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THE LATE WISCONSIN AND HOLOCENE DEVELOPMENT
OF THE ST. JOSEPH RIVER DRAINAGE BASIN,
SOUTHWEST MICHIGAN AND NORTHERN INDIANA

presented by

Kevin A. Kincare

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of the requirements for the

Ph.D. degree in Geological Sciences

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THE LATE WISCONSIN AND HOLOCENE DEVELOPMENT OF THE ST. JOSEPH RIVER DRAINAGE BASIN, SOUTHWEST MICHIGAN AND NORTHERN INDIANA.

By

Kevin A. Kincare

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ABSTRACT

THE LATE WISCONSIN AND HOLOCENE DEVELOPMENT OF THE ST. JOSEPH RIVER DRAINAGE BASIN, SOUTHWEST MICHIGAN AND NORTHERN INDIANA.

By

Kevin A. Kincare

The St. Joseph River and its tributaries drain approximately 11,137 km2 in southwestern Michigan and northcentral Indiana. It originates from Baw Beese Lake near Hillsdale, Michigan through a total length of 316 kilometers with a head of 158.8 meters above base level in Lake Michigan at Benton Harbor, Michigan.

Evidence for the events that shaped the St. Joseph River drainage basin is seen in the longitudinal profile of the river which can be used as a proxy for events that occurred in the drainage basin given a proper understanding of how disequilibrium in the drainage basin changes the slope of the profile. The longitudinal profile of the St. Joseph River has four nickpoints that divide the profile into 5 distinct sections: a base-level affected section at the mouth, a stream-capture section, a central section where proglacial lakes overflowed to the south, a convex-upward section immediately downstream of flow loss due to stream capture, and a final section that was a former tributary that became the main stem due to the aforementioned stream capture.

The longitudinal profile shows evidence that the St. Joseph River drainage basin was sequentially constructed from the distal areas to the proximal areas. Glacial retreat added sections to the basin in step-wise fashion as the glacier

margin moved basinward. The accepted pattern of regional slope and stream-network development corresponding to a pre-existing structural or stratigraphic pattern in combination with continuous extension via headward erosion does not apply. The drainage basin was constructed in discrete parts during glacial retreat. Each discrete part has a distinct depositional history separate from each subsequently added segment. The relative positions of the ice margins also imparted controls on the volume of meltwater into the distal part of each basin. In addition, the ancestral St. Joseph River contains underlying deglacial terrain originating within the Lake Michigan, Huron/Erie, and Saginaw lobes that all contributed to the final drainage-basin geomorphology.

Early deglacial drainage was initially south toward the Wabash River and later west to the Kankakee River before attaining its final configuration draining to Lake Michigan. Glaciers acted as dams and ponded water in proglacial lakes that overflowed to the south. Post-glacial drainage basin development was controlled by Holocene lake-level fluctuations, stream capture, and the inherited hydraulic gradients of glacial flow regimes. Two major terraces document the final meltwater pulses through the drainage basin both originated from the Huron/Erie lobe. The St. Joseph River did not drain into the Lake Michigan basin until the Calumet phase of glacial Lake Chicago, 1500 years after previously thought.

ACKNOWLEDGEMENTS

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For the aforementioned delays in finishing this tome I can only blame myself. However, people who know me and yet continued to dangle irresistible baubles like book chapters, field trips, journal articles, and teaching opportunities in front of me, should have known better.

This could not have even been attempted without Valerie Haan, to whom I owe everything. She found me at my worst, and allowed me to be my best.

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Chapter 1

INTRODUCTION AND SCOPE OF WORK

The St. Joseph River and its tributaries drain approximately 11,137 km2 (4,300 miles2) in southwestern Michigan and northcentral Indiana (Figure 1). It originates at Baw Beese Lake, one kilometer southeast of Hillsdale, Michigan. The river flows a total of 316 kilometers (196 miles) with a head of 158.8 meters (521 feet) above its base level in Lake Michigan at St. Joseph, Michigan. The St. Joseph River drainage basin encompasses an area that was occupied by three separate lobes of the Laurentide ice sheet (Figure 2), the Lake Michigan, Saginaw, and Huron/Erie lobes.

The St. Joseph River contains five sections with disparate characteristics of gradient, sinuosity, terrace development, depositional history, geomorphology, and discharge. These characteristics have not been researched adequately to elucidate the causes of their distribution throughout the drainage basin. The hypothesis being examined in this dissertation is that a description of the geology of the drainage basin along with construction of the longitudinal profile of the river and its terraces will allow the writer to interpret the development of the drainage basin. A river in equilibrium has a characteristic concave upward profile that is generally due to the decreasing depth/width ratio that causes progressive flattening downstream (Hack 1957, Leopold et al. 1964). Interruptions in a smooth longitudinal profile can be caused by a variety of agents both natural (e.g. resistant outcrops, base level changes) and manmade (e.g. dams, hydraulic mining). A profile interruption is known as "nickpoint" (Thornbury 1969, p. 110).

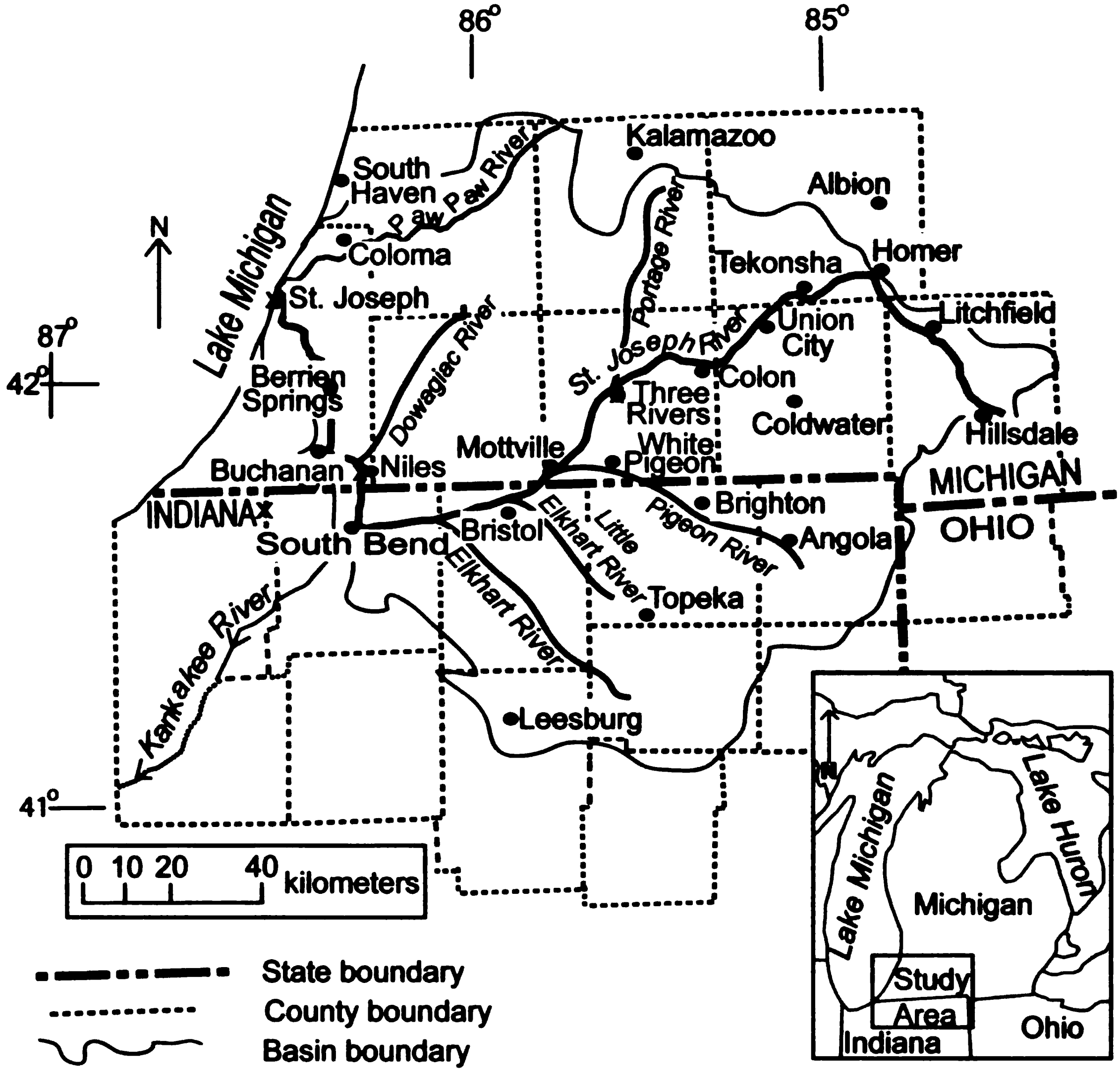

Figure 1. Location map of the St. Joseph River drainage basin.

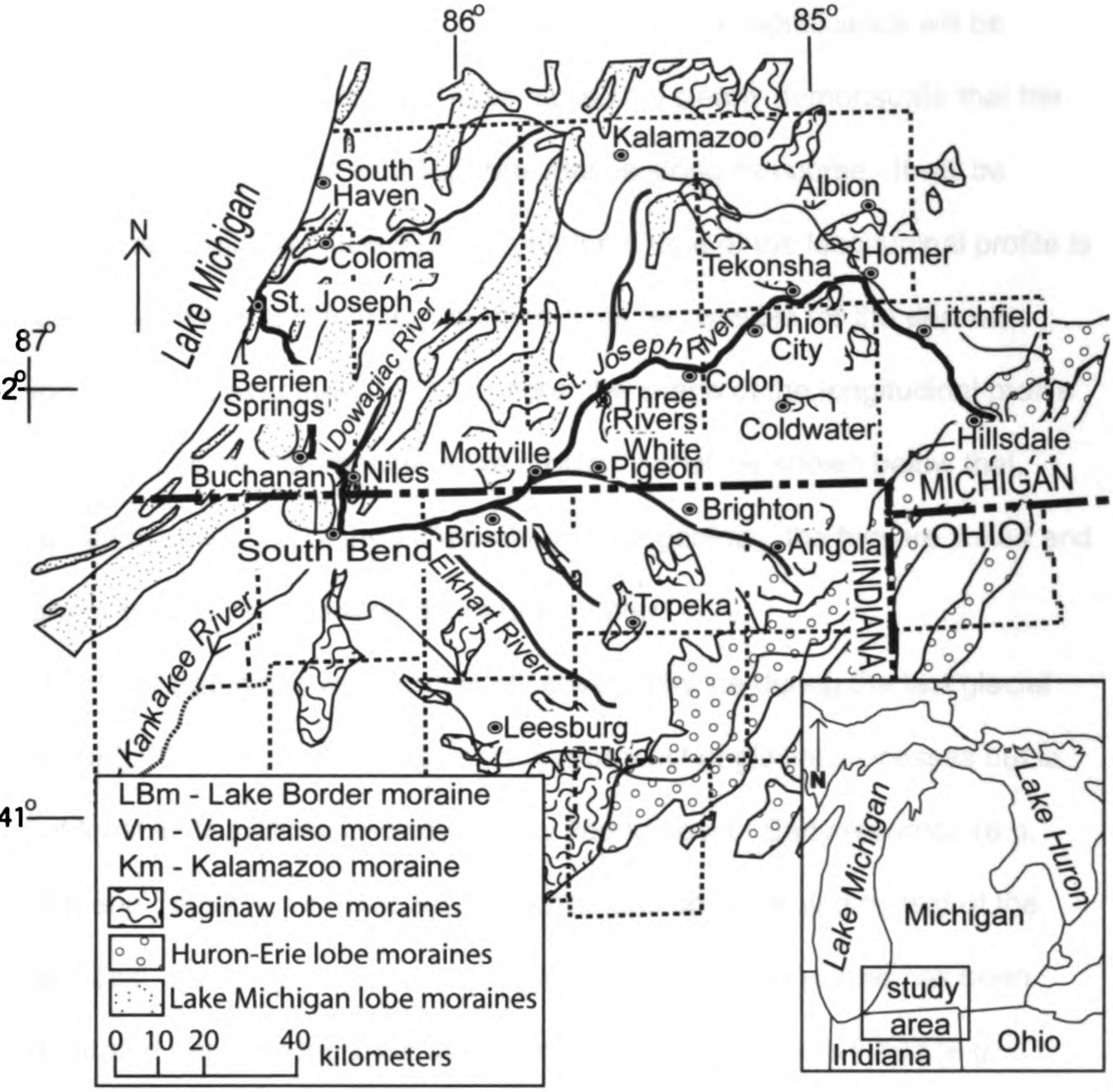

Figure 2. Map of the moraines that occupy the St. Joseph River basin by glacial lobe (after Farrand 1974).

Examining nickpoints in longitudinal profiles can yield important data on the base level and depositional history of a river. The longitudinal profile of the St. Joseph River (Figure 3) contains four nickpoints whose significance will be examined in detail in Chapter 7. These nickpoints clearly demonstrate that the river is not in equilibrium at four separate places along its course. It will be demonstrated throughout these pages that the shape of the longitudinal profile is a valuable predictive tool to interpret the events responsible for the deposits through which the river flows. Furthermore, the shape of the longitudinal profile indicates the type of disturbance to the profile. It shall be shown below that, through an examination of the nickpoints and the geology, the two are linked and the former can be used as a predictor of the latter.

The drainage basin was completely covered by ice during the last glacial maximum (LGM) and therefore is not only a product of glacial processes but is also, at least in its surficial geology, a very young basin. Early research (e.g. Leverett and Taylor 1915) described synchronous advance and retreat of the glacial lobes. Recently, asynchronous glacial advance and retreat has been demonstrated between the Lake Michigan and Saginaw glacial lobes (e.g. Kehew et al. 1999).

Initial deglaciation of the basin probably began when the Saginaw lobe retreated from the LGM to the southern margin of the drainage basin about 20,000 calendar years B.P. As deglaciation proceeded, different ice-margin locations controlled meltwater sources, outlets, and the configuration of depositional environments. The drainage basin therefore, presents a unique

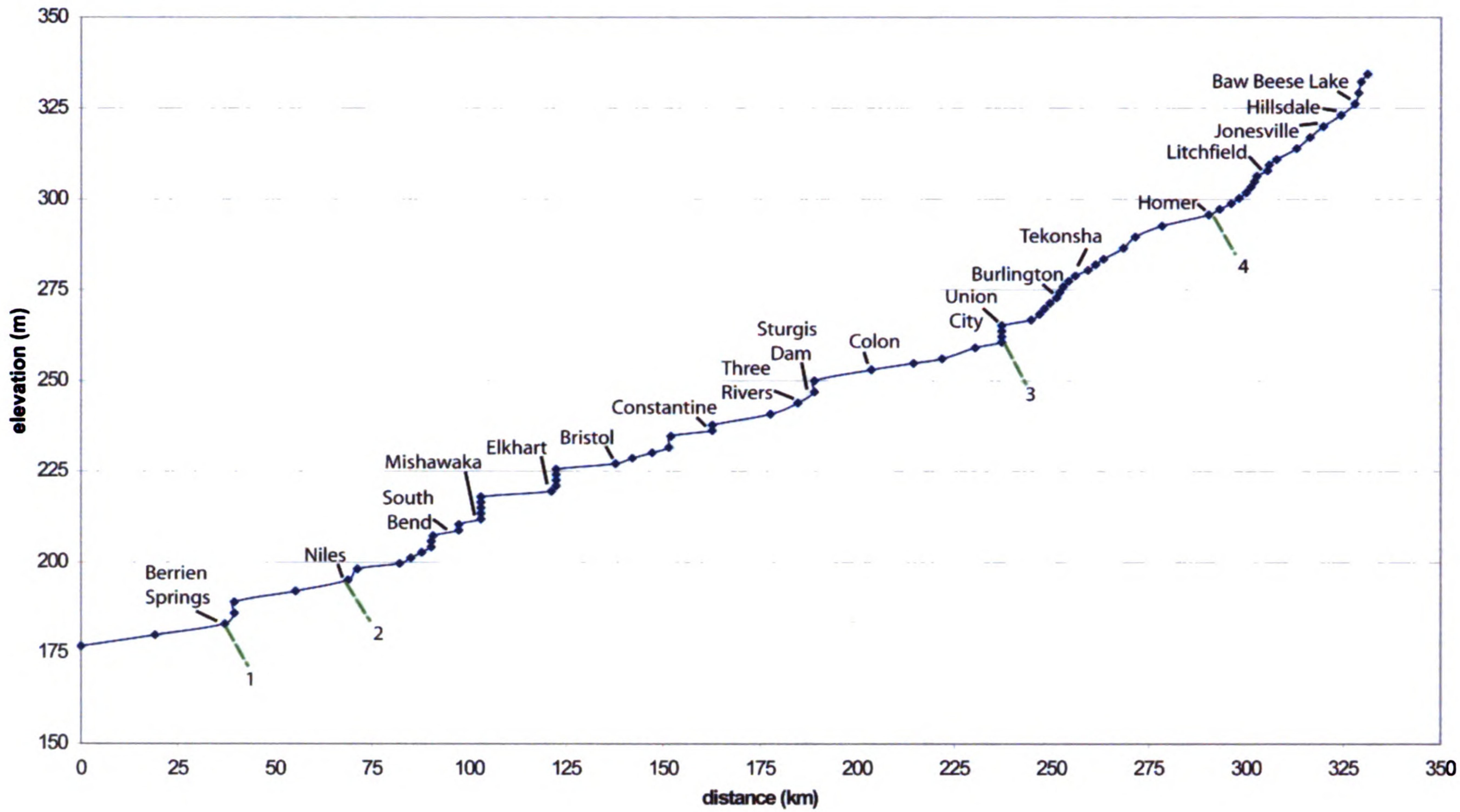

Figure 3. Longitudinal profile of the St. Joseph River. Numbers show the location of the nickpoints.

opportunity to study the position and dynamics of the respective lobes through the contribution of meltwater and resultant outwash and lacustrine deposits throughout the basin.

Little more is known about the St. Joseph River drainage basin than was briefly described by Leverett (1908) and Leverett and Taylor (1915). Other than the recognition that the basin was once part of the Kankakee River drainage basin (Leverett 1908), the basin contained glacial Lake Dowagiac (Leverett 1908), and a description of landforms discernable at a scale of 1:1,000,000 (Leverett and Taylor 1915), few other details were available. This is unfortunate because this northern outlet drainage basin had extensive boundaries with all three glacial lobes that were responsible for the glacial landforms that exist throughout southern Michigan, western Ohio, Indiana, and eastern Illinois. This study was undertaken because a careful examination of the St. Joseph River drainage basin would lead to important discoveries concerning the cross-cutting relationships between glacigenic deposits of the three glacial lobes and therefore yield a better chronology of deglaciation than currently exists for these glacial lobes. The availability of 7.5 minute U.S. Geological Survey quadrangles and 30 m digital elevation model (DEM) data also provide an unprecedented ability to reexamine landforms and deposits previously thought to be properly interpreted. These will be discussed in Chapter 7.

Lakes are among the most ephemeral of geologic features (Thornbury 1969). Lacustrine conditions are often associated with glaciation, particularly during glacial retreat when drainage is blocked by ice on one side and constructional landforms of the previous ice stand on the other. Therefore, the record of glacial retreat is important for establishing a sequence of glacial-lake levels upon which a hypothesis can be based for the origin of observed erosion and sedimentation features and their cross-cutting relationships. The series of glacial lakes also provide the base level for fluvial systems that develop in a drainage basin and knowledge of both further enhance our ability to provide an event chronology. Chapter 3 provides a summary of the existing chronology of glacial advances and retreats during the Pleistocene epoch. These advances and retreats controlled the outlets (and therefore the levels) of the proglacial lakes. Chapter 4 presents a summary of the current state of knowledge on the extent and surface altitude of the proglacial and post-glacial Great Lakes. This is the record of base-level changes for the St. Joseph River.

Chapter 5 presents a literature review summarizing the concept of base level, affects of base-level changes to fluvial geomorphology, and a summary of sedimentary features expected in deglaciated fluvial environments subjected to base-level changes. This chapter also explains why the present state of knowledge is inadequate to explain the development of drainage basins in glaciated terrain. Chapter 5 presents new and compiled information on the geology of the St. Joseph River drainage basin based on an examination of nickpoints, sedimentology, and geomorphology. Chapter 6 summarizes and discusses current theory regarding the development of drainage networks and how it is inadequate to fully understand the manner in which glaciated drainage basins are constructed.

Time reported in this text is in calendar years while that reported in parentheses is in radiocarbon years BP (Before Present). Conversion from the last 47,000 radiocarbon years to calendar years was achieved using a radiocarbon calibration curve published by Fairbanks et al. (2005). Conversion from 47,000 to 50,000 radiocarbon years to calendar years was achieved by extrapolation of the radiocarbon calibration curve.

Chapter 2

METHODOLOGY

The process of adjusting to a new base level often causes an abrupt stream gradient change or "interrupted profile" (Thornbury 1969, p. 109). A nickpoint is the location on a longitudinal profile where the gradient changes. Where nickpoints are not due to geologic or man-made structure, they typically mark the headward or upstream limit of regrading to a new base level (Butcher 1989). The presence and location of nickpoints in rivers can give an indication of base-level history, provided that equilibrium has not been reached throughout the basin.

Longitudinal profiles of the St. Joseph River were drawn for both channel and valley distance vs. altitude by using 7.5 minute USGS quadrangles. Conversions were made for incompatible scales and measurement systems (e.g. 1:24,000 vs. 1:25,000 scales and english system vs. metric system). Profiles were done by hand using a Brunton rolling map-measuring tool and checked against the distance-measuring tool in ArcView 3.2. Topographic maps were downloaded from the Michigan Center for Geographic Information (www.michigan.gov/cgi) and projected in Michigan GeoRef meters. Graphs were plotted using a computer spreadsheet. Locations of man-made structures were noted on the profile.

The thickness and character of the cut and fill section in a river valley is also an indicator of the duration and amplitude of the base-level event. The St. Joseph River carved a 55 m (180 ft) deep valley which was subsequently filled with 43 m (140 ft) of transgressive, valley-fill sediments (Kincare 2000). The fill sediments of the terraces will be described in order to separate the cut and fill sequences from other periods of lower lake levels from the Chippewa low.

Cores of the valley-fill sequence were obtained with split-spoon rods and direct-push methods. Cores were logged for grain-size trends, presence of organics (for datable material) and any evidence of depositional environment. Cores were drilled in four locations along the axis of the drowned valley of the St. Joseph River to recover sediments and place them in a chronological context.

Mapping of glaciofluvial and glaciolacustrine sediments in the St. Joseph River drainage basin was performed to:

- show the areal distribution of surficial geologic units that cover areas greater than 20 hectares,
- generate a lithologic description of all map units and their expected range of variability.

The surficial geologic map of glaciofluvial and glaciolacustrine sediments was produced by field investigation of surficial materials observed in natural exposures, road cuts, building and construction excavations and hand-auger holes dug by the investigator. In addition, use was also made of aerial photography, topographic maps, and county soil surveys.

Chapter 3

GLACIAL HISTORY OF THE GREAT LAKES REGION

Ever since T.C. Chamberlin (1883) published the first map to show the extent of drift (glacial deposits) in the mid-continent region of North America, glacial geologists have attempted to divide and subdivide the drift in order to unravel the glacial history of the region. This work has lead to several time-stratigraphic classifications of the last (Wisconsin) glaciation, most of which apply to specific areas of the region. For example, Willman and Frye (1970) developed a widely accepted classification for the Lake Michigan-Illinois area (Table 1) based on the time-sequence stratigraphy of drift units, buried soils and ^{14}C (radiocarbon) age determinations (chronostratigraphy). In their classification, the last glacial stage is divided into five substages, some representing glacial and some nonglacial conditions. Another widely accepted classification (Table 1), by Dreimanis and Karrow (1972), applies to the eastern-northern Great Lakes area and is also based on time-sequence stratigraphy of drift units and ^{14}C age determinations, but not buried soils. In this classification the last glacial stage is divided into three substages that are further subdivided into stadials representing glacial conditions and interstadials representing nonglacial conditions.

A difficulty inherent with the above classifications is that time intervals (stages, substages, stadials, and interstadials) have time-parallel boundaries that do not necessarily correspond to the timing of events across a large area. For example, a glacial advance in the eastern Great Lakes area during the Port Bruce Stadial may have resulted in a ~1,500 year interval of ice cover in southern

Table 1. Illinois and Ontario Chronostratigraphic Classifications[1]

Illinois (Willman and Frye 1970)	Ontario (Dreimanis and Karrow 1972)
Wisconsin Stage	Wisconsin(an) Stage
Valderan Substage[2]	late Wisconsin(an) Substage
	Driftwood Phase or Stadial
	North Bay Interval or Interstadial
	Valders Phase or Stadial
Twocreekan Substage	Two Creeks Interval or Stadial
Woodfordian Substage	Port Huron Phase or Stadial
	Mackinaw Interval or Interstadial
	Port Bruce Phase or Stadial
	Erie Interval or Interstadial
	Nissouri Stadial
Farmdalian Substage	middle Wisconsin(an) Substage
	Plum Point Interstadial
Altonian Substage	Cherrytree Stadial
	Port Talbot Interstadial
	early Wisconsin(an) Substage
	Guildwood Stadial
	St. Pierre Interstadial
	Nicolet Stadial

1: After Karrow et al. (2000)
2: Renamed Greatlakean Substage by Evenson et al. (1976)

Ontario, but the same advance may have resulted in only a ~500 year interval of ice cover in northern Ohio. Likewise, a temporary retreat of the glacier margin in the Lake Michigan-Illinois area during the Farmdalian Substage may have resulted in a soil-forming interval in northern Illinois lasting several 1000 years, but this same soil-forming interval may have lasted less than 1000 years in eastern Wisconsin. To overcome this problem Johnson et al. (1997) and Karrow et al. (2000) have recently proposed a time-stratigraphic classification system for the last glaciation that includes diachronic-time divisions and better represents the timing of events associated with what we know was a dynamic ice margin.

The Great Lakes before glaciation

Little is known about climate in the eastern part of the mid-continent prior to Late Quaternary glaciations because much of the geologic record has been removed by ice-sheet erosion or is now deeply buried beneath glacial sediments. However, some clues remain. For example, in northern Indiana remnants of Pliocene (5.3-1.8 Ma) frogs, pond turtles, fishes, birds, snakes, and small and large mammals have been found in sinkholes buried beneath glacial deposits (Holman 1998, Farlow et al. 2001). Collectively, these remnants indicate a dry, open, prairie-like or savanna environment. In other areas, analyses of the bedrock topography buried beneath glacial deposits indicate that during the Pliocene major rivers flowed where the Great Lakes are today (Figure 4).

Age and extent of the glacial deposits

It is not known when the eastern part of the mid-continent was first glaciated. However, evidence from Ohio, Indiana and Illinois suggests that

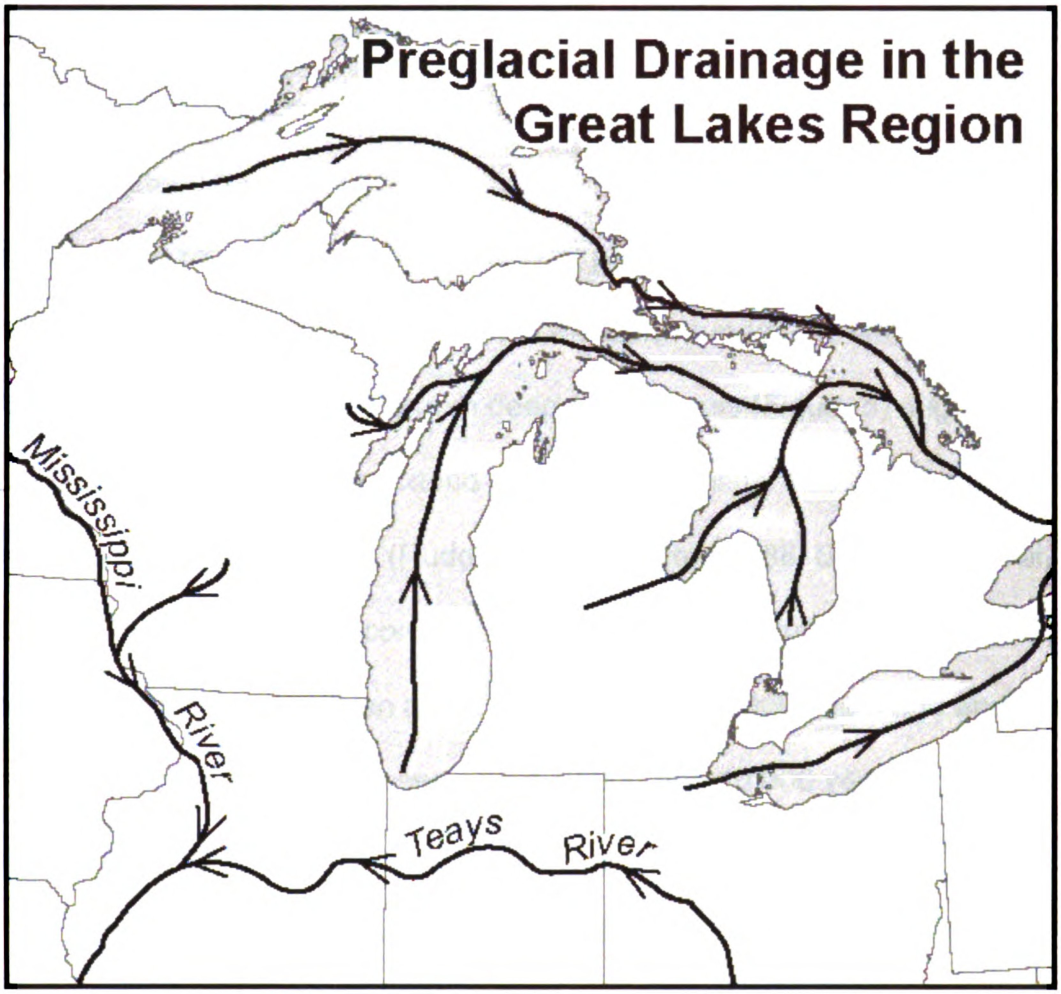

Figure 4. Preglacial drainage of the Great Lakes region during the late Cenozoic before glaciation. Probable pattern is based on known bedrock structures and geophysical reconnaissance of buried valleys. After Horberg and Anderson (1956), Hough (1958), and Flint (1971).

sometime prior to 780,000 years ago glacier ice extended, over much of the Great Lakes area and dammed several northward-draining valleys in Illinois, Indiana, and Ohio, forming lakes (Fullerton 1986, Johnson 1986). Paleomagnetic data from some of the sediments deposited in these lakes are magnetically reversed and therefore must predate 780,000 years ago (Matuyama-Brunhes boundary) - when the Earth's magnetic field last reversed. Also, oxygen-isotope records from deep ocean cores (Figure 5) suggest that glaciations large enough to extend into the mid-continent occurred many times over the last 800,000 years (Ruddiman and Raymo 1988, Shackleton et al. 1988). The ocean-core records also reveal that the penultimate glaciation occurred sometime between about 202,000 and 132,000 years ago and that the last glaciation occurred between about 79,000 and 10,000 years ago (Ruddiman and Raymo 1988, Shackleton et al. 1988).

The penultimate glaciation in the eastern part of the mid-continent is generally referred to as the Illinois glaciation (Chamberlin 1896, Leverett 1899) and corresponds to Marine Isotope Stage (MIS) 6 of the marine oxygen-isotope record (Follmer 1983, Curry and Pavich 1996). It probably reached its maximum extent about 150,000 years ago and is represented by an extensive drift sheet, in places exceeding tens of meters in thickness (Figure 6). Because glacial ice is usually channeled through lowlands and major river valleys, the southern margin of the drift sheet is generally lobate, except where uplands of the Appalachian Mountains obstructed ice flow. Where exposed, the drift sheet is highly weathered with soil development up to several meters in thickness (Ruhe 1974).

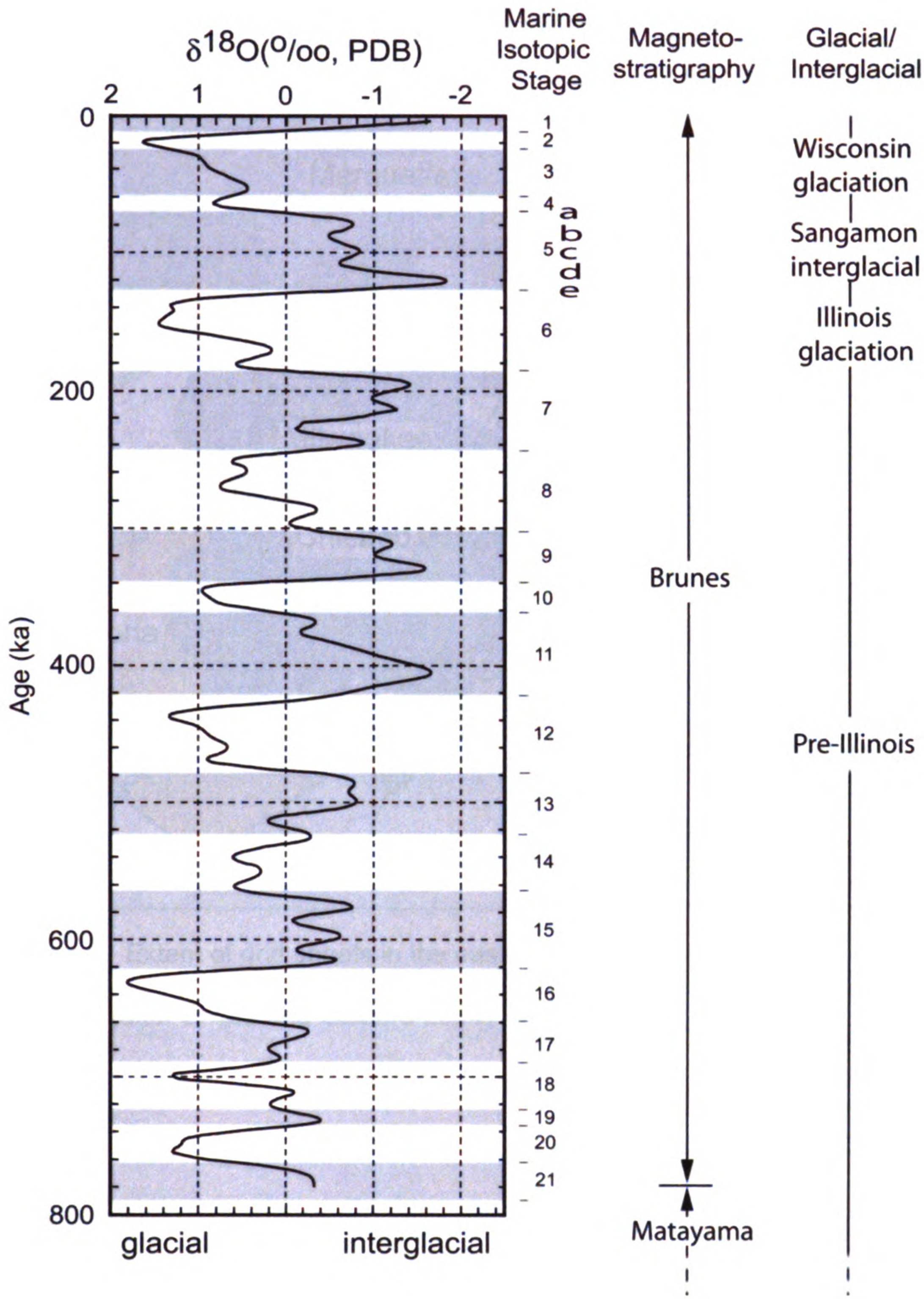

Figure 5. Record of $\delta^{18}O$ in deep-sea cores corresponding to global climate.

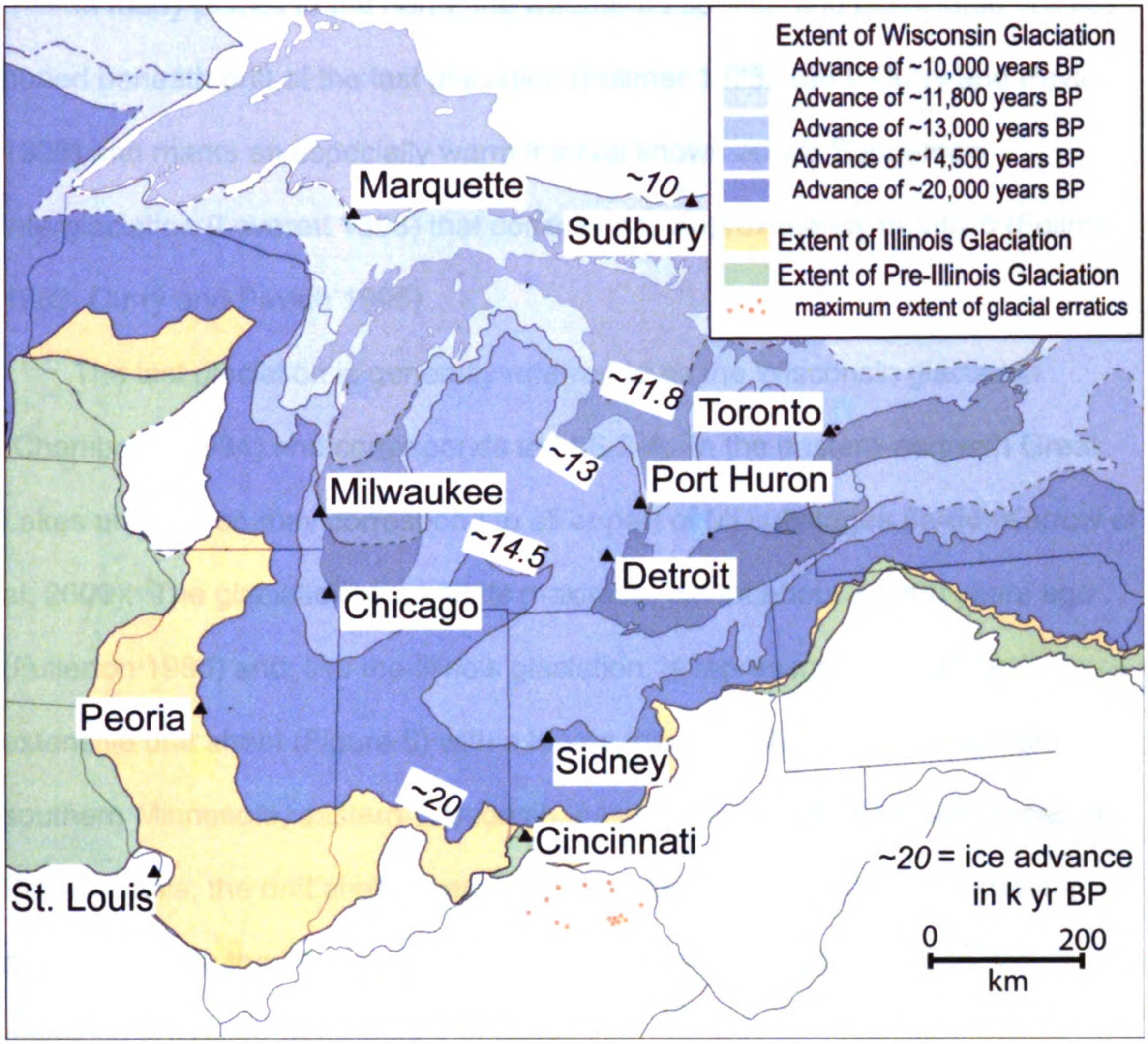

Figure 6. Extent of drift sheets in the eastern part of the mid-continent of North America (after Ehlers and Gibbard 2004).

In many places to the north, the weathered surface and associated soil lies buried beneath drift of the last glaciation (Follmer 1978, 1983, Curry and Pavich 1996) and marks an especially warm interval known as the Sangamon interglaciation (Leverett 1898) that corresponds approximately to MIS 5 (Follmer 1983, Curry and Pavich 1996).

The last glaciation is generally referred to as the Wisconsin glaciation (Chamberlin 1894) and corresponds to MIS 2-4. In the eastern-northern Great Lakes area it also may correspond to all or part of MI substages 5a-5d (Karrow et al. 2000). The glaciation reached its maximum extent about 23,900 years ago (Fullerton 1986) and, like the Illinois glaciation, is represented by a thick, extensive drift sheet (Figure 6) with a lobate margin. With the exception of southern Minnesota, eastern Wisconsin, eastern Ohio, New York and places in Pennsylvania, the drift sheet does not extend as far south as the drift sheet associated with the Illinois glaciation. It also is not as weathered, with soil development rarely more than 1 m thick.

Drift sheets older than those associated with Wisconsin and Illinois glaciations also occur in the eastern part of the mid-continent (Figure 6). However, their number and ages are uncertain (Hallberg 1986); therefore they are not attributed to particular glaciations but collectively referred to as Pre-Illinoian. In parts of Kansas, Nebraska, Wisconsin, Iowa, and Missouri as well as locally in southwestern Ohio, one or more Pre-Illinoian drift sheets extend beyond the limits of the Wisconsin and Illinois glaciations and are characterized by a very high degree of weathering (Willman and Frye 1970). In Pennsylvania this drift is generally very thin and patchy and often consists of only a few scattered erratic boulders on bedrock. Highly weathered erratic boulders also have been reported in northern Kentucky and their presence suggests that one or more Pre-Illinoian glaciations must have extended into northern Kentucky (Teller and Goldthwait 1991).

Temporal record of the Wisconsin glaciation

The diachronic classification of the Wisconsin glaciation shown in Figure 7 was first proposed by Johnson et al. (1997) and Karrow et al. (2000) and has been pieced together from a number of stratigraphic sections. The smallest time division in the classification is a phase. It usually is applied locally and defined by a referent such as a unit of till, outwash, lacustrine sediment, loess, peat, or buried soil. The next higher order time division is a subespisode - applied more regionally and defined by one or more referent units. The highest order time subdivision is an episode. It applies throughout an entire region, in this case the eastern part of the mid-continent, and is defined by one or more referent units. Of particular importance are referent units such as buried soils, organic beds and fossiliferous sediments because they can reveal valuable information about climatic conditions during ice-free intervals. Also, they can be dated using the ^{14}C method as far back as about 53,000 years ago if they contain organic materials such as wood, plant fragments, peat, and shells (Fairbanks et al. 2005). The discussion that follows highlights the important events of each of the major time units in the eastern part of the mid-continent, as shown in Figure 7.

Sangamon Episode

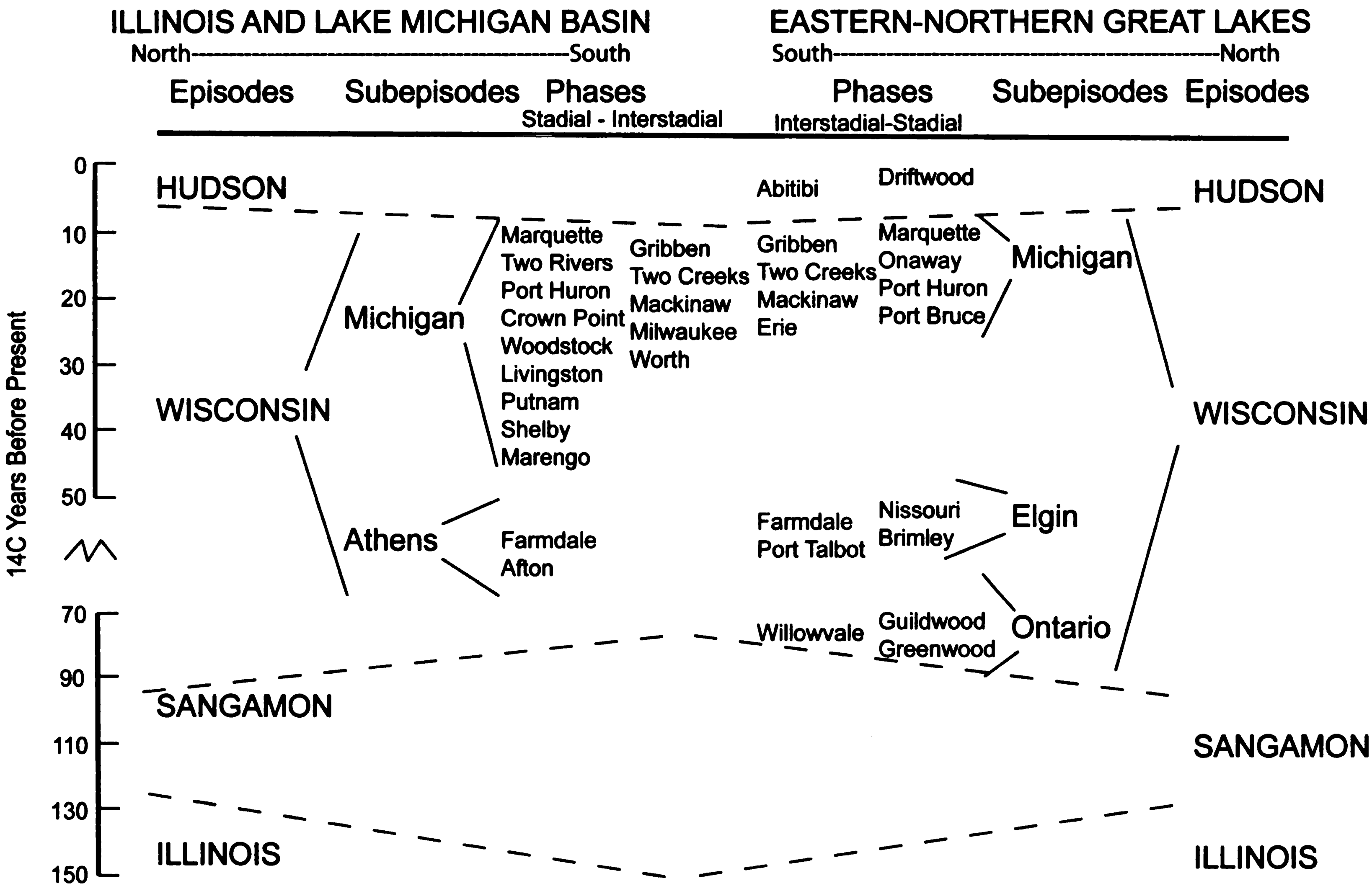

Figure 7. Comparison of time-distance diagrams for the eastern and western Great Lakes regions.

The Don Formation exposed near the city of Toronto, Ontario serves as the referent for the Sangamon Episode (last interglacial) in the eastern-northern Great Lakes area and probably represents a time span limited to MIS 5e - about 130,000-115,000 years ago (Karrow et al. 2000). It consists of fluvio-lacustrine sediments and includes fossils that indicate a climate as warm or warmer than present (Terasmae 1960, Eyles and Williams 1992). In the Lake Michigan-Illinois area, on the other hand, the Sangamon Geosol (buried soil) exposed near the city of Athens in central Illinois serves as the referent for the Sangamon Episode (Johnson et al. 1997). It probably represents a time span limited to MIS 5 - about 130,000-75,000 years ago (Follmer 1983). Pedological analyses of the geosol also suggest that the climate in the Lake Michigan-Illinois area during the early part of the Sangamon Episode was warmer than present (Follmer 1983). The lateral extent of the geosol is considerable and correlative buried soils have been reported throughout much of the upper Mississippi River basin (Curry and Follmer 1992, Hall and Anderson 2000), including northern Illinois. Recently, ^{10}Be data suggest a correlative buried soil in northern Illinois developed from at least 155,000 to 55,000 years ago (Curry and Pavich 1996).

Wisconsin Episode

Eastern-Northern Great Lakes area

In the eastern-northern Great Lakes area, the Wisconsin Episode (last glaciation) is divided into three subepisodes: Ontario, Elgin and Michigan (Figure 7). The Ontario is the oldest and is named after the province of Ontario where the best stratigraphic evidence for the subepisode is located (Karrow et al. 2000).

It includes three phases with material referents in the Toronto area, mainly tills and glaciolacustrine sediments, and represents a time of cold climate and extensive ice cover, probably corresponding to MIS 4 and perhaps part or all of substages 5a-5d (Karrow et al. 2000; Figure 7). During the Greenwood Phase, ice extended well into the Ontario basin but during the Willowvale Phase it withdrew, probably into the eastern end of the basin (Karrow et al. 2000). During the Guildwood Phase the ice margin readvanced westward across the Ontario basin and possibly into the north-central part of the Erie basin (Dreimanis 1992). Little is known about where the ice margin stood in the Huron basin during the early part of the Greenwood and Willowvale phases, but during the Guildwood Phase it probably extended into part or all of Georgian Bay.

To date, no Ontario Subepisode sediments have been reported in Michigan. However, there are reports of buried organic material that have yielded infinite ^{14}C dates (Eschman 1980, Winters et al. 1986). Some of this material may have accumulated during the Ontario Subepisode when parts of southern Ontario were covered by glacier ice, or earlier, e.g., during the Sangamon Episode.

The Elgin Subepisode (Figure 7) is named after Elgin County, on the north shore of Lake Erie. Here, the best evidence for the episode occurs, mainly in the form of lacustrine and organic sediments (Karrow et al. 2000). The Elgin Subepisode includes the Port Talbot, Brimley and Farmdale phases and represents a time of moderated (warmer) climate and significant ice contraction. Of particular significance is the Port Talbot Phase, defined by pollen- and macrofossil-bearing lacustrine sediments located near Port Talbot along the north shore of Lake Erie. Pollen from this site indicates initial warm and dry climatic conditions (but cooler than an interglacial), followed by climatic cooling and possibly a forest-tundra environment (Berti 1975). Peat and wood found within the sediments have yielded a number of calibrated ^{14}C ages ranging from about 51,200 to 47,300 years ago (Dreimanis et al. 1966, Dreimanis and Karrow 1972). Calibrated ^{14}C ages of >49,000 years also have been obtained from a buried soil and overlying fossiliferous sediments in the city of Guelph (Karrow et al. 1982). Analyses of the fossils indicate a cooler and dryer climate than present; it is possible that the buried soil and sediment is associated with the Elgin Subepisode, or alternatively the Ontario Subepisode or some earlier time (Karrow et al. 1982).

In Michigan, sites of buried organic material consisting of wood, muck, and marl have yielded calibrated ^{14}C ages ranging from about 51,900 to 42,300 years ago (Eschman 1980, Winters et al. 1986). Of particular interest is material reported near Kalkaska in northwestern Lower Michigan (Winters et al. 1986). It includes a pollen record that shows vegetation evolving from a cold, open forest into a closed boreal forest about 40,300 years ago and suggests a climate possibly influenced by an ice margin in southern Ontario (Winters et al. 1986). Also of interest is organic material exposed in the banks of the Black River in southeastern Lower Michigan, which has been dated at about 51,900 years (Eschman 1980). It contains a mixed terrestrial and aquatic fauna that clearly has boreal affinities (Karrow et al. 1997). These buried organic materials confirm that Lower Michigan was not covered by ice during the Elgin Subepisode (Port Talbot phase), although the ice margin was likely not far to the north.

The Michigan Subepisode (Figure 7) is the last subepisode of the Wisconsin Episode and is so named because the landscape of Michigan is dominated by glacial sediments deposited near the end of the Wisconsin glaciation (Johnson et al. 1997, Karrow et al. 2000). The subepisode includes 11 phases defined mainly by till units found in Michigan, Ohio, and Ontario. It marks a period of cold climate and maximum expansion of ice, followed by warming and ice-margin retreat. The Nissouri Phase at the beginning of the subespisode is of particular significance because it was during this phase that the ice margin advanced rapidly out of the Erie and Huron basins to cover all of Ontario and Michigan, as well as much of Indiana and Ohio. Upon reaching its most southerly position near Cincinnati, Ohio, the margin overrode and buried a forest that has yielded several calibrated ^{14}C ages that average about 23,700 years ago (Lowell et al. 1990).

Following the ice advance during the Nissouri Phase, the ice margin retreated northward towards the Erie and Huron basins, only to periodically readvance multiple times. The exact timing of these readvances is not well known, mainly because of the paucity of in situ datable organic material, especially wood, during this cold climatic interval. A notable exception, however, is the readvance associated with the Onaway Phase. It buried a bryophyte (moss) bed near Cheboygan, Michigan that has yielded several calibrated ^{14}C ages averaging about 13,600 years ago (Larson et al. 1994).

Lake Michigan-Illinois area

In Lake Michigan-Illinois area, the Wisconsin Episode is subdivided into the Athens and Michigan Subepisodes (Figure 7). The Athens Subepisode includes the Alton and Farmdale phases and is based on a record of loess and buried soils exposed near the city of Athens in central Illinois (Johnson et al. 1997). The Michigan Subepisode has nine phases based mainly on till units found in Illinois, Wisconsin, and northern Michigan (Hansel and Johnson 1992). In general, the phases of the Michigan Subepisode are generally concurrent with those in the eastern-northern Great Lakes area (Karrow et al. 2000).

The Alton Phase was a time of transition from interglacial to periglacial conditions in central Illinois and its referent material is the Roxana Silt (Johnson et al. 1997). Little is known about the ice margin position during the Alton Phase, but it may have extended as far south as north-central Wisconsin and central Minnesota, where it contributed massive quantities of silt into the upper-Mississippi drainage system (Grimley 2000). The following Farmdale Phase was a time of significant reduction in loess supply, probably because of ice margin retreat from the upper-Mississippi drainage system (Grimley 2000) and a concurrent return to a milder climate. The Farmdale Geosol and Robein Silt exposed along the banks of Farm Creek, near Peoria, Illinois (Frye and Willman 1960), serve as the referent for the Farmdale Phase. Based on radiocarbon dates at and near the exposure, the Farmdale Geosol is believed to have developed from about 33,400 to 26,500 years ago (Willman and Frye 1970, Johnson 1976). The geosol also is found at other localities in Illinois and western

Indiana (Curry and Follmer 1992, Hall and Anderson 2000); in eastern Indiana and Ohio it is referred to as the Sidney Soil (Hall and Anderson 2000).

At a number of localities in Lower Michigan buried organic material, in places meters thick, has yielded calibrated ^{14}C ages ranging from 51,000 to 28,700 years ago (Eschman 1980; Winters et al. 1986). The presence of this material indicates that much of the Lower Peninsula was cool but ice-free during the Alton and Farmdale phases and that climate at that time was probably wetter than in central Illinois.

As in the eastern-northern Great Lakes area, the Michigan Subepisode in the Lake Michigan-Illinois area marks a period of cooling and expansion of ice, eventually followed by general warming and retreat of the ice margin. Maximum glacier-ice expansion occurred early in the Michigan Subepisode during the Shelby Phase. At that time the ice margin advanced out of the Lake Michigan basin south to as far as Peoria, Illinois, where it overrode and buried a moss layer that has yielded a calibrated ^{14}C age of about 23,500 years ago (Follmer 1979).

After advancing to Peoria, the ice margin began to slowly retreat northward, though briefly readvancing a number of times. The exact timing of each readvance is uncertain because of the paucity of in situ datable organic materials. However, the timing of the readvance associated with the Two Rivers Phase is well documented because advancing ice overrode and buried a forest near Two Creeks Wisconsin that has yielded an average calibrated ^{14}C age of about 13,600 years (Broecker and Farrand 1963, Leavitt and Kalin 1992, Kaiser 1994).

The timing of the readvance associated with the Marquette Phase is also well established, because glacial sediments associated with that readvance buried a forest near Marquette, Michigan that has yielded an average calibrated ^{14}C age of about 11,500 years (Lowell et al. 1990, Pregitzer et al. 2000). The last known ice readvance was the Cochrane readvance in northern Ontario about 9,400 years ago. After this, the ice sheet split into two large remnants east and west of Hudson Bay and was entirely gone by about 6,800 years ago (Dyke and Prest 1987).

Hudson Episode

Post glacial time is represented by the Hudson Episode (Figure 4), named so because Hudson Bay is dominated by marine, fluvial and paludal sediments deposited since deglaciation (Johnson et al. 1997, Karrow et al. 2000). As yet, there is no particular referent for the Hudson Episode but a possible candidate is marine sediments associated with the Tyrrell Sea that occupied Hudson Bay after retreat and breakup of the last ice sheet (Shilts 1984, Dredge and Cowan 1989). In the eastern part of the mid-continent, deposits associated with the Hudson Episode include lacustrine sediments deposited in the Great Lakes since deglaciation (Colman et al. 1994b, Rae et al. 1994), as well as within smaller inland lakes. Other deposits include loess, dune sand, fluvial deposits, and organic accumulations in swamps and bogs.

Chapter 4

RECORD OF PROGLACIAL AND POST-GLACIAL LAKES

Origin and development of lake basins

The Great Lakes are one of Michigan's most distinctive geographic features. The familiar shape of the Great Lakes is however, a recent phenomenon. Their present form is the result of a number of factors such as glacial erosion and deposition, isostatic depression and subsequent rebound due to glacial-ice load, distribution of glacial meltwater, and changing lake outlets. The shape and location of each Great Lake has been largely determined by the underlying geology. Most of the bedrock beneath each lake basin is easily-eroded Paleozoic sedimentary deposits with the exception of Lake Superior which is a structural basin underlain by complex Middle Proterozoic rocks (Hough 1958). Therefore, prior to Late Cenozoic Ice Age, each lake basin was probably already a river valley draining to an ancestral St. Lawrence River (Figure 4). The structural bedrock highs of the Kankakee arch south of Chicago, and the Findley arch on the west side of Lake Erie, would have effectively prevented the pre-glacial Great Lakes drainage basin from draining south to the Mississippi River.

Recent evidence on climate, gleaned from cores of sea-floor sediment and ice caps, indicate that at least 40 separate glaciations occurred during the last 2.75 million years (Ruddiman and Raymo 1988). Each glaciation that was extensive enough to reach Michigan probably further eroded the lake basins and altered the Great Lakes. Most of the glacial sediments in Michigan were deposited during the Wisconsin Episode and the majority of these were deposited during the final phase of deglaciation (<20,000 years ago). Older glacial sediments are found to the south of Michigan in Illinois, Indiana, and Ohio, demonstrating that earlier glaciations advanced well to the south of Michigan. However, we have not found any definitive evidence yet for older glacial deposits in Michigan, although this may change with advances in dating techniques and additional field research. Despite the glaciers being prone to erosion each time they covered Michigan, it still has some of the thickest glacial deposits in North America, up to 365 m thick in northern lower Michigan (Rieck and Winters 1993, Soller 1998). Given this fact, further discussion below will center on deposits of the late Wisconsin glaciation.

Glaciers and lakes

Leverett (1899) and Leverett and Taylor (1915) published the most detailed compilations of field data on glacial-lake phases in the Great Lakes basin. Later chronologies using new and reinterpreted data have been compiled by Hough (1958), Fullerton (1980), Karrow and Calkin (1985), Schneider and Frasier (1990), and Ehlers and Gibbard (2004). The implications of this work for questions of fluvial responses to base-level change include looking at how far inland the effect of base level is traceable. This can also be extended to questions of sediment load and the formation of low-stand deltas.

Three major lobes of the Laurentide ice sheet moved across and covered Michigan from out of the overdeepened lake basins and are named after those basins, the Lake Michigan, Saginaw, and Huron/Erie lobes (Figure 8). The farthest extent occurred during the Last Glacial Maximum (LGM) and was well to the south of the St. Joseph River basin in Indiana and Ohio (Figure 9). The ice at the LGM began to retreat around 21.5 k years ago. Evidence from the cross-cutting relationships of former ice margins and related outwash shows that the lobes did not move synchronously with each other (Mickelson et al. 1983, Kehew et al. 1999). The Saginaw lobe probably retreated from south-central Michigan first, leaving fine-grained deposits in the St. Joseph River valley, from a lake that was trapped by the Lake Michigan and Huron/Erie lobes which were still much farther south (Figure 10). At this time, before the Great Lake basins opened up, meltwater drainage was into the Mississippi River system via the Wabash River and the Kankakee River.

As noted in chapter 3, the retreat of the ice from the LGM did not proceed as a long, steady retreat but was uneven and punctuated by a series of rapid retreats and readvances. This complicated series of events led to numerous proglacial and post-glacial lakes within each Great Lake basin. At times these lakes were separate and at other times they were confluent or connected by river channels.

Lakes can produce a definitive set of features based on particular types of depositional and erosional patterns. The low energy environments of lakes generally lead to fine-grained deposits, e.g., silt and clay, throughout most of the area they cover. Two major exceptions to this are shoreline areas (high energy due to wave action) and river mouths (where deltas are typically built). A lake, by

Figure 8. Map of ice margin about 19,000 years ago, showing the three main ice lobes that covered Michigan and the Saginaw lobe reentrant.

Figure 9. Farthest extent of Late Wisconsin ice advance. This occurred about 23,000 yrs ago in Illinois and Ohio, but later at other parts of the Laurentide ice margin.

Figure 10. Map of ice margin about 18,000 years ago, showing the Saginaw reentrant with meltwater drainage overflowing to the Wabash and Kankakee Rivers.

definition, has the same surface altitude at all points. On this basis, shorelines from ancient lakes can be traced from place to place, even if intervening parts of the shoreline no longer exist due to erosion or burial. Deltas can also show a former lake-surface altitude where a river entered a lake, as well as showing the off-shore connection to sediments of deeper, quiet-water deposits. Lakes have the power to erode as well. Many shorelines are not traced by beach deposits, but by wave-cut bluffs (strandlines) eroded into adjoining headlands, as well as their spits which may contain the eroded material. River channels that are now dry or have rivers too small to have cut the valley it inhabits provide evidence of old glacial-lake outlets. Each bit of evidence, separately or in combination, is used to map ancient lake margins, while their cross-cutting relationships tell us the order in which they occurred.

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Early lakes of the Lake Erie/Huron basin

The history of the lakes in the Erie and Huron basins are so closely tied together that they must be discussed together. Retreat of the Huron/Erie lobe from the Fort Wayne moraine around 17,400 years ago formed glacial Lake Maumee, the first important proglacial lake of the Great Lakes basin (Figure 11,

Figure 11. Map of ice margin about 17,500 years ago, showing glacial Lake Maumee I (M1).

Table 2). There were three phases of glacial Lake Maumee (I, II, and III) which initially filled the lowland west of Lake Erie and spilled over the drainage divide to the Wabash River near present day Fort Wayne, IN (Calkin and Feenstra 1985). The Maumee I shore (the altitude of the outlet and associated beaches) was at 244 m above sea level and, as the Huron/Erie lobe continued its retreat, extended from Fort Wayne into Michigan through Lenawee County to Macomb County. After the Huron/Erie lobe ice readvanced to the Defiance moraine, glacial Lake Maumee was present only in north-eastern Ohio. Ice retreat again allowed expansion of glacial Lake Maumee into southeastern Michigan. But the retreat also uncovered a lower outlet, lowering Maumee I to the Maumee II altitude of 232 m (Figure 12). Shore features of Maumee II are not well developed in Michigan and are known mostly from Ohio (Calkin and Feenstra 1985). There is no outlet at this altitude across the Defiance moraine, so it is thought that the ice advance ending Maumee II (and building the Flint moraine) buried the outlet channel (Leverett and Taylor 1915). With the lower outlet buried, the previous and higher (244 m) outlet was used again. Ice retreat from the Flint moraine opened up a channel near Imlay City (Lapeer County) allowing Maumee III to stabilize at 238 m and drain eastward into the Saginaw lowlands. This well-developed channel is clearly visible on a topographic map of Lapeer County, extending from the glacial Lake Maumee plain to the Flint River.

Ice retreat of the Saginaw lobe from the Flint moraine also started opening up the Saginaw lowlands in central Michigan, forming Early Lake Saginaw (Figure 12). This lake drained westward via the glacial Grand River, through the

Table 2: Post-glacial lakes and levels from 17,000 to 3,000 years calBP in the Great Lakes basins (elevations in parentheses given in meters ASL).

Cal BP	Superior	Michigan	Huron	Erie
				Maumee I (244)
		Glenwood I (195)		Maumee II (232)
17000			Early Saginaw (225)	Maumee III (237)
			Arkona (216-211)	
		Mackinaw (170?)	post-Arkona low (?)	Ypsilanti (<166)
16000		Glenwood II (195)	Saginaw (218-211)	Whittlesey (225)
15000			Warren I&II-Wayne (210-199)	
			Warren III (206-203)	
			Grassmere-Lundy (195-189)	
		Calumet? (189)	Early Algonquin	
14000				Early Erie (<159)
		Two Creeks (170?)	Kirkfield (173?)	
		Calumet (189)		
		Toleston? (184)		
13000		Main Algonquin (184)		
	Duluth (331)			
	Duluth (331)	Algonquin (184)		
		Algonquin (descending)		
12000				
	Duluth (331) & Minong (220)	Algonquin (descending)		
		Chippewa (75)	Stanley (66)	
11000				
	Minong (220)			
		Mattawa flood I		
10000				
		Mattawa flood II		
			-closed basins?-	
9000		Olson forest bed (153)		
	-Superior basin confluent with Huron basin-			
8000				
			Sanilac forest bed (165)	
7000				
6000	Nipissing (184)			
		-coastal dune building begins-		
5000				
4000				
	Algoma (181)			
3000				

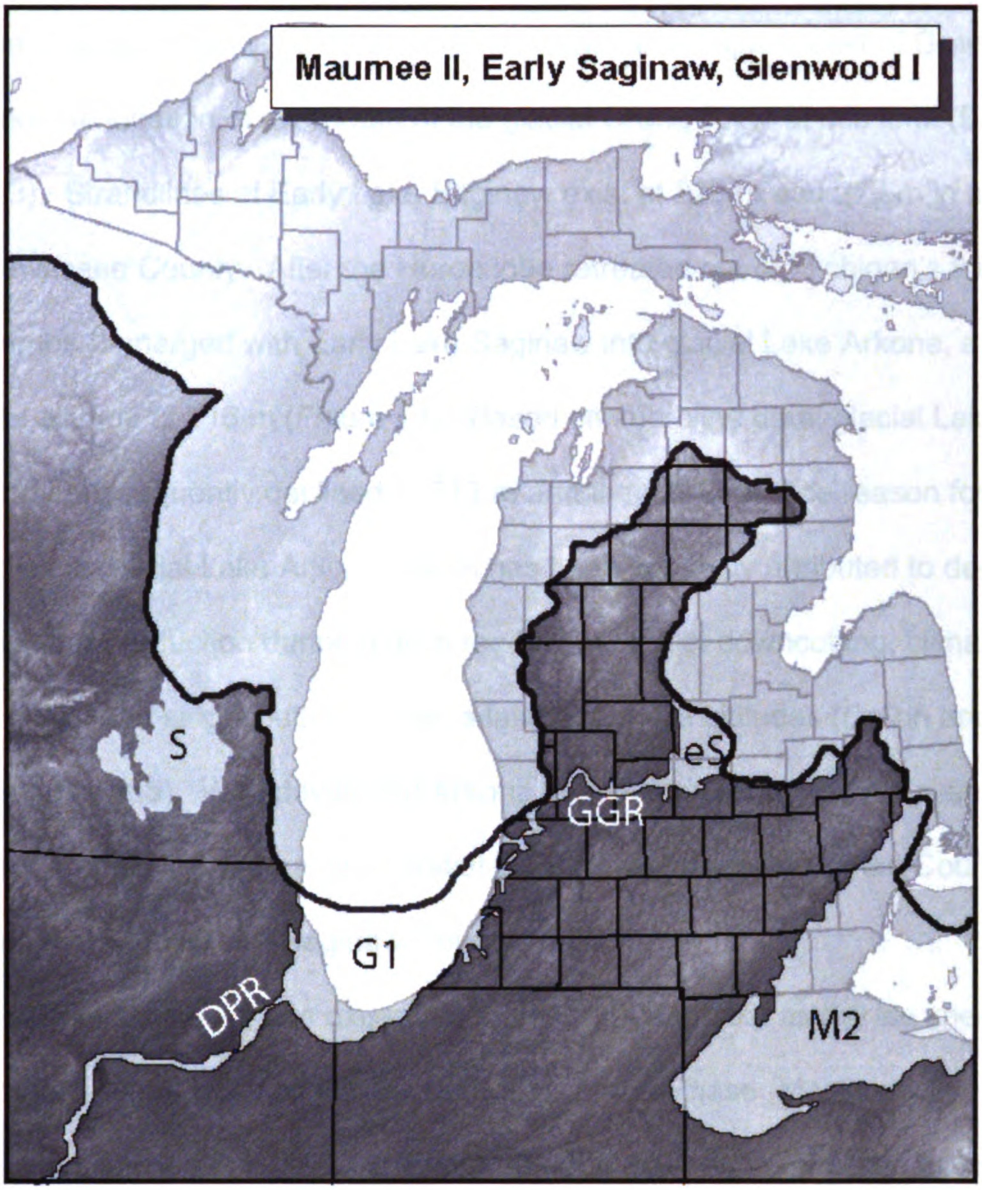

Figure 12. Map of ice margin about 17,100 years ago, showing Lake Maumee II (M2), Early Lake Saginaw (eS), and the Glenwood I phase of glacial Lake Chicago (G1). At this time, the glacial Grand River (GGR) drained Early Lake Saginaw into Lake Chicago and the Des Plaines River (DPR) drained Lake Chicago to the Mississippi River drainage basin. Lake Scuppernong (S) is shown in front of the Green Bay lobe in Wisconsin (Clayton 1997).

Maple River valley at the village of Maple Rapids, to eventually empty into glacial Lake Chicago, Glenwood I phase. The Allendale delta, just east of Grand Rapids marks the location of the mouth of the glacial Grand River at this time (Bretz 1953). Strandlines of Early Lake Saginaw exist at 225 m and 222 m in northern Shiawassee County. After the Huron lobe retreated out of Michigan's thumb, Maumee III merged with Early Lake Saginaw into glacial Lake Arkona, at an initial altitude of 216 m (Figure 13). Based on shoreline data, glacial Lake Arkona subsequently declined to 213 m and then 212 m. The reason for the decline in glacial Lake Arkona's level has been variously attributed to decrease in meltwater production during glacial readvance, outlet downcutting, climate, and the ability of a single outlet to drain a lake at several altitudes (Calkin and Feenstra 1985). Well-developed Arkona beaches exist on the lake plain east of Maple Rapids in southeastern Gratiot County, northeastern Clinton County, and along the Shiawassee-Saginaw County line.

Glacial Lake Arkona expanded to the north and east as the ice sheet retreated during a period known as the Mackinaw Phase (Mackinaw interstade of Eschman and Karrow (1985)). Eventually, the ice margin retreated far enough into southern Ontario that it opened isostatically-depressed outlets in southeastern Georgian Bay, draining the Saginaw lowlands. This event left the Lake Erie basin with no meltwater source (the St. Clair River was left dry at the time) as well as an isostatically depressed outlet in the vicinity of the Niagara River. As a result, a very low level lake named Lake Ypsilanti existed for a short

Figure 13. Map of ice margin about 16,500 years, ago showing Lake Arkona (A) and the Glenwood I phase of glacial Lake Chicago (G1).

time that only had drainage from local rivers (Calkin and Feenstra 1985) and probably occupied about half the area that Lake Erie does today.

The Port Huron glacial advance about 16,000 years ago (Blewett and Winters 1995) ended the Mackinaw Phase and not only closed the outlet in Ontario but also separated the Saginaw lowlands off from southeastern Michigan (Figure 14). The Port Huron moraine cuts off glacial Lake Arkona beaches in southeastern Michigan from glacial Lake Arkona beaches in the Saginaw lowlands, proving that they preceded the Port Huron glacial advance. Thus separated from the Saginaw valley, a large lake with well-developed beach ridges at 225 m called glacial Lake Whittlesey formed in the Lake Erie basin (Leverett and Taylor 1915). Glacial Lake Whittlesey drained across the Thumb through a channel near Ubly, in Huron County. The Ubly channel ends at a delta (now cut by the Cass River 4 km upstream of Caro in Tuscola County) where it emptied into glacial Lake Saginaw at 212 m, which itself overflowed into the glacial Grand River and eventually into the Glenwood II lake phase of glacial Lake Chicago (contributing more sediment to the Allendale delta). Glacial Lake Whittlesey shorelines in Michigan appear along the eastern edge of the Defiance moraine from Lenawee County through to Sanilac County.

Retreat from the Port Huron maximum exposed lower outlets and reconnected the proglacial lakes of southeastern Michigan with the Saginaw lowland, ending glacial Lake Whittlesey and forming glacial Lakes Warren I and II at 210-205 m, respectively. The Warren lakes drained through the glacial Grand River to glacial Lake Chicago (there is debate as to whether Glenwood II or

Figure 14. Map of ice margin about 15,900 years ago, showing Lake Whittlesey (W), Lake Saginaw (S), and the Glenwood II phase of glacial Lake Chicago (G2).

Calumet existed at the time), and were followed by Lake Wayne at 201-199 m, when ice retreat exposed a lower outlet to the east through New York State (Fullerton 1980). Closure of the eastern outlet returned drainage to the glacial Grand River for glacial Lake Warren III at 206-203 m, based on evidence in north-central New York and incision of the Allendale delta (Muller 1977, Fullerton 1980).

The eastern outlet apparently reopened during ice retreat allowing the lower glacial Lake Grassmere (195 m) to follow glacial Lake Warren III (Calkin and Feenstra 1985). Two other drops in lake level stabilized briefly at glacial Lakes Lundy (189 m) and Elkton (187 m). After this event around 14,400 years ago, the Lake Erie basin was separated from direct meltwater input and dropped to the low, Early Lake Erie level. As the outlet at Niagara River rose due to isostatic uplift, the level of Lake Erie slowly rose over the ensuing 14,000 years, to its present level.

The Lake Huron basin lake, still in contact with the retreating Port Huron ice margin, probably dropped to the 184 m level of Early Lake Algonquin, most likely draining south via an outlet at Port Huron (Deane 1950). Direct evidence for this lake has not been found because subsequent lake stages at the same altitude obscured its beaches (Eschman and Karrow 1985). Continued ice retreat past the Straits of Mackinac however, resulted in extension of Early Lake Algonquin into the Lake Michigan basin, eventually leading to the Two Creeks low lake phase (Figure 15) (Hansel et al. 1985). When the retreating ice margin opened a lower outlet near Fenelon Falls southeast of Georgian Bay, Early Lake Algonquin

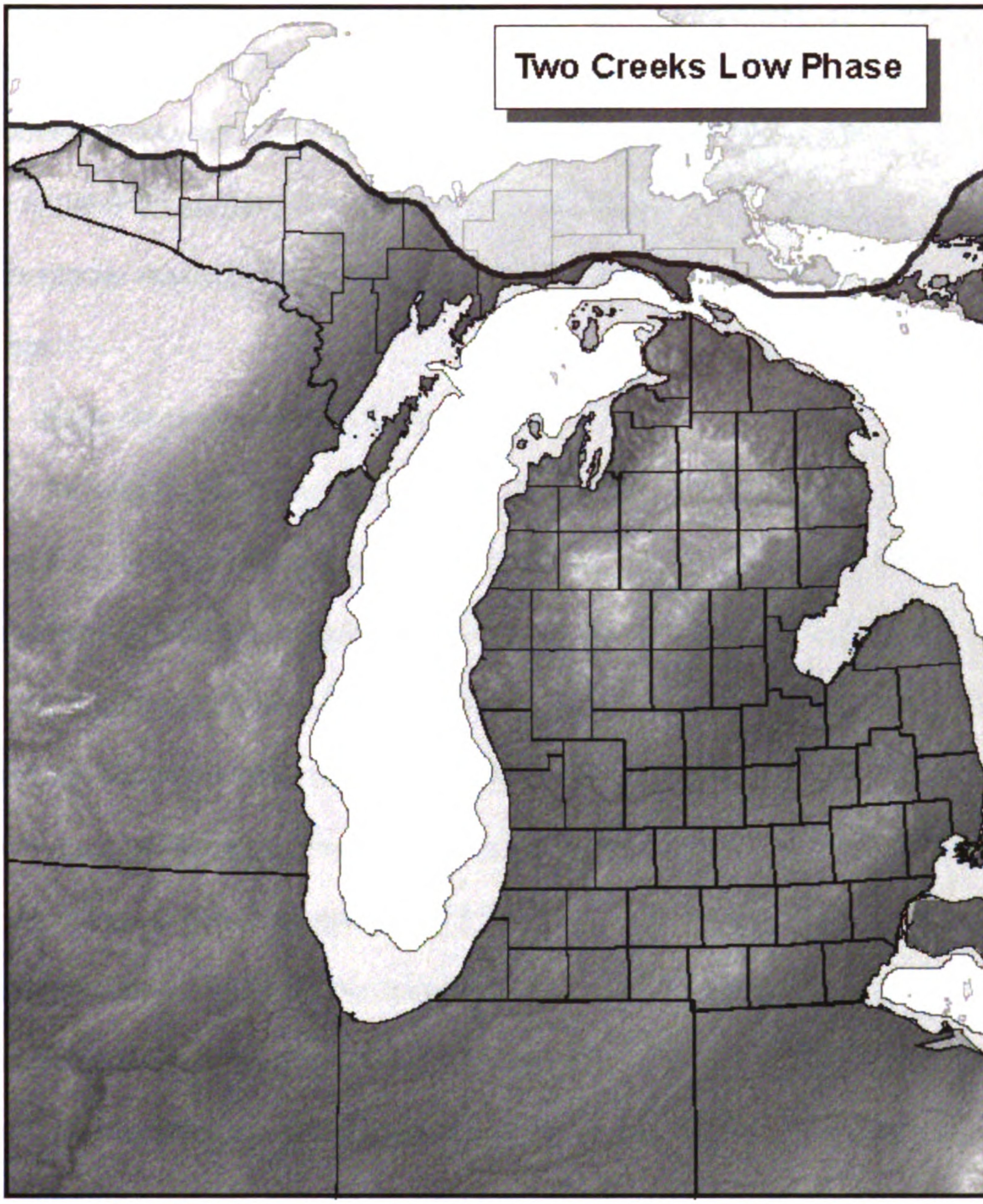

Figure 15. Map of ice margin about 14,000 years ago, showing low lake levels associated with the Twocreekan interstadial (Two Rivers Phase).

dropped an even lower lake phase - the Kirkfield lake phase. The Kirkfield lake phase ended and returned to the Algonquin level when its outlet either rebounded or was covered by a glacial advance. The latter seems more likely since the timing appears to coincide with the Two Rivers Phase glacial advance. This glacial readvance also separated the Lakes Michigan and Huron basins again.

Early lakes of the Lake Michigan basin

As long as the area of the Straits of Mackinac were covered by ice, lake level fluctuations in the Lake Huron basin did not affect events occurring in the Lake Michigan basin, except for some drainage down the Glacial Grand River. Recall that the Saginaw lobe retreated into south-central Michigan long before the Lake Michigan lobe retreated into the Lake Michigan basin. Meltwater from south-central Michigan, therefore, initially overflowed down the Wabash River, along with overflow from glacial Lake Maumee. Ice retreat to the Valparaiso moraine allowed much of the drainage from south-central Michigan to be redirected SW via the Kankakee River, (an Illinois River tributary), and also formed several ice-marginal lakes between the ice margin and the Kalamazoo moraine (Figures 10, 11). Only after ice retreated from the Valparaiso position toward the Lake Border moraine, probably after 17,500 years ago (Farrand and Eschman 1974), did meltwater finally form a lake within the Lake Michigan basin. Known as the Glenwood I phase of glacial Lake Chicago (Figure 13), it stood at 195 m, based on an extensive spit in the Chicago area, or about 18 m above current lake level. Glacial Lake Chicago drained to the south into the Des Plaines and Illinois Rivers (known as the Chicago outlet) to the Mississippi River.

Prolonged ice recession from the Lake Border moraine during the Mackinaw Phase eventually formed a lower level Mackinaw lake phase (also referred to as the intra-Glenwood low lake phase) in the Lake Michigan basin. This lake drained eastward into the Huron basin via an outlet at or near the Straits of Mackinac (Monaghan and Hansel 1990). The altitude of this lake is not known, but Hansel et al. (1985) cited a lack of shore-feature evidence as reason to believe that the Mackinaw lake phase water level was lower than present. Monaghan and Hansel (1990) reported a 16,400 year age from wood at the base of a spit near present lake altitude in Berrien County. The rising lake level forming the spit was caused by closure of the Mackinac Straits by the Port Huron ice advance and marks the end of the Mackinaw Phase. Isolation of the Lake Michigan basin returned drainage to the Chicago outlet, beginning the Glenwood II phase of glacial Lake Chicago at 195 m (Figure 14). The Glenwood II phase of glacial Lake Chicago lasted until at least 15,150 years ago (Karrow et al. 1975). Hough (1963) believed that erosion of the Chicago outlet initiated two later lake phases at lower levels. The first was the Calumet phase at 189.0 m; the second was the Toleston lake phase at 184.5 m. Both of these phases were initially based on beaches, spits and wave-cuts near the original Glenwood lake phase spit. Calumet beaches have been traced into southwestern Michigan. The St. Joseph River also built a large delta in Berrien County during the Calumet lake phase (Kincare 2007). Tracing Toleston lake phase beaches is difficult because two subsequent lake phases were at the same altitude. There exists debate as to when the Calumet phase was initiated, with some researchers arguing it was prior to development of the Two Creeks Forest bed found in eastern Wisconsin that has been dated at 13,760 years ago (Bretz 1951, 1959; Eschman and Farrand 1970, Kaiser 1994). Others have argued it was later (Hough 1958) and possibly related to the amount of discharge entering the lake (Hansel and Mickelson 1988). Some also doubt the existence of a Toleston lake phase entirely (Hansel et al. 1985).

Retreat of the Lake Michigan lobe ice margin from the Port Huron moraine and the opening of the Straits of Mackinac eventually allowed glacial Lake Chicago to drop to the level of Early Lake Algonquin in the Huron basin and then to the level of the Kirkfield phase. During the Kirkfield phase, lake level in the Lake Michigan basin must have been lower than present to allow for the growth of the Two Creeks forest, near present lake level.

Glacial readvance then covered the Straits of Mackinac outlet again during the Two Rivers Phase, burying the Two Creeks forest bed beneath glacial till. The advance again isolated the Lake Michigan basin from the Huron basin, causing lake level in the Lake Michigan basin to rise to 189 m, to the level of the Calumet lake phase of glacial Lake Chicago, and returning drainage to the Chicago outlet (Figure 16). The timing of the ice margin advance during the Two Rivers Phase is defined in Michigan at about 13,700 years ago by radiocarbon dates on the Cheboygan bryophyte bed in Cheboygan County (Larson et al. 1994). This was the last oscillation of ice into the Lower Peninsula of Michigan

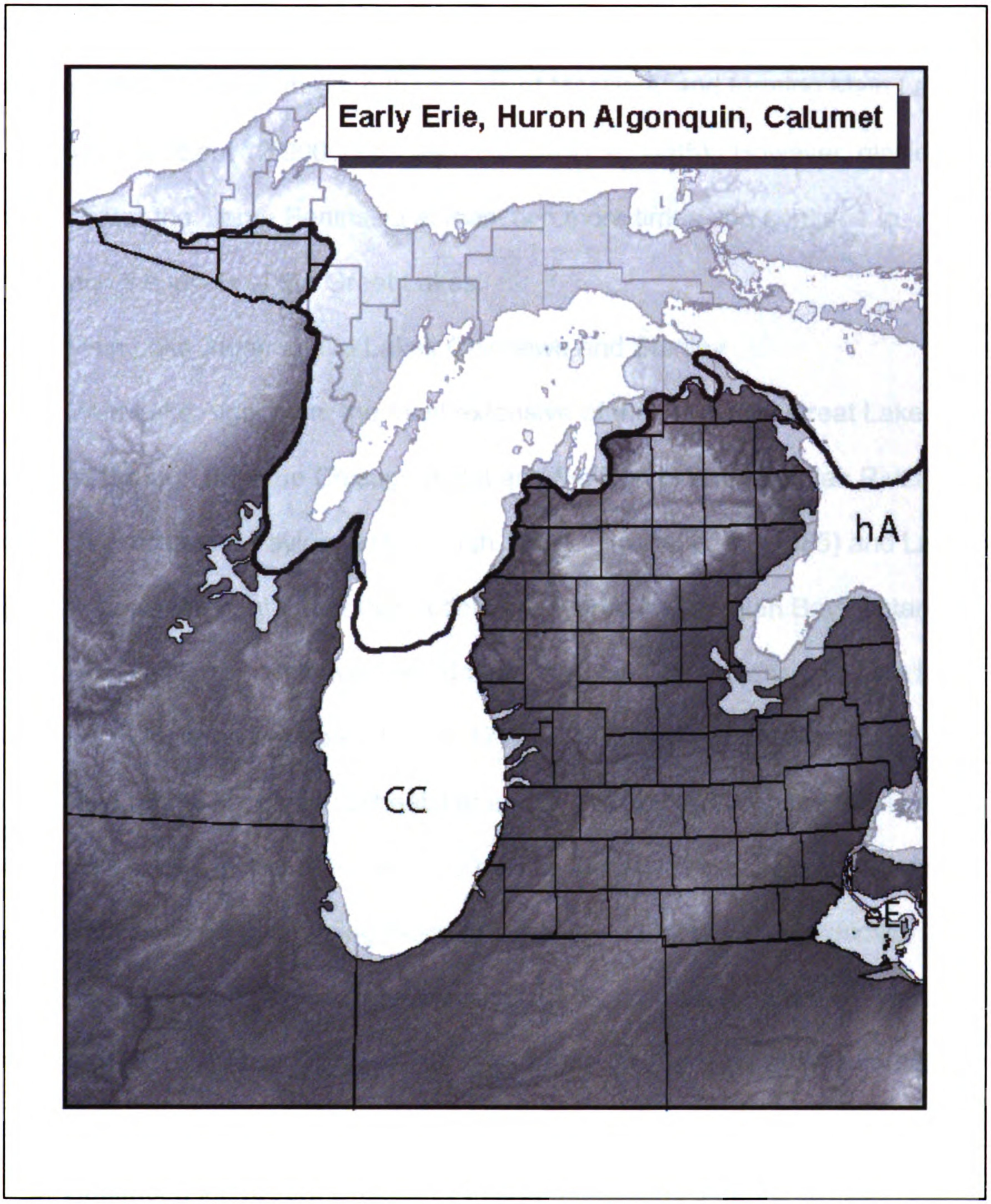

Figure 16. Map of ice margin about 13,700 years ago, showing Early Lake Erie (eE), Huron Lake Algonquin (hA) and the Calumet phase of glacial Lake Chicago (CC).

and its departure across the Straits of Mackinac ended glacial Lake Chicago, allowing free drainage through the Straits of Mackinac and forming Main Lake Algonquin at about 12,900 years ago (Hansel et al. 1985). However, glaciers would enter the Upper Peninsula at least two more times and continue to influence the levels of the Great Lakes.

Main Lake Algonquin to Lakes Chippewa and Stanley

Main Lake Algonquin, the most extensive of the proglacial Great Lakes, drained through both the Chicago outlet and the Port Huron (St. Clair River) outlet (Leverett and Taylor 1915, Hough 1958). Hansel et al. (1985) and Larsen (1987), however, contended that outlets uncovered in Georgian Bay, Ontario by glacial retreat, also may have been drainage paths. Regardless, this lake left an indelible mark upon Michigan (Figure 17) and its shorelines are marked by well-developed wave-cut bluffs (Schaetzl et al. 2002), beaches (Futyma 1981), and extensive spits (Krist and Schaetzl 2001). As the ice retreated, Lake Algonquin expanded northward, covering much of the eastern half of the isostatically depressed northern peninsula, with many islands protruding above the water surface (Schaetzl et al. 2002). The Lake Algonquin shoreline eventually reached at least as far as 65 km north of Sault Ste. Marie (Farrand and Drexler 1985).

Continued ice retreat eventually began uncovering a series of lower outlets east of Georgian Bay - similar to outlets used during the earlier Kirkfield Phase. These outlets were isostatically depressed by the weight of the ice, allowing for drainage across the divide through Ontario. Thus, several post-Main Lake

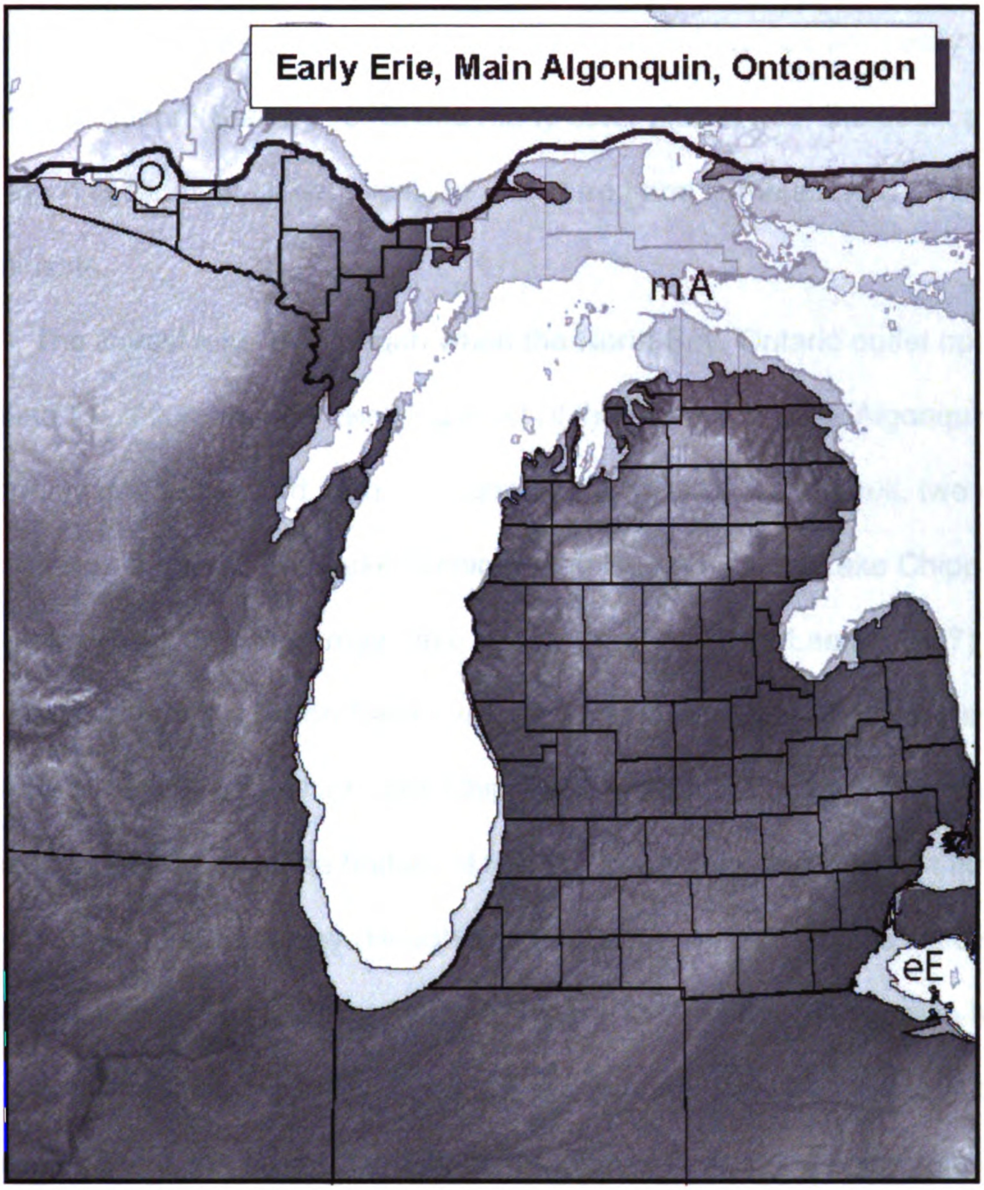

Figure 17 Map of ice margin about 13,200 years ago, showing Main Lake Algonquin (mA), Early Lake Erie (eE), and Lake Ontonagon (O).

Algonquin phases temporarily stabilized at successively lower altitudes as each new outlet was opened. Some of these lower beaches are easily visible on Mackinac Island (Stanley 1945) and many other places near the coast along the northern sections of Lakes Michigan and Huron and in the eastern Upper Peninsula.

The lowest lake level began when the North Bay, Ontario outlet opened around 11,200 years ago, allowing most of the volume of Lake Algonquin to suddenly drain eastward, through Canada (Figure 18). As a result, two very low, small lakes formed in the Lake Michigan and Huron basins. Lake Chippewa (in the Lake Michigan basin) may have been as low as 70 m (Larsen 1987) while Lake Stanley (in the Huron Basin) was perhaps about 45 m (Eschman and Karrow 1985). The level for Lake Chippewa was controlled by a river channel eroded into bedrock at the bottom of the Straits of Mackinac, where it flowed eastward into Lake Stanley (Hough 1958). Today, we refer to this valley as the Mackinac Gorge. Only about 800 years separated the Main Algonquin high level from the Chippewa low level. Recently, Lewis et al. (2007) suggested Lake Stanley may have only dropped to 127 m, basing their argument on the altitude seismic reflections and an erosion surface seen in off-shore cores. They also have suggested that by 9000 years ago Lakes Chippewa and Stanley were temporarily closed basins, i.e. they had no outward drainage, and that an early Holocene dry climate played a part in the inability of lake levels to keep pace with isostatically rising outlets.

Figure 18. Map of ice margin about 11,600 years ago, during the Marquette glacial readvance (Gribben Phase), showing Early Lake Erie (eE), Lake Stanley (S), Lake Chippewa (C), and Lake Minong (M).

Opening the Lake Superior basin

In some ways the development of Lake Superior is simpler than the other Great Lakes, because the Superior basin was the last to be deglaciated. However, complicating its development are the facts that the basin twice had separate eastern and western lakes and was subjected to inflows from glacial Lake Agassiz - the largest North American glacial lake, covering a large area in Manitoba and western Ontario, Canada at the time of the inflows. The Lake Superior basin also experienced the largest amount of isostatic depression (and hence, rebound) of the Great Lake basins, and its outlet at Sault Ste. Marie, being on the eastern edge of the basin, is rebounding more than its western margins.

The oldest existing shorelines in the Superior basin were formed after the Two Rivers Phase ice advance, roughly 12,900 years ago. At this time, glacial Lake Duluth formed at the western end of the basin as well as the much smaller glacial Lake Ontonagon in Gogebic, Ontonagon, and Houghton Counties (Figure 17). Glacial Lake Ontonagon's outlet was at 403 m, draining to the SW through Wisconsin (Leverett 1929). Glacial Lake Duluth overflowed south through outlets in Minnesota and Wisconsin at an altitude of around 331 m. As glacial Lake Duluth was expanding northward along the western shore (and extinguishing glacial Lake Ontonagon), glacial Lake Algonquin was also expanding northward across the eastern Upper Peninsula. Once the retreating ice margin cleared the Keweenaw Peninsula and the Huron Mountains, glacial Lake Duluth merged with glacial Lake Algonquin -which was at a lower level. It is not known how far north glacial Lake Algonquin eventually extended, because the subsequent Marquette Phase ice readvance wiped away its northern shorelines as far south as Alona Bay, Ontario (Farrand and Drexler 1985). Following the merger of the two lakes, the level of glacial Lake Algonquin did not remain stable but slowly fell due to the progressive opening of lower outlets to the east. A drainage divide at Sault St. Marie, however, prevented the water level in the Superior basin from falling to the extremely low level of Lake Stanley in the Huron basin, establishing a new lake in the Superior basin named glacial Lake Minong. It was around this time that a pathway also may have opened for water from glacial Lake Agassiz (Moorhead phase) to escape east into the Superior basin and drain into the North Atlantic Ocean (Fisher 2003). Farrand and Drexler (1985) pointed out that Minong shorelines are 40 m above the present outlet at Sault Ste. Marie. A barrier to drainage, perhaps a moraine across Whitefish Bay (Saarnisto 1974), must therefore have existed during glacial Lake Minong, to hold this water up. During the peak of the Marquette Phase ice readvance, glacial Lake Minong was quite small, pinned into the SE corner of the Superior basin. But as the ice retreated, glacial Lake Minong expanded to the north and west. When the ice cleared the Keweenaw Peninsula, glacial Lake Duluth in the west merged with glacial Lake Minong in the east. Shorelines from glacial Lake Minong are found on Isle Royale, and are the highest shorelines along the north shore of Lake Superior.

During the Marquette Phase ice readvance around 11,580 years ago (Figure 18), the advancing ice margin separated the eastern and western portions of the Lake Superior basin, leaving a much diminished glacial Lake

Minong in the east and reestablishing glacial Lake Duluth in the west. The advance even briefly squeezed glacial Lake Duluth out of Michigan, leaving a renewed glacial Lake Ontonagon in the western corner of the Upper Peninsula. As the ice margin subsequently retreated, glacial Lake Duluth began expanding northward again, subsuming glacial Lake Ontonagon (Figure 19). When the ice margin pulled back from the Huron Mountains in Marquette County, eastern outlets opened up and again allowed the level of glacial Lake Duluth to fall (Farrand and Drexler 1985). Discharge from glacial Lake Duluth initially went south from Munising via the AuTrain-Whitefish channel and then found a lower outlet on the north side of the Marquette Phase moraines in Marquette and Alger Counties toward glacial Lake Minong (Leverett 1929, Blewett 1994).

The Nipissing transgression

As soon as the North Bay, Ontario outlet was uncovered by glacial retreat about 11,200 years ago, leading to the rapid draining of glacial Lake Algonquin and forming the low-level Lakes Chippewa and Stanley, the outlet started to rebound. The slow but steady rise of the outlet caused the level of these low lakes to rise as well. Without any other influences, the rising lake level should have been a long, uninterrupted asymptotic curve; fast at first and slowing down over time. But because the Marquette Phase ice retreat had opened outlets from glacial Lake Agassiz into the Lake Superior basin, glacial Lake Agassiz floodwaters again discharged into the Superior basin, this time quickly overwhelming glacial Lake Minong which was at the time overflowing into the

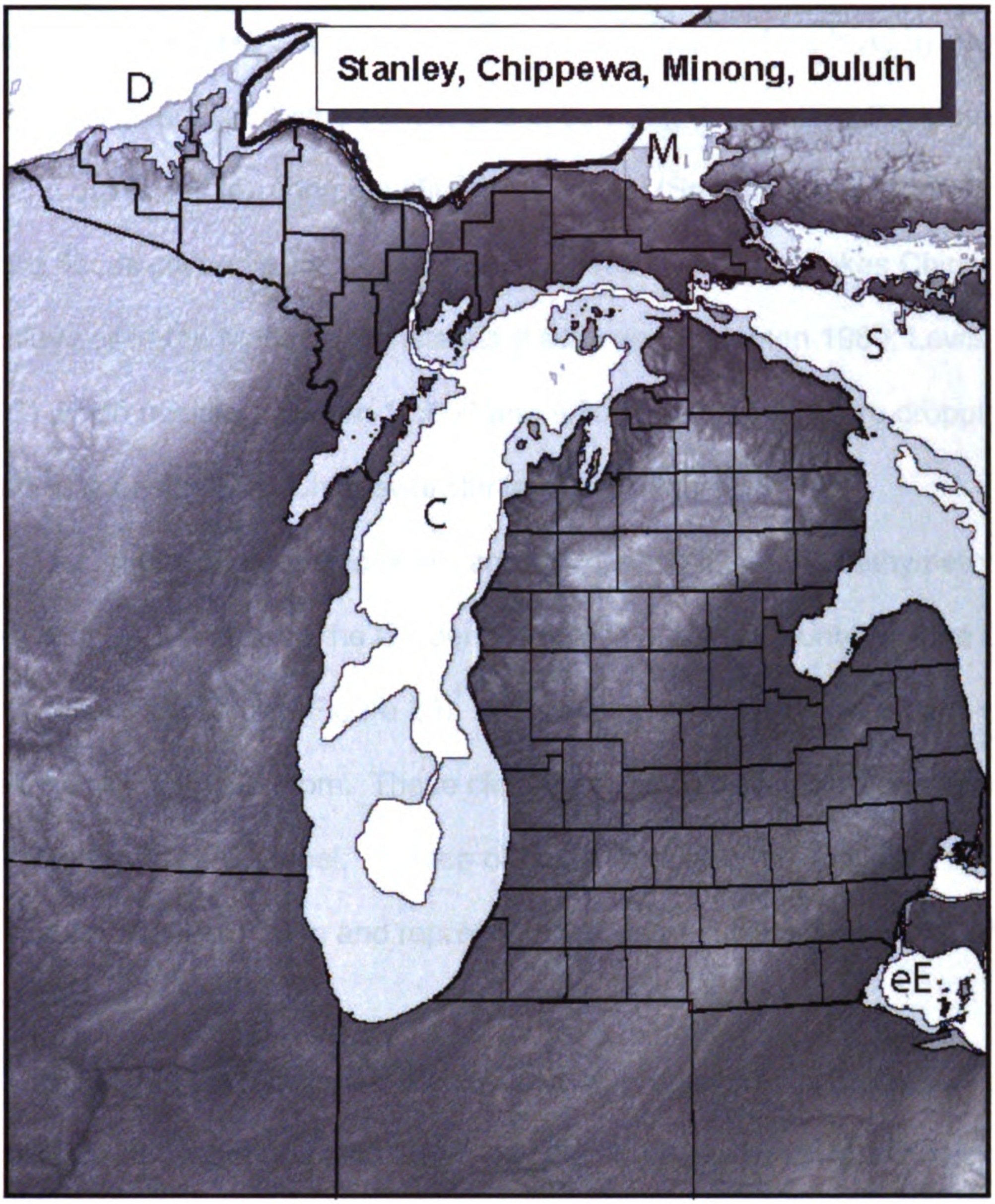

Figure 19. Map of ice margin about 11,200 years ago, during retreat of the Marquette ice, showing the rising Lakes Stanley (S) and Chippewa (C) and Lakes Duluth (D) and Minong (M).

Lake Michigan and Huron basins. At least two floods from glacial Lake Agassiz tore through the Lake Superior basin, the first raising the lake level by as much as 35 m in a very short time period, and in so doing downcutting the glacial Lake Minong outlet and lowering the Minong lake level (Safarudin and Moore 1999). These floods caused a series of temporary rises in levels of Lakes Chippewa and Stanley called the Mattawa highstands (Lewis and Anderson 1989, Lewis et al. 2007) which peaked between 10,600 and 9,300 years ago, before dropping back down to the rising post-Chippewa/Stanley level (Figure 19).

The results of these floods are actually observable in the bathymetry of Lake Michigan - between the Garden Peninsula of Delta County and the Door Peninsula of Wisconsin (Figure 20). A distinct channel ending in a delta can be seen along the lake-bottom. These clearly extend directly from the onshore AuTrain-Whitefish Channel. The top of the delta (called the Whitefish fan) is at an altitude of about 126 m and represents the water surface at the time of deposition.

Recall that the Chippewa-Stanley lake level was already gradually rising due to isostatic rebound of the North Bay outlet, before flood inflows from the Mattawa highstands occurred. It is likely that the Agassiz floods also eroded the St. Marys River down to bedrock, allowing the rising Chippewa-Stanley Lakes to merge with a post-Minong lake in the Superior basin.

Several drowned forests have been found under the present day upper Great Lakes, attesting to the once low Chippewa-Stanley levels in the Michigan and Huron basins. Wood samples from the Olson drowned forest site (Figure

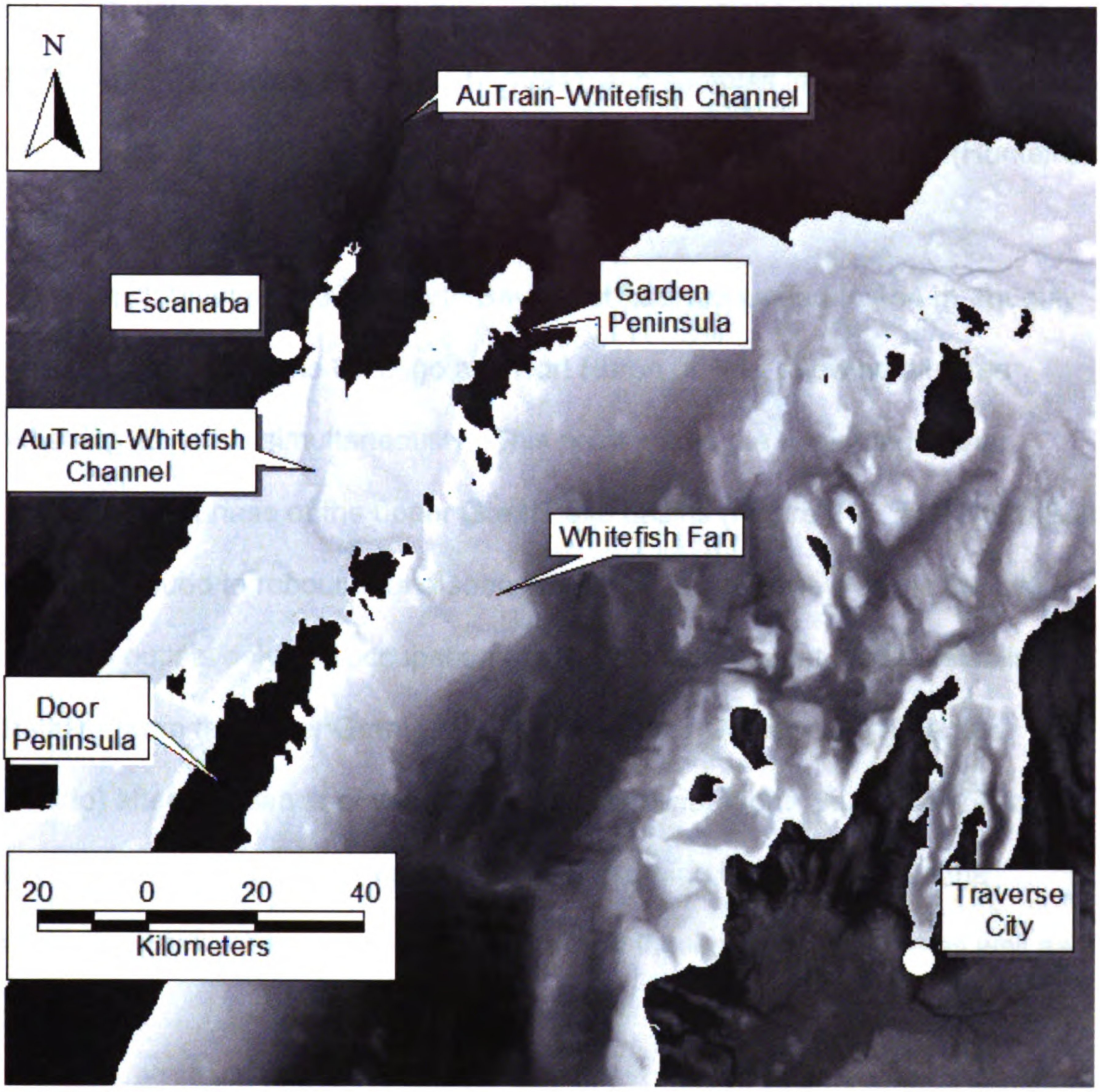

Figure 20. Digital elevation model (DEM) of the northwest edge of Lake Michigan, showing the Au Train-Whitefish channel descending from onshore to offshore and connecting with the submerged Whitefish fan. DEM data courtesy of the Michigan Center for Geographic Information (www.michigan.gov/cgi).

21), 25 m under Lake Michigan near Chicago, has yielded an average age of 9,155 years (Schneider and Popadic 1994), and samples from the Sanilac drowned forest dated to about 7,350 years ago in 12.5 m water depth (Hunter et al. 2006).

By 6,300 years ago the North Bay outlet had rebounded to 184 m, roughly the same altitude as the Chicago and Port Huron outlets, allowing all three outlets to be active simultaneously. This point marks the beginning of the Nipissing lake phase of the upper Great Lake basins (Figure 21). The North Bay outlet continued to rebound, and soon thereafter all drainage was via the two southern outlets only. Reoccupation of the Port Huron outlet also reestablished drainage from the upper Great Lakes into the lower Great Lakes (Erie and Ontario) after thousands of years of hydrologic separation. Studies by Fraser et al. (1990) indicate that the Nipissing lake phase high level (as well as the subsequent fall in lake level) was also strongly influenced by climate as well as by rebound.

Nipissing lake phase shorelines appear in many places all around coasts of Lakes Superior, Michigan, and Huron and are often the first prominent shoreline above present lake level (Hough 1958). They are second only to Algonquin shorelines in strength of development and have a more widespread geographic occurrence. Nipissing shorelines appear within a few meters of 184 m as far north as Traverse City and then gradually increase to 197 m at Sault Ste. Marie.

Figure 21. Map of Nipissing phase (N) highstand about 6,000 years ago. The locations of the Olson (o) and Sanilac (s) drowned forest beds and early Lake Erie (eE) are shown.

The modern Great Lakes

The glacial sediments of the Port Huron outlet were more susceptible to erosion than the bedrock at the Chicago outlet. Thus, the Port Huron outlet was gradually lowered by discharge from the Huron basin, allowing it to eventually take all the waters from the upper Great Lakes. The lake level stabilized long enough for a poorly-developed beach system to form around 3,400 years ago. Known as Lake Algoma, its altitude was 181.4 m. This event probably represents a temporary shift to a cooler and/or wetter paleoclimate and a subsequent rise in lake level (Fraser et al. 1990). Lake Algoma ended around 2,300 years ago (Baedke and Thompson 2000).

Rebound of the St. Marys River caused the Lake Superior basin to separate from Lake Huron around 2,280 years ago (Farrand and Drexler 1985). At this time the modern Great Lakes achieved their present configuration. Studies of beach ridges in embayments around Michigan and Indiana suggest that lake levels have fluctuated on both a 160 year and a 30 year cycle since the end of Lake Algoma (Baedke and Thompson 2000). Historical records show lake level has varied by about 1.2 m since 1880 A.D. (Fraser et al. 1990).

Chapter 5

BASE LEVEL AND THE GRADED STREAM

The concept of base level

The concept of base level in regard to valley development was proposed by Powell (1875). Base level is the limit of land reduction a stream is capable of producing. Powell (Ibid. p. 24) described sea level as the grand base level, "...below which the dry lands cannot be eroded." Malott (1928 p. 83) said that "our judgment of the base level of any locality is our knowledge of the stream and valley gradients from the locality in question thru [sic] to the sea." Clearly then, an examination of the effects of base-level changes must include characterization of the slope (gradient) of the stream reach(es) in question. With the availability of high quality 7.5 minute quadrangles, this type of examination is certainly easier than it was in Powell's day.

Malott (Ibid.) further defined the three types of base level he claimed that Powell had introduced: ultimate base level, local base level, and temporary base level. Ultimate base level is Powell's grand base level. Local base level is the level to which an interior drainage is graded (see discussion below). Temporary base level is an inland barrier to erosion. This can be in the form of a lake, resistant bedrock (which may be manifested as a waterfall), or an artificial barrier such as a dam. Malott (Ibid.) used Niagara Falls as an example to illustrate that Lake Erie is clearly not graded to the sea. An inclined plane representing the gradient of the St. Lawrence River up to Lake Erie would lie well below the lake. The back-cutting of Niagara Falls will eventually drain Lake Erie and the streams in its drainage basin will then have to adjust to a lower base level. This new level may be Lake Ontario or a greatly reduced lake in the center of the Lake Erie basin. Thornbury (1969) argued that the concept of sea level as the ultimate base level controlling erosion for a continent is not very useful when studying areas distant from the sea. He defined local base level for a valley as "...the present level of the valley to which it is tributary" (Ibid. p. 106). This seems to the writer a more practical approach to the study of a stream valley.

Blum and Tornqvist (2000) point out that there is recognition of two types of base level; stratigraphic and geomorphic. The above discussion was based on arguments of geomorphic base level, that is, the real surface of the limit of erosion. Stratigraphic base level is the upper limit of deposition, which is partially controlled by changes in geomorphic base level. It is also controlled by commonly described fluvial variables such as sediment supply, stream power and channel morphology.

The concept of the graded stream

The concept of a graded stream has been attributed to the work of Gilbert (Rubey 1952, Thornbury 1969, Mackin 1948), though related ideas concerning the stability of natural rivers were introduced a century earlier (Chorley 2000). From Gilbert's (1877) observations of stream patterns during his surveys in the Western Territories, he reasoned that the energy of a stream is used in the transportation of debris. When the amount of debris is less than the energy expended in its transport, erosion (corrasion) will occur. If the load exceeds the energy of the stream, deposition of load will occur until energy and load are balanced. Gilbert (Ibid. p. 112) said that this process "...tends to establish a single, uniform grade." But he also recognized that, through the course of a stream, local conditions may introduce change which would "... result in inequalities of grade ... proportioned to the [bed] resistance" and the velocity of joined tributaries (Ibid. p. 113). Gilbert (Ibid. p. 113) also believed that the flood stage of a stream was responsible for determining the grade because it "...overpower[s] any influence ... exerted at a low stage."

Davis (1902) incorporated the graded stream concept in his theory of the cyclical development of stream valleys. His premise was that stream valleys undergo a progression of stages: youth, maturity and old age. Each of these stages was represented by certain characteristics of stream velocity, channel shape and sinuosity, debris load, and valley wall slope. Youthful valley slopes were steep and the streams were straight with coarse debris. Old valleys were said to have very low slopes whose streams meandered widely with a fine debris load. This ultimately becomes Davis's peneplain, where erosion reduces the land to low relief with the occasional relict knob of resistant bedrock (Thornbury 1969). Davis's geomorphic cycle is not held in favor today (Dury 1966) for reasons beyond the late nineteenth century aversion to convulsive events. While the characteristics he cited are important in the development of stream morphology, it appears from later work (e.g., Leopold and Bull 1979, Rubey 1952, Thornbury 1969, Mackin 1948) that features which should be representative of certain "Davisian" ages can appear under multiple origins and morphologies. It is unfortunate that the assignment of certain morphologic features to a place in a cycle is so strongly attached to a system that, independent of causation, would otherwise afford good descriptive power.

Gilbert (1877) introduced two descriptive terms that reappear throughout the literature: competence and capacity. The former denoted the largest grain size that a stream can carry. The stream was then said to be "... competent to such debris" (Ibid. p. 110). It should be noted here that, regardless of available energy, a stream may carry only those grain sizes that are in supply. He regarded capacity as the volume of load per volume of water. As the volume of water increases, either the capacity of the stream for a given grain size increases, or the "... competence increases, and larger [grain sizes] are lifted" (Ibid. pp. 110-111).

Many artificial flume experiments (e.g., Gilbert 1914, Leopold and Wolman 1957, Wolman and Brush 1961) and field studies (e.g., Rubey 1952, Mackin 1948, Leopold and Bull 1979) have been performed in efforts to quantify the factors that control stream morphology. Leopold and Wolman (1957) summarized these to include: discharge, sediment volume, grain-size distribution, velocity, slope, channel width, depth, and roughness. Said variables are either dependent or independent of stream action. For example, a stream cannot control the climate (and therefore its discharge) nor can it control the amount of sediment supplied by the drainage basin. But it can adjust features like slope and channel form to accommodate the discharge and load. Leopold et al. (1964) refined these factors to include those that were semi-dependent. Flow resistance is partly a function of debris size and the kinds of sedimentary structures the stream builds within the channel. Similarly, debris size, while being limited by supply is also subject to reduction by abrasion within the channel. Interestingly, Leopold et al. (1964) also considered river gradient to be semi-dependent, given a stream's propensity to erode or aggrade where necessary to achieve an efficient gradient. It may be questionable as to how far to draw the concept of semi-dependency. A stream may produce sedimentary structures or alter its depth/width ratio to influence roughness, but most of the factors involved in gradient are set in motion by tectonics, climate, and base level; that is, outside the realm of the stream's control.

Gilbert's (1914) work showed that the largest grain size put in motion by a stream (competence) varies with the sixth power of the water velocity. He also concluded that the channel shape (depth/width ratio) was an important factor in transport of load. Rubey (1938) refined Gilbert's sixth power rule to apply strictly to bed velocity, which varies throughout the stream due to factors such as friction loss along the channel perimeter. In that regard, Leopold and Wolman's (1957 p. 71) flume studies showed that "... the width of the band of moving sediment was often less than the width of the channel." Given this, it is clear that multiple conditions can exist in a stream channel simultaneously. This implies that while a graded stream may be an equilibrium condition, it is certainly not a homogenous condition.

The compilation of experimental and empirical data allowed workers like Mackin (1948) and Rubey (1952) to constructively refine the concept of the graded stream. Mackin (1948) said that a graded condition is an equilibrium reached when the slope at all locations along a stream is just enough to carry the load being supplied, given the existing circumstances. A graded stream's "... defining characteristic is that any change in any of the controlling factors will cause a displacement of the equilibrium in a direction that will tend to absorb the effects of the change" (Ibid. p. 471). Mackin used the valley of the Shoshone River, Wyoming, as an example of a stream that shows evidence of having alternately been at grade and downcut (degraded) during its development. The Shoshone has a slope of more than 5.7 m/km (30 ft/mile) for both terraces and present channel that does not vary when crossing boundaries between soft and hard bedrock. Yet Mackin (Ibid.) observed that even with its steep slope, the Shoshone was at grade during terrace construction and not eroding its channel. Otherwise, its slope would have varied between bedrock units with differing resistance to corrasion. The terraces were formed by lateral corrasion of the confining walls of the valley. Terraces formed in this manner are known as strath terraces (Bull 1991). The thin alluvium covering the bedrock floor was formed by lateral shifting of meanders (see discussion below).

Rubey (1952) stated that, just like the slope, the cross-section of a stream can be graded. Rubey (1952, p. 129) considered the concept of adjusted cross sections "... no less vital to a proper understanding of the laws of river work." After studying the lower Illinois River, Rubey (Ibid. p. 129) noted the "remarkable stability" of its features. He concluded that its present characteristics represented a graded condition. Yet, unlike the Shoshone, it has a gradient of less than 0.024 m/km (1.5 inches/mile). Base level for the Illinois River is its mouth at the Mississippi River. The Illinois River has undergone several episodes of raised base level at its mouth due to aggradation of the Mississippi River. The aggradation was a response to glacial debris and meltwater inputs during Pleistocene deglacial events (Schumm and Brackenridge 1987). The graded condition of the Illinois River was maintained by shifting to a greater proportionate depth/width ratio (form ratio) than seen in other streams in the region. Channel adjustment allowed the Illinois River to maintain "... essential equilibrium ... despite the extreme flatness" (Rubey 1952, p. 136).

Dury (1966) and Kesseli (1941) objected to the use of the term graded stream saying that it offered little due to conflicting definitions. Dury (1966) noted that, by virtue of carrying debris, a graded stream must be party to erosion and a player in landscape reduction. It cannot at the same time and by definition, neither erode nor aggrade. Kesseli (1941) felt that the constantly changing nature of discharge and load through the temporal and spatial variations of climate and landscape rendered invalid any suggestion of a stream in equilibrium. It seems inescapable that, regardless of differences in semantics, streams in varied circumstances will attain the form that allows the most efficient means of doing the work that nature asks of them.

The Pleistocene glaciated landscape of the Great Lakes is between 18,000 and 9,000 years old. This is a very short time over which for streams to equilibrate. In fact, these landscapes are so far from equilibrium that the term coined for poorly drained glaciated terrains is "deranged," (Thornbury 1969). On top of this young landscape are the base-level changes discussed previously.

The record of these changes is contained within the channel and valley forms and the sediments of the streams in the Great Lakes watershed.

Adjustments to Grade by Shifts in Base Level and Sediment Supply

One of the important tools for interpreting changes in rivers is the longitudinal profile. Hack (1957 p. 45) stated that "longitudinal profiles...are intimately related to [the basin's] topography" and therefore "a geomorphological analysis of a region based on a comparison of long profiles is of value." Much has been written about this aspect of rivers, though credit is often given to Mackin (1948) for establishing principles and Hack (1957) and Leopold and Maddock (1953) for quantification. Some workers though, consider the use of longitudinal profiles as “some kind of voodoo” (Grahame Larson, personal communication). Mackin (1948, p. 485) considered the longitudinal profile to be an interacting series of segments that, in a graded stream, were each adjusted to the "controlling conditions" within each segment. Taken together, the segments reflect the factors that control the slope of the profile. Mackin (1948, p. 491) perhaps erred in his consideration of some of the factors as non-systematic "geographic circumstance," such as the general concave upward aspect of the profile. In humid regions, such as the Great Lakes, there are few factors more systematic than the downstream increase in discharge due to both the contribution of tributary flow and the increasing interception of ground-water discharge of largely unconfined aquifers. As a consequence of the general increase in discharge downstream, the depth/width ratio decreases, thus decreasing the slope (Hack 1957). This is in fact the natural force behind the well-known characteristic of the concave upwards longitudinal profile (Leopold et al. 1964). While Hack's studies were performed on bedrock streams in Virginia and Maryland, his map investigations of other non-arid watersheds in North America bore out his results (though not the constants in his empirical equations).

Hack (1957, p. 74) went on to state that "one of the most important [factors] causing characteristic differences" in profiles between drainage basins is grain size. Inasmuch as Michigan's watersheds, particularly along the Lake Michigan coast, contain a wide-variety of grain sizes supplied by the ubiquitous glacial deposits, their general morphologies should be intrinsically comparable.

Rubey (1952, p. 134) stated that the slope at various points along the profile reflected the work that upstream factors imposed while the profile's altitude was controlled by "base level downstream." A recent review by Blum and Tornqvist (2000) implies that Rubey was on the right track, though perhaps only partly correct. Factors other than base level can influence the profile's shape. For instance, with a drop or rise in sea level, the shape taken by the profile's lower segment will depend partly on the pre-existing shape of the shelf exposed or covered, respectively (Posamentier 2001). Also, a high sedimentation rate is capable of causing progradation even during a fall in base level (Leeder and Stewart 1996). Posamentier (2001) stated that observed rates of nickpoint migration indicate that long distances can equilibrate to a new base level within a very short time period. Rates of 40 meters/yr were observed in the Sunda Straits area, considered by Posamentier (2001) to be instantaneous in the geologic record. Posamentier (Ibid.) went on to say that shelf and coastal plains, when fully exposed, can be incised by base-level declines of short duration. He described late Pleistocene incision extending more that 200 km from the shelf edge, though non-incised systems were more common than incised, the latter requiring lowstands of greater amplitude. The relation of Posamentier's (Ibid.) study to Great Lakes basins is of interest. The amplitude of the base-level changes is up to 30% of the entire relief of the drainage-basin and the time scale is an order of magnitude more rapid and yet, as we shall see, some sections of the river achieved equilibration.

While it has long been known that sediment production is tied to climate, the response of alluvial architecture to different inputs can be varied (Schumm 1991). Bull (1991) noted that most fluvial landscapes are polygenetic, that is, they contain the signatures of multiple causal factors. The study of fluvial geomorphology in the Great Lakes is fortunate then to have the dominant factor as the late Wisconsin glaciation imprinted with a mere 10-15 k years of fluvial activity. That is perhaps another reason to study these rivers, to look at the development of a drainage basin with a limited set of variables active over a short time period.

A problem clouding our ability to sort out the relative roles of upstream and downstream factors is the lag time for effects that start in one part of a drainage basin to equilibrate across the entire basin. "In most fluvial systems, upstream and downstream reaches are out-of-phase with respect to aggradation and degradation…" (Ethridge et al. 1998, p. 18). Shanley and McCabe (1994) stated that climate, tectonism, and in areas close to the sea, eustasy, have major influence over sediment supply. However, these controls are somewhat interdependent and their individual influences are difficult to separate. Ethridge et al. (1998) also called sediment supply a limiting factor in alluvial architecture, but did not regard it as an allocyclic control because it was a controlled by climate, tectonism, and base-level fluctuations.

Climate, as regards the development of soils, may be one of the more reliable factors available for alluvial chronostratigraphy. Soil development is well-studied and in-situ, i.e. one does not have to wait 10-100 ka for a soil to form as one does for a base-level change to migrate upstream or a tectonic event to migrate down to the sea. A soil horizon may become more difficult to distinguish in the ancient record however. "The challenge of sequence stratigraphic studies of continental strata…is to discriminate depositional patterns that reflect autocyclic changes from … allocyclic…" (Shanley and McCabe 1994, p.545).

Climate plays an important role upstream as the arbiter of sediment supply as it is "one of the dominant controls on changes in sediment delivery to depositional basins through time" (Blum and Tornqvist 2000, p. 39). Blum and Tornqvist (Ibid., p. 38) stated that further work must develop an "understanding of the relative importance of upstream vs. downstream controls at both the stratigraphic and alluvial architecture scales."

As the effective base level for a river is changed, its drainage and flow characteristics may also change. A lowering of base level can cause rejuvenation of the river by creating an increase in its slope, thereby increasing available energy. Raising the base level can lower the energy in the system by decreasing its slope. These effects will be felt upstream until equilibrium has been restored and a new grade is established (Thornbury 1969, Leopold and Bull 1979).

A stream which has been subjected to a lowered base level may have part of its lower reaches adjusted to the new base level. At the same time the upper course will, for a time, remain adjusted (graded) to the old base level. That portion of the river between these reaches will be in the process of adjusting to the new base level. This type of abrupt stream gradient change is known as an "interrupted profile" (Thornbury 1969, p. 109). The headward or upstream limit of regrading to the new base level is the "nickpoint" (Ibid., p. 110). Base-level changes are not the only cause of nickpoints and interrupted profiles. A structural feature, such as a resistant bedrock layer, may cause a gradient change. Dams constructed along the course of a stream alter its natural gradient by creating artificial nickpoints.

Changes that an alteration in energy level can bring about are numerous. An increase in energy can cause the onset of erosion along the affected stream channel. The process of downcutting can lead to the terrace formation, entrenchment of the stream in an incised channel, and loss of contact with other stream valley features (e.g., point bars, flood plains, etc.) of the previous flow regime (Ibid.). Increasing erosion and stream competence can increase its sediment load and may lead to deposition on those point bars that are still in contact with the stream or new point bars may be established. The depositional sequence may show an increase in grain size before returning to the typical fining upward sequence characteristic of point bars (Thornbury 1969, Reineck and Singh 1980).

An increase in base level has the obvious effect of the water level rising and drowning the lower portion of a stream. Of major importance, however, is the loss of stream competence in carrying its sediment load. Hence, the stream will deposit those particles it is no longer capable of transporting. The stream may adjust for the loss of competence by increasing its slope (Mackin 1948). Deposition may take place on point bars, in the river channel (which increases slope), and in flooded reaches below the new base level. Hypothetical depositional sequences for these areas of the stream must consider the loss of competence. Point bars may experience increased deposition with particle size either fining upward or being of uniform size (Knighton 1998).

A graded river flows in dynamic equilibrium and does not deposit significant amounts of sediment in its channel (Mackin 1948, Rubey 1952). The deposits within the channel are referred to as bedload sediment. Bedload is moved by a stream within the channel and transported mainly by rolling action (Bagnold 1960, Reineck and Singh 1980). Bedload deposits account for only a small percentage of total stream deposits. A loss of competence, as in the waning phases of a flood, may form a fining upward sequence in the channel. Deposition within the channel raises its altitude and could be indicative of loss of competence. A rise in base level can cause loss of competence. Channel deposits, however, are particularly subject to erosion should competence subsequently increase, (e.g., lowered base level), and may not be preserved.

Areas below the new base level may contain a fining-upward sequence (estuarine) of sediment deposited in the water. Deposits would include silt and clay-size particles which are now able to settle out in the still water. It should be noted that this could change to a coarsening upward sequence if the new accommodation space is filled and near-shore or deltaic deposits advance over marine, lacustrine, and estuarine deposits. These may, in time, even be supplanted by fluvial (topset) deposits to complete the coarsening upward sequence. Flooded areas that were formerly floodplains will receive deposition of only silt and clay, as normally happens during floods. As such, this type of deposit may be indistinguishable from normal flood-plain deposits.

The type, thickness and distribution of stream deposits will depend on several factors. The rate and amplitude of base-level changes will affect the amount of concomitant deposition or erosion. The rate at which base level rises and energy decreases will determine when and where progressively smaller particles will be deposited. Conversely, competence will increase proportionately to the energy added by the base-level decline. It bears repeating, that particle size is also controlled by the sizes available for erosion in the drainage basin.

Leopold and Wolman (1957) characterized the plan view patterns of river channels in alluvial streams as being braided, meandering or straight. They observed that these three patterns lie on a continuum from one form to the other and that "... each occurs in nature through the whole range of possible discharges" (Ibid., p. 72). Braided streams contain alluvial islands around which the flow divides into 2 or more channels. They will also have steeper slopes for a given discharge and greater discharge at a given slope than straight or meandering streams (Ibid.). They believed that a sinuosity (channel length/valley length) of 1.5 or greater without question signifies a meandering stream, though they admit that the "... value is an arbitrary one" (Ibid., p. 60).

Michigan streams lack continuous channel division around alluvial islands and are not braided in any observable sections. Michigan streams contain both straight and meandering reaches. There are a few instances where Michigan streams become anastomozing for very short stretches (~0.5 km) where they enter an estuary or a kettle depression, but these are inconsequential and not part of a larger flow pattern of the river. Therefore, the following discussion will concentrate only on the depositional features of meandering streams.

Point bars are a dynamic depositional form of meandering rivers. Point bars have been the subject of intensive research to describe the factors which control their deposition (see for example: Bagnold 1960, Guy et al. 1966, Jopling 1966, McGowen and Garner 1970). Controlling factors include variations in discharge, river channel morphology, sediment type and concentration, and river channel gradient (McGowen and Garner 1970, Simons and Richardson 1962). When analyzing an ancient deposit to determine its origin, factors that controlled deposition may no longer exist. Evidence can be gathered to deduce the influence of some of these factors.

Depositional environments of the point-bar facies vary widely in both the horizontal and vertical direction. Vertical variations usually involve the upward fining of sediment size. Depth of water on a point bar tends to decrease with time (in regard to the bankfull stage) as semi-continuous deposition raises its surface altitude. Shallow water has less available energy and combined with the lack of larger grain sizes in the upper part of the water column are the primary reasons for the vertical variation in grain size. Departures from a constant rate of flow can produce minor changes in the fining upward sequence but will not alter the overall trend.

McGowen and Garner (1970) presented a model of successive sedimentary structures seen in vertical cross section. The sequence may have channel lag gravel (bedload) at its base, depending on the position of the channel before the meander sequence began. The structures at the base were formed by water with the highest available energy levels in the stream. Typically, they are large scale trough or megaripple cross set stratification. Above that, small scale trough set or thin foreset cross stratification may be seen. A transition zone from the large scale (megaripple or large trough) type to the small scale (small ripple or small trough) type stratification is usually present. The trend will be toward the lower energy structures. The upper part of the sequence will have small ripples, climbing ripples, and small trough set cross stratification. A mud drape and vegetation structures such as root tubes may cap the sequence.

Depositional environments with horizontal (i.e. contemporaneous) relationships can be more variable than the vertical sequence. The thalweg is the line connecting the lowest points along the course of a stream and is the zone in a stream where the current is the strongest. As the river current begins to curve around the meander bend, centripetal force throws the thalweg to the outer margin of the curve (Bagnold 1960). This causes the more rapid flow vectors to impinge upon the outer bank and the slower flow vectors across the inner bank. This is the basic principle behind the formation of point bars (Ibid.). The energy loss on the inner bank causes a local loss of competence that results in deposition. The thalweg moves farther from the inner bank as the river flows around the meander bend, causing the energy level on the inner bank to decrease in the downstream direction as well. This results in a decreasing grain size downgradient along the point bar. It is important to note that these structures are only deposited when the stream is at a high enough stage to inundate the point bar.

Meandering streamflow adds more complexity in horizontal relationships. High-water stages can temporarily inundate the point bar and temporarily create a short-term high energy environment. During flood stage, the thalweg is wider; coarser grains and higher energy sedimentary structures can move over point bar. A two-tier point bar may develop (McGowen and Garner 1970). The lower tier is formed by the mean flow velocity; the upper tier is formed by peak flow periods. Flood chutes may also develop in the low area along the rear of the point bar (McGowen and Garner 1970, Knighton 1998). These chutes funnel some of the overflow behind the more fully developed, topographically higher portion of the point bar. The low area initially develops from the beginning stages of point-bar formation. At that time, the current pattern that creates point bars is not fully developed and sedimentation is low. Subsequent high water flows cause chute bars to be deposited in the flood chute. Grain size and sedimentary structures in chute bars are dependent on the strength of the flood. McGowen and Garner (1970) indicated that sand waves, foreset ripple, and climbing ripple lamination are the most common chute bar structures. Waning flood strength creates a fining upward sequence, often with a mud drape and vegetation structures on top. Successive flood periods can create alternating fining upward sequences within a composite chute bar.

Simons and Richardson (1962) described in detail the relationships between flow regimes, sedimentary structures, and grain size. They performed experiments with an artificial flume in which they could vary the water depth, flow velocity, and grain size. They distinguished two flow regimes based on velocity: tranquil flow (lower regime) and rapid flow (upper regime). Rapid flow is observed in streams with high velocity currents and steep gradients. Rieneck and Singh (1974) defined the transition between the two by the equation:

$$F = v/(gh)^{1/2}$$

where:

F - Froude number

v - flow velocity

g - acceleration due to gravity

h - height of water column

A Froude number >1 is in the rapid flow regime. Because the rivers in Michigan do not fit the rapid flow category by either velocity, gradient or Froude number, rapid flow need not be discussed here. Though it bears mention that rapid flow conditions probably existed during time of glacial meltwater.

Tranquil flow sedimentary structures are the dominant forms of fine-grained meander belt streams (McGowen and Garner 1970, Reineck and Singh 1980). Tranquil flow sedimentary structures (from lowest energy to highest energy) consist of small ripples, small trough sets, megaripples, and large trough sets. Grain-size distribution exerts a partial control on the type of structures that will form. When the mean grain size is greater than 0.6 mm (coarse sand), small ripples will not form. When the critical velocity is reached for particle motion on such a sand bed, the first structure that is formed will be megaripples (Simons and Richardson 1962). Structures formed in the tranquil flow regime were found to be independent of the depth of water (Guy et al. 1966, Simons and Richardson 1962). The controlling factors were the velocity of flow and, to the degree previously mentioned, grain size. Hence, depth of water in an ancient deposit laid under tranquil flow conditions cannot be estimated with a reasonable degree of accuracy.

Interpretation of depositional environment through time can be made by examining facies changes in cross-section. Examination of patterns of alluvial sedimentation and the location of permeable deposits vs. impermeable deposits continues to be a research subject (e.g. Bowen and Weimer 2003). Bridge and Leeder (1978) modeled stacking patterns in subsiding basins (analogous to base-level changes in non-subsiding basins), interpreting that stacking patterns of fluvial-sand bodies (typically channel and/or point-bar deposits in a floodplain) increased in amalgamation with decreasing rates of sediment production. The idea being that as normal avulsion processes proceed, sand bodies become vertically stacked closer together with decreasing accumulation rates of floodplain deposits. The model results also showed that amalgamation tended to increase with increasing ratio of channel width to floodplain width and varied inversely with rates of aggradation and avulsion frequency. Heller and Paola (1996) reconsidered fluvial amalgamation by linking avulsion frequency with sedimentation rate. This yielded a more realistic model of fluvial-stacking patterns where a high sedimentation rate would produce amalgamation when coupled with a high rate of subsidence.

Ouchi (1985) simulated depressions and uplifts in meandering alluvial rivers using sand flume models. He concluded that, for meandering rivers, increasing gradient tends to increase sinuosity with the reverse being true as well. Where the increase in gradient was caused by uplift, the uplift axis tends to be a site of erosion. The gradient increase also raises the stream power, allowing the river to handle the increased load from the incision. Beyond the uplift however, the river tends to aggrade because the increased stream power is lost, and with it load capacity. A depression increases the gradient on the upstream side with a decrease on the downstream side. If the increased gradient causes incision, the extra load will be lost with aggradation in the middle and downstream stretches. With enough loss of gradient, decreasing sinuosity can advance into channel anastomosing and ponding. Howard (1967, p. 2246) stated that analysis of stream patterns was useful for interpreting underlying structures, "particularly in areas of low relief."

Chapter 6

DEGLACIAL DRAINAGE DEVELOPMENT

The literature does not offer a wealth of information on the subject of drainage-basin development. Much more research is available on channel development and erosion/sedimentation styles, as detailed above. Horton (1945) discussed parallel patterns created as rill erosion forms from overland flow on a virgin surface. These subsequently develop cross-grading as micropiracy allows random rills to take over drainage of neighboring rills. In this way, drainage nets develop. In fact, there appears to be an order of self-similarity in the way drainage nets grow. This is seen experimentally in random-walk models (Leopold et al. 1963), deterministic models (Willgoose et al. 1991), and physical models (Schumm et al. 1987) and can even be observed on any freshly graded construction site after a rain event. However, it seems unrealistic to assume that a homogenous surface can exist in nature at or even remotely near the drainage-basin scale. As stated by Leopold et al. (1964, p. 420-421) "...there are few areas that, on a regional scale, provide a clean slate on which the channel network can develop unhindered or unconstrained by differences in rock type, structure, or initial topographic irregularities."..."In nearly all cases there are some inherited characteristics which will influence subsequent development or change and, furthermore, it is rare indeed that lithology and structure are so simple that homogeneity is characteristic of any large area."

Ruhe (1952) studied drainage density on glacial tills of the Des Moines lobe. He discovered that drainage density and stream length were positively correlated to the amount of time the till sheets were exposed to erosion. He further postulated that the rate of propagation of drainage pathways was greatest during the first 20,000 years of exposure. Thornbury (1969) dealt with the development of stream valleys by examining the dual controls of valley widening and valley lengthening. The latter said to proceed mainly by the process of headward erosion. In this way drainage basins may be consolidated and/or expanded. Thornbury was careful to point out that no major stream develops in such a way, but that "many minor tributary valleys were so formed" (Thornbury 1969, p. 104).

The development of drainage basins is said to generally proceed from proximal to distal reaches of the river. This process is dominated by headward erosion. Schumm (2005) characterized drainage basins as having three zones. The distal zone is the area where erosion occurs and the majority of entrained sediment is produced. The central or transfer zone is in equilibrium and characterized by input and output of sediment being equal. The proximal zone is the sediment sink where the fluvial system reaches base level and deposits its load in estuaries, deltas, fans, etc. This classification system recognizes the difficulty in casting an entire drainage basin into a single category given the variety of controls that can exist. It also recognizes that the upper reaches of the river generally have a positive sediment budget and are the prime agents of headward erosion.

The existing literature does not deal with one of the basic facts of drainage basins developed in glacial landscapes. In many glaciated areas (Michigan and northern Indiana in particular) drainage basins are often sequentially constructed from their distal areas to the proximal areas. Glacial retreat adds sections to the basin in step-wise fashion as the glacier margin moves basinward. The common pattern of regional slope and stream-network development corresponding to a pre-existing structural or stratigraphic pattern (Knighton 1998) does not apply. Rivers draining today into the Lake Michigan basin were constructed in discrete parts during retreat of the Lake Michigan lobe. Each discrete part has a unique depositional history that may not match that of the subsequently added segment. In addition, the relative positions of ice margins of the Huron/Erie and Saginaw lobes also imparted controls on the volume of meltwater into the distal part of each basin. This was particularly the case for the ancestral St. Joseph River whose drainage basin contains deglacial terrain lying within areas formerly occupied by the Lake Michigan, Huron/Erie, and Saginaw lobes. A realistic model of post-glacial drainage basin development must reflect abrupt contrasts of sediment properties and slope gradients related to each successive ice-marginal position.

Chapter 7

CHARACTERISTICS OF THE ST. JOSEPH RIVER

The St. Joseph River and its tributaries (excepting the Paw Paw River which joins the main stem at its mouth) drain approximately 11,137 km2 (4,300 miles2) in southwestern Michigan and northcentral Indiana (Figure 1). It originates from Baw Beese Lake, one kilometer southeast of Hillsdale, Michigan. The river flows a total of 316 kilometers (196 miles) through a decline of 158.8 meters (521 feet). The Great Lakes Basin Commission (1975) framework study quotes the decline at 174.7 meters (570 ft). This appears to have been calculated by the height of the adjacent drainage divide, not the fall of the river itself.

The surface material in the drainage basin is almost entirely unconsolidated glacial deposits. Glacially derived sediment provides the full range of sediment sizes for erosion and transport. Moraines can provide clay through boulder-size particles. Lacustrine deposits are typically composed of clay, silt and fine sand. Outwash plains contain mostly sand and gravel. The only known natural bedrock outcrops are in the upper St. Joseph River basin in the vicinity of Hillsdale and Branch Counties, Michigan. In Hillsdale County, between the towns of Hillsdale and Jonesville, the Mississippian Marshall Sandstone crops out in places from under a thin mantle of diamict. The Marshall is composed of medium to coarse sand with shale interbeds in this area (Catacosinos et al. 2001) and is compacted and partially carbonate cemented. The type-location of the Mississippian Coldwater Shale outcrops along the Coldwater River near the town of Coldwater. The Coldwater Shale is generally described as an illite/kaolinite dominated shale, grayish in color, containing a minor amount of silt (Harrell et al. 1991). Several upland creeks are also reported to contain outcrops of Marshall Sandstone and Coldwater Shale in these same counties (Martin and Straight 1950, Michigan Geological Survey, unpublished outcrop location report).

The St. Joseph River basin became ice-free in several steps. The Saginaw glacial lobe retreated from Indiana to Michigan first to the Sturgis moraine and then to the Tekonsha and Kalamazoo moraines (Figure 10). This effectively deglaciated the middle portions of the drainage basin (from south to north) while the lower and upper reaches were still beneath the Lake Michigan and Huron/Erie lobes, respectively. At this time (about 15,000 ^{14}C B.P.) drainage must have overflowed to the south into the Wabash River basin, though this fact appears to be absent from the literature. The cross-cutting relation by which Saginaw lobe landforms cross Lake Michigan and Huron/Erie lobe features leads some to believe that the latter two advanced at least partly across the former before it had fully retreated. Kehew et al. (1999) interpreted the palimpsest relationship between linear, linked-depression features commonly associated with the Saginaw lobe that appeared within landforms deposited by the Lake Michigan lobe was a clear indication that the latter actually overran landforms of the former. By the time the Saginaw lobe had retreated from the Tekonsha and Kalamazoo moraines the Lake Michigan lobe still occupied the outer Kalamazoo moraine. The Miami Highway Ridge (Figure 22) represents at least one stand of the glacier that must have blocked drainage to the Kankakee River.

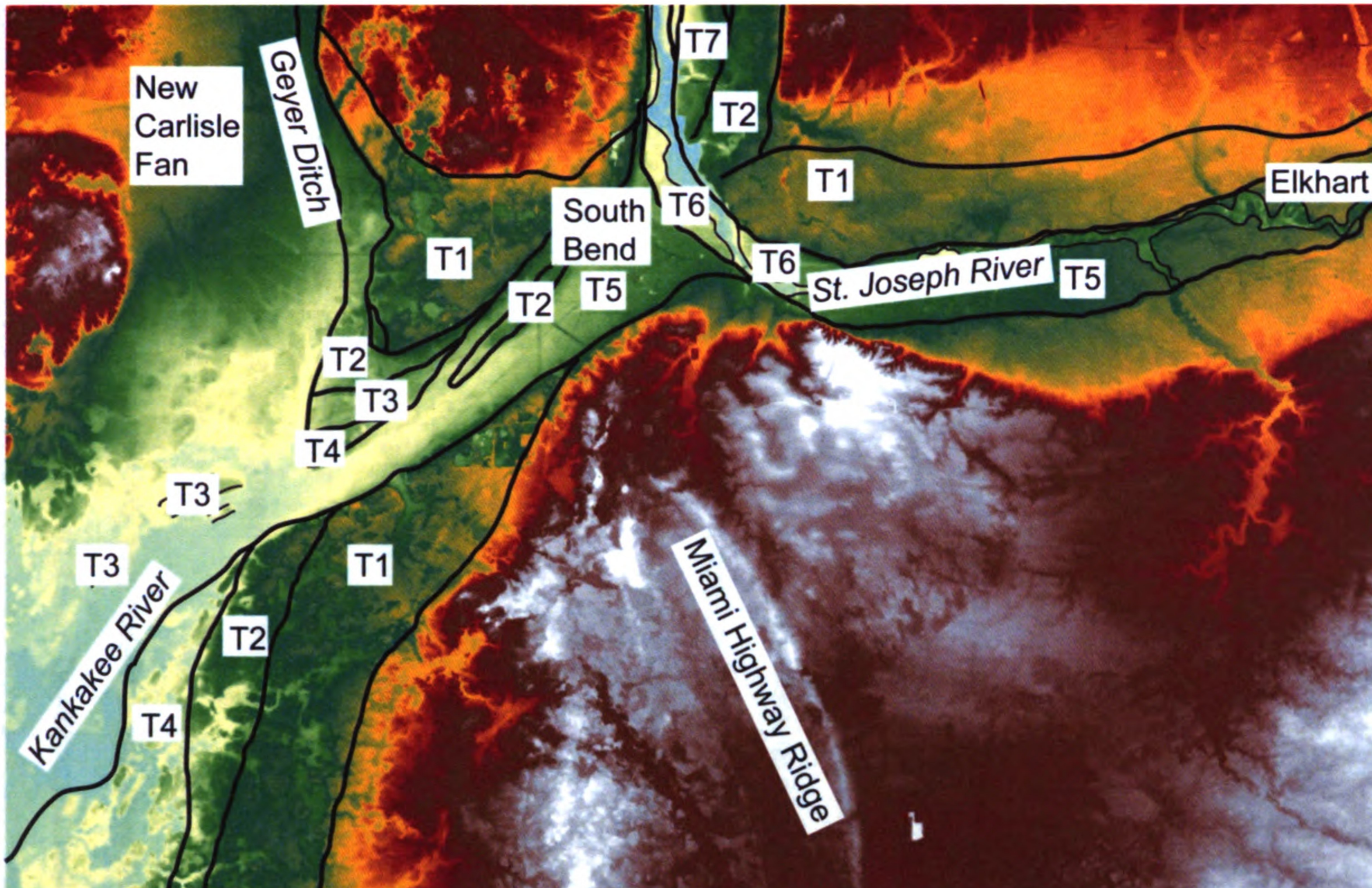

Figure 22. Terraces of the St. Joseph River in the vicinity of South Bend, Indiana as seen on 5 meter DEM (courtesy of Indiana Geological Survey).

When the Lake Michigan lobe pulled out to the inner Kalamazoo moraine the drainage from the St. Joseph River basin switched from the Wabash to the Kankakee River. A well-developed terrace on the west shoulder of the Maxinkuckee moraine (Brown 2003) southwest of South Bend, Indiana attests to water flowing at an altitude of 233 m ASL (Figure 22). It is possible that drainage across this terrace, early in its development, may have still been directed towards the Wabash River. A glacier at the Maxinkuckee moraine strongly suggests that a lake, also draining south to the Wabash River, must have existed in the basin east of South Bend, Indiana. In fact, Brown (2003) reported lacustrine deposits in the upland south of South Bend associated with a morainal ridge of the Maxinkukee moraine. These lacustrine deposits, at an altitude between 262-259 m further attest to southward drainage to the Wabash River when ice of the Lake Michigan lobe blocked westward drainage.

Once the glacier retreated to the Valparaiso moraine, the St. Joseph drainage basin flowed through the low saddle in South Bend (currently occupied by the Dixon West Place Ditch at 221 m) to the Kankakee River rather than overflowing across a sill blocked by ice (Ekblaw and Athy 1935). Deposits attributed to this period of high flow in the Kankakee River are found through to the Illinois River (McKay et al. 2005).

North of South Bend, in the area that would later become Dowagiac River (a tributary of the lower St. Joseph River), glacial Lake Dowagiac (Russell and Leverett 1908) formed in the low lands between the inner Kalamazoo moraine and the Valparaiso ice margin (Figure 23). This lake overflowed south into the

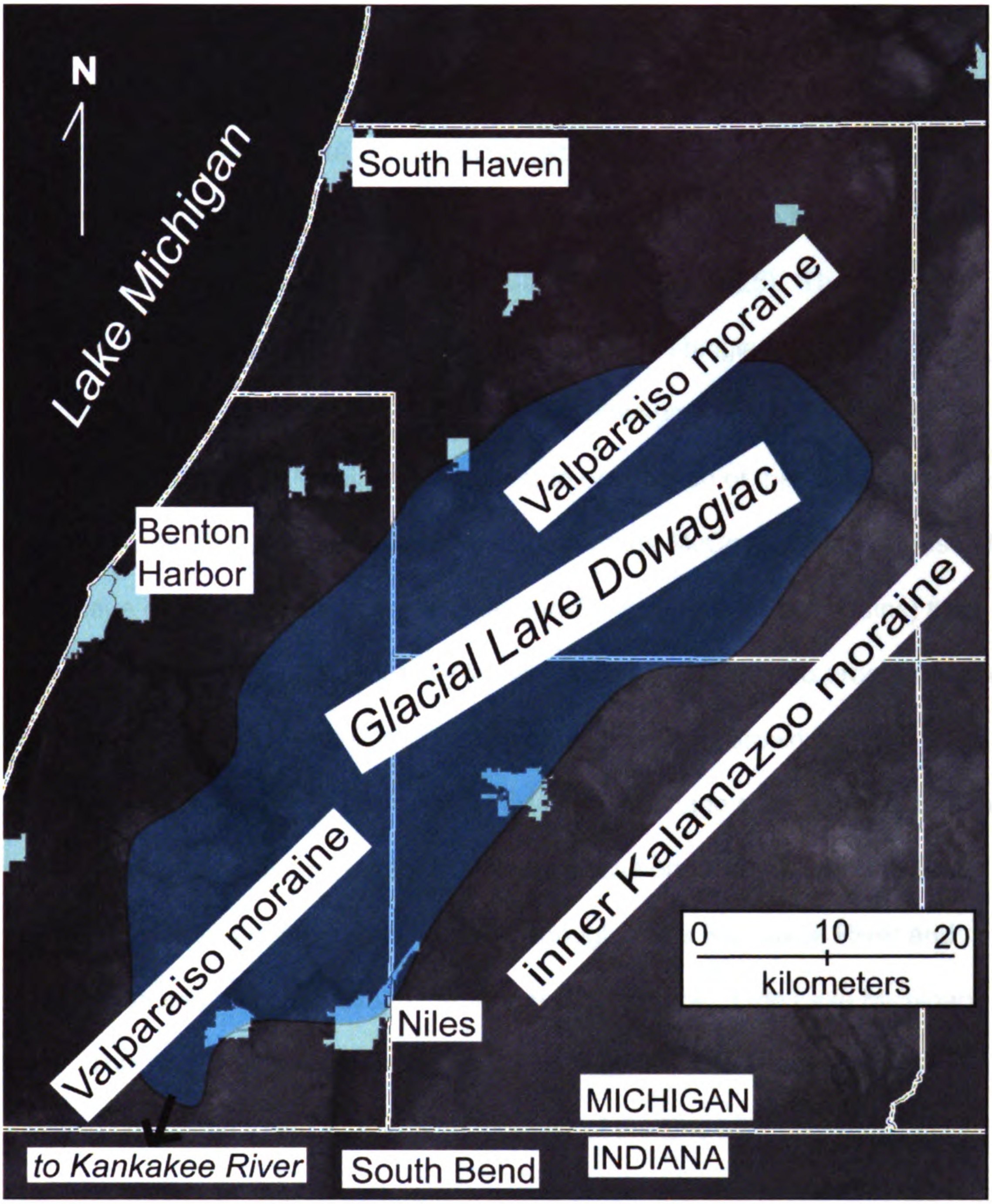

Figure 23. Map showing the probable extent of glacial Lake Dowagiac between the inner Kalamazoo moraine and the Valparaiso moraine. Base map is the U.S. Geological Survey 30 m DEM.

Kankakee drainage, likely utilizing the sill between McCoy Creek (south of Buchanan) and Geyer Ditch (west of South Bend) which is at an altitude of 222.5 m (Figure 24). Glacial Lake Dowagiac probably also used the southeastern Barron Lake outlet at an altitude of 227 m that passes between Niles and South Bend before the McCoy Creek outlet was available (Figure 25). The Valparaiso moraine in Berrien County is composed of a series of shingled, fluvio-deltaic complexes that were built out into glacial Lake Dowagiac until the ice margin retreated into the Lake Michigan basin prior to occupation of the Lake Border ice margin (Stone 2002, Stone et al. 2003). South of the McCoy Creek sill (presently a drainage divide between the Kankakee and St. Joseph drainage basins) the Valparaiso moraine is predominantly fluvial because there was no proglacial lake in which lacustrine and deltaic sedimentation could occur (Brown 2003).

Terraces in the upper and central part of the drainage basin indicate that the last input of water and sediment with a direct glacial origin came from the Erie lobe. This water appears to have traveled though to the Kankakee River and therefore should be correlative to the Valparaiso moraine of the Lake Michigan lobe. This runs counter to the existing chronology that has the Valparaiso moraine correlated to the Fort Wayne moraine of the Erie lobe (Martin 1955, Fullerton 1980). In fact, asynchronous movement of the glacial lobes in Michigan has become an overriding theme of evidence being gathered recently (e.g. Brown et al. 2006, Kehew et al. 2005).

The standard geologic history for the area also holds that the St. Joseph River entered the Lake Michigan drainage basin when the ice retreated from the

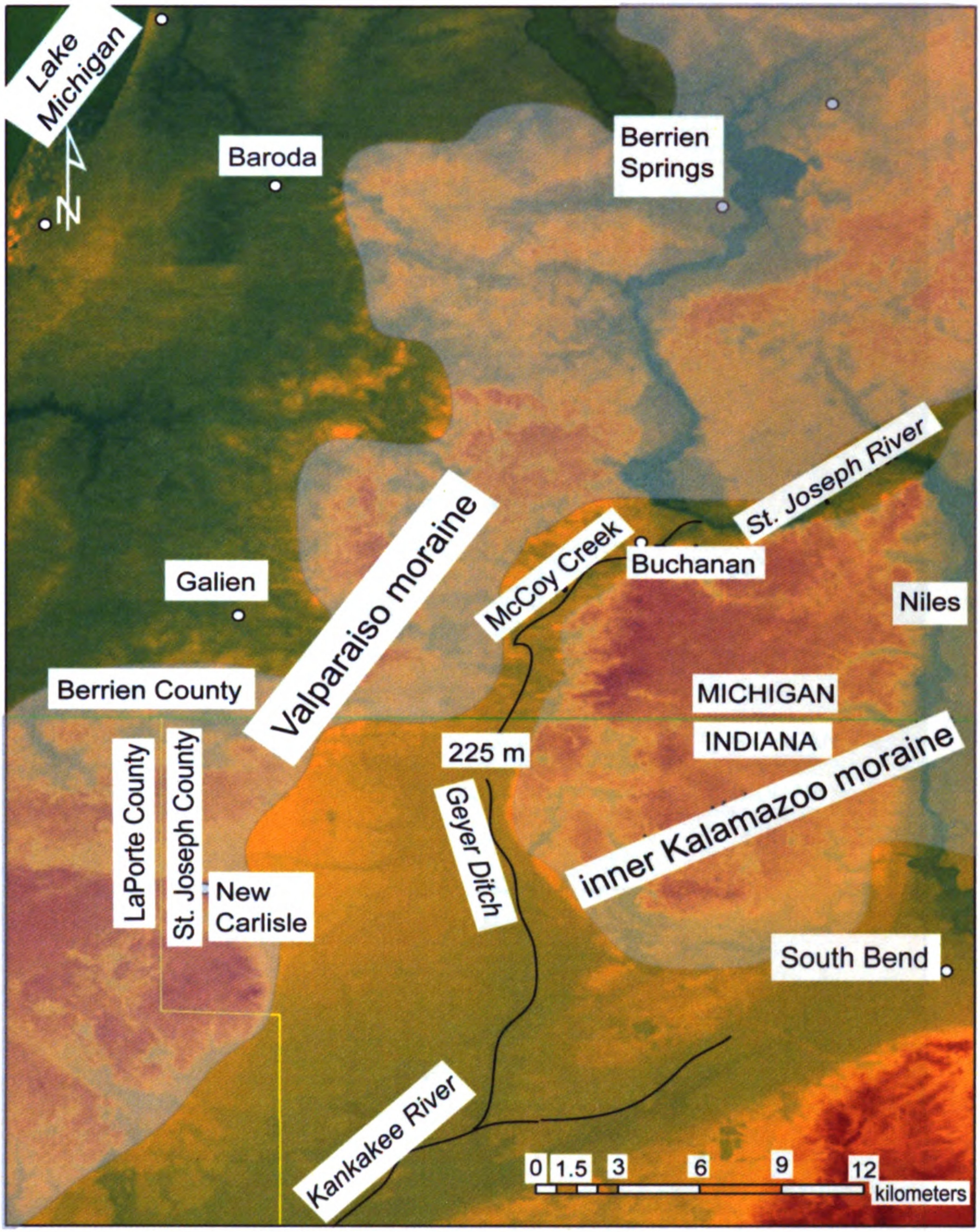

Figure 24. Map of the McCoy Creek outlet at Buchanan, Michigan of glacial Lake Dowagiac. The height of land between the present St. Joseph River basin and the Kankakee River is 225 m at the divide between McCoy Creek and Geyer Ditch. Base map is the U.S. Geological Survey 30 m DEM.

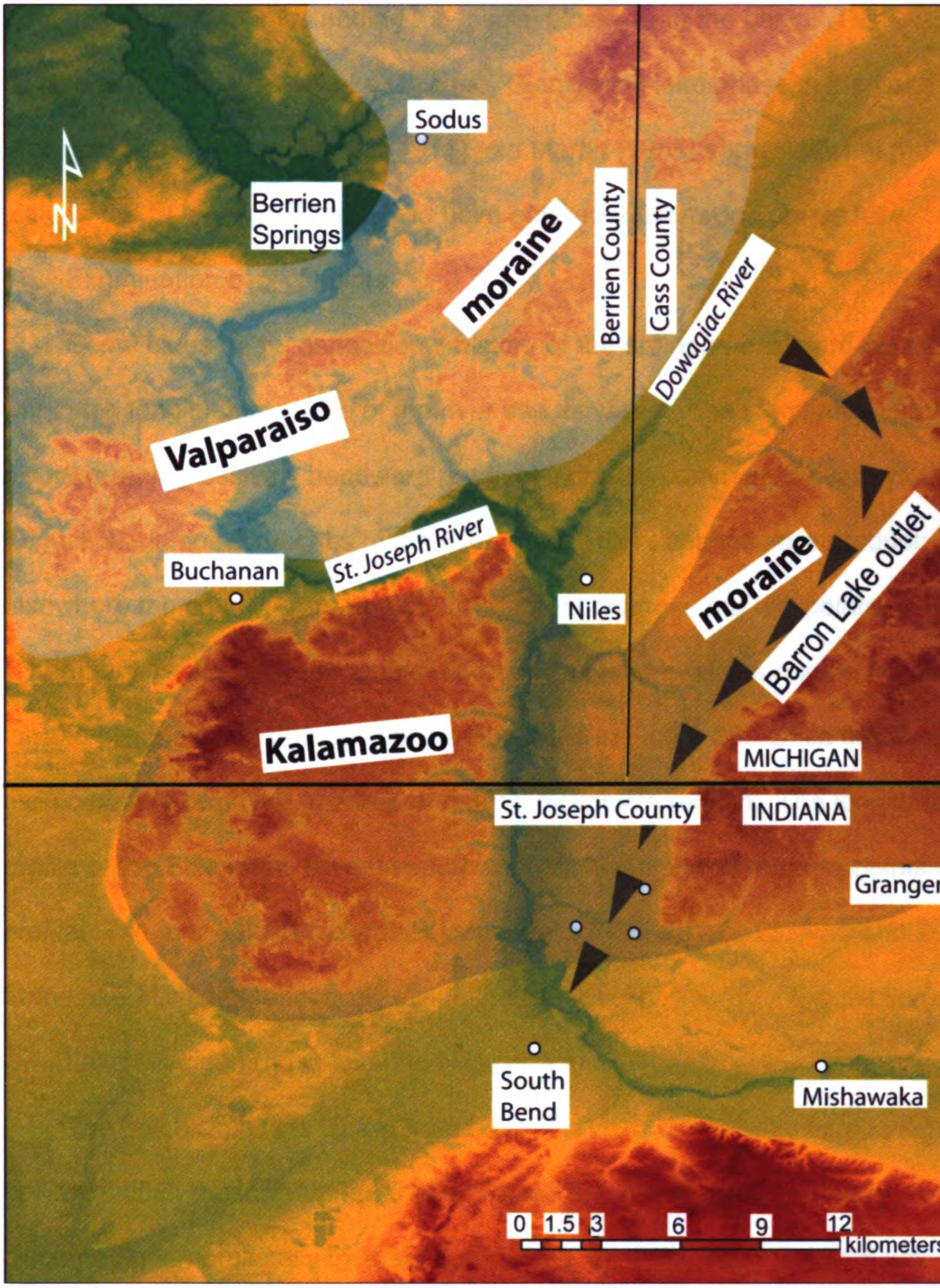

Figure 25. Map of the Barron Lake outlet of glacial Lake Dowagiac east of Niles, Michigan. This outlet was used before the McCoy Creek outlet became available. Base map is the U.S. Geological Survey 30 m DEM.

Valparaiso moraine, thus ending the glacial blockade of the basin. This coincides with the glacial Lake Chicago, Glenwood I lake phase in the Lake Michigan basin. Both Leverett (1911) and Martin (1955) show deltas at the altitude the St. Joseph River would have entered a Glenwood phase lake.

Examination of Nickpoints

Some clues to the depositional history of the drainage basin are contained within the longitudinal profile of the river and its valley. The longitudinal profile displays a progressively headward gradient increase common to river systems (Figure 3). In some reaches of the river, the gradient changes are gradual; elsewhere the changes are abrupt. The abrupt changes are nickpoints (Thornbury 1969). Anthropogenic nickpoints caused by dams can be accounted for and eliminated by interpolation as illustrated in Figure 26. Once these cultural nickpoints are removed, the true or natural nickpoints remain. The longitudinal profile of the St. Joseph River in southwest Michigan shows four nickpoints in its profile. The four natural nickpoints located at 37 km, 68 km, 237 km, and 293 km from Lake Michigan, effectively divide the St. Joseph River into five sections.

Lake Michigan (base level) to Nickpoint 1

The first nickpoint is 37 km from Lake Michigan and is seen as a sudden upward shift in the longitudinal profile (Figure 3). The nickpoint is easily discerned geomorphically by a sudden narrowing of the valley from 1.6 km downstream to less than 0.5 km upstream. The average gradient of this section is 0.16 m/km. Below the nickpoint, the lowest section of the river is characterized by a distinctly non-classical valley with a meandering river (sinuosity 1.35)

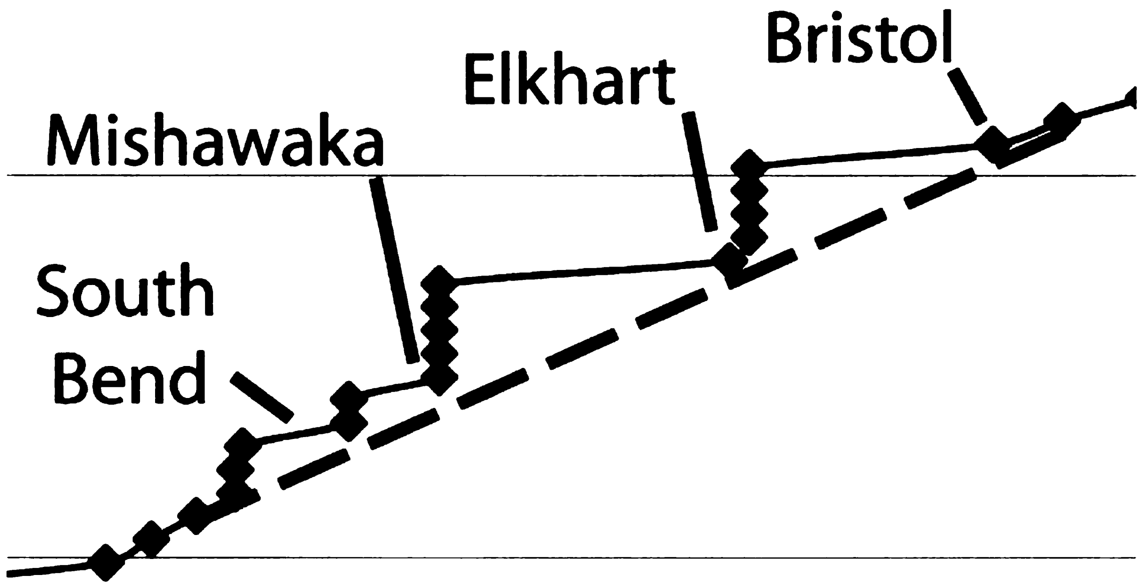

Figure 26. Interpolation of original stream gradient through artificial dam ponds. Blue diamonds indicate location of topographic contours crossing the stream channel. In this case, contours are 5 ft.

confined by steep, 12 m tall valley walls (Kincare 2007). Fluvial features within the walls of this incised valley include scroll-type point bars, a well-developed meander pattern, oxbow cutoffs, and a set of terrace fragments just above the level of the active floodplain. The walls of the valley truncate glacial deposits of the outer Valparaiso and Lake Border moraines as well as terrace deposits graded to glacial Lake Chicago (Kincare 2000). This nickpoint is interpreted to have resulted from headward erosion during the Chippewa low lake phase.

The Rock Hearth site (Figure 27), an archeological site on the modern floodplain 3 km downstream from Berrien Springs, Michigan, yielded a ^{14}C age of 3740 +/- 80 years B.P. (Garland 1984). The material for this date was recovered at a depth of less than 0.5 meters. This showed that very little alluviation of the floodplain has occurred since aboriginal occupation of the Rock Hearth site.

There is a set of terrace remnants sitting on the edge of the floodplain in several places (Figure 27). These terrace remnants are about 3.7 m above the present floodplain within the confines of the incised valley. Excavations by the author in the terrace remnant next to the Rock Hearth site revealed cross-bedded coarse sand and fine gravel in several sets ranging from 10 to 70 cm thick (Figure 28). Flanking the coarse terrace deposits is an apron of silty, fine sand. In the saddle between the terrace and the confining valley wall there is a 1.2 meter thick, single foreset-bed of very well-sorted, medium sand (Figure 29). This bed was capped with 0.3 m of slightly organic fine sandy silt and clay. In combination these deposits indicate that the terrace is a point bar with a distal

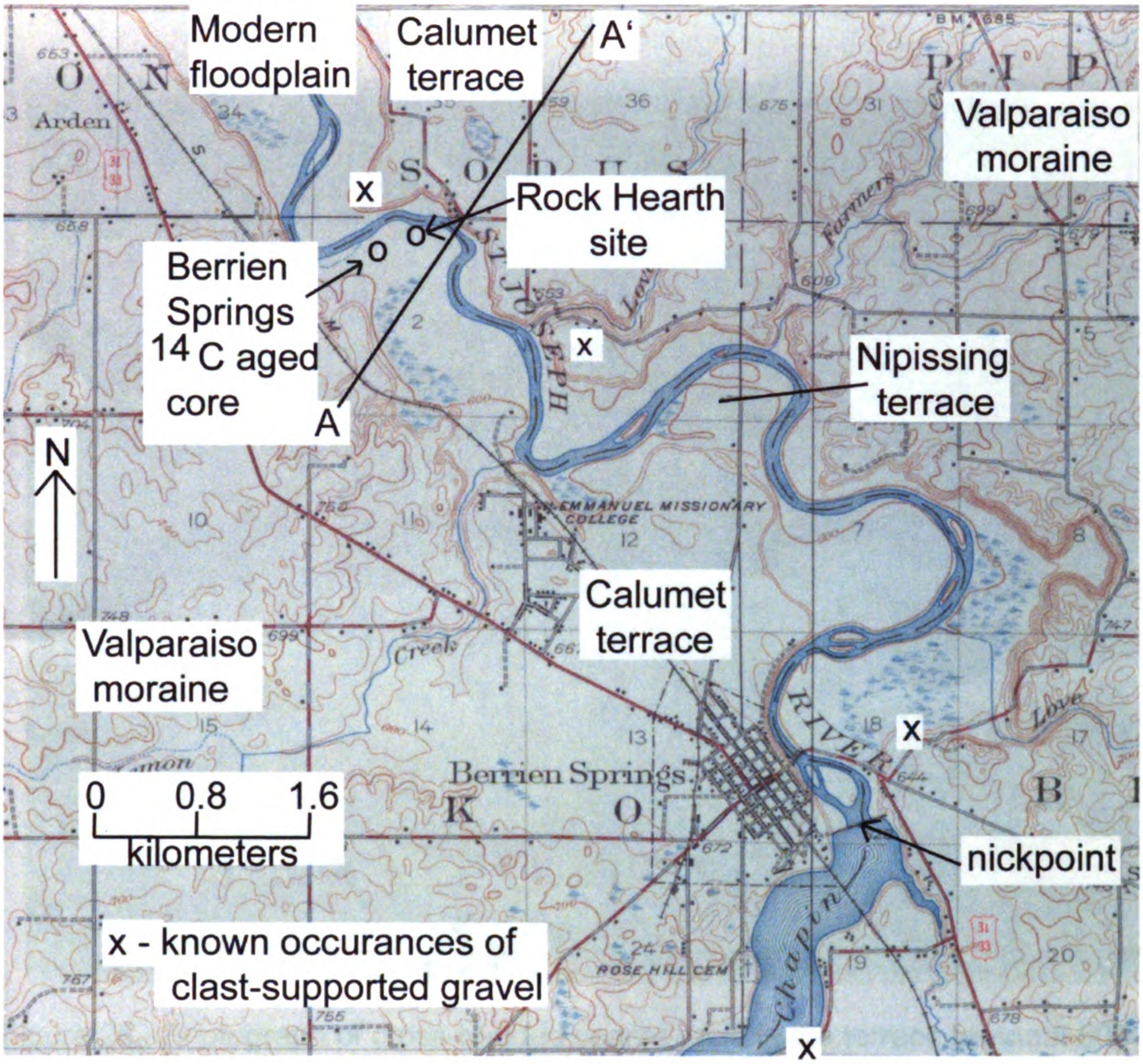

Figure 27. Map showing terrace remnants above the modern floodplain of the lower St. Joseph River. Base map is from the Benton Harbor 15 minute quadrangle.

Figure 28. Photograph of cross-bedded coarse sand in the terrace remnant 500 m west of the Rock Hearth site. Film canister in upper right is 5 cm tall.

Figure 29. Photograph of single foreset bed (overlain by 4 sets of current-rippled medium sand and a mud drape) of very well-sorted medium sand 500 m west of the Rock Hearth site. Exposed face is 1 m tall.

flood chute surrounded by floodplain deposits. It can be further concluded that normal fluvial processes were active between 2.5 - 4.5 meters above the modern floodplain. It has been here established that the present valley did not exist before the downcutting caused by the Chippewa regression. The valley was refilled with sediment during the Nipissing transgression. Therefore, the above deposit is interpreted to be the fluvial deposits left by the river during readjustment to the declining base level from the Nipissing phase high level to the lower modern level. The Rock Hearth site radiocarbon age indicates that this readjustment was concluded no later than 3740 +/- 80 years B.P.

Cores for this project near the same location went through 27 m of valley-fill sediment before reaching the underlying glaciolacustrine deposits (Figure 30). ^{14}C dating of organic remains in the cores (location shown on Figure 27) yielded the following dates:

TABLE 3. ^{14}C ages and altitudes from Berrien Springs core.

Sample ^{14}C age	sample altitude	material
5320 +/- 70 BP (Beta 175284)	179.5 m ASL	wood
5610 +/- 60 BP (Beta 175285)	178.0 m ASL	wood
5830 +/- 60 BP (Beta 175286)	177.0 m ASL	peat
6390 +/- 110 BP (Beta 175282)	173.8 m ASL	peat
6860 +/- 80 BP (Beta 175283)	172.0 m ASL	org. sed.

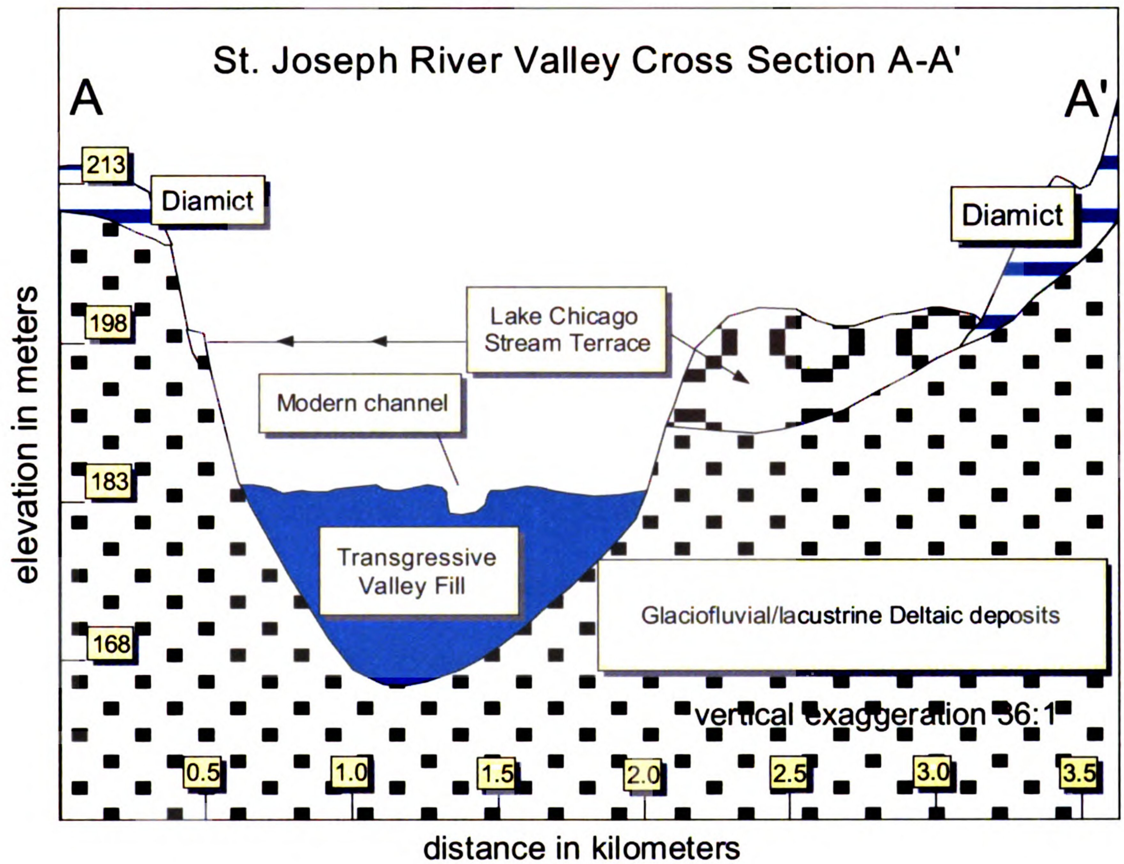

Figure 30. Cross-section of lower St. Joseph River valley sediments. Section location shown on Figure 27.

The floor of the incised valley is relatively flat, as are most of the rivers that drain into Lake Michigan. This is due to the massive incision event during the Lake Chippewa lowstand followed by partial filling during the subsequent Nipissing trangression. Cores and electromagnetic soundings taken from the floodplain indicate about 20 - 30 m of valley fill (Figure 30). With valley walls averaging about 15 m above the floodplain the total incision is as much as 45 m. Sediment removed just from the portion of the St. Joseph River valley that is presently subaerial accounts for some 1.16 million m^3 of sediment, some of which likely resides in a lowstand delta approximately 30 km off-shore. The altitude of the bottom of the incised valley also requires that the valley continue offshore from the present mouth of the river. The 30 m bathymetric contour is about 12 km offshore (Figure 31).

Downstream of Berrien Springs, core data indicate that in the 1,500 yrs preceding the Nipissing high stand 7.5 m of valley fill was deposited. Excavations of Nipissing age terrace remnants shows active point bar accretion at altitudes 4 m above where floodplain and point bars are forming today. Therefore, 4 m of degradation in the floodplain due to post-Nipissing base-level reduction is postulated (Kincare 2007). Dates on organic material that accumulated in oxbows also appear to date to an early post-Nipissing regression time when the St. Joseph River was mildly rejuvenated (Garland 1984).

Most of the city of Benton Harbor is built upon a flat plain of sand and gravel called the Fairplain deposit (Stone 2002). This deposit sits 15 - 18 m above the incised valley of the St. Joseph River. Its average altitude is about 189 m, rising

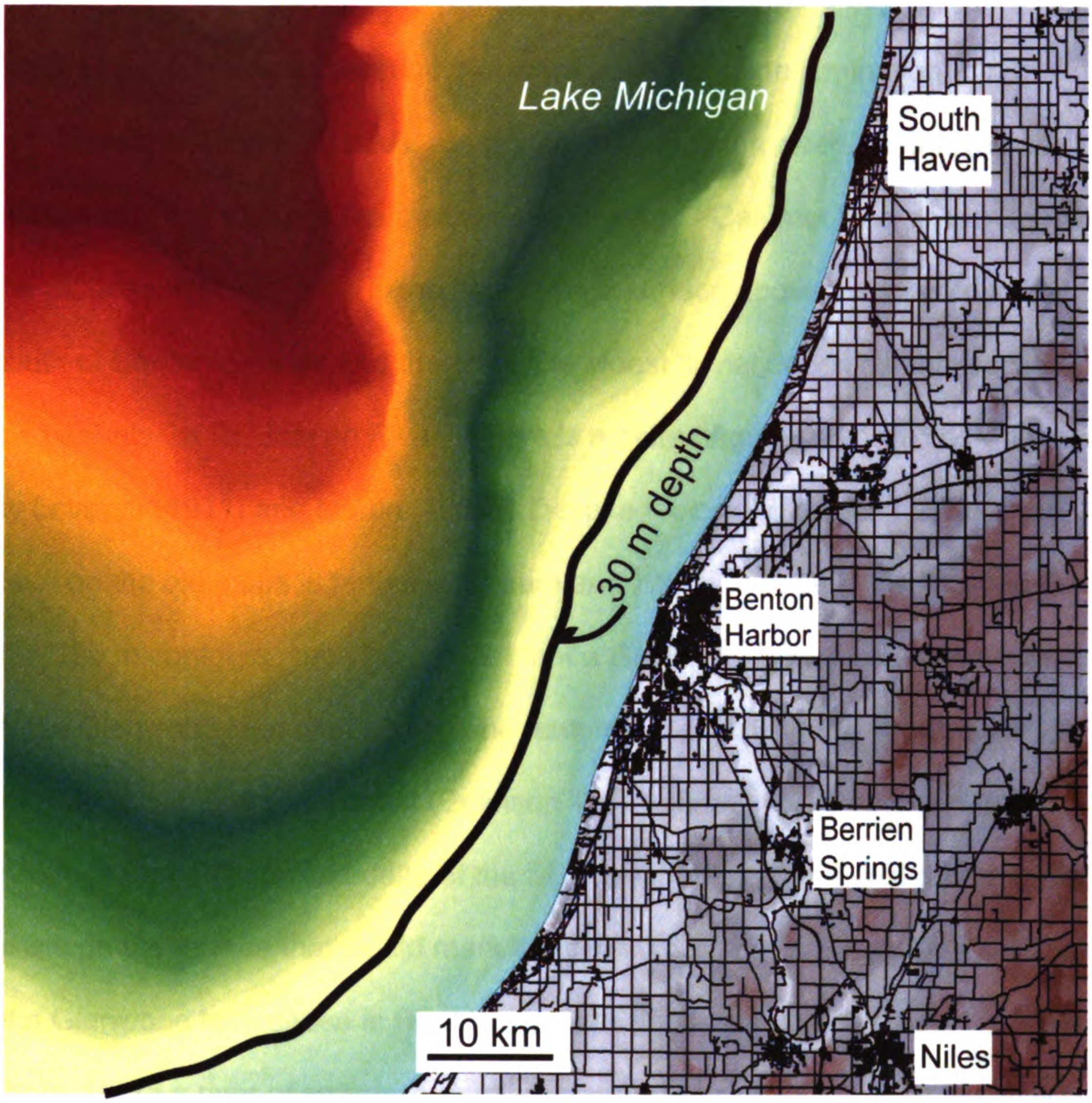

Figure 31. Bathymetry offshore from the present mouth of the St. Joseph River with 30 m bathymetric contour indicated.

at about 0.35 m/km upstream toward Berrien Springs (Figure 32). Exposures in Benton Harbor show a coarsening-upward trend from fine lamination to ripple-drift lamination to foreset beds. Based on altitude, interpreted base-level, and stratigraphy this must be the glacial Lake Chicago - Calumet lake phase delta of the St. Joseph River. This is the only terrace surface above the incised valley south of Berrien Springs and therefore the oldest fluvial deposit in this section of the post-glacial St. Joseph River. There is a delta (Figure 33) depicted on maps by Leverett (1911) and Martin (1955). Despite their small scale, it is clear the delta on the old maps is farther upstream and intended to be drawn around the 198-195 m contours, implying a belief that a Glenwood lake phase delta exists. While Glenwood lake phase shoreline features are found in the area parallel to and inland of the present coast (Evenson 1973), there is presently no strong evidence that a Glenwood delta of the St. Joseph River exists and no nickpoint exists on the terrace that would mark the delta if it existed. Given the existence of a Glenwood lake phase in the area, a delta should exist at all rivers along its shore where a river of sufficient discharge and load existed. A large Glenwood delta exists 130 km to the north on the Grand River near Allendale. The Grand River has a similar size drainage basin to the St. Joseph River. Therefore, conditions for building a Glenwood delta were possible. There is also no nickpoint or incision at the Glenwood phase altitude along the terrace (Figure 34) as would be required when the base level dropped from Glenwood phase to the Calumet phase. This statement is corroborated by the existence of terrace fragments created by the drop in altitude from Nipissing lake phase to modern

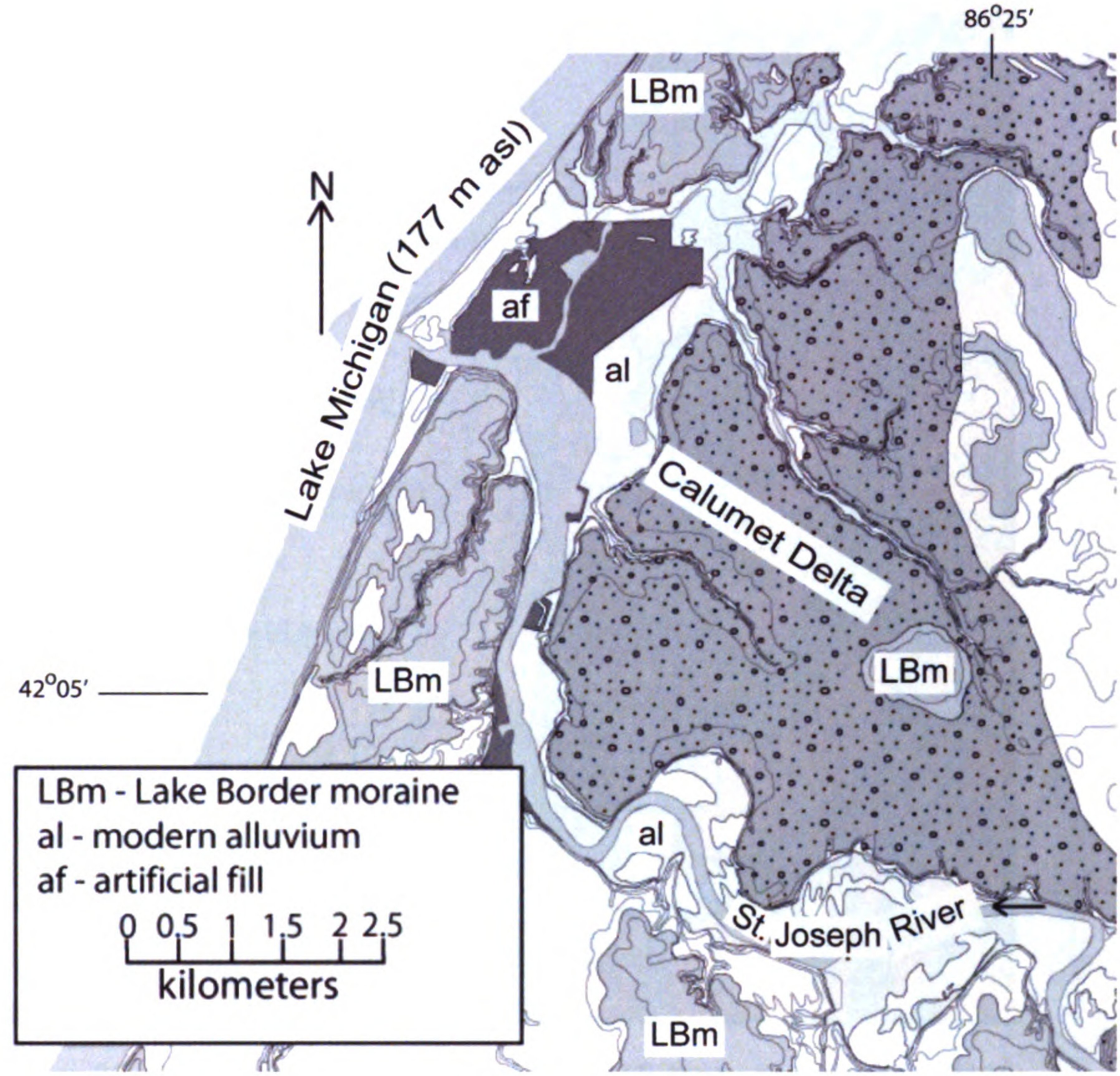

Figure 32. North-west Benton Harbor 7.5 minute quadrangle showing Calumet delta of glacial Lake Chicago.

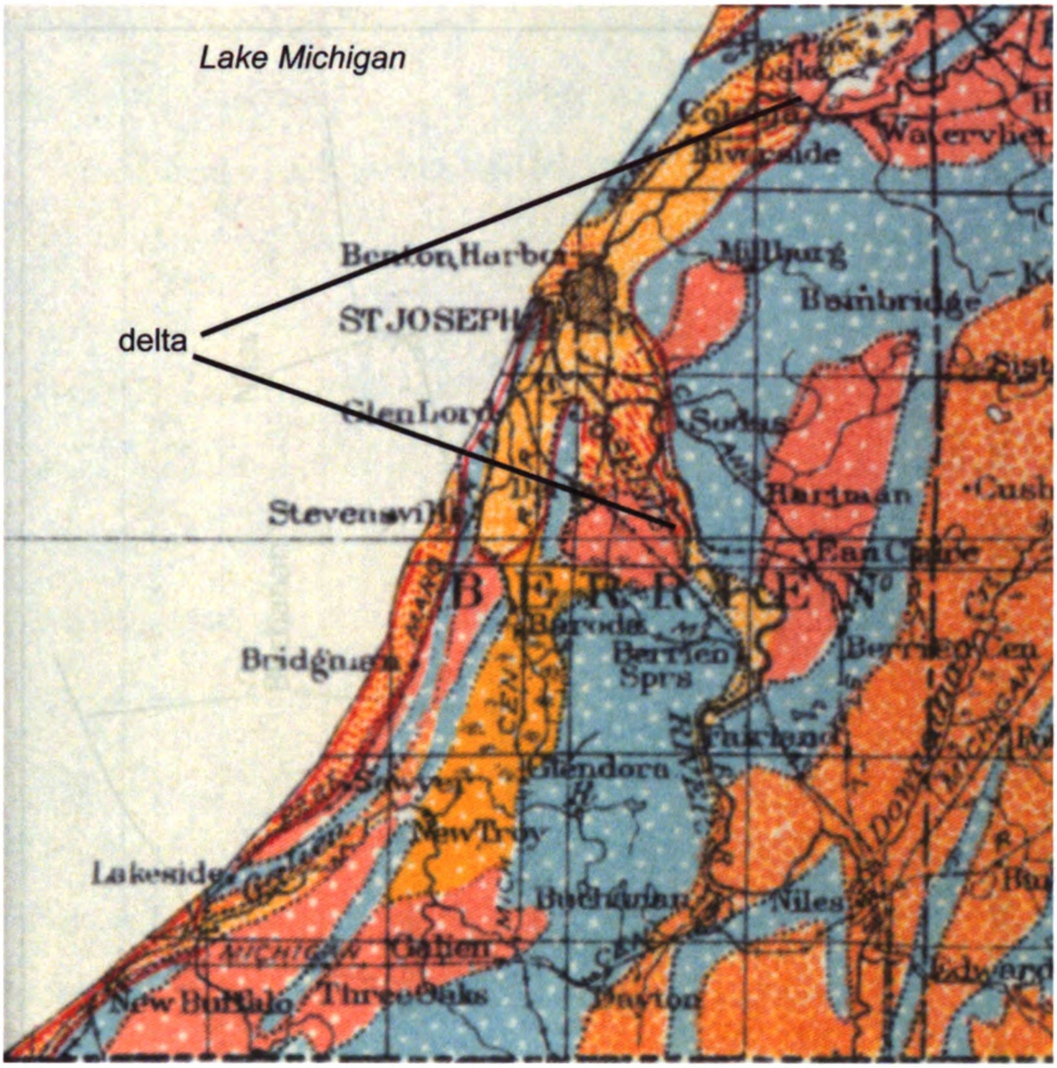

Figure 33. Section of glacial geology map of Michigan by Leverett (1911) indicating deltas at the glacial Lake Chicago, Glenwood phase. Dashed red lines indicate delta splaying outward from an origin in the vicinity of the 195 m (640 ft) contour just downstream of Berrien Springs as well as at Coloma on the Paw Paw River.

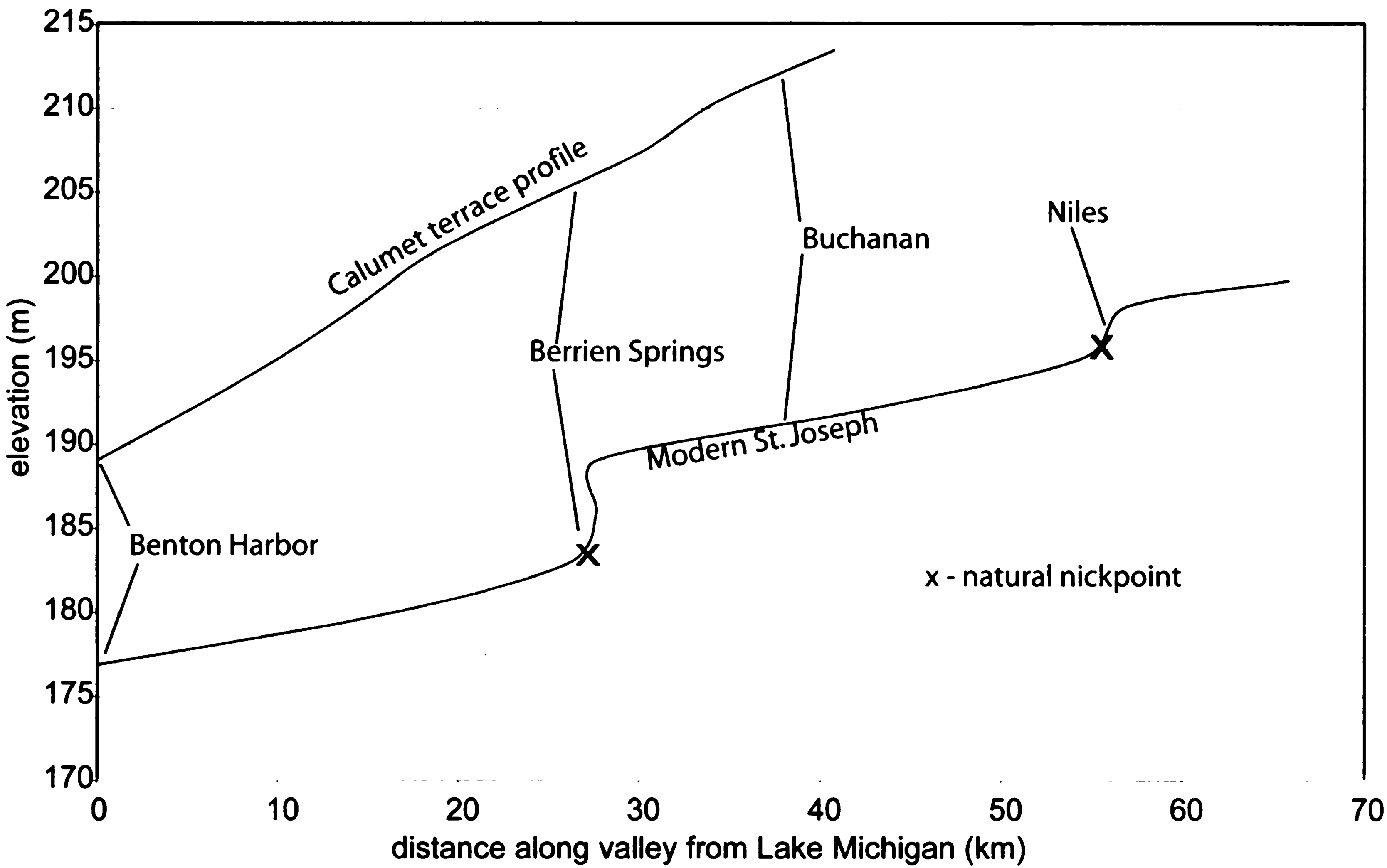

Figure 34. Longitudinal profile of part of the St. Joseph River with the glacial Lake Chicago, Calumet phase terrace superimposed.

lake phase, an equivalent base-level drop from Glenwood lake phase to Calumet lake phase.

The implications of this are twofold. One, this interpretation constrains the entrance of the St. Joseph River into the Lake Michigan basin until after the Glenwood lake phase and before the end of the Calumet lake phase. The broad, well-defined extent of the delta indicates the river began depositing early in the Calumet lake phase. Two, the upper St. Joseph River must have continued to discharge into the Kankakee River throughout the Glenwood lake phase, thus giving us an idea of how long it took for the kettles to melt out and for the northerly drainage out of South Bend to consolidate.

A 2 m thick layer of clast-supported gravel, characteristic of high discharge, exists in the lower portion of the terrace graded to the Calumet delta (Figure 35). Flow direction from cross-bedding in the gravel indicates a generally north to northwest trend, as it should for flow bound for glacial Lake Chicago. The gravel is seen in several outcrops between Benton Harbor and Niles, Michigan. A few locations show glaciolacustrine deposits directly below the clast-supported gravel of the terrace deposits. This type of open-work gravel is often seen as a facies in the ice-contact outwash deposits but rarely, if ever, in post-glacial alluvial deposits in Michigan. The reason for this is a general lack of stream power in these relatively small Great Lakes drainage basins typified by low gradients surrounded by high-porosity surficial materials and dependant on ground-water discharge for as much as 90% of base flow (Grannemann et al. 2000). This clast-supported gravel leads this writer to the conclusion that it marks an event in

Figure 35. Photograph of the clast-supported gravel in the lower portion of the terrace graded to glacial Lake Chicago, Calumet phase observed along Lake Chapin Road in Section 2, Buchanan Township, Berrien County. Camera lens cap is 5 cm diameter.

this part of the drainage basin; an event with a high discharge. The terrace and delta evidence indicate that the St. Joseph River was not draining to the Lake Michigan basin until early in the Calumet lake phase. Backcutting, aided by ground-water sapping of a tributary southward from Niles combined with melting of kettles associated with the inner Kalamazoo moraine between Niles and South Bend created the conditions for stream piracy of the St. Joseph River from the Kankakee River basin into the Lake Michigan basin. It is proposed that this occurred as an event with a sudden discharge that entered a valley graded to a small stream (probably equivalent of Dowagiac Creek), eroded into the glaciolacustrine deposits of glacial Lake Dowagiac, and left the high terrace whose base consists of a 2 m thick layer of clast-supported gravel.

Nickpoint 1 to Nickpoint 2

The first nickpoint is situated just below the artificial nickpoint of the Lake Chapin dam. In fact, this hydroelectric dam site was chosen in 1903 because of the narrowing of the valley at that position. Above the nickpoint, the St. Joseph River has a straight channel (sinuosity of 1.06). The flood plain is narrow to nonexistent in the section within the Valparaiso Moraine. The valley walls, though not as steep as in the previous river section, encroach upon the river; valley width is no greater than 0.5 km. While it is true that part of this section is within the Chapin Dam pond, it is clear from Figure 36 that the valley would be barely wider than the river even without the dam. The gradient above the first nickpoint increases to 0.30 m/km (1.6 feet/mile). In this stretch the river has one major tributary and several short tributaries. The short tributaries cut sharp

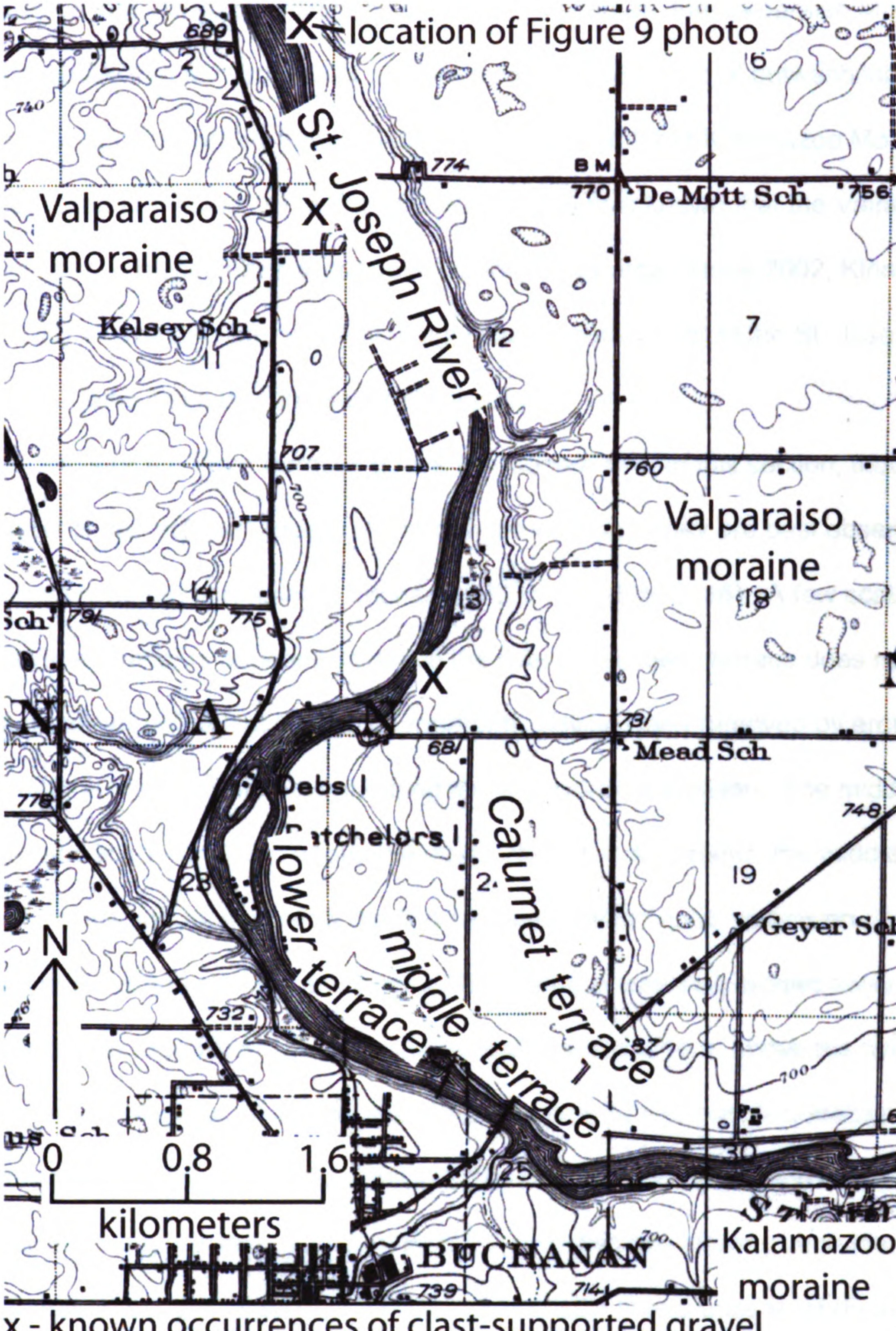

Figure 36. Map showing a portion of the St. Joseph River between Buchanan and Berrien Springs with terraces above the present river channel.

valleys into the surrounding uplands. The river enters the Valparaiso Moraine 8 km above the dam from what had been considered a strip of outwash/lacustrine sand deposits that lay between the Valparaiso and Inner Kalamazoo Moraines (Leverett and Taylor 1915, Farrand 1982). It is now known that the Valparaiso moraine is glaciodeltaic to glaciolacustrine in this area (Stone 2002, Kincare 2000). Dowagiac Creek, the first significant tributary river of the St. Joseph River, joins just below the second nickpoint.

There are three terraces above the modern river in this section, though none of the terraces appear all through the section. They are best observed east of the river just northeast of Buchanan, Michigan (Figure 36). A few scattered terrace remnants appear 6 m above the river. This lower terrace does not appear below the first nickpoint, apparently having been removed by erosion from the major incision event during the Chippewa regression. The middle terrace is about 10.5 m above the river. Though in fragments, the middle terrace also does not appear below the first nickpoint (based on an extension of its apparent gradient). Any evidence of its existence was also eroded away by the Chippewa regression. The highest terrace lies about 18 m above the river and is the most extensive of the three and was cut directly into the glaciodeltaic sediments of the Valparaiso moraine (Figure 34). The highest terrace can be traced through to the lower section of the river that has already been identified as graded to the glacial Lake Chicago, Calumet lake phase base level (Figure 32).

It has long been known that the St. Joseph River, below South Bend, Indiana owes its flow to the process of river piracy away from the Kankakee River system (Russell and Leverett 1908, Ekblaw and Athy 1925). After the drainage of glacial Lake Dowagiac but before the piracy of the Kankakee River, the area below Niles, Michigan was drained by the ancestral Dowagiac River. This section of the river runs through the Valparaiso moraine, now known to be a fluviodeltaic deposit (Stone et al. 2003) rather than the classical definition of a moraine as a constructional deposit of unsorted glacial till (Flint 1971). Fluviodeltaic deposits in the Valparaiso moraine of Berrien County tend to be shingled (Stone 2002). That is, they were deposited as laterally overlapping deltas in a glacial lake. This process was repeated as the glacier retreated from distal to proximal positions. Former ice-front positions are marked in some cases by a series of kettle chains along an ice-marginal slope (Figure 37). Areas between adjacent deltas are often marked by low drainage ways and kettle chains. An examination of the morphology of glacial deposits in this section of the St. Joseph River shows several potential paths the river could have taken (Figure 37). The present course was likely the lowest path through ice-marginal kettle chains and lateral drainage ways available at the time of capture.

Just north of Niles, the St. Joseph River (near the confluence of Dowagiac River) has carved high bluffs on the west and north sides of the westward turn of the river (Figure 37). These bluffs, are further (though not definitive in and of themselves) evidence of the above described river piracy that occurred as a sudden event rather than a long term transfer of drainage. In particular, the large amplitude, short radius meander bend is unusual for an unconstrained meander in either this or the lower section of the river.

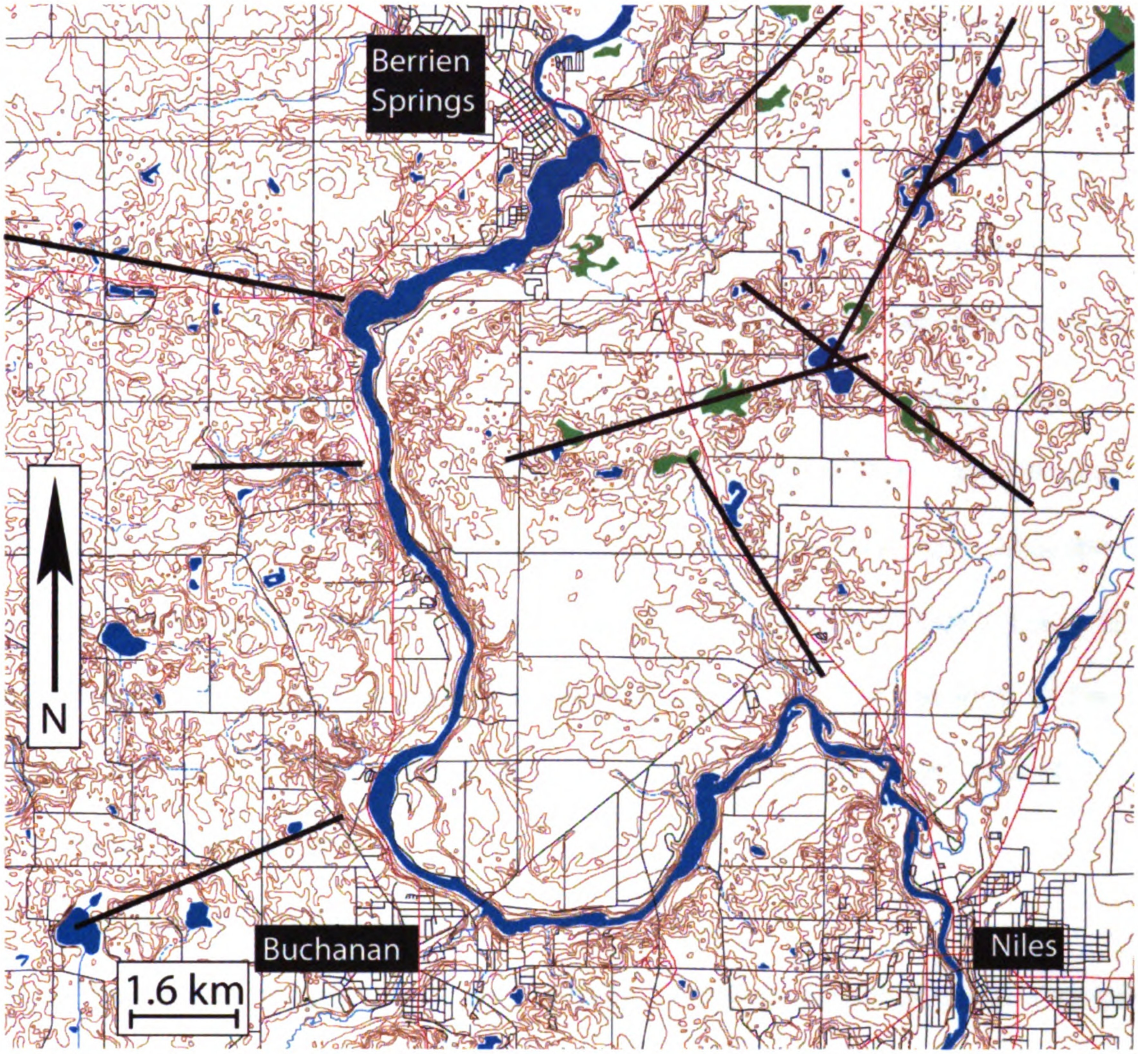

Figure 37. Map of a portion of the St. Joseph River between Niles and Berrien Springs showing kettle chains and lateral drainageways (marked by black lines).

In addition to the three terraces described above, a single fragment of a fourth terrace was discovered during fieldwork west of the St. Joseph River 3.5 km north of Buchanan on the east side of Redbud Trail (Figure 36). This terrace was partially excavated by the landowner for fill elsewhere on his property. Several measurements by the writer on cross-beds (m=170°, n=8) all showed a southbound flow direction. This definitively shows that the St. Joseph River valley held, at an altitude of 219 m (32 m above the present river), southbound flow away from the Lake Michigan basin, toward the Kankakee basin, after the end of glacial Lake Dowagiac. The western edge of this terrace has a scarp 18 m in height carved into fluviodeltaic sediments of glacial Lake Dowagiac of the Valparaiso moraine. This is the only location where this high, south-bound terrace has been observed.

Nickpoint 2 to Nickpoint 3

The second nickpoint is located at Niles, Michigan, 62 km above the mouth of the river (Figure 3) at an altitude of about 195.1 m. At this point, the gradient increases to 0.00047 (2.5 feet/mile). Although the second nickpoint is at the same level as glacial Lake Chicago, Glenwood phase, this is a coincidence. At the time of the Glenwood lake phase, the river valley had not yet been incised to its present altitude. The 195 m contour likely crossed the valley in the vicinity of Berrien Springs where that contour crosses the Calumet terrace. This third river section maintains the low sinuosity of the previous section to the vicinity of Elkhart, Indiana. At this location, the density of significant tributaries begins to increase and the sinuosity increases to about 1.3 without a change in slope.

In the area of the second nickpoint, there is a change in the surficial geology. The river flows through a valley at the westernmost extension of the Inner Kalamazoo moraine in Berrien County, Michigan. This part of the moraine, like the younger Valparaiso moraine, is composed of fluviodeltaic sediments. The river crosses at a gap between the distal end of the westernmost part of the moraine and the next older head of outwash to the east. It is here proposed that this is the path (from Niles, Michigan to South Bend, Indiana) that a small, local creek took to slowly backcut, probably with the aid of ground-water sapping and the melting of ice blocks into kettle depressions, to capture of the upstream section of the St. Joseph River basin from the Kankakee River basin into the Lake Michigan basin.

The location of the capture can be seen geomorphically as an "elbow of capture" (Easterbrook 1999, p. 153) in the city of South Bend, Indiana where the west flowing rivers turns north near the current location of the US-23 bridge over the St. Joseph River (Figure 38). The river formerly flowed west through what is now Dixon West Place Ditch to the Kankakee River. Upstream of the elbow of capture is a large area of the (here named) former glacial Lake Elkhart that probably extended upstream at least to White Pigeon, Michigan. The altitude of the terrace in South Bend, existence of probable wave-cut bluffs south of the main channel between South Bend and Bristol, Indiana, the requirement of southward drainage pathway during blockage of the present drainage pathway, the existence of fine sediments in water-well logs and observed by the writer in terraces above 244 m (800 ft) on the Little Elkhart River upstream of Bristol,

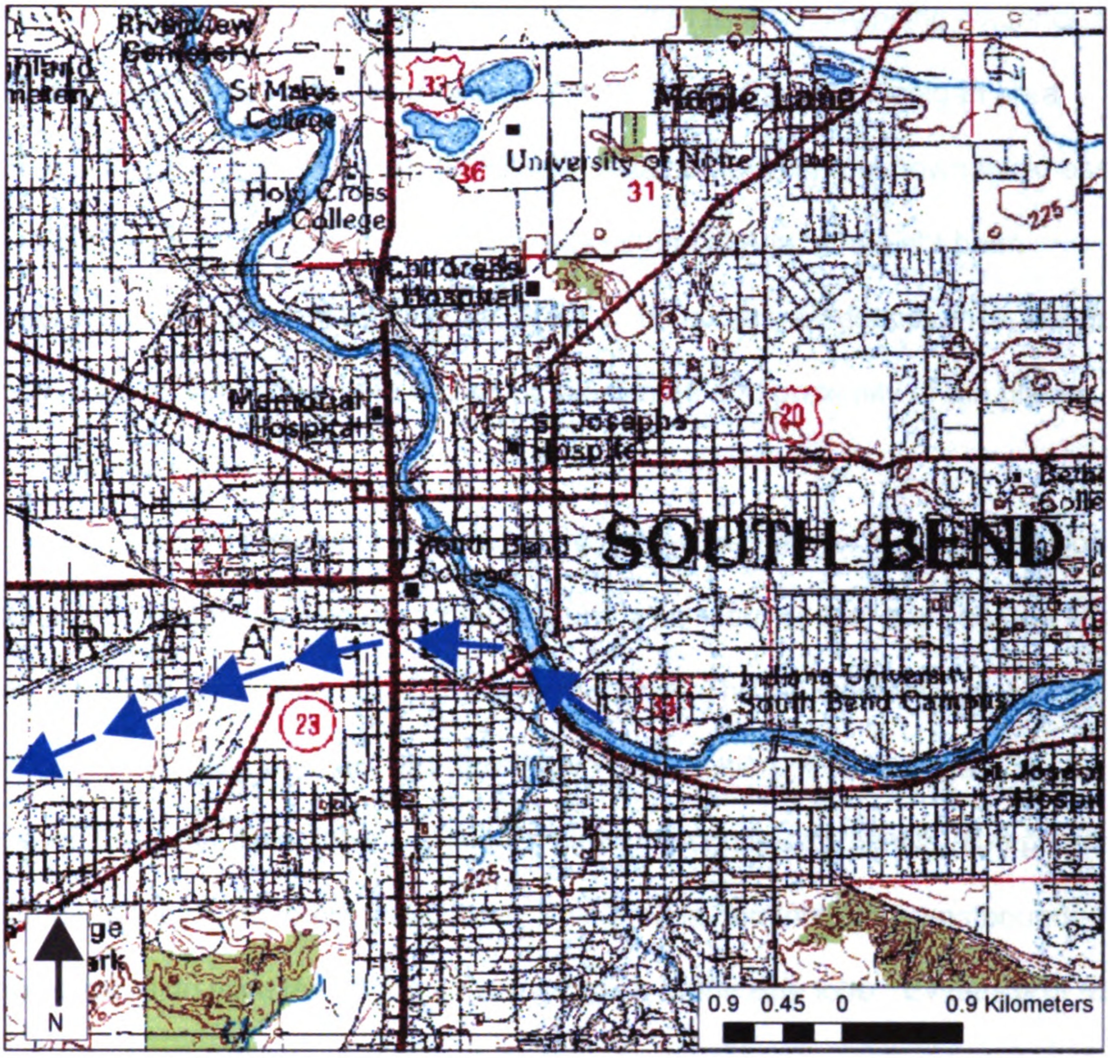

Figure 38. Map showing “elbow of capture” in South Bend, Indiana where the St. Joseph River was captured by a north-flowing river away from its previous westward path to the Kankakee River. Base map is the South Bend 1:100,000 scale topographic map. Contours in meters.

Indiana and deltas with apices around 250 m (820 ft) are convincing evidence for Glacial Lake Elkhart. However, test pits in the White Pigeon, Michigan area indicate that the apex of the White Pigeon delta contains an unknown thickness of topset sand and gravel and the fluvial/lacustrine interface has not been observed in this area. However, the topset beds contain evidence of ice-wedge casts (Figure 39), attesting to the very cold climate and proximity of the glacier during deposition.

This lake was certainly dammed by ice to the west when the Lake Michigan lobe occupied the Maxinkuckee moraine and also requires that the Saginaw lobe had retreated north from the Middlebury moraine. The elevation of this lake is difficult to establish due to a probable later reoccupation of the Wabash/St. Joseph drainage divide by ice of the Erie lobe which altered the sill. The present altitude of the divide is 250 m altitude. This area is dominated by westbound fans originating from the Mississinawa moraine of the Erie lobe. Evidence of the former limits and southern spillway(s) of much of the southern extent of glacial Lake Elkhart is buried beneath these fans.

The northern boundary of glacial Lake Elkhart is marked by several previously unmapped fan/deltas, up to 4.5 km long, originating from southbound flow off the highly collapsed upland of the outer Kalamazoo moraine and the Lake Michigan lobe arm of the Sturgis moraine. Kehew et al. (1999) have described the moraine area as Saginaw lobe terrain overrun by the Lake Michigan lobe based on palimpsest tunnel valleys. Clearly, this upland was still actively draining during glacial Lake Elkhart while the landforms immediately to

Figure 39. Photograph of permafrost wedges developed in loess overlying fluvial top-set beds of the White Pigeon delta, section 7, White Pigeon Township. Shovel to right of center is 0.46 m long.

the south were fully formed. Tunnel valleys south of Mottville, Michigan collapsed after deposition of the fan/delta as seen in the truncated contours shown in Figure 40. Unpublished water-well and geophysical data collected by the US Geological Survey, Water Resources Division (Randall Bayless, personal communication) corroborate a fining from north to south as well as a fining with depth in the St. Joseph River valley between Mottville, Michigan and Bristol, Indiana.

Further upstream between Three Rivers and Constantine, Michigan, a tunnel valley (as discussed by Kehew et al. 1999) crosses the St. Joseph River creating a localized flat gradient. This can be seen in the geomorphology where a delta exists in the present river channel (Figure 41). It is recommended that a core be taken at this delta as it may contain a partial sedimentary history of the river. It is apparent, with increased accuracy beyond the 10 ft contour lines currently available, that this location would present a nickpoint in the river's gradient. However, the gradient upstream returns to the gradient below Constantine, Michigan. Further examination beyond a marker of palimpsest geomorphology and a storehouse of sediment would be of limited utility.

East of the river itself, numerous westbound channels can be observed cutting across a large delta that occupies nearly the entire Constantine 7.5 minute quadrangle (Figure 42). The delta was deposited in glacial Lake Elkhart while the channels that cut the delta were incised by the latest outwash from the last advance of the Erie lobe that also deposited the Topeka fan to the southeast. Pigeon River is incised into the delta while the smaller Fawn River was diverted

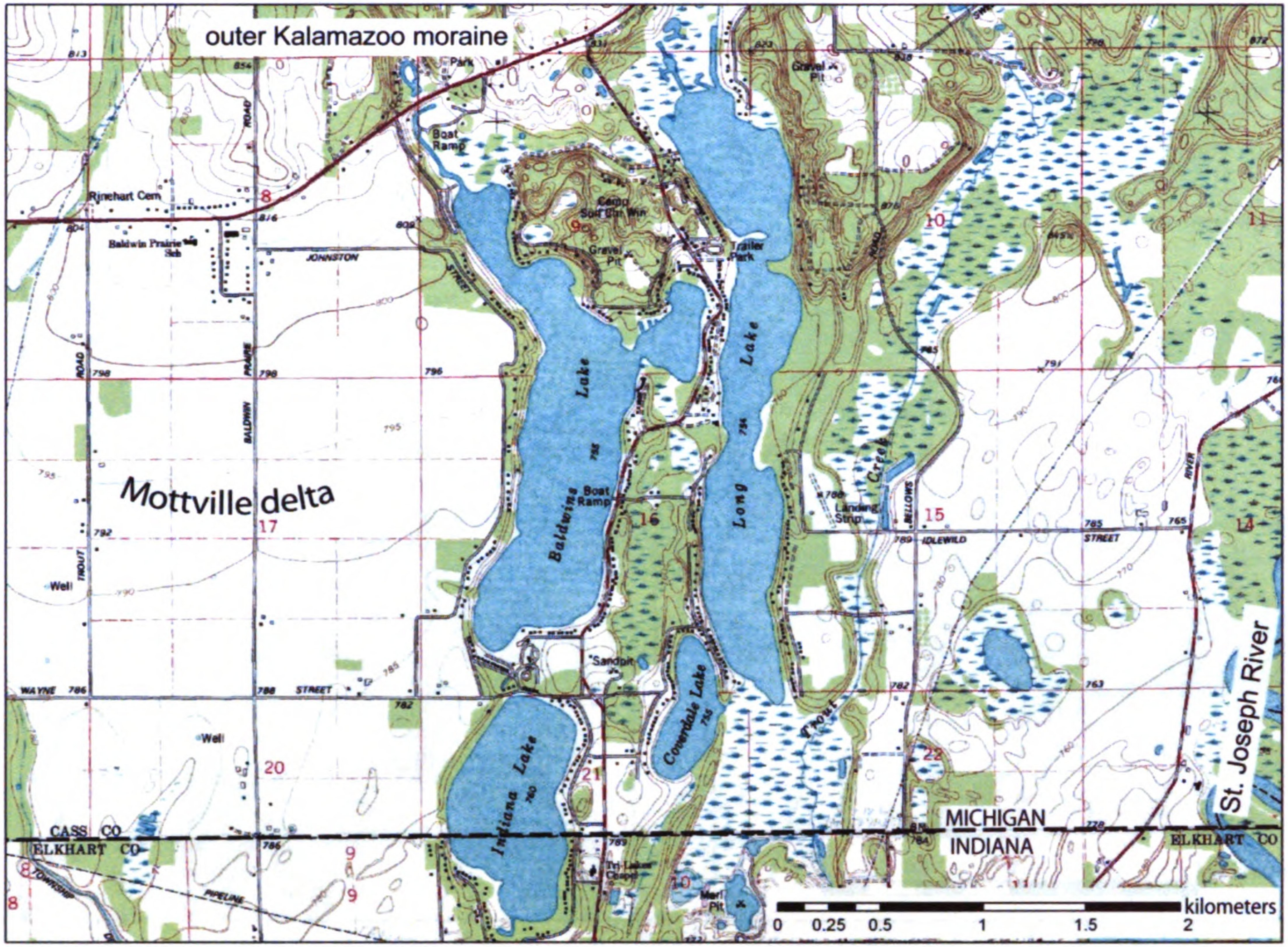

Figure 40. Map showing a collapsed delta of the former glacial Lake Elkhart. Lakes are former ice blocks that melted after the delta formed. Base map is the Mottville 7.5 minute quadrangle. Contours in feet.

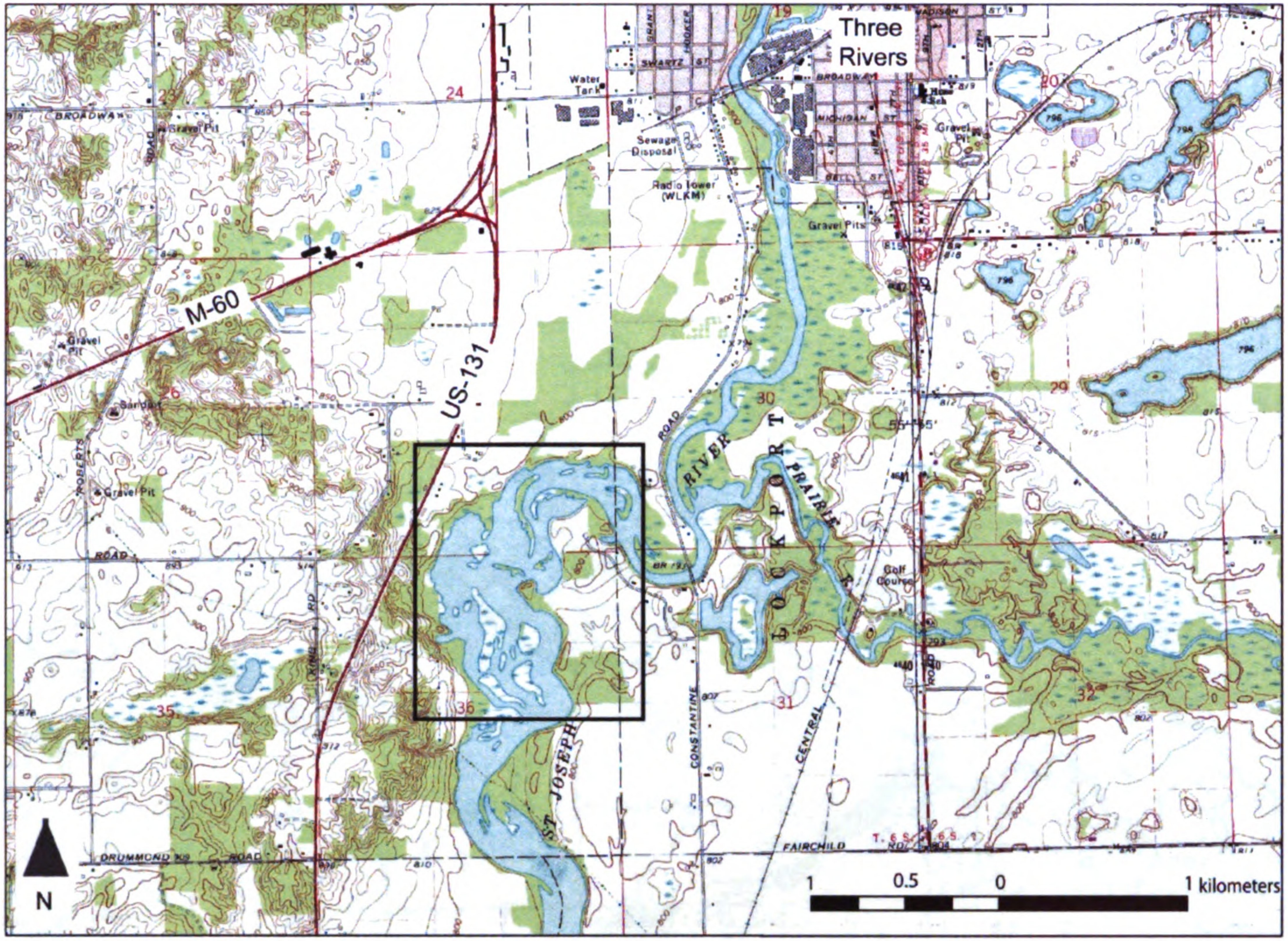

Figure 41. Map showing the St. Joseph River crossing collapse zone (black outline) between Three Rivers and Constantine causing a localized flat gradient. River flows south at this location. Base map from the Three Rivers East and Three Rivers West 7.5 minute quadrangles. Contours are in feet.

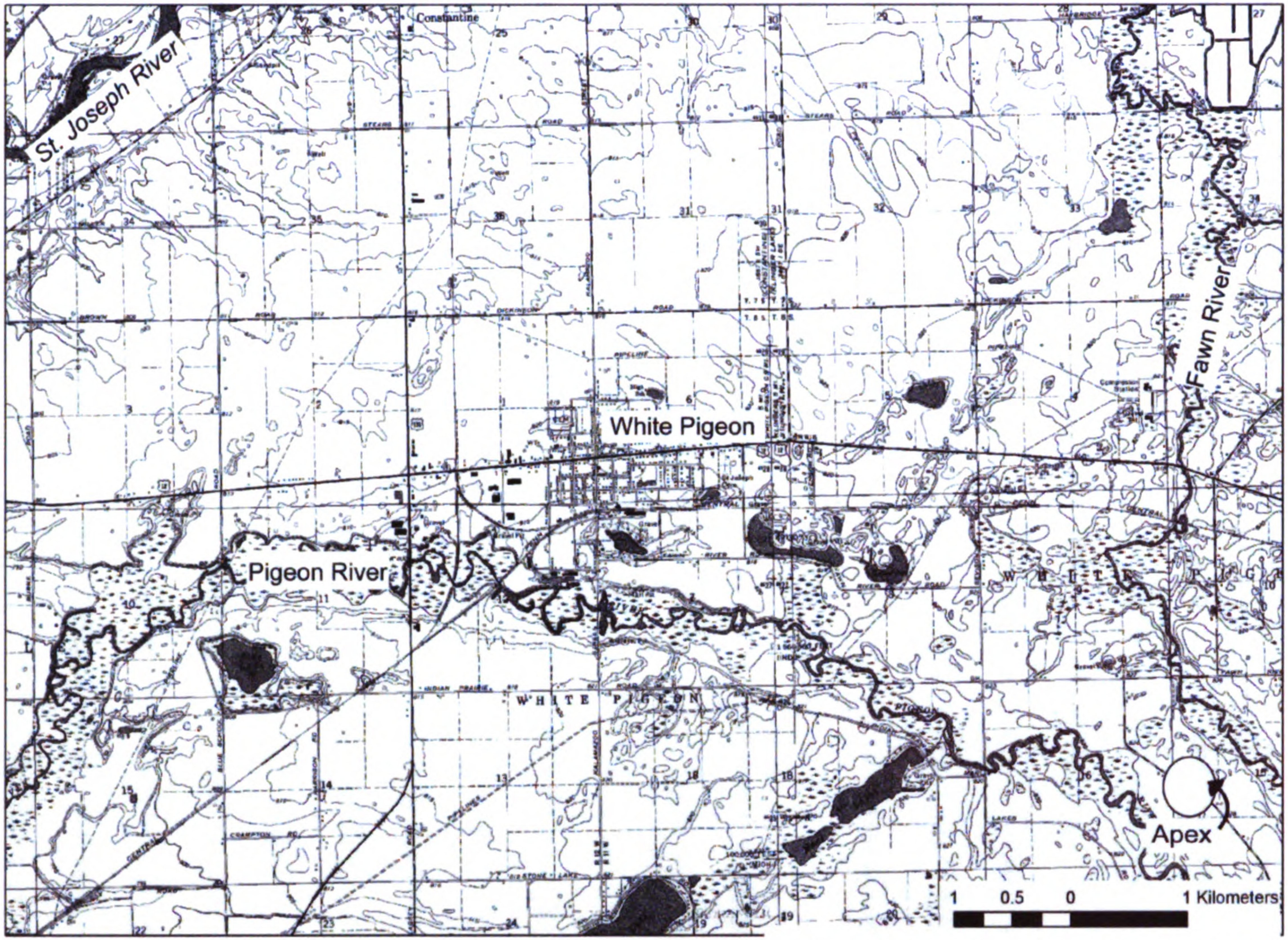

Figure 42. Map showing the White Pigeon delta of glacial Lake Elkhart. Apex is where Fawn River diverges from the Pigeon River. Base map from the Constantine and Sturgis 7.5 minute quadrangles. Contour interval 10 feet.

around the delta to the north. A similar pattern can be seen where the Little Elkhart River emerges from the dissected upland east of Bristol, Indiana (Figure 43). A delta exists between 244-250 m altitude which is dissected on its west side by a braided stream channel at 237 m altitude. These examples again demonstrate that glacial Lake Elkhart was either gone or much reduced at the time of the last outwash event whose terraces cut glacial Lake Elkhart features. It is apparent that the lake had an altitude near 244 m as well as a lower altitude at the above described terrace west of South Bend at 231 m.

North of Three Rivers, Michigan is the junction of southbound outwash deposits (presently drained by the Portage and Rocky Rivers) from Kalamazoo, Michigan and westbound outwash deposits (Figure 44) from Mendon. There is evidence that the westbound deposits have eroded the southbound deposits, thereby establishing their order by a cross-cutting relationship. This is seen best in the former drainageway of Nottawa Creek which, rather than its present channel that turns south to join the St. Joseph River, followed the southern margin of the Tekonsha moraine north of Mendon, Michigan, through the Garman Foster Drain to Portage River. The gradient of Portage River at this location is three times steeper than the section immediately above the junction despite no change in the gradient of the outwash deposit itself. This same drainageway can be followed east from Nottawa Creek through Spencer Creek (Figure 45) all the way to Union City, a total distance of 34 km.

Thirteen kilometers upstream from Mendon is the village of Colon, Michigan. Here the St. Joseph River crosses another tunnel valley through a short section

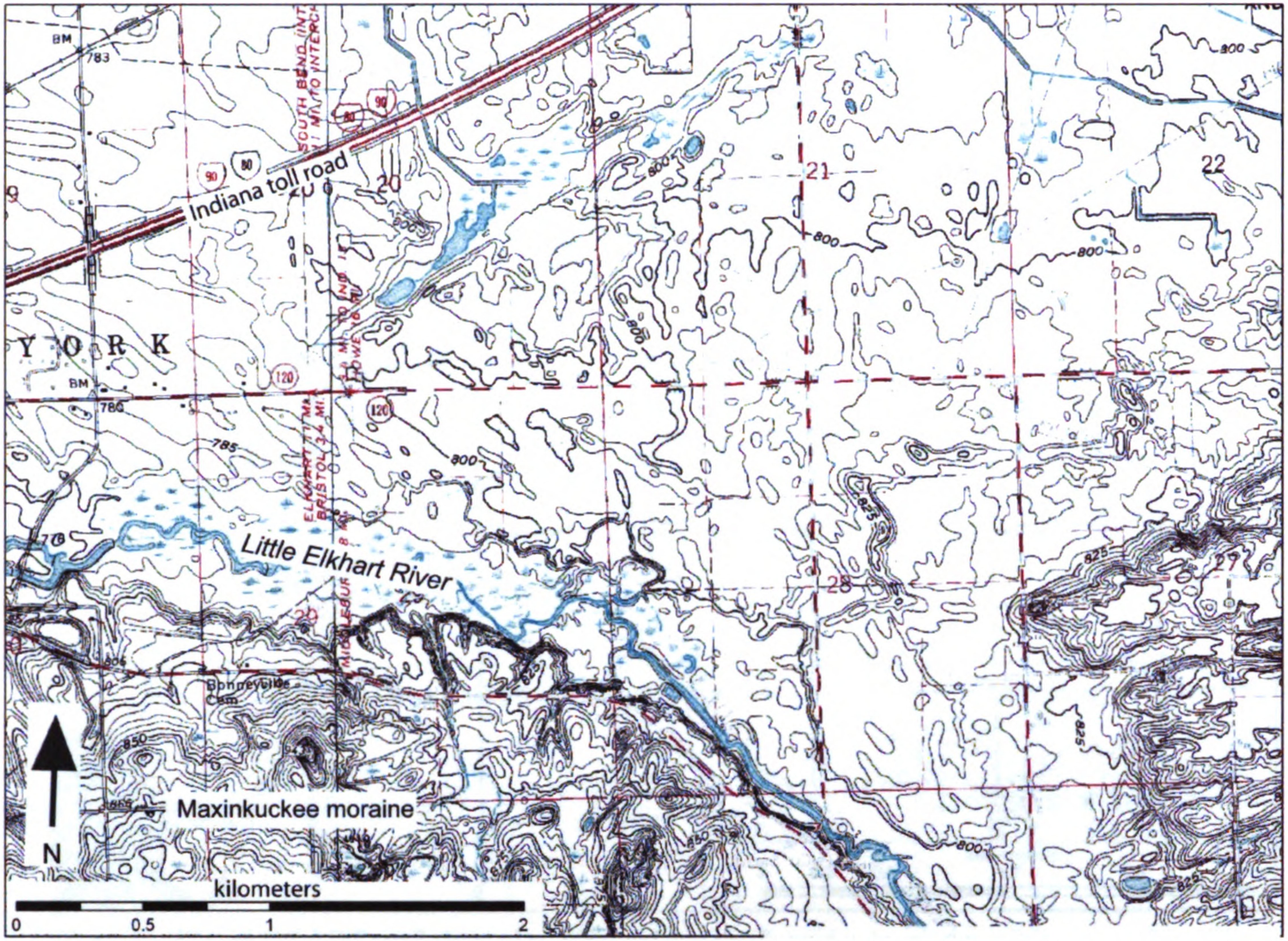

Figure 43. Map showing the Little Elkhart River delta of glacial Lake Elkhart 3 km east of Bristol, Indiana. Base map is the Bristol and Middlebury 7.5 minute quadrangles. Contour interval is 5 feet.

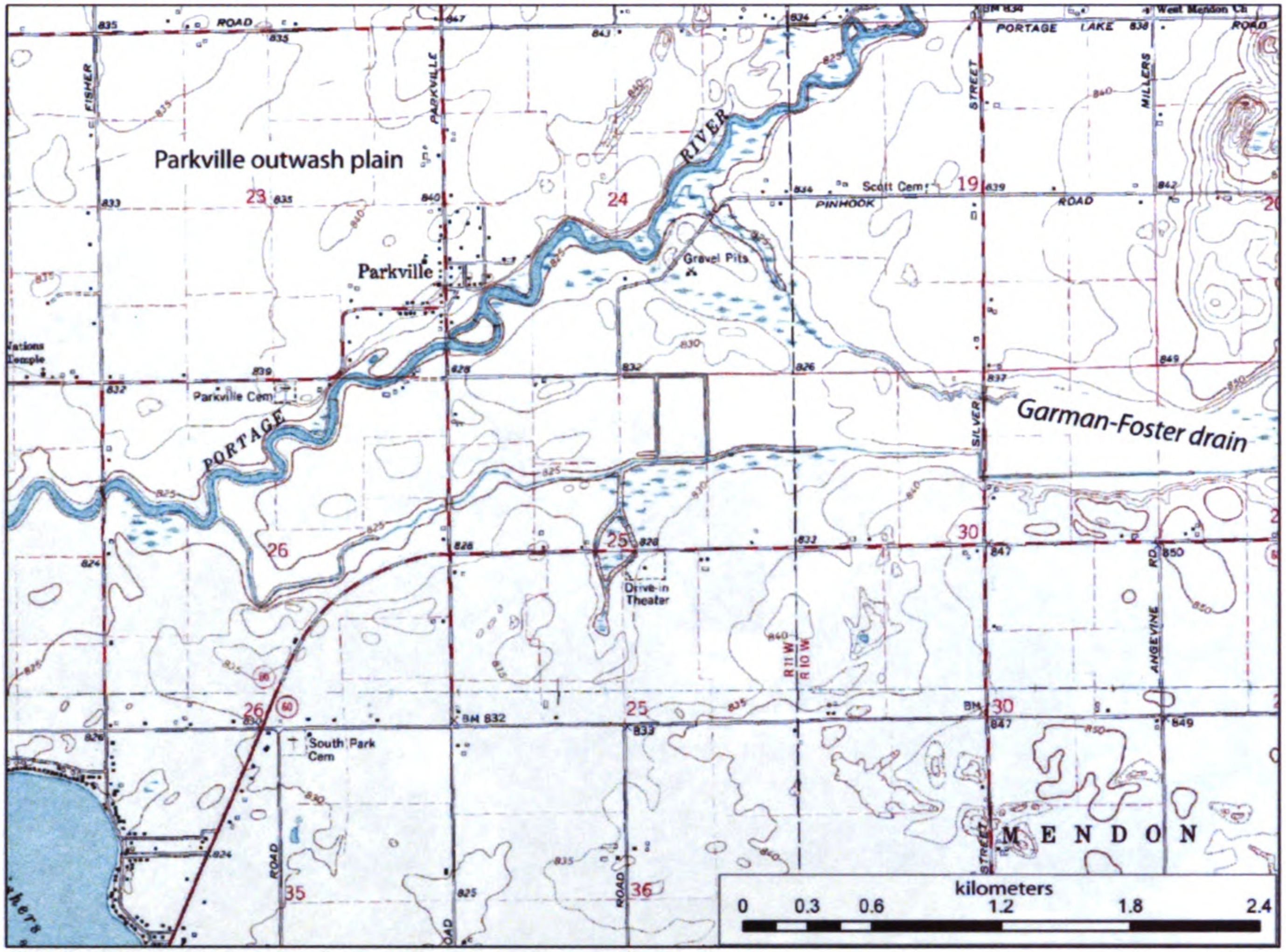

Figure 44. Map showing the confluence of southbound and westbound outwash deposits north of Three Rivers, Michigan. The grade of Garman-Foster drain lies 3 m below the Parkville outwash grade. Base map is the Three Rivers East and Mendon 7.5 minute quadrangles. Contour interval is 5 feet.

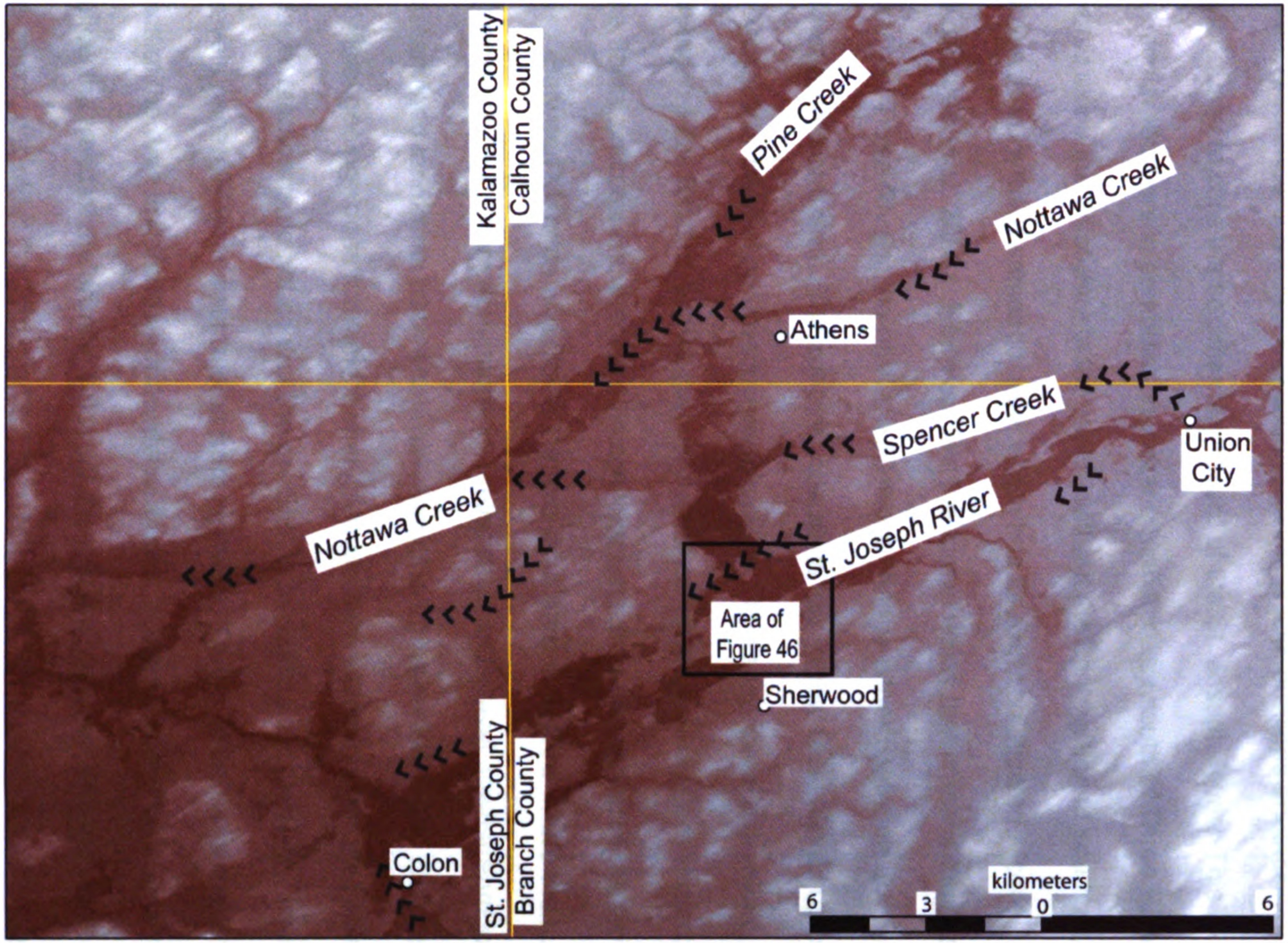

Figure 45. Map of the Spencer Creek-Nottawa Creek paleodrainage. Arrows follow former meltwater channels. Base map is the 30 meter DEM (courtesy US Geological Survey).

of flat gradient that creates Sturgeon Lake. This lake provides another opportunity to core the sedimentary record of the St. Joseph River. This particular location, in combination with the tunnel valley south of Three Rivers, would show the difference in the record without the contribution of southbound outwash from the Kalamazoo area, which has its own complicated record (Kozlowski et al. 2005). However, like the tunnel valley that crosses the river south of Three Rivers, the gradient does not change either upstream or downstream of Colon. The zone of collapse is too narrow to be picked up by the 10 ft contour interval on the topographic maps. The absence of a gradient change at this location shows that the tunnel valley did not have any impact on the fluvial process and therefore its collapse postdates deposition of the terrace.

Between Mendon and Colon the river also enters an area where the surficial geology changes from dominantly low-relief sand and gravel downstream to streamlined landforms (drumlins) with diamicton at the surface. The change in surficial geology does not yield a change in gradient indicating that the underlying glacial landforms do not control the gradient in this section of the river. The river travels generally from west to east from Homer to Three Rivers while the long axis of the drumlins trends northeast-southwest. Given that the river does not follow the trend of the drumlin field and the gradient does not change in this segment, it is clear that the gradient was determined by the event that deposited the outwash and dissected the drumlin field.

Upstream of Colon is also a partially collapsed terrace composed of sand and gravel of unknown thickness that is about 4.5 m above the modern floodplain

(Figure 46). At the nickpoint in Union City the terrace is at an altitude of about 277 m and ends at about 262 m just west of Colon. The main terrace joins a terrace associated with Nottawa Creek (Figure 45). South of the St. Joseph River in southern Mendon and Leonidas Townships the above terraces grade into a flat expanse covering over 33 km^2 at an altitude of about 260 m (Figure 47). In addition, several upland remnants, generally associated with the Sturgis moraine (Leverett and Taylor 1915) all have slope breaks along their margins between 259 and 263 m altitude (Figure 48). These mark shorelines of (here named) glacial Lake Nottawa. It is not likely that this lake lasted a long time, as it did not leave as much evidence as presently exists for glacial Lake Elkhart. However, if as currently thought, this was an outwash plain (Kehew et al. 1998, Farrand 1982) rather than a glacial lake, the slope breaks would be the result of fluvial erosion. If this were the case they would show a downstream decrease in height rather than the uniform elevation observed. It is recommended that additional information, particularly sedimentological, be collected in this area.

Nickpoint 3 to Nickpoint 4

The third nickpoint evident from the profile is near Union City, Michigan, 226 km above the mouth of the St. Joseph River at an altitude of about 270 meters (885 feet). The gradient increases abruptly to 0.0028 (15 ft/mile). The river does not cross into a different glacial terrain at this location. Union City is still within an area of streamlined landforms interpreted as drumlins (Brown et al. 1999, Fisher and Taylor 2002) with sand and gravel occupying the low areas between drumlin clusters. However, the river does cross the Tekonsha moraine

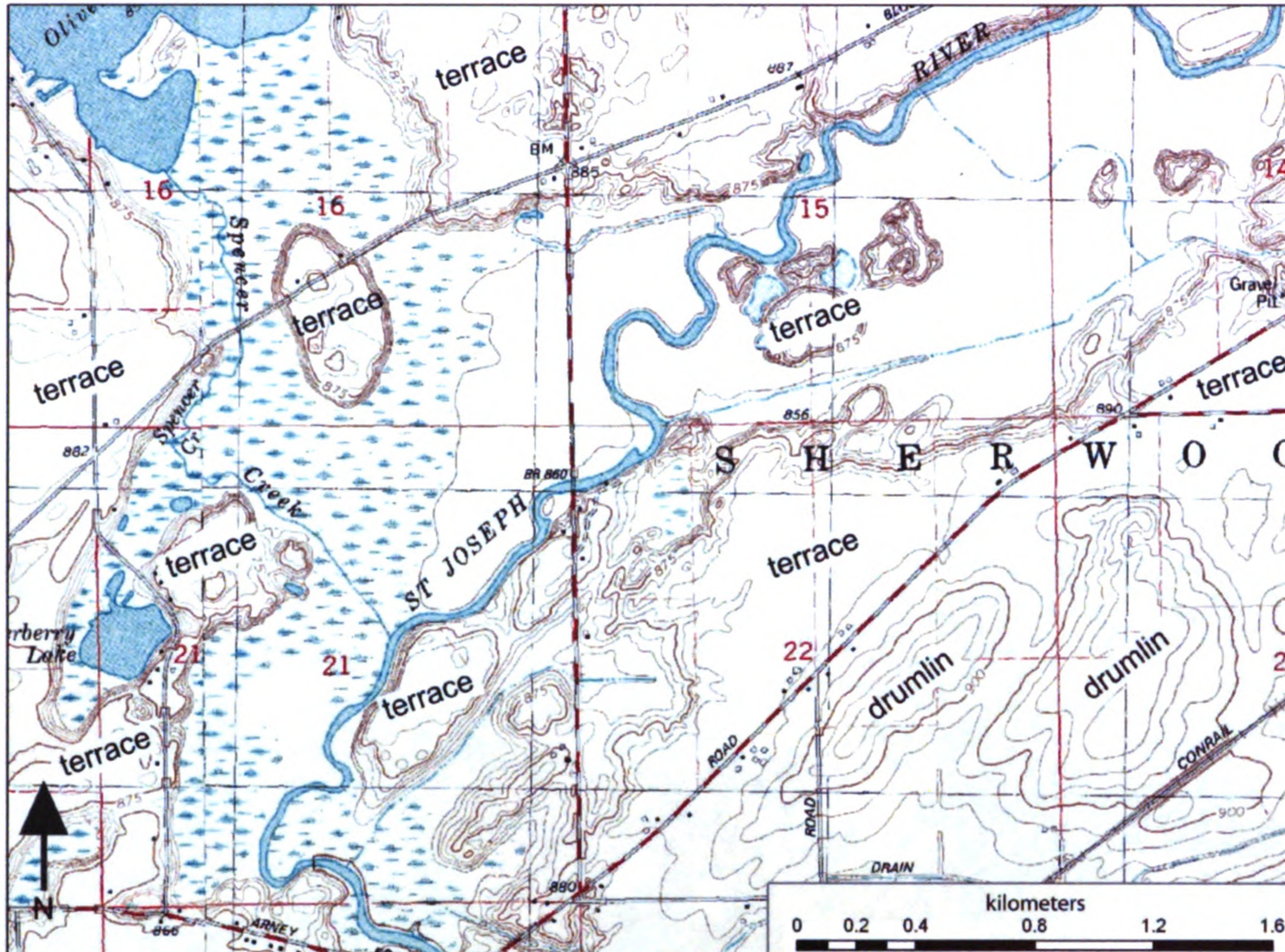

Figure 46. Map of partially collapsed terrace between Colon and Union City, Michigan. Tops of intact terrace fragments are on grade with uncollapsed portion. Base map is the Colon and Union City 7.5 minute quadrangles. **Contour interval is 5 feet.**

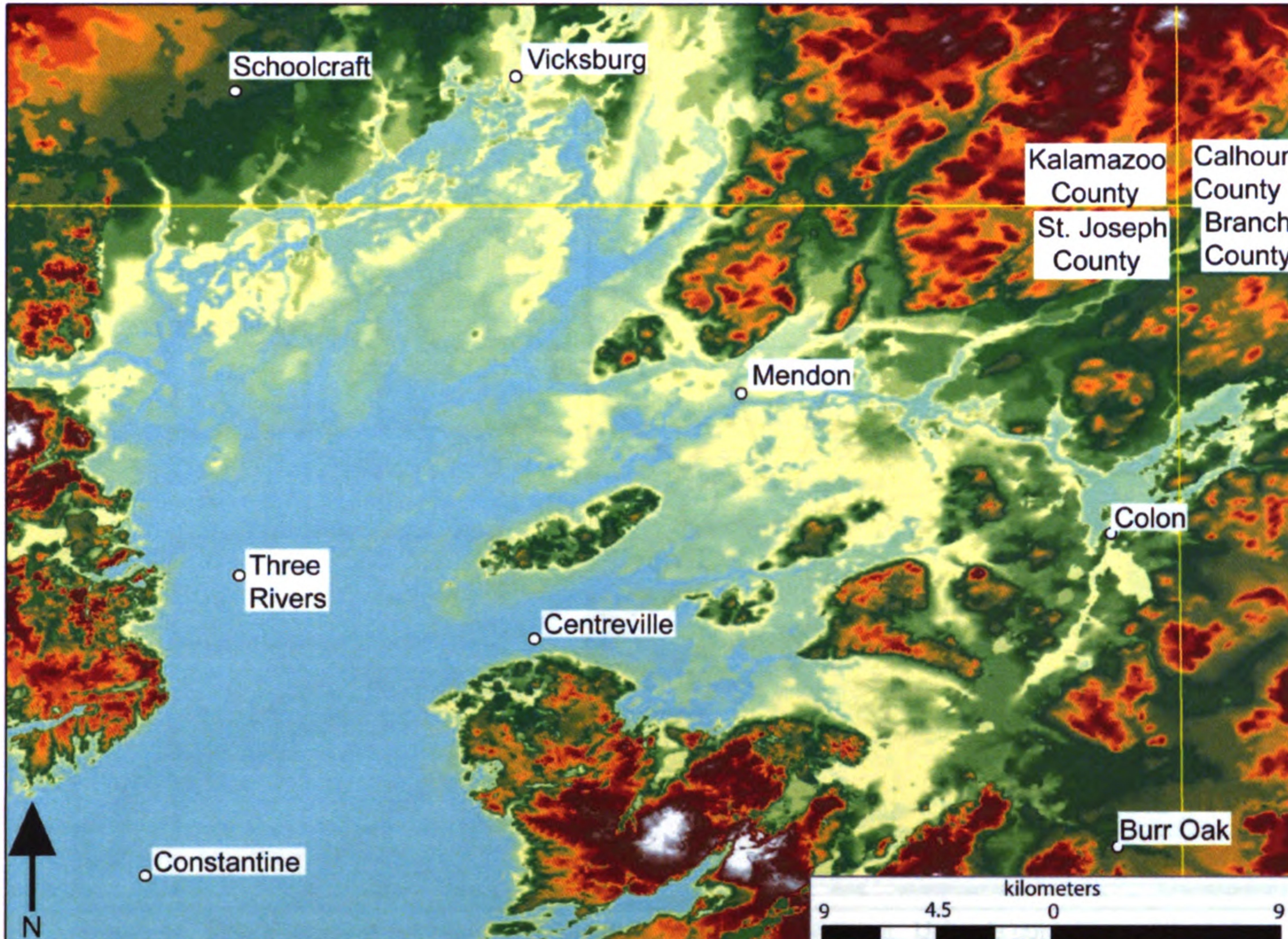

Figure 47. Map of glacial Lake Nottawa, St. Joseph County, Michigan. Base map is 30 meter DEM (courtesy U.S. Geological Survey) flooded to an elevation of 260 meters. Light blue color is water depth > 2 meters.

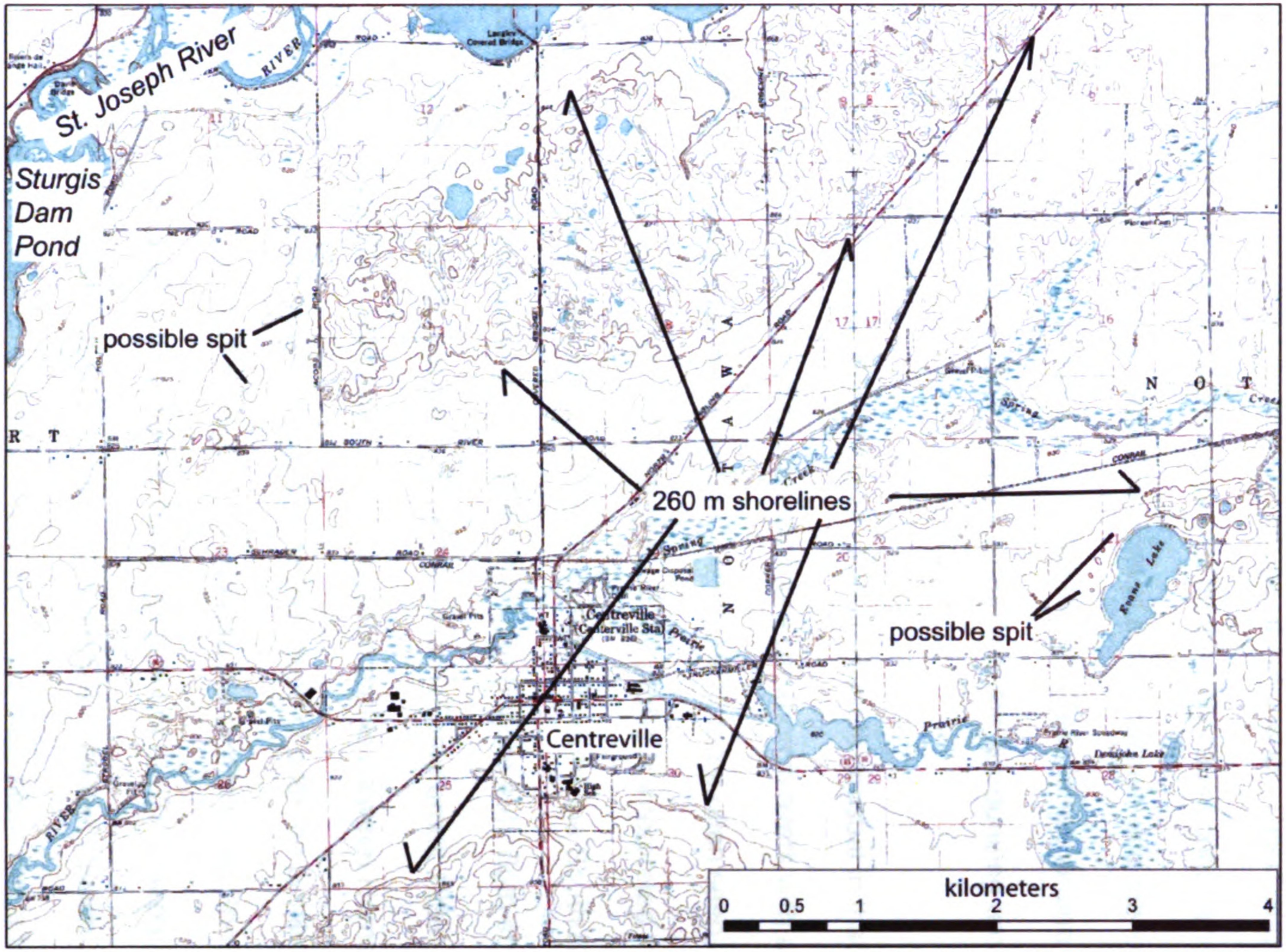

Figure 48. Map showing shorelines along islands in glacial Lake Nottawa. Uplands contain consistant slope breaks at 260 m (850 ft) with flat intervening areas. Slope breaks contain gravel concentrations suggesting wave action. Base map is the Three Rivers East 7.5 minute quadrangle. Contour interval is 10 ft.

10 km upstream of Union City near the village of Tekonsha. This change may have an effect on the gradient, but it cannot be the only cause considering the lack of spatial agreement between the location of the moraine and the location of the nickpoint and recent questions as to whether the Tekonsha moraine even represents a true landform sufficiently different from the surrounding area (Fisher and Taylor 2002).

The profile for this section of the St. Joseph River is unusual in that it is convex upward (Figure 3), a characteristic more common in rivers with an increasing load/discharge relationship (Carlston 1969) more often seen in arid regions (Schumm et al. 1987). This feature of the profile is interpreted to be a result of piracy of the headwaters of the St. Joseph River by the South Branch Kalamazoo River at the village of Homer, Michigan (Figure 49). Both streams have the geomorphic feature of an elbow of capture (Easterbrook 1999) and the Kalamazoo River does not have a convex-upward profile between Homer and Albion, ruling out the underlying geology as the cause of both the profile shape and the right-angle turns of the rivers. In addition, the profile of the South Branch Kalamazoo River upstream of Homer fits the gradient of the St. Joseph River downstream of Homer without a nickpoint (Figure 50). The last section of the St. Joseph River from Homer to Hillsdale that is now the main stem was a tributary before capture and joins the river at the last nickpoint. Originally a short tributary, the South Branch Kalamazoo River backcut south from Albion, Michigan until it intercepted the main stem of the St. Joseph River and captured the latter's flow upstream of Homer. This event reduced the flow of the St. Joseph River

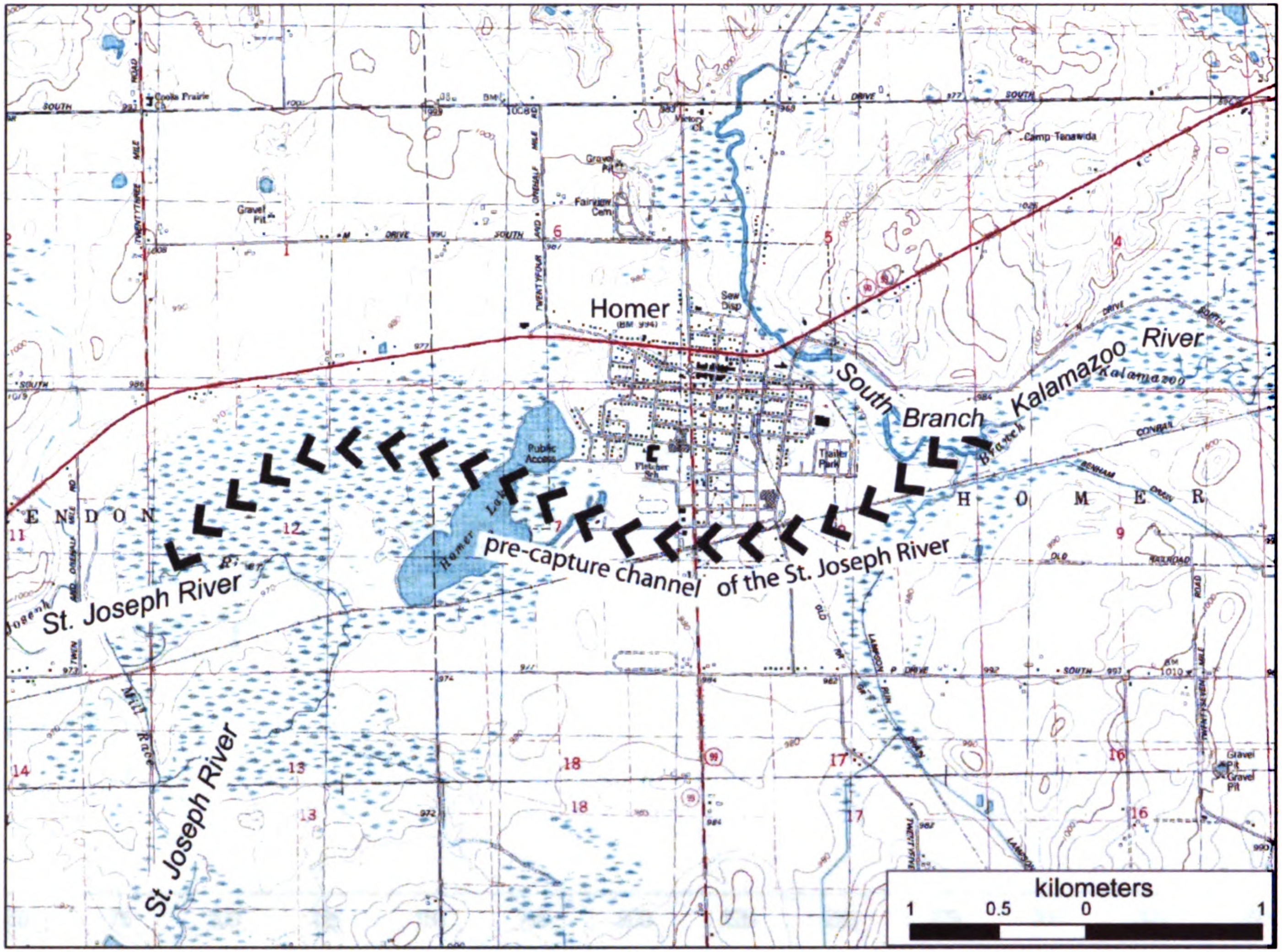

Figure 49. Map showing the point of capture of the St. Joseph River by the South Branch Kalamazoo River at Homer, Michigan. Arrow shows former channel. Base map is Southwest Albion 7.5 minute quadrangle. **Contour interval is 10 ft.**

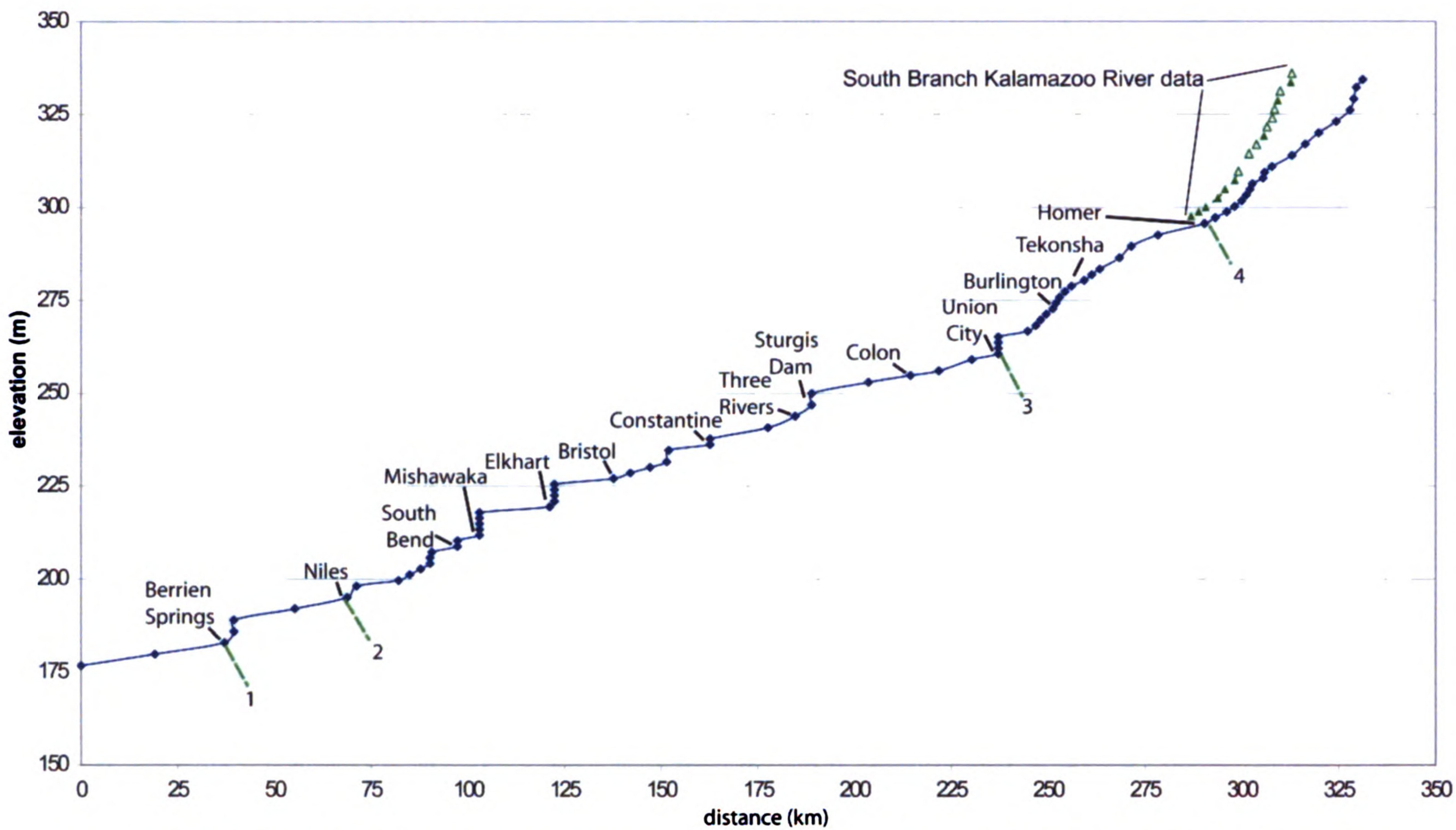

Figure 50. Longitudinal profile of the St. Joseph River with the captured section of the South Branch Kalamazoo River appended. Green triangles represent data upstream from Homer on the South Branch Kalamazoo River.

immediately west of Homer by about half (based on the relatively equal drainage basins for each river upstream of Homer), thereby causing the increase in the load/discharge ratio. This ratio disparity was returned to conditions typically associated with humid rivers (i.e. concave upward profile) in Union City where a significant tributary (Coldwater River) joined the St. Joseph River.

The tributary Coldwater River joins the St. Joseph River from the south in Union City (Figure 51). The terrace described in the previous section continues up both the St. Joseph River and the Coldwater River. Both rivers are incised about 9 m below the 1.6 km wide terrace. The terrace can be traced upstream along the St. Joseph River from an altitude of 277 m in Union City to about 288 m at Homer. However, it becomes more difficult to trace because the modern floodplain is no longer incised at Tekonsha (17 km upstream of Homer) and cross-sections of the terrace stratigraphy can no longer be easily observed in the valley walls. The terrace appears to persist at least to Litchfield, Michigan at the distal edge of the Mississinawa moraine of the Huron-Erie glacial lobe.

The terrace on Coldwater River can be traced upstream to Hodunk, Michigan where it splits and also continues up Hog Creek. Digital orthophoto quadrangles (DOQ) show evidence of the longitudinal bars of braided streams on the terrace surface (Figure 52). Both terraces continue through to the drainage divide at the Mississinawa moraine in southwestern Hillsdale County, Michigan. The headwaters of Swan Creek, which joins the St. Joseph River at Colon, is 7 km west of Coldwater. The terrace appears to split at this location as well, part being traceable northwest up Coldwater River and part southwest along Swan

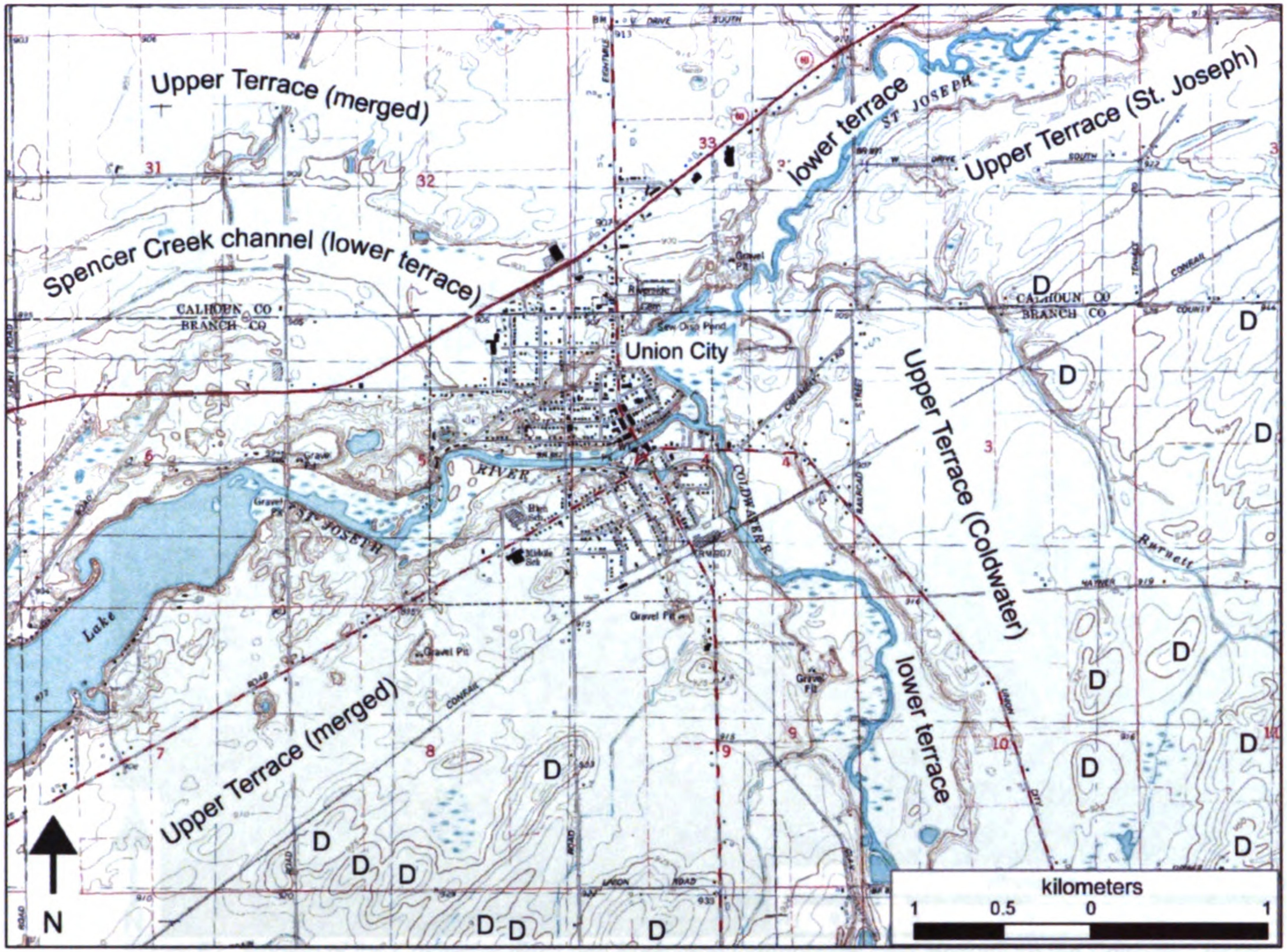

Figure 51. Map showing terraces of the Coldwater St. Joseph Rivers at Union City, Michigan. Base map is **the Union City 7.5 minute quadrangle. Contour interval is 5 ft. Drumlins are labelled "D."**

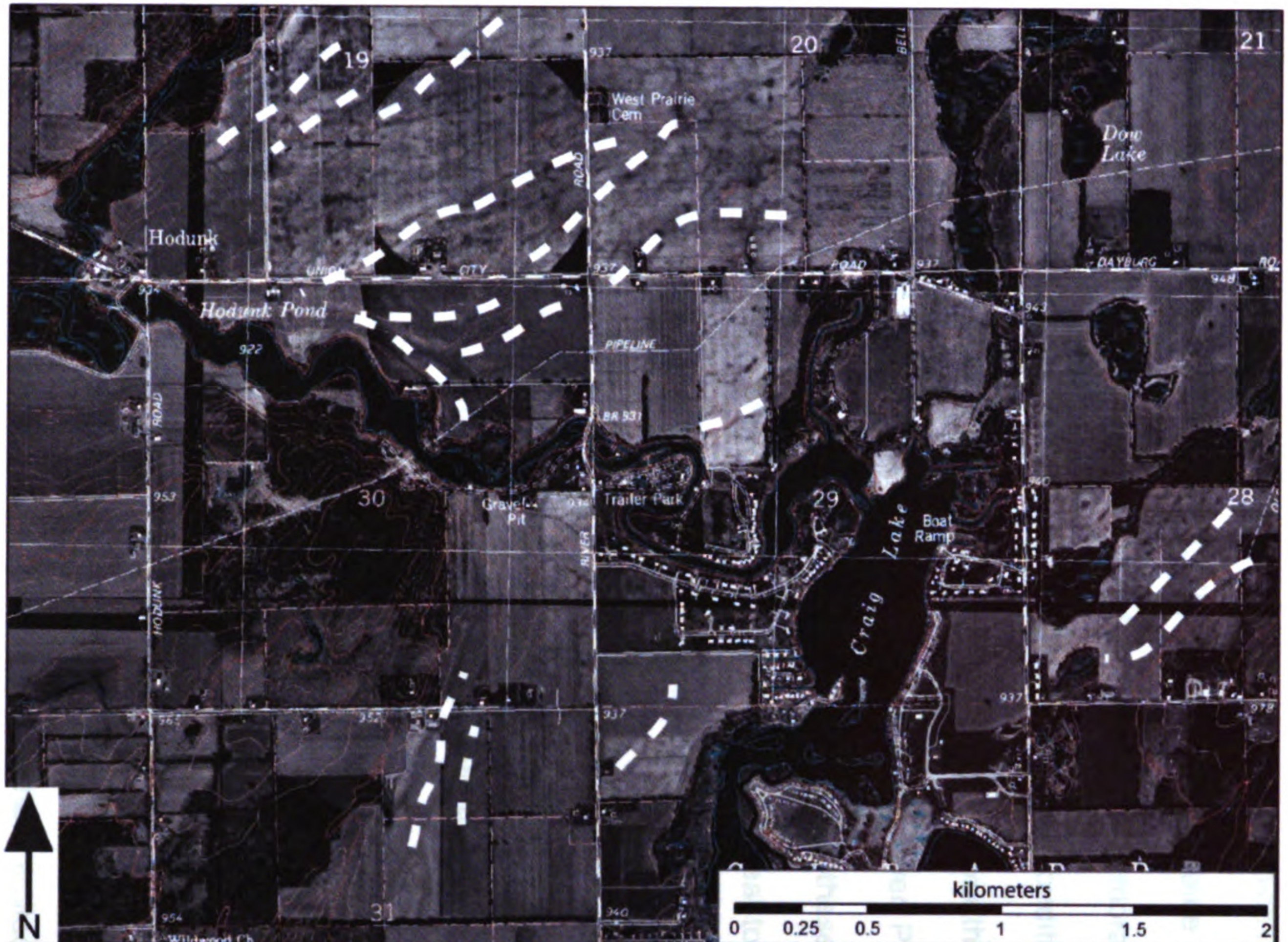

Figure 52. Portion of the Coldwater West and Burlington Orthophotos near Hodunk on the Coldwater River showing braided-stream morphology on the terrace surface. Several channels are noted with a dashed line. **Contour interval is 5 ft.**

Creek. To the south of the Swan Creek drainage is Prairie River, which joins the St. Joseph River 2 km south of Three Rivers. The headwaters of Prairie River appear to originate in a fan on the distal margin of the Mississinawa moraine on the south side of Coldwater Lake, 13 km south of Coldwater. This condition is repeated at the headwaters of other rivers tributary to St. Joseph River that rise along the distal margin of Mississinawa moraine, including Fawn River, Pigeon River, Little Elkhart River and Elkhart River. The event that created these fan and terrace deposits seems to have waned to the north. There appear to be no fans north of Coldwater Lake. This has interesting implications for the mechanics of this advance of the Huron-Erie lobe. It could add weight to the idea that the Huron-sourced part of the glacier was sliding past existing deposits of the Huron-Erie/Saginaw interlobate zone, generating little meltwater until reaching a more open area south of Coldwater Lake, as generally described by Brown et al. (2006). This is proposed as a fruitful area for future study.

Nickpoint 4 to Headwaters at Baw Beese Lake

Glacial deposits in this section of the St. Joseph River are relatively thin, from 30 m to absent in a few places. Hillsdale is the southern limit of a cuesta formed by the resistant Marshall Formation. This cuesta may have limited eastward movement of the Huron-Erie lobe as well as westward movement of the Saginaw lobe. The location of the interlobate moraine along this cuesta is likely no coincidence.

Evidence for the reason for a nickpoint at Homer was discussed in the section above. This last section of the river has a concave-upward profile, considered more normal for the humid eastern part of North America (Thornbury 1969). This section does not have a single gradient but rises smoothly upstream of Homer to a maximum of 1.05 m/km before the river reaches its headwaters at Baw Beese Lake just southeast of Hillsdale, Michigan. Baw Beese Lake appears to be the most distal in a northwest-trending series of lakes being called "tunnel valleys" or "tunnel channels" in the literature (e.g. Clayton et al. 1999, Kehew et al. 1999). While it is not necessary to describe their origin for the present discussion, these features have the morphology of linear chains of kettle lakes within a narrow belt of collapsed-glacial deposits (Figure 53). Some tunnel valleys also contain eskers implying at least some time of formation beneath the ice (Clayton et al. 1999). The orientation of the tunnel channel is said to be generally perpendicular to the margin of the glacier with which it is associated. The proximal end of the Baw Beese Lake tunnel valley lies outside the St. Joseph River basin in the St. Joseph River of the Maumee River basin (yes, it's true, they actually gave adjoining river basins the same name). The northwest orientation of this tunnel channel indicates a likely origin from the Huron-Erie glacial lobe as far east as the Wabash moraine. However, unlike the headwaters of the tributaries described in the previous section, the headwaters of the St. Joseph River appear to extend beyond the distal margin of the Mississinawa moraine to its proximal side as does its original main stem, the South Branch Kalamazoo River as well as the only other major river in southeast Michigan that flows through to Lake Michigan, the Grand River. The headwaters of all three of these rivers are only 20 km apart.

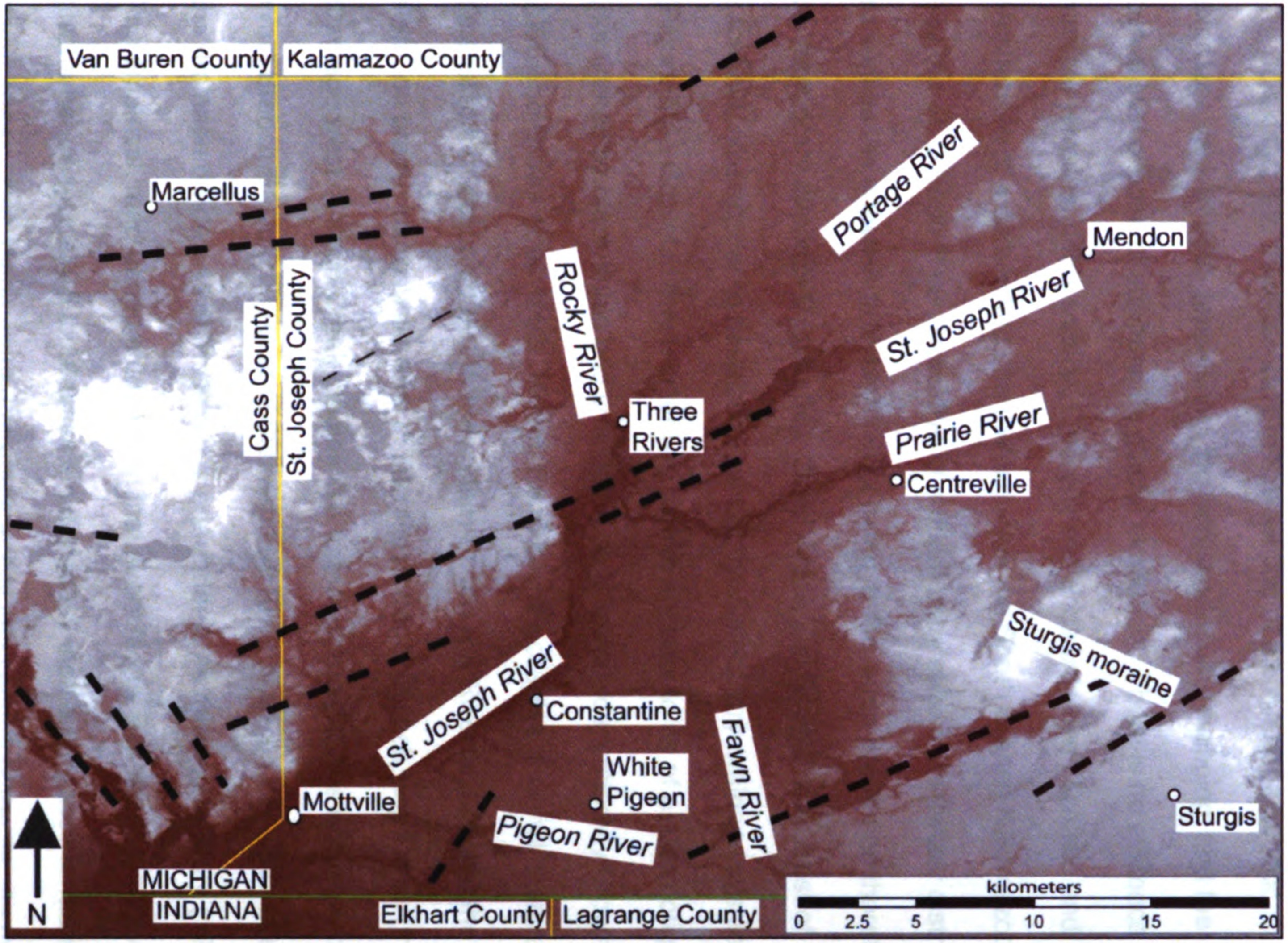

Figure 53. Map showing morphology of tunnel valleys in the vicinity Three Rivers, Michigan. Base map is the 30 meter DEM (courtesy U.S. Geological Survey). Dotted lines show linear chains of collapsed topography.

It is difficult to trace terraces in this section of the river due to poor exposure, lack of post-depositional incision, inadequate mapping, and the pervasive collapse of surrounding glacial landforms. There is no evidence of multiple terraces in this section. There is evidence between Litchfield and Mosherville of a connecting channel between the South Branch Kalamazoo River and the St. Joseph River, indicative of a distributary style of meltwater distribution as well as the outwash from both rivers being at the same grade and therefore contemporary deposits (Figure 54). Sand Creek and Soup Creek, west of Litchfield, also appear to have been part of the distributary-drainage system (Figure 55). Gravel in these latter two creeks appears to be of eastern origin, (dominated by black shale and limestone) as generally described by Brown et al. 2006 and Brown et al. 1999. In fact, Sand Creek occupies a tunnel channel parallel to that occupied by the main stem St. Joseph River and rejoins the main stem, on grade, just downstream of Litchfield. South Branch Hog Creek (the next drainage west of Soup Creek, begins in a tunnel channel that is occupied by the West Branch St. Joseph River of the Maumee River basin on the other side of the drainage divide. All of these rivers are underfit, as outlined by Dury (1965). The empirical, map-based study found that rivers fed by Pleistocene-meltwater discharge are typically 5 to 10 times smaller today than during their formation (Dury 1964). The above named streams in this last section of the St. Joseph River have a ratio of former channel width/current channel width ranging from 5-1 to 50-1. The lower range of ratios are seen at rare bedrock constrictions observed by the writer 1 km upstream of Jonesville, Michigan and constrictions caused by diamict-covered drumlins as seen in South Branch Hog Creek at Section 28, Butler Township, Branch County.

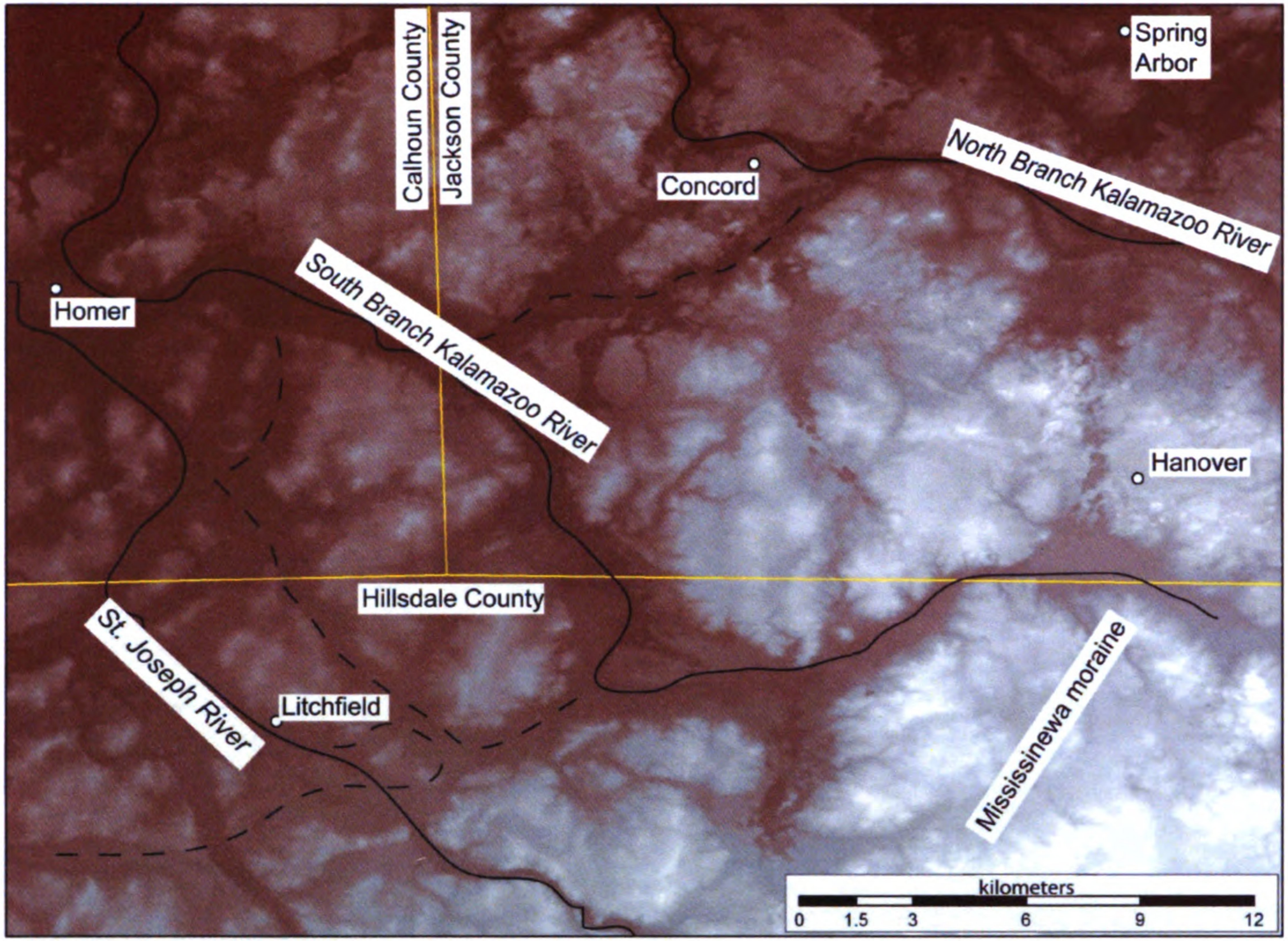

Figure 54. Map showing connecting channels between the North Branch Kalamazoo River, South Branch Kalamazoo River, and St. Joseph River, indicating a distributary style of meltwater distribution. Base map is the 30 meter DEM (courtesy of the U.S. Geological Survey).

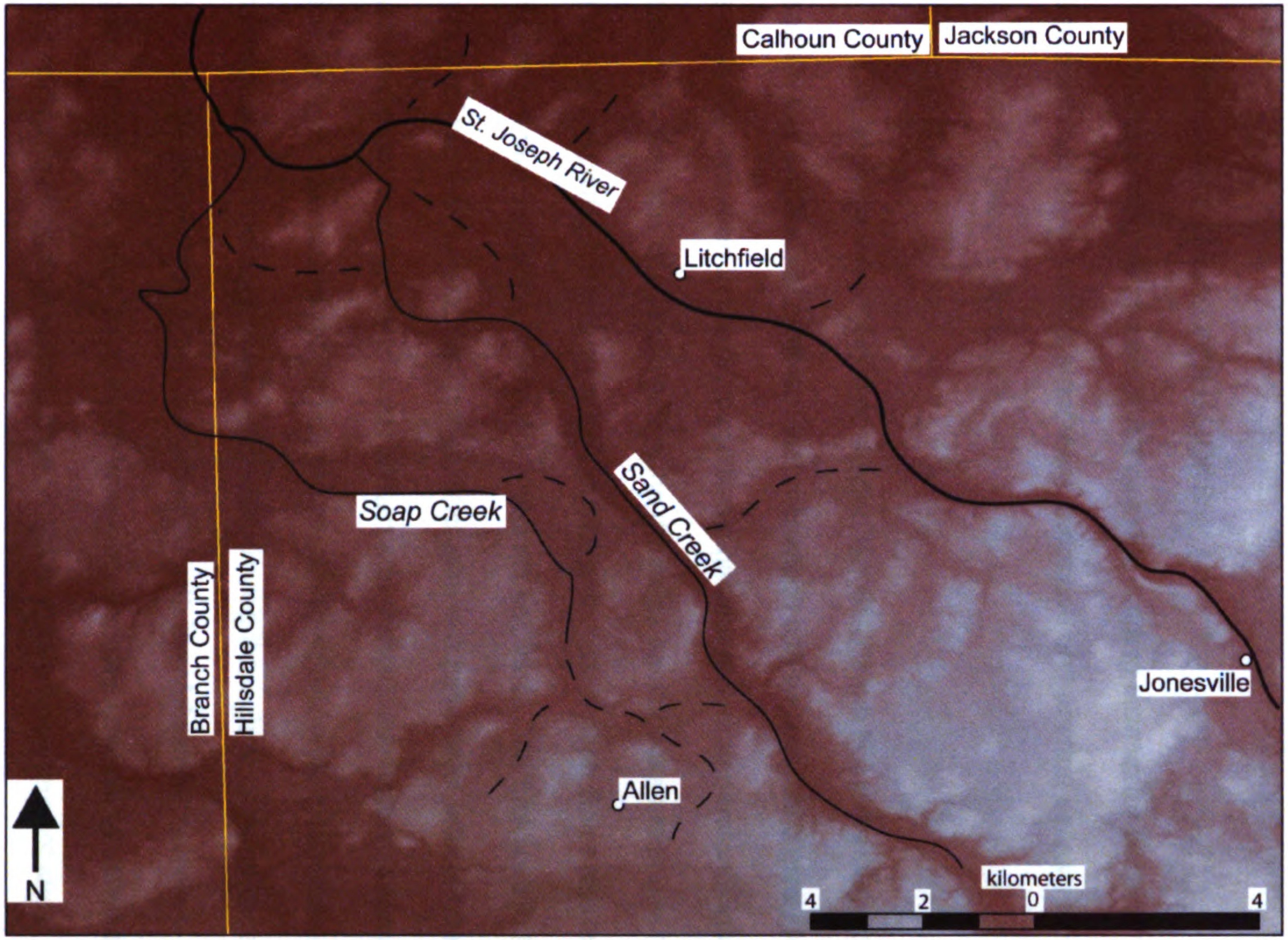

Figure 55. Map showing the Sand and Soup Creek terraces west of Litchfield, Michigan. These tributaries are highly underfit. They also contain dry channels (dashed lines) that once connected all three in an anastomozing system. Base map is the 30 meter DEM (courtesy US Geological Survey).

Chapter 8

SUMMARY AND CONCLUSIONS

Nothing under heaven is softer or more yielding than water but when it attacks things hard and resistant there is not one of them that can prevail.

Tao Te Ching, Chapter 78,

240 B.C.

Evidence for the events that shaped the St. Joseph River drainage basin is seen in the longitudinal profile of the river which can be used as a proxy for events that occurred in the drainage basin given a proper understanding of how disequilibrium flow conditions within the drainage basin causes changes the slope of the profile. The longitudinal profile of the St. Joseph River has four nickpoints that divide the profile into 5 distinct sections. The first section from the mouth of the river to Berrien Springs, Michigan is a base-level affected section with a distinct morphology of a flat bottom with steep valley walls caused by erosion and subsequent refilling during the Chippewa-Nipissing regression and transgression. The Chippewa valley was eroded through an older terrace that overlies glaciodeltaic sediments associated with the Valparaiso moraine. This geomorphology is expressed in the longitudinal profile as a low gradient section that ends in a distinct nickpoint. The profile of the terrace does not show any evidence of the drop in water level from the Glenwood to Calumet phases of glacial Lake Chicago, which led to the additional hypothesis that the stream capture of the St. Joseph River basin drainage into the Lake Michigan basin occurred after the Glenwood phase of glacial Lake Chicago gave way to the

Calumet phase. The discovery of the clast-supported gravel deposit underlying as the basal unit of the terrace sediments further supports the hypothesis.

The second section is also short and was carved by flow from stream-capture and contains the same Calumet phase terrace as the previous section as well as the basal unit of clast-supported gravel. This section also contains two lower terraces that were eroded away below the nickpoint. It has a slightly steeper gradient and is clearly graded to a higher base level. The nickpoint below did not proceed farther upstream of Berrien Springs, Michigan and left this section of the river graded to a previous base level. The low Chippewa base level rose to the high Nipissing base level and the river no longer had the grade to finish equilibrating this section and the evidence is seen in the longitudinal profile.

The third section of longitudinal profile is the longest section. This central section of the drainage basin was the first part to be deglaciated by the retreat of the Saginaw lobe. However, the Lake Michigan lobe and the Huron Erie lobe still occupied the western and eastern parts of the drainage basin, respectively. This fact has gone unreported in the literature. Two previously unknown proglacial lakes (here named) at three separate levels formed in this section, descending in level as the ice margins that formed the dams retreated out of the St. Joseph River drainage basin, opening successively lower outlets. The longitudinal profile reflects the gradient of several deltas that fill the valley and the channels that subsequently cut through them to the current drainage divide of the Kankakee River.

The fourth section of the St. Joseph River shows an intriguing convex upward profile typically seen in arid regions. This feature of the profile is interpreted to be a result of piracy of the headwaters of the St. Joseph River by the South Branch Kalamazoo River at the village of Homer, Michigan. This event led to an increasing load/discharge relationship due to loss of normal flow for the load being delivered to the river. The longitudinal profile not only shows this phenomenon but, when extended to the former main stem, demonstrates that the pirating river has a profile more at grade with its former downstream section than does the present main stem of the St. Joseph River.

The final section is, as stated above, that of a former tributary that became the main stem due to the aforementioned stream capture. It has a rising gradient that is typical of a headwaters section, but meets the next lower section at an abrupt nickpoint that is normally seen in tributary rivers. A headwater section at grade would approach the lower section with a slowly declining slope.

This study has demonstrated that longitudinal profiles contain information useful in the examination of rivers and in the search for patterns in morphology, deposition, and erosion. The shape and form of longitudinal profiles is determined to a large degree by the events that shaped the river basin. While not an end unto themselves, the proper study of a river basin should include the evaluation of longitudinal profiles. These studies are predicated on the existence of high quality elevation data (either topographic maps or DEMs) as well as knowledge of the events that shaped the deposits of the drainage basin.

The record of the Wisconsin glaciation in the eastern part of the mid-continent is well represented by glacial and nonglacial sediments, buried soils and organic deposits. It shows that glacial ice initially advanced southward and southwestward over southern Ontario and extended into the western part of the Ontario basin, the eastern part of the Erie basin, and probably into the northern part of the Huron and Superior basins by about 70,000 years ago. It also shows the ice margin remained and oscillated there for several tens of thousands of years until about 31,200 years ago, after which it advanced over the rest of the Great Lake basins as far south as Illinois, Indiana, Ohio and Pennsylvania. After reaching its southern limit near the Ohio River about 23,500 years ago, the ice margin retreated northward. The retreat however was interrupted by a number of major readvances. Not until just after about 11,500 years ago did the ice margin finally retreat north of the Great Lakes.

The oscillation of the margins of the individual glacial lobes was uneven and punctuated by a series of rapid retreats and readvances that created a complicated series of glacial lakes. At times these lakes were separate and at other times they were confluent or connected by river channels. Each lake-level change was a change in base level for the rivers draining into the Great Lakes. There has not been adequate time for the St. Joseph River to re-equilibrate to base-level changes, leaving its longitudinal profile with a major nickpoint 27 km upstream from the mouth in the vicinity of Berrien Springs, Michigan. The Berrien Springs nickpoint was the only nickpoint evident in the profile caused by base-level variation, despite all the base-level changes documented for the basin. The base level changes were, however, responsible for abandonment of two terraces downstream of the first nickpoint and three terraces above the first nickpoint. Nickpoints were also caused by river piracy, flow-regime changes due to tributary input, local base-level adjustments, and most importantly the stepwise addition of pieces of the drainage basin by asymmetric advance and retreat of the individual glacial lobes.

The water level of the Lake Michigan basin is currently 177 m above sea level. Around 11,200 calendar years B.P., the lake level in the Lake Michigan basin had dropped to its lowest level in prehistory, about 70 m above sea level. This low level (Lake Chippewa) had profound effects on the rivers flowing directly into the basin. Extreme low lake level rejuvenated the river, causing massive incision of up to 43 m in a valley no more than 1.6 km wide. The incision is seen 27 km upstream of the present shoreline.

As lake level rose from the Chippewa low, the St. Joseph River lost competence and its estuary migrated back upstream. Floodplain and channel sediments partially refilled the recently excavated valley leaving a distinctly non-classical morphology of steep sides with a broad, flat bottom. During this study the valley walls of the lower St. Joseph River were measured at 12-18 m tall while borings revealed up to 30 m of infill sediment below the modern floodplain. Sediment removed from the St. Joseph River valley during the Chippewa phase lowstand is a massive volume of sediment, some of which likely resides in a lowstand delta approximately 30 km off-shore in Lake Michigan.

The active floodplain below Niles, Michigan, is inset into an upper terrace and delta graded to the Calumet lake phase of glacial Lake Chicago. In the lower portion of the terrace stratigraphy a 1.5-2.0 m thick section of clast-supported gravel marks the entry of the main St. Joseph River drainage above South Bend, Indiana into the Lake Michigan basin. This gravel layer represents the consolidation of drainage that probably occurred during final melting out of ice-marginal kettle chains allowing stream piracy to proceed between Niles and South Bend.

Winters et al. (1986) stated that the valleys across moraines in west Michigan originated as a result of draping of the glacial debris over preexisting valleys in the bedrock. Accepting this, it would not be genetically correct to say that the St. Joseph River "cut" across or into the moraines, it simply took advantage of an existing low as soon as it became ice free. This theory seems plausible in that the river already had a southwest outlet to the Kankakee River at South Bend, Indiana when the ice blocked the present valley across the Valparaiso moraine (Figure 8). However, the Valparaiso moraine in this location is composed of fluvial, lacustrine, and deltaic deposits (Stone 2002) that filled, not draped, preexisting topography. Rather, in the case of the St. Joseph River, it seems clear from the surficial geology that the river took advantage of kettle chains, low areas behind "heads of outwash," and low areas between the shingled deltas to cross the Valparaiso moraine to enter the Lake Michigan drainage basin. If this was the case, the connection to the Lake Michigan basin was not instantaneous with removal of the ice barrier, but would have been delayed at least as long as it took to melt the ice that was in the kettles, a process that can take centuries (Johnson 1992) and for normal stream processes to backcut through the kettled areas to pirate a stream on the other side of the drainage divide.

River basins that were developed within glaciated areas, particularly those with base level inside the glacial maximum, have not been properly evaluated in the context of modern geomorphology. These basins were constructed sequentially as portions of the basin are added during glacial retreat. They may drain toward multiple basins during development and certainly have drained toward multiple base levels. The context of headward erosion from a main stem toward a hinterland does not apply. While it has been long recognized that glaciated basins have undergone vast changes in flow regime from the Pleistocene to the Holocene (e.g. Dury 1964), inadequate knowledge of events in the history of the drainage basins has led to an incomplete understanding of the deposits in the basins.

Two previously unknown glacial lakes, glacial Lake Elkhart and glacial Lake Nottawa, are proposed on the basis of this study. Glacial Lake Elkhart has been noted by the existence of three deltas, the easternmost centering around White Pigeon, all of whose apices are near 244 m altitude. This lake likely discharged to one of a series of terraces west of South Bend. The earlier glacial Lake Nottawa existed east of Three Rivers at an altitude of around 260 m. The evidence for this, though incomplete, is from the anomalous termination of two, well-formed terraces (originating from the St. Joseph River and Nottawa Creek) into a broad, flat area of diamicton covered by a thin layer of fine sand that covers a large part of Nottawa Township, St. Joseph County. In addition, several uplands within the flat area have sharp slope breaks between 250 and 260 m. These slope breaks are interpreted to be wave-cut bluffs. Glacial Lake Nottawa likely drained south to the Wabash River drainage basin.

The distal margin of the Mississinawa moraine is incised by many channels that are all graded to the same level. This is so despite the fact that many of them are presently occupied by underfit rivers that are not capable of regrading their channels. As such, the gradient of their longitudinal profiles is inherited from a former flow regime. These channels have the form of an anastomozed system that owes its origin to a single event, the advance of the Huron-Erie lobe to the Mississinawa moraine. These channels merge with upper terraces of Nottawa Creek and the St. Joseph River downstream of Union City as well as the terrace of the Coldwater River that also merges with the St. Joseph River upper terrace at Union City. Therefore, cross-cutting relationships dictate that occupation of the Mississinawa moraine occurred before lowering the water level of the glacial Lake Nottawa to glacial Lake Elkhart which then requires that the Lake Michigan lobe occupies the Kankakee River basin.

Several recent papers (Sjogren et al. 2002, Fisher and Taylor 2002, Kehew et al. 1999, Brennand and Shaw 1994) have attributed the stratigraphy and geomorphology of the drumlinized area between nickpoint 3 and the headwaters of the St. Joseph River to a single advance of the Saginaw glacial lobe. In particular, the features seen as a "linear depression or series of linked depressions" (Sjogren et al. 2002, p. 41) are called tunnel channels and shown as evidence of subglacial flooding. The stratigraphy of the area is said to consist of bedrock or subglacial till at the bottom, sand and gravel in the middle, and a carapace of diamict at the top. This stratigraphy occurs when glacial advance erodes the bedrock and leaves a layer of subglacial till at the bottom of the pile. Subsequently, subglacial flooding, which is pressurized due to the confining influence of the overlying ice, erodes channels into the subglacial till and/or bedrock. Pressurized flow tends to be turbulent and creates vortices. The "linked depression" aspect of the tunnel channels is a result of pothole erosion with intra-depression areas, or hummocks, being pre-existing topography left as non-eroded or incompletely eroded remnants by upward-deflected turbulent flow. If the process proceeds long enough, even these "hummocks" between potholes will eventually be smoothed out, leaving a flat-bottom tunnel channel. Finally, the confining glacial-ice stagnates and melts away, leaving a carapace of diamict that owes its origin to a supraglacial melt-out process.

Sjogren et al. (2002, Figure 12) show a topographic map and ground-penetrating radar (GPR) transect of a tunnel channel and hummock in the St. Joseph River between Colon and Union City in Section 16 of Sherwood Township (Figure 56). This area was previously addressed in this report and is also shown in Figure 43. This hummock is not a remnant of pre-existing or sub-glacial deposit eroded by turbulent sub-glacial discharge. It is, in fact, a remnant of the terrace of the St. Joseph River. Its flat top is on grade with the elevation of the portions of the terrace upstream and downstream of the collapse zone

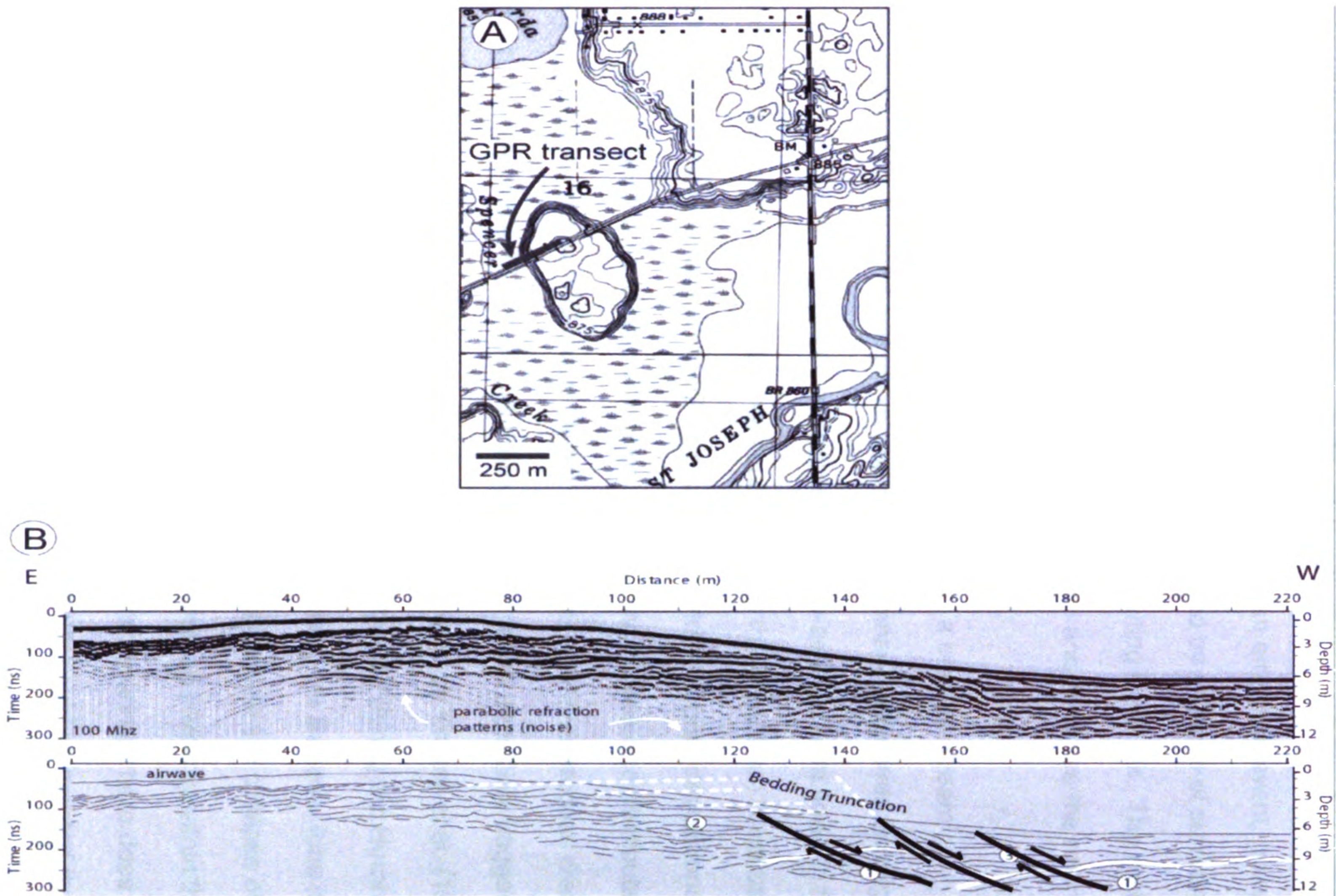

Figure 56. Graphics from Sjogren et al. (2002, Figure 12) showing "steeply dipping reflector," here reinterpreted as lisric faulting (heavy black lines and arrows) commonly seen in areas of ice-block meltout collapse.

associated with the tunnel channel. The "steeply dipping reflector" (Sjogren et al. 2002, p. 47) in the GPR transect, was ascribed to an erosional event. However, further analysis reveals that it is a listric fault caused by collapse of surrounding terrace sediments down into the void left by the melting ice block. The rest of the terrace is still there under the surface of the swamp and therefore the remnant is not a sub-glacial erosional landform.

Sjogren et al. (2002) were correct in ascribing a braided-stream environment to the sand and gravel seen in the remnant. However, its origin was not from sub-glacial deposition, but rather from sub-aerial stream flow. In fact, further observation of tunnel-channel hummocks in the St. Joseph River basin show many more terrace remnants that also resulted from post-depositional collapse, rather than sub-glacial erosion (e.g. Nottawa Creek and its tributary Pine Creek, as well as Coldwater River). Therefore, whatever process created the tunnel channels, they were still there, filled with ice (a possibility suggested for an area of the outer Kalamazoo moraine to the west by Kehew et al. (1999)), when the terrace of the St. Joseph River was constructed during the Huron-Erie lobe's stand at the Mississinawa moraine. At that time, the Saginaw lobe itself had retreated north of the St. Joseph River basin; otherwise the collapse of the terrace would have been regionally pervasive rather than primarily within the tunnel channels. It should be noted that the analysis presented here does not deny the potential that the origin of linked depressions could be created by the tunnel channel process as described by Sjogren et al. (2002). However, the examples of landforms in the St. Joseph River basin used in their analysis cannot be used to illustrate their hypothesis.

There is still the issue of the "thin, sandy diamicton....blanketing the area" that has been interpreted as a result of "melt-out of stagnant ice" (Sjogren et al. 2002, p. 49). This deposit is here interpreted to be a cryoturbated-loess deposit. The loess owes its origin to the dewatering and subsequent deflation of both glacial Lake Nottawa immediately south of the Sjogren et al. (2002) study area as well as glacial Lake Elkhart. Similar deposits were previously observed (Kincare 2000, Stone et al. 2003, Stone 2002) and attributed to the dewatering of glacial Lake Baroda and subsequent deflation of its exposed lake-bottom sediments in Berrien County, Michigan. Where the loess cover was less than 1 m in thickness, cryoturbation processes caused the underlying sand and gravel to be frost-heaved up into the loess, creating a bimodal grain-size distribution that looks very much like a supraglacial melt-out deposit. Permafrost and cryoturbation features have been observed by the writer in the White Pigeon area (Figure 39) and patterned ground has been observed in Branch County from aerial photography (Figure 57).

On the basis of the lake levels interpreted for glacial Lake Nottawa and glacial Lake Elkhart, three ice-marginal positions are proposed in Figures 58-60. Figure 58 shows the proposed positions of the Lake Michigan and Huron-Erie lobes. There are no sills to the south capable of holding up a lake at an elevation of 260 m. Therefore, the outlet to southbound overflow of glacial Lake Nottawa was over the ice at the merged boundary between the Lake Michigan and

Figure 57. Aerial photograph showing patterned ground characteristic of permafrost in Batavia Township, Branch County, Michigan. Downloaded from Google Earth™ on 12/1/2008.

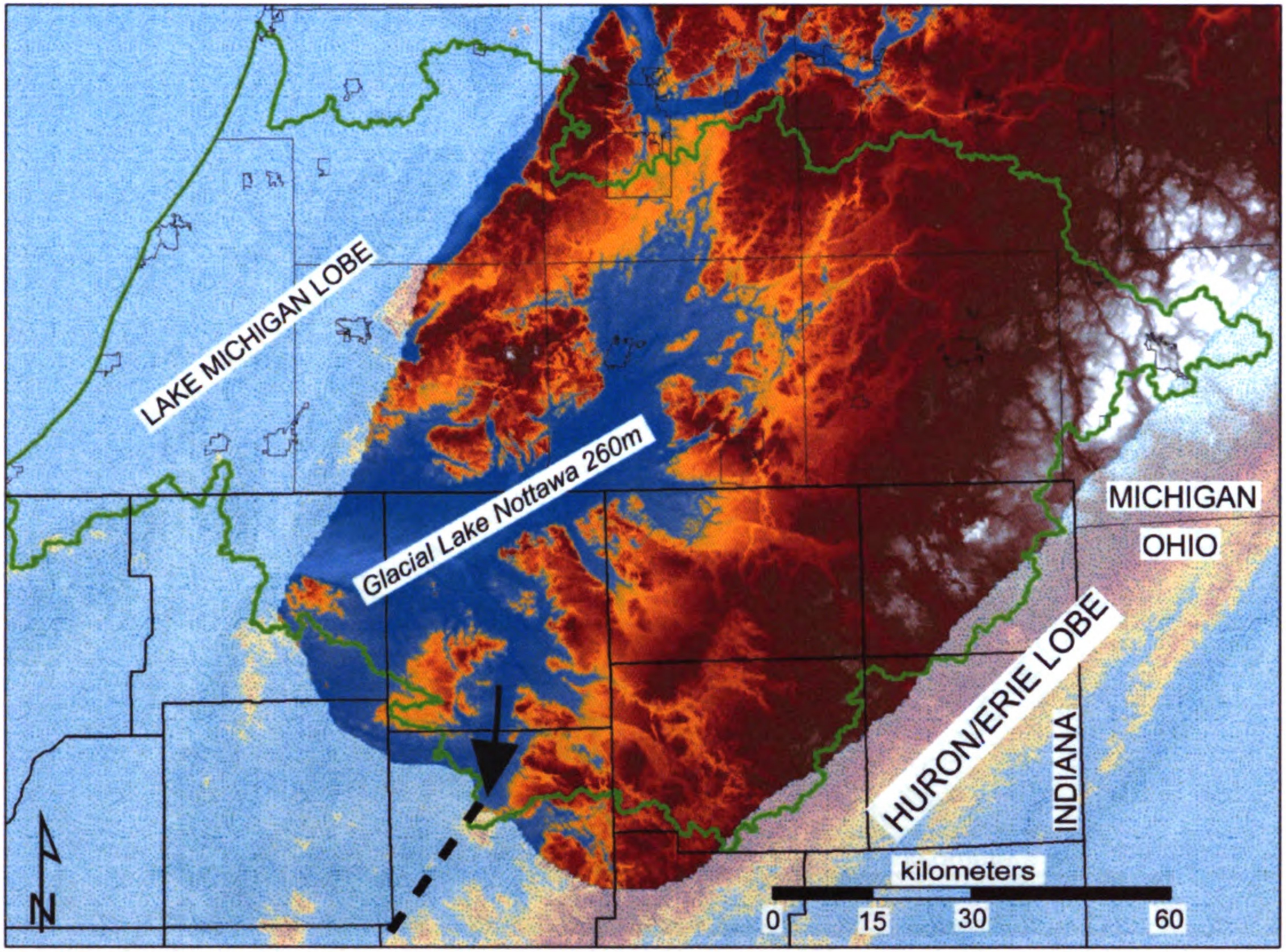

Figure 58. Proposed ice margin confining glacial Lake Nottawa at 260 m altitude. The overflow was in the **vicinity of the seam between the Lake Michigan and Huron/Erie glacial lobes (dashed line).**

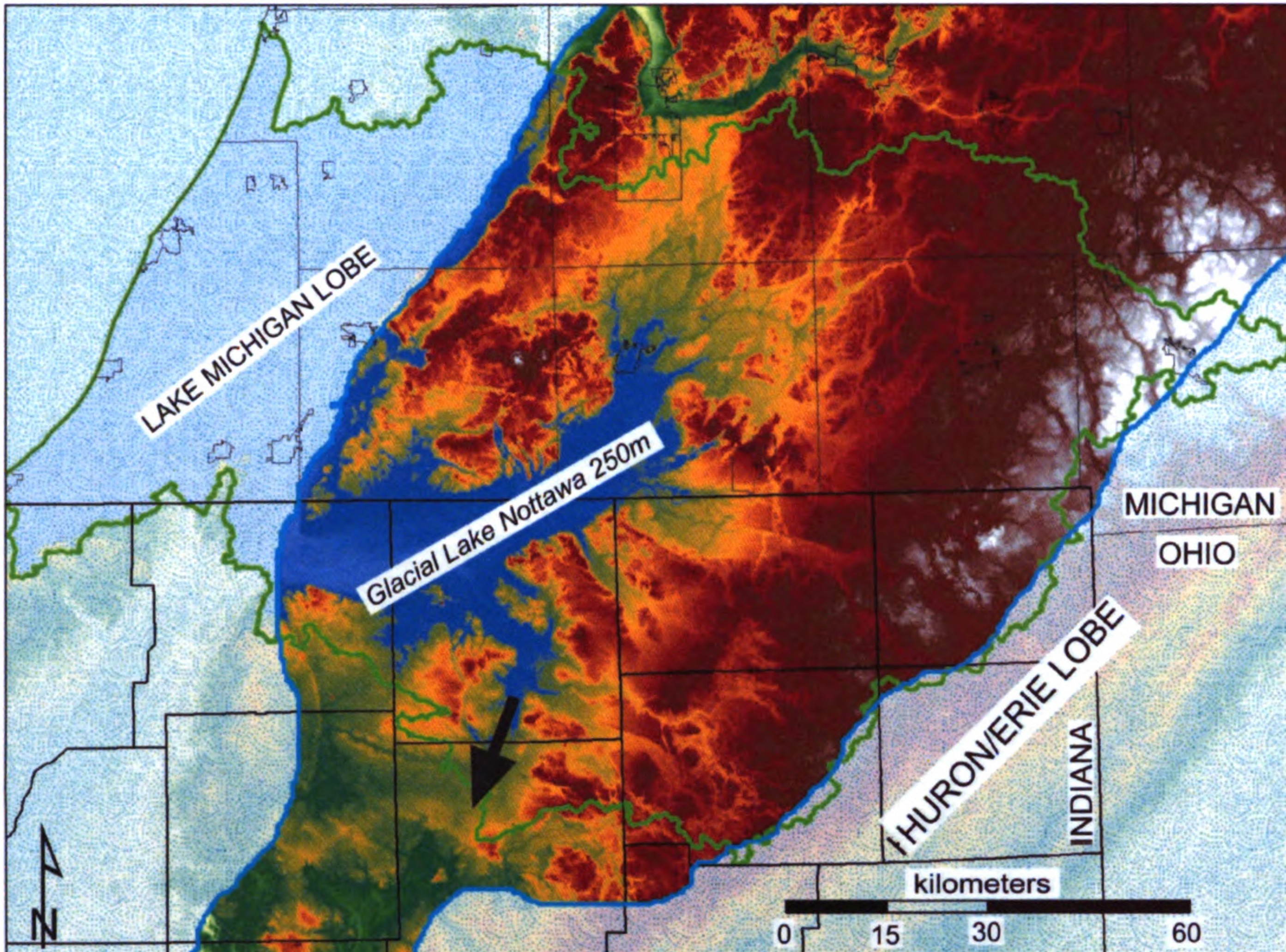

Figure 59. Proposed ice margin confining glacial Lake Nottawa at 250 m altitude. The overflow was in the south to the Wabash River from the vicinity of Leesburg, Indiana.

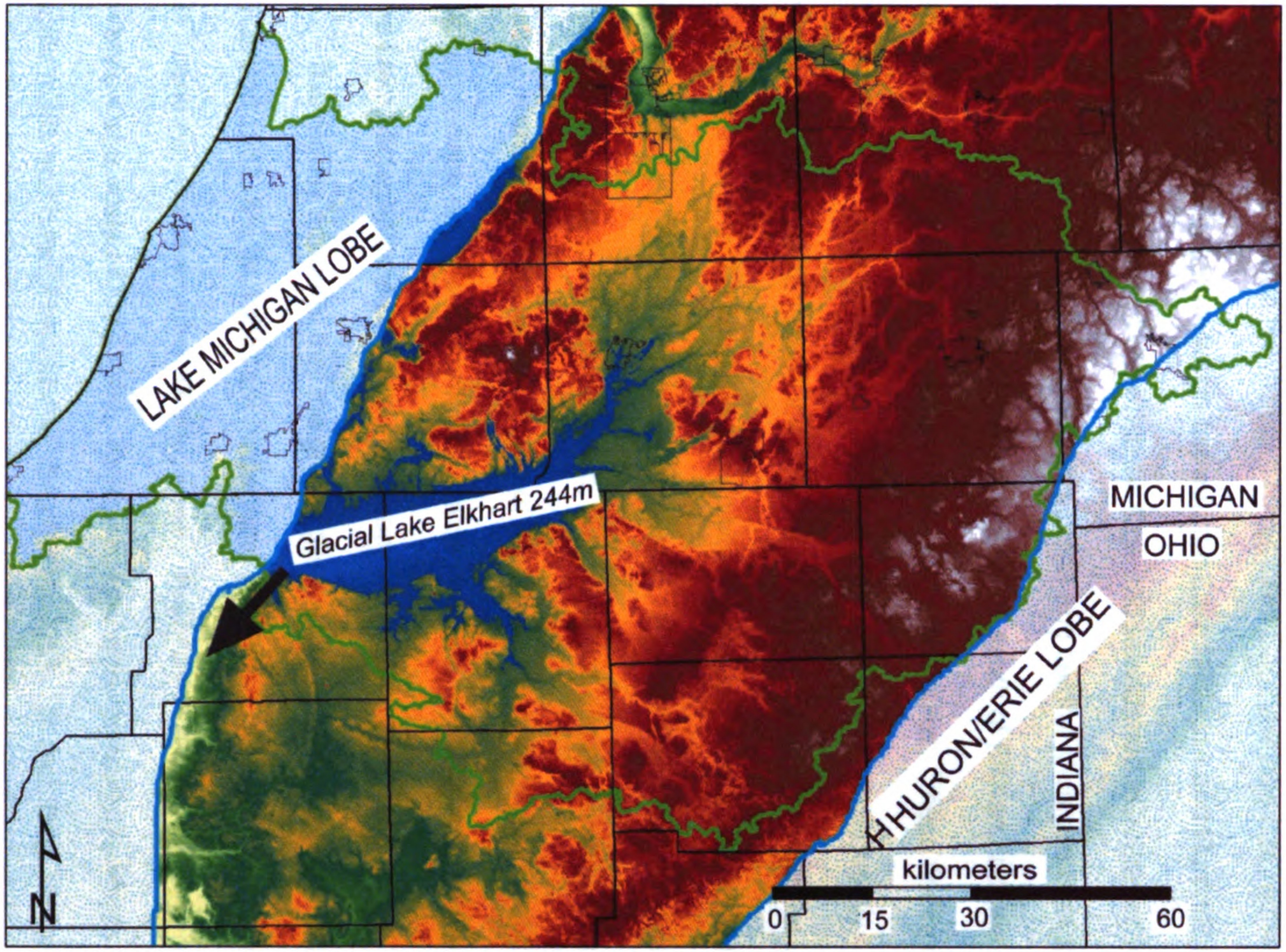

Figure 60. Proposed ice margin confining glacial Lake Elkhart at 244 m altitude. The overflow was south to to the Wabash River across the highest terrace southwest of South Bend, Indiana.

Huron/Erie lobes toward the Wabash River. The Saginaw lobe had retreated from this area long before which allowed the Lake Michigan and Huron/Erie lobes to advance into the space the Saginaw lobe had abandoned. Cross-bedded sand and gravel with southerly flow directions and central Michigan lithologies are found in gravel pits throughout the highlands south of Elkhart, Indiana and are buried the diamicts with western Ohio shale lithologies. These findings attest to a landscape formerly occupied by Saginaw lobe ice and subsequently reoccupied by Huron/Erie lobe ice. Figure 59 shows the outlet for glacial Lake Elkhart over land to the south (in the vicinity of Leesburg, Indiana) after the Lake Michigan and Huron/Erie lobes had separated. Figure 60 shows the retreat of the Lake Michigan lobe to South Bend, Indiana which allows drainage of glacial Lake Elkhart around the upland south of the city and the development of the highest terrace on its west flank with drainage bound for the Wabash River (as the Lake Michigan lobe continued to occupy the Kankakee River basin.

It is readily apparent that the glacial geology maps of the state of Michigan (Farrand 1982, Martin 1955) are inadequate to address resource issues in the 21st century. These small-scale maps were designed to and still provide a fine overview at a scale of 1:500,000. However, they were not compiled from large-scale maps and cannot be used for detailed analysis. In fact, they are based on an outdated model of synchronous movement of glacial lobes and simplistic dynamics of generation of outwash at one time ice-margins with static base-level drainage into the same basins used by present drainage. Each of these three issues is now known to be false and this knowledge should open up a wide range of research possibilities for the foreseeable future. Given the increased dependence of Michigan's economy on the natural resources found in the glacial deposits, it is imperative that this research begin as soon as possible.

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