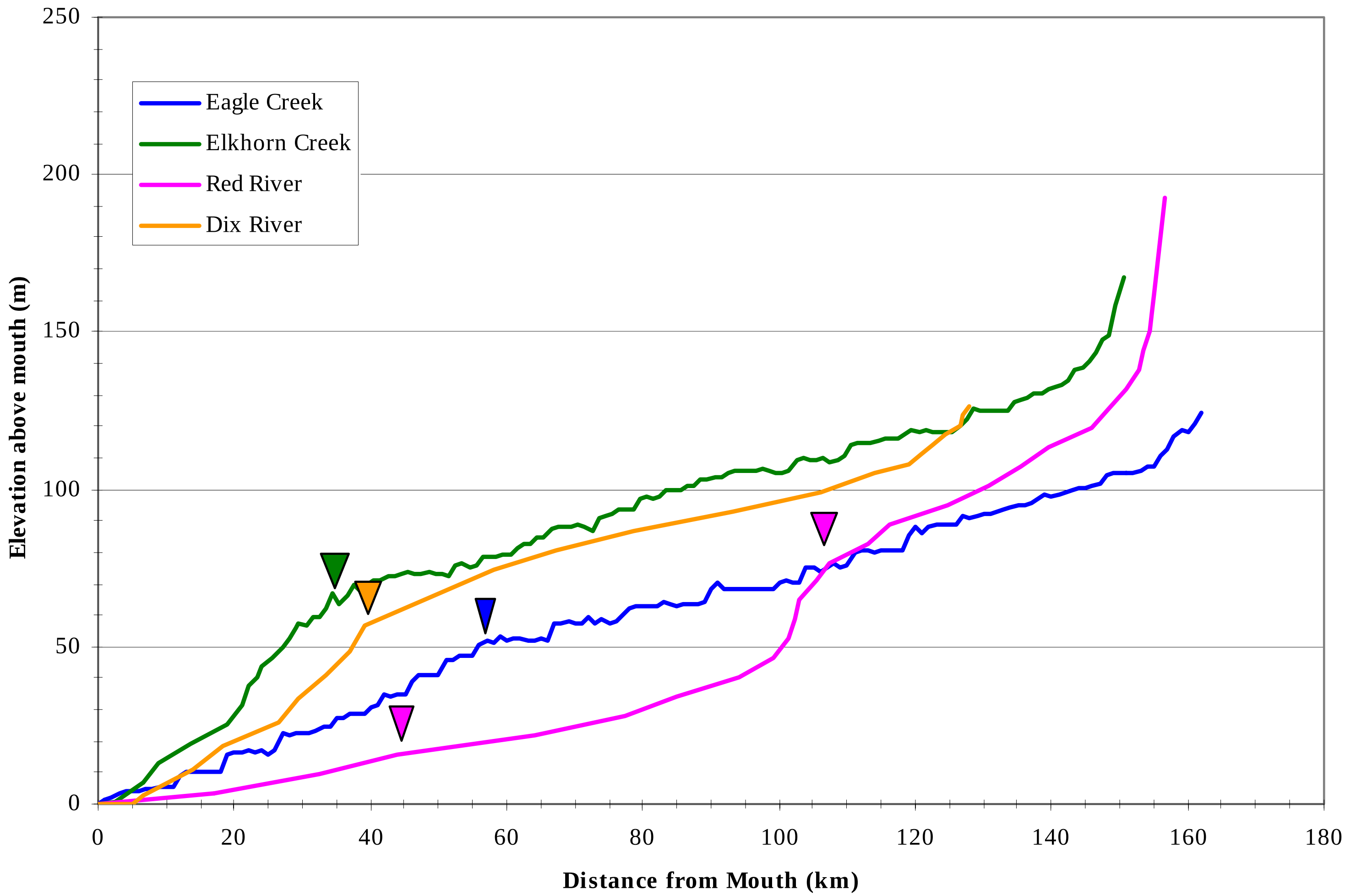


Figure 27. Longitudinal profiles of Kentucky River tributaries with drainage areas greater than 1,000 km^2. Triangles denote the location of knickpoints. Drainage area for each stream is listed in Table 3.

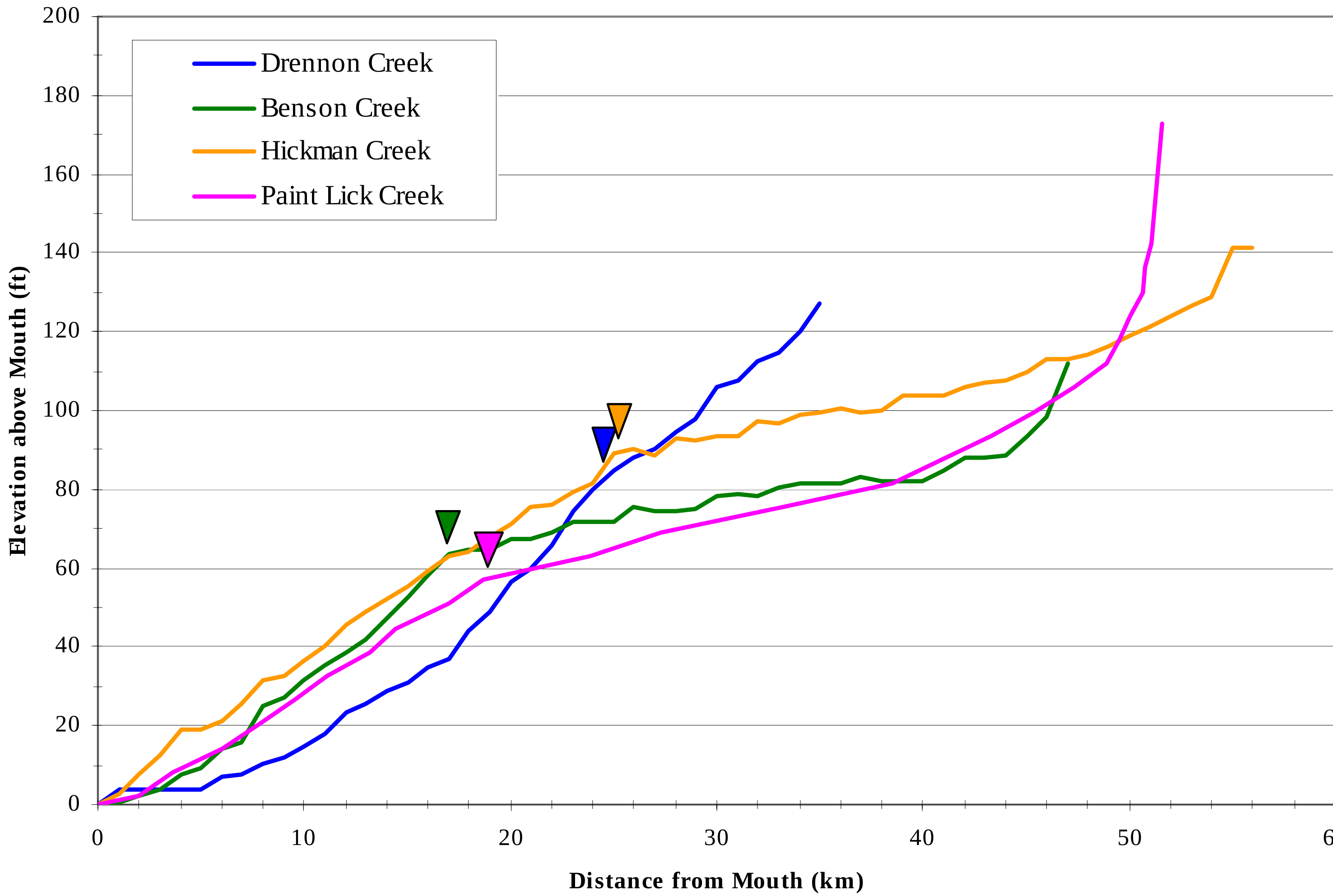


Figure 28. Longitudinal profiles of Kentucky River tributaries with drainage areas between 250 and 285 km^2. Triangles denote the location of knickpoints. Drainage area for each stream is listed in Table 3.

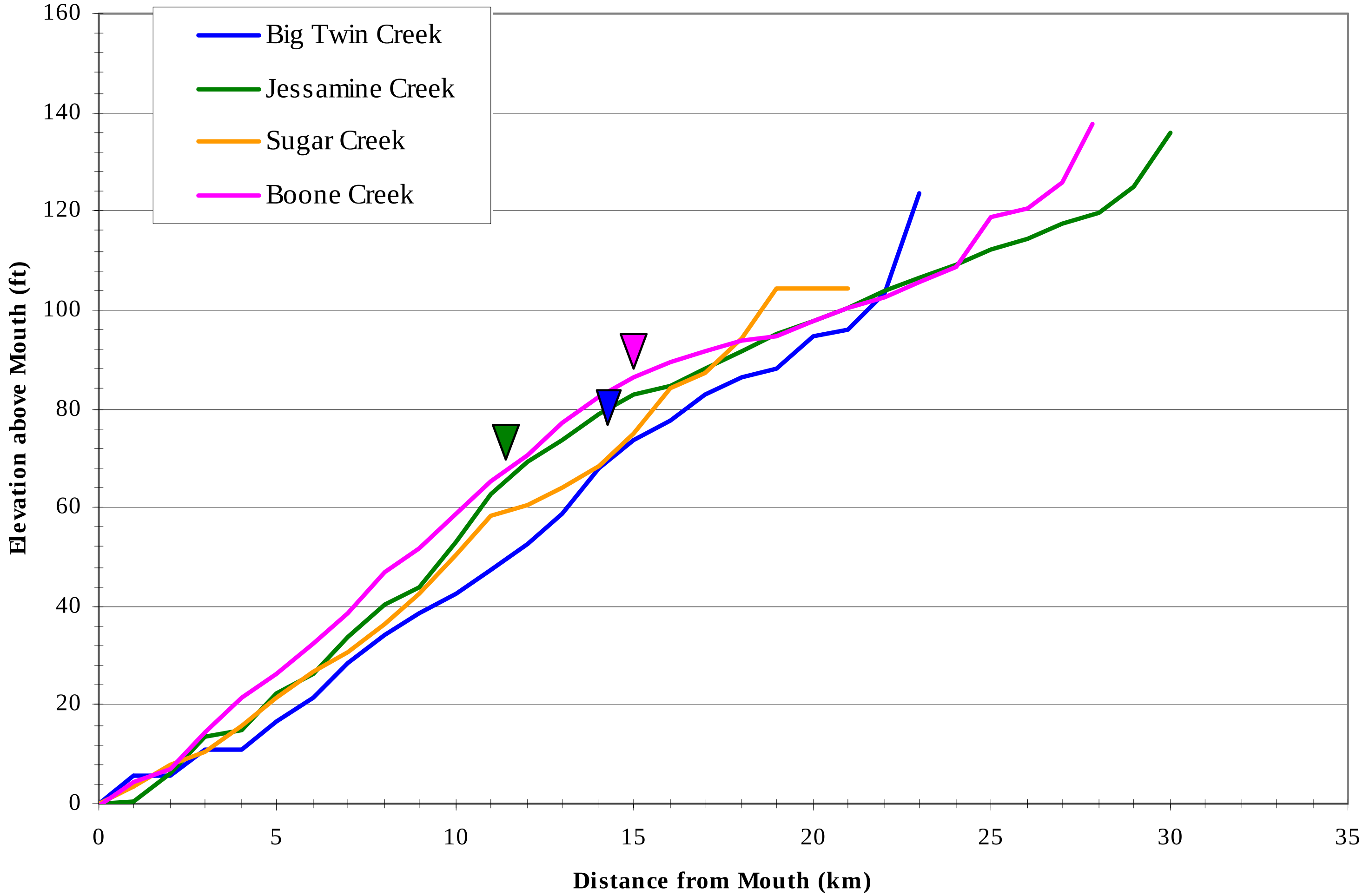


Figure 29. Longitudinal profiles of Kentucky River tributaries with drainage areas between 100 and 200 km^2. Triangles denote the location of knickpoints. Drainage area for each stream is listed in Table 3.

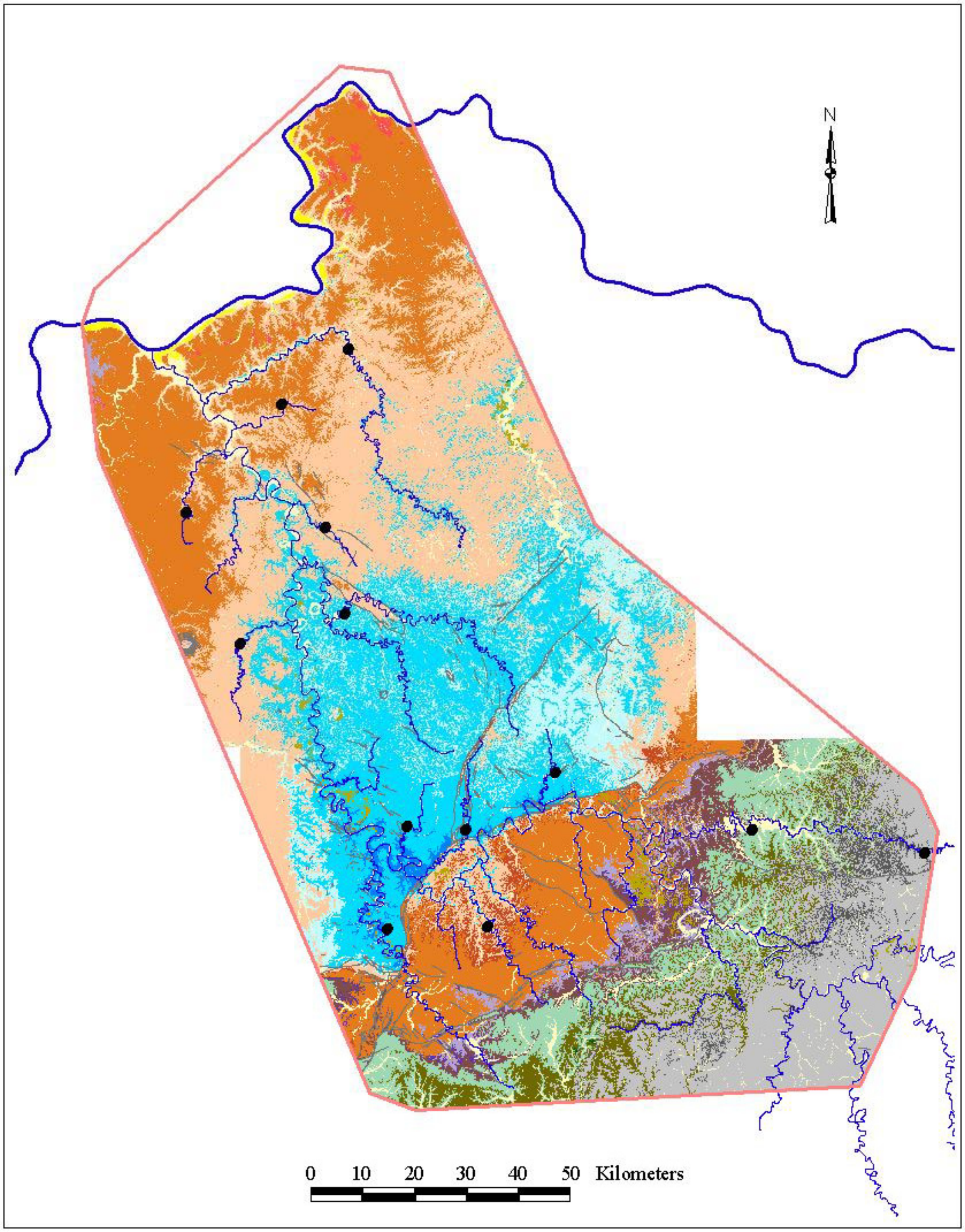


Figure 30. Distribution of knickpoints (shown by black dots) on major tributaries in relation to bedrock geology. Most knickpoints are close to, but do not directly coincide with, bedrock contacts.

As one component of his study of the response of the Kentucky River to base-level change, Warwick (1985) surveyed 16 small tributaries of the Kentucky River. Each tributary he examined had a drainage area less than 2.5 km^2. Although Warwick (1985) acknowledged lithologic control on knickpoint development, he inferred the progressive migration of a major knickpoint up the Kentucky River as the major controlling factor in knickpoint development in tributary streams. He cited progressive decrease in distance of the knickpoints from the mouths of the tributaries as evidence for this interpretation (Figure 31). Statistical regression of his data yields an R^2 value of 0.47. Warwick (1985) measured knickpoint distances to the bottom of the knickpoint slope. This study re-evaluated Warwick's profiles using the top of each of the knickpoints. In several streams, no clear knickpoint was identifiable in the profile data. Regression of the revised knickpoint-distance data yields an even lower R^2 of 0.27 (Figure 31).

To further test Warwick's hypothesis, a similar analysis of the knickpoints on selected larger tributaries of the Kentucky River (drainage area >200 km^2) was undertaken. Profiles of the selected tributaries were constructed from digital topographic data. Knickpoints were identified on the basis of the prominent changes in slopes of the longitudinal stream profiles (Figures 27, 28, 29). The distance from the mouth of the tributary to the knickpoint and the drainage area of the tributary basin above the knickpoint were measured using GIS software. The results show no correspondence between distance from the mouth of the Kentucky River and the distance of a knickpoint from the mouth of the tributary (Figure 32; $R^2 = 0.002$). On the basis of these results, distance upstream from the mouth of the Kentucky River is inferred to be poorly related, if at all, to the distribution of knickpoints along tributary streams. This refutes the conclusions of Warwick (1985).

Fluvial erosion and bedrock quarrying are most effective with high flow, and drainage area is a useful proxy for maximum potential flows (Knighton, 1998). On the basis of this concept, knickpoint-distance data from the larger tributaries (drainage area > 20 km^2) were compared to the drainage area above the knickpoint in the tributary basin. The results are plotted in Figure 33; regression of the data yields R^2 equal to 0.8709. A similar comparison of the revised data from Warwick (1985) yields an R^2 of 0.564 (Figure 34). These results suggest a strong correlation between drainage area and knickpoint migration, especially in the larger tributary drainage basins. The distribution of data relative to the regression line on Figure 33 is also apparently related to lithology. Data points which plot higher on the graph, with larger

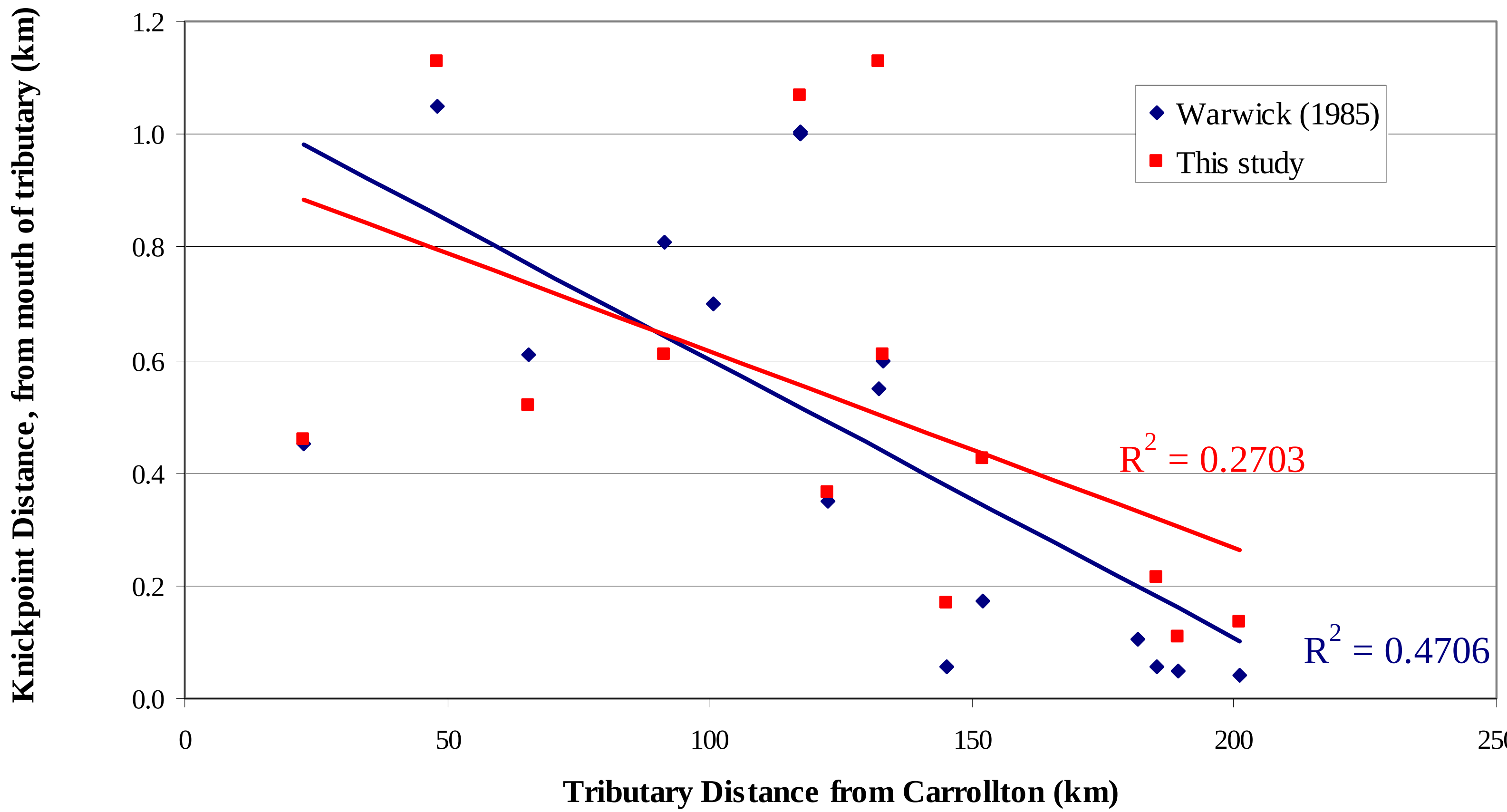


Figure 31. Comparison of knickpoint distance versus tributary distance for small tributaries (drainage area less than 2.5 km^2) as measured by Warwick (1985) and as revised by the current study.

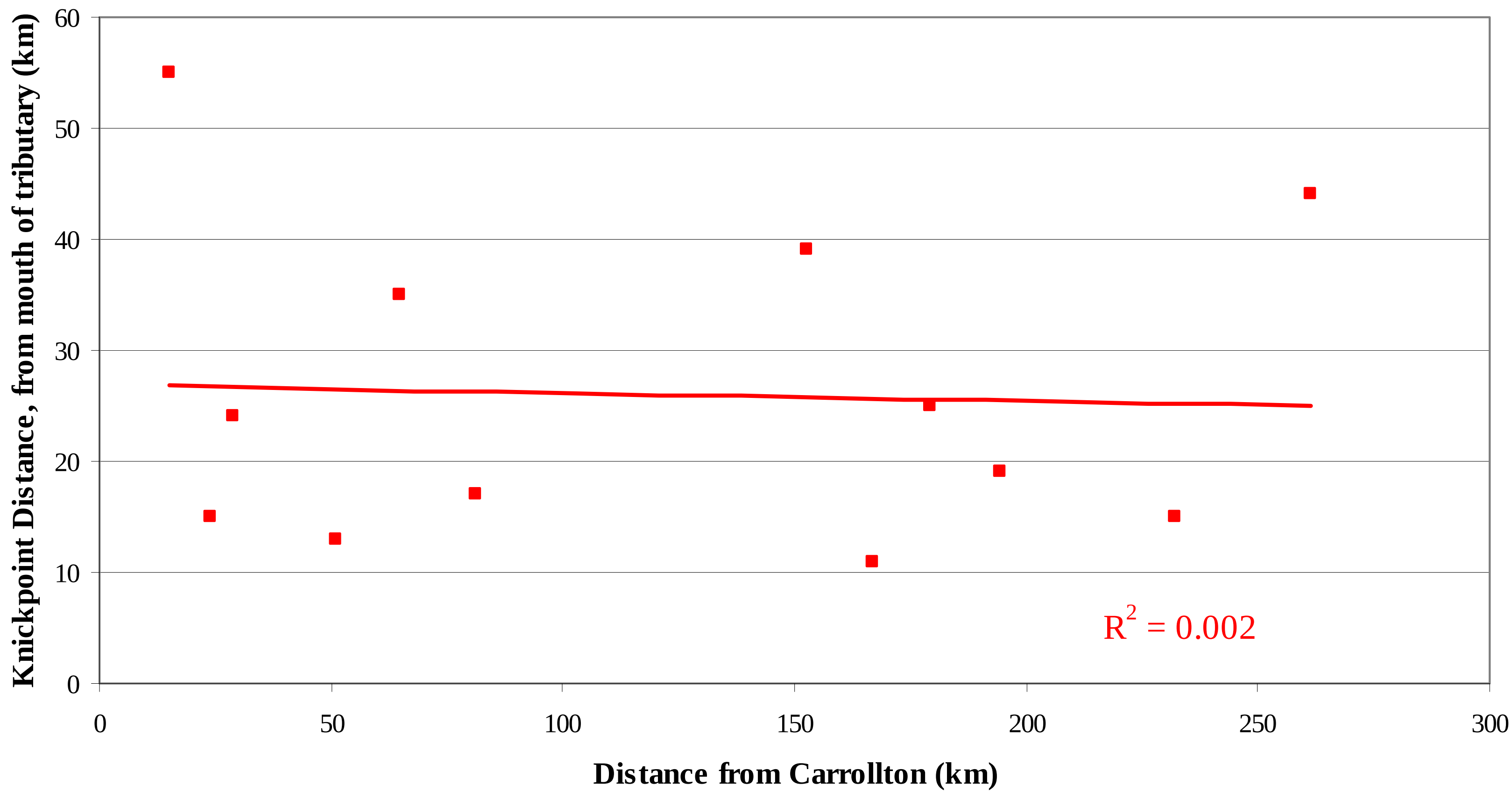


Figure 32. Comparison of knickpoint distance versus tributary distance for larger tributaries as measured in the current study.

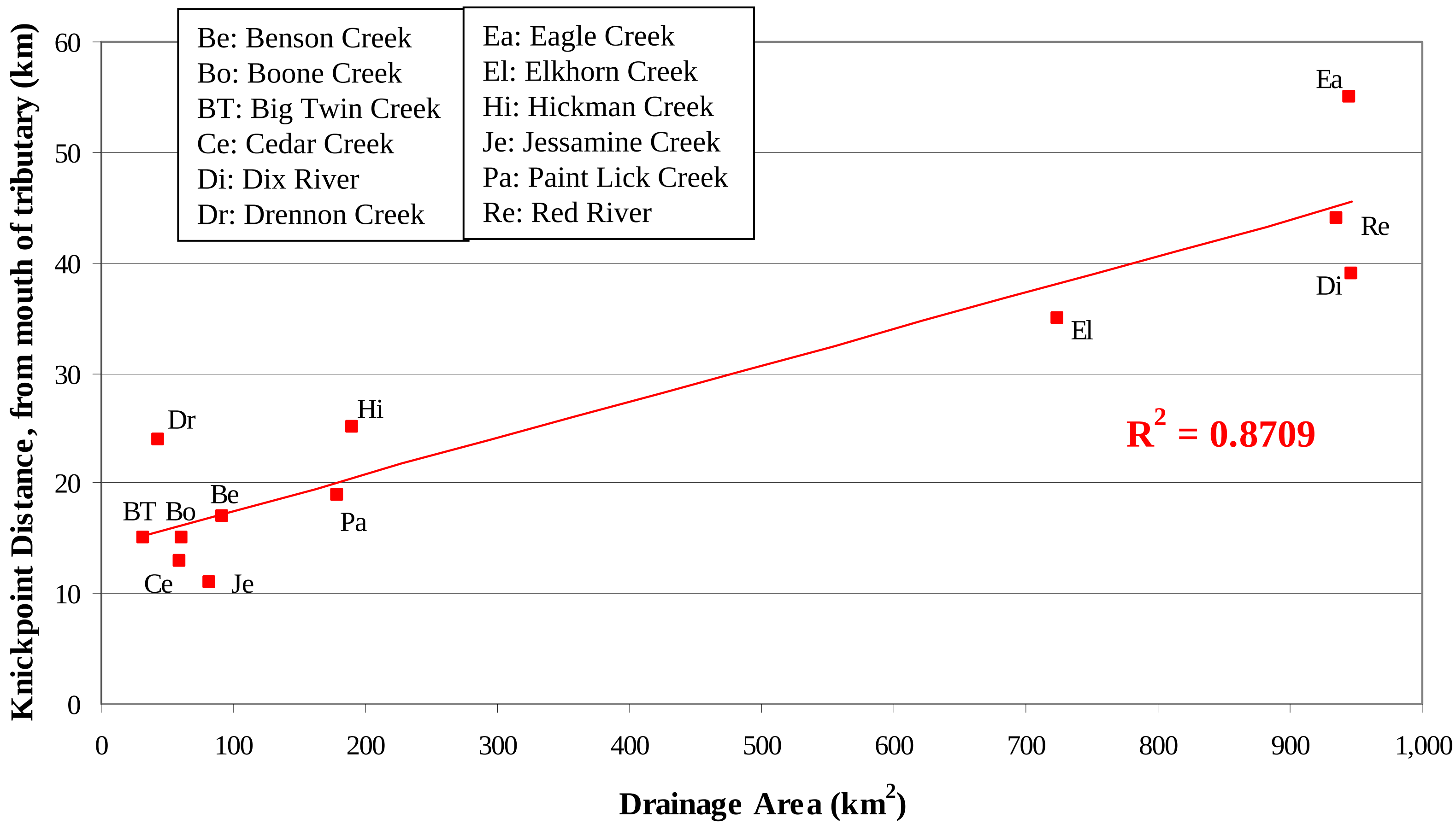


Figure 33. Comparison of knickpoint distance versus drainage area for larger tributaries as measured in the current study.

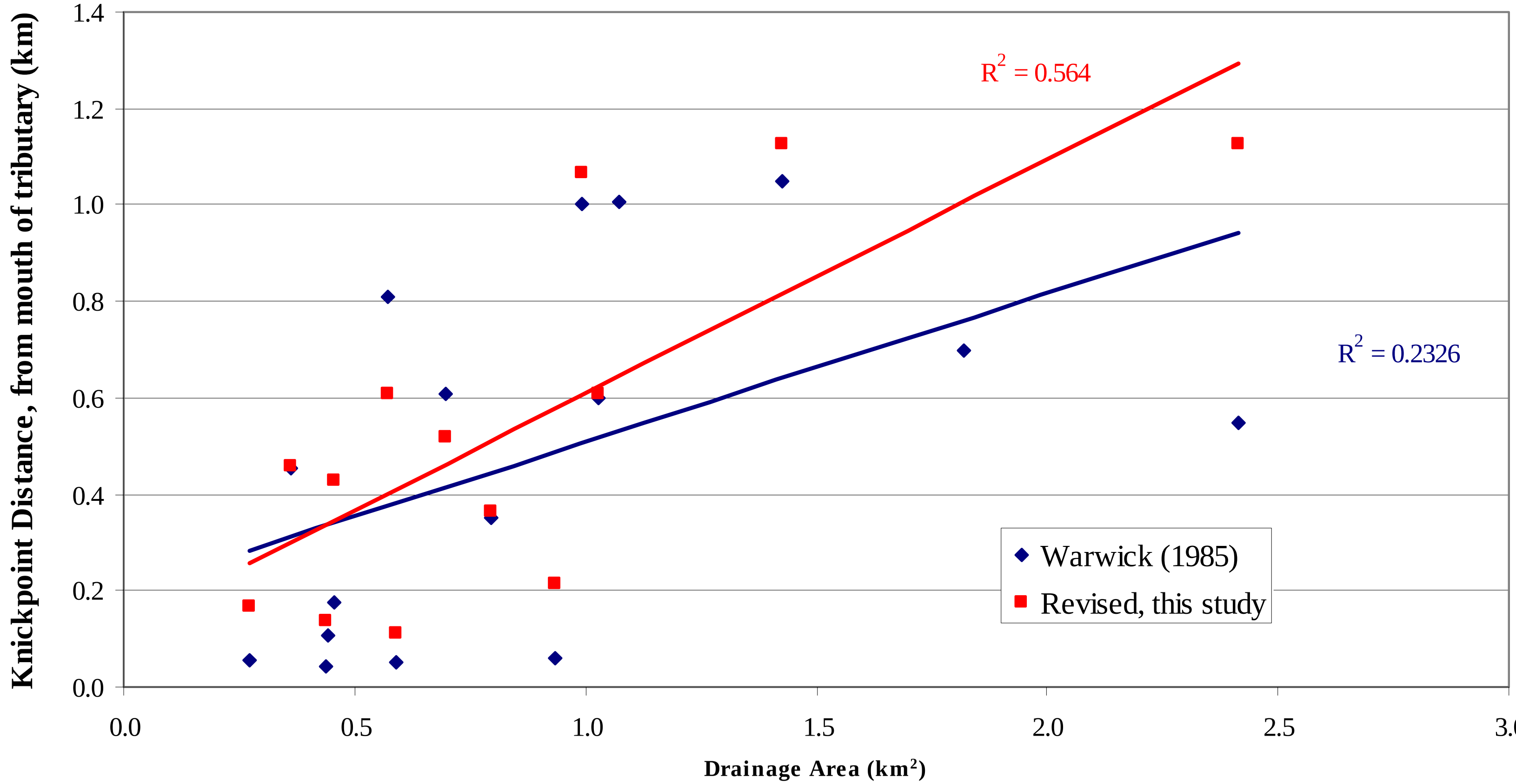


Figure 34. Comparison of knickpoint distance versus drainage area for small tributaries as measured by Warwick (1985) and as revised by the current study.

knickpoint distances, represent knickpoints developed on bedrock containing more shale than those plotting lower on the graph. This suggests that knickpoint migration is primarily controlled by the upstream drainage area of the tributary basin, and is modified by the underlying lithology.

Flood Frequency and Magnitude

The U.S. Army Corp of Engineers (USACoE) has calculated flood frequency graphs for the Kentucky River (Appendix 4), and has a database of high-water mark information for historical floods. This data can be used to identify the parts of the river valley within the active flood zone, and thus potentially susceptible to Holocene modification. The maximum recorded flood crests, 1937 below Lock and Dam No. 8 and 1913 above, are in general agreement with the 0.2%-annual probability (500-year flood) flood calculated by the USACoE (Figure 35). Whether this prediction of flood frequency is accurate or not, the close association with historical flood-of-record information provides a plausible estimate of the fluvially active part of the Kentucky River valley: those areas susceptible to inundation during the highest flows. When considered with the knowledge of significant land-use and engineering modifications of the valley, any surficial deposits within the zone of the flood-of-record are likely to be geomorphically active, and therefore responsive to modern controls. The deposits in this active zone are likely to be thin, discontinuous, and have complex relationships to older deposits. Other than archaeological data, no age control is available for these deposits. Therefore, a detailed chronological analysis of these deposits will not be included in this study.

The stream-profile data maintained by the USACoE include profiles of the stream thalweg and the low-bank along the course of the stream valley. Because the low-bank profile dips below the modern engineered pools upstream from each dam, it is assumed that this represents a pre-engineering surface or a remnant/reflection of such. The generalized thalweg profile and the low-bank profile both have an average gradient of 0.15 m/km. The flood profiles have a very similar average gradient of 0.14 m/km.

Valley Morphology

Valley Width

Although valley width is a basic geometric descriptor of valley morphology, it is a difficult and arbitrary parameter to quantify in alluviated valleys. To maintain consistency, this

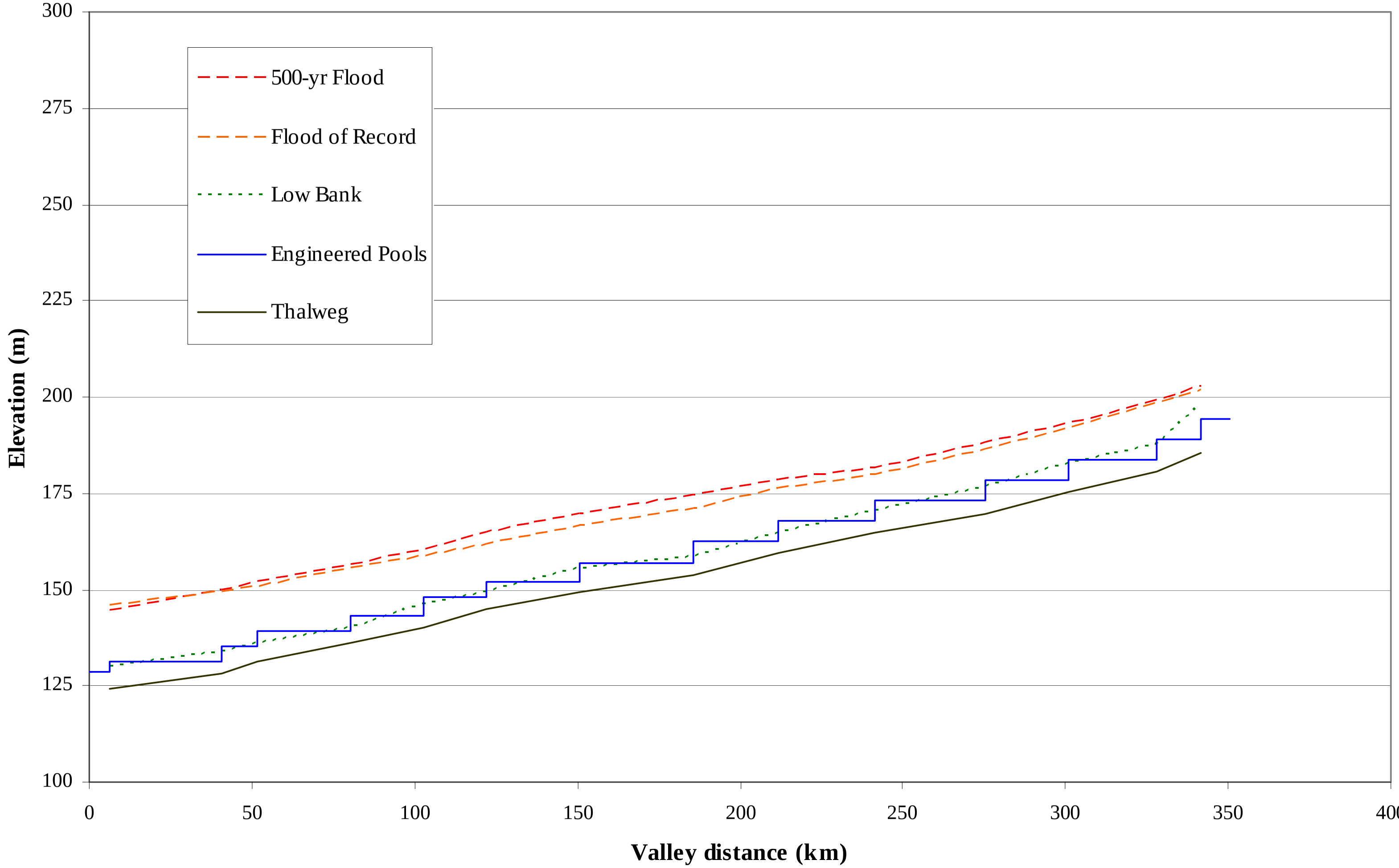


Figure 35. Longitudinal profile of the Kentucky River showing the profile of the calculated 500-year flood, and the flood of record, as recorded by the US Army Corps of Engineers.

study used the width of mapped alluvium as illustrated on USGS 1:24,000 geologic maps (referenced in Appendix 1); most mappers appear to have used slope-break along valley margins as a proxy for the boundaries of alluvial deposits. The measurement of the width of mapped alluvium used here (Figure 36) is a measure of valley width at the level of current alluvial fill. Valley profiles buried beneath the fill are not readily available, and other measurements above the level of the fill would be arbitrary and potentially inconsistent.

Figure 37 shows the width of mapped alluvium along the Kentucky River in the study area. The widths are variable in detail, but general patterns emerge along the length of the stream. Seven general zones of width-pattern can be readily differentiated (Figure 37). In the lower 33 km of the river valley (zone W1), the alluvium is broad, averaging more than 1500 m in width. The sharp valley-width decrease at 10 to 20 km is related to terrace/fan development at the mouth of a tributary valley. In zone W2, the width of alluvial deposits averages about 525 m, but is variable and decreases gradually upstream. In zone W3, the valley and mapped alluvium are quite narrow, averaging only 185 meters wide and decreasing gradually upstream as in area W2. In zone W4, the width averages 240 meters, but varies in a somewhat cyclic pattern in which amplitudes gradually decrease upstream. In zone W5 the width is greater, and is generally more consistent than in zone W4; the alluvial fill averages 380 meters wide. Zone W6 has a much greater valley width, averaging 700 meters wide, but locally is more than 1 km wide. In zone W7, the alluvial fill is much narrower and less internally variable, averaging only 200 meters wide.

Sinuosity

The main stem Kentucky River is a meandering stream along the entire course through the study area. Although the river trend is generally linear, the valley meanders along the linear trend, and the river meanders with the valley, to varying degrees (Figure 38). Two methods can be employed to illustrate or quantify the meandering of the river within the valley. One is to measure the excess river distance by subtracting the cumulative river distance from cumulative valley distance (Figure 38). When river distance is plotted against valley distance (Figure 39), flatter sections on the graph indicate areas where the valley meanders coincide with river meanders closely following the bends of a narrow valley. Steeper segments on the graph indicate those stretches of the river where the stream meanders more broadly within the river valley, and

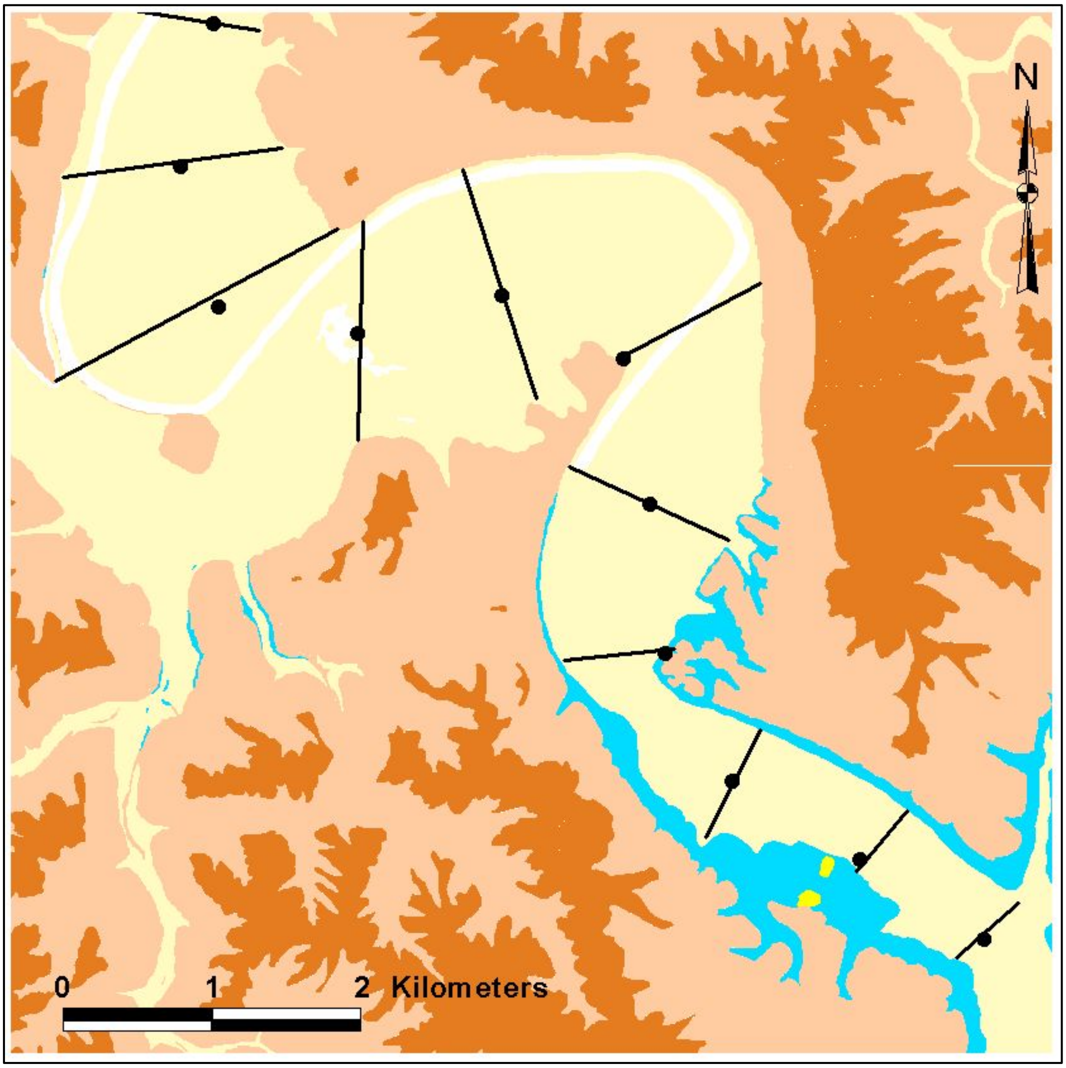


Figure 36. Map illustrating method for measured alluviated valley width. Black lines show locations of measurements at 1-km intervals along the valley. Colors indicate different geologic units: yellow, alluvium; blue, Lexington Limestone; pink, Kope Formation; orange, Upper Ordovician rocks.

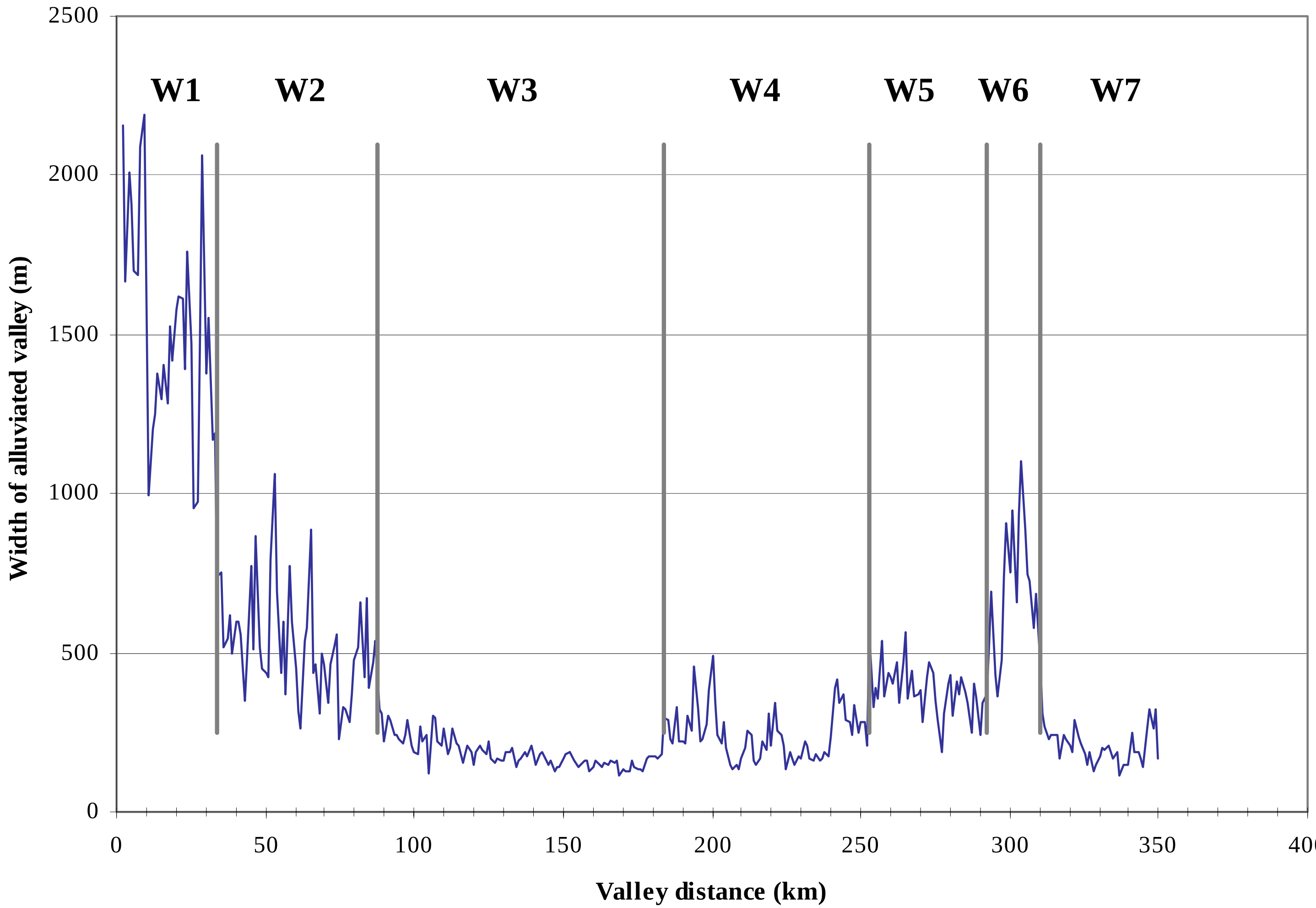


Figure 37. Width of mapped alluvium along the Kentucky River valley, showing inferred width zones.

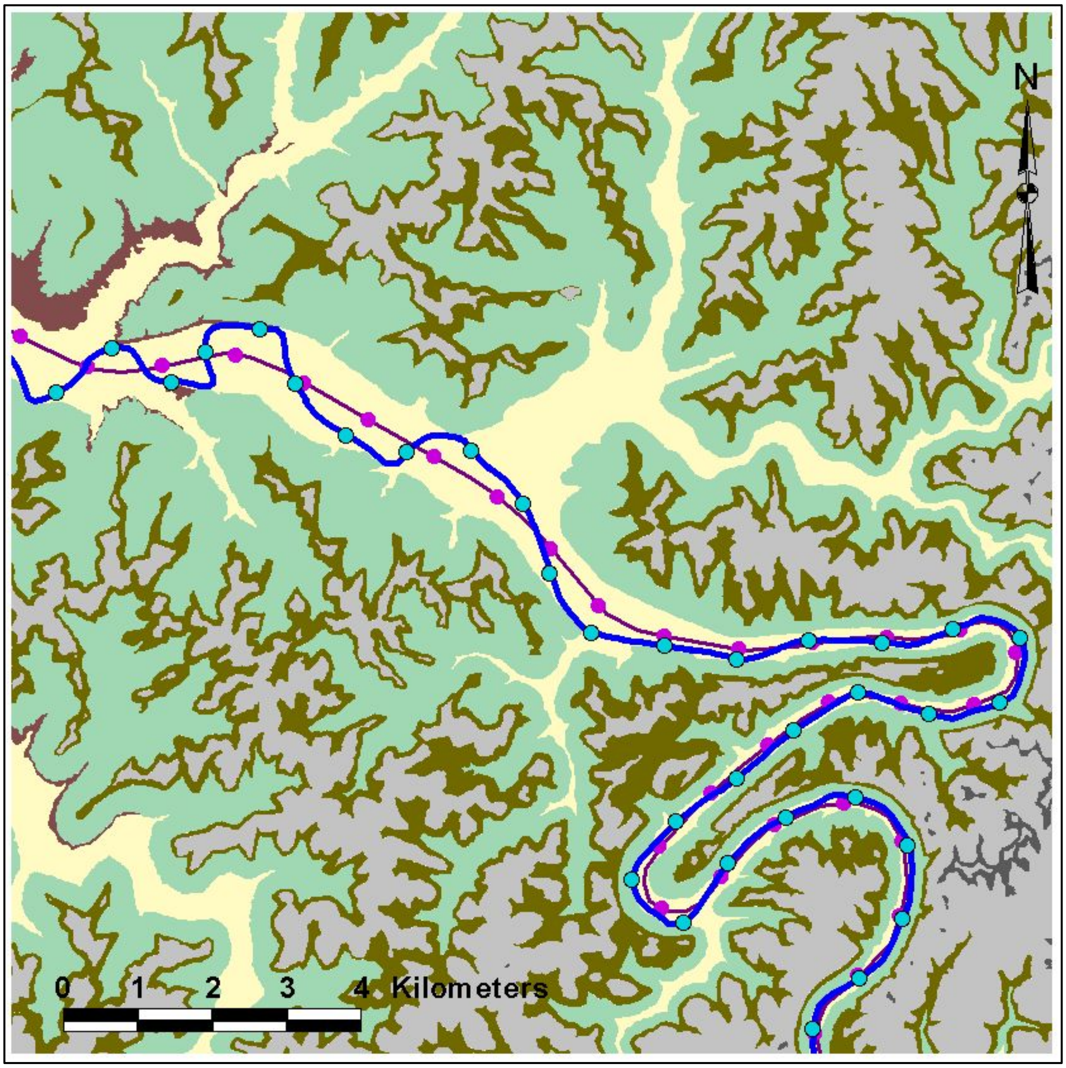


Figure 38. Map illustrating variation in valley distance (purple line) and river distance (blue line). Colors indicate different geologic units: yellow, alluvium; brown, New Albany Shale; light green, Borden Formation; dark green, Slade Formation; gray, Pennsylvanian rocks.

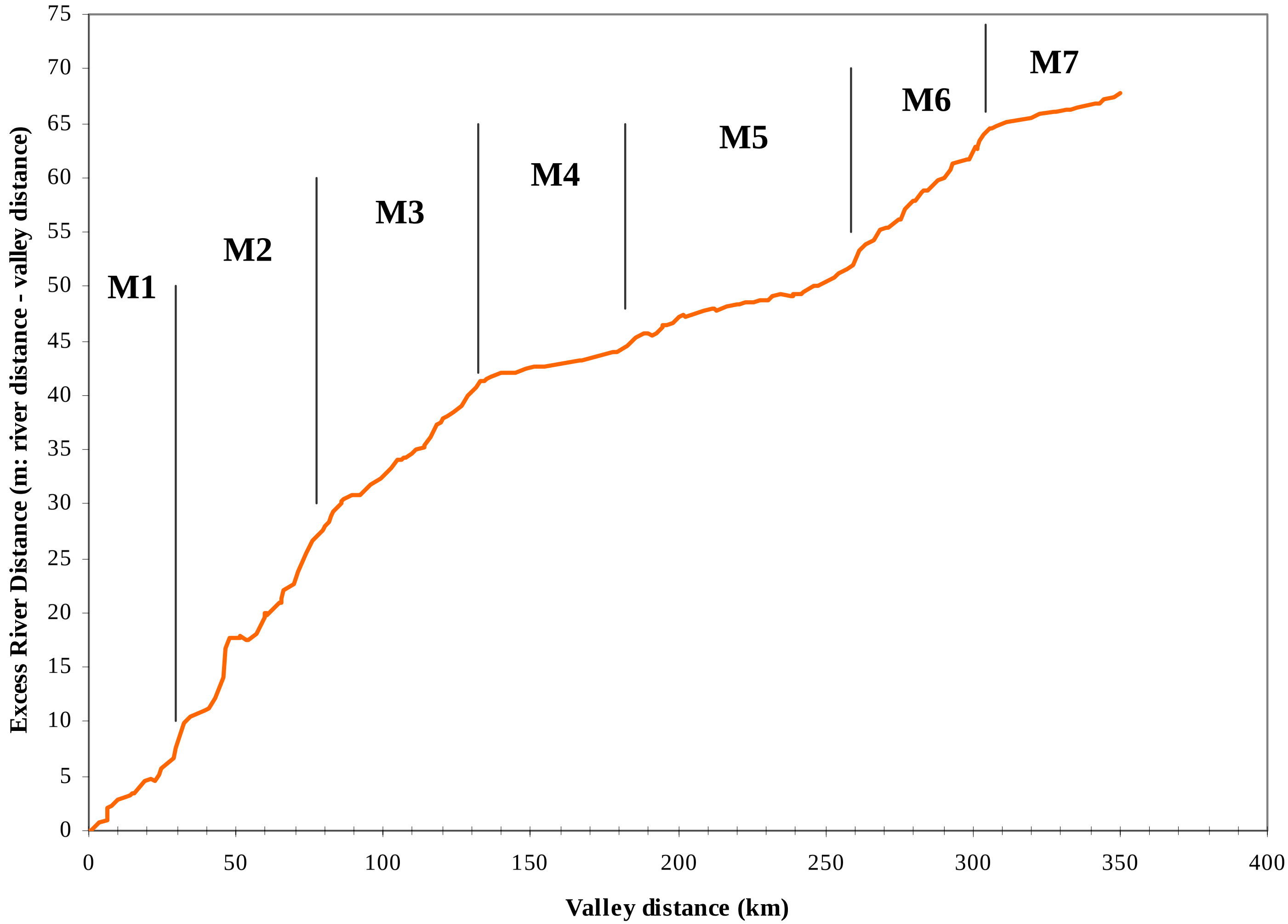


Figure 39. Cumulative excess river distance (vs valley distance) as one measure of sinuosity of the Kentucky River, showing inferred sinuosity zones.

thus river distance increases more dramatically with valley distance. Valley width and sinuosity are related in that higher sinuosity requires a wider valley. The patterns which emerge from each of these methods can be separated into zones of particular behavior.

Seven general zones of sinuosity pattern can be differentiated on Figure 39. In zone M1, the distances diverge almost linearly. In zone M2, the divergence is greater but more variable. A more linear trend characterizes zone M3. In zone M4, the two distances diverge much more gradually; river meanders closely follow valley meanders in a narrow valley. In zone M5, the two distances also diverge only gradually. In zone M6, the values diverge markedly; the river valley meanders in a moderately wide valley. Zone M7 shows minimal divergence between the river and valley distances; river meanders closely follow valley meanders in this narrow stretch of the valley.

Another, and more traditional, method to quantify the sinuosity of the river within the valley is to examine the local ratio of river distance to valley distance for limited stretches of the river (Figure 40). This is a different visualization of sinuosity, so comparable zones can be delineated using this technique. Figure 40 compares the sinuosity ratio to the sinuosity zones identified using the excess-river-length method shown in Figure 39. Zones M1 and M3 have comparable values of sinuosity ratio. Zone M2 includes the highest sinuosity ratio values in the study area, with values greater than 2.0. Zone M4 includes very low values approaching 1.0. The ratio values in zone M5 are somewhat variable, but show a pattern distinct and intermediate between the low values of zone M4 and the higher values of zone M6. Zone M7 has low sinuosity ratios comparable to zone M4.

Valley Morphology Styles

Considered together, valley width and sinuosity can be used to identify eight different valley-morphology styles along the course of the main stem Kentucky River (Figure 41). These valley styles are also apparent in a map view of the basin (Figure 42). Names for the valley styles are derived from 7.5-minute quadrangles in which the river typifies the parameters that distinguish the style. The importance of fluvial versus colluvial processes in slope development and maintenance varies between morphology styles, and is a direct result of different lithologies in the valley walls.

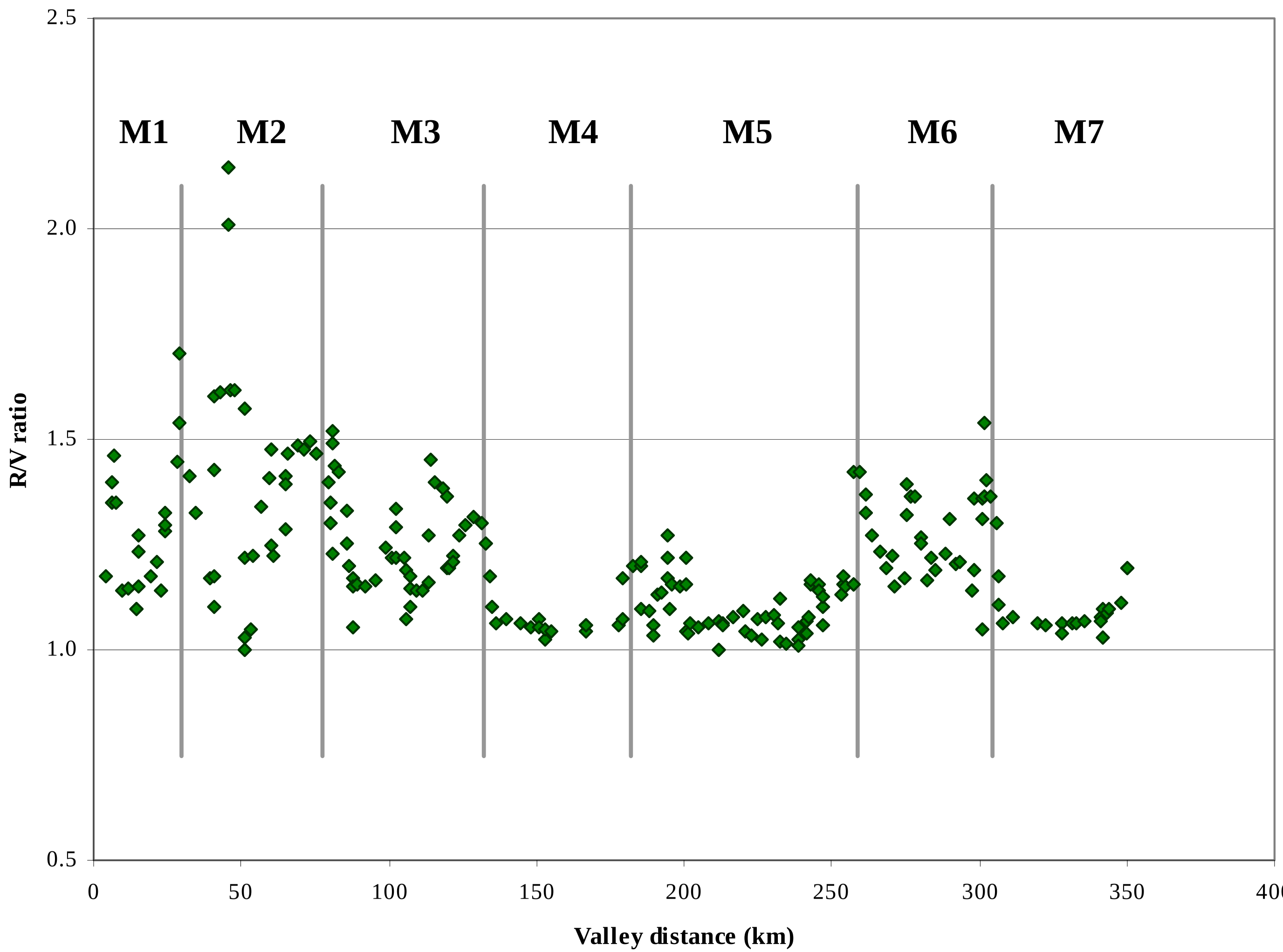


Figure 40. Sinuosity ratio (river distance/valley distance) as another measure of sinuosity of the Kentucky River, showing inferred sinuosity zones from Figure 39.

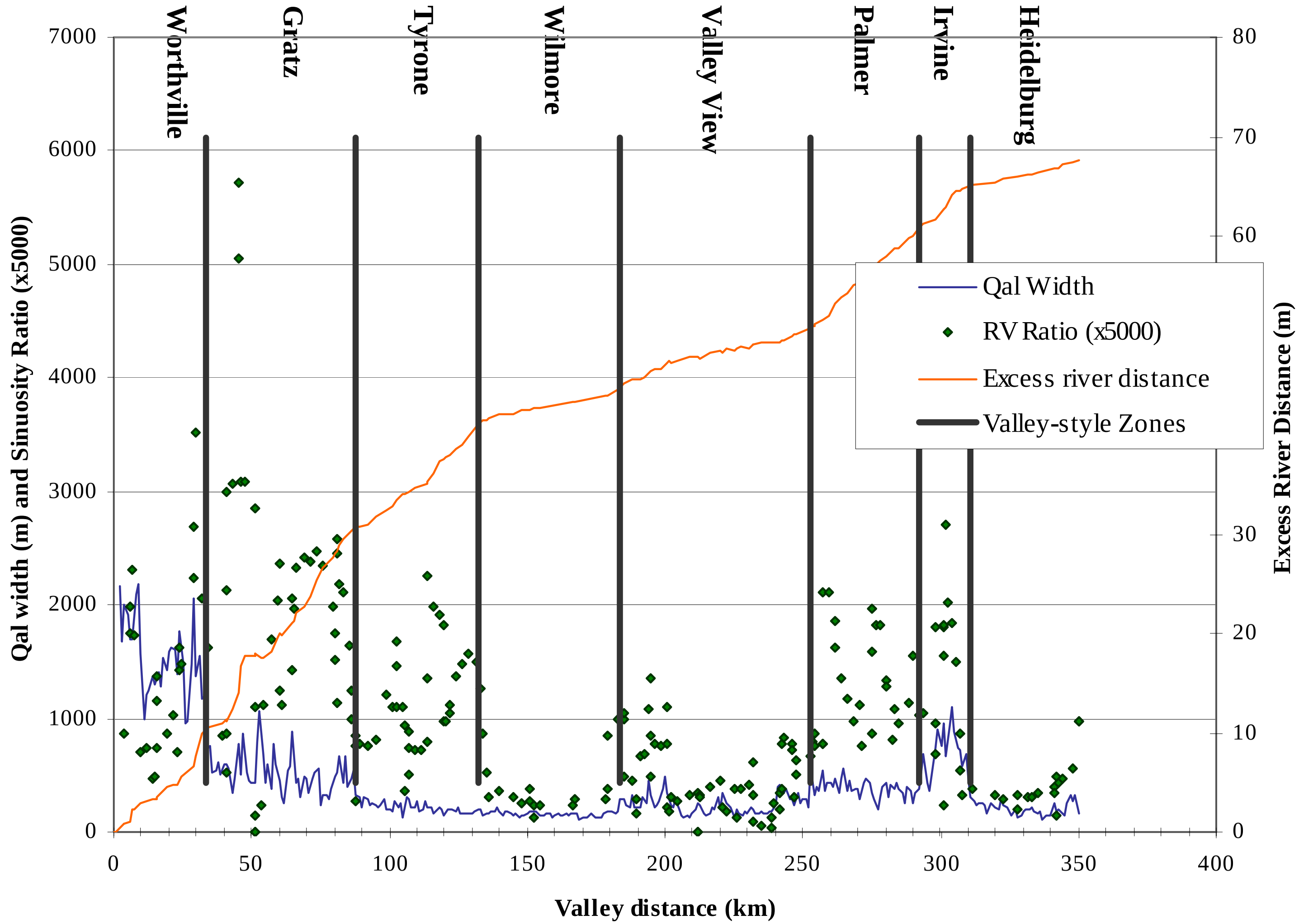


Figure 41. Valley-style regions of the Kentucky River valley derived from alluvial-width, excess-river-distance, and sinuosity-ratio data.

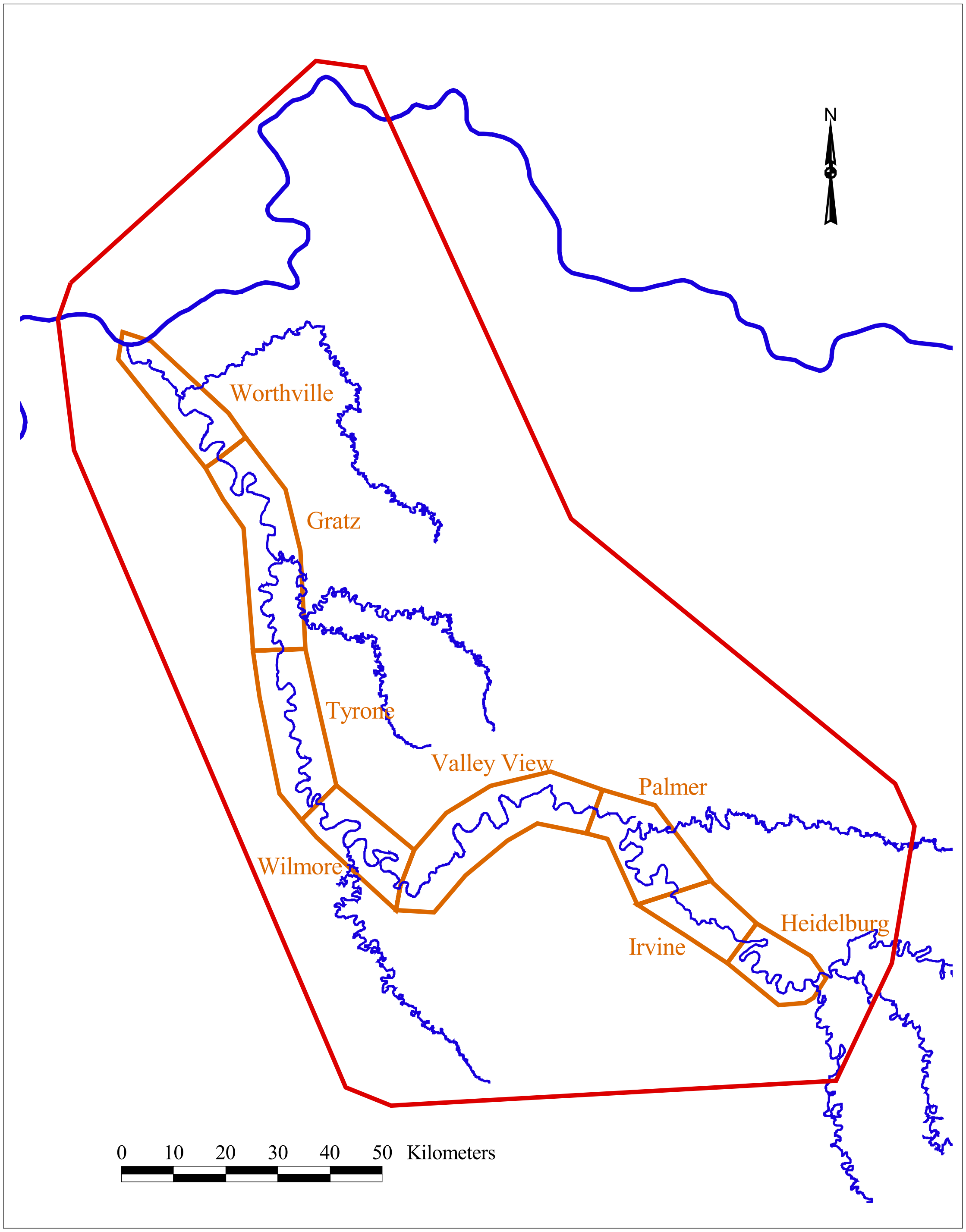


Figure 42. Delineation of valley morphology styles identified in this study.

A relatively wide valley with steep and symmetrical valley walls characterizes the Worthville style (Figures 43, 44a). Although the river meanders within the broad valley, individual reaches of the river are predominantly linear. Shale and limestone of the Kope Formation dominate the valley walls and offer little resistance to lateral or vertical erosion. The symmetrical slopes of the valley walls reflect the colluvial processes dominating the maintenance of those slopes; the river "cleans" colluvial material from footslopes and allows the colluvial processes to dominate. Abandoned meanders are present, but bedrock benches and valley-side terraces are rare, as they are easily undermined and removed by colluvial processes.

Moderate valley widths and more curvilinear river reaches characterize the Gratz style (Figure 44b, 45). This valley style is distinguished from the Worthville style by narrower valley widths and higher sinuosity ratio. Valley walls in the Gratz style are steep along cut-bank slopes of the valley, but less steep on slip-off slopes and on bedrock benches near the valley bottom. The upper slopes are underlain by shales and limestones of the Kope Formation, whereas the lower slopes are eroded into the thin-bedded limestones of the Lexington Limestone. The Lexington Limestone is more resistant to erosion than the Kope. Bedrock benches are typically preserved on the Lexington Limestone, reflecting slower colluvial slope degradation in the more resistant limestone, enabling preservation of the benches. Abandoned meanders are more common in the Gratz style than in the Worthville style.

The Tyrone style has much narrower valleys than the Worthville or Gratz styles, with asymmetric valley walls in meanders, and symmetric valley walls in the intervening straight reaches (Figure 46, 47a). The Tyrone style is distinguished form the Gratz style on the basis of narrower valley widths, lower sinuosity ratios, and a slight change in the excess-length measure of sinuosity. The insides of meanders in the Tyrone style exhibit well-developed slip-off slopes. Stream reaches are somewhat curvilinear as in the Gratz style. The Lexington Limestone and Tyrone Limestone dominate the valley stratigraphy. These units are relatively resistant to erosion, and colluvial processes are relatively slow. The landforms are primarily sculpted by fluvial erosion, and only minimally modified by colluvial processes.

The lower valley walls in the Wilmore style are symmetrical, whereas the upper valley walls are similar to that of the Tyrone style (Figure 47b, 48). The Wilmore style is differentiated form the Tyrone style by very low sinuosity ratios and low excess-length sinuosity. The lower

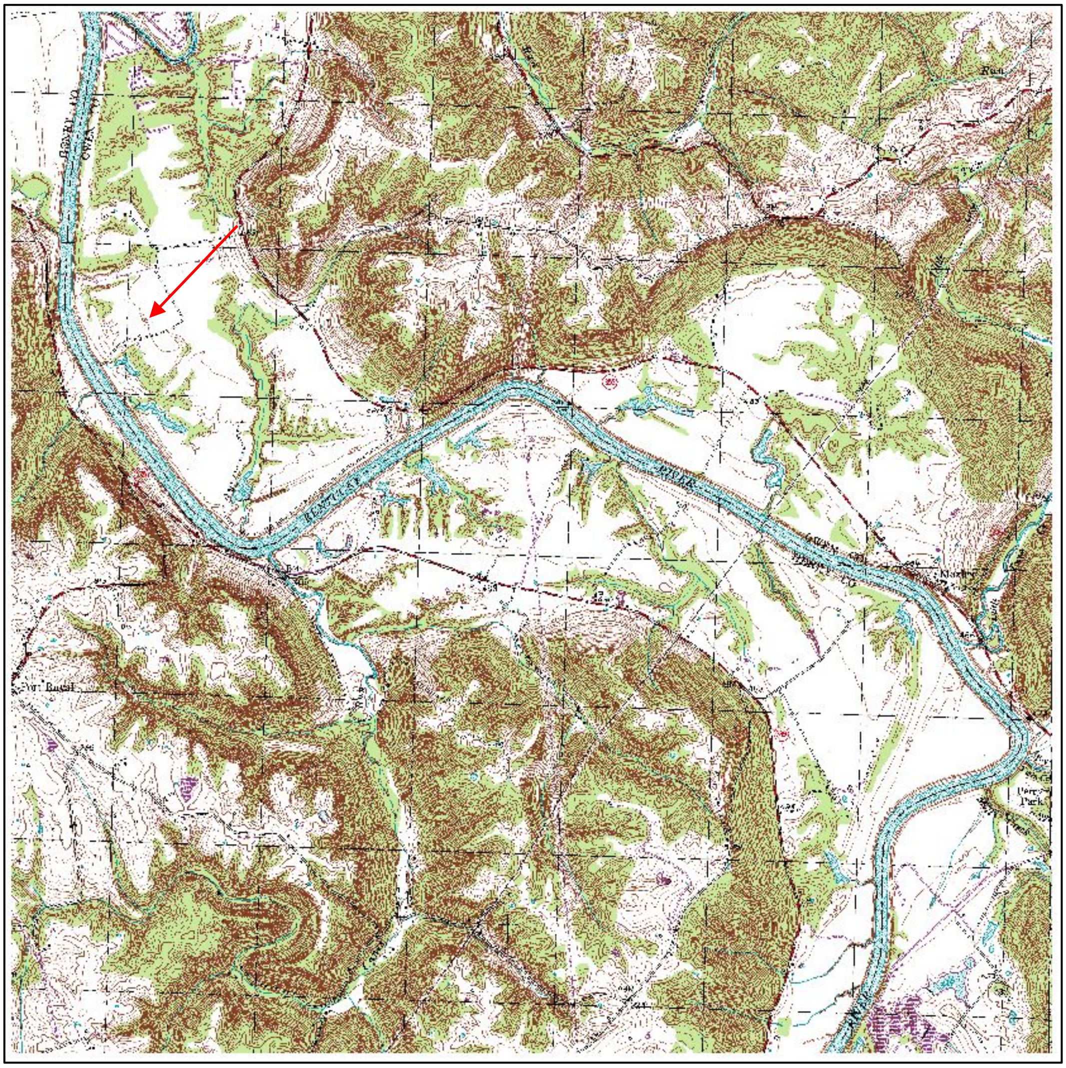


Figure 43. Topographic map from the Worthville 7.5-minute quadrangle, showing the Worthville style of river sinuosity and valley morphology. Red arrow shows the location and direction of view in Figure 44a.

A

B

Figure 44. Photographs of valley morphology styles downstream from Frankfort. A: View across Kentucky River valley near Worthville. Location shown on Figure 43. Kentucky River flows left to right in the tree line behind the barns in the distance. B: View of the Gratz morphology style in Clements Bottom. Location shown on Figure 45.

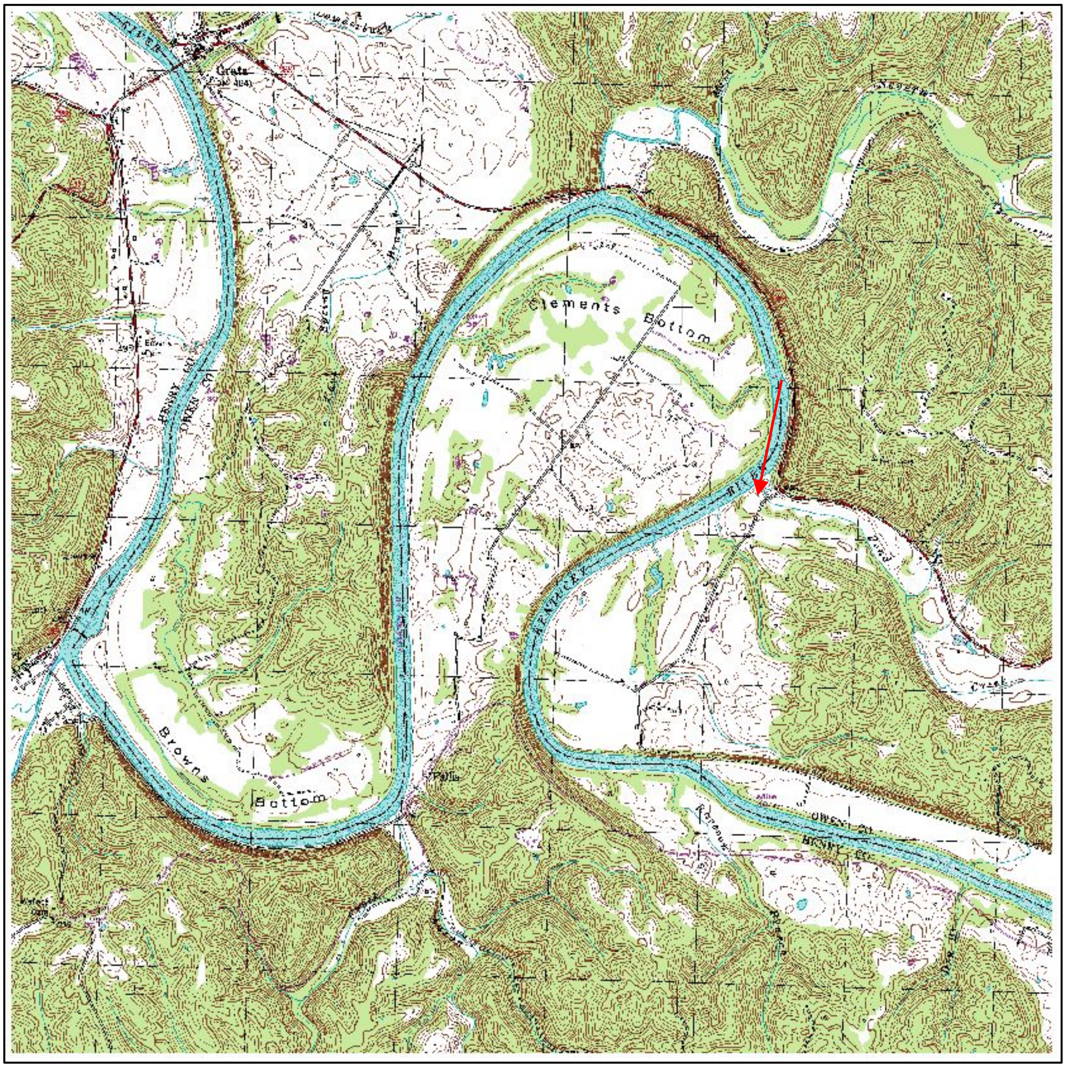


Figure 45. Topographic map from the Gratz 7.5-minute quadrangle, showing the Gratz style of river sinuosity and valley morphology. Red arrow shows the location and direction of view in Figure 44b.

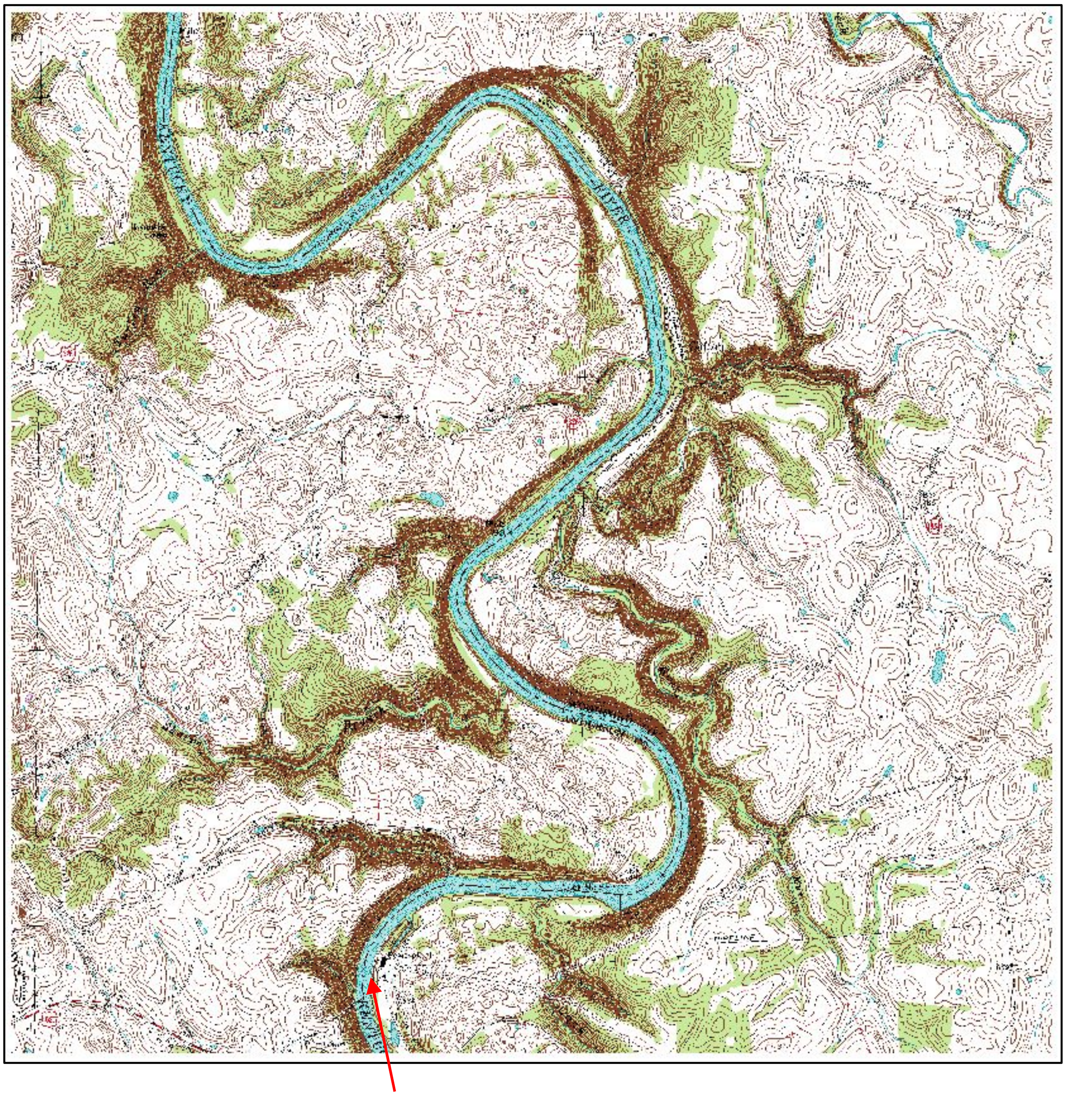


Figure 46. Topographic map from the Tyrone 7.5-minute quadrangle, showing the Tyrone style of river sinuosity and valley morphology. Red arrow shows the location and direction of view in Figure 47a.

A

B

Figure 47. Photographs of valley morphology styles upstream from Frankfort. A: View along Kentucky River valley near Tyrone. Location shown on Figure 46. B: View of the Wilmore morphology style from the US 68 bridge ("Brooklyn Bridge") over the Kentucky River (photo by Brandon Nuttall).

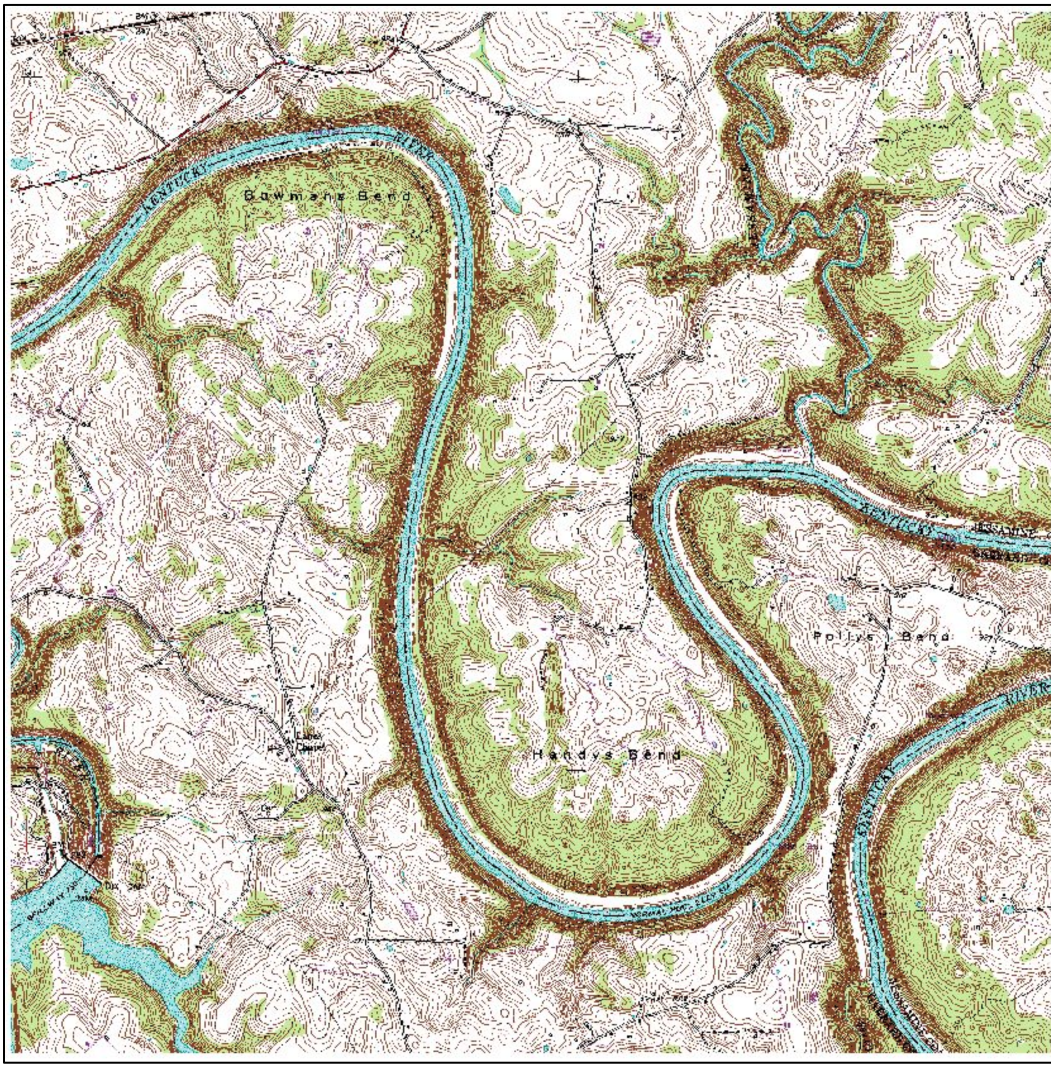


Figure 48. Topographic map from the Wilmore 7.5-minute quadrangle, showing the Wilmore style of river sinuosity and valley morphology. Photograph in Figure 47b was taken west of this map area.

valley is very narrow, and is eroded into the massive Camp Nelson Limestone, which is very resistant to both lateral and vertical erosion. The upper valley walls are in the Tyrone Limestone and lower Lexington Limestone, and closely resemble the morphology of the Tyrone style. Stream reaches are curvilinear, similar to the Tyrone style.

The Palmer style (Figure 49) is similar in valley width and morphology to the Gratz style (Figure 45). The valley is wider than the Wilmore and Tyrone styles, and is comparable to the Gratz style. The valley is asymmetric with bedrock benches and less steep slopes on slip-off slopes. Sinuosity is comparable to the Tyrone style. Colluvial processes maintain the steep slopes. The bedrock is comprised of Upper Ordovician and Silurian carbonates and shales, which vary widely in resistance to erosion. Bedrock benches are typically formed on resistant rocks. Stream reaches are linear, as in the Worthville style.

The Valley View style (Figure 50) is an alternating combination of the Wilmore and Palmer valley styles, where the river crosses and recrosses the Kentucky River fault zone into different lithologies. The Valley View style has widths and sinuosity intermediate between the Wilmore and Palmer styles.

A wide symmetrical valley characterizes the Irvine style (Figure 51). The Irvine style is distinguished by a wide valley compared to adjacent styles. The valley walls are dominated by the Borden Formation, and undermining of overlying lithologies by erosion of the weak Nancy Member of the Borden leads to colluvial processes dominating the steep valley slopes. Stream reaches are linear, but shorter than in the otherwise similar Worthville style.

A narrow symmetric valley characterizes the Heidelburg style (Figure 52). The Heidelburg style has comparable width and sinuosity to the Wilmore style. The valley walls in the Heidelburg style are underlain by massive siltstones of the Cowbell Member of the Borden Formation, which is less susceptible to erosion than the underlying Nancy Member.

Fluvial Deposits and Associated Features

Glacial Deposits and Erratics

Although previous workers had recognized glacial deposits in northern Kentucky, the USGS-KGS geologic mapping program provided the most thorough description and delineation of the deposits. The following description is largely summarized from USGS geologic-

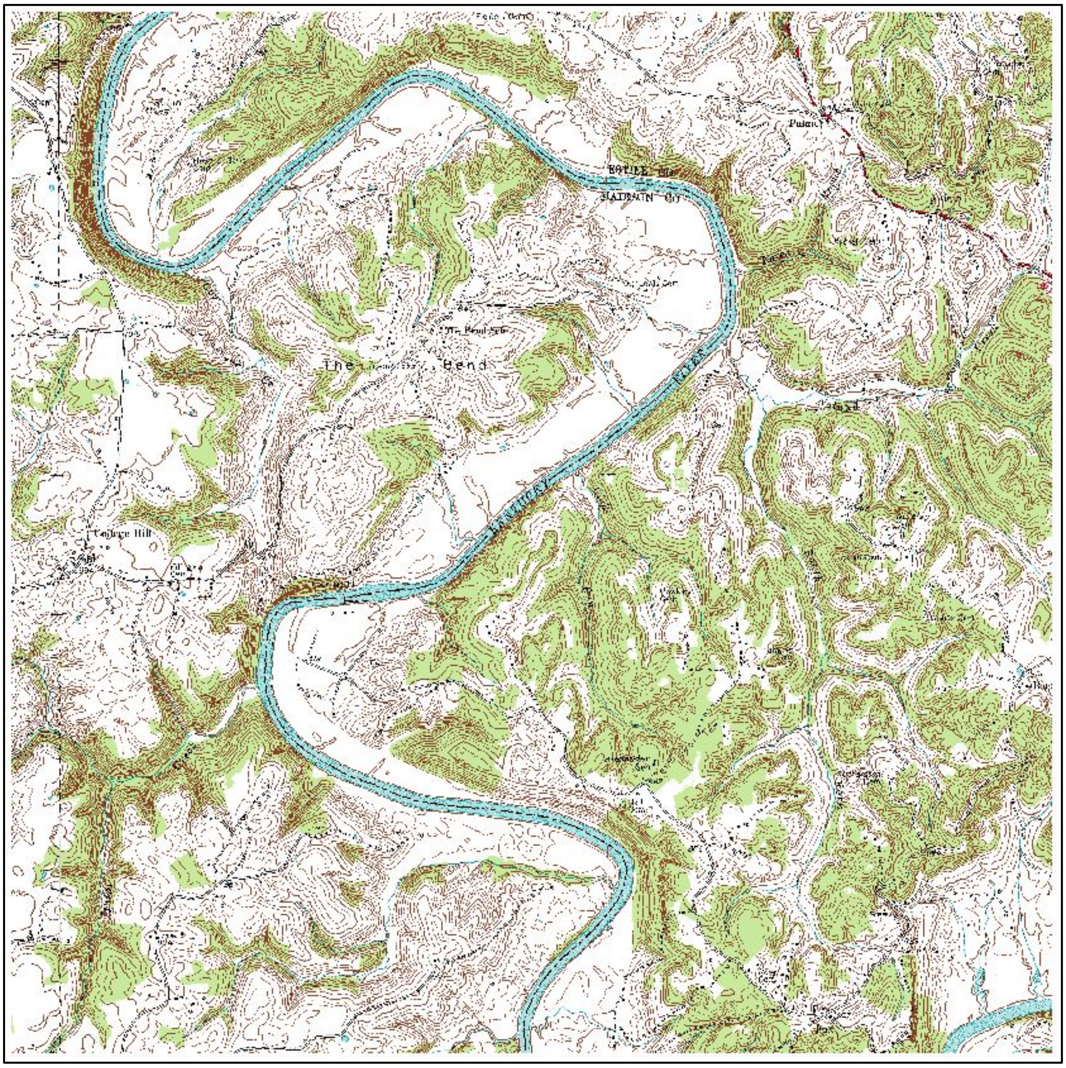


Figure 49. Topographic map from the Palmer 7.5-minute quadrangle, showing the Palmer style of river sinuosity and valley morphology.

Figure 50. Topographic map from the Valley View 7.5-minute quadrangle, showing the Valley View style of river sinuosity and valley morphology.

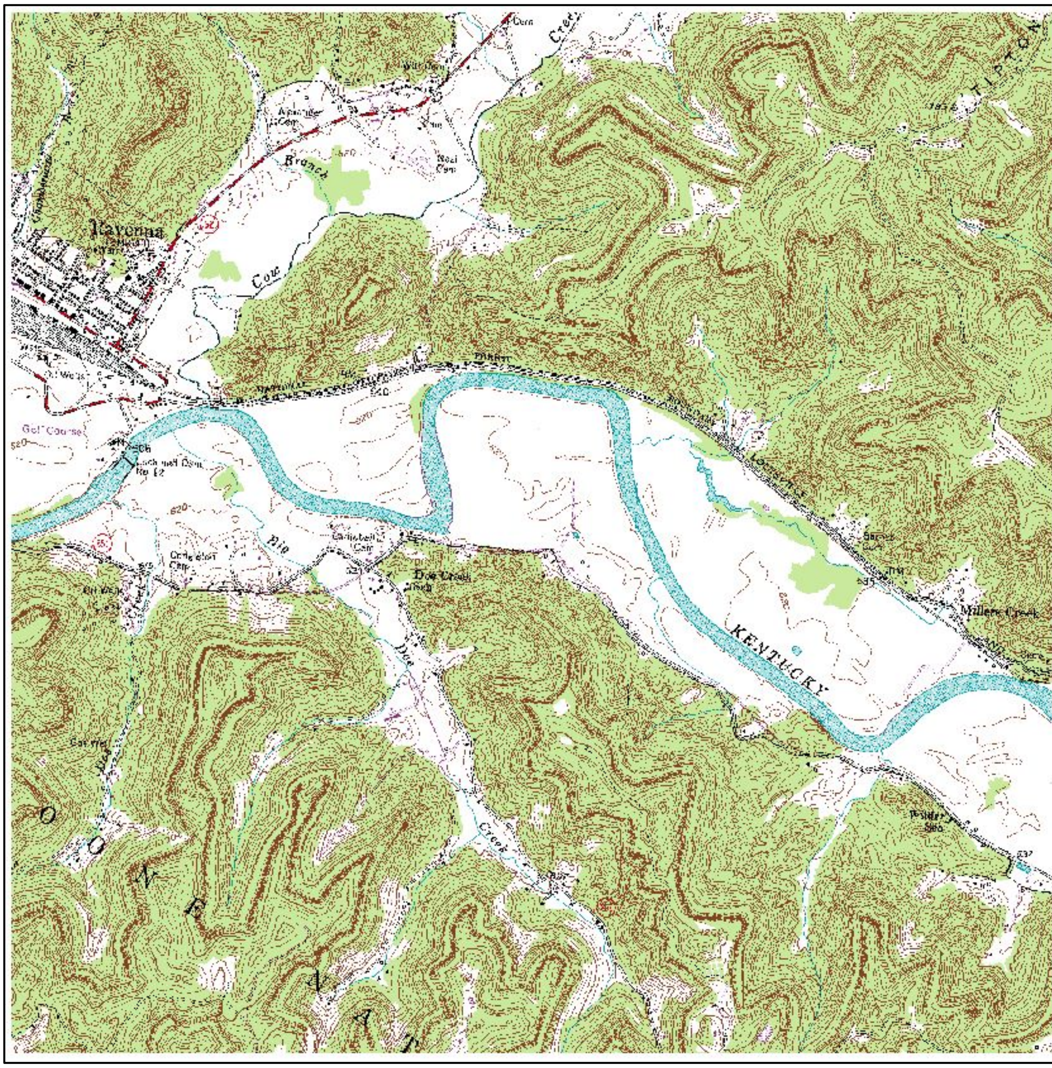


Figure 51. Topographic map from the Irvine 7.5-minute quadrangle, showing the Irvine style of river sinuosity and valley morphology.

Figure 52. Topographic map from the Heidelburg 7.5-minute quadrangle, showing the Heidelburg style of river sinuosity and valley morphology.

quadrangle maps. Figure 53 illustrates a stratigraphic column of the glacial deposits and related strata mapped in northern Kentucky, and Figure 54 shows the distribution of those deposits. Two different units of glacial till/drift have been mapped in northern Kentucky. Both are composed of silty clay, and contain pebbles, cobbles, and boulders of limestone, quartz, chert, and igneous and metamorphic rocks. One drift unit is found only on uplands, and generally no lower than 665 to 670 ft in elevation. It is deeply weathered, and contains limonite pellets. This drift unit ("pre-Illinoisan drift" on Figure 54) overlies Old Kentucky River fluvial deposits in the paleovalleys between Carrollton and Cincinnati (Swadley, 1971) and is commonly separated from the fluvial sediment by a lacustrine fill dominated by laminated calcareous clay (Swadley, 1971; Ettensohn, 1974). The other drift unit ("Illinoisan drift" on Figure 54) is typically found below 650 ft, but locally as high as 720 ft in elevation. It is more sandy, less leached, and is locally cemented by calcite. This less-leached drift is stratigraphically superimposed on the more deeply weathered drift in the valley of Fourmile Creek in the Vevay North and South geologic quadrangle map. The less leached drift has been assumed to be of Illinoisan age, and the more deeply weathered is thus pre-Illinoisan. In the Kentucky River basin, these drift deposits are confined to the area northwest of Eagle Creek.

Presumed glacial erratics of exotic non-sedimentary lithologies are found across northern and northeastern Kentucky, south of the mapped glacial limit (Figure 55) (Ray, 1969). Boulders of igneous and metamorphic rocks were found at elevations as high as 850 ft in the headwaters of the Licking River basin (Leverett, 1929). They are found as high as 985 ft in Lewis County. Two of these boulders, the Epworth and Farmers boulders, were studied by Jillson (1924a, 1924b, 1925). Various authors have speculated about their origin (Ray, 1969), but most regional glacial histories only treat them in passing (e.g., Goldthwait and others, 1981). Leverett (1929) inferred that these erratics were rafted to their locations and deposited during times of glacial impoundment of the area stream valleys.

Wisconsin glaciers did not reach Kentucky, but Wisconsin outwash is found along the Ohio River valley. The outwash is composed of crossbedded gravel, sand, and silty clay. The gravel includes well-rounded pebbles and cobbles of limestone, siltstone, quartz, chert, granite, schist, gniess, coal, and fine-grained igneous and metamorphic rocks. Total thickness of the outwash is as much as 140 ft. The upper 5 to 25 ft of the deposits is dominantly sandy. The

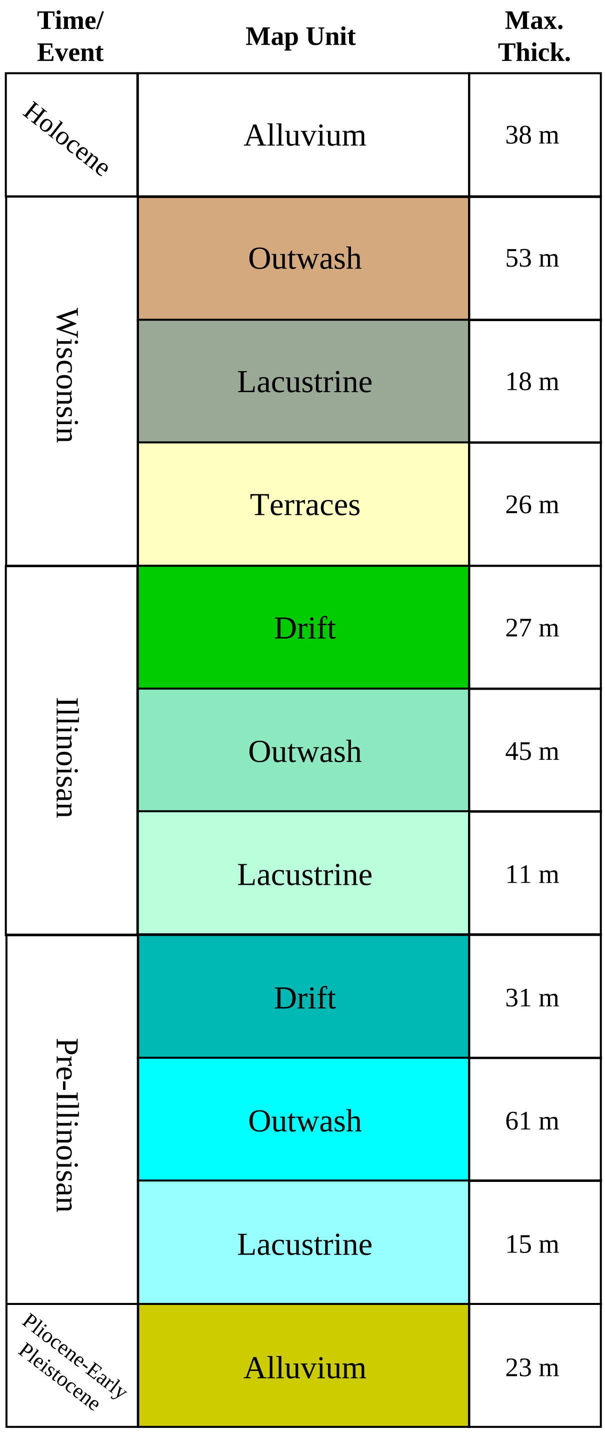


Figure 53. Stratigraphic column of glacial deposits and related stratigraphy in northern Kentucky. Modified from Sparks and others (2002).

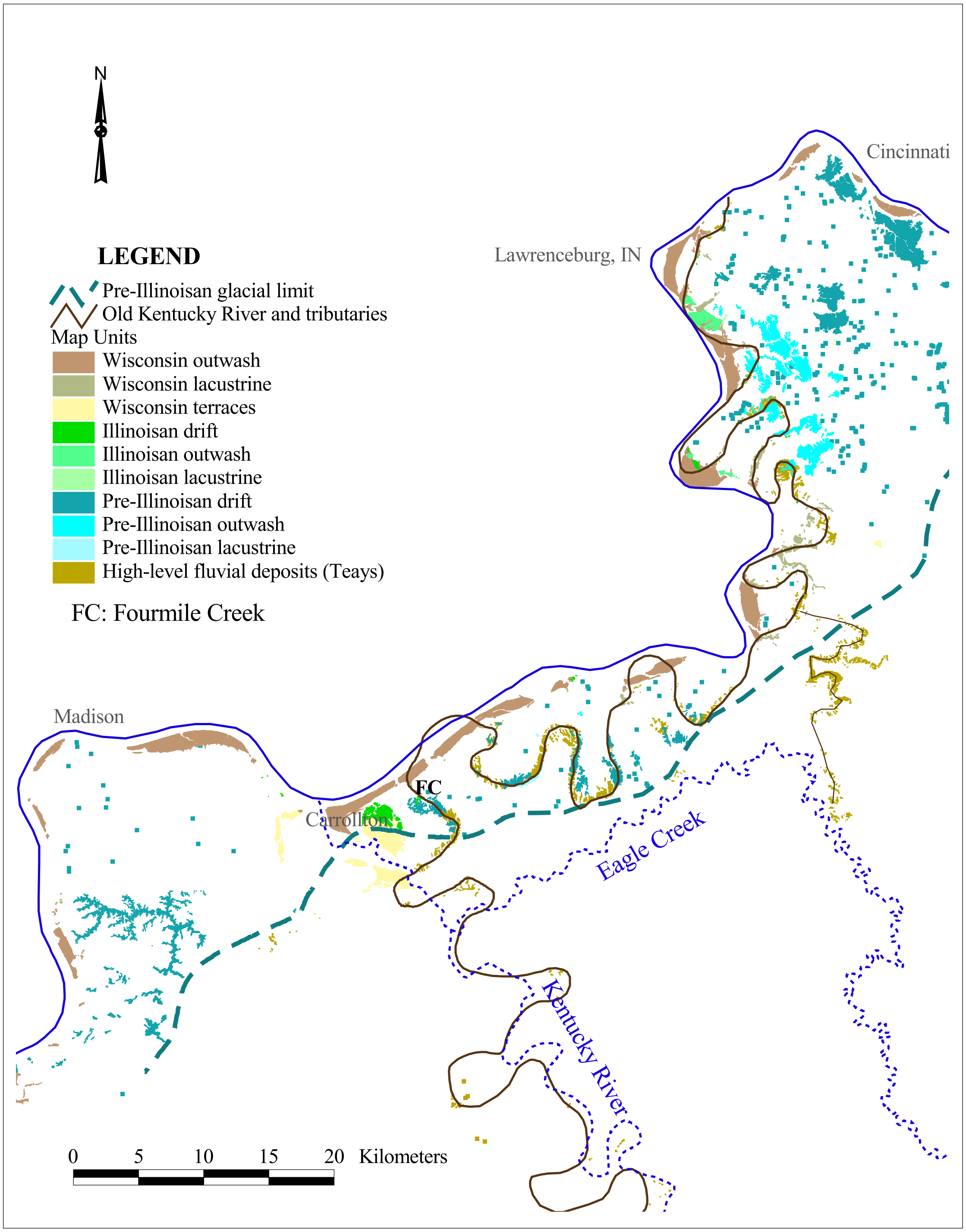


Figure 54. Distribution of glacial deposits and related features near the mouth of the modern Kentucky River. Dashed blue line shows modern streams in the Kentucky River valley

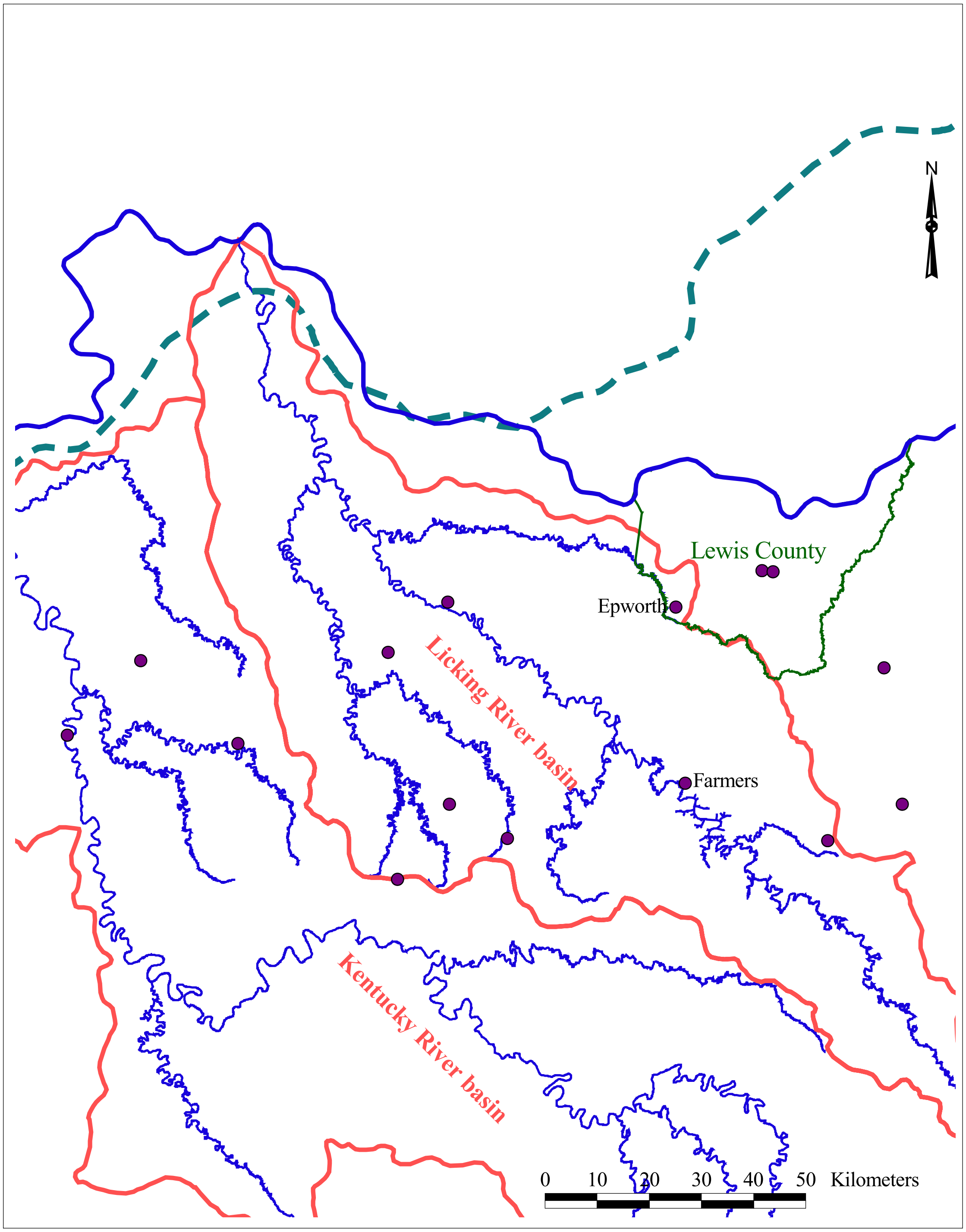


Figure 55. Distribution of glacial erratics (purple dots) beyond the mapped Pre-Wisconsin glacial limit (blue dashed line). Data from Leverett (1929). Modern drainage divides shown by red lines.

highest elevation of the Ohio River outwash terrace is approximately 500 ft in the area around Carrollton.

Valley-Bottom Fluvial Deposits

The valley fill of the Kentucky River valley consists of silty clay, clayey silt, and fine sand. The valley-fill materials contain scattered lenses of gravel, including pebbles and cobbles of chert, quartz, and coal, geodes, and slabs of limestone, siltstone, and sandstone. Some terraces exhibit cut-and-fill structures. Prior to construction of the lock-and-dam system, deposition of coarse gravel bars at the mouths of many tributary streams had caused shallow riffles along the Kentucky River, which were hazardous to navigation. Limited concentrations of degraded woody debris and organic matter can be found locally in the floodplain sediment. The valley fill ranges to a maximum of 83 feet (25 m) in thickness near Carrollton, but is generally significantly thinner upstream (Figure 56).

Two different materials of inferred glacial origin are associated with the valley-bottom fluvial deposits. A prominent series of Wisconsin outwash terraces along the Ohio River valley is underlain by a thick succession of gravel and sand (Figure 57a). The tops of these terraces have elevations as high as 152 m (500 ft) near Carrollton (Figure 56). Ice-deposited sediment of Wisconsin age has not been identified in Kentucky. A ridge of till inferred to be Illinoisan age (elevation 182 m (595 ft)) separates the Kentucky River valley from the Ohio River valley northeast of Carrollton. Glacial drift of inferred Illinoisan age has been mapped nearby at elevations as high as 215 m (700 ft) (Figure 56). This Illinoisan till/drift is less leached than older, pre-Illinoisan drift and till, and overlaps the older drift just north of Carrollton. Illinoisan outwash may be present along the sides and margins of the Ohio River valley, but is lithologically indistinguishable from the younger Wisconsin deposits. Thick layers of stiff, laminated to massive, blue-gray silty clay (“blue clay”) comprise much of the valley-fill material in the Kentucky River valley below Lock and Dam No 2, as documented by core logs obtained from the USACE and the USGS (Figure 44). Published geologic maps and reports (Appendix 1; Kane, 1972; Ryder, 1975; Johnson and Parrish, 1999) have mapped similar deposits in this part of the Kentucky River valley and in smaller tributaries of the Ohio River.

Although the valley-bottom deposits covering the bedrock valley are potentially subject to recent land-use-driven modification, some elements of the valley fill are probably remnants or

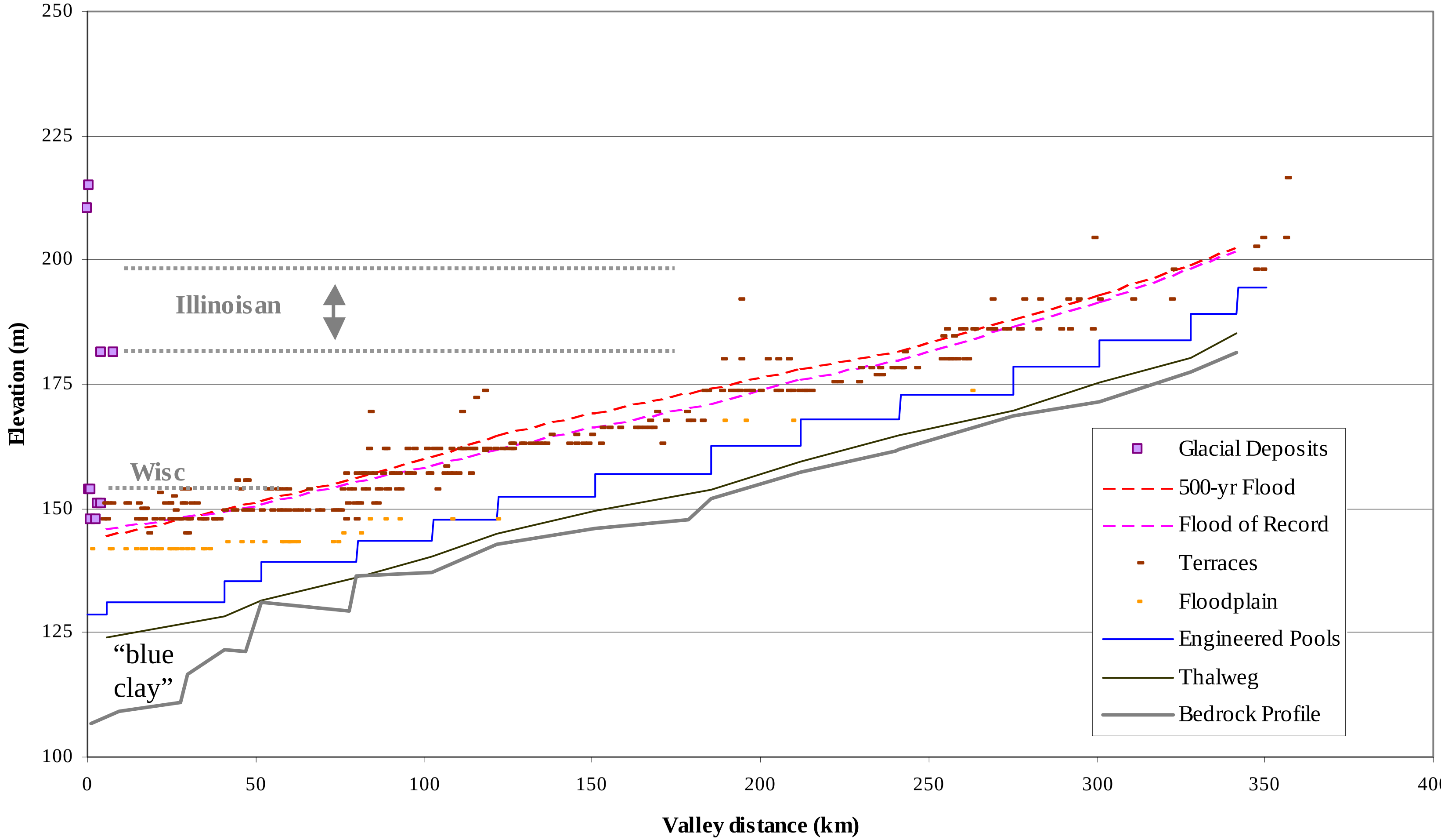


Figure 56. Longitudinal profile of the Kentucky River valley showing elevations of valley-bottom deposits and elevations of glacial deposits near Carrollton. The stair-step pattern of floodplain and terrace elevations is an artifact of the 10-ft or 20-ft (3-m or 6-m) contour intervals of topographic maps used to extract the elevations.

A

B

Figure 57. Photograph of valley-bottom terraces. A: Margin of gravel pit in Wisconsin outwash terrace, in Mexico Bottom, Indiana. Dashed line represents contact between sandy floodplain deposits and underlying gravel-dominated outwash deposits. B: View of terraces along Pond Creek between Gratz and Monterey. The arrow indicates the top of the terrace accordant with the Wisconsin outwash deposits downstream.

relicts of pre-settlement landforms. These can be identified as those standing above the geomorphically active flood zone, or as those buried below the active thalweg of the stream (Figure 58). As far upstream as Lock and Dam No. 3, elevations of terraces (~490 to 500 ft; 149 to 152 m) are accordant with Wisconsin glacial outwash terraces in the Ohio River valley (Figure 57b). Other terraces farther upstream are below the elevation of the highest mapped Illinoisan drift, and may represent similar terraces related to Illinoisan impoundment. Detailed stratigraphic data from these terraces was not available for this study. Other surficial terrace and "floodplain" levels are within the active flood zone, and are most likely modified to some degree by the land-use and engineering effects of the last two centuries (Figure 58). The thick layers of blue clay mapped in the Kentucky River valley below Lock and Dam No. 3, and in other smaller tributaries of the Ohio River, probably represent glacial impoundment during Wisconsin glacial outwash events along the Ohio River.

High-Level Fluvial Deposits

Descriptions of the high-level terrace and paleovalley deposits can be found in Campbell (1898), Foerste (1906), Jillson (1943b, 1944a, 1944b, 1945a, 1945b, 1946a, 1946b, 1946c, 1946d, 1947, 1948a, 1948b, 1950), and on the geologic-quadrangle maps referenced in Appendix 1. These descriptions have been supplemented by field work in the current study (Figure 17). The high-level fluvial deposits contain clayey silt, sandy silty clay, and sand, and are generally deeply weathered. The deposits are characterized in the field by the presence of diagnostic well-rounded pebbles of quartz and chert, as well as broken or eroded quartz geodes as much as 1 foot (30 cm) in diameter (Figure 59a). The geodes are found only downstream of Irvine. Slabs of sandstone as much as 2 ft (60 cm) in diameter are found in many of the deposits in the study area (Figure 59b). Gravel is found throughout the deposits, but typically is most abundant at the base of each deposit. The deposits are locally limonite cemented, and contain sparse to common limonite pellets. The deposits range from thin veneers on surfaces and benches to more than 50 feet (16 m) thick in less-eroded parts of paleovalleys.

<u>Irvine Formation</u>

Campbell (1898) described and formally named the Irvine Formation. The Irvine Formation includes unconsolidated sand, gravel, and clay on upland terraces and broad hilltops in the vicinity of Irvine (Figures 60, 61a). He inferred a fluvial origin for the deposits and

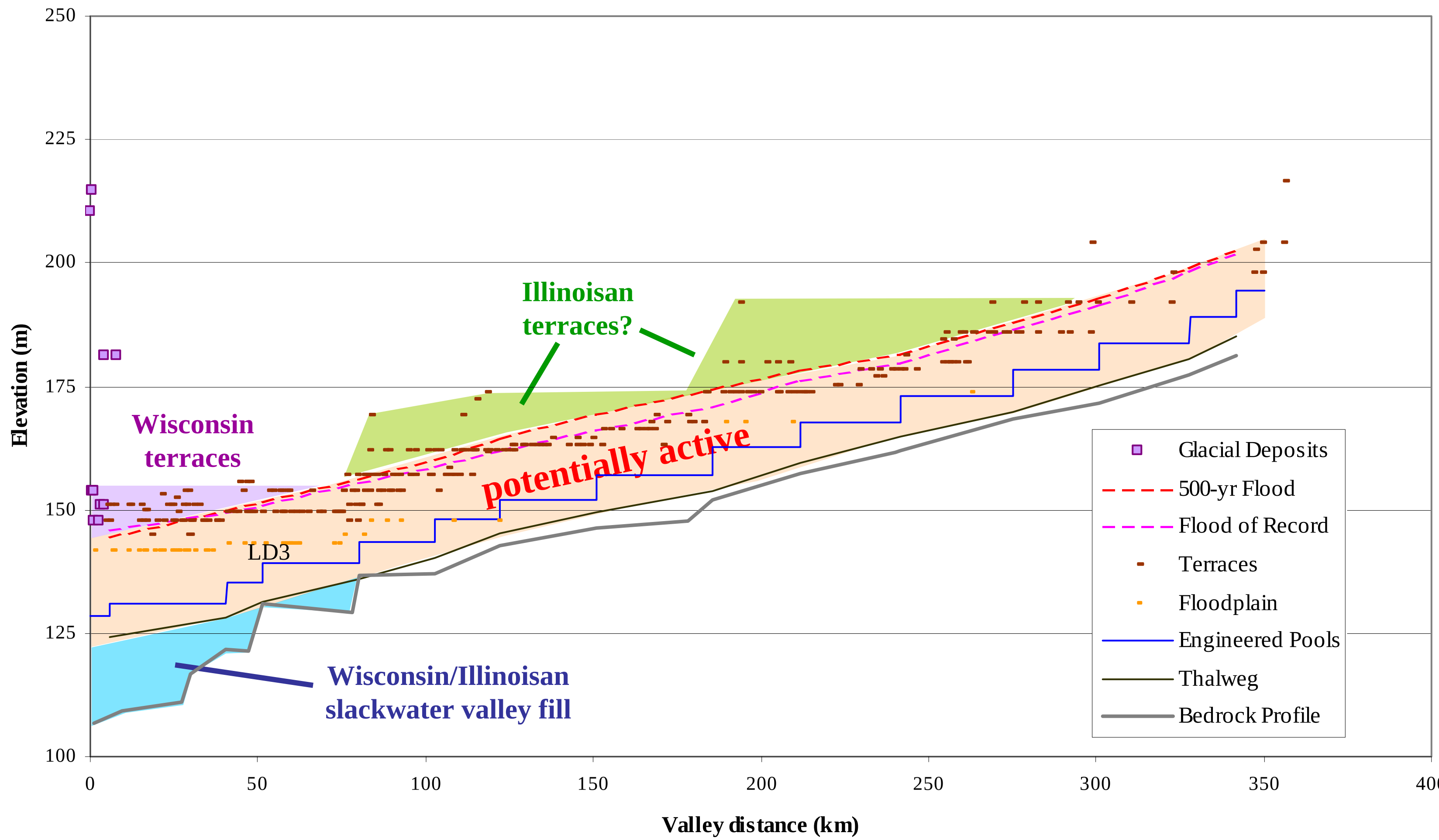


Figure 58. Interpreted longitudinal profile of valley-bottom deposits in the Kentucky River valley.

A

B

Figure 59. Photographs of diagnostic high-level fluvial sediments. A: Typical high-level fluvial gravel including chert and quartz pebbles and geodes in a sandy matrix near Camp Nelson. B: Sandstone slab found in fluvial deposit with chert and quartz pebbles near US 127 on the Franklin-Anderson County line.

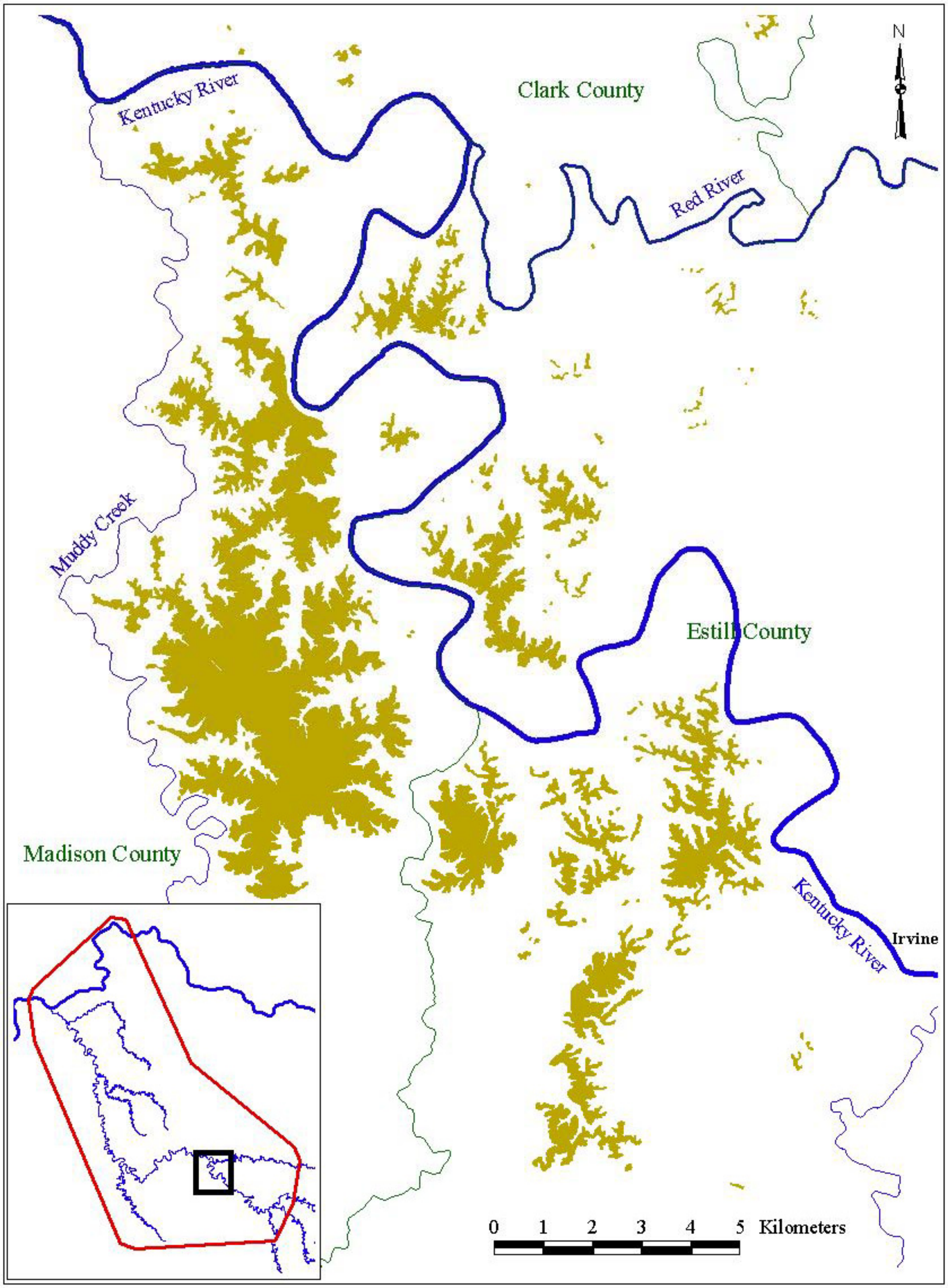


Figure 60. Distribution of the Irvine Formation of Campbell (1898), shown in tan.

A

B

Figure 61. Photographs of high-level fluvial deposits. A: Irvine Formation sand exposed in a borrow pit at Rice Station in Estill County. B: High-level fluvial sand, silt, and clay exposed in a construction site in Stringtown, near Lawrenceburg, Kentucky.

tentatively assigned a Late Tertiary age for the Irvine deposits on the basis of geomorphic position. Foerste (1906) expanded the description of the Irvine deposits, identifying additional exposures of the sand, gravel, and clay in the Clark, Madison, Estill County area. Subsequent workers have correlated other high-level fluvial deposits downstream (Figure 61b) with the Irvine (e.g., McFarlan, 1943, Granger and Smith, 2000), but no formal extensions of the Irvine Formation have been made beyond the type area. The KGS-USGS geologic mapping project restricted the formal Irvine Formation to the area shown in Figure 60.

Paleovalleys and High-Level Abandoned Meanders

The fluvial deposits are associated with high-level arcuate abandoned meanders and paleovalleys, or are in isolated bodies on benches and terraces. The high-level abandoned meanders and paleovalleys are within 9 km of the modern stream valley, and typically within 6 km (Figure 62). Many of the high-level abandoned meanders are larger in diameter than the present valley-bottom counterparts, and are mostly distributed along the stretch of the Kentucky River from the mouth of the Dix River almost to Carrollton. Other high-level abandoned meanders are in four isolated locations between Camp Nelson and the mouth of the Red River (Figure 62).

Valley widths, as mentioned previously, can be a difficult and arbitrary parameter to quantify in alluviated valleys. The width of high-level abandoned meanders was estimated by identifying paleovalleys with relatively intact valley walls which contained fluvial deposits spanning the valley width. The width of the fluvial deposits was measured perpendicular to the valley axis. The values obtained in this way are only estimates, because the fluvial deposits and valley walls potentially have been subjected to erosion and modification. The Old Kentucky River paleochannels northeast of Carrollton range from 1000 to 1400 m wide. A paleochannel between Tyrone and Salvisa is 660 to 750 m wide. The Warwick Abandoned Channel of Jillson (1947) is 870 to 1030 m wide. The Hickman Abandoned Channel of Jillson (1948a) is 900 to 100 m wide. The paleochannels mapped in the Winchester quadrangle (Appendix 1) range from 570 to 890 m wide.

Linear Paleochannels

Near Lawrenceburg in Anderson County and Levee in Montgomery County, linear paleochannel systems are oriented nearly perpendicular to the main valley (Figure 62). These

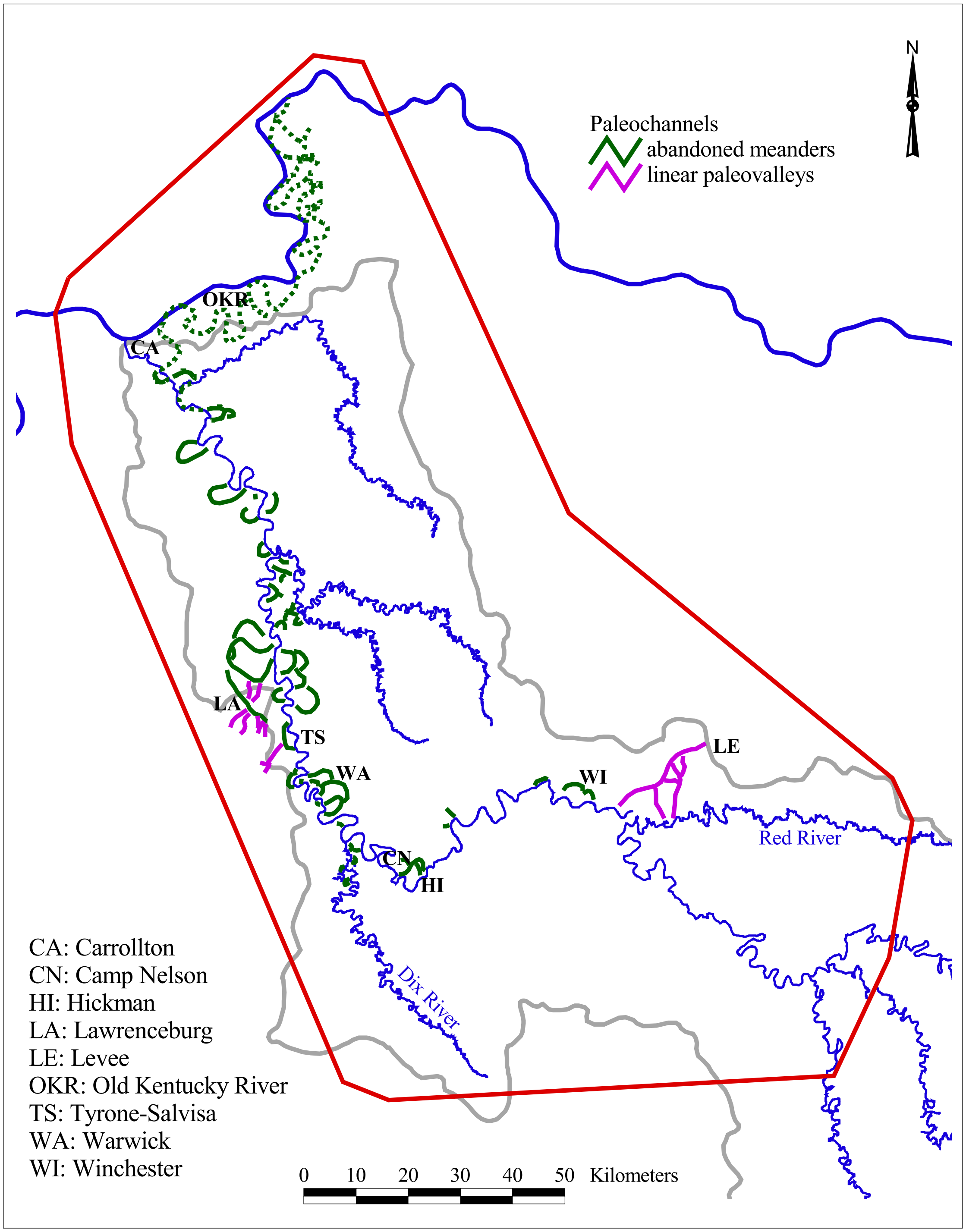


Figure 62. Distribution of high-level abandoned meanders and linear paleochannels in the Kentucky River valley.

paleovalleys extend from the main valley of the Kentucky River or from high-level abandoned meanders to the divide of the Kentucky River drainage basin. This study has confirmed the observation of previous workers (e.g., Leverett, 1929) that the channels have local veneers and deposits of quartz pebbles and sand. Each paleochannel has a distinct topographic expression as a stream valley, but contains no modern stream readily apparent for development of the valley. The Lawrenceburg and Levee paleochannels correspond to the lowest point along the divide between the respective drainage basins.

The Lawrenceburg paleochannels extend across the modern divide between the upper Salt River valley and the Benson Creek watershed in the main Kentucky River valley (Figure 63). The cols along the divide in each of the Lawrenceburg paleochannels have elevations ranging from 242 m (795 ft) to 248 m (815 ft). Leverett (1929) and workers from the KGS-USGS mapping program noted fluvial sediment along the divide between the Kentucky River and the upper Salt River, which was confirmed by field work during this study (Figure 17).

The Levee paleochannel trends to the southwest from the Licking River divide (elevation 270 m (890 ft)) (Figure 64, 65). The lower end of the paleochannel is a hanging valley above the Kentucky River near Upper Howards Creek at an elevation of 240 m (790 ft). Another paleochannel trends south-southeast from the main paleochannel toward the Red River near Clay City, and ends as a hanging valley at an elevation of 228 m (750 ft). Both segments of the Levee paleochannel are incised below the elevation of nearby Irvine Formation deposits, which have minimum elevations of 255 m (835 ft). With the exception of the lower paleovalley near the Red River, few fluvial deposits were mapped in this paleochannel system by the KGS-USGS geologic mapping program. Geologic mapping on the Levee and Hedges quadrangles (Appendix 1) noted the presence of sand and quartz pebbles in the paleochannel, but did not delineate the deposits. Field work conducted during the course of this study confirms the presence of rounded quartz and chert pebbles in the paleochannel from the Licking River divide to the community of West Bend (black dots on Figure 64).

The Lawrenceburg and Levee paleochannel systems are interpreted here to be spillways between adjacent basins. This interpretation is made on the basis of the geomorphic expression as paleovalleys, the coincidence with low cols along drainage divides, and the existence of fluvial sediment near the drainage divides. The Levee and Lawrenceburg spillways would have

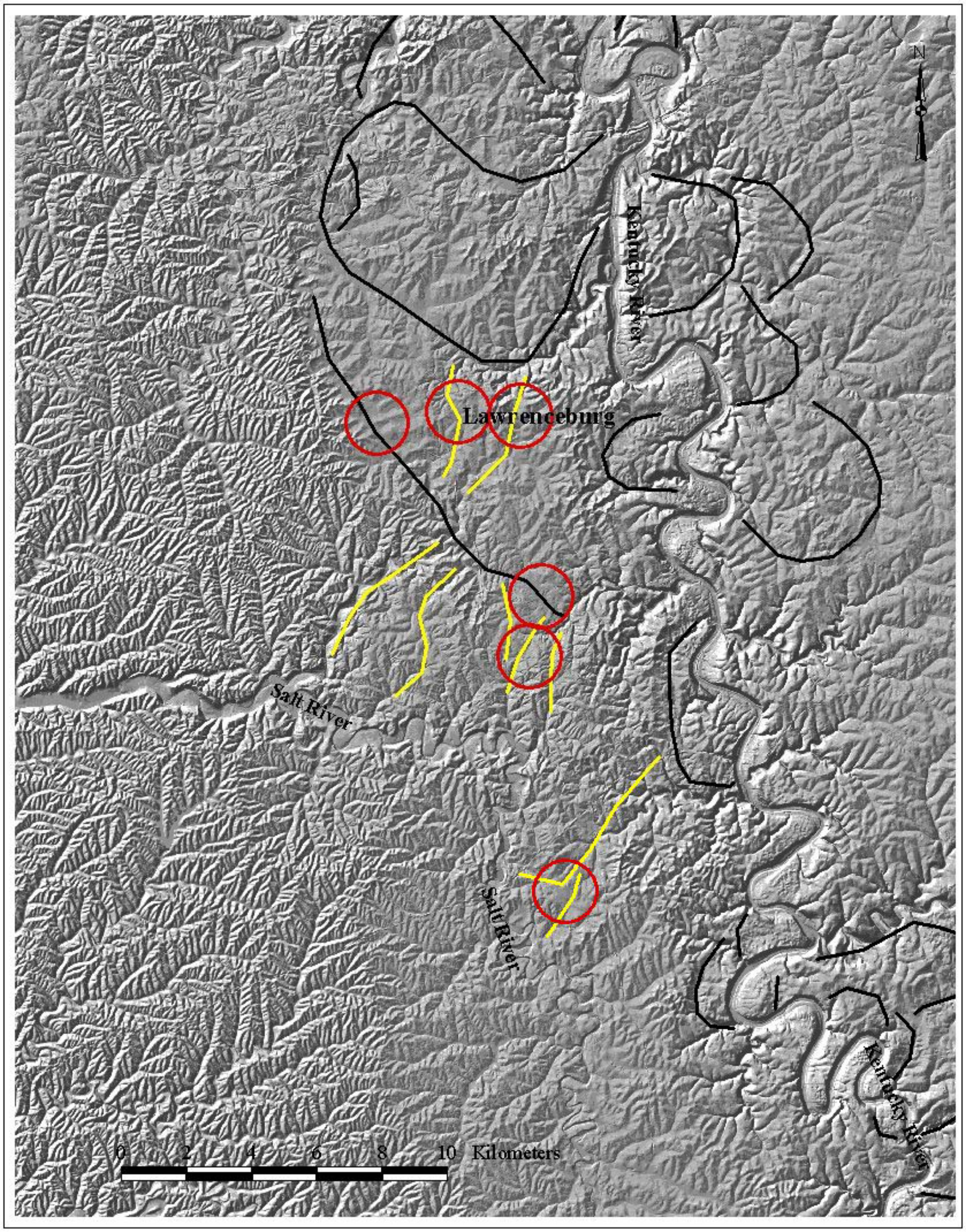


Figure 63. Map showing abandoned meanders (black lines) and linear paleochannels (yellow lines) near the Kentucky River and Salt River drainage divide. Red circles indicate locations where paleochannels cross the drainage divide.

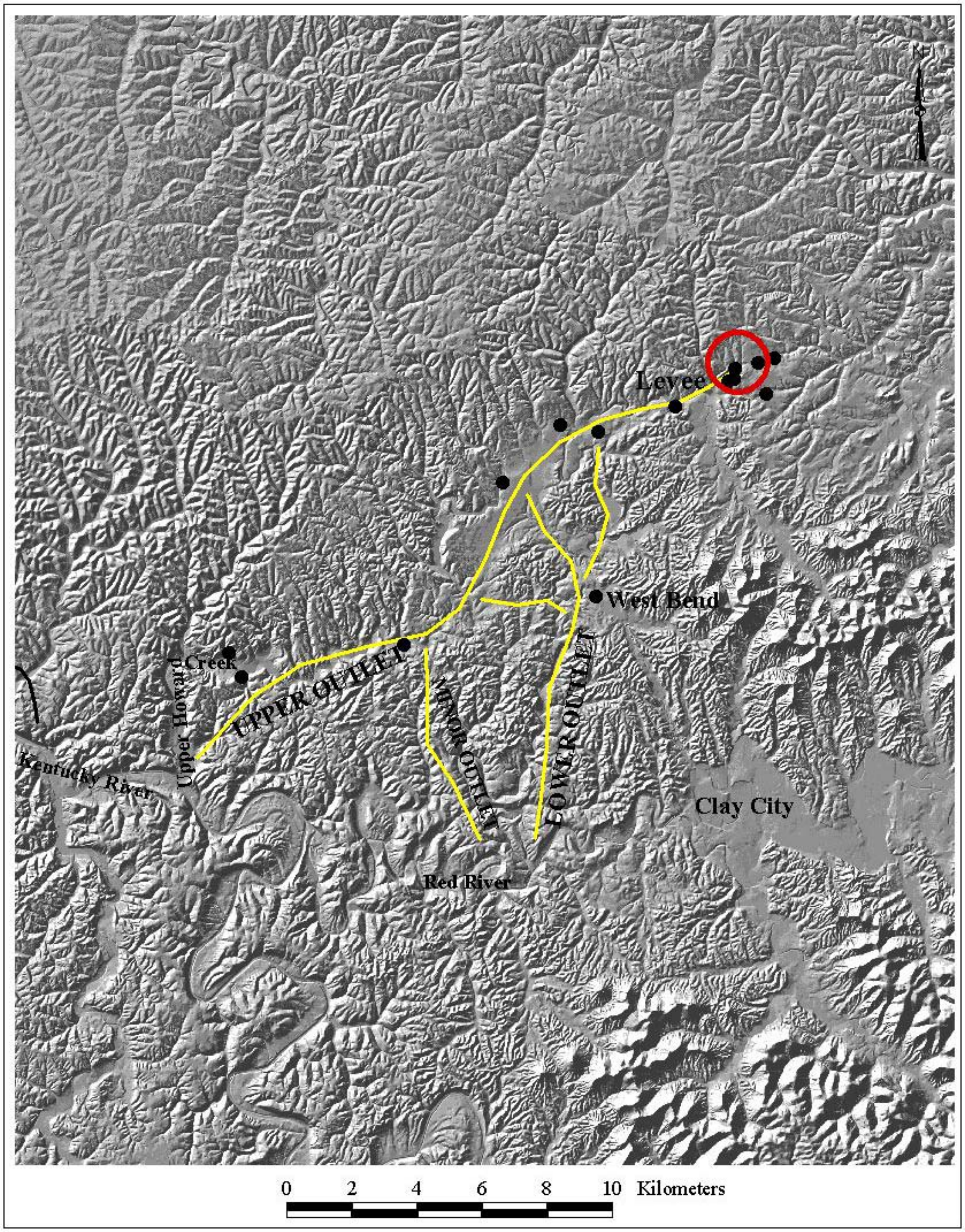


Figure 64. Map showing linear paleochannels (yellow lines) near Licking River divide. The red circle indicates the location of the col in the divide. Black dots indicate confirmed locations of quartz gravel.

A

B

Figure 65. Photographs of Levee paleochannel. A: View along paleochannel from near col, looking southwest over community of Levee. Flow was away from the viewer. B: View from divergence of upper and lower outlets, looking northeast. Flow was toward the viewer, and diverged right (lower outlet) and left (upper outlet) past the viewer. Arrow in each photo indicates the same upland, viewed from opposite directions.

been active only during times of glacial impoundment of the Licking River basin and Kentucky River basin, respectively. The Levee spillway is here inferred to have released impounded water from the Licking River valley into the Kentucky River valley. The Lawrenceburg spillways are inferred to have released impounded water from the Kentucky River valley into the lower Salt River valley.

<u>Bedrock Benches</u>

Numerous bedrock benches or terraces/straths are identifiable from topographic maps of the river valley. Many of these are similar in topographic expression to terraces with mapped fluvial deposits (Figure 66). In several places, however, the fluvial deposits are present as a thin veneer or were overlooked by geologic mappers. Whereas the number of benches identified from topographic maps in the Kentucky River valley is quite large, comprehensive field checking was impractical for this study. Only a small sample of bedrock benches was examined in the field; the presence of quartz and chert gravel was confirmed at each of these sites (Figure 17).

<u>Profile of Fluvial Deposits and Features</u>

The distribution of the Irvine Formation, other high-level fluvial deposits, paleochannels, the outlets of inferred spillways, and bedrock benches is shown in the longitudinal profile on Figure 67. In general, the fluvial deposits and paleochannels are found only at elevations of 200 m (650 ft) or higher in the Kentucky River basin. The “floor” of the main trend of fluvial deposits and paleochannels is graded upstream with a slope very similar to that of the modern stream. The deposits generally grade from the Old Kentucky River paleochannels near Carrollton (valley-km 10, elevation 200 m) (Swadley, 1971) up to the upper Levee spillway outlet mapped in Clark County (elevation 240 m). Local exceptions are in three isolated areas of the valley (valley-kilometers 50 to 140, 180 to 200, and 250 to 300) where bedrock benches and fluvial deposits are found below this grade. Above this grade, no distinct or discrete profiles are evident in the data (Figures 67 and 68).

Both ends of the “floor” of fluvial features have associations with a glacial event, discussed earlier in this chapter. Swadley (1971) mapped lacustrine sediment and southbound outwash deposits over the northbound Old Kentucky River deposits near Carrollton. Swadley (1971) inferred the sedimentary succession to represent a reversal of flow in the Old Kentucky River caused by disruption of Teays River drainage downstream by a pre-Illinoisan glacial

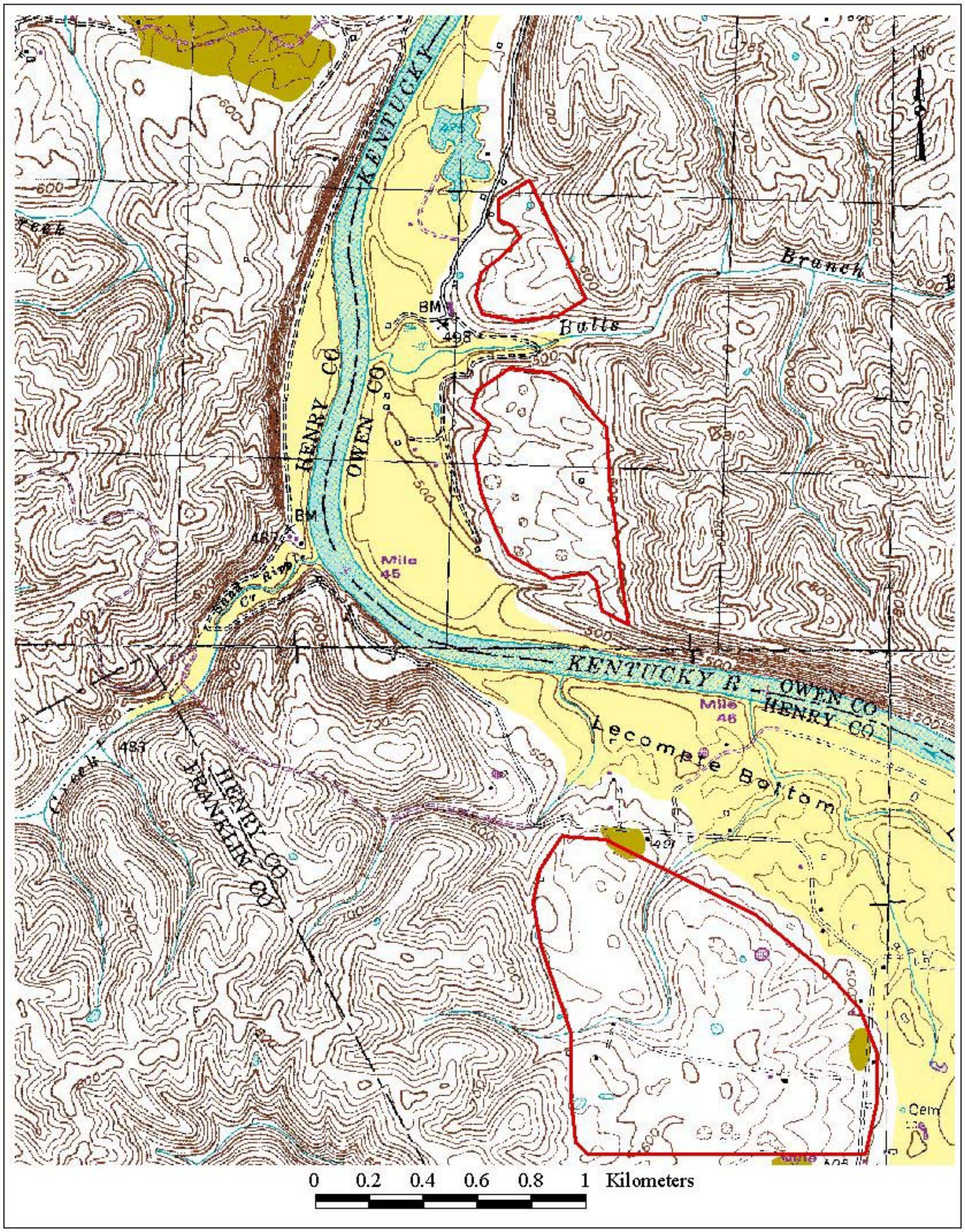


Figure 66. Detail of Gratz and Polsgrove 7.5-minute topographic quadrangles, showing prominent bedrock benches (outlined in red). Valley-bottom alluvium is shown in light yellow and high-level fluvial deposits shown in dark tan.

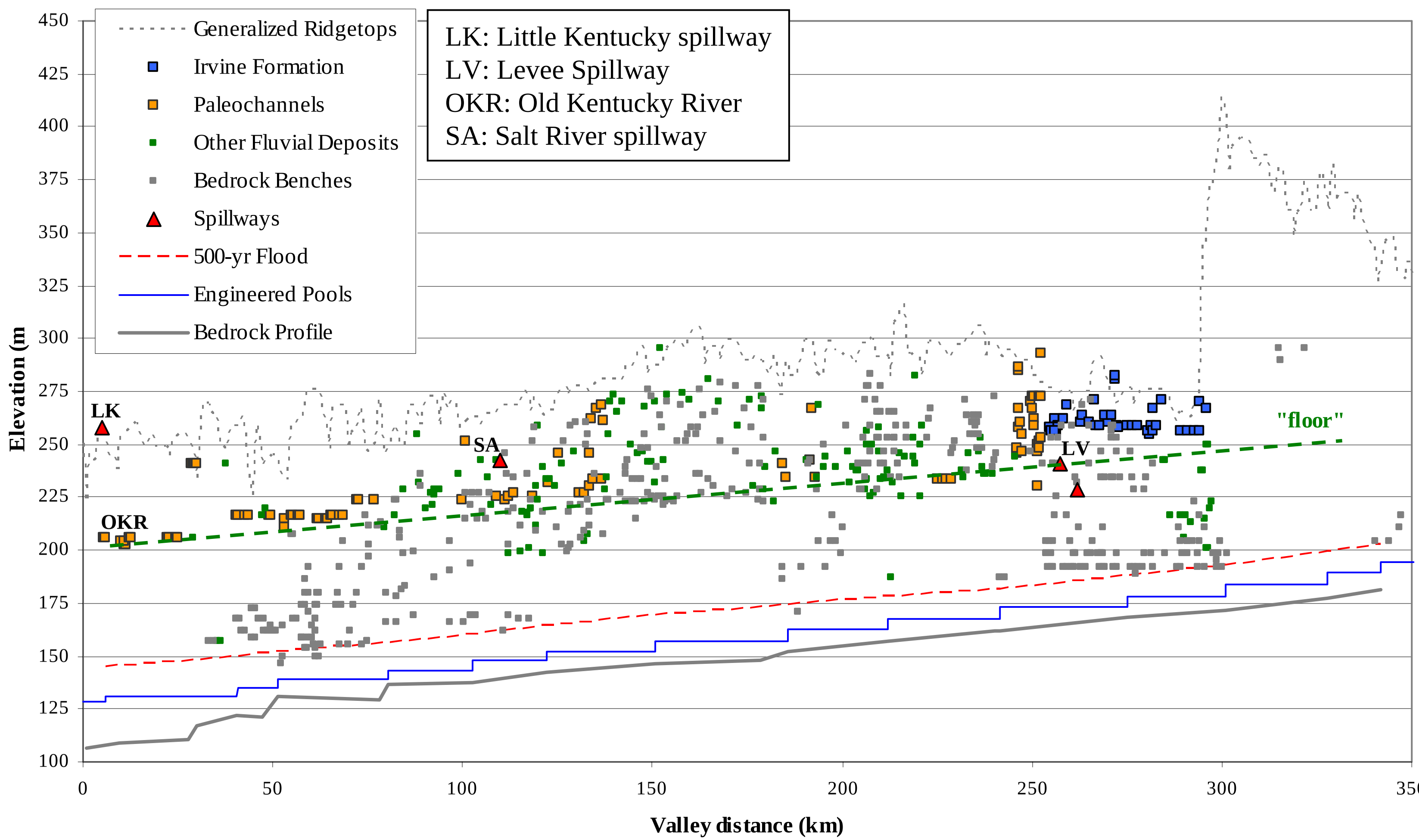


Figure 67. Longitudinal profile of the Kentucky River valley showing elevations of paleochannels, high-level fluvial deposits, including the Irvine Formation as mapped by Campbell (1898), bedrock benches, and spillways draining into and out of the Kentucky River basin.

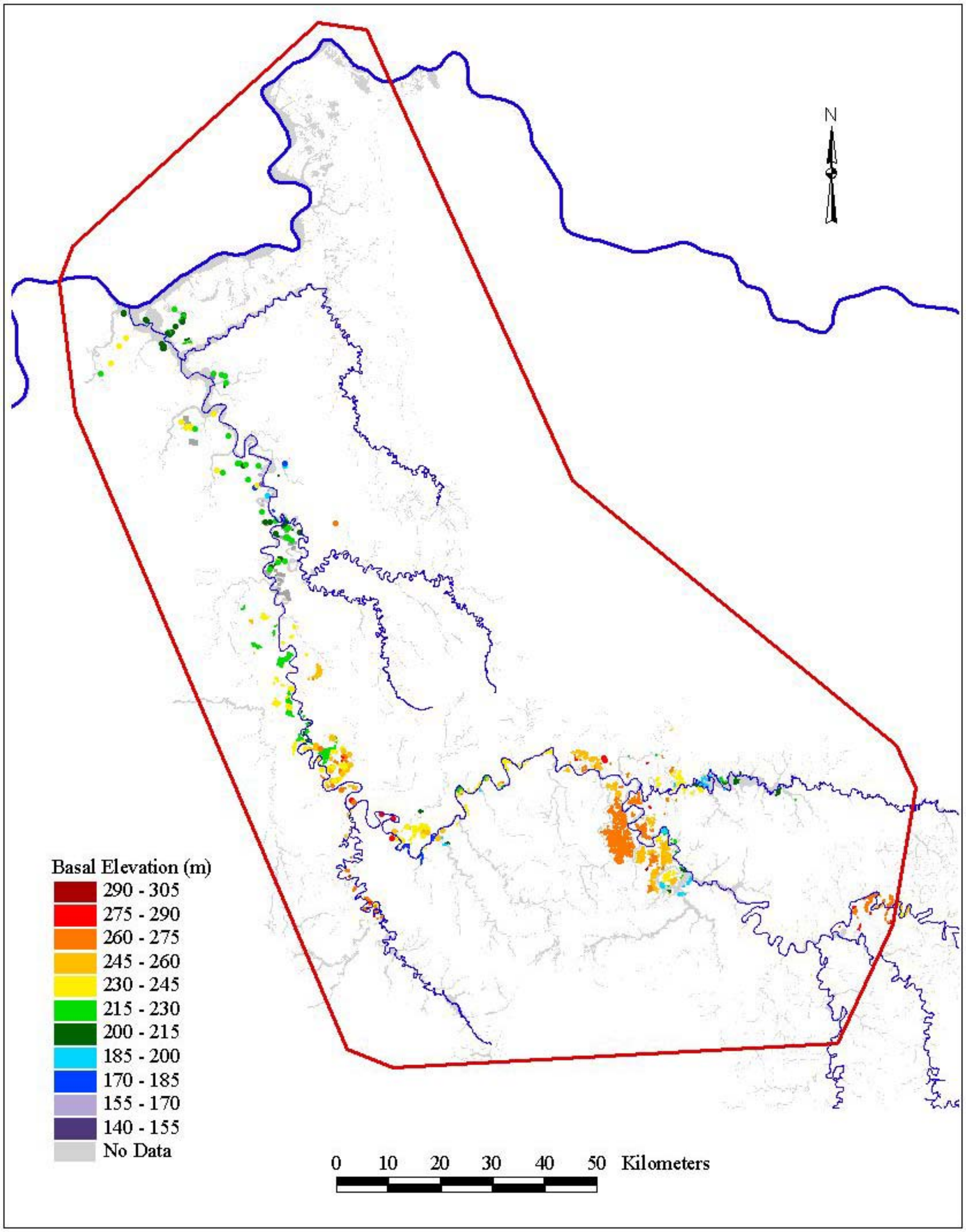


Figure 68. Basal elevation of high-level fluvial deposits in the Kentucky River valley.

advance. The Levee spillway is inferred by this study to represent an outlet of water overflowing from the Licking River basin during times of glacial impoundment. The coincidence of these two glacial interpretations at opposite ends of the “floor” suggests that the two sites represent the effects of a single glaciation on different ends of the Kentucky River valley. The “floor” therefore would represent the final position of the Old Kentucky River prior to disruption of the system by destruction of the Teays River system and integration of the Ohio River system by a pre-Illinoisan glacial advance.

Fluvial Deposits and Bedrock Stratigraphy

The valley-morphology styles identified and discussed earlier in this chapter correspond to the bedrock stratigraphy exposed in the valley walls. The Wilmore style corresponds with the Camp Nelson Limestone. The Tyrone style corresponds with all but the top few meters of the Tyrone Limestone. The Gratz style corresponds to the upper Tyrone Limestone and the Lexington Limestone. The Worthville style corresponds generally to the Kope and Clays Ferry Formations. The Palmer style corresponds to Upper Ordovician units above the Kope/Clays Ferry Formations, overlying Silurian units, and the Devonian Boyle Dolomite. The Irvine style mainly corresponds to the Nancy Member of the Borden Formation, which is dominated by clay shale. The Heidelburg style corresponds to the resistant beds of the Cowbell Member of the Borden Formation, the Slade Formation, and the Pennsylvanian Corbin Sandstone.

The bedrock lithologies constrain the colluvial processes, as well as the fluvial erosion rates and thresholds, and thus control the development of landforms and valley morphology. If bedrock lithology is a dominant control on valley morphology, the distribution of high-level fluvial deposits and features should be analogous to the corresponding valley-bottom morphology styles suggested by the enclosing bedrock stratigraphy. To test this hypothesis, this study compares selected high-level fluvial deposits and paleovalleys to the associated present-day valley-bottom morphology style, on the basis of bedrock stratigraphy. The valley morphology styles discussed above correspond to major stratigraphic units (Figures 36, 38). Figure 69 shows the projection of the potential valley morphology styles in the Kentucky River valley, on the basis of bedrock stratigraphy shown in Figure 6.

The paleovalley of the Old Kentucky River northeast of Carrollton (Swadley, 1971) is eroded into Upper Ordovician stratigraphic units of limestone and limestone and shale (Figure

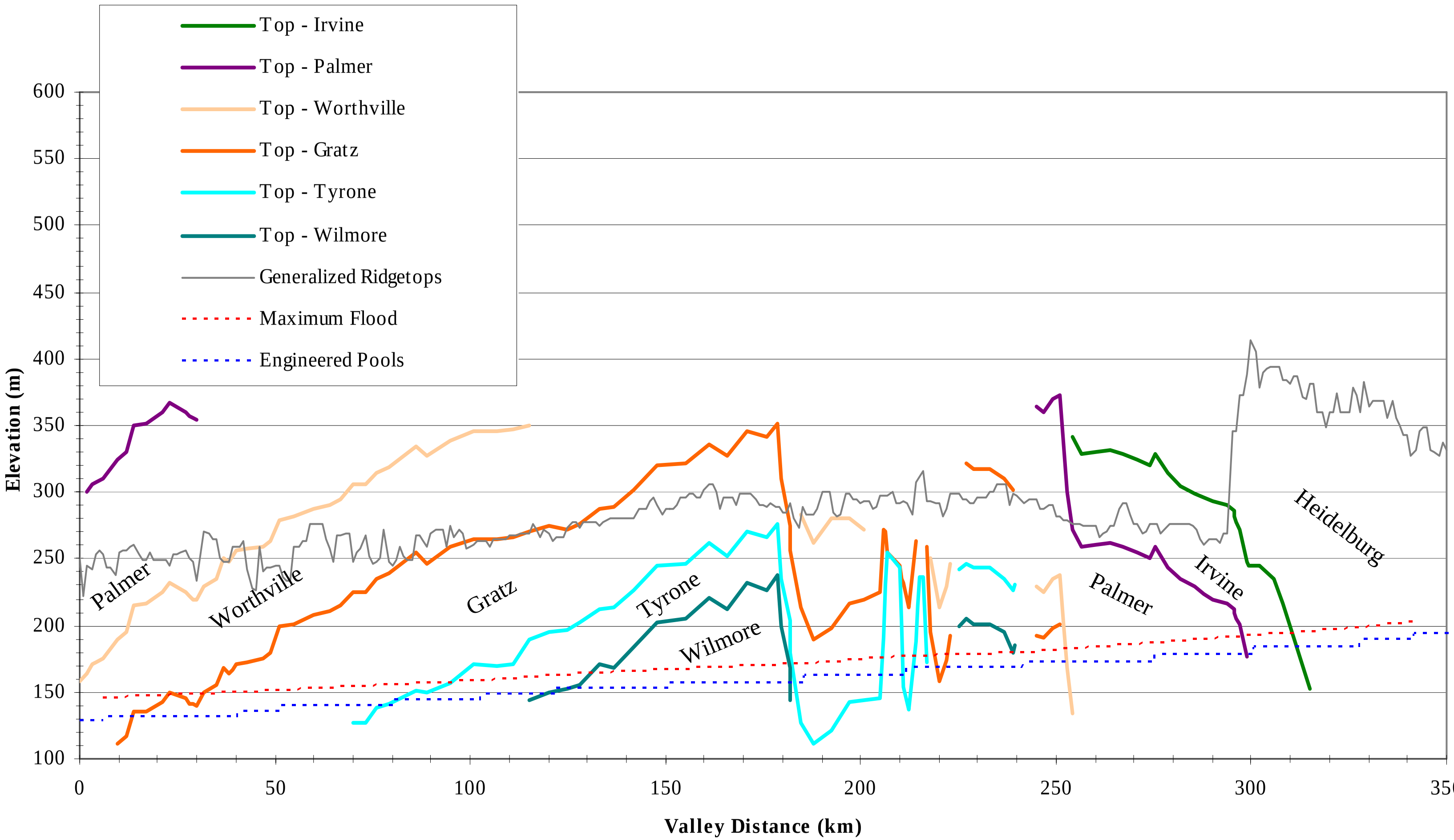


Figure 69. Projection of valley morphology styles using bedrock stratigraphy.

70). Multiple paleovalleys eroded into Upper Ordovician rocks were mapped in the Winchester quadrangle (Appendix 1). The valley-bottom morphology style associated with these lithologies is the Palmer style. The Old Kentucky River paleovalleys are 650 to 800 m wide, and have a rounded meander valley pattern. The Old Kentucky River paleovalley walls are asymmetric in meander bends. The elevation of high-level fluvial deposits on the inside of meander bends is consistent with slip-off slopes. The Winchester paleovalleys are arcuate and range from 650 to 850 m wide. The higher levels of fluvial deposits in the Winchester paleovalleys are located on the inside of meander bends, which is consistent with slip-off slopes. The characteristics of both sets of paleovalley examples are consistent with the modern valley morphology of the Palmer style (Figure 70).

A series of paleovalleys between Tyrone and High Bridge is eroded into the Lexington Limestone. The bedrock stratigraphy corresponds to the Gratz morphology style. The paleovalleys range from 650 to 1000 m wide, which is consistent with the Gratz style. One paleovalley is subparallel to the main valley, whereas the others in this area are arcuate. In the Gratz style, the valley-bottom abandoned meanders are arcuate and comparable in size to the High Bridge-area paleovalleys, and the Pond-Cedar Creek valley has a cutoff to the main valley that is similar to the subparallel paleovalley near Lawrenceburg (Figure 71).

The Irvine Formation includes broad deposits of sand, gravel, and clay as much as 2300 m wide preserved on uplands; no distinct paleovalleys are preserved. The deposits primarily rest on New Albany Shale; the Nancy Member of the Borden Formation would thus have comprised the valley walls for these deposits. The corresponding present-day valley-bottom morphology analog is the Irvine style. The valley-bottom fluvial deposits in the Irvine style are similarly broad in the present Kentucky River valley, with valley widths as much as 1100 m. In a similar bedrock setting in the Red River, the alluviated valley ranges to 1700 m wide (Figure 72). The basal elevation of Irvine Formation deposits is lower to the east, as is the lower contact of the Nancy Member of the Borden Formation. Although no distinct paleovalley is available for comparison, the distribution of the Irvine Formation deposits in relation to the underlying bedrock stratigraphy is consistent with lithologic control on a paleovalley.

The stratigraphic projection of valley morphology styles into the upland can also help to explain the entrenched meander patterns of the Tyrone and Wilmore styles (Figure 73). The

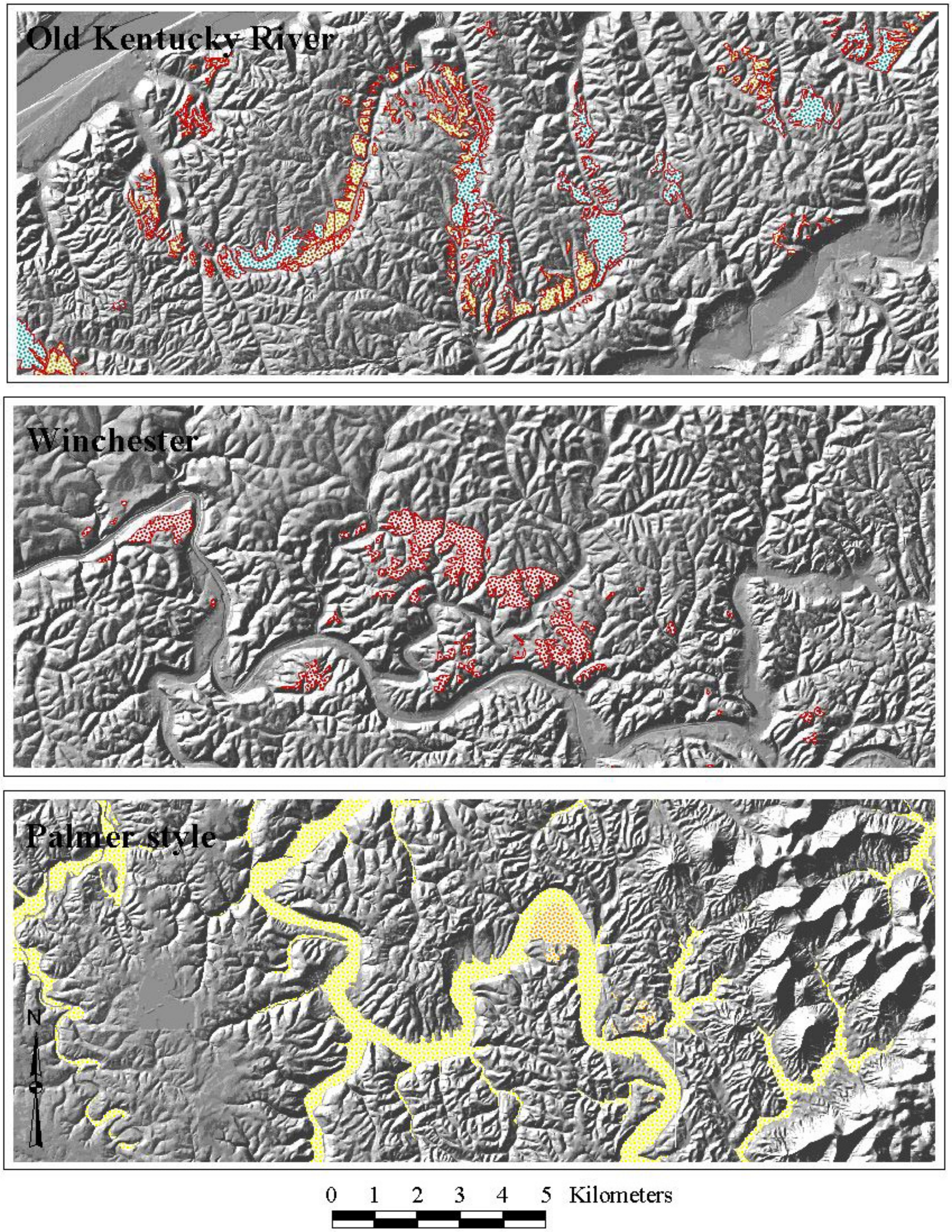


Figure 70. Comparison of paleochannels in the Old Kentucky River and near Winchester with the Palmer valley-bottom morphology style. Fluvial deposits are indicated by the dotted patterns.

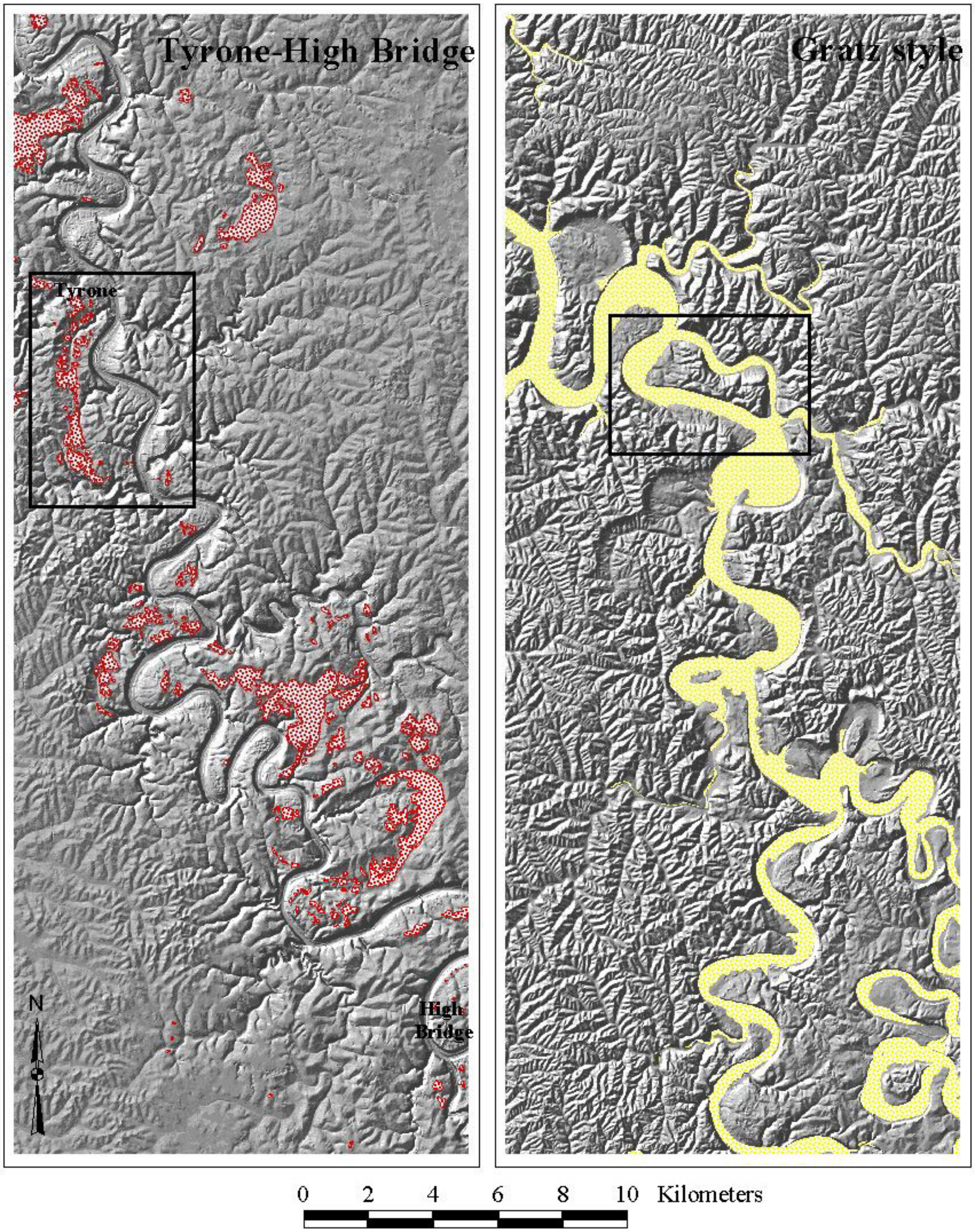


Figure 71. Comparison of paleochannels between Tyrone and High Bridge with the Gratz valley-bottom morphology style. Fluvial deposits are indicated by dotted patterns. The subparallel paleovalley (left) and the Cedar-Pond Creek capture are outlined with black boxes.

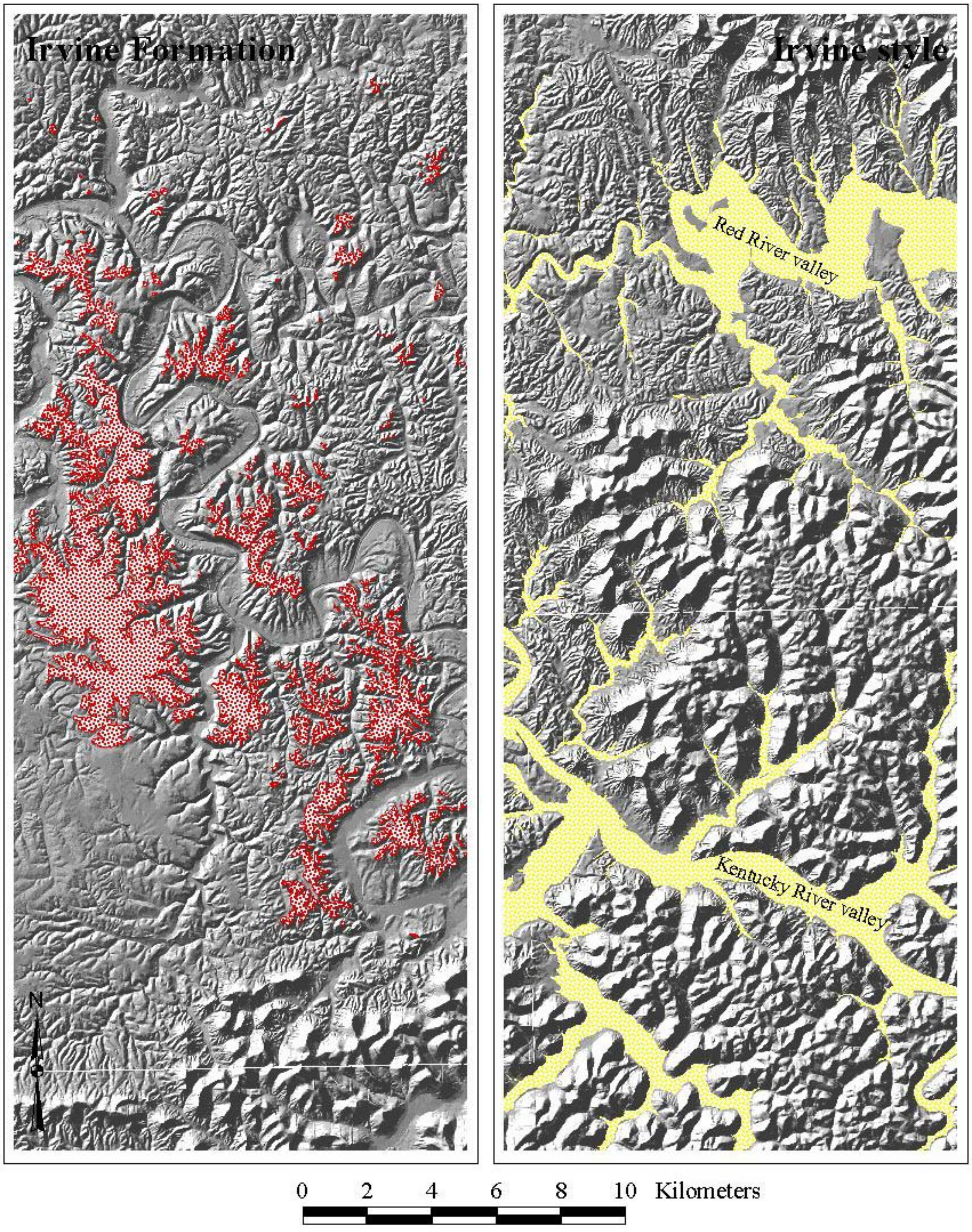


Figure 72. Comparison of the distribution of Irvine Formation deposits with the Irvine valley-bottom morphology style. Fluvial deposits are indicated by dotted patterns.

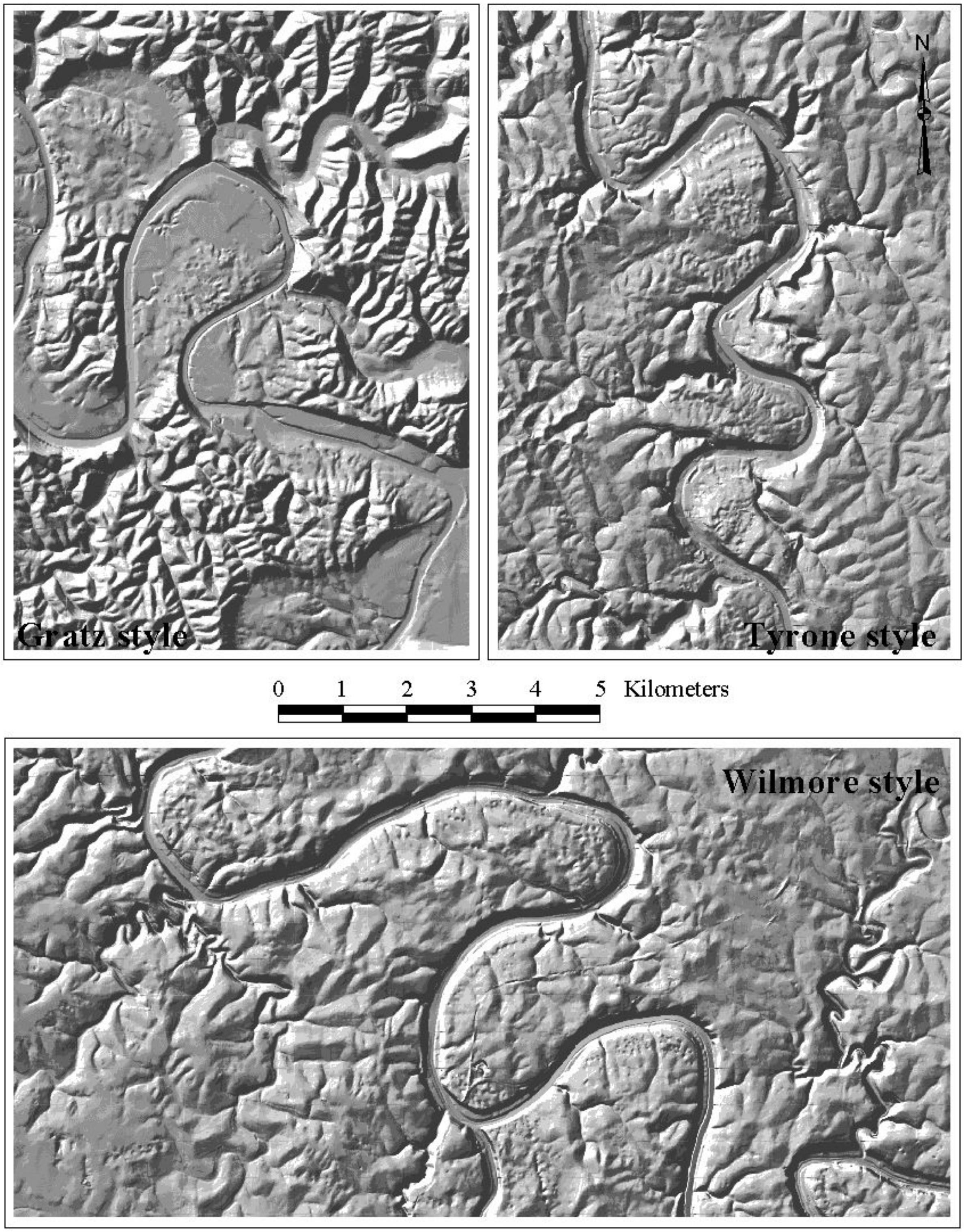


Figure 73. Comparison of meanders in the Gratz, Tyrone, and Wilmore morphology styles.

bedrock lithologies of both styles are relatively resistant to lateral erosion, and should resist the development of meandering stream courses. The Gratz pattern, however, is characterized by well-rounded, and well-developed meanders (Figure 73). Well-developed meanders in the Gratz valley morphology style that eroded into the underlying Tyrone style would have been "locked" into place by the lateral resistance of the Tyrone, and further constrained by the more-resistant underlying Wilmore style. The meanders preserved in the Tyrone and Wilmore styles are inherited from overlying, preceding landscapes which were controlled by their enclosing stratigraphy, and the modern valleys in those styles are dominated by vertical erosion, and limited lateral erosion.

The morphology and distribution of preserved upland fluvial deposits and features is similar to those in comparable stratigraphic settings in the modern river valley. This reinforces the concept that the development of landforms and valley morphology is controlled by the lithology of the bedrock units. The consistency of valley-bottom morphology with the distribution of high-level deposits and paleovalleys suggests that lithology has been the dominant control on landform morphology through the time period represented by the observed deposits.

Supra-upland Deposits

Jillson (1963) identified potential fluvial deposits well away from the modern and upland deposits of the Kentucky River. At a series of sites trending generally northwest from Boonesboro toward Monterey (Figure 74), Jillson found quartz pebbles that he inferred as demonstrating an eastern Kentucky provenance for the deposits. At four sites, he also identified granite cobbles and pebbles (Figure 74). As shown in Jillson's (1963) frontispiece photograph, these granite cobbles appear to be relatively unweathered, water-modified, faceted pebbles; he reports them to be 2.5 to 3.5 inches (6 to 9 cm) in diameter. Jillson offered these pebbles as evidence of a through-flowing, very ancient ("Mesozoic"), paleo-Kentucky River with headwaters in the Blue Ridge Mountains of southern Virginia or northwestern North Carolina, along the modern trend of the modern Wautauga River and North Fork of the Kentucky River. The relatively fresh nature of Jillson's central Kentucky granite pebbles argues against a very-old depositional age, because survival of intact granite cobbles at the surface in Kentucky since the Mesozoic seems improbable but is possible. A search of the bibliographic database GeoRef did

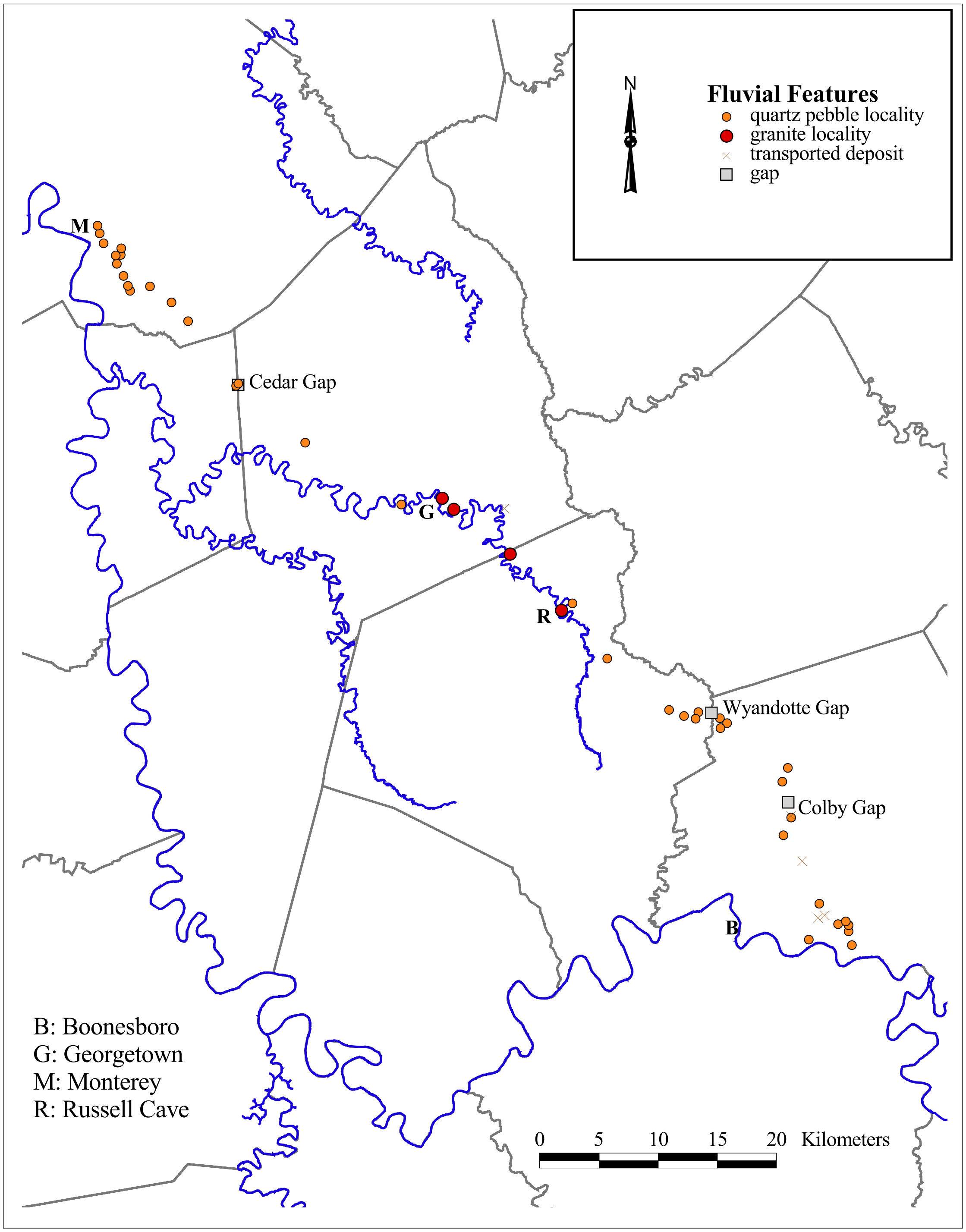


Figure 74. Distribution of fluvial features identified by Jillson (1963) across the central Kentucky upland.

not produce any sources documenting the weathering rate of granite in a temperate humid climate that could resolve this issue.

Adjacent Streams

Licking River

Master streams in the Licking River basin generally trend to the northwest (Figure 75). Luft (1980, 1986) summarized the most recent geologic mapping in the Licking River valley. He identified a lacustrine-over-fluvial sedimentary pattern similar to that described by Swadley (1971) for the Old Kentucky River. A fluvial deposit mantles a bedrock strath at roughly 190 m (620 feet) and is graded northward into Ohio. This strath joins the Old Kentucky River strath north of Cincinnati, and Luft correlated it with the Teays age Parker Strath. This strath and associated fluvial deposits are covered by Illinoisan drift in southern Ohio. In Kentucky, the fluvial deposit is overlain by a clay-dominated lacustrine unit named the Claryville Clay. Two other units are mapped above the Claryville: one is a gravel lag at approximately 250 m (820 ft), and the other is a lacustrine-type clay beneath sand and gravel at some sites at 240 to 270 m (785 to 885 ft) in Campbell County, Kentucky. Below the Parker Strath, Luft (1980) identified scattered deposits mantling valley walls and meander cores and referred them to Illinoisan outwash blocking the Licking River valley. A set of high terraces grades to the Wisconsin outwash in the Ohio River valley, and lower terraces are within the modern flood zone. Luft (1986) considered the composition of gravel in “Teays age” and older fluvial deposits along the South Fork of the Licking River, and observed that they included lithologies not present in bedrock of the South Fork valley, such as quartz and geodes. He hypothesized that the ancient Kentucky River originally flowed along the trend of the South Fork of the Licking River.

Salt River

The Salt River, west of the Kentucky River basin, has a series of north-northwest flowing tributaries, feeding into a westward-flowing master stream (Figure 75). The upper Salt River makes a sharp bend from the north-northwest to a more westerly trend near Lawrenceburg, Kentucky, and passes between relatively high north-trending ridges (Figure 76). Tributary streams flanking the high ridge are distinctly barbed, trending opposite the flow direction of the main stream. The upper Salt River drainage basin is the only component of the Salt River

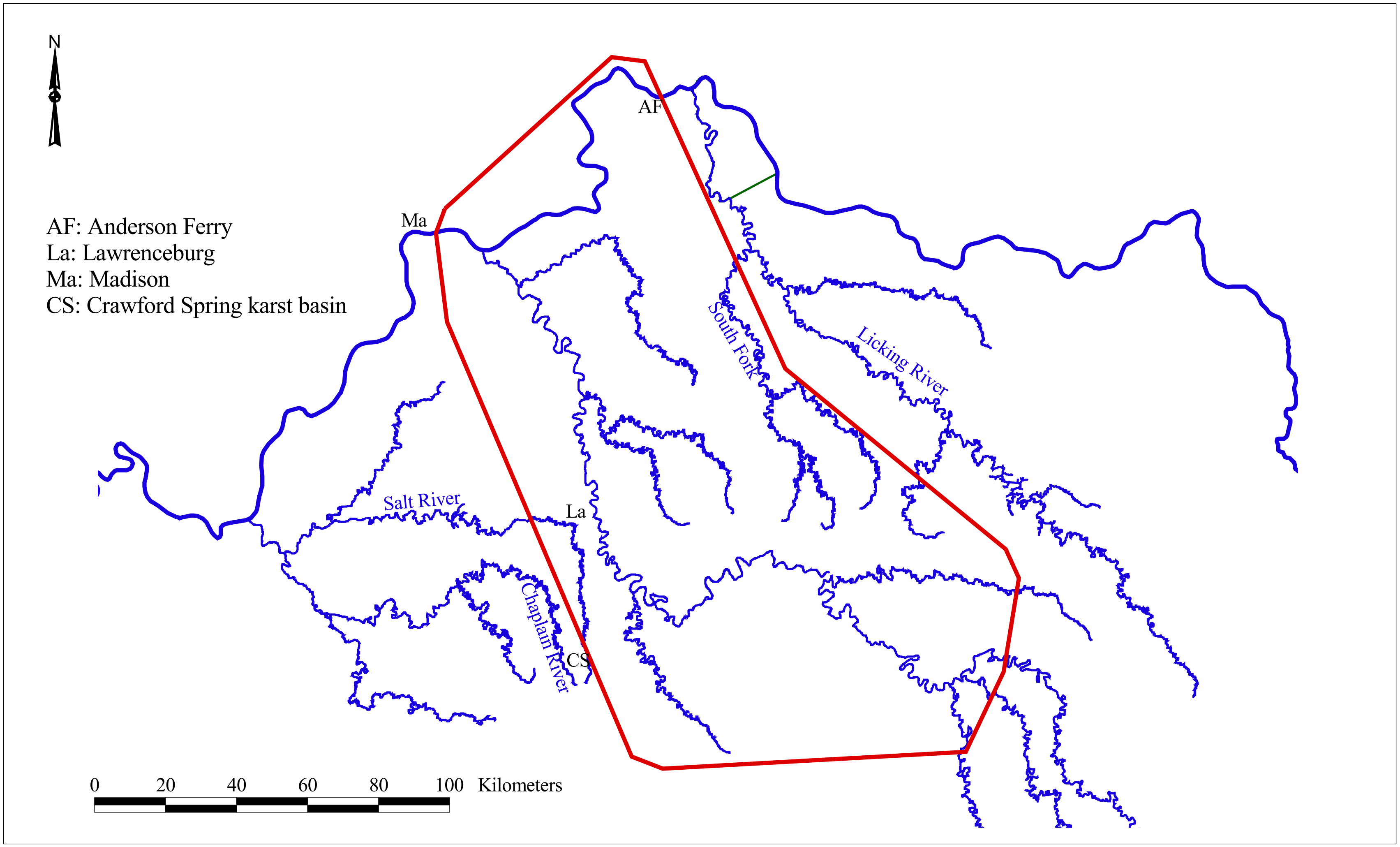


Figure 75. Map of stream valleys adjacent to the Kentucky River discussed in this study. Campbell County shown in green.

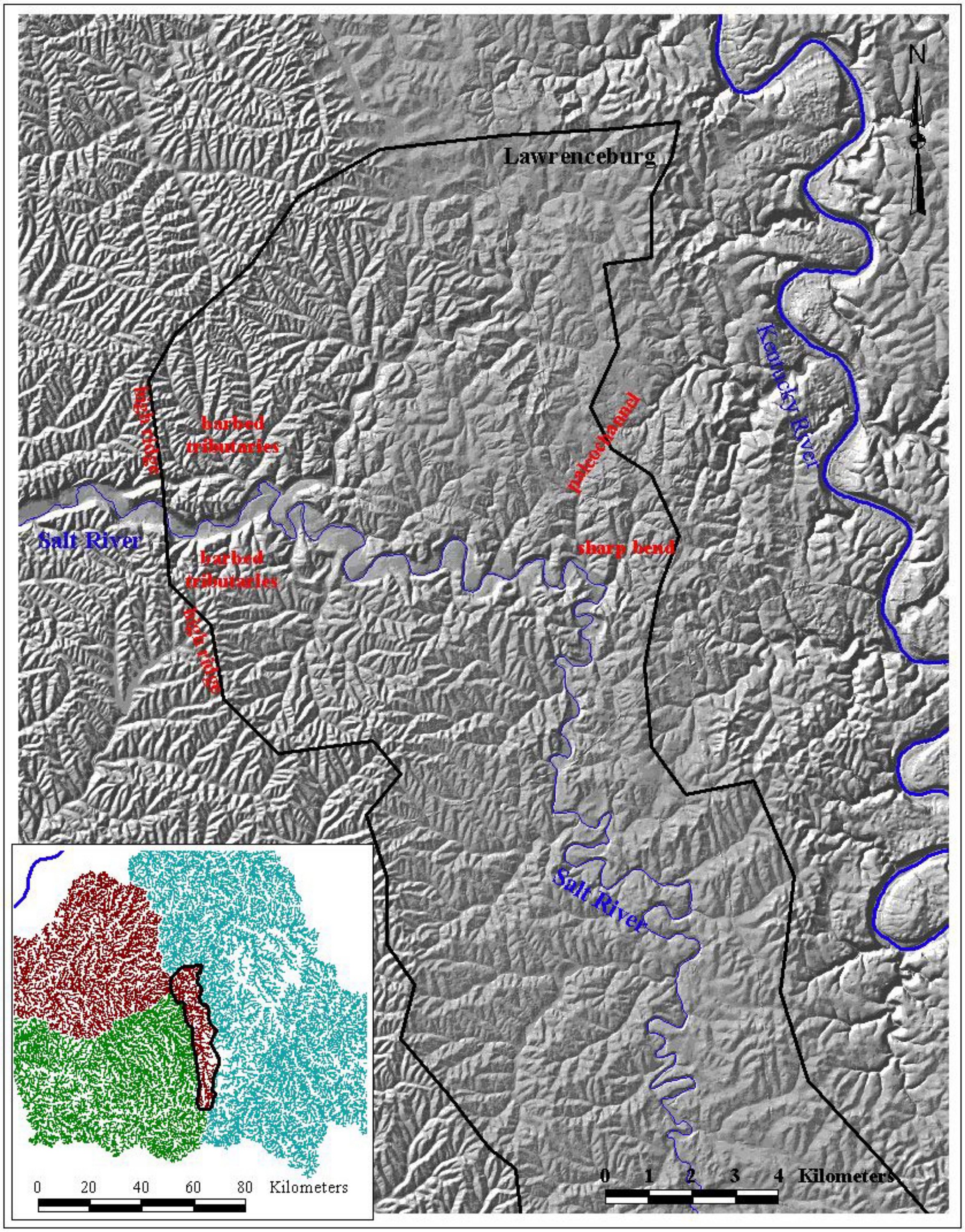


Figure 76 Detail view of topography near the point where the Salt River captured an ancient tributary of the Old Kentucky River. The inset compares the areas of the Salt River basin (red), the Rolling Fork basin (green), and the Kentucky River basin (blue). The modern drainage area of the upper Salt River is outlined in black on both maps.

drainage basin that is east of the relatively high north-northwest trending ridge. North of the sharp bend of the Salt River, a paleovalley can be traced into the Benson Creek watershed and high-level abandoned meanders of the Kentucky River. Fluvial sediments have been mapped along the divide between the two streams. Leverett (1929) and Jillson (1943) interpreted these data as evidence of a capture of an old tributary of the Kentucky River by headwaters of the ancient Salt River. The lower Salt River has a distinctly anastomosing pattern for approximately 5 km downstream of the high dividing ridge, in contrast to a meandering channel pattern elsewhere along the course of the upper and lower Salt River. The upper Salt River was most likely a tributary of the Kentucky River prior to capture by the lower Salt River. The modern upper Salt River has a drainage area of roughly 445 km^2.

Recent KGS dye-trace experiments have indicated that the upper Salt River is losing flow through a karst conduit system to Crawford Spring along the Chaplain River (Figure 77, 78) (R. Paylor, personal communication, 2003). The karst flow passes under the intervening surface drainage of Quirks Run, which also loses flow to the karst system (Currens and others, 2003). The conduit system decreases approximately 23 m (75 ft) in elevation over a straight-line distance of 7.8 km, which yields a maximum gradient of 3.0 m/km. This is identical to the dip of bedrock across this same area as shown on the geologic quadrangle maps (Appendix 1). The conduit flow in the Crawford Spring groundwater basin passes under the high dividing ridge separating the upper Salt River drainage basin and the Rolling Fork drainage basin to the west (Figure 77). The Salt River swallets and Crawford Spring are in the lowest tongue of the Tanglewood Member of the Lexington Limestone, between the Grier and Brannon Members. The site of the inferred upper Salt River capture is in a similar stratigraphic position as the Crawford Spring groundwater basin. A tongue of the Tanglewood Member of the Lexington Limestone is exposed below Clays Ferry Formation, and the bedrock dips to the west at 2.25 m/km. The distance from the northbound upper Salt River to the west side of the high dividing ridge is roughly 10.8 km, compared to the 7.8-km distance of the modern Crawford Spring conduit system.

It is possible that a karst conduit system comparable to the Crawford Spring groundwater basin may have caused or assisted the capture of the upper Salt River by the lower Salt River. Other possible mechanisms of capture include headward erosion or groundwater sapping of the early Salt River (Pederson, 2001) or overtopping of the divide by water during glacial

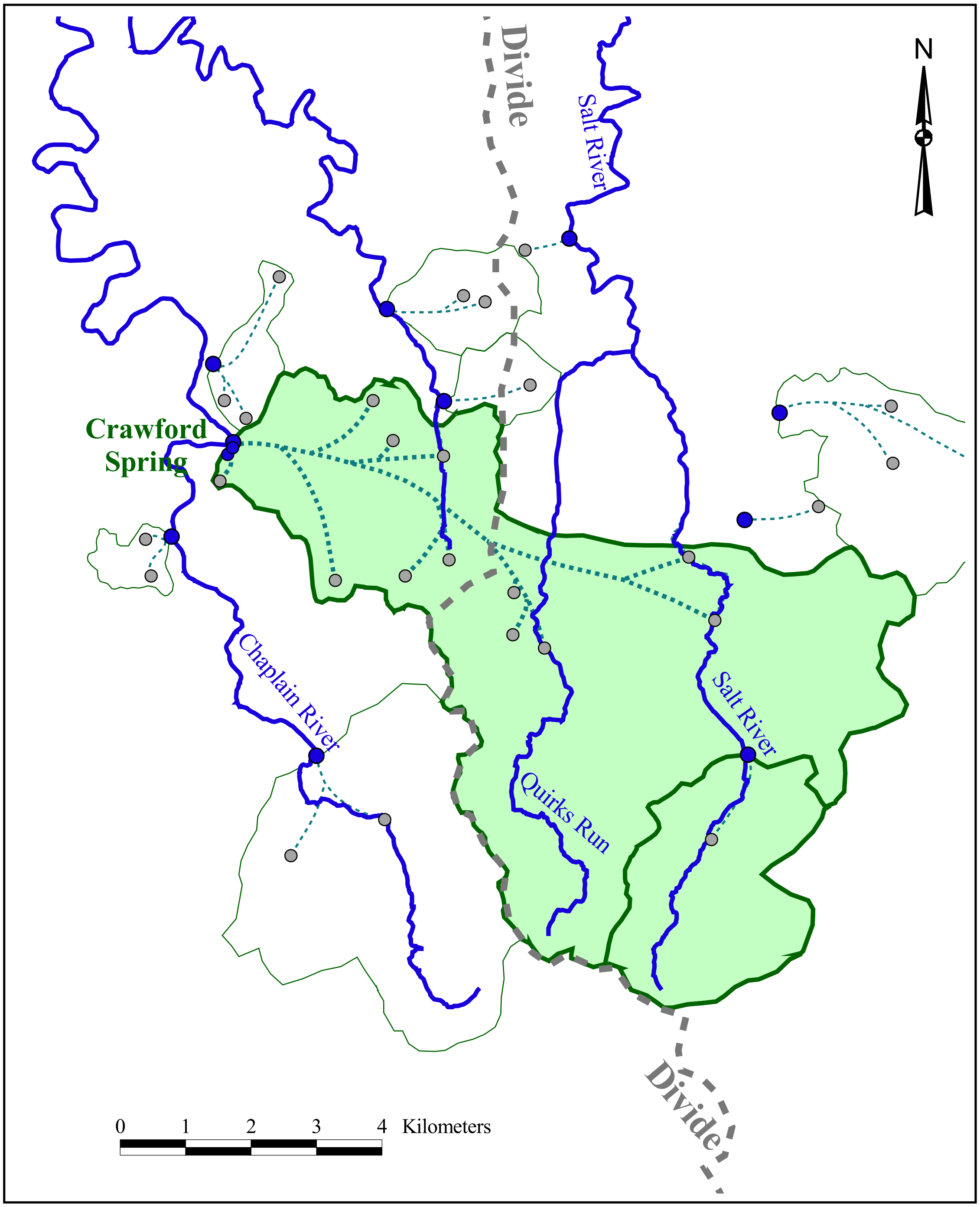


Figure 77. Map of Crawford Spring groundwater basin, showing results of Kentucky Geological Survey dye-trace experiments in the southern Bluegrass region. Crawford Spring basin is shown in green. Blue circles indicate springs; gray circles are dye-injection locations; dashed blue lines are inferred flow paths; dashed gray line is surface drainage divide between the Chaplain River and Salt River basins. Modified from Currens and others (2003).

Figure 78. Photograph of Crawford Spring

impoundment of the Kentucky River basin. The mechanism of this capture was not a focus of this study, and detailed field work to resolve the timing and mechanism of capture was not conducted.

Teays and Ohio Rivers

The bedrock topography of the ancient Teays River system was mapped by Ohio Division of Geological Survey (2003), Gray (1982), and Horberg (1950). The buried bedrock valley of the Teays River system can be traced from the glacial margin in southeast Ohio, through west-central Ohio, across central Indiana, and across Illinois (Figure 79). The Teays paleovalley and the modern Ohio River valley converge, or are in close proximity, near the modern confluence of the Ohio and Mississippi Rivers at Cairo, Illinois. The sedimentary deposits in the Teays River valley and its tributaries have been summarized by Rhodehamel and Carlston (1963), Bigham and others (1991), Bleuer (1991), Kempton and others (1991), Luft (1980), Swadley (1971), and Teller and Goldthwait (1991). Typically, fluvial deposits of gravel, sand, and silt are overlain by lacustrine silts and clays.

The distance from Carrollton, Kentucky, to Cairo, Illinois, by way of the buried Teays River valley is approximately 1250 km. The distance from Carrollton to Cairo along the modern Ohio River valley is only 600 km. A generalized profile of the Teays River valley was constructed using data from the bedrock topography maps (Figure 80). The average gradient of the Teays is approximately 0.110 m/km, compared to a gradient of 0.152 m/km for the modern Kentucky River thalweg. The “floor” of the high-level fluvial deposits in the Kentucky River valley that extends from the Old Kentucky River paleochannels near Carrollton to the upper outlet of the Levee spillway has an average gradient of 0.157 m/km, nearly identical to that of the modern Kentucky River thalweg.

A series of high-level gravel deposits was identified along the Ohio River during the KGS-USGS geologic mapping project. These sediments are characterized by sand deposits with abundant chert gravel covered with a smooth brown patina. The deposits have been mapped as the Ohio River Formation in southern Indiana (Wayne, 1960), as the Luce Gravel near Owensboro (Ray, 1965), as the Lafayette Gravel in the Jackson Purchase area (Potter, 1955a, 1955b), and as unnamed high-level fluvial deposits by Theis (1929) and the KGS-USGS geologic mapping program. The elevations of these deposits are shown in the longitudinal

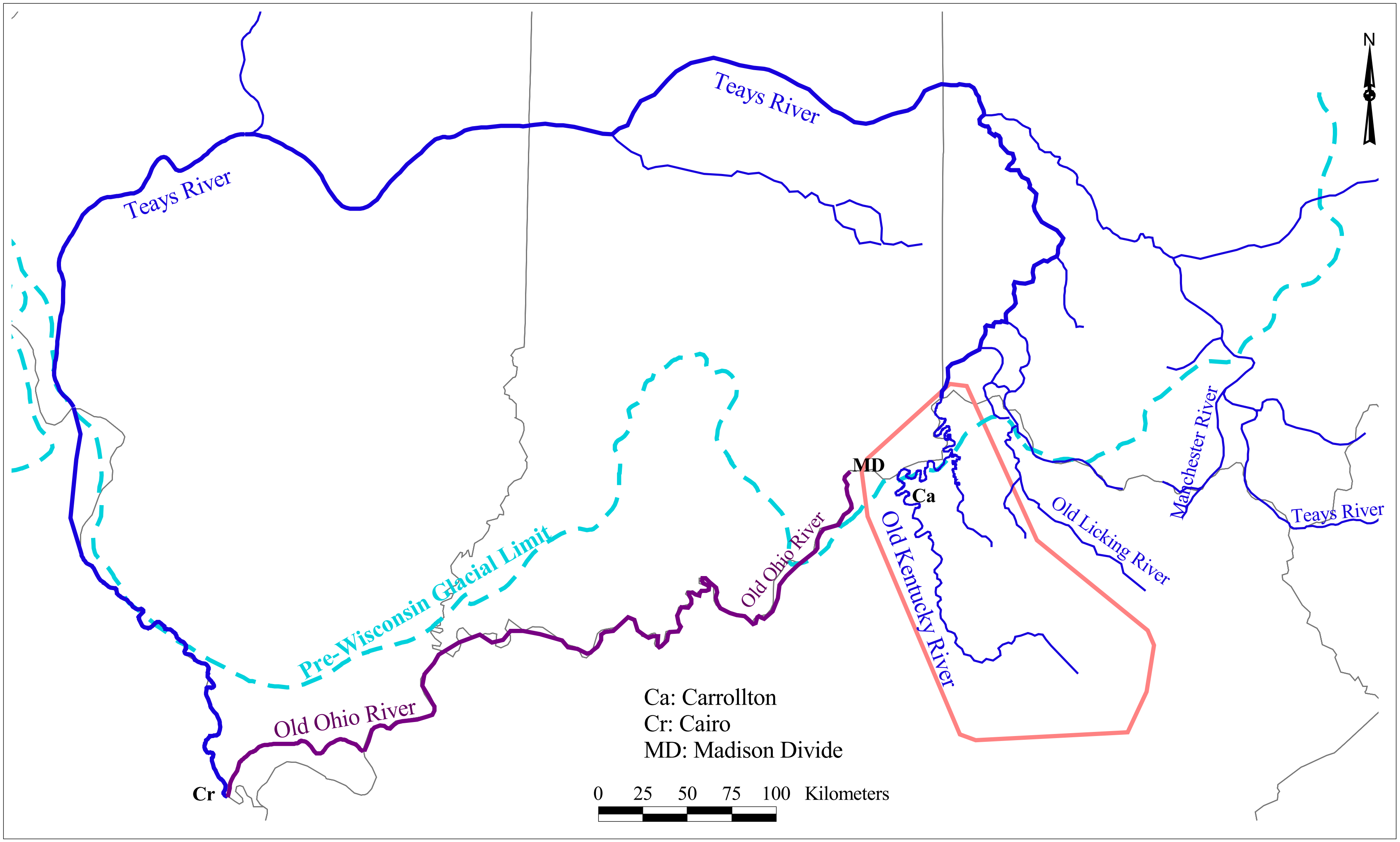


Figure 79. Teays River system paleodrainage and Pre-Wisconsin glacial limits. References in text.

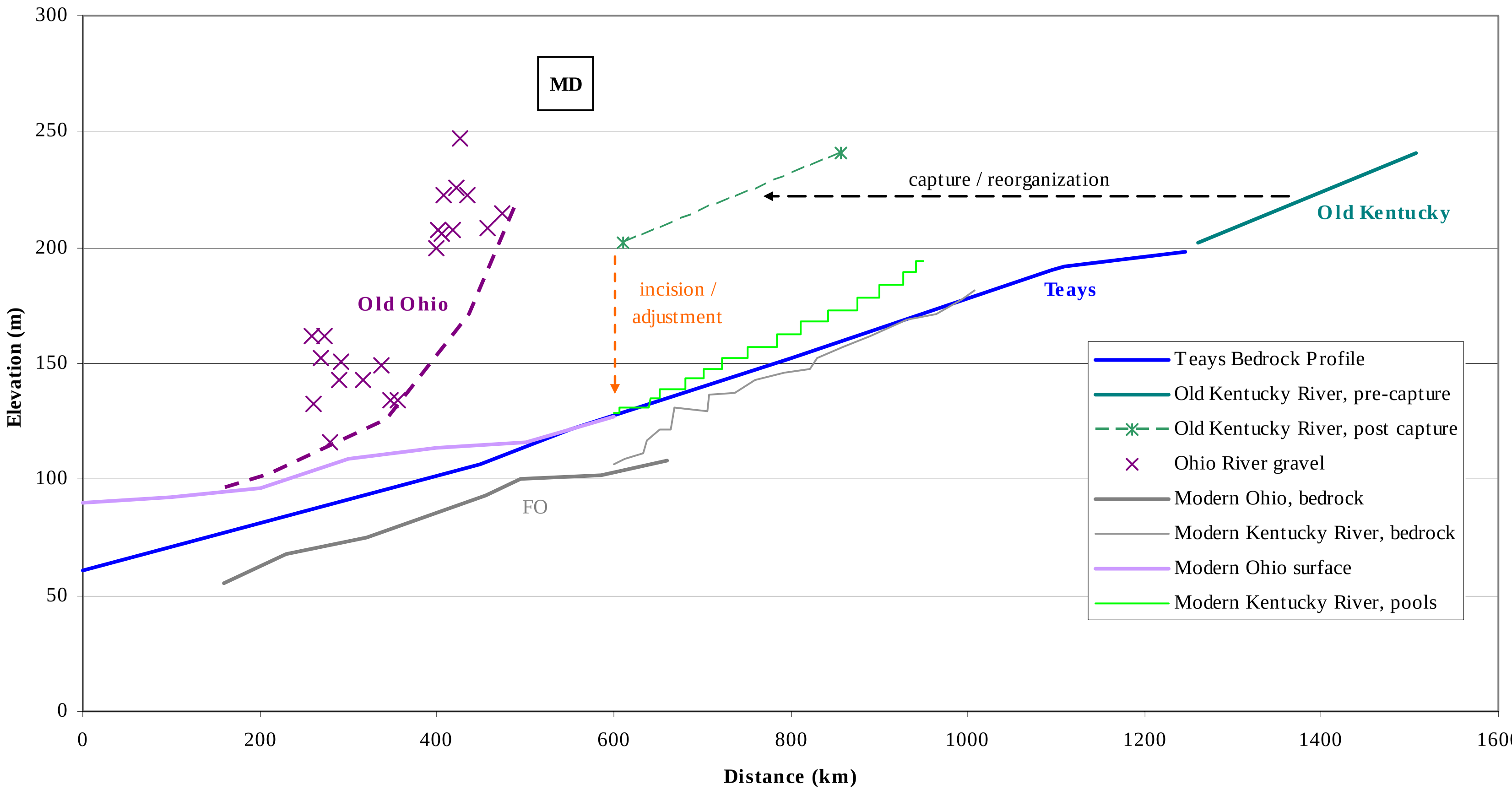


Figure 80. Comparison of the Teays River and Old Kentucky River profiles with the bedrock and surface profiles of the modern Ohio River and Kentucky River. Dashed green line shows the projected profile of the Old Kentucky River immediately following capture by the Old Ohio River. MD = Madison divide; FO = Falls of the Ohio.

profile in Figure 80. These deposits may represent either Ohio River deposition, or may be related to major tributaries such as the Salt River, Green River, and Cumberland River. The distribution of deposits grades toward the ancient divide between the Ohio and Teays at Madison, Indiana. Whether directly or indirectly related to the Ohio River, the distribution of these deposits probably provides a general profile for the Ohio River during the Late Tertiary or early Pleistocene.

When the Ohio River was integrated following the destruction of the Teays River system, the distance from the mouth of the Kentucky River to Cairo, Illinois, decreased by roughly 650 km. The combination of the modern Ohio River and the modern Kentucky River closely parallels the Teays profile as shown on Figure 80. Projection of the Old Kentucky River from its position at the head of the Teays profile to the upland position "above" the modern Kentucky River suggests that the Kentucky River incised to the current level to reach a comparable grade as the ancient Teays River. Base-level adjustment following drainage reorganization of the Ohio River can account for most of the incision observed in the Kentucky River valley between the Old Kentucky River and the modern stream profile.

CHAPTER THREE: GEOCHRONOLOGY, MODELING, AND CONSTRAINTS

Time Scale and Climate

A temporal framework is necessary for consideration of landscape evolution or process rates. This study does not collect new geochronologic or paleoclimate data. Published reports provide a general framework in which the results of the current study can be placed. The results of this study can be used as the foundation for specific targeted geochronologic studies of the Kentucky River basin.

The late Cenozoic time scale is based primarily on oxygen isotopes, geomagnetic polarity, and fossil data. Many of the sediments in terrestrial settings lack abundant diagnostic fossils for biostratigraphic correlation, so dating late Cenozoic sediments in the study area primarily relies on paleomagnetism and cosmogenic radionuclide dating. Even though oxygen isotopes are not typically measured in these sediments because of issues of modern contamination through weathering, etc, many studies relate their findings to the oxygen isotope chronology, so it will be included here. Richmond and Fullerton (1986, and references therein) summarized the Pleistocene glacial and associated stratigraphy of North America, and related the stratigraphy to the oxygen-isotope and geomagnetic polarity time scales for the late Cenozoic. A summary of the regional time scale is shown in Figure 81.

Oxygen has two isotopes, ^{16}O and the less abundant ^{18}O. The lighter ^{16}O is preferentially fractionated in water vapor during evaporation of sea water, leaving the ocean relatively enriched in ^{18}O. The fractionation is enhanced by colder climates, as water molecules with an ^{18}O atom will not have the kinetic energy to vaporize during evaporation. Variations in the isotopic ratio through time, thus, are inferred to reflect global or regional variations in climate. The cyclic variations of oxygen-isotope ratios observed in foraminifera have been dated and assigned sequential numeric stage designations to form a time scale for the late Cenozoic (Shackleton and Opdyke, 1976; Imbrie and others, 1984; Pillans and others, 1998; Kitamura and others, 2001).

The Earth's magnetic polarity has varied from normal (similar to modern) to reversed through time (Easterbrook, 1999). When sediments are deposited, especially those in quiet, low-energy environments, they are affected by the ambient magnetic field at the time of deposition. If the depositional energy is low enough to allow it, magnetically susceptible minerals will be oriented in accordance with the magnetic field during deposition, and thus the rocks preserve the

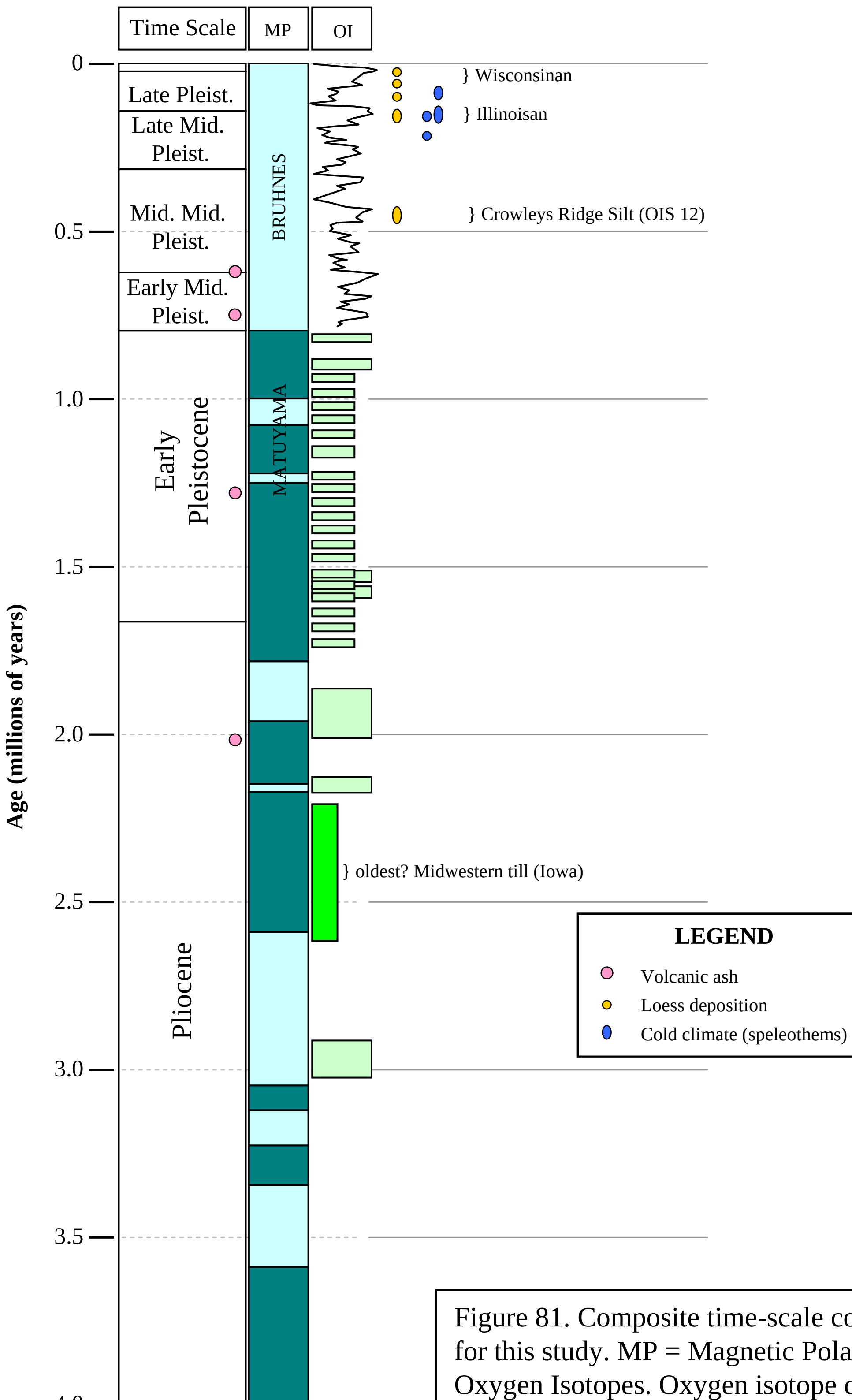


Figure 81. Composite time-scale compiled for this study. MP = Magnetic Polarity; OI = Oxygen Isotopes. Oxygen isotope curve (SPECMAP) shown for 0 to 0.75 Ma. Cold-interval oxygen isotope stages shown in green for 0.75 to 3.0. Ma.

orientation of the magnetic field at the time of deposition. Fine-grained clastic sediments and volcanic materials have recorded the ambient magnetic field as they were deposited and lithified. By dating these materials and noting the magnetic polarity, a chronology of magnetic reversals has been established for the late Cenozoic (Mankinen and Dalrymple, 1979). Berggren and others (1995) provided revised ages for the late Cenozoic geomagnetic polarity time scale (Figure 81).

Loess deposition in the central United States has been inferred to be genetically related to episodes of glacial outwash aggrading major river valleys, when silt was blown off the valley trains and deposited in adjacent uplands. The age of a loess deposit thus approximately represents the time of glaciation. Forman and Pierson (2002) summarized the geochronology of loess deposition in the Mississippi and Missouri River valleys of the central U.S., largely on the basis of thermal luminescence techniques. After a sample of silicate sediment is shielded from sunlight (i.e. buried), ionizing radiation from naturally occurring radioactive isotopes produces free electrons which are subsequently trapped in crystal defects. The introduction of light or heat allows the free electrons to recombine with the sediment, and very faint luminescence emissions are produced in predictable quantities (Forman and Pierson, 2002).

Forman and Pierson (2002) bracketed the ages of four distinct episodes of loess deposition and identified older deposits beyond the range of their dating techniques. Peoria Loess deposition, associated with the Wisconsin glaciation, was dated at 25 to 12 ka. The Roxana Silt, representing Middle Wisconsin glaciation, was deposited 60 to 30 ka. The Teneriffe Silt represents Eowisconsin and is dated at 100 to 80 ka. The Loveland Silt is inferred to represent Illinoisan glaciation, and is dated at 180 to 140 ka. Older loess at Crowley's Ridge was beyond the range of the dating techniques, but Grimley and others (2003) inferred the Crowleys Ridge Silt to correspond to oxygen isotope stage 12 (480 to 430 ky) on the basis of soil-development characteristics of the overlying Yarmouth geosol.

Harmon and others (1978) examined oxygen and hydrogen isotope ratios in growth rings in Mammoth Cave speleothems, dated by uranium-series disequilibrium methods, to estimate climate variations during growth of the speleothems. The record spans from 100 ka to 230 ka, and they identified cold climate conditions at 215 to 195 ka and 160 to 130 ka with an intervening warm period. Thompson and others (1976) conducted a similar study on speleothems in West Virginia, and identified cold conditions at 160 to 110 ka and again at 95 to 55 ka. Hooghiemstra (1984) noted the oldest cold climate signature since 3.5 Ma occurring between 2.5 and 2.2 Ma in a pollen record from Colombia. Groot (1991) examined palynological evidence of Miocene through early Pleistocene climate changes along the Atlantic Coast of North America, and inferred the oldest "colder than modern" interval between 2.5 and 2.0 Ma. Boellstorff (1978) used fission-track dating on an adjacent ash deposit to estimate the age of the oldest sampled till in Iowa at 2.2 ± 0.2 Ma.

Cosmogenic Radionuclide Dating

Background

Traditional radiometric isochron techniques such as K-Ar and U-Pb systems do not work well on very young materials. Application of carbon-14 dating is generally limited to materials younger than 70 ka. Materials from much of the Pleistocene (from 1.5 to 0.01 Ma) are not datable by traditional techniques. Unfortunately, many geomorphological problems, including this study, are directly related to Pleistocene events or deposits. Recent advances in the application of cosmogenic isotopes to geomorphology (Bierman, 1994; Cerling and Craig, 1994; Granger and others, 1996) have enabled the investigation of Pleistocene geomorphological rates and allow rate-based discussion of landform evolution.

Cosmic rays, primarily protons and alpha particles with very high kinetic energies, continually bombard the Earth from outer space. When these particles impact another atom, they may create a spallation reaction in the nucleus of the impacted atom. The spallation reaction involves the release of neutrons and a subsequent reduction in the atomic number. The released neutrons can themselves impact adjacent atoms, causing secondary spallation reactions. In the atmosphere, oxygen, nitrogen, and argon are common targets for the cosmogenic collisions, because of their abundance in the atmosphere.

Some of the cosmic rays, or secondary neutrons produced from spallation reactions in the atmosphere, impact the surface of the Earth causing terrestrial or "in-situ" production of cosmogenic isotopes. These collisions modify the near-surface atoms, producing a variety of nuclides. Many of these nuclides are not stable isotopes and quickly decay. Others, such as ^{26}Al and ^{10}Be have longer half-lives, and may be used for geochronologic applications.

In-situ production of ^{10}Be results primarily from spallation reactions in oxygen. Beryllium has one naturally occurring stable isotope (^{9}Be) and two cosmogenic isotopes (^{7}Be and ^{10}Be). ^{7}Be has a short half-life of 53 days, and is only useful for very short-term studies tracing atmospheric processes and interaction with the lithosphere. ^{10}Be has a half-life of 1.6 m.y., and is more useful for longer term geological and geomorphological studies. Aluminum has only two naturally occurring isotopes: ^{27}Al is stable and ^{26}Al is a cosmogenic radionuclide with a half-life of 0.72 m.y. Spallation reactions with silicon atoms produce ^{26}Al.

^{10}Be is produced both in the atmosphere and in-situ, and can be produced in a variety of minerals. Many of the minerals are susceptible to chemical weathering, which can mix atmospheric ^{10}Be with in-situ ^{10}Be. Quartz, however, is a relatively abundant surficial mineral that is not highly susceptible to chemical weathering. Thus, most if not all of any ^{10}Be measured in quartz should be from in-situ cosmogenic production. Cosmogenic processes also produce ^{26}Al in quartz, providing a second system to test interpretations derived from the ^{10}Be system.

The production of in-situ cosmogenic isotopes is controlled by numerous factors, many of which are not particularly well understood. The factors may all be very difficult to measure or quantify for geologic materials. The key element of cosmogenic production is the flux of neutrons from the atmosphere, which is difficult to measure and is subject to numerous other difficult-to-quantify processes. Earth’s magnetic field shields low latitude areas from part of the cosmogenic flux because the geomagnetic field conducts cosmogenic rays away from the equator and toward the magnetic poles, creating a latitudinal dependence on production rates (Lal, 1991; Dunai, 2000; Desilets and others, 2001). To complicate matters, both the solar production of cosmic rays and the Earth’s magnetic field vary with time.

Other factors are more readily quantified. The depth of the target area within the atmosphere affects the in-situ production rate, because cosmic rays and secondary neutrons lose energy with each subsequent collision as they penetrate the atmosphere. The neutron flux decreases exponentially as the rays penetrate deeper into the atmosphere. The in-situ material being impacted by the cosmic rays also affects the production rate. The size of the nuclei (reaction cross section) affects the probability of a collision occurring, and the depth into the material also affects the production rate through attenuation of the neutron flux.

The accumulation of in-situ cosmogenic isotopes is also affected by the erosion rate. If material is being eroded quickly, the accumulated radionuclides will be removed, reducing the concentration remaining in the surficial material. If erosion rates are slow relative to cosmogenic production, then the radionuclides will accumulate within the surface material.

The cosmogenic radionuclides produced in quartz, ^{26}Al and ^{10}Be, are both radioactive and decrease in abundance with time. Eventually, after approximately 10 m.y., the quartz becomes "saturated" with ^{26}Al and ^{10}Be, as the radioactive loss of these nuclides balances their cosmogenic production. Burial of the quartz, and thus isolation from the cosmogenic source, allows for a burial age to be calculated by comparing the relative abundances of ^{26}Al and ^{10}Be, as the two isotopes decay at different rates. The burial age is based only on the decay of the two isotopes, and is thus not confounded by uncertainties in cosmogenic isotope production rates. The burial age calculation does assume, however, that the sediment received an adequate dose of cosmogenic radiation prior to burial, that it was isolated by burial from cosmogenic influence, and that it has not been exhumed and reburied.

For a given in-situ target undergoing no erosion, the concentration of ^{26}Al and ^{10}Be are each dependent on time and on the production rate for each radionuclide. The production rates for each radionuclide at a given location can be estimated on the basis of latitude and elevation (depth within the atmosphere). Such non-erosional settings are rare in natural environments, however. For sites undergoing steady-state erosion, which may be a reasonable assumption over long enough periods of time, the concentration of each radionuclide depends on the erosion rate at the site and on the production rate for the radionuclides. The production rates for each radionuclide are dependent on the same cosmogenic flux, so the ratio of ^{26}Al to ^{10}Be relative to the concentration of ^{10}Be in the target varies systematically with erosion rate. A "pre-burial" erosion rate, or sediment-source area denudation rate, can be estimated using this technique. When the target is buried, and isolated from cosmogenic production, the two radionuclides decay systematically; the ^{26}Al:^{10}Be ratio is dependent on the duration of burial and the initial concentrations of ^{26}Al and ^{10}Be at the time of burial. Measurement of the concentrations of ^{26}Al and ^{10}Be can be plotted on a graph similar to Figure 82. The curved horizontal lines represent progressive burial ages. The colored near-vertical lines delineate the decay paths for targets derived from areas with specific erosion rates (Figure 82). Post-burial erosion rates can be estimated by comparing the geomorphic setting of sediment with a measured burial age, and

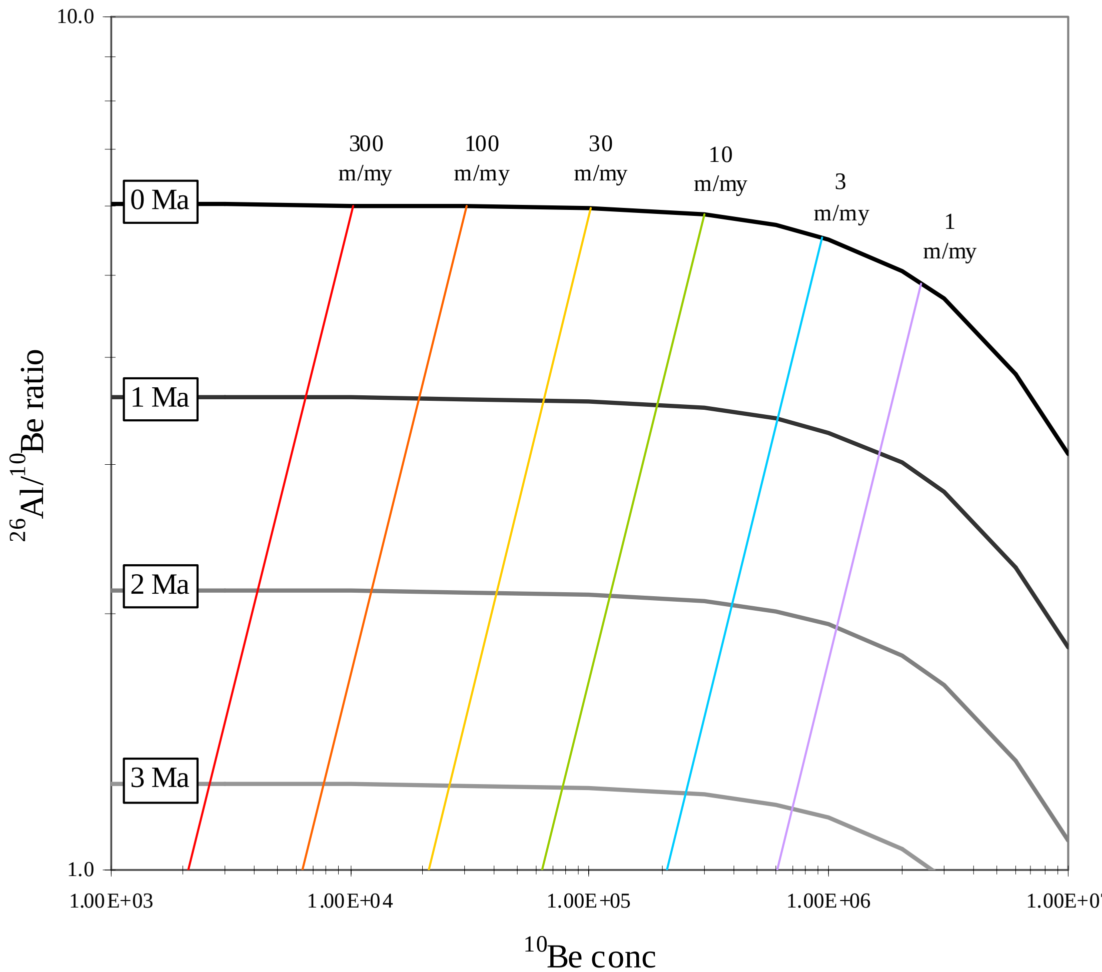


Figure 82. Plot of ^{10}Be concentration versus $^{26}Al/^{10}Be$ ratio. This diagram is used to interpret age and erosion rates from cosmogenic radionuclide studies. Samples at the surface will plot somewhere along the “0 Ma” line, depending on the erosion rate on that surface. Slower erosion rates produce higher concentrations of ^{10}Be. After a sediment is buried and shielded from additional cosmogenic nuclide production, the radiogenic nuclides will decay and the sample values will migrate down the graph parallel to the colored “erosion rate” lines through time. Modified from Granger and others, 1996.

comparing the data with that from sediment in a different (i.e. higher or lower) geomorphic position. Burial ages from sediment in a vertical succession of stream terraces can provide an estimate of the incision rate of a stream.

Two differing erosion rates are commonly pursued in the landscape-evolution literature. Basin denudation rates, or landscape erosion rates, measure the wholesale or blanket removal of material from across an entire landscape. Fluvial incision rates, however, measure the localized vertical erosion rate of a stream within its valley, which may exceed the landscape-erosion rate in adjacent uplands by an order of magnitude or more and may be quite variable through time. A third, and less considered, measure of erosion is that of lateral erosion by a stream valley. The differing concepts of erosion, and their variable rates, must be taken into account when interpreting measured data.

Cosmogenic Burial-Age and Erosion-Rate Studies

Granger and others (2001) analyzed ^{26}Al and ^{10}Be from quartz pebbles in Mammoth Cave to estimate the burial age (cosmogenic shielding age) of quartz sediments (Figure 83). Certain passages in Mammoth Cave reflect a phreatic origin, and are graded toward terrace levels along the Green River (Palmer, 1981). As the Green River progressively incised its valley, higher levels of the cave system were abandoned in favor of lower ones. Granger and others (2001) assumed a genetic link between local incision events along the Green River and regional changes in the Ohio River system. Although several studies have made a similar assumption (Miotke and Palmer, 1972; Palmer, 1981), the link has not yet been demonstrated. Numerous resistant lithologies are present in the bedrock of the valley downstream from Mammoth Cave, and it is possible that the incision events represent the passage of successive knickpoints migrating upstream. However, assuming the former hypothesis is true, the incision events would have some significance for timing of regional events. The data set collected by Granger and others (2001) at Mammoth Cave offers a unique opportunity to constrain the timing of incision events at that particular location along the Green River, regardless of the cause of incision.

The cosmogenic "shielding" dates measured by Granger and others (2001) represent the time at which the pebbles entered the cave system, and not necessarily when they were deposited at the sample locations in the cave. Thus the oldest dates in a passage do not necessarily represent the minimum age of the passage because some sediment may have been inherited from

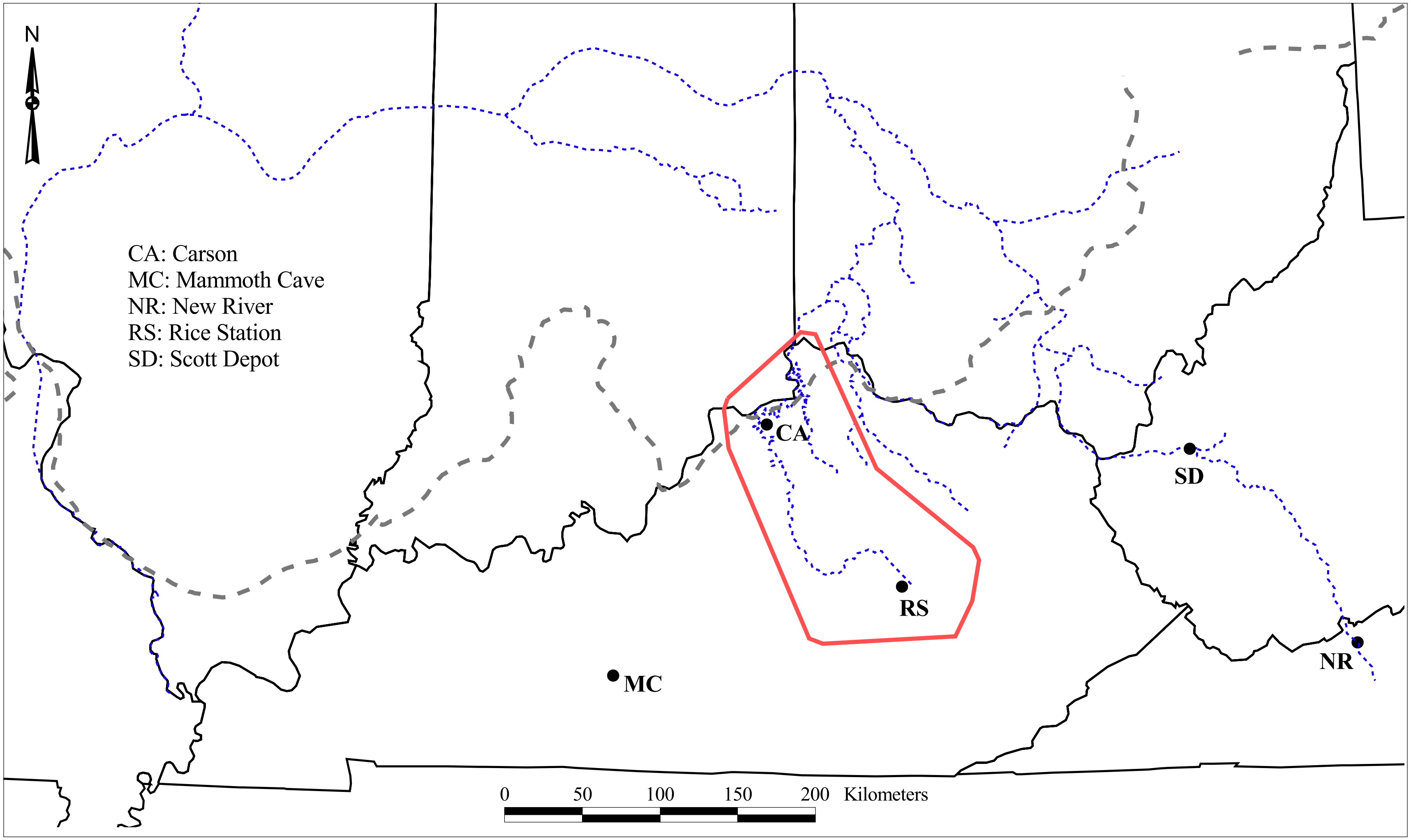


Figure 83. Location of cosmogenic radionuclide study sites discussed in the text.

older levels, although Granger and others (2001) carefully selected their sample sites away from major cave-passage intersections to minimize the effects of sediment inheritance and reworking within the cave. The youngest coarse-grained sediment found in each level represents the maximum age of abandonment of that level, and thus provides an approximate chronology of incision events along the Green River.

Figure 84 shows a profile of the ages and elevations of the sediment measured in different levels of Mammoth Cave. On the basis of these data, Granger and others (2001) inferred incision events to have occurred at approximately 0.75, 1.1, 1.45, and 2.0 Ma (Figure 84). By considering these data in terms of the elevation of the samples, Granger and others (2001) suggested average fluvial incision rates of 30 m/m.y. since 2 Ma. However, as noted by Granger and others (2001), fluvial incision probably occurred during abrupt downcutting events. Fluvial erosion rates were probably highly variable through the Plio-Pleistocene. Maximum fluvial erosion rates for the Green River at Mammoth Cave may be as high as 75 to 100 m/my, on the basis of 15 to 20 m incision events occurring over relatively short (~ 0.2 m.y.) time scales. The ^{26}Al:^{10}Be ratio of the samples suggests upland, sediment-source erosion rates of 2 to 7 m/m.y. for the past 3.5 m.y. (Figure 85) (Granger and others, 2001).

In a similar study, Granger and others (1997) used ^{26}Al and ^{10}Be in quartz-rich cave sediment to infer a burial or “shielding” age of gravels deposited in caves during downcutting of the New River in Virginia (Figure 83). Using a standard regression technique of age vs. elevation, they inferred an average fluvial incision rate of 27.3 ± 4.5 m/my since 1.5 Ma, comparable to the estimate of Granger and others (2001) for the Green River at Mammoth Cave. The data are insufficient to identify punctuated periods of incision comparable to the Mammoth Cave study. The ^{26}Al:^{10}Be ratio of the samples suggests sediment-source erosion rates of 2.5 to 9 m/m.y. for the past 1.5 m.y. (Figure 85) (Granger and others, 1997).

Granger and Smith (1998) measured burial ages using ^{26}Al and ^{10}Be in quartz sediment underlying lacustrine deposits at three unspecified locations in Kentucky and West Virginia. They reported an average age of Teays valley sediments or equivalents to be 1.13 ± 0.16 Ma. Granger (personal communication, 2001) also used ^{26}Al and ^{10}Be to estimate burial ages of sediment at Carson, Kentucky, and Scott Depot, West Virginia (Figure 83). The samples for this study were collected roughly 10 m below the surface of exposed fluvial deposits. The age

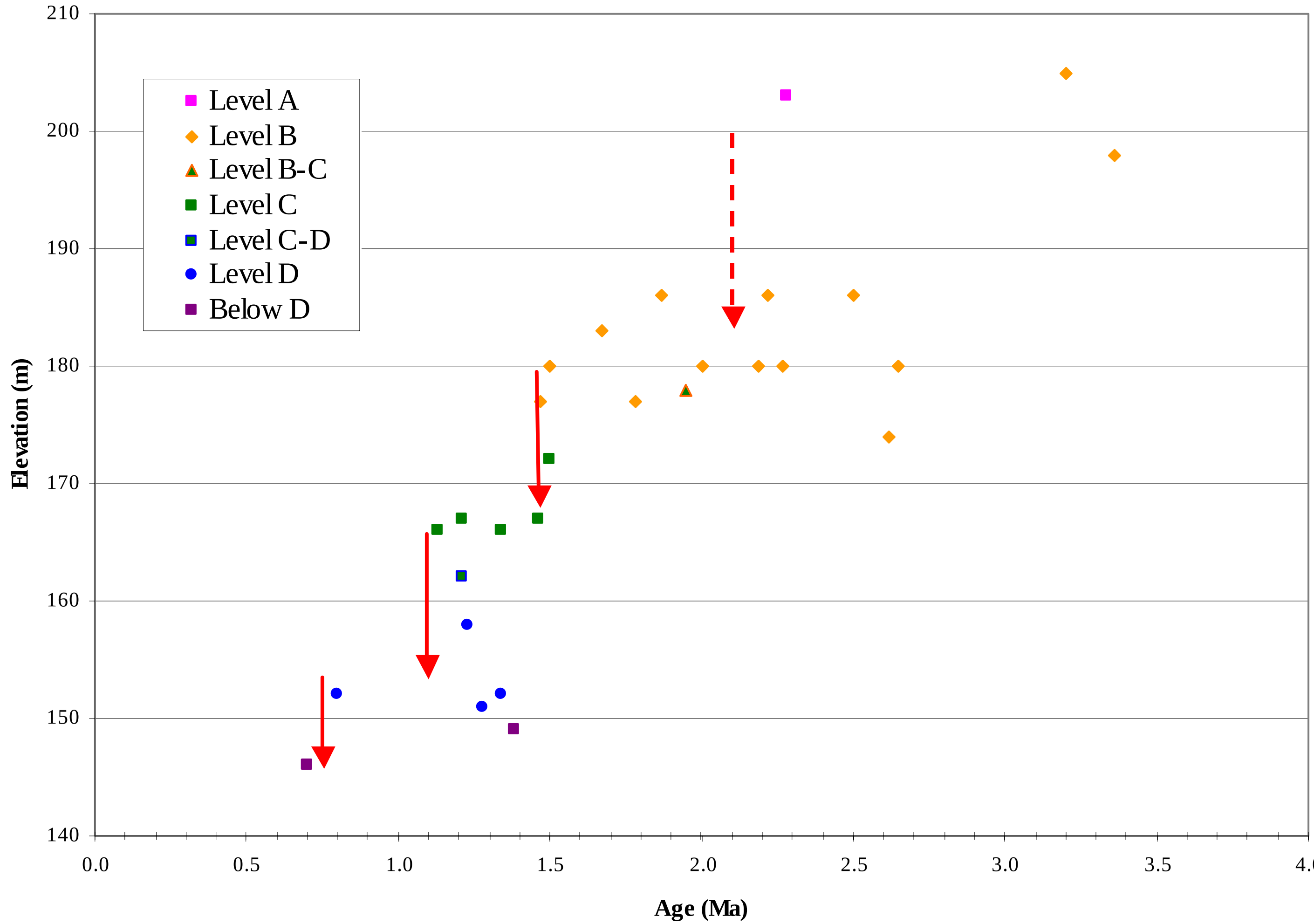


Figure 84. Comparison of elevation versus age for Mammoth Cave sediments as determined by Granger and others (2000). Red lines denote inferred episodes of incision by the Green River.

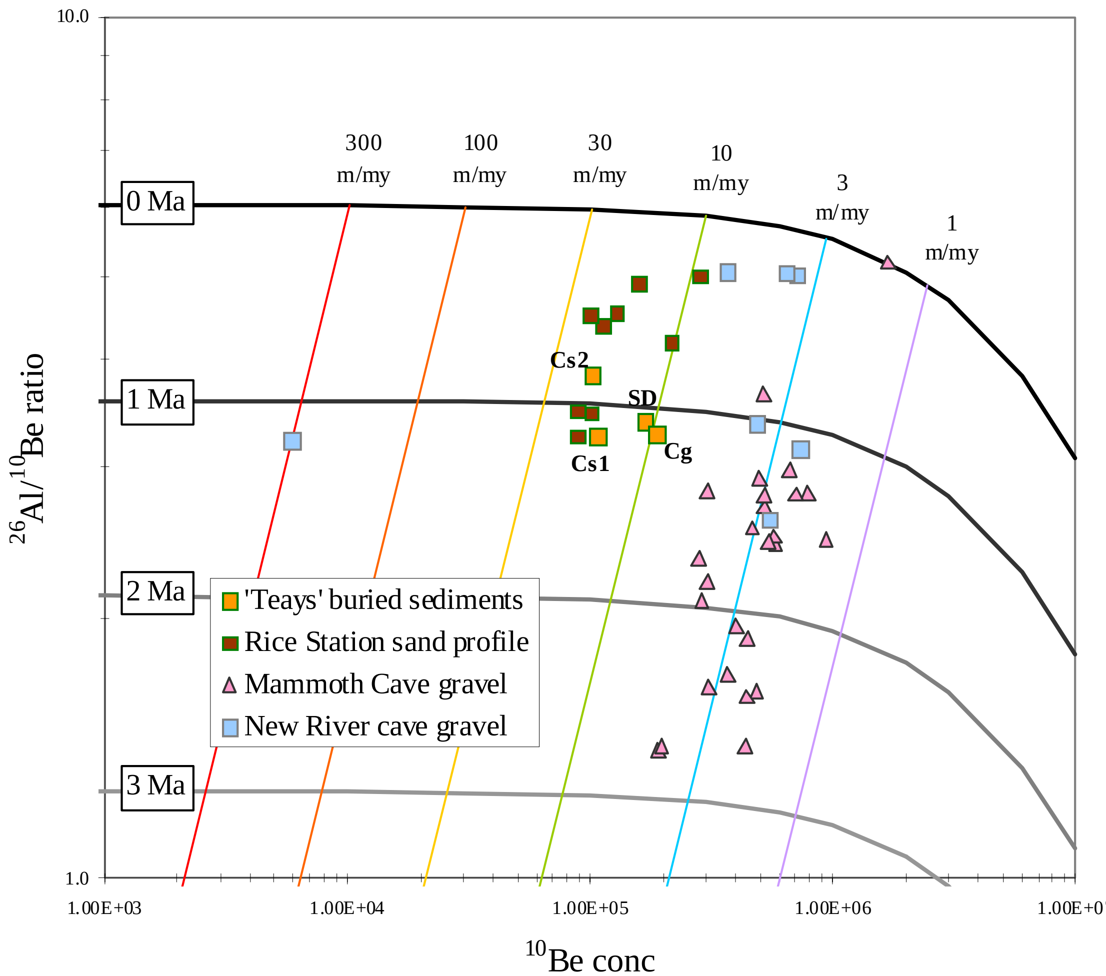


Figure 85. An erosion-age plot of data in studies discussed in the text. Mammoth Cave data from Granger and others (2001); New River data from Granger and others (1997); Rice Station data from Granger and Smith (2000); Teays sediment data from Granger (personal communication, 2001). Cg, Carson gravel, Cs1 and Cs2, Carson sand; SD, Scott Depot.

estimates reported by Granger (personal communication, 2001) do not agree with the estimates as plotted on Figure 85. The reported burial age for gravel from the Carson site is 1.43 $^{+0.32}/_{-0.31}$ Ma. Sand samples from the Carson site have reported burial ages of 1.93 $^{+0.74}/_{-0.57}$ Ma and 1.36 $^{+0.57}/_{-0.55}$ Ma. The Scott Depot gravel has a reported burial age of 1.30 $^{+0.43}/_{-0.40}$ Ma. The data as plotted on Figure 85 suggest burial ages of 0.9 to 1.2 Ma.

Granger and Smith (2000) measured the concentration of ^{26}Al and ^{10}Be in a 10-m profile of Irvine Formation sand near Rice Station in Estill County, Kentucky, at an elevation of 275 m (900 ft) (Figure 83). They used an empirical approach to model various factors—including sediment density, sediment-source erosion rate, surface erosion rate, and time—to produce a best fit model age for concentrations of ^{10}Be in the profile. Granger and Smith (2000) examined the influence of varying the different parameters on an idealized production/isotope-concentration curve. Granger and Smith (2000) then substituted what they considered to be geologically reasonable values for sediment density (1.8 g/cm^2), pre-burial sediment-source area erosion rate (50±5 m/m.y.), and modern surface erosion rate (6.2±0.2 m/m.y.). They reported the resulting estimated model age to be 1.50 $^{+0.32}/_{-0.25}$ Ma.

Granger (personal communication, 2001) noted an earlier estimate of the Rice Station model age to be 1.74 $^{+0.41}/_{-0.33}$ Ma, but that it had been revised on the basis of unspecified "additional data" to the published estimate of 1.50 Ma. The data at Rice Station, as plotted on Figure 85, would suggest much younger ages of 0.4 to 1.2 Ma. The ^{10}Be profile for the site is well within the depth of cosmogenic influence of roughly 10 m for this setting. Thus, the younger plotted ages for Rice Station, compared to the older modeled ages for the data, most likely result from the post-burial cosmogenic production of ^{10}Be.

The data from the Rice Station, Carson, and Scott Depot sites collectively suggest sediment-source erosion rates of 10 to 30 m/m.y. (Figure 85). The sediment-source erosion rates are similar for studies in comparable settings: 2 to 9 m/m.y. for "shielding" ages in caves, and 10 to 30 m/m.y. for "exposed" fluvial deposits. This may represent either a systematic flaw in the assumptions of these studies, which would be consistent with the age discrepancies in the exposed fluvial samples. These sites may have undergone multiple episodes of cosmogenic production, which would render ages and erosion rates estimated by the current techniques to be unreliable. Renewed production of cosmogenic radionuclides would create anomalously young ages for the samples. The plotted erosion rates would represent an average of inherited sediment-source $^{26}Al:^{10}Be$ ratios and $^{26}Al:^{10}Be$ ratios consistent with the comtemporary erosion rate of the fluvial sediment. Alternatively, the data may actually represent a genuine natural trend of higher sediment-source erosion rates in the Kentucky and West Virginia parts of the Teays River system. If so, the plotted data suggest a gradual increase in sediment-source erosion rates through time.

The Carson site (elevation 201 m) is located very near the "floor" of high-level fluvial deposits in the Kentucky River valley, estimated to be 200 m near Carson. The Rice Station deposit (elevation 275 m) is approximately 35 m higher than the upper outlet of the Levee spillway (240 m), which is the upstream end of the "floor," and should therefore be older than the Carson deposit (Figure 86). Assuming a nearly coincident age of the two sites (1.43 and 1.36 Ma as reported at Carson, and 1.50 at Rice Station), as reported by Granger and Smith (2000) and Granger (personal communication, 2001) would suggest incision rates of nearly 500 m/my. The development of numerous abandoned meanders and fluvial deposits between the two levels argues against an extreme fluvial incision rate for the interval above the "floor." Using the plotted age of the Carson site gravel (1.1 Ma) and the discarded age of Rice Station, 1.74 Ma, results in calculation of a more moderate erosion rate of 55 m/my. Figure 87 illustrates the range of possible ages for the Rice Station and Carson deposits, and shows the ages suggested by Figure 85.

Lateral Erosion Rates

The lateral erosion rate is a key variable in determining valley morphology. Valleys constrained by lithologies resistant to lateral erosion have different morphologies from those underlain by lithologies more susceptible to lateral erosion. Fluvial incision rates as large as 100 m/m.y. have been estimated for Kentucky streams, as discussed in the previous sections. No direct estimates of lateral erosion rate have been published for Kentucky streams. The closest study of this sort is that by Brakenridge (1984, 1985) along the Duck River, a tributary of the Tennessee River.

Brakenridge (1984, 1985) used trenching and radiocarbon dating to investigate the late Pleistocene and Holocene floodplain stratigraphy of the Duck River in west central Tennessee. The Duck River has developed an "in-grown" pattern of incised meanders with asymmetric

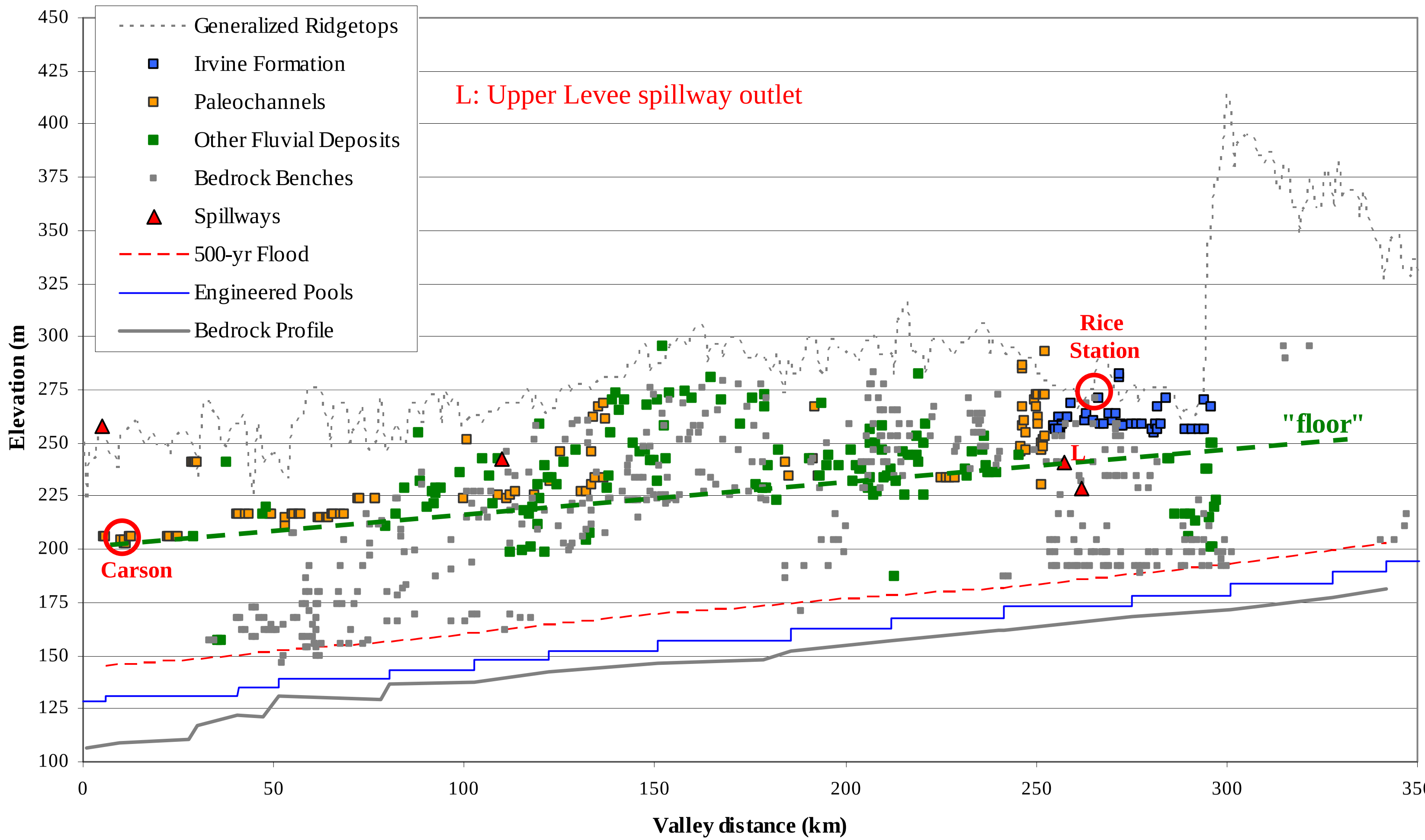


Figure 86. Longitudinal profile of the Kentucky River valley showing elevations of paleochannels, high-level fluvial deposits, including the Irvine Formation as mapped by Campbell (1898), bedrock benches, and spillways draining into and out of the Kentucky River basin. Cosmogenic geochronology studies discussed in the text are shown by red circles.

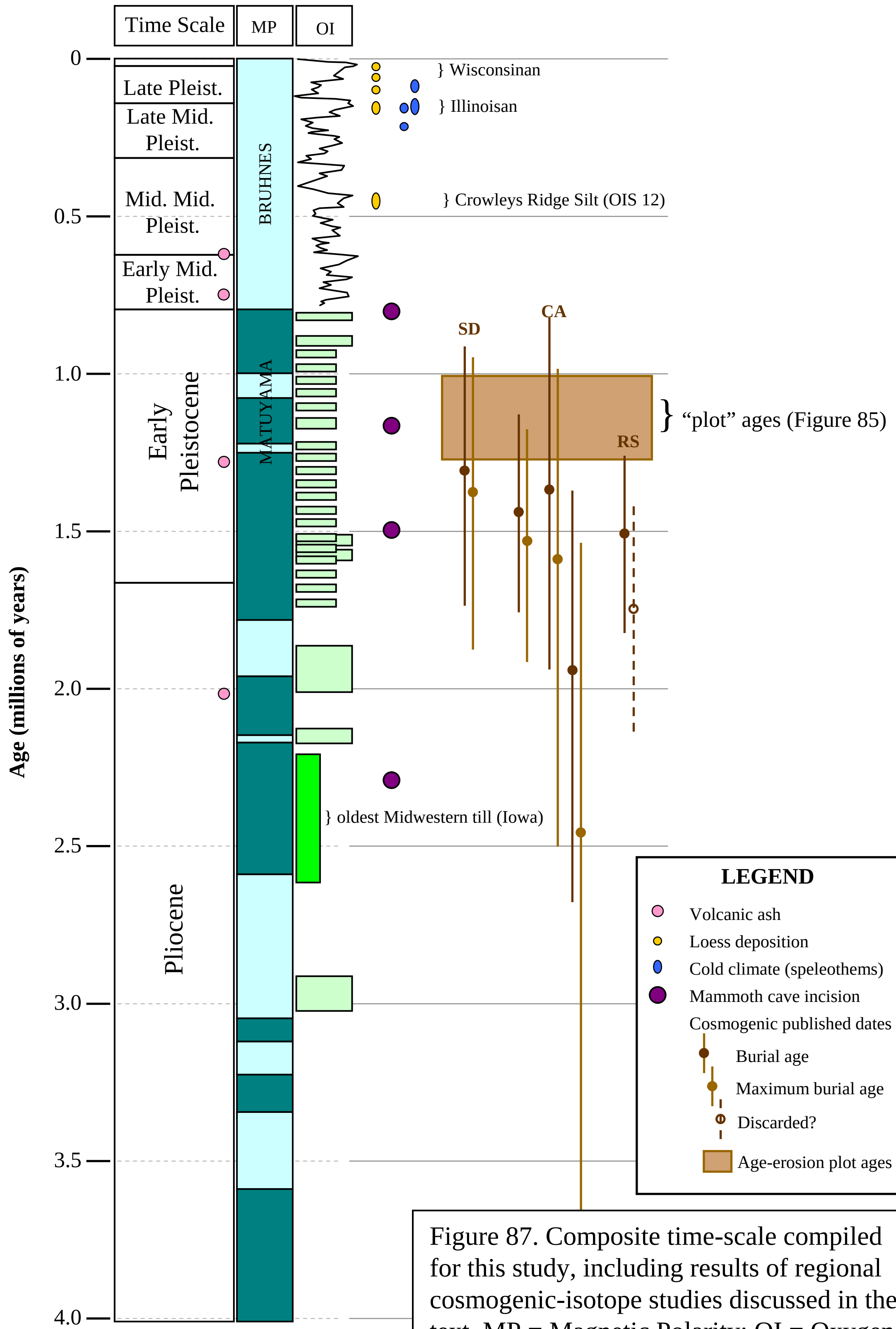


Figure 87. Composite time-scale compiled for this study, including results of regional cosmogenic-isotope studies discussed in the text. MP = Magnetic Polarity; OI = Oxygen Isotopes; CA = Carson; RS = Rice Station; SD = Scott Depot.

valley walls and slip-off slopes eroded into Ordovician Lebanon Limestone (thin-bedded limestone with shale partings) and Ridley Limestone (thick-bedded limestone with minor dolostone beds). The lithologies are similar to those in the Tyrone Limestone and Oregon Formation, respectively, in the Kentucky River valley. The stratigraphy and lithologies are comparable to those associated with the Tyrone style of valley morphology in this study. On the basis of comparison of the two rivers on 1:250,000 topographic maps, the Duck River and Kentucky River have comparable meander amplitudes and wavelengths. Detailed measurements of alluvial-valley widths are not available for the Duck River.

Using detailed stratigraphy of the floodplain material to reconstruct channel positions during the last 30,000 years, Brakenridge (1985) estimated lateral stream migration rates of 6 to 19 m/ky (6,000 to 19,000 m/m.y.), and lateral bedrock-cliff erosion rates of 5 to 14 m/ky (5,000 to 14,000 m/m.y.) with negligible vertical erosion during that same time. These short-term, lateral rates are two to three orders of magnitude greater than vertical incision rates measured elsewhere in the region, and discussed in the previous sections. It should be noted, however, that lateral erosion was focused only where cut banks of the Duck River impinged upon the bedrock walls of the valley, and were neither spatially nor temporally continuous through the study area.

Paleomagnetic Data

Bonnett and others (1991) observed normal and reversed magnetic polarity in samples from Minford Silt Member of Teays Formation (impoundment phase of Teays Valley). The magnetically reversed measurements were from less weathered samples. They attributed the normally magnetized measurements to weathering, and thus “resetting” in a normal magnetic setting. They inferred deposition of the Minford Silt to have occurred during the Matuyama reversed polarity chron (0.79 to 1.60 Ma). The Calcutta Silt is another fine-grained valley fill unit in valleys northeast of the Teays system. Lessig (1961, 1963, 1964, cited in Fullerton, 1986) noted that the Calcutta Silt is older than the Minford Silt on the basis of stratigraphic, geomorphic, and pedologic criteria.

Bleuer (1991), in his review of the stratigraphy of materials filling the buried Teays-age Lafayette valley in Indiana, noted that the oldest till (West Lebanon Till Member) rests upon pro-glacial sediments exhibiting reversed magnetization. Teller and Last (1981) performed preliminary paleomagnetic investigations of the Claryville Clay in the Licking River valley as part of a more comprehensive study. They noted that the Claryville Clay is pre-Illinoisan in age on the basis of geomorphic and stratigraphic relationships. All of their samples yielded normal remnant magnetism. Ettensohn (1974) noted that most of the Claryville Clay is weathered, and thus the paleomagnetic signature of these samples may have been reset by post-Matuyama weathering.

Schmidt (1982) examined the paleomagnetic polarity of fine-grained sediments in Mammoth Cave. He identified reversed polarity in samples between 170 and 208 m (560 and 680 ft) in elevation in the cave system. He inferred these sediments to have been deposited during the Matuyama reversed polarity epoch. The subsequent cosmogenic dating on cave gravel by Granger and others (2001) supports that hypothesis. Pease and others (1994) and Pease and Gomez (1997) correlated a similar pattern of paleomagnetic signatures in cave sediments in southern Indiana with those in Mammoth Cave.

Glacial Loading and Crustal Flexure

The current study examines landforms and fluvial deposits in the context of modern elevations. However, the advance of Pleistocene continental ice sheets introduced a significant load on the lithosphere of the region. The underlying asthenosphere is displaced as the lithosphere is depressed by a mass of overlying ice. Understanding the potential rate and magnitude of lithospheric deflection can be useful to reconstruct environments during a glacial advance. The rate of downward displacement is constrained by the viscosity of the asthenophere. If the ice remains in place long enough, isostatic equilibrium will be achieved. The mass of the displaced asthenosphere will be equal to the mass of the displacing ice. The equations discussed below are from, or are modified from Turcotte and Schubert (1982). Explanation of symbols is in Table 4.

$$\rho_m g h_m = \rho_i g h_i$$

Reorganizing the equation shows that the displacement of the lithosphere (equal to the displacement of the asthenosphere) is proportional to the thickness of the ice sheet and the density ratio of ice to asthenosphere.

$$h_m = (\rho_i/p_m)h_i$$

Table 4. Explanation of symbols used in flexure calculations

Symbol	**Parameter**	**Units**
ρ_i	density of ice	kg/m^3
ρ_m	density of the mantle	kg/m^3
g	gravitational acceleration	m/s^2
h_m	displacement of the mantle*	m
h_i	thickness of ice	m
h	thickness of lithosphere	m
D	flexural rigidity	$(kg.m^2)/s^2$
E	Youngs modulus	$kg/(m.s^2)$
υ	Poissons ratio	
α	flexural parameter	m
w	lithospheric displacement	m
τ_r	characteristic response time	s
μ	viscosity	$(kg.s)/m^2$
λ	wavelength	m
w_b	height of forebulge	m
x_b	distance to forebulge	m
x_0	width of proximal trough	m

** or maximum lithospheric defelection*

The plate beyond the margins of the ice is bent in response to the ice load, and a trough is formed proximal to the ice front. The flexural rigidity of the plate incorporates the plate thickness, Poisson's ratio, and Young's modulus:

$$D = Eh^3 / (12 (1 - \upsilon^2))$$

The flexural parameter is determined by

$$\alpha = [4D / ((\rho_m - 1)g)]^{0.25}$$

The deflection equation describes the ideal response of a homogeneous plate; the maximum displacement is assumed to be the magnitude of the isostatic adjustment of lithosphere at/near the ice margin.

$$w = h_m e^{(-x/\alpha)} (\cos(x/\alpha) + \sin(x/\alpha))$$

Rearrangement of the equation will allow for determination of the ideal half-width of the trough proximal to the ice, which is constrained by the flexural parameter

$$x_0 = 3\pi/4 * \alpha$$

The distance to the crest of the associated forebulge is also constrained by the flexural parameter, and is described by

$$x_b = \pi * \alpha$$

The amplitude of the bulge is directly related to the thickness of the ice sheet:

$$w_b = - h_m e^{-t} = -0.0432\ h_m$$

Essentially, the flexural properties of the underlying lithosphere determine the wavelength of the crustal response to the ice load, whereas the thickness of the ice sheet determines the amplitudes of the surface deflections. The crust can be assumed to have had reasonably similar properties in the Pleistocene as now, but the thickness of glacial ice is not well known or constrained.

Figure 88 illustrates a variety of estimates for the equilibrium flexure proximal to a hypothetical ice sheet. The estimates assume that ice thickness ranges between 1000 and 1500 m, and effective flexural thickness is 35 to 40 km. The ice-proximal trough would be 190 to 225 km wide. The maximum depression at the ice margin would be approximately 30% of the ice thickness. The crest of the forebulge would be 260 to 300 km from the ice front, and would be no

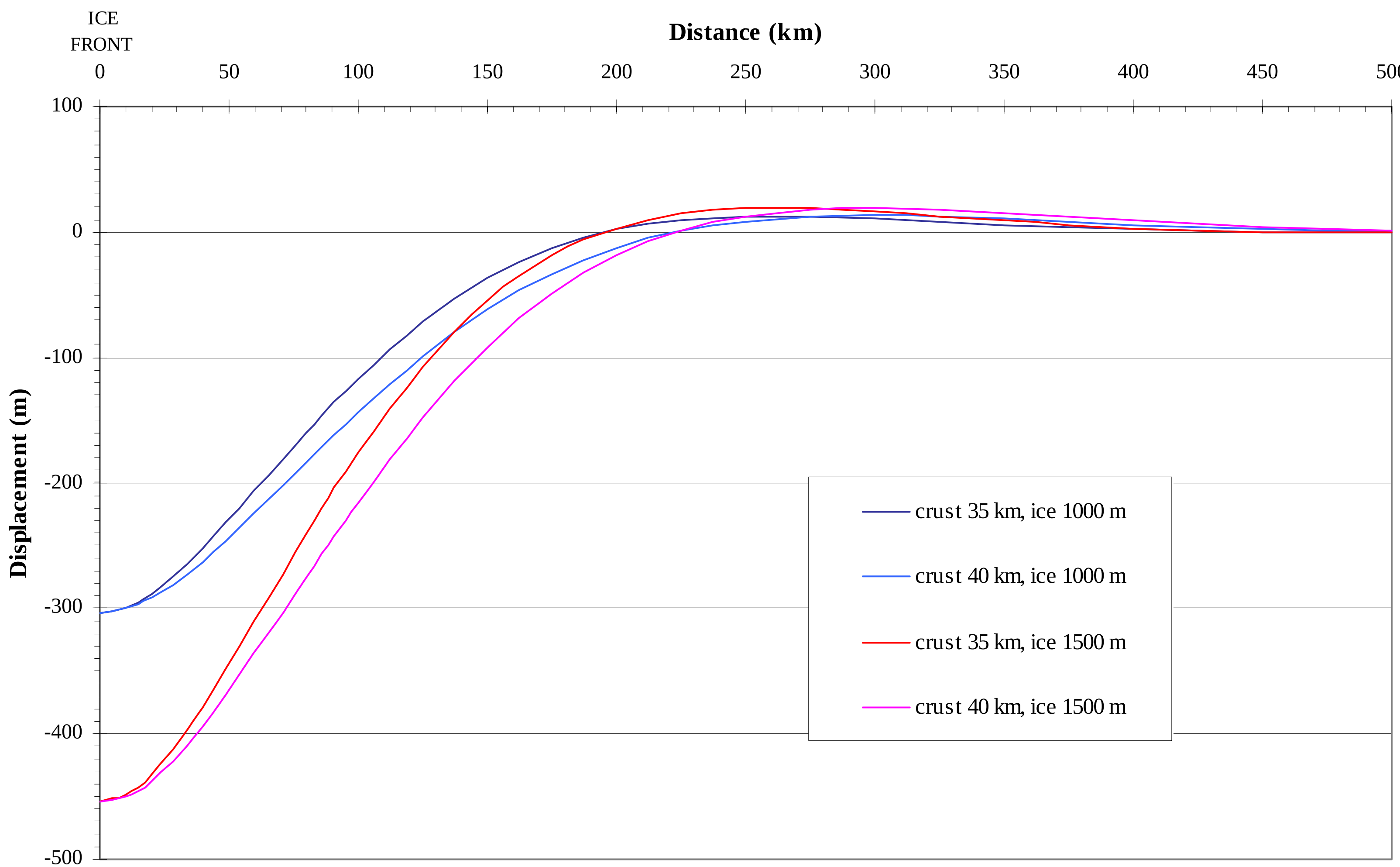


Figure 88. Results of lithospheric flexural modeling performed for this study. Thickness of the lithosphere (“crust”) and thickness of the ice load were the primary variables to affect the results.

more than 20 m high. The amplitudes of responses should be less if the ice sheet does not remain in place long enough for equilibrium conditions to develop. Figure 89 shows the maximum flexure across central Kentucky using the maximum limit of pre-Illinoisan glacial deposits, and assuming an ice thickness of 1000 m.

Application of these equations assumes that the ice was stable at its maximum extent for a length of time sufficient for isostatic equilibrium to be reached. The required response time is constrained by the viscosity of the displaced lithosphere. The characteristic time of the flexural response is given by

$$\tau_r = (4\pi\mu)/(\rho_m g\lambda)$$

The exponential flexural response as a function of time is described by

$$w = h_m\, e^{(-t/\tau)}$$

The thickness of the responsible ice sheet is unknown, so the exponential portion of the equation can be used to model the percent response of the system, independent of ice thickness. The flexural response as a function of time can be graphed as shown in Figure 90. The timing and duration of pre-Wisconsin glacial maxima is poorly constrained, and the actual flexural situation was probably much more dynamic. An incompletely developed trough-bulge system migrated with the advance and retreat of glacial ice. The potential dynamic responses of flexure associated with a moving or fluctuating ice load are beyond the scope of this study.

The configuration of the maximum preserved extent of glacial ice for Illinoisan and pre-Illinoisan glaciers would have completely blocked the mouth of the Licking River valley. The amount of time the glacier remained at this maximum would determine the magnitude of crustal depression proximal to the ice front, and thus affect the extent of the resulting ice-dammed lake in the Licking River valley. If the depression reached a maximum, the topographic divides adjacent to the ice front would be depressed significantly below the original elevations; ice-marginal flow over the divides would prevent extensive lakes and lake deposits. The Claryville Clay and associated lacustrine deposits have been well documented (Ettensohn, 1974; Luft, 1980; Teller and Last, 1981), supporting the model of significant slackwater deposition in the Licking River valley. A spillway between the Licking and Kentucky River basins, at a modern elevation of 270 m (890 ft), has been documented by this study near the community of Levee in

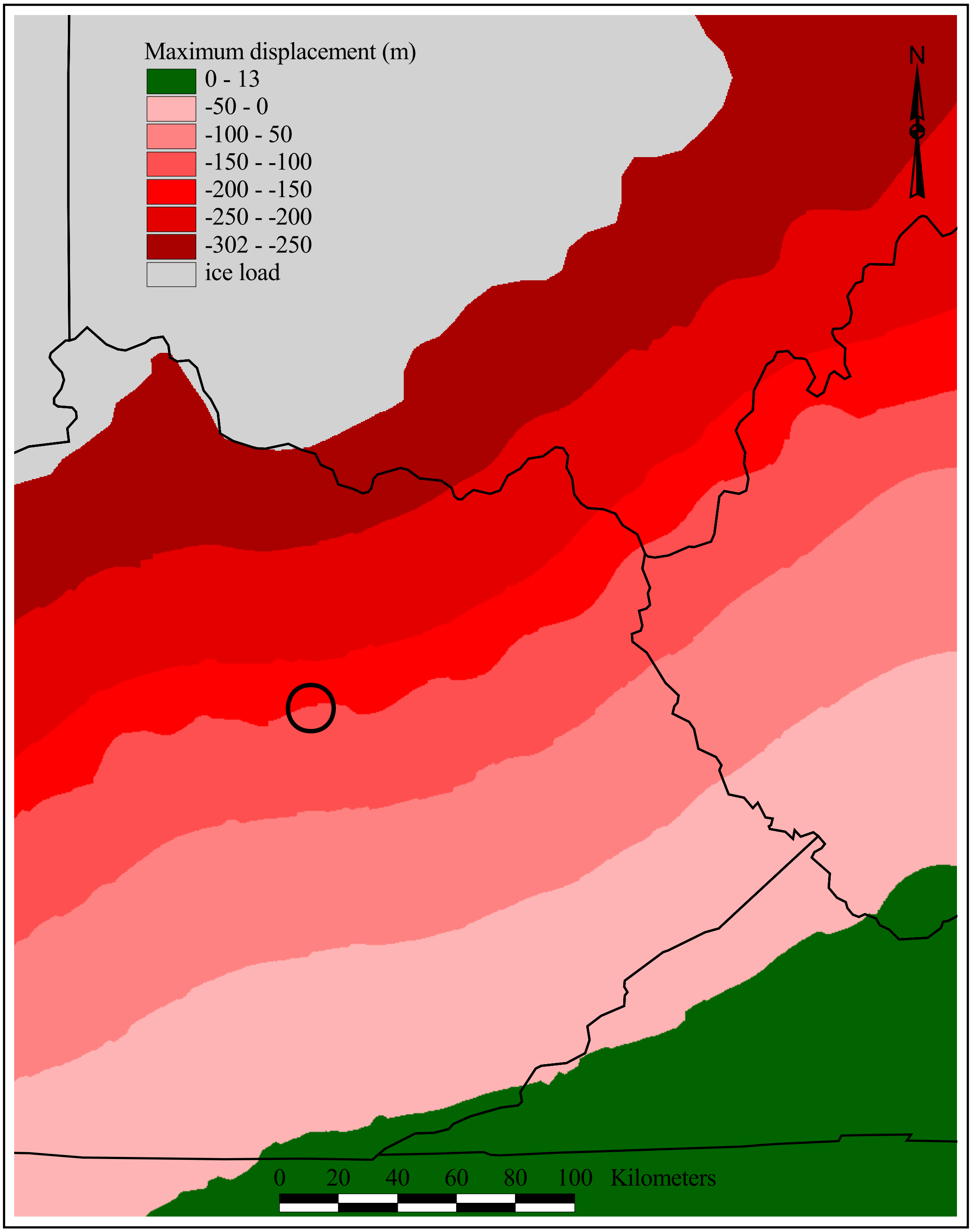


Figure 89. Magnitude of displacement modeled for a 1000-m thick ice sheet in isostatic equilibrium at the mapped pre-Wisconsin glacial limit. Black circle indicates the location of the Levee spillway.

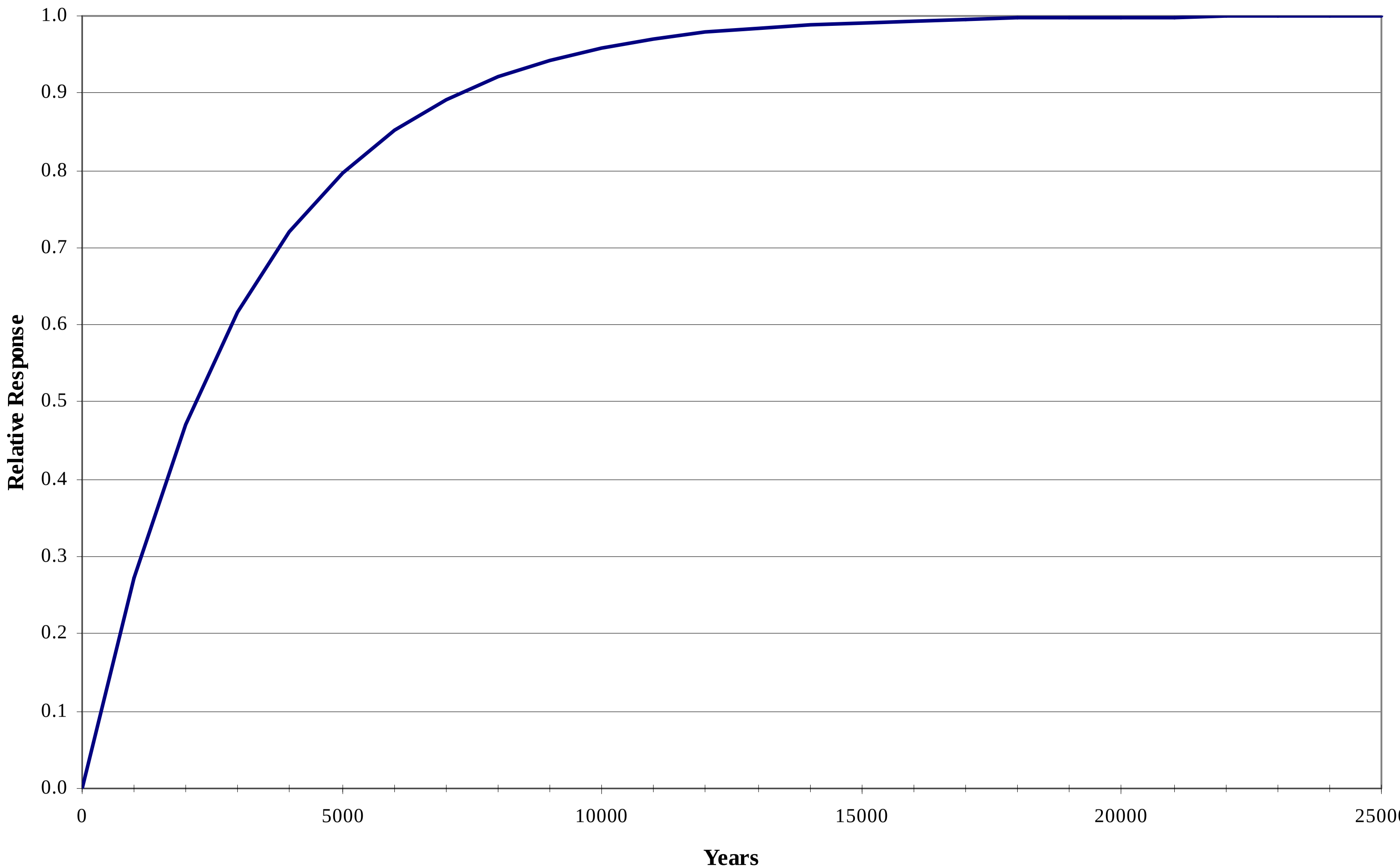


Figure 90. Estimated time of isostatic response of the lithosphere.

Montgomery County. As such, it provides a minimum extent to which the impounded water must have reached upvalley in order for the spillway to have formed and existed.

Digital elevation models can be used to compare different potential scenarios of glacial flexure in central Kentucky. Figure 91 shows the estimated extent of a lake on the modern topography with no glacial flexure imposed, filling the lake to the elevation of the Levee spillway. Figure 92 shows the same extent with flexure imposed, filling the lake to the level of significant overflow of the ice-proximal divide. The hypothetical lake, assuming flexural depression, does not reach the Levee spillway.

Using modern topography as a base, and filling the areas upstream of Cincinnati to an elevation of 270 m (890 ft) results in a hypothetical impoundment of approximately 3 million hectares in extent and 1,000 billion cubic meters in volume. The modern average flow of the Ohio River past Cincinnati is approximately 100 billion cubic meters per year. This suggests that the extensive lake behind any ice blocking the Ohio and Licking River systems would have filled on a decade timescale, which is nearly instantaneous in comparison to the millennial response of crustal flexure. The syn-glacial flow of the Ohio River and the syn-glacial precipitation and runoff regime are not known. However, an order of magnitude difference between the modern and glacial values would result in estimates of lake impoundment on 1- to 100-year time scales. Thus, a lake could have rapidly formed and extended far upstream as in Figure 91, and gradually evolved with progressive lithospheric flexure to more resemble Figure 92. As crustal flexure occurred, different spillway exits were activated for the impounded waters progressively downstream, until the water drained across the ice front in a deeply flexed setting, assuming that the ice stayed in position long enough for this scenario to completely develop.

Overburden Reconstruction and Paleogeography

To consider rates of fluvial erosion in context, the total overburden removed from the study area and the rates at which it was removed must be estimated. Much of the Paleozoic stratigraphic section, exposed elsewhere in Kentucky, has been eroded from the crest of the Cincinnati arch in central Kentucky. Using thicknesses from geologic maps and from subsurface data (Figure 5), hypothetical thicknesses of these stratigraphic units can be projected over the arch. Silurian strata are known to pinch out beneath a Middle Devonian unconformity along the axis of the arch (Peterson, 1981; McDowell, 1983) and are not included in the reconstruction.

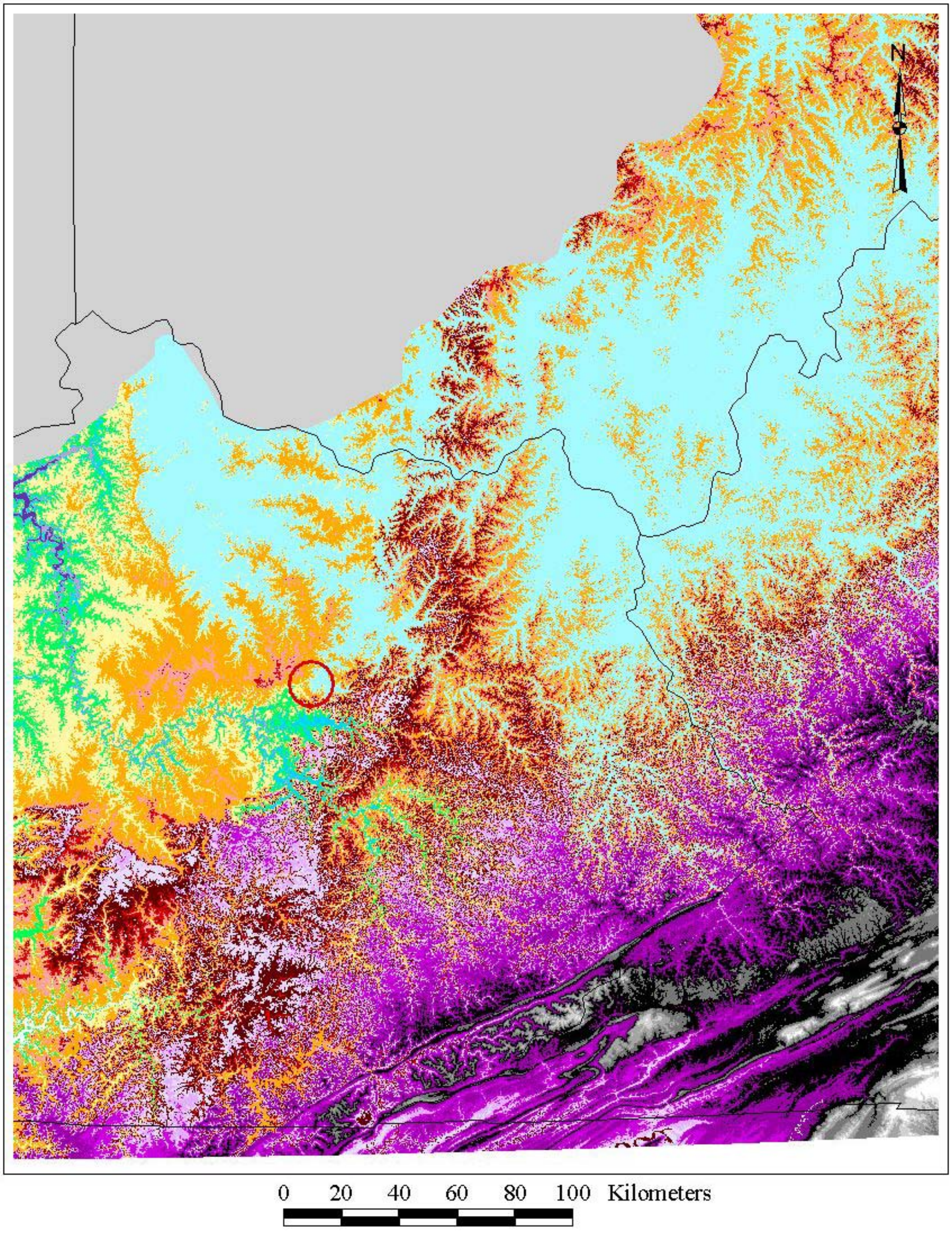


Figure 91. Estimated extent of lake, shown in light blue, resulting from impoundment of river valleys by a generalized pre-Wisconsin glaciation at the mapped glacial limit. No isostatic flexure is imposed on the modern topography. Red circle indicates the location of the Levee spillway.

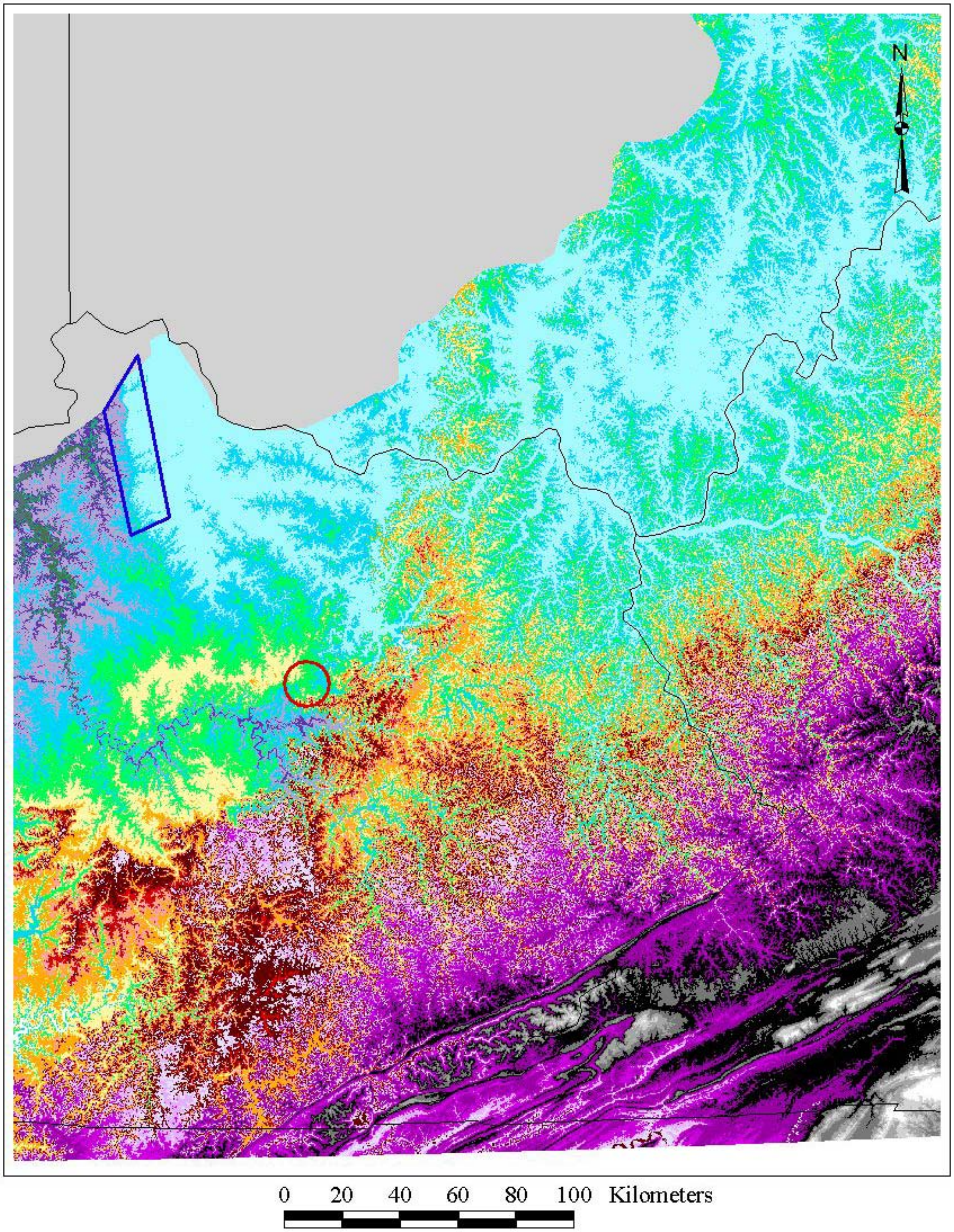


Figure 92. Estimated extent of lake, shown in light blue, resulting from impoundment of river valleys by a generalized pre-Wisconsin glaciation at the mapped glacial limit. Isostatic flexure is imposed on the modern topography. Red circle indicates the location of the Levee spillway. Blue box indicates area of extensive spillover of ice-marginal divide.

Figure 93 uses maximum estimated thickness of non-Silurian units (by assuming the units do not thin onto the arch) to extrapolate a maximum thickness of cover across the arch. Assuming some of these stratigraphic units thin toward the arch, the extrapolation should provide a maximum thickness of removed sedimentary cover.

Because these projections are relatively arbitrary, independent means of confirmation can be used to constrain the estimated thicknesses. Thermal maturity indicators such as the conodont alteration index (Epstein and others, 1977; Harris and others, 1978; Ryder, personal communication, 2004), vitrinite reflectance (O'Hara and others, 1990; Williams, 1984), and fixed carbon (Williams, 1984) can be used to estimate the depth to which a sediment body was buried, after making assumptions about the thermal gradient. Vitrinite reflectance can be used to roughly constrain the thermal gradient, if enough data are available (O'Hara and others, 1990). The available thermal maturity data have comparable spatial trends, suggesting that a reasonable estimate of overburden can be made (Harris and others, 1978). The "restored" land surface for conodonts from Paleozoic rocks, on the basis of conodont data in Epstein and others (1977), Harris and others (1978), and Ryder (personal communication, 2004), and the tops of the relevant stratigraphic units are shown in Figure 94. Adding from 425 to 700 m of overburden onto the existing topography provides the best fit to the thermal maturity data. The 700-m thermal maturity reconstruction is in close agreement with the stratigraphic thickness extrapolation estimate.

The youngest bedrock units preserved in the Appalachian and Illinois basins are Permian in age. During the Late Pennsylvanian and early Permian, the study area was along the western margin of an overfilled foreland basin (Figure 95) (Donaldson and others, 1985; Chesnut, 1994; Greb and others, 2002). The study area experienced a fluctuating coastal margin between fluvio-deltaic sediments being transported northwest from the Appalachian highlands on the southeast and marine incursions from the west (Heckel, 1995). Sandstones of the Upper Pennsylvanian Conemaugh and Monogahela Formations in northeastern Kentucky trend to the northwest (Rice and others, 1979), subparallel to the dominant drainage directions of modern streams and perpendicular to structural trends of the Alleghanian orogen (Figure 95). Paleodrainage systems in the Pennsylvanian of the Illinois basin trend to the west (Wanless and others, 1963). As basin depocenters migrated northward when the basin overfilled (Greb and others, 2002), fluvial systems carried sediment across the filled Appalachian basin into the Illinois basin. The fluvial

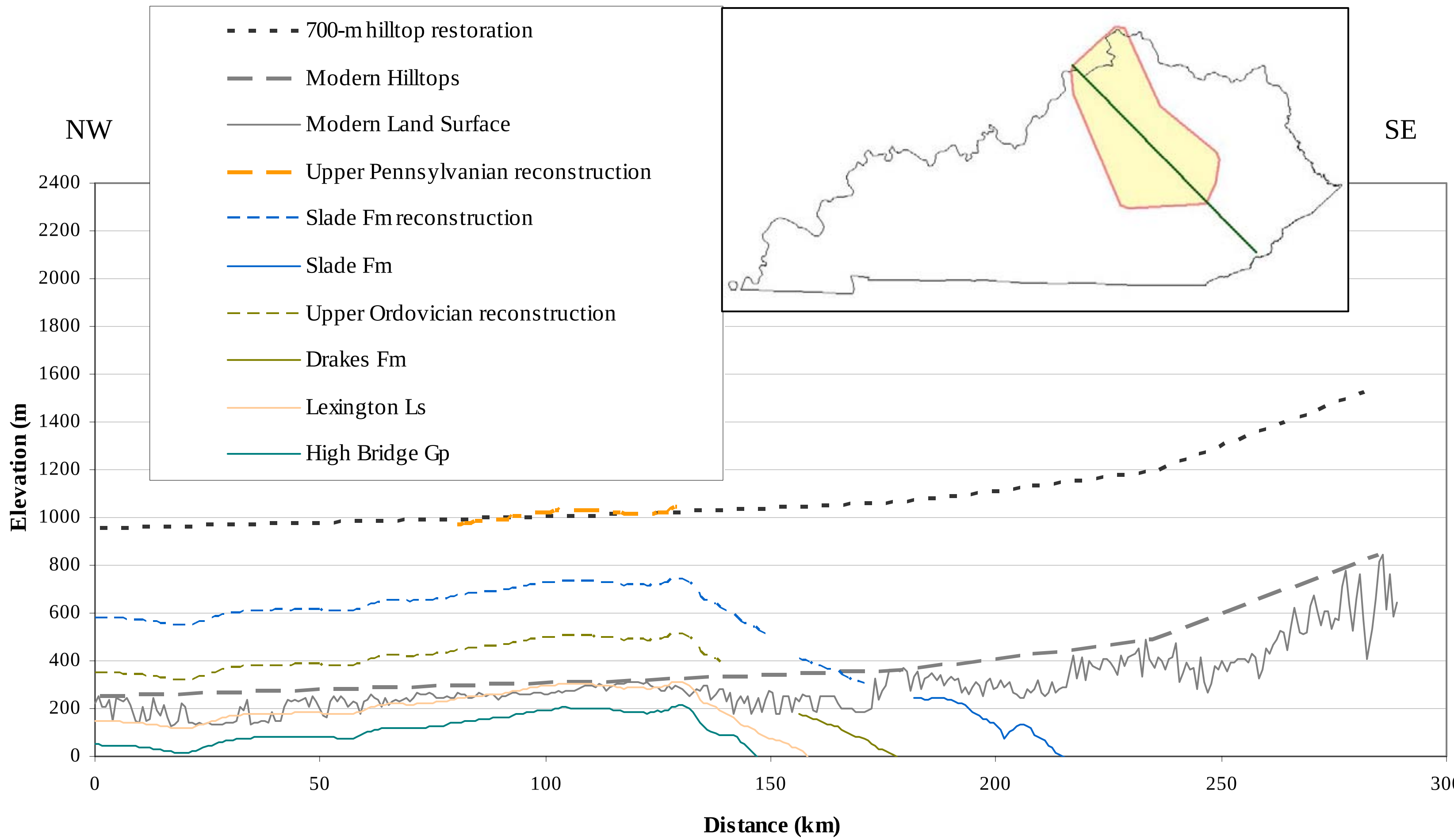


Figure 93. Overburden estimate using extrapolation of stratigraphic thickness and assuming stratigraphic units do not decrease in thickness toward the Cincinnati arch. Lines denote tops, or projected tops, of stratigraphic units.

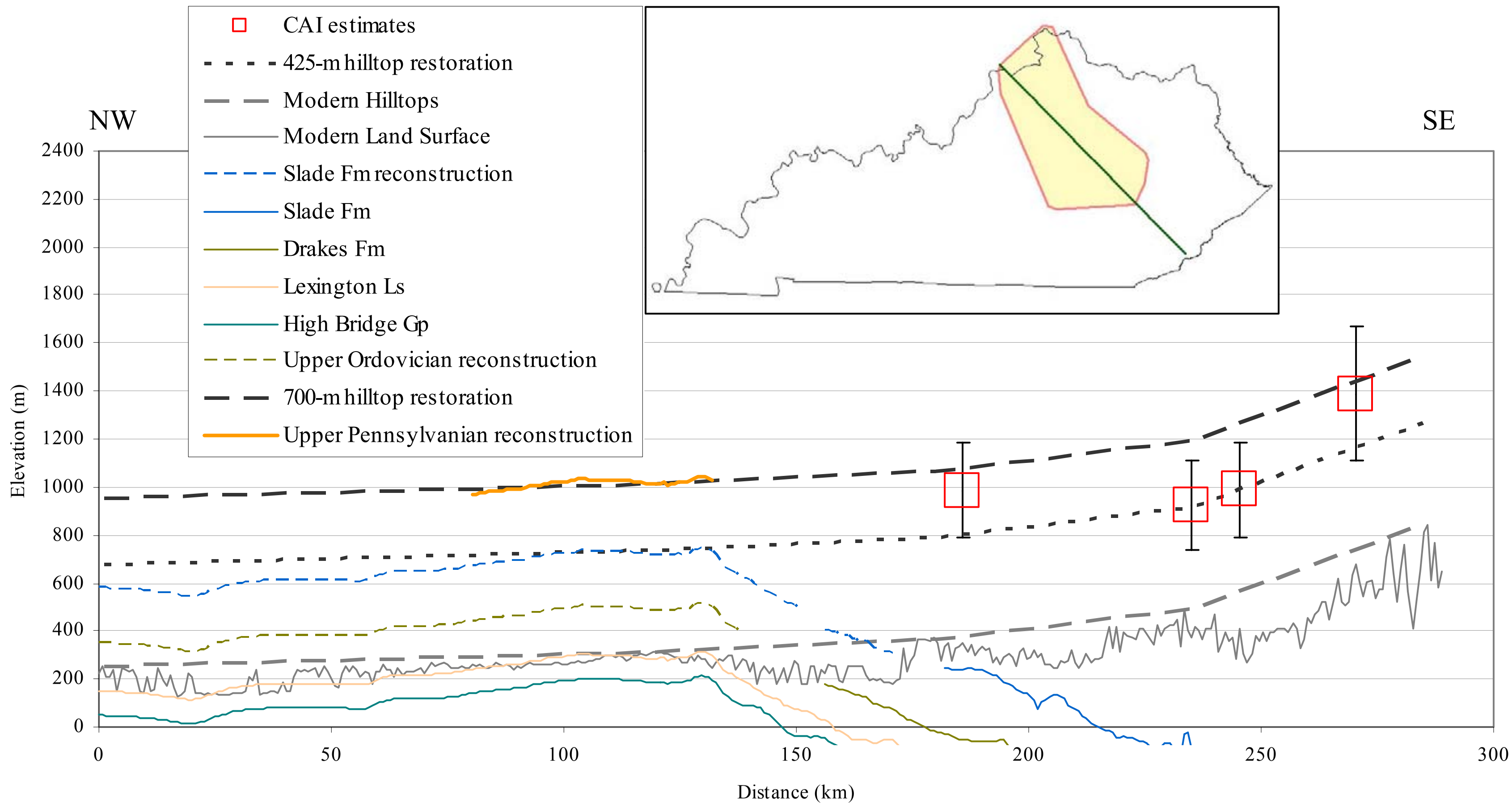


Figure 94. Comparison of overburden estimates based on stratigraphic reconstruction with overburden estimates from conodont alteration studies (Epstein and others, 1977; Harris and others, 1978; Ryder, personal communication, 2004) and supported by thermal maturity studies on eastern Kentucky coal beds. CAI values are estimated overburden following Epstein and others (1977) with error bars representing overburden uncertainty of 20%. CAI overburden estimates are measured from Ordovician and Mississippian strata.

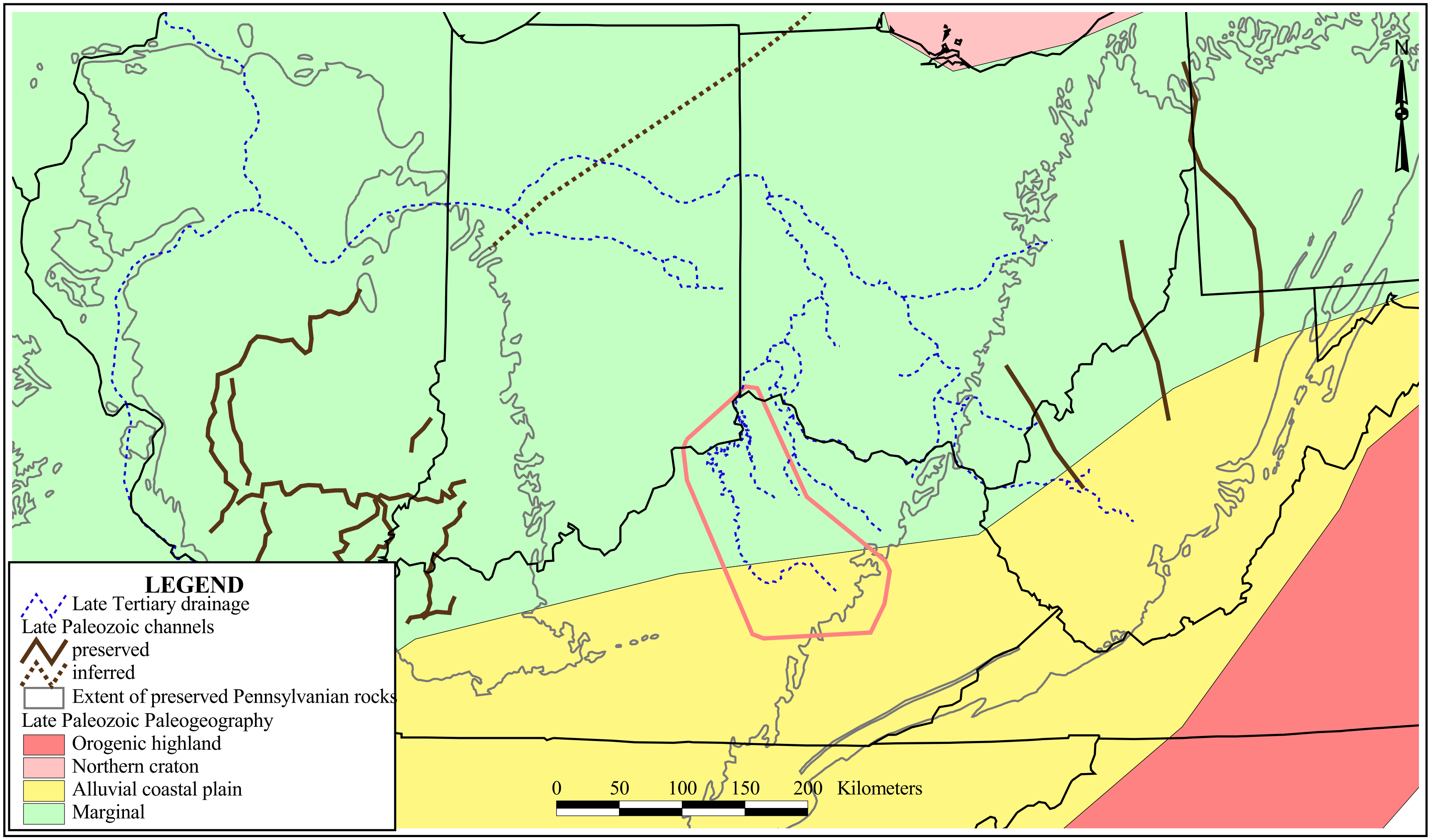


Figure 95. Paleogeography and paleodrainage during late Paleozoic time, references in text.

transport direction from southeast to northwest would have produced a depositional surface graded toward the northeast across the study area.

Assuming deposition ceased and erosion began in the latest Pennsylvanian, the landscape has evolved to its present configuration during approximately 285 m.y. The 700-m reconstruction suggests a long-term average landscape erosion rate of roughly 2.45 m/my, whereas the 425-meter reconstruction would imply a long-term average landscape erosion rate of approximately 1.5 m/my. These rates are consistent with the 2 to 7 m/my estimate of Granger and others (2001) for the Mammoth Cave area.

Fission track data analyze crystal-scale damage from the decay of naturally occurring radioactive isotopes to estimate the thermal history of a sample. The tracks created by the release of alpha particles are healed at selected temperatures. For a known concentration of radioactive isotopes, a statistical analysis of track lengths and frequency in a crystal can provide an estimate of the thermal history of the sample. Roden and others (1993) and Boettcher and Millikin (1994) have produced fission-track thermal histories in the study area (Figure 96). Roden and others (1993) considered data from zircons and apatites in Ordovician bentonites along the crest of the Cincinnati arch; they inferred that cooling has gradually accelerated since 250 Ma. Boettcher and Milliken (1994) examined fission tracks from Pennsylvanian sandstones near Pine Mountain in eastern Kentucky. They inferred a more complex thermal history, with cooling from 140° beginning at approximately 170 Ma. The cooling curve shows a slower cooling rate between 80 and 30 Ma, and then accelerating cooling from 30 Ma until the present.

Estimates of overburden thickness from fission-track data require assumptions about the thermal gradient of the study area. O'Hara and others (1990) estimated Pennsylvanian thermal gradients as large as 60°/km on the basis of vitrinite reflectance data. Price (1978) measured modern geothermal gradients of approximately 15°/km. Therefore, the thermal gradient of the study area has likely changed significantly through time, making quantitative overburden estimates difficult from the available fission track data.

Implications for Landscape Evolution

Although most landscape reconstructions in central Kentucky begin with an inferred late Tertiary regional erosional peneplain (Fenneman, 1939; Jillson, 1930, 1945a, 1963; Thornbury, 1965; Straw, 1968; and Warwick, 1985), the existence of a peneplain is an interpretation derived

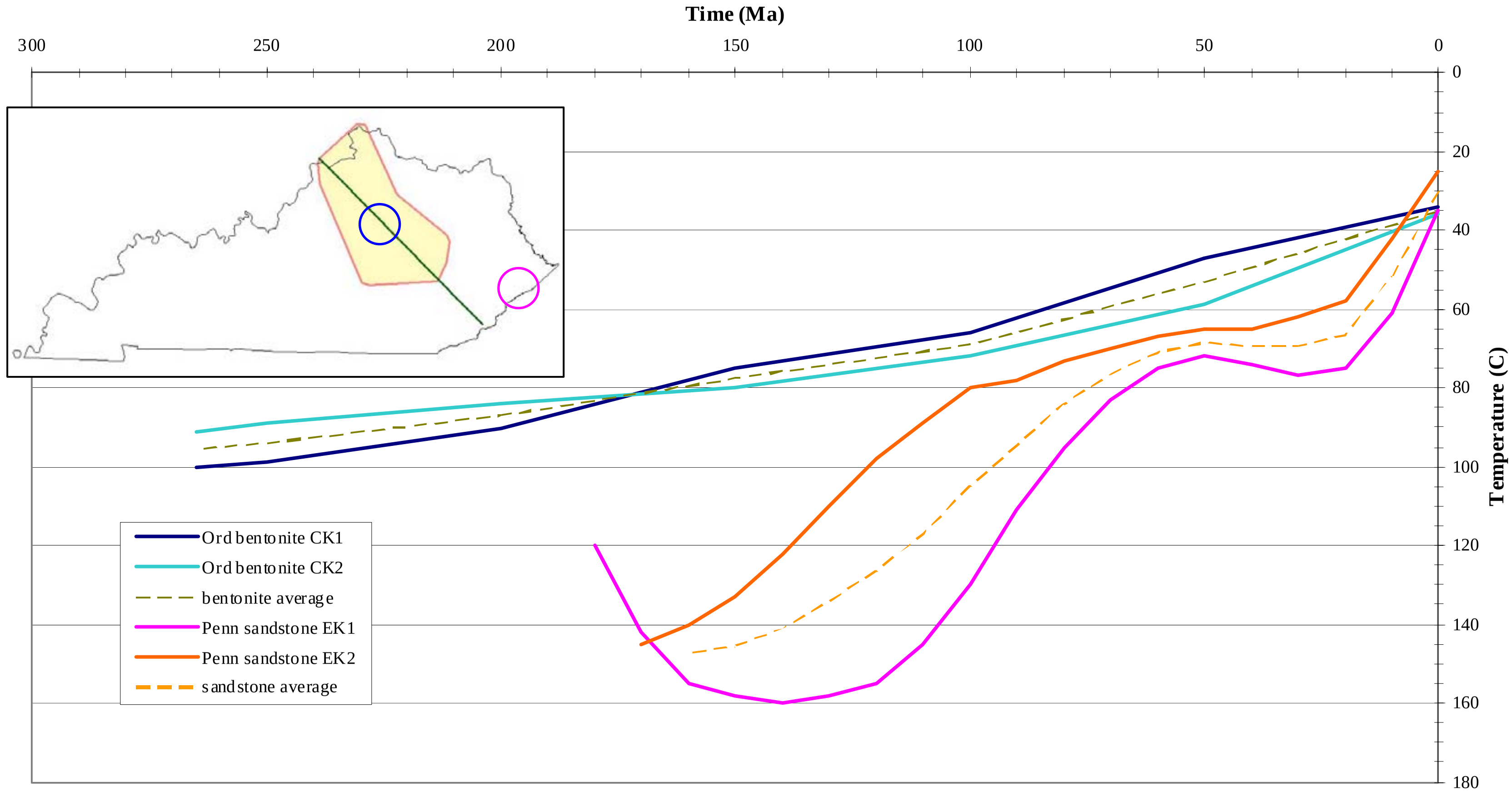


Figure 96. Thermal histories interpreted from fission-track studies on zircons and apatites in Ordovician bentonites (blue circle, Roden and others, 1994) and apatites in Pennsylvanian sandstones (pink circle, Boettcher and Millikin, 1995).

from hypothetical landscape-evolution models and is not based upon empirical data. A more temporally distant, but empirically justified geomorphic datum would be the transition from deposition along a fluvio-deltaic coastal plain during much of the Pennsylvanian to a transportational and subsequent erosional regime during the Late Pennsylvanian or Early Permian. Details of the surface are impossible to determine, but the general pattern of an overfilled foreland basin graded to the northwest and transporting sediment into the Illinois basin provides a general conceptual model for the starting point of inferring landscape evolution trends in central Kentucky. This northwest-graded landscape surface has been progressively eroded since the late Paleozoic or early Mesozoic; accelerated erosion began in the Oligocene or Miocene (Hulver, 1997).

CHAPTER FOUR: IMPLICATIONS AND DISCUSSION

Relative Chronology

This study has undertaken to develop the framework within which subsequent targeted geochronological studies could be conducted. Limited geochronological studies have been conducted in the area, as discussed in Chapter 3, and these data are used to constrain the general timing and rates of events. The data available for interpreting the geologic history of the basin decrease steadily with age. With each progressive step back in time, fewer data and clues are available to reconstruct the ancient landscapes. On the basis of the 1.5 to 1.74 Ma age estimates for the Irvine Formation, which is mapped near the upland summits, only the last 2 m.y. of fluvial history is recorded in the existing profiles and landscape. However, a speculative reconstruction of the Late Pennsylvanian to Permian landscape across the study area can be reasonably inferred from bedrock geology and thermal maturity data.

Early Depositional Slope

During much of Pennsylvanian time, the central Appalachian basin was an active foreland basin, receiving sediment from the rapidly rising and eroding Alleghanian orogeny (Donaldson and others, 1985; Chesnut, 1994; Heckel, 1995). A fluvio-deltaic coastal plain developed with rivers carrying sediment northwestward away from the active mountain belt. These sediments spilled across the Cincinnati arch, contributing thinner deposits to the Illinois basin (Greb and others, 2002). As the Appalachian basin filled and depocenters migrated northward toward the Dunkard basin in Late Pennsylvanian and Early Permian time (Donaldson and others, 1985; Greb and others, 2002), the study area became less of a depositional center and more of a transportation zone, where rivers flowed across the young coastal plain, carrying sediment to the Illinois basin (Figure 95).

The Pennsylvanian-Permian land surface has long since been eroded away, and thus all evidence that could be used to directly reconstruct that surface is lost. Nevertheless, some general constraints can be reasonably inferred from the depositional environments represented in the Appalachian and Illinois basins. Whereas the surface of the coastal plain had been recently deposited by active river systems, the entire land surface should have mimicked, in general, something akin to the ideal equilibrium profile of a river system. The absolute elevation of this former land surface above the modern landscape is not known, but the landscape reconstruction described in Chapter 3 helps to constrain the range of possibilities. A liberal stratigraphic reconstruction and maximum estimates from thermal maturity indices, such as conodont alteration, supported by vitrinite reflectance, suggest that a maximum of 700 meters of sediment has been removed from the crest of the Cincinnati arch. The bulk of the thermal maturity data, however, suggests the amount of removed overburden could be as little as 450 to 500 meters (Figure 94). The limited amount of overburden implies very slow average landscape erosion rates since the Permian.

Early to Mid Tertiary Record Absent

The eroded interval between late Paleozoic and late Tertiary time offers no direct evidence of past fluvial profiles or landscape evolution rates or conditions. The only clues to the conditions existing during this time come from unroofing and denudation studies from fission-track dating and similar techniques. These studies suggest slow and steady erosion since the Permian (Roden and others, 1993) or initially slow erosion, with a significant increase in erosion rate during mid-Tertiary time (Boetcher and Millikin, 1994; Hulver, 1997).

Numerous hypotheses put forward by previous workers (e.g. Jillson, 1963; Luft, 1986) require fluvial profiles that project well above the modern land surface, and are thus more than likely untestable. Luft's (1986) possible connection between the South Fork of the Licking River and the Kentucky River projects above the modern land surface in the Kentucky River basin. The profiles on the Licking River side closest to the Kentucky River divide show a distinct steepening, suggesting that this is the head of that particular fluvial profile. The divide is the highest part of the modern landscape, and any deposits representing the connection between the two streams, or defining an ancient longitudinal profile of the connection, would predate any features currently observable in the landscape.

Jillson's (1963) "Mesozoic course" of the Kentucky River has only three possible intact valley segments: the Colby, Wyandotte, and Cedar gaps (Figure 74). The distribution of quartz pebbles found across Jillson's central Kentucky profile for this stream shows a distinct convex up profile, ranging over nearly 100 m of relief. The sites range higher in elevation in a "downstream" direction as inferred by Jillson (1963) as far as Wyandotte gap; the elevations are progressively lower "downstream" from Wyandotte gap. If the quartz-pebble locations do indeed mark a paleocourse of the Kentucky River, the course significantly predates any features in the modern landscape, and the pebbles in the deposits have been let down with significant bedrock eroded from beneath them. As the topography has been inferred to have been shaped by significant fluvial erosion of bedrock, broad distribution of such long-lived residual deposits is unlikely.

Early Glaciations

The first indication of major climatic cooling and the first record of continental glaciation in the Midwestern states are recorded between 2.5 and 2.0 Ma (Hooghiemstra, 1984; Groot, 1991). This first cold climatic interval would have dramatically affected vegetation. The long preceding warm climates should have led to very mature vegetational communities, wherein only a few species tolerant of dramatic climate shifts may have been present. If so, this first glacial stage would have led to decreased vegetation cover, and increased erosion, as local ecosystems adapted to colder climates with the advance of glaciers. This would in turn have resulted in increased erosion of surficial residuum, and mobilization of sediment long held within mature soils. This could potentially explain the sudden flux of brown-patina chert found in the Lafayette Gravel and similar deposits—Irvine Formation, Old Kentucky River gravels, Ohio River Formation, Luce Gravel, etc—around the Ohio River valley.

The origin of the Teays River system (Tight, 1903) is unclear. Gray (1991) speculated that an early glaciation formed the Teays system, reorganizing an older system flowing to the Great Lakes area. This hypothesis remains untested. An alternative hypothesis, similarly untested, is that the Teays drainage is a remnant of older drainage systems that parallel the late Paleozoic drainage systems suggested by bedrock sandstone trends. Either way, the Teays River system was well developed enough to be significantly entrenched into the surrounding bedrock (Gray, 1982, 1991; Ohio Division of Geological Survey, 2003) before it was disrupted by a later glaciation (Swadley, 1971).

The age of the glaciation that disrupted the Teays River was poorly constrained before recent cosmogenic isotope dating of fluvial deposits (Granger, personal communication, 2001). The till of the disrupting glaciation is associated with reversed-polarity lacustrine sediment (inferred to be >0.78 Ma), whereas younger sediments filling the Teays River valley, including 6 younger tills and associated deposits, show normal magnetization (<0.78 Ma). Similar reversed-polarity lacustrine sediment in the unglaciated Teays valley in West Virginia, the Minford Silt, overlies fluvial sediment that was cosmogenically dated at 1.3 Ma (Granger, personal communication, 2001), which falls within the time period of the Matuyama reversed polarity epoch. Comparable dates of 1.36 to 1.43 Ma (Granger, personal communication, 2001) for fluvial deposits in the Old Kentucky River system that are below lacustrine and outwash deposits also fall within the Matuyama epoch, which ended at approximately 0.78 Ma. These data combined suggest that the glacier that blocked and diverted the Teays River did so sometime between 1.3 and 0.78 Ma.

A total of seven tills have been identified in the Teays Valley, and five tills have been identified just south of the Wisconsin glacial limit in Decatur County, Indiana. The evidence from Luft (1980), Ettensohn (1974), Swadley (1971), Teller (1970), and this study all support the interpretation of only two major episodes of glaciation in northern Kentucky, the Illinoisan and one pre-Illinoisan advance. The two pre-Illinoisan tills identified by Leighton and Ray (1965) in northern Kentucky probably represent two temporally related advances of ice (Teller, 1970) because there is no significant and reliable geomorphic distinction between the two. The joint KGS-USGS mapping program recognized only one unit of pre-Illinoisan drift in the area. If so, only the oldest pre-Illinoisan till identified in Decatur County (located south of the buried Teays valley) advanced to the northern Kentucky area, and subsequent pre-Illinoisan ice was limited to the abandoned Teays valley and points north.

Spillways

A paleochannel has been identified extending from the Kentucky River valley to a divide between the Kentucky and Licking River basins, and is interpreted here as a spillway that was active during times when the Licking River was impounded by glacial ice. The spillway has two outlets: one near the mouth of Upper Howards Creek along the Kentucky River at an elevation of 240 m, and another emptying into the Red River valley near Clay City at an elevation of 228 m (Figure 64). Both outlets of the spillway are geomorphically lower, and thus younger, than the Irvine Formation, which has an estimated age of 1.5 to 1.74(?) Ma (Granger and Smith, 2000; Granger, personal communication, 2001).

The spillway outlets are at the upstream end of the "floor" of most fluvial features for the Kentucky River valley. The upper outlet is on one end of the projected floor, but the lower is also relatively close to the profile trend of the "floor" (Figure 67). The downstream end of the "floor" coincides with the Old Kentucky River deposits near Carrollton, which are overlain by glacial deposits (Swadley, 1971). The glacial impoundment that created the original spillway and that is represented by the upper spillway outlet most likely corresponds in age to the glacial deposits found in the Old Kentucky River valley downstream, on the basis of their coincidence in the longitudinal profile of the Kentucky River and on the proximity of the mouth of the Licking River valley to the Old Kentucky River deposits. The lower spillway outlet might represent a later activation of the spillway during the same glacial episode, but this is not testable without more detailed geochronologic data. The lower spillway outlet is a hanging valley more than 50 m above the modern Red River valley. On the basis of the distribution of Illinoisan and pre-Illinoisan glacial deposits, the Licking River valley would have been impounded, and the spillway thus active, during both phases of glaciation.

The upper Salt River above Lawrenceburg, Kentucky, was inferred to have originally been a tributary of the Old Kentucky River and to have been captured by the lower Salt River (Leverett, 1929), and this interpretation is supported by this study (see discussion in Chapter 2). The elevation of the modern divide suggests that the capture occurred when the upper Salt River was at an elevation of about 242 m (795 ft). This is approximately 30 m above the floor of fluvial deposits in this stretch of the Kentucky River valley. However, if the downstream along-tributary distance to the main valley is considered, coupled with the above-knickpoint slope of modern tributaries, the capture elevation is within 10 m of the estimated floor of fluvial deposits (Figure 97). The timing of the capture is not firmly constrained, but on the basis of the position in the tributary profile, it most likely occurred shortly prior to, if not contemporaneous with, the glacial impoundment of the Kentucky River represented by the fluvial-lacustrine-glacial transition of sediments in the Old Kentucky River valley near Carrollton. Karst capture, analogous to the modern Crawford Spring karst groundwater basin in Boyle County (Paylor, personal communication, 2001), may have assisted in the capture process.

Considering the significant deposit of Illinoisan drift at Carrollton (Figure 54), it is a reasonable hypothesis that the Kentucky River was impounded at least briefly during the Illinoisan. The lacustrine sediment in the Old Kentucky River valley near Carrollton clearly suggests that the valley was significantly impounded during the pre-Illinoisan glacial advance into the area. For both Illinoisan and pre-Illinoisan lakes, the impounded water would have had

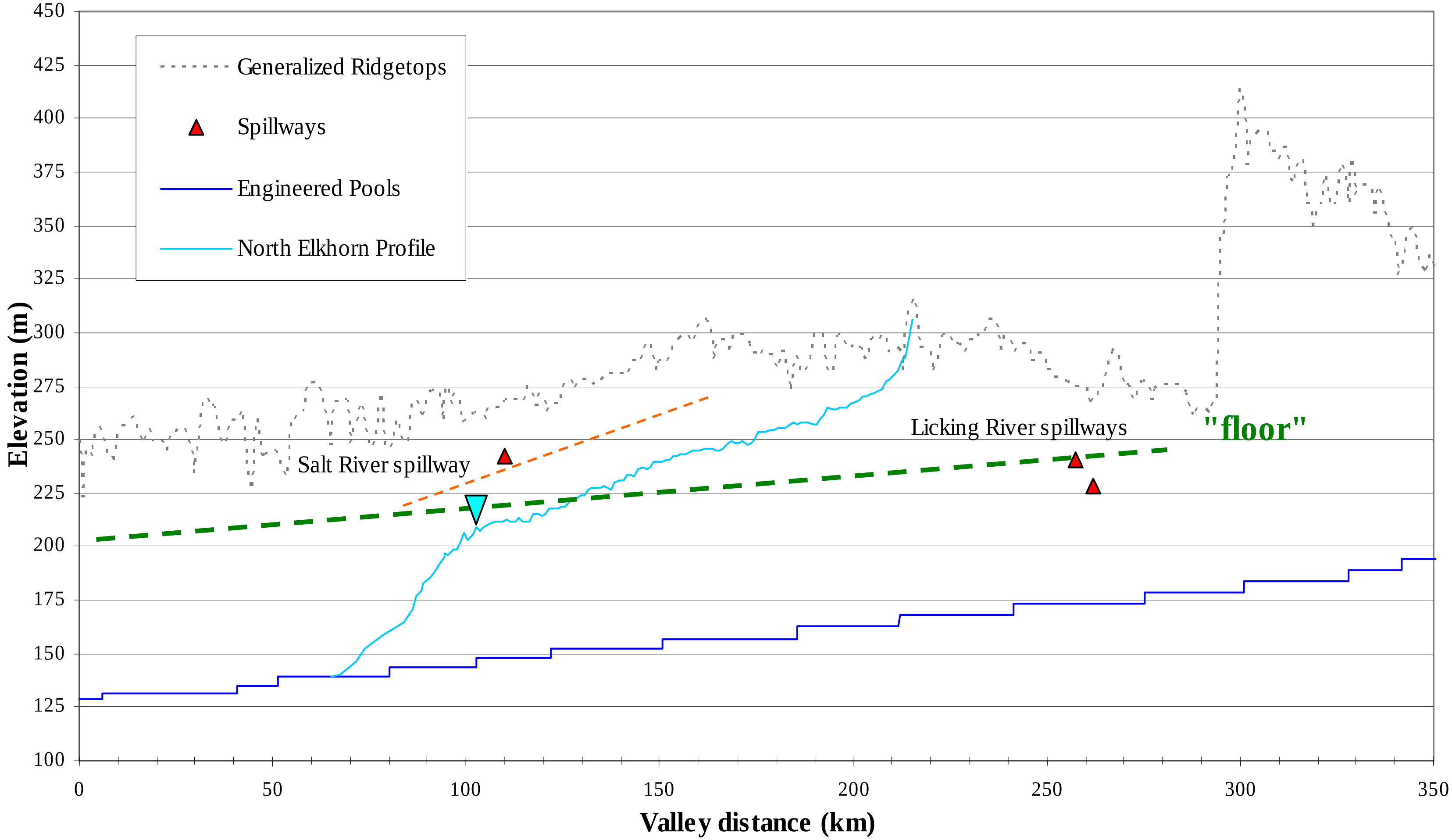


Figure 97. Comparison of Salt River spillway elevation with the “floor” of fluvial deposits (dashed green line) and the projected tributary grade (dashed orange line). Projected tributary grade is parallel to gradient of major tributary streams (e.g. North Elkhorn Creek) above the knickpoints discussed in Chapter 2 (light blue triangle).

to find an escape route into the Salt River valley or the Green River valley. The lowest point along the southwestern divide of the Kentucky River valley is where the upper Salt River makes a sharp bend to the west south of Lawrenceburg. The divide between the modern Kentucky and Salt River valleys is approximately 242 m (795 ft) at this point. Any impoundment of the Kentucky River valley would likely have drained across the divide at this point, and been released through the lower Salt River valley. This is approximately 20 meters above the “floor” of fluvial deposits in the area.

Other spillways are identifiable near the glacial fronts in northern Kentucky (Figure 67). The Licking River spillway near Walton is higher in elevation than that in Montgomery County, and would have been activated only after glacial loading had induced sufficient crustal flexure to tilt the landscape toward the ice and open flow across the downstream spillway. No geomorphic outlets for the spillway are preserved in the Kope Formation on the Kentucky River side of the Walton spillway. Another spillway is inferred at the headwaters of the Little Kentucky River in Trimble County at an elevation of approximately 256 m (840 ft). This spillway would have also been activated only after significant tilting, and also lacks a clear downstream outlet.

Kentucky River Incision

Bedrock erosion logically requires the absence of fluvial deposits between the river and the bedrock it is eroding. Thus, an actively or aggressively eroding bedrock stream is unlikely to leave behind extensive fluvial deposits. The distinct “floor” of fluvial features in the profile of the Kentucky River valley suggests an abrupt change in behavior of the river. Above the floor, a diverse array of paleochannels, fluvial terrace deposits, and bedrock benches is preserved (Figure 67). No distinct, through-going valley profiles are readily traceable in the areas above the floor. Profile trends of paleochannels, fluvial deposits, and bedrock benches appear to closely parallel stratigraphic contacts, suggesting lithologic control of origin and preservation. Below the floor, the fluvial record is locally dominated by stratigraphically limited bedrock benches with few or no mapped fluvial deposits (Figure 67).

The lack of discreet, traceable, graded profiles above the “floor” suggests that the river has undergone slow, steady erosion with no prolonged hiatuses while fluvial deposits and paleochannels above the floor developed. Adequate lateral erosion existed to allow for deposition and preservation of fluvial deposits, but enough vertical erosion was active to continually incise the river valley without producing unique graded profiles of fluvial features. The lack of fluvial deposits and presence of only limited bedrock benches, below the "floor" suggests that vertical bedrock erosion was the dominant process, and sediment was quickly swept from the system and not preserved as fluvial deposits or terraces. The bedrock benches below the floor represent limited lateral erosion in the Gratz and Palmer morphology styles which are optimum for (1) developing and (2) preserving those features. The "floor" of the fluvial deposits, thus, represents a change in behavior of the river from slow, steady vertical-and-lateral erosion to rapid vertical incision.

At the downstream end of the "floor," fluvial deposits have been mapped by Swadley (1971) in meandering Old Kentucky River paleovalleys with a northward graded bedrock strath. Granger (personal communication, 2001) dated the Old Kentucky River fluvial deposits near Carrollton at 1.36 to 1.43 Ma. These southerly derived fluvial sediments are overlain by lacustrine deposits, which in turn are overlain by northerly derived glacial outwash and drift. This sequence of deposits is comparable to those found in the Teays River and Old Licking River valleys. These sedimentary sequences represent the youngest vestiges of the Teays River system, and represent the sequence of events coeval with the abandonment of the Teays and integration of the early Ohio River. The upstream end of the "floor" corresponds to the older spillway outlet along Upper Howards Creek. The spillway represents spillover of glacially impounded water from the Licking River valley. The coincidence of the two glacially related events along the "floor" at opposite ends of the valley profile supports the hypothesis that the two events are coeval and genetically related. The projection of the Salt River capture near Lawrenceburg to very near the "floor" suggests that the capture and overflow from the Kentucky River into the Salt River valley was contemporaneous.

The "floor" of fluvial deposits in the profile of the Kentucky River valley, thus, probably represents the approximate profile of the river just prior to the integration of the early Ohio River, and the abandonment of the Teays River system. The new drainage organization provided a much shorter flow path to the vicinity of Cairo, Illinois (Figure 79). The flow from Carrollton through the Teays system to Cairo was roughly 1200 km, whereas the route down the early Ohio was only about 600 km. The level of drainage systems at Cairo remained unchanged, while the route of Kentucky River flow to that point was shortened by roughly half the distance. As discussed above, the "floor" of fluvial features represents the onset of rapid bedrock incision of the Kentucky River. The “floor” is associated with glacial deposits that are inferred to correspond to the disruption of the Teays River and the organization of the early Ohio River. When the Kentucky River flow was rerouted from the Teays River system into the newly formed Ohio River system, the elevation difference to Cairo was the same, but the distance was cut approximately in half (Figure 80). The profile of the modern Ohio River and the modern Kentucky River, as measured from Cairo, closely parallels the profile of the Teays River measured from the same point. The Kentucky River incision corresponds very well with the 80 to 100 meters of incision required to adjust the Old Kentucky River to a position 600 km closer to the Cairo along the Teays River valley. Thus, the Kentucky River is inferred to have incised to its currently level in response to the drainage reorganization of the Teays River system into the Ohio River system. The available geochronology does not constrain the rate of incision, or whether a knickpoint migrated up the river. If the younger Licking River spillway, which is 12 meters lower than the older one, represents a second phase of the same pre-Illinoisan glaciation that reorganized the regional drainage, it suggests rapid incision within a comparable glacial cycle. Such very rapid stepped incisions are compatible to the punctuated incision Granger and others (2001) identified at Mammoth Cave (Figure 84).

Illinoisan and Wisconsin Alluviation

Illinoisan drift has been identified along the flanks of the modern Ohio River valley, suggesting that most of the modern bedrock valley morphology was in place by the onset of Illinoisan glaciation roughly 180 ka. The Illinoisan drift plug at the mouth of the Kentucky River near Carrollton, and the Illinoisan drift overriding the pre-Illinoisan in the upland nearby suggests that a substantial thickness of Illinoisan ice reached at least as far south as the mouth of the Kentucky River valley. The convex and irregular profile of the buried bedrock valley of the Kentucky River downstream from Frankfort suggests that bedrock incision had not yet reached an ideal concave profile, and thus was still active when it was interrupted by alluviation associated with the Illinoisan glaciation. A series of valley-bottom terraces above the active flood zone, well up-valley from the better-documented Wisconsin terraces (Figure 58), may illustrate lacustrine deposition associated with Illinoisan impoundment of the Kentucky River valley. On the basis of the distribution of Ohio River valley deposits assigned to the Illinoisan, the Illinoisan glaciation would most likely have reactivated the spillways between the Licking, Kentucky, and

Salt River basins by impounding the flow from those basins. Geochronologic data do not currently exist to confirm or deny this.

Wisconsin glaciers did not reach northern Kentucky, but their effects have been noted in the lower reaches of the Kentucky River valley (Kane, 1972). The Wisconsin glaciation produced a sediment-choked outwash valley-train deposit in the Ohio River valley. The high flow volumes and rapid alluviation associated with this valley train led to shallow impoundment of tributary valleys. The maximum elevation of Wisconsin terrace deposits in Ohio River valley near Carrollton is approximately 152 m (500 ft). Spillover of high-suspended-load water from the outwash flows into the mouths of tributary valleys led to the deposits of ubiquitous silty clays, commonly refered to as “blue clay,” in the bottoms of most tributary valleys (Ray, 1974; Kane, 1972). A series of terraces at 149 to 152 m (490 to 500 ft) in elevation extends up the Kentucky River valley as far as the mouth of Elkhorn Creek, where they merge with the active terraces of the modern river. The levels of the tributary-mouth lakes thus formed would have fluctuated with the magnitude of outwash flows in the main Ohio River valley. Although less data exist for the Illinoisan, the older glaciation should have had similar effects when the ice was not at its maximum extent and Illinoisan outwash was flowing down the Ohio River valley.

Reconciliation of Erosion Rate Estimates

The incision history inferred by this study and the limited geochronological data available provide limits on erosion rates and rates of landscape evolution in the study area. The estimated overburden across the Cincinnati arch (450 to 700 m) has been removed since the end of deposition of the Alleghanian clastic wedge in eastern Kentucky, which occurred 250 to 290 Ma. This results in an average upland erosion rate of 1.6 to 2.8 m/m.y. for the study area since late Paleozoic time. Integrating regional fluvial incision of 50 to 150 m into this calculation suggests long-term average fluvial incision rates of 1.7 to 3.4 m/m.y. over the same time period.

The observations of this study, however, are limited primarily to Plio-Pleistocene deposits, on the basis of the limited cosmogenic isotope dating available for the study area. The published and unpublished dates (Granger and Smith, 2000; Granger, personal communication, 2001) for the Irvine Formation at Rice Station (ele. 275 m) range from 1.8 to 1.5 Ma, and the burial ages of the Old Kentucky River fluvial sediments near Carson range from 1.3 to 1.45 Ma (Granger, personal communication, 2001). The valley is inferred to have incised 90 to 100 m after deposition of the Old Kentucky River sediment and prior to the onset of Illinoisan glaciation (0.15 to 0.18 Ma). This yields average fluvial incision rates of 69 to 98 m/m.y. during most of the Pleistocene at Carson. At Rice Station, a cosmogenic production model age of 1.5 to 1.8 m.y. since deposition of the Irvine Formation, with 110 to 120 m of bedrock erosion prior to the Illinoisan glaciation (Figure 86). This yields average fluvial erosion rates of 67 to 91 m/m.y. at Rice Station, strikingly comparable to the rates estimated downstream at Carson.

The projected grade of the Old Kentucky River deposits (the "floor" of fluvial features, projected from the Carson site through the upper/older spillway outlet at Upper Howards Creek) is 30 to 35 m below the bedrock strath of the Rice Station deposit (Figure 86). The uncertainties in age estimates (1.8 to 1.5 Ma for Rice Station, 1.3 to 1.45 Ma for Carson and the "floor") yield estimated average fluvial erosion rates of 50 to 700 m/m.y. The highest values are extreme in the context of this study and other incision estimates in the region, and are considered unrealistic; the lower values are slightly lower than the estimates for the entire valley. More precise age estimates are thus necessary to constrain Late Pliocene to Early Pleistocene (1.8 to 1.2 Ma) erosion rates in the study area.

Whereas the preservation of fluvial features and deposits suggests a sudden and dramatic shift in erosion rate/style at 1.3 to 1.45 Ma, the limited and low-resolution geochronologic data currently available do not conclusively resolve any change in erosion rates within fluvial deposits observed in this study. However, the difference in estimated fluvial erosion rates since 1.8 Ma (67 to 98 m/m.y.) versus since 250 Ma (1.7 to 3.4 m/m.y.) suggest that erosion rates were not constant through time in this area. Restoration of the estimated removed overburden (450 to 700 m) with the estimated Plio-Pleistocene erosion rates (67 to 98 m/m.y.) suggests that fluvial erosion could have produced the observed denudation and incision in 4.6 to 12.7 m.y. Fission-track dating of eastern Kentucky sandstones suggests that uplift/denudation in the region has not been uniform through time, and that an acceleration of uplift/denudation began at 50 to 30 Ma (Figure 79) (Boetcher and Milliken, 1994; Hulver, 1997). The presence of Cretaceous through Eocene (ca. 70 to 45 Ma) marine and marine-marginal sediment in far western Kentucky (Olive, 1980) suggests that the region had not undergone significant regional uplift until after the Eocene. If uplift and major fluvial incision in the Kentucky River valley began only at 30 to 50 Ma, then erosion rates over that time period would be 10 to 28 m/m.y. A plausible hypothesis that remains to be tested, but is beyond the scope of this study, is that fluvial erosion rates remained low (<10 m/m.y.) from late Paleozoic until mid-Tertiary time, and then accelerated (10 to 100 m/m.y.) to produce the modern landscape. Currently visible topographic features would be no older than Pliocene in age in this scenario.

Controls on Fluvial Evolution

Lithologic Control

This study has documented lithologic control as the dominant control on landform evolution in the Kentucky River valley. Variations in drainage density correspond to different bedrock lithologies (Figure 23). The distribution of abandoned meanders and the morphology of paleochannels correspond to the bedrock geology. Although most knickpoints are not directly associated with bedrock geology, two prominent knickpoints in the Kentucky River valley, one at Lock and Dam No. 4, and one at 105 km on the Red River, correspond to underlying lithology. Contributing drainage area is the dominant factor in determing the distance of knickpoints from the main Kentucky River valley, but that relationship is modified by lithologic control (Figure 33).

This study documented lithologic control on valley morphology, and identified eight distinct valley morphology styles in the study area (Figures 41, 42). These styles correspond to the dominant bedrock lithology in the valley walls, and are controlled by the shale content and bedding thickness of the bedrock. Shale content and bedding thickness control the lateral and vertical erodibility of bedrock units. The distribution of high-level fluvial deposits in the Kentucky River valley corresponds to the valley morphology styles suggested by bedrock stratigraphy.

Relict Landforms and Inheritance

Some of the landforms in the study area represent past events that are no longer active, but even the inherited landforms are dependent on bedrock lithology for preservation. Wisconsin alluviation of the Ohio River resulted in slackwater deposition in the lower Kentucky River valley. A series of terraces near the mouth of the river at approximately 150 m, well above the modern flood zone, represent the top of the lacustrine deposits. Illinoisan ice may have also impounded the river, and led to deposition of terraces above the modern flood zone well upstream. Post-glacial (Holocene) erosion by the stream has not yet reworked or destroyed these deposits.

The paleochannels associated with the high-level fluvial deposits represent abandoned ancient meanders of the river. The development of these high-level abandoned meanders was controlled by bedrock stratigraphy in the same manner as the valley-bottom meanders. Bedrock benches represent fragments of stream-cut straths as the river meandered while incising the valley. The benches are preserved only on lithologies that are sufficiently resistant to lateral erosion to inhibit the undercutting of the benches by subsequent meander migration. These benches are not preserved along sections of the valley dominated by shaly litholgies and the associated colluvial slope processes.

The entrenched, symmetrical meanders in the Wilmore valley morphology style were inherited from overlying valley morphology styles more conducive to the development of broad, rounded meanders, as in the Gratz style (Figure 73). As the river eroded from the Gratz style into the underlying, more resistant lithologies, the meanders were superposed onto the subjacent rocks. The high resistance of the rocks in the Wilmore and Tyrone styles preserves the meander pattern by preventing lateral erosion from modifying or destroying the superposed meanders.

Glaciation and Drainage Reorganization

Glaciation and associated drainage changes had significant and abrupt impacts on the evolution of the Kentucky River. Swadley (1971) documented reversal of flow in the Old Kentucky River associated with a pre-Illinoisan glaciation. This glaciation led to destruction and abandonment of the Teays River system and the establishment of the integrated Ohio River system. This resulted in the abandonment of the Old Kentucky River paleovalleys. The drainage reorganization resulted in a shortening of the longitudinal profile of the Kentucky River, and subsequent incision is related to the adjustment of the Ohio and Kentucky Rivers to the new profile. Most of the high-level fluvial deposits and paleochannels preserved in the Kentucky River valley were formed prior to the onset of rapid adjustment and incision of the Kentucky River. The incision of the river was either uniformly rapid, or accomplished by a knickpoint migrating quickly up the main valley. Knickpoints in tributaries to the Kentucky River do not show any relationship to the distance from the mouth of the Kentucky River that would suggest slow migration of a knickpoint upstream.

The pre-Illinoisan glaciation that destroyed the Teays River also resulted in slackwater impoundment of the Licking River and the rest of the drainage area upstream from the Kentucky River. This extensive lake rapidly filled and spilled over the divide into the Kentucky River valley near Levee in Montogomery County. As the weight of the ice sheet slowly depressed the lithosphere, the distal spillway was abandoned in favor of other shorter lived spillways more proximal to the ice front.

Implications for Landscape Evolution

The primary implication of this study for considerations of landscape evolution in the study area is the dominance of bedrock lithology as a control on the distribution of landform patterns. The successful projection of modern valley morphologies into stratigraphically equivalent sections of the upland suggests that the association of landform patterns with large-scale lithologic units persists through time. Large-scale lithologic units control the evolution of hillslope morphologies, erosion rates, and styles, and thus determine the characteristics of resulting valley morphology style. The concept that bedrock lithology controls landform development is by no means new, but this study affirms the pervasive—but not exclusive—dominance of bedrock lithology in control of central Kentucky landscape evolution. Inheritance of older forms is a significant factor in the distribution of selected landforms or morphologies, but even the inheritance is an indirect reflection of bedrock lithology. The inherited landforms are preserved because of modification-limiting characteristics of the underlying bedrock, such as resistance to lateral erosion.

The recognition of bedrock-controlled valley-morphology styles in a region of reasonably well-documented stratigraphy conceptually allows for a temporal and spatial reconstruction of landform patterns. By examining active landform processes along the profile of the modern stream, one can infer the progression of landform patterns through time. The along-stream progression of landforms would represent a series of landscape snapshots through time for a given point along the river. For example, an examination of karst landforms from the Worthville area to the Camp Nelson area potentially could illustrate the progressive development of karst landforms in the Palisades.

The distribution and preservation of upland fluvial features supports the projection of valley morphology styles into the upland on the basis of enclosing stratigraphy. Preserved fluvial features and deposits are consistent with modern analogs elsewhere in the valley in comparable stratigraphic positions. This suggests that the deposits formed and were preserved with enclosing valley morphologies and relief, similar to corresponding modern valley analogs. Numerous fluvial features and deposits in the study area are preserved just below the level of the modern ridgetops, suggesting projection of restored stratigraphy well above the surface of the modern hilltops (Figure 67).

The reconstructed stratigraphy of the study area, supported by physical extrapolation, thermal maturity, and fission-track estimates, suggests between 450 and 700 m of overburden has been removed from the study area since late Paleozoic time, most of which probably was removed since late Tertiary time. The late Paleozoic fluvio-deltaic depositional surface remained little changed until mid-Tertiary time, when accelerated erosion and incision began to dissect the ancient landscape. The available fission-track dating studies in the area show relatively slow and uniform unroofing of the area, with accelerated unroofing since mid-Tertiary time. The evidence suggests a long period of Mesozoic and early Cenozoic stability, followed by accelerating erosion which has continued to the present. The erosion rates derived from this study (~60 to 100 m/m.y.), and the limited overburden suggested by stratigraphic reconstruction and published thermal maturity studies, suggest a landscape that has evolved rapidly during late Tertiary and Quaternary time, inherited from an older landscape little changed since late Paleozoic and Mesozoic time.

Summary and Conclusions

This study utilized published geologic and topographic data, as well as field observations and extensive compilation and comparison of digital data, to examine the fluvial record preserved in the Kentucky River valley in central Kentucky. Numerous fluvial features including abandoned paleovalleys, fluvial terraces and deposits, bedrock benches, and relict spillways between adjacent river valleys were identified during the course of the study.

This study developed a relative chronology for the evolution of the Kentucky River during the Pliocene and Pleistocene. The oldest preserved fluvial deposits, located near the modern upland summits, are approximately 2 My old or late Pliocene in age. The record of deposits and paleovalleys suggests slow steady erosion until an Early Pleistocene glacial advance. At that time, a glacier blocked the flow of the Kentucky and Licking Rivers into the

Teays River system, causing the flow to be incorporated into the newly reorganized Ohio River system. The Kentucky River incised its valley to accommodate the shorter flow distance produced by the drainage reorganization. The Illinoisan and Wisconsin glaciations caused impoundment of the Kentucky River and deposition of fine-grained slackwater sediment in the valley bottom. Holocene responses of the river to human activity were beyond the scope of this study.

The morphology of the modern valley was examined, and eight separate zones of distinct geomorphic style were delineated along the modern Kentucky River valley. Bedrock lithology is the dominant control on valley morphology and, thus, on the distribution and preservation of fluvial deposits and features in the study area. Rock units vary in susceptibility to vertical and lateral erosion and in control on the style and relative dominance of colluvial processes active along the valley walls.

Because valley morphology coincides with bedrock lithology, the modern valley morphology styles can be projected into the upland on the basis of similar bedrock stratigraphy. Modern and upland fluvial deposits and valleys in comparable lithologic contexts have developed comparable morphologies. The application of bedrock-controlled valley morphology can be a useful tool in resolving landscape evolution questions in this area.

Some stream trends are inherited from the late Paleozoic drainage during the Alleghanian orogeny. The original drainage directions persisted through time, being only locally modified by captures and adjustments related to bedrock lithologic contrasts. More recent inheritance of valley morphology has resulted from the erosion of the river from one lithology down into another lithology with differing erosional susceptibility, thus superposing the meander patterns of the overlying valley style onto the underlying lithology.

One major drainage reorganization has been recognized by previous workers, and confirmed in this study. An Early Pleistocene, pre-Illinoisan glacial advance (between 1.3 and 0.8 Ma) disrupted the northward flow of the Old Kentucky River toward the Teays River system. This led to organization of the early Ohio River and a resulting southwestward flow of water from the Kentucky River through the Ohio River system. This greatly reduced the distance to base-level, and led to a change in erosional style for the Kentucky River. Erosion rate may have increased also, but the resolution of available geochronological data is insufficient to precisely quantify rates. Stream captures and paleocourses previously hypothesized by Luft (1980, 1986) and Jillson (1963) are not testable in the current study; the record required to consider these questions has been removed by erosion.

This study inferred limited eroded overburden in the study area, on the basis of stratigraphic extrapolation and published thermal maturity and fission-track data. The limited overburden implies very slow average upland erosion rates since the end of late Paleozoic deposition. Fission-track studies suggest fluvial erosion rates have apparently increased through time, consistent with the results of this study.

The successful projection of valley morphologies on the basis of bedrock stratigraphy, the history of erosion suggested by fission track data and the results of this study, as well as soil thickness and development, all argue against the existence of a mid to late Tertiary, low-relief, regional erosional surface. This study instead hypothesizes that the apparent accordance of ridge-top elevations in the study area is a reflection of a fluvially downwasted late Paleozoic depositional surface.

APPENDIX 1

USGS 7.5-minute Geologic Quadrangle maps utilized for this study

Appendix 1. USGS 7.5-minute Geologic Quadrangle maps utilized for this study.

Quadrangle Name(s)	Author(s)	Year	GQ No.	100K Quad.	KGS Digitizer
ALCORN	Charles L. Rice	1972	GQ-963	Harrodsburg	H. Nelson
ALEXANDRIA	Gibbons, A. B.	1971	GQ-926	Falmouth	A. Harper
AUSTERLITZ	William F. Outerbridge	1975	GQ-1245	Lexington	X. Yang
BEATTYVILLE	Gordon W. Weir, Richard E. Eggleton	1978	GQ-1483	Irvine	T. Sparks, D. Curl
BEDFORD	W.C. Swadley	1977	GQ-1409	Madison	M. Tyra
BEREA	Gordon W. Weir	1967	GQ-649	Harrodsburg	H. Nelson
BERLIN	Stanley J. Luft	1975	GQ-1256	Falmouth	H. Nelson
BERRY	Stanley J. Luft	1975	GQ-1284	Falmouth	Q. Zhang
BETHLEHEM	W.C. Swadley	1977	GQ-1436	Madison	M. Thompson
BIGHILL	G.W. Weir, K.Y. Lee, P.E. Cassity	1971	GQ-900	Harrodsburg	H. Nelson
BOONEVILLE	Gordon W. Weir	1978	GQ-1479	Hazard	W. Andrews., J. Patton
BRECKINRIDGE	Roberts M. Wallace	1976	GQ-1344	Lexington	H. Nelson
BRODHEAD	J.L. Gualtieri	1967	GQ-662	Somerset	M. Murphy
BROOKSVILLE	William F. Outerbridge	1971	GQ-905	Falmouth	H. Nelson
BRYANTSVILLE	Don E. Wolcott, Earle R. Cressman	1971	GQ-945	Harrodsburg	D. Carey, C. Hettinger
BUCKEYE	Don E. Wolcott	1970	GQ-843	Harrodsburg	D. Carey
BURLINGTON-ADDYSTON	A.B. Gibbons	1972	GQ-1025	Cincinnati	S. Cordiviola
BUTLER	Stanley J. Luft	1972	GQ-982	Falmouth	H. Nelson
CAMPBELLSBURG	A.B. Gibbons, W.C. Swadley	1976	GQ-1364	Madison	M. Thompson
CAMPTON	T. Dennis Coskren, Harry P. Hoge	1978	GQ-1502	Irvine	V. Sullivan, D. Curl
CARLISLE	Lawrence U. Blade	1978	GQ-1450	Lexington	C. Hettinger
CARROLLTON	W.C. Swadley	1976	GQ-1281	Madison	M. Thompson
CENTERVILLE	E.R. Cressman, S.P. Kanizay	1967	GQ-653	Lexington	H. Nelson
CLAY CITY	George C. Simmons	1967	GQ-663	Irvine	L. Morris, D. Curl
CLAYSVILLE	Stanley J. Luft	1976	GQ-1341	Falmouth	H. Nelson
CLINTONVILLE	W.C. MacQuown Jr.	1968	GQ-717	Lexington	M. Murphy
COBHILL	Donald C. Haney	1976	GQ-1347	Irvine	M. Murphy, D. Curl
COLETOWN	Douglas F.B. Black	1967	GQ-644	Harrodsburg	H. Nelson

Quadrangle Name(s)	Author(s)	Year	GQ No.	100K Quad.	KGS Digitizer
CORNISHVILLE	E.R. Cressman	1973	GQ-1135	Harrodsburg	M. Thompson
COVINGTON	Stanley J. Luft	1972	GQ-955	Cincinnati	S. Cordiviola
COWCREEK	William F. Outerbridge	1978	GQ-1448	Hazard	T. Sparks, J. Patton
CRAB ORCHARD	J.L. Gualtieri	1967	GQ-571	Somerset	M. Murphy
CYNTHIANA	Roberts M. Wallace	1976	GQ-1333	Lexington	M. Thompson
DANVILLE	E.R. Cressman	1972	GQ-985	Harrodsburg	Q. Zhang
DE MOSSVILLE	Stanley J. Tuft	1970	GQ-862	Falmouth	A. Harper, T. Sparks
DELAPLAIN	Roberts M. Wallace	1977	GQ-1426	Lexington	H. Nelson
ELLISTON	W.C. Swadley	1972	GQ-994	Falmouth	R. Duncan, M. Thompson
EMINENCE	Stanley J. Luft	1977	GQ-1385	Louisville	
EZEL	G.N. Pipiringos, S.C. Bergman, V.A. Trent	1968	GQ-721	Irvine	M. Murphy D. Curl
FALMOUTH	Stanley J. Luft	1972	GQ-1037	Falmouth	X. Yang
FELICITY	R.H. Osborne, W.F. Outerbridge, M.P. Weiss	1973	GQ-1063	Falmouth	Q. Zhang
FORD	Douglas F.B. Black	1968	GQ-764	Harrodsburg	H. Nelson
FRANKFORT EAST	J.S. Pomeroy	1968	GQ-707	Lexington	H. Nelson
FRANKFORT WEST	Frank B. Moore	1975	GQ-1221	Lexington	H. Nelson
FRANKLINTON	A.B. Gibbons	1976	GQ-1330	Louisville	H. Nelson
FRENCHBURG	Harry P. Hoge	1977	GQ-1390	Irvine	H. Nelson, D. Curl
GEORGETOWN	Earle R. Cressman	1967	GQ-605	Lexington	H. Nelson
GLENCOE	W.C. Swadley	1974	GQ-1154	Falmouth	H. Nelson
GLENSBORO	E. R. Cressman	1976	GQ-1355	Louisville	C. Hettinger
GOFORTH	Stanley J. Luft	1971	GQ-925	Falmouth	M. Thompson
GRATZ	Frank B. Moore	1977	GQ-1359	Lexington	M. Thompson
HALLS GAP	Gordon W. Weir	1972	GQ-1009	Somerset	X. Yang
HARRODSBURG	John W. Allingham	1972	GQ-1020	Harrodsburg	H. Nelson
HAZEL GREEN	W.B. Cashion	1963	GQ-266	Irvine	M. Murphy, D. Curl
HEDGES	Douglas F.B. Black	1975	GQ-1235	Harrodsburg	M. Murphy
HEIDELBERG	Douglas F.B. Black	1977	GQ-1340	Irvine	M. Murphy, D. Curl
HUSTONVILLE	Richard Q. Lewis Sr., Alfred R. Taylor	1971	GQ-916	Somerset	C. Hettinger
INDEPENDENCE	Stanley J. Luft	1969	GQ-785	Falmouth	M. Tyra

Quadrangle Name(s)	Author(s)	Year	GQ No.	100K Quad.	KGS Digitizer
IRVINE	Harry P. Hoge, Perry B. Wigley, Fred R. Shawe	1976	GQ-1285	Irvine	L. Morris, D. Curl
JOHNETTA	J.L. Gualtieri	1968	GQ-685	Somerset	Q. Zhang
JUNCTION CITY	Leonard D. Harris	1972	GQ-981	Harrodsburg	Q. Zhang
KEENE	E.R. Cressman	1965	GQ-440	Harrodsburg	M. Thompson
KELAT	Stanley J. Luft	1974	GQ-1172	Falmouth	Q. Zhang
KIRKSVILLE	Robert C. Greene	1965	GQ-452	Harrodsburg	H. Nelson
LANCASTER	Gordon W. Weir	1971	GQ-888	Harrodsburg	H. Nelson
LANDSAW	Wallace R. Hansen, John E. Johnston	1963	GQ-201	Irvine	V. Sullivan, D. Curl
LAUREL	J.J. Kohut, S. Luft, M.P. Weiss	1973	GQ-1075	Falmouth	Q. Zhang
LAWRENCEBURG	W.C. Swadley	1975	GQ-1026	Lexington	C. Hettinger
LAWRENCEBURG (IND)-AURORA-HOOVEN	W.C. Swadley	1972	GQ-989	Cincinnati	S. Cordiviola
LAWRENCEVILLE	E.R. Cressman	1972	GQ-1204	Falmouth	H. Nelson
LEESBURG	Roberts M. Wallace	1976	GQ-1328	Lexington	H. Nelson
LEIGHTON	Donald C. Haney, Charles L. Rice	1978	GQ-1495	Irvine	L. Morris, D. Curl
LEVEE	Robert C. McDowell	1978	GQ-1478	Irvine	L. Morris, D. Curl
LEXINGTON EAST	Ernest Dobrovolny, William MacQuown Jr.	1968	GQ-683	Lexington	D. Curl
LEXINGTON WEST	Robert D. Miller	1967	GQ-600	Lexington	Q. Zhang
LITTLE HICKMAN	Don E. Wolcott	1969	GQ-792	Harrodsburg	E. Ciszak
MADISON EAST	A.B. Gibbons	1978	GQ-1471	Madison	M. Tyra
MADISON WEST	W.C. Swadley	1978	GQ-1469	Madison	
MARETBURG	Seymour O. Schlanger	1965	GQ-338	Somerset	X. Yang
MASON	Stanley J. Luft	1976	GQ-1311	Falmouth	H. Nelson
MCBRAYER	E. R. Cressman	1973	GQ-1079	Harrodsburg	E. Ciszak
MCKEE	Gordon W. Weir, Martin D. Mumma	1973	GQ-1125	Hazard	H. Nelson, J. Patton
MEANS	Gordon W. Weir	1976	GQ-1324	Irvine	L. Morris, D. Curl
MIDWAY	J.S Pomeroy	1970	GQ-856	Lexington	H. Nelson
MILLERSBURG	Norman P. Cupples, William F. Outerbridge	1974	GQ-1219	Lexington	H. Nelson
MOBERLY	Robert C. Greene	1968	GQ-664	Harrodsburg	J. Miller
MONTEREY	Frank B. Moore	1977	GQ-1400	Lexington	M. Thompson

Quadrangle Name(s)	Author(s)	Year	GQ No.	100K Quad.	KGS Digitizer
MOSCOW	Stanley J. Luft, R.H. Osborne, Malcolm P. Weiss	1973	GQ-1069	Falmouth	Q. Zhang
MT OLIVET	Roberts M. Wallace	1977	GQ-1404	Falmouth	A. Harper
NEW CASTLE	A.B. Gibbons	1977	GQ-1431	Louisville	
NEW COLUMBUS	Frank B. Moore	1978	GQ-1492	Lexington	J. Patton
NEW LIBERTY	A.B. Gibbons, W.C. Swadley	1976	GQ-1348	Falmouth	H. Nelson
NEW RICHMOND	A.B. Gibbons, J.J. Kohut, M.P. Weiss	1975	GQ-1228	Falmouth	M. Tyra
NEWPORT-WITHAMSVILLE	A.B. Gibbons	1973	GQ-1072	Cincinnati	S. Cordiviola
NICHOLASVILLE	W.C. MacQuown Jr.	1968	GQ-767	Harrodsburg	H. Nelson
NORTH MIDDLETOWN	Charles T. Helfrich	1977	GQ-1444	Lexington	M. Murphy
NORTH PLEASUREVILLE	Warren L. Peterson	1976	GQ-1346	Louisville	H. Nelson
OWENTON	W.C. Swadley	1975	GQ-1237	Falmouth	H. Nelson
PAINT LICK	Gordon W. Weir	1969	GQ-800	Harrodsburg	H. Nelson
PALMER	George C. Simmons	1967	GQ-613	Harrodsburg	V. Sullivan
PANOLA	Robert C. Greene	1968	GQ-686	Harrodsburg	J. Miller
PARIS EAST	William F. Outerbridge	1974	GQ-1167	Lexington	X. Yang
PARIS WEST	William F. Outerbridge	1974	GQ-1162	Lexington	X. Yang
PARKSVILLE	Samuel L. Moore	1978	GQ-1494	Harrodsburg	J. Patton
PATRIOT-FLORENCE	W.C. Swadley	1969	GQ-846	Falmouth	M. Tyra
PERRYVILLE	E.R. Cressman	1974	GQ-1185	Harrodsburg	T. Sparks, B. Nuttall
PIQUA	Roberts M. Wallace	1978	GQ-1425	Lexington	T. Sparks
POLSGROVE	Frank B. Moore	1977	GQ-1349	Lexington	M. Thompson
POMEROYTON	Gordon W. Weir, Paul W. Richards	1974	GQ-1184	Irvine	H. Nelson, D. Curl
RICHMOND NORTH	George C. Simmons	1967	GQ-583	Harrodsburg	H. Nelson
RICHMOND SOUTH	Robert C. Greene	1966	GQ-479	Harrodsburg	E. Ciszak
RISING SUN-ABERDEEN	W.C. Swadley	1971	GQ-929	Falmouth	M. Thompson
SADIEVILLE	Moore and Wallace	1978	GQ-1486	Lexington	H. Nelson
SALVISA	E.R. Cressman	1968	GQ-760	Harrodsburg	H. Nelson
SANDERS	W.C. Swadley	1973	GQ-1095	Falmouth	H. Nelson
SANDGAP	J.L. Gualtieri	1973	GQ-1100	Somerset	M. Solis
SCRANTON	Donald C. Haney, Norman C. Hester	1978	GQ-1488	Irvine	H. Nelson, D. Curl
SHADY NOOK	Roberts M. Wallace	1975	GQ-1261	Lexington	H. Nelson
SHAWHAN	Norman P. Cuppels	1954	GQ-1122	Lexington	H. Nelson
SHELBYVILLE	E.R. Cressman	1975	GQ-1258	Louisville	M. Thompson

Quadrangle Name(s)	Author(s)	Year	GQ No.	100K Quad.	KGS Digitizer
SIDEVIEW	Lawrence V. Blade	1976	GQ-1356	Lexington	R. Duncan
SLADE	Gordon W. Weir	1974	GQ-1183	Irvine	H. Nelson, D. Curl
SMITHFIELD	Stanley J. Luft	1977	GQ-1371	Louisville	
STAMPING GROUND	Frank B. Moore	1977	GQ-1430	Lexington	H. Nelson
STANFORD	Fred R. Shawe, Perry B. Wigley	1974	GQ-1137	Harrodsburg	C. Hettinger
STANTON	Gordon W. Weir	1974	GQ-1182	Irvine	T. Sparks, D. Curl
STURGEON	Gordon W. Weir	1978	GQ-1455	Hazard	H. Nelson, J. Patton
SWITZER	Frank B. Moore	1975	GQ-1266	Lexington	H. Nelson
TALLEGA	Douglas F.B. Black	1978	GQ-1500	Irvine	V. Sullivan, D. Curl
TYRONE	E.R. Cressman	1964	GQ-303	Lexington	H. Nelson
UNION	W.C. Swadley	1969	GQ-779	Falmouth	M. Thompson
UNION CITY	George C. Simmons	1967	GQ-585	Harrodsburg	H. Nelson
VALLEY VIEW	Robert C. Greene	1966	GQ-470	Harrodsburg	E. Ciszak
VERONA	W.C. Swadley	1969	GQ-819	Falmouth	H. Nelson
VERSAILLES	Douglas F.B. Black	1964	GQ-325	Lexington	H. Nelson
VEVAY NORTH-VEVAY SOUTH	W.C. Swadley	1973	GQ-1123	Madison	M. Tyra
WADDY	E.R. Cressman	1975	GQ-1255	Louisville	M. Thompson
WALTON	Stanley J. Luft	1973	GQ-1080	Falmouth	H. Nelson
WILDIE	J.L. Gualtieri	1968	GQ-684	Somerset	X. Yang
WILLIAMSTOWN	Stanley J. Luft	1973	GQ-1104	Falmouth	R. Duncan
WILMORE	E.R. Cressman, S.V. Hrabar	1970	GQ-847	Harrodsburg	E. Ciszak
WINCHESTER	Douglas F.B. Black	1974	GQ-1159	Harrodsburg	M. Murphy
WORTHVILLE	A.B. Gibbons	1975	GQ-1265	Madison	M. Tyra
ZACHARIAH	Douglas F.B. Black	1978	GQ-1452	Irvine	V. Sullivan, D. Curl

APPENDIX 2

Digital data sets utilized in this study

Appendix 2. Digital data sets utilized in this study

Theme	Bedrock and quaternary geology
Title	Digitally Vectorized Geologic Quadrangles (DVGQ)
Format	vector, polygon/line/point
Scale/resolution	1:24,000
Database size (this study)	540 MB, Kentucky River basin
Produced by	Kentucky Geological Survey
Acquired from	Kentucky Geological Survey
Website	http://www.uky.edu/KGS

Theme	Land-surface elevation
Title	Digital Elevation Model (DEM)
Format	raster/grid
Scale/resolution	10 meter
Database size (this study)	280 MB, Kentucky River basin
Produced by	U.S. Geological Survey
Acquired from	Kentucky Office of Geographic Information Systems
Website	http://ogis.state.ky.us

Theme	Land-surface elevation
Title	Shuttle Radar Topography Mission (SRTM)
Format	raster/grid
Scale/resolution	90 meter
Database size (this study)	325 MB, Illinois, Indiana, Kentucky, West Virgina
Produced by	National Aeronautics and Space Administration
Acquired from	U.S. Geological Survey
Website	http://srtm.usgs.gov

Theme	Topographic maps
Title	Digital Raster Graphic (DRG)
Format	image
Scale/resolution	1:24,000
Database size (this study)	approx. 1,440 MB, Kentucky River basin
Produced by	U.S. Geological Survey
Acquired from	Kentucky Office of Geographic Information Systems
Website	http://ogis.state.ky.us

Theme	Soils
Title	Soil Survey Geographic database (SSURGO)
Format	Vector, polygon/point
Scale/resolution	1:12,000
Database size (this study)	240 MB, Kentucky River basin
Produced by	U.S. Department of Agriculture
Acquired from	U.S. Department of Agriculture
Website	http://www.ncgc.nrcs.usda.gov/branch/ssb/products/ssurgo/

Theme	Streams
Title	National Hydrologic Dataset (NHD)
Format	Vector, line
Scale/resolution	1:24,000
Database size (this study)	40 MB, Kentucky River basin / 164 Mb statewide
Produced by	U.S. Geological Survey
Acquired from	U.S. Geological Survey
Website	http://nhd.usgs.gov/

APPENDIX 3

Comparison of different measures of distance in the Kentucky River valley

Appendix 3. Comparison of different measures of distance in the Kentucky River valley

Feature	USACoE mi	USACoE km	NHD mi	NHD km	Valley mi	Valley km
Lock and Dam No 1	4.0	6.4	4.3	7.0	3.7	6.0
Mouth of Eagle Creek	11.0	17.7	11.6	18.6	9.5	15.3
Mouth of Big Twin Creek	17.4	28.0	18.1	29.2	14.9	24.0
Mouth of Drennon Creek	21.0	33.8	21.9	35.3	17.8	28.7
Lock and Dam No 2	31.0	49.9	32.2	51.9	25.2	40.6
Mouth of Sixmile Creek	31.1	50.0	32.3	52.0	25.3	40.8
Mouth of Severn Creek	35.8	57.6	37.0	59.5	28.2	45.4
Mouth of Cedar Creek	41.8	67.3	42.8	68.8	31.8	51.2
Lock and Dam No 3	42.0	67.6	43.1	69.3	32.0	51.5
Mouth of Flat Creek	48.4	77.9	49.7	80.0	37.3	60.1
Mouth of Elkhorn Creek	51.9	83.5	53.4	86.0	40.4	65.0
Lock and Dam No 4	65.0	104.6	67.2	108.2	49.8	80.2
Mouth of Benson Creek	65.8	105.9	68.0	109.5	50.5	81.2
Mouth of Glenns Creek	71.4	114.9	73.8	118.7	54.7	88.1
Lock and Dam No 5	82.2	132.3	84.5	136.0	63.8	102.6
Mouth of Bailey Run	84.6	136.1	86.9	139.9	65.7	105.8
Mouth of Grier Creek	85.5	137.6	87.9	141.5	66.6	107.3
Mouth of Gilbert Creek	89.7	144.3	92.4	148.7	70.6	113.6
Mouth of Clear Creek	94.7	152.4	97.7	157.2	74.3	119.7
Lock and Dam No 6	96.2	154.8	99.4	159.9	75.7	121.9
Lock and Dam No 7	117.0	188.3	120.2	193.4	93.7	150.9
Mouth of Dix River	118.2	190.2	121.4	195.3	94.9	152.7
Mouth of Jessamine Creek	127.4	205.0	130.7	210.3	103.8	167.0
Kentucky River at Camp Nelson	135.1	217.4	138.1	222.2	110.7	178.2
Mouth of Hickman Creek	135.3	217.7	138.7	223.2	111.3	179.2
Lock and Dam No 8	139.9	225.1	143.4	230.7	115.2	185.5
Mouth of Sugar Creek	142.6	229.4	146.3	235.4	117.9	189.8
Mouth of Paint Lick Creek	146.0	234.9	149.6	240.8	120.9	194.5
Mouth of Silver Creek	150.3	241.8	154.0	247.8	124.6	200.6
Lock and Dam No 9	157.9	254.1	161.4	259.8	131.6	211.9
Mouth of Tates Creek	158.1	254.4	161.9	260.6	132.3	212.9
Mouth of Boone Creek	170.6	274.5	174.8	281.3	144.3	232.2
Mouth of Lower Howard Creek	174.6	280.9	178.9	287.9	148.3	238.7
Lock and Dam No 10	176.4	283.8	180.7	290.9	150.1	241.5
Mouth of Otter Creek	177.3	285.3	181.6	292.2	150.8	242.8
Mouth of Twomile Creek	179.3	288.5	183.8	295.8	152.7	245.8
Mouth of Fourmile Creek	180.4	290.3	184.8	297.5	153.7	247.4
Mouth of Muddy Creek	184.8	297.3	189.4	304.9	157.8	253.9
Mouth of Upper Howard Creek	187.2	301.2	191.9	308.9	159.9	257.4
Mouth of Red River	190.8	307.0	195.6	314.8	162.5	261.5
Lock and Dam No 11	201.0	323.4	205.9	331.4	171.0	275.3
Mouth of Drowning Creek	205.1	330.0	210.0	338.0	174.0	280.1
Mouth of Station Camp Creek	218.5	351.6	223.4	359.6	185.1	297.9
Lock and Dam No 12	220.9	355.4	226.0	363.7	187.0	300.9
Mouth of Cow Creek	221.2	355.9	226.2	364.1	187.3	301.4
Mouth of Millers Creek	225.8	363.3	230.7	371.3	190.6	306.7
Lock and Dam No 13	239.9	386.0	244.8	393.9	203.7	327.9
Mouth of Sturgeon Creek	248.7	400.2	253.7	408.3	212.2	341.5
Lock and Dam No 14	249.0	400.6	253.9	408.7	212.4	341.8
Confluence of North and South Forks	254.8	410.0	259.7	418.0	217.7	350.3

APPENDIX 4

Flood frequency for the Kentucky River calculated by the U.S. Army Corps of Engineers

Appendix 4. Flood frequency for the Kentucky River calculated by the U.S. Army Corps of Engineers

Dist mi	Dist km	Thalweg ele	Freq-98%	Freq-50%	Freq-20%	Freq-10%	Freq-5%	Freq-2%	Freq-1%	Freq-0.2%
1.00	1.64	123.6	135.00	136.60	138.06	139.33	141.03	142.15	142.72	144.02
2.00	3.28	123.7	135.36	136.91	138.34	139.60	141.28	142.39	142.96	144.23
3.00	4.92	124.0	135.74	137.23	138.62	139.88	141.51	142.57	143.12	144.36
3.95	6.48	124.1	135.98	137.48	138.86	140.12	141.70	142.72	143.26	144.48
3.98	6.53	124.1	135.98	137.48	138.86	140.12	141.70	142.72	143.27	144.48
3.98	6.53	124.2	135.96	137.46	138.85	140.10	141.69	142.71	143.25	144.47
3.98	6.53	124.2	135.96	137.46	138.85	140.10	141.69	142.71	143.25	144.47
4.00	6.56	124.3	135.88	137.39	138.80	140.06	141.65	142.68	143.22	144.44
4.00	6.56	124.2	135.88	137.40	138.80	140.06	141.66	142.68	143.22	144.44
4.08	6.69	124.2	136.00	137.46	138.83	140.08	141.66	142.68	143.22	144.44
4.19	6.87	124.3	136.01	137.46	138.82	140.07	141.66	142.69	143.23	144.46
4.41	7.23	124.4	136.14	137.57	138.93	140.21	141.86	142.91	143.46	144.70
5.76	9.45	124.5	136.78	138.14	139.44	140.65	142.08	143.07	143.60	144.81
6.91	11.33	124.7	137.06	138.40	139.69	140.88	142.26	143.22	143.76	144.95
8.00	13.12	125.0	137.36	138.67	139.98	141.17	142.53	143.49	144.03	145.21
9.00	14.76	125.2	137.64	138.94	140.27	141.43	142.72	143.66	144.19	145.34
10.00	16.40	125.3	137.85	139.14	140.50	141.64	142.88	143.80	144.32	145.46
10.95	17.96	125.6	138.09	139.40	140.77	141.86	143.04	143.94	144.45	145.57
12.00	19.68	125.6	138.30	139.60	140.99	142.06	143.22	144.12	144.64	145.74
13.00	21.32	125.8	138.64	139.95	141.36	142.44	143.58	144.49	145.02	146.12
14.00	22.96	125.9	138.83	140.15	141.55	142.62	143.74	144.64	145.17	146.28
15.00	24.60	126.1	139.05	140.36	141.77	142.85	143.97	144.88	145.41	146.53
16.00	26.24	126.2	139.29	140.60	142.02	143.10	144.21	145.12	145.66	146.79
17.00	27.88	126.3	139.49	140.80	142.25	143.34	144.43	145.34	145.89	147.02
18.00	29.52	126.5	139.66	140.96	142.43	143.52	144.61	145.51	146.06	147.18
19.00	31.16	126.6	139.91	141.21	142.70	143.77	144.82	145.70	146.25	147.35
20.00	32.80	126.8	140.12	141.45	142.96	144.03	145.06	145.96	146.51	147.62
21.00	34.44	126.9	140.33	141.68	143.20	144.27	145.30	146.21	146.77	147.87
22.00	36.08	127.1	140.45	141.81	143.35	144.42	145.44	146.35	146.91	148.01
23.00	37.72	127.2	140.65	142.00	143.55	144.63	145.64	146.54	147.11	148.21
24.00	39.36	127.3	140.82	142.17	143.74	144.82	145.84	146.76	147.32	148.41
25.00	41.00	127.4	141.05	142.39	143.98	145.08	146.11	147.04	147.63	148.75
26.00	42.64	127.7	141.23	142.58	144.19	145.30	146.34	147.29	147.88	149.00
27.00	44.28	127.7	141.38	142.74	144.37	145.49	146.53	147.48	148.07	149.18
28.00	45.92	127.9	141.49	142.85	144.50	145.62	146.65	147.61	148.20	149.31

Dist mi	Dist km	Thalweg ele	Freq-98%	Freq-50%	Freq-20%	Freq-10%	Freq-5%	Freq-2%	Freq-1%	Freq-0.2%
29.00	47.56	128.0	141.59	142.96	144.62	145.73	146.78	147.74	148.34	149.46
30.00	49.20	128.1	141.72	143.10	144.77	145.90	146.95	147.93	148.54	149.68
30.85	50.59	128.3	141.83	143.23	144.90	146.03	147.08	148.06	148.68	149.82
30.99	50.82	128.4	141.82	143.22	144.90	146.02	147.07	148.05	148.67	149.82
30.99	50.82	128.4	141.81	143.21	144.88	146.01	147.05	148.04	148.65	149.80
30.99	50.82	128.4	141.81	143.21	144.88	146.01	147.05	148.04	148.65	149.80
31.00	50.84	128.4	141.77	143.17	144.85	145.98	147.03	148.02	148.63	149.78
31.00	50.84	128.4	141.77	143.17	144.85	145.98	147.03	148.02	148.63	149.78
31.11	51.02	128.5	141.88	143.28	144.95	146.08	147.11	148.10	148.71	149.86
32.00	52.48	128.6	141.92	143.32	145.00	146.12	147.17	148.17	148.79	149.94
33.00	54.12	128.9	142.09	143.50	145.19	146.32	147.37	148.36	148.99	150.14
34.00	55.76	129.2	142.27	143.69	145.39	146.53	147.57	148.57	149.19	150.35
35.00	57.40	129.4	142.41	143.84	145.56	146.70	147.75	148.77	149.40	150.58
36.00	59.04	129.7	142.53	143.98	145.70	146.86	147.91	148.92	149.55	150.72
37.00	60.68	130.0	142.66	144.12	145.86	147.02	148.07	149.08	149.71	150.88
38.00	62.32	130.2	142.79	144.26	146.01	147.17	148.22	149.24	149.87	151.04
39.00	63.96	130.5	142.97	144.46	146.23	147.40	148.46	149.47	150.11	151.28
40.00	65.60	130.8	143.26	144.77	146.56	147.72	148.75	149.75	150.38	151.53
41.00	67.24	131.1	143.42	144.92	146.69	147.85	148.88	149.87	150.50	151.64
41.90	68.72	131.3	143.56	145.08	146.85	148.00	149.03	150.03	150.65	151.80
41.99	68.86	131.4	143.56	145.08	146.85	148.00	149.03	150.02	150.65	151.79
41.99	68.86	131.5	143.54	145.06	146.83	147.99	149.01	150.01	150.63	151.78
41.99	68.86	131.5	143.54	145.06	146.83	147.99	149.01	150.01	150.63	151.78
42.00	68.88	131.5	143.43	144.98	146.77	147.94	148.97	149.97	150.60	151.74
42.00	68.88	131.5	143.43	144.98	146.77	147.94	148.97	149.97	150.60	151.74
42.02	68.91	131.5	143.53	145.03	146.79	147.95	148.98	149.97	150.60	151.74
43.00	70.52	131.7	143.90	145.36	147.09	148.25	149.28	150.29	150.92	152.08
44.00	72.16	131.9	144.23	145.69	147.40	148.53	149.54	150.53	151.15	152.30
45.00	73.80	132.0	144.61	146.05	147.73	148.83	149.81	150.78	151.39	152.53
46.00	75.44	132.3	144.90	146.36	148.04	149.12	150.09	151.06	151.67	152.80
47.00	77.08	132.4	145.05	146.48	148.15	149.23	150.18	151.14	151.75	152.87
48.00	78.72	132.6	145.22	146.67	148.35	149.43	150.40	151.37	151.99	153.12
49.00	80.36	132.7	145.40	146.86	148.53	149.60	150.56	151.52	152.13	153.25
50.00	82.00	132.9	145.61	147.07	148.75	149.82	150.78	151.75	152.37	153.50
51.00	83.64	133.1	145.76	147.22	148.89	149.97	150.93	151.91	152.53	153.66
52.00	85.28	133.3	145.94	147.41	149.09	150.17	151.13	152.10	152.71	153.84
53.00	86.92	133.5	146.04	147.51	149.20	150.28	151.25	152.23	152.86	154.00
54.00	88.56	133.7	146.17	147.65	149.34	150.43	151.40	152.40	153.02	154.18

Dist mi	Dist km	Thalweg ele	Freq-98%	Freq-50%	Freq-20%	Freq-10%	Freq-5%	Freq-2%	Freq-1%	Freq-0.2%
55.00	90.20	133.8	146.33	147.82	149.52	150.63	151.61	152.62	153.25	154.41
56.00	91.84	133.9	146.52	148.00	149.70	150.80	151.77	152.77	153.40	154.56
57.00	93.48	134.1	146.72	148.21	149.93	151.03	152.01	153.03	153.66	154.81
58.00	95.12	134.4	146.87	148.37	150.07	151.18	152.15	153.17	153.79	154.95
59.00	96.76	134.7	146.97	148.46	150.15	151.25	152.23	153.24	153.87	155.03
60.00	98.40	134.9	147.14	148.63	150.31	151.42	152.40	153.41	154.04	155.20
61.00	100.04	135.1	147.30	148.77	150.42	151.52	152.49	153.50	154.12	155.28
62.00	101.68	135.3	147.48	148.97	150.62	151.73	152.71	153.73	154.36	155.55
63.00	103.32	135.6	147.63	149.11	150.76	151.87	152.85	153.88	154.52	155.71
64.37	105.57	134.4	147.84	149.35	151.01	152.14	153.14	154.20	154.85	156.08
64.70	106.11	134.4	147.87	149.37	151.03	152.17	153.17	154.23	154.89	156.12
64.96	106.53	134.5	147.92	149.42	151.09	152.23	153.23	154.30	154.95	156.20
64.99	106.58	136.1	148.02	149.54	151.21	152.35	153.35	154.41	155.06	156.30
64.99	106.58	136.3	148.00	149.52	151.20	152.34	153.34	154.40	155.05	156.29
65.00	106.60	136.4	148.00	149.52	151.20	152.34	153.34	154.40	155.05	156.29
65.00	106.60	136.6	147.94	149.47	151.16	152.31	153.31	154.38	155.03	156.27
65.00	106.60	136.8	147.94	149.48	151.16	152.31	153.31	154.38	155.03	156.27
65.10	106.76	136.9	148.02	149.52	151.19	152.32	153.32	154.39	155.05	156.29
66.23	108.62	131.4	148.20	149.74	151.44	152.56	153.55	154.60	155.25	156.49
66.43	108.95	135.2	148.22	149.77	151.48	152.60	153.59	154.65	155.30	156.55
66.45	108.98	135.7	148.20	149.75	151.46	152.58	153.57	154.63	155.27	156.52
66.70	109.39	135.7	148.25	149.82	151.53	152.65	153.64	154.70	155.34	156.58
66.72	109.42	134.0	148.23	149.81	151.52	152.64	153.62	154.68	155.32	156.57
66.83	109.60	134.3	148.24	149.82	151.54	152.65	153.64	154.70	155.34	156.58
67.11	110.06	135.3	148.34	149.94	151.67	152.80	153.78	154.83	155.48	156.73
67.35	110.45	133.8	148.41	150.01	151.77	152.89	153.88	154.94	155.59	156.84
68.00	111.52	135.5	148.49	150.09	151.89	153.02	154.00	155.06	155.71	156.98
68.57	112.45	137.9	148.58	150.17	152.01	153.14	154.12	155.18	155.83	157.09
69.46	113.91	138.0	148.77	150.33	152.22	153.34	154.30	155.34	155.98	157.23
70.12	115.00	135.4	148.90	150.46	152.39	153.50	154.46	155.51	156.14	157.38
70.66	115.88	136.5	148.98	150.54	152.51	153.61	154.57	155.60	156.23	157.47
70.68	115.92	136.5	148.98	150.55	152.52	153.62	154.57	155.61	156.24	157.48
70.68	115.92	135.9	148.97	150.53	152.50	153.60	154.55	155.59	156.21	157.44
70.69	115.93	135.9	148.98	150.54	152.50	153.60	154.55	155.59	156.22	157.45
70.69	115.93	135.8	149.00	150.57	152.54	153.65	154.60	155.64	156.27	157.51
70.74	116.01	135.8	149.01	150.58	152.56	153.66	154.62	155.66	156.29	157.53
70.74	116.01	135.4	149.00	150.57	152.54	153.64	154.59	155.63	156.26	157.49
70.75	116.03	135.4	149.00	150.57	152.54	153.64	154.59	155.63	156.26	157.50

Dist mi	Dist km	Thalweg ele	Freq-98%	Freq-50%	Freq-20%	Freq-10%	Freq-5%	Freq-2%	Freq-1%	Freq-0.2%
70.75	116.03	135.3	149.02	150.59	152.57	153.67	154.63	155.67	156.30	157.55
70.78	116.08	135.3	149.03	150.60	152.58	153.68	154.64	155.68	156.31	157.56
71.15	116.69	136.1	149.11	150.70	152.71	153.82	154.77	155.82	156.46	157.71
71.64	117.49	137.6	149.19	150.80	152.83	153.93	154.88	155.93	156.57	157.82
72.32	118.60	136.6	149.31	150.94	152.99	154.09	155.04	156.08	156.71	157.96
74.60	122.34	137.3	149.69	151.41	153.51	154.59	155.52	156.56	157.18	158.43
76.00	124.64	136.0	150.01	151.76	153.89	154.95	155.87	156.89	157.51	158.76
77.90	127.76	135.9	150.41	152.16	154.36	155.40	156.32	157.34	157.96	159.22
78.80	129.23	137.7	150.59	152.37	154.59	155.64	156.55	157.58	158.20	159.47
79.30	130.05	137.3	150.72	152.52	154.74	155.78	156.69	157.72	158.34	159.62
81.00	132.84	138.4	151.07	152.97	155.17	156.20	157.10	158.13	158.76	160.04
82.07	134.59	140.4	151.25	153.20	155.38	156.42	157.33	158.37	159.02	160.33
82.19	134.79	140.5	151.28	153.24	155.42	156.46	157.37	158.42	159.06	160.38
82.19	134.79	140.5	151.26	153.22	155.41	156.45	157.36	158.41	159.05	160.37
82.19	134.79	140.5	151.26	153.22	155.41	156.45	157.36	158.41	159.05	160.37
82.20	134.81	140.5	151.10	153.14	155.36	156.41	157.32	158.37	159.02	160.33
82.20	134.81	140.5	151.10	153.14	155.36	156.41	157.32	158.37	159.02	160.33
82.31	134.99	140.6	151.36	153.24	155.41	156.44	157.35	158.39	159.04	160.35
83.40	136.78	139.6	151.60	153.42	155.65	156.69	157.60	158.65	159.30	160.63
84.80	139.07	138.8	152.05	153.78	156.11	157.16	158.08	159.14	159.80	161.17
86.00	141.04	139.0	152.36	154.05	156.48	157.53	158.44	159.51	160.18	161.56
88.90	145.80	139.4	153.08	154.79	157.27	158.32	159.24	160.32	161.01	162.42
90.50	148.42	140.7	153.33	155.09	157.59	158.64	159.56	160.66	161.36	162.80
91.90	150.72	140.5	153.65	155.54	158.00	159.06	160.00	161.11	161.82	163.27
93.40	153.18	141.5	154.04	156.02	158.45	159.51	160.45	161.58	162.29	163.76
95.90	157.28	142.5	154.53	156.63	158.99	160.05	161.00	162.14	162.86	164.35
96.07	157.55	144.8	154.63	156.73	159.08	160.14	161.09	162.23	162.95	164.45
96.19	157.75	145.0	154.63	156.74	159.08	160.14	161.09	162.23	162.96	164.45
96.19	157.75	145.0	154.63	156.74	159.08	160.14	161.09	162.23	162.96	164.45
96.19	157.75	145.0	154.63	156.74	159.08	160.14	161.09	162.23	162.96	164.45
96.20	157.77	145.0	154.63	156.74	159.08	160.14	161.09	162.23	162.96	164.45
96.20	157.77	145.0	154.63	156.74	159.08	160.14	161.09	162.23	162.96	164.45
96.22	157.80	145.0	155.09	156.74	159.08	160.14	161.09	162.23	162.96	164.45
96.33	157.98	145.1	155.55	156.94	159.17	160.22	161.15	162.28	163.01	164.49
96.60	158.42	143.3	155.59	156.97	159.19	160.24	161.18	162.31	163.03	164.52
100.20	164.33	143.9	156.74	158.14	160.21	161.26	162.20	163.30	164.00	165.45
102.90	168.76	144.4	157.37	158.80	160.83	161.89	162.83	163.92	164.62	166.08
105.00	172.20	143.5	157.76	159.22	161.24	162.31	163.24	164.33	165.03	166.49

Dist mi	Dist km	Thalweg ele	Freq-98%	Freq-50%	Freq-20%	Freq-10%	Freq-5%	Freq-2%	Freq-1%	Freq-0.2%
107.20	175.81	143.3	158.27	159.76	161.76	162.83	163.74	164.83	165.53	166.98
109.50	179.58	144.9	158.93	160.46	162.44	163.49	164.41	165.51	166.21	167.67
111.40	182.70	144.8	159.43	160.98	162.92	163.97	164.88	165.97	166.67	168.14
113.20	185.65	147.0	159.95	161.48	163.41	164.44	165.34	166.43	167.13	168.59
114.00	186.96	144.6	160.13	161.65	163.58	164.61	165.51	166.59	167.30	168.76
114.60	187.94	146.6	160.26	161.78	163.72	164.75	165.65	166.74	167.45	168.92
115.90	190.08	147.4	160.52	162.02	163.98	165.00	165.90	167.00	167.70	169.18
116.60	191.22	147.3	160.68	162.18	164.14	165.16	166.06	167.15	167.86	169.33
116.89	191.70	149.0	160.77	162.28	164.24	165.27	166.17	167.27	167.98	169.47
117.02	191.91	149.0	160.77	162.28	164.24	165.27	166.17	167.27	167.98	169.47
117.02	191.91	149.1	160.77	162.28	164.24	165.27	166.17	167.27	167.98	169.47
117.02	191.91	149.1	160.77	162.28	164.24	165.27	166.17	167.27	167.98	169.47
117.03	191.93	149.1	160.77	162.28	164.24	165.27	166.17	167.27	167.98	169.47
117.03	191.93	149.1	160.77	162.28	164.24	165.27	166.17	167.27	167.98	169.47
117.06	191.98	149.1	160.77	162.28	164.24	165.27	166.17	167.27	167.98	169.47
117.20	192.21	149.1	161.18	162.55	164.41	165.41	166.29	167.37	168.07	169.53
117.90	193.36	149.2	161.42	162.78	164.63	165.63	166.51	167.58	168.27	169.73
118.40	194.18	147.0	161.53	162.89	164.74	165.74	166.61	167.68	168.38	169.85
120.40	197.46	146.3	162.28	163.60	165.48	166.44	167.28	168.30	168.97	170.36
121.60	199.42	148.4	162.67	163.99	165.84	166.84	167.67	168.68	169.33	170.69
123.60	202.70	148.2	163.13	164.45	166.27	167.31	168.13	169.12	169.76	171.09
125.60	205.98	149.8	163.58	164.91	166.74	167.79	168.61	169.58	170.20	171.52
127.30	208.77	150.2	163.95	165.27	167.12	168.15	168.96	169.93	170.54	171.84
128.00	209.92	149.3	164.08	165.40	167.26	168.28	169.09	170.05	170.65	171.94
129.90	213.04	149.6	164.52	165.84	167.70	168.76	169.57	170.53	171.13	172.40
132.00	216.48	151.6	165.07	166.40	168.24	169.32	170.13	171.08	171.67	172.91
133.80	219.43	149.5	165.43	166.77	168.63	169.71	170.51	171.46	172.04	173.28
134.70	220.91	151.1	165.62	166.96	168.84	169.91	170.72	171.67	172.25	173.48
135.10	221.56	150.9	165.69	167.02	168.92	169.98	170.79	171.74	172.32	173.54
135.25	221.81	153.8	165.71	167.05	168.95	170.02	170.83	171.77	172.35	173.58
136.30	223.53	151.9	165.98	167.32	169.21	170.26	171.07	172.00	172.57	173.78
137.40	225.34	152.4	166.22	167.55	169.46	170.51	171.31	172.25	172.82	174.02
139.78	229.24	153.9	166.59	167.92	169.83	170.86	171.67	172.60	173.16	174.36
139.89	229.42	154.0	166.59	167.92	169.83	170.86	171.67	172.60	173.16	174.36
139.89	229.42	154.0	166.59	167.92	169.83	170.86	171.67	172.60	173.16	174.36
139.89	229.42	154.0	166.59	167.92	169.83	170.86	171.67	172.60	173.16	174.36
139.90	229.44	154.0	166.59	167.92	169.83	170.86	171.67	172.60	173.16	174.36
139.90	229.44	154.0	166.59	167.92	169.83	170.86	171.67	172.60	173.16	174.36

Dist mi	Dist km	Thalweg ele	Freq-98%	Freq-50%	Freq-20%	Freq-10%	Freq-5%	Freq-2%	Freq-1%	Freq-0.2%
139.92	229.47	154.0	166.59	167.92	169.83	170.86	171.67	172.60	173.16	174.36
140.40	230.26	153.1	167.14	168.24	170.06	171.06	171.84	172.75	173.30	174.48
141.70	232.39	154.1	167.51	168.63	170.41	171.39	172.16	173.06	173.60	174.76
142.40	233.54	155.4	167.73	168.86	170.60	171.57	172.34	173.23	173.77	174.92
142.90	234.36	153.8	167.98	169.10	170.78	171.72	172.47	173.34	173.87	174.99
145.20	238.13	153.0	168.70	169.85	171.45	172.36	173.09	173.94	174.45	175.53
146.10	239.60	155.5	168.96	170.13	171.71	172.61	173.34	174.18	174.69	175.76
147.70	242.23	153.9	169.44	170.62	172.13	172.99	173.68	174.49	174.97	175.99
149.40	245.02	156.8	169.97	171.18	172.68	173.52	174.21	174.99	175.47	176.47
150.70	247.15	156.1	170.42	171.64	173.10	173.92	174.59	175.35	175.81	176.77
153.00	250.92	156.6	170.99	172.23	173.68	174.50	175.17	175.93	176.38	177.33
155.90	255.68	155.8	171.62	172.88	174.30	175.12	175.79	176.55	176.99	177.92
157.39	258.12	159.4	171.97	173.25	174.65	175.47	176.14	176.91	177.35	178.28
157.49	258.28	159.5	172.02	173.29	174.70	175.52	176.20	176.96	177.40	178.33
157.49	258.28	159.4	172.02	173.29	174.70	175.52	176.20	176.96	177.40	178.33
157.49	258.28	159.4	172.02	173.29	174.70	175.52	176.20	176.96	177.40	178.33
157.50	258.30	159.4	172.02	173.29	174.70	175.52	176.20	176.96	177.40	178.33
157.50	258.30	159.4	172.02	173.29	174.70	175.52	176.20	176.96	177.40	178.33
157.52	258.33	159.4	172.02	173.29	174.70	175.52	176.20	176.96	177.40	178.33
157.90	258.96	159.2	172.28	173.50	174.87	175.67	176.34	177.09	177.53	178.45
158.30	259.61	158.4	172.37	173.58	174.94	175.74	176.39	177.14	177.57	178.48
160.30	262.89	158.4	172.75	173.96	175.29	176.08	176.73	177.46	177.89	178.79
162.30	266.17	157.7	173.12	174.34	175.66	176.44	177.08	177.81	178.24	179.13
163.50	268.14	158.6	173.36	174.57	175.88	176.65	177.28	178.00	178.43	179.31
165.20	270.93	158.4	173.64	174.86	176.16	176.92	177.54	178.24	178.66	179.53
166.30	272.73	158.6	173.90	175.13	176.42	177.16	177.77	178.47	178.89	179.74
167.60	274.86	159.2	174.19	175.42	176.71	177.45	178.06	178.76	179.18	180.04
170.00	278.80	160.0	174.73	175.96	177.22	177.95	178.56	179.25	179.68	180.54
172.10	282.24	159.5	175.21	176.42	177.65	178.36	178.96	179.65	180.08	180.93
173.90	285.20	161.0	175.48	176.70	177.93	178.63	179.24	179.93	180.36	181.21
174.50	286.18	161.6	175.59	176.82	178.05	178.76	179.37	180.06	180.50	181.36
175.50	287.82	161.3	175.80	177.03	178.24	178.94	179.54	180.23	180.66	181.50
176.24	289.03	164.7	175.92	177.15	178.36	179.06	179.66	180.34	180.78	181.63
176.39	289.28	164.8	175.97	177.20	178.41	179.11	179.71	180.39	180.83	181.67
176.39	289.28	164.8	175.97	177.20	178.41	179.11	179.71	180.39	180.83	181.67
176.39	289.28	164.8	175.97	177.20	178.41	179.11	179.71	180.39	180.83	181.67
176.40	289.30	164.8	175.97	177.20	178.41	179.11	179.71	180.39	180.83	181.67
176.40	289.30	164.8	175.97	177.20	178.41	179.11	179.71	180.39	180.83	181.67

Dist mi	Dist km	Thalweg ele	Freq-98%	Freq-50%	Freq-20%	Freq-10%	Freq-5%	Freq-2%	Freq-1%	Freq-0.2%
176.42	289.33	164.8	175.98	177.20	178.41	179.11	179.71	180.39	180.83	181.67
177.00	290.28	164.9	177.03	177.76	178.79	179.42	179.98	180.62	181.03	181.85
178.00	291.92	165.1	177.41	178.13	179.11	179.70	180.23	180.84	181.23	182.01
179.00	293.56	165.3	177.79	178.52	179.46	180.03	180.53	181.11	181.48	182.23
180.00	295.20	165.5	178.13	178.88	179.79	180.33	180.81	181.37	181.74	182.47
181.00	296.84	165.7	178.45	179.22	180.11	180.63	181.10	181.64	182.00	182.71
182.00	298.48	165.9	178.79	179.59	180.48	181.00	181.46	182.00	182.35	183.04
183.00	300.12	166.1	179.12	179.94	180.81	181.33	181.79	182.32	182.67	183.35
184.00	301.76	166.3	179.34	180.17	181.05	181.57	182.03	182.57	182.92	183.60
185.00	303.40	166.5	179.46	180.28	181.14	181.64	182.10	182.63	182.97	183.65
186.00	305.04	166.7	179.69	180.53	181.40	181.91	182.36	182.89	183.24	183.91
187.00	306.68	166.9	179.88	180.74	181.61	182.13	182.59	183.12	183.47	184.15
187.72	307.86	167.1	179.99	180.84	181.71	182.22	182.67	183.20	183.54	184.21
189.00	309.96	167.3	180.21	181.09	181.98	182.51	182.98	183.52	183.88	184.56
190.00	311.60	167.5	180.41	181.31	182.22	182.76	183.24	183.79	184.15	184.84
191.00	313.24	167.7	180.63	181.55	182.48	183.03	183.52	184.08	184.44	185.13
192.00	314.88	167.9	180.77	181.71	182.67	183.23	183.73	184.31	184.68	185.38
193.00	316.52	168.1	181.01	182.00	183.00	183.60	184.11	184.70	185.08	185.82
194.00	318.16	168.3	181.16	182.19	183.22	183.82	184.35	184.95	185.35	186.11
195.00	319.80	168.5	181.33	182.38	183.43	184.05	184.58	185.20	185.60	186.39
196.00	321.44	168.7	181.49	182.56	183.64	184.26	184.81	185.44	185.85	186.67
197.00	323.08	168.9	181.65	182.74	183.83	184.46	185.00	185.63	186.04	186.85
198.00	324.72	169.2	181.86	182.98	184.11	184.75	185.31	185.96	186.39	187.24
199.00	326.36	169.3	182.00	183.14	184.28	184.93	185.50	186.17	186.60	187.47
200.00	328.00	169.5	182.13	183.28	184.43	185.08	185.66	186.33	186.78	187.65
200.84	329.38	169.7	182.27	183.45	184.63	185.30	185.89	186.58	187.03	187.93
200.99	329.62	169.8	182.27	183.45	184.63	185.30	185.89	186.58	187.03	187.93
201.00	329.64	169.9	182.27	183.45	184.63	185.30	185.89	186.58	187.03	187.93
201.00	329.64	169.9	182.27	183.45	184.63	185.30	185.89	186.58	187.03	187.93
201.00	329.64	169.9	182.27	183.45	184.63	185.30	185.89	186.58	187.03	187.93
201.00	329.64	170.0	182.27	183.45	184.63	185.30	185.89	186.58	187.03	187.93
201.02	329.67	170.0	182.36	183.45	184.63	185.30	185.89	186.58	187.03	187.93
202.00	331.28	170.1	183.22	184.06	185.20	185.86	186.45	187.15	187.61	188.52
203.00	332.92	170.4	183.49	184.38	185.54	186.21	186.80	187.50	187.96	188.88
204.00	334.56	170.6	183.78	184.72	185.91	186.59	187.21	187.93	188.42	189.39
205.00	336.20	170.8	184.04	185.03	186.25	186.95	187.59	188.35	188.86	189.88
206.00	337.84	171.1	184.20	185.22	186.46	187.17	187.82	188.59	189.10	190.13
207.00	339.48	171.4	184.40	185.43	186.66	187.37	188.01	188.77	189.28	190.29

Dist mi	Dist km	Thalweg ele	Freq-98%	Freq-50%	Freq-20%	Freq-10%	Freq-5%	Freq-2%	Freq-1%	Freq-0.2%
208.00	341.12	171.6	184.61	185.67	186.91	187.62	188.26	189.03	189.54	190.55
209.00	342.76	171.9	184.81	185.88	187.11	187.81	188.45	189.21	189.72	190.72
210.00	344.40	172.2	185.02	186.10	187.32	188.03	188.66	189.42	189.91	190.91
211.00	346.04	172.5	185.25	186.35	187.56	188.26	188.89	189.64	190.13	191.13
212.00	347.68	172.8	185.45	186.57	187.81	188.52	189.16	189.92	190.42	191.41
213.00	349.32	173.0	185.63	186.74	187.95	188.64	189.28	190.03	190.52	191.52
214.00	350.96	173.2	185.90	187.04	188.27	188.98	189.63	190.37	190.86	191.84
215.00	352.60	173.5	186.13	187.28	188.51	189.22	189.85	190.59	191.08	192.06
216.00	354.24	173.8	186.33	187.50	188.74	189.46	190.11	190.87	191.36	192.36
217.00	355.88	174.1	186.58	187.77	189.02	189.74	190.40	191.16	191.66	192.68
218.00	357.52	174.3	186.75	187.93	189.17	189.89	190.54	191.30	191.80	192.81
219.00	359.16	174.7	186.89	188.07	189.31	190.04	190.70	191.47	191.97	192.98
220.00	360.80	175.0	187.09	188.29	189.54	190.22	190.83	191.55	192.03	193.02
220.79	362.10	175.2	187.23	188.45	189.73	190.43	191.05	191.79	192.27	193.24
220.89	362.26	175.2	187.20	188.42	189.68	190.37	190.99	191.71	192.17	193.11
220.89	362.26	175.3	187.17	188.38	189.65	190.33	190.95	191.66	192.12	193.05
220.89	362.26	175.3	187.17	188.38	189.65	190.33	190.95	191.66	192.12	193.05
220.90	362.28	175.3	187.17	188.38	189.65	190.33	190.95	191.66	192.12	193.05
220.90	362.28	175.4	187.17	188.38	189.65	190.33	190.95	191.66	192.12	193.05
220.90	362.28	175.4	187.17	188.38	189.65	190.33	190.95	191.66	192.12	193.05
220.92	362.31	175.4	187.22	188.45	189.73	190.43	191.06	191.79	192.27	193.23
222.00	364.08	175.5	187.81	188.93	190.14	190.81	191.43	192.16	192.63	193.60
223.00	365.72	175.7	188.50	189.60	190.64	191.23	191.78	192.44	192.88	193.79
224.00	367.36	176.0	189.01	190.02	190.91	191.44	191.96	192.58	193.01	193.91
225.00	369.00	176.2	189.35	190.25	191.08	191.58	192.08	192.69	193.11	193.99
226.00	370.64	176.5	189.54	190.39	191.19	191.68	192.17	192.77	193.18	194.06
227.00	372.28	176.8	189.77	190.58	191.35	191.83	192.30	192.88	193.28	194.14
228.00	373.92	177.1	190.02	190.82	191.59	192.05	192.51	193.08	193.47	194.30
229.00	375.56	177.3	190.28	191.12	191.94	192.41	192.89	193.47	193.87	194.71
230.00	377.20	177.6	190.55	191.44	192.31	192.81	193.30	193.90	194.32	195.18
231.00	378.84	177.9	190.83	191.76	192.67	193.20	193.71	194.34	194.76	195.65
232.00	380.48	178.1	191.12	192.09	193.05	193.60	194.14	194.79	195.23	196.15
233.00	382.12	178.6	191.38	192.38	193.37	193.94	194.48	195.15	195.60	196.54
234.00	383.76	179.2	191.65	192.66	193.68	194.26	194.83	195.52	195.98	196.95
235.00	385.40	179.8	191.84	192.87	193.90	194.49	195.06	195.75	196.22	197.19
236.00	387.04	180.1	192.04	193.08	194.12	194.72	195.30	196.00	196.47	197.46
237.00	388.68	180.1	192.29	193.37	194.46	195.08	195.69	196.42	196.91	197.94
238.00	390.32	180.1	192.49	193.58	194.69	195.32	195.93	196.67	197.17	198.21

Dist mi	Dist km	Thalweg ele	Freq-98%	Freq-50%	Freq-20%	Freq-10%	Freq-5%	Freq-2%	Freq-1%	Freq-0.2%
239.00	391.96	180.1	192.75	193.87	195.02	195.68	196.30	197.05	197.56	198.62
239.75	393.19	180.1	192.90	194.04	195.21	195.88	196.51	197.28	197.80	198.88
239.89	393.42	180.2	192.93	194.06	195.24	195.91	196.54	197.32	197.83	198.92
239.89	393.42	180.4	192.93	194.06	195.24	195.91	196.54	197.32	197.83	198.92
239.90	393.44	180.6	192.93	194.06	195.24	195.91	196.54	197.32	197.83	198.92
239.90	393.44	180.7	192.93	194.06	195.24	195.91	196.54	197.32	197.83	198.92
239.90	393.44	180.9	192.93	194.06	195.24	195.91	196.54	197.32	197.83	198.92
239.92	393.47	181.0	192.93	194.06	195.24	195.91	196.54	197.32	197.83	198.92
241.00	395.24	181.1	193.44	194.56	195.72	196.38	197.01	197.77	198.28	199.36
242.00	396.88	182.0	193.77	194.90	196.09	196.75	197.39	198.17	198.68	199.78
243.00	398.52	183.2	194.09	195.24	196.44	197.12	197.76	198.55	199.08	200.20
244.00	400.16	184.4	194.40	195.56	196.75	197.42	198.06	198.85	199.38	200.49
245.00	401.80	185.6	194.88	196.04	197.25	197.94	198.59	199.39	199.93	201.07
246.00	403.44	185.5	195.28	196.43	197.63	198.31	198.96	199.76	200.30	201.43
247.00	405.08	184.6	195.68	196.83	198.03	198.71	199.36	200.17	200.71	201.85
248.00	406.72	184.5	196.08	197.25	198.47	199.17	199.84	200.66	201.21	202.39
248.90	408.20	185.2	196.36	197.52	198.75	199.45	200.12	200.94	201.50	202.69
248.99	408.34	185.3	196.42	197.59	198.82	199.53	200.20	201.02	201.59	202.77
248.99	408.34	185.5	196.42	197.59	198.82	199.53	200.20	201.02	201.59	202.77
248.99	408.34	185.7	196.42	197.59	198.82	199.53	200.20	201.02	201.59	202.77
249.00	408.36	185.9	196.42	197.59	198.82	199.53	200.20	201.02	201.59	202.77
249.00	408.36	186.1	196.42	197.59	198.82	199.53	200.20	201.02	201.59	202.77
249.01	408.38	186.3	196.90	197.59	198.82	199.53	200.20	201.02	201.59	202.77
250.00	410.00	186.5	198.25	199.05	199.92	200.53	201.15	201.92	202.46	203.59
251.10	411.80	187.0	198.51	199.36	200.27	200.89	201.51	202.29	202.82	203.95
252.40	413.94	188.5	199.03	199.93	200.89	201.50	202.11	202.87	203.40	204.51
254.00	416.56	190.0	199.72	200.63	201.57	202.16	202.73	203.45	203.95	205.02
254.58	417.51	189.4	199.93	200.87	201.84	202.43	202.99	203.70	204.19	205.25
254.85	417.95	189.4	200.00	200.95	201.94	202.52	203.09	203.80	204.29	205.34
255.80	419.51	191.5	200.32	201.29	202.30	202.89	203.45	204.14	204.61	205.60

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