HISTORIC CHANNEL CHANGES IN THE MUSKEGON RIVER, NORTH-CENTRAL MICHIGAN, USA

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A THESIS

Submitted to
Michigan State University
in partial fulfillment of the requirements
for the degree of

Geography - Master of Science

2015

ABSTRACT

HISTORIC CHANNEL CHANGES IN THE MUSKEGON RIVER, NORTH-CENTRAL MICHIGAN, USA

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This study assesses channel patterns over the Muskegon Rivers entire length during the historic record by comparing stream patterns from the initial land survey taken between the 1830s and 1850s with aerial photos acquired in 1938 and 2012. The research focuses on the following questions: (1) What is the nature of changes in channel position between the three sample times? (2) What is the extent of change in channel width throughout the system from the initial survey to the present? (3) What are the key variables that may have caused channel changes? To assess these questions a combination of field and laboratory analyses were conducted. River positions at each time period were georectified in ArcGIS10 for comparative purposes. The stream was surveyed by kayak during the summer of 2013 in order to collect modern width measurements along section lines and to analyze channel changes since the initial land survey.

Results indicate that numerous changes have occurred in the Muskegon River with respect to channel width and meander pattern during the historic period. Numerous migrating cutbanks and cutoff meanders were found in the upper half of the system where the channel slope is consistently low (averaging 0.3 m per km) and sinuosity is high (~1.8 – 3.2). The stream appears to have narrowed in the lower part of the system since 1836, whereas it changed little in the upper part. I hypothesize that channel narrowing in the lower reaches is related to dam placement and/or devegetated bank slopes and tributary systems that increase sediment yields.

ACKNOWLEDGEMENTS

A lot of people have been generous with their time, friendship, and support to get me to this point in my career. I would like to begin by thanking my undergraduate mentors and advisors, John Anderton, Susy Ziegler, Walt Loope, Robert Legg, Cam Fuess, and Ronald Sundell. Their guidance in my early college career was and still is invaluable and much appreciated. Their support and guidance exposed me to academic research, conferences, and the graduate application process. Another thank you goes out to Daniel Dermyer, Student Support Services, and the Ronald E. McNair Post-baccalaureate Achievement Program, all of which has encouraged me to pursue a graduate career and has given me the resources, support, and determination to do so. I cannot thank them enough for the advice and time they have given me during my time at NMU.

I am much obliged for the guidance of Randy Schaetzl who has given me countless hours of his time here at MSU. His relentless support for my admittance to the graduate program at MSU is greatly appreciated. I am thankful for the opportunity to learn soils and geomorphology from him in the classroom and in the field, which has provided me a vast amount of knowledge and countless new skills. I also thank him for allowing me to participate in research with him.

Other faculty in the MSU Geography Department, such as Dave Lusch, Morris Thomas, David Campbell, and Catherine Yansa deserve my upmost gratitude. I would also like to thank the Geography Department staff, especially Sharron Ruggles, Claudia Brown, and Judy Reginek who have been extremely helpful and friendly to me.

I would like to thank my fellow graduate geographers: Dan Kowalski, Mike Luehmann, Jay Strahan, Brad Miller, Phil Wernette, Albert Fulton, Alison Keener, and Caitlin Clark. Mike and Brad welcomed me into the department and both were great role models for an entering master's student. Jay and Phil were great to have around for technical help and any questions I had regarding GIS. A special mention goes to Dan Kowalski who has been a devoted friend. I also thank him for braving the three day, 85-mile-kayak between Houghton Lake and Evart, which proved to be meandering, mosquito infested, poison-ivy-rich, and log-jam cluttered. A thank you must also go out to Albert, Alison, and Caitlin who were always around for good conversation and entertainment.

I am thankful to my parents Jim and Laurel Michalek and grandparents Art and Marion Ritter, who have provided logistical and emotional support throughout the years. Their support and love has provided me everything I needed to complete this degree. A lot of my determination comes from my grandmother Marian Ritter, who was always a strong believer in hard work and education. I wish she was here to see my progress from a stubborn child who wanted to quit elementary school, to a well-educated graduate student.

To my friend Brandon Peterson, thank you for kayaking the Newaygo County section of the Muskegon River with me. I hope you enjoyed your first kayaking experience. I would also like to thank Jennifer Rosso, who has been a loving friend. She has been a never ending source of cheer, inspiration, advice, and support. I am also grateful for her help kayaking the fast white waters between Evart and High Banks Park.

I would also like to thank the financial aid received from the MSU Department of Geography, the family of Stephen Ripplinger, and the Fillmore C.F. Earney Scholarship, funded my Fritz Nelson, which has helped fund my endeavors and I am grateful for their support.

Last, and certainly not least, I would like to thank my graduate advisor, Alan Arbogast. His academic and life guidance has been invaluable and humorous at times. I appreciate and value the opportunity to have been his research assistant for two years. His prior work on the Muskegon River set the foundation for my study and his continued interest of the system has provided me with the tools and resources to look at the Muskegon's historic channel changes. I cannot thank him enough for the advising he has provided me while also tending to other students and fulfilling his duty as chair of the Geography Department.

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CHAPTER 1

INTRODUCTION

Flowing water is responsible for most geomorphic change on Earth (Leopold et al., 1964; Morisawa, 1985; Pettis and Foster, 1985; Knighton, 1998). The types of landforms produced by flowing water range from tiny rills to large valleys. Within stream valleys a variety of fluvial landforms may be present, such as terraces, oxbows, and the floodplain. These features are indicative of past and present fluvial behavior and adjustments that have occurred in the stream over time.

Many studies in North America have examined fluvial landforms to record how streams have changed over time (e.g., Fisk, 1944; Smith, 1996; Hereford, 2002; Arbogast et al., 2008; Bettis et al., 2008; Ingram, 2008; Leigh, 2008; Stinchcomb et al., 2012; Hall and Peterson, 2013). These studies primarily focus on prehistoric adjustments that occurred during the Late Pleistocene and Holocene. Some of these studies have demonstrated that terraces rapidly formed throughout the period (e.g., Blum and Valastro, 1989; Hall, 1990; Martin, 1992; Arbogast and Johnson, 1994; Smith, 1996; Baker et al., 2000; Daniels and Knox, 2005; Bettis et al., 2008; Vandenburghe, 2014), while others have examined changes in channel pattern (e.g., Fisk, 1944; Smith, 1996) and floodplain stratigraphy (e.g., Bettis et al., 2008; Hall and Peterson, 2013).

Although most fluvial studies have focused on prehistoric adjustments, there has also been a variety of work on changes that have occurred since Euro-American settlement (e.g., Bryan, 1928; Happ, 1944; Schumm and Litchty, 1963; Burkham, 1972; Knox, 1977; Graf, 1978; Magilligan, 1985; Martin and Johnson, 1987; Webb et al., 1991; Johnson, 1994;

Dominick and O'Neill, 1998; Van Steeter and Pitlick, 1998; Ruhlman and Nutter, 1999; Grams and Schmidt, 2002; Juracek, 2002; VanLooy and Martin, 2005; Galster et al., 2008; Ghoshal et al., 2010; McBride et al., 2010; White et al., 2010; Swanson et al., 2010; Dean and Schmidt, 2011; Skorko et al., 2011; Wallick et al., 2011; Horn et al., 2012; Lecce, 2013; Mossa, 2013). These studies have examined the degree of change that has occurred within a fluvial system by using historic records such as photography, survey notes, and personal accounts. These studies have demonstrated that stream systems have been highly sensitive to alterations in climate, vegetation, and anthropogenic change during the historic period. Within the U.S., most of this research has been conducted in three regions, the American Southwest, Central Great Plains, and the Driftless Area of the Midwest.

There are many reasons why channel adjustments have occurred in these regions. In a study in the American Southwest, for example, Bryan (1928) used land surveys and personal accounts to evaluate the widening and deepening of the Rio Grande River in eastern New Mexico. Bryan (1928) concluded that the introduction of livestock in the region initiated a period of erosion that would last for decades. Later work in the region by Webb et al. (1991) indicated extensive arroyo formation on Kanab Creek in northern Arizona. Webb et al. (1991) suggested that a series of large floods, in conjunction with poor land use practices, increased the rate of runoff in the region.

Similar work by Schumm and Lichty (1963), in the Central Great Plains suggested that periods of channel widening and narrowing had occurred between the late 1800s and early 1900s on the Cimarron River in southwestern Kansas. Their work suggested that channel width adjusted several times (ranging from ~15.2 to 365.7 m) due to periods of flooding followed by widespread drought. Similar work in the neighboring Medicine Lodge

River basin was done by Martin and Johnson (1987). Here the authors noticed channel narrowing and the encroachment of riparian vegetation, which they attributed to a decrease in seasonal precipitation. Johnson (1994) also described channel narrowing and increased channel vegetation in the Platte River system of Nebraska, which are thought to be associated with increased irrigation and damming.

Finally, studies in the Driftless Area reveal that an increase in agricultural development and deforestation may have accelerated the rate of aggradation in the regional streams. Work on the Platte River in southwestern Wisconsin by Knox (1977), indicated that stream widths had increased during the historic record. Knox (1977) suggested that changes in channel width were closely associated with flood events since settlement, which likely increased sediment transport in the basin. A similar study by Trimble (2009), also documented an increase in sediment yield on Coon Creek in southwestern Wisconsin. Trimble (2009) associated the increase in sedimentation to deforestation, an increase in agriculture, and pasturing during the historic record.

1.1 Origin of the Research Problem

Although abundant work has been done in some parts of the United States, very little historic channel change research has been conducted in Michigan, which lies in the core of the Great Lakes region. Like the regions previously discussed, Michigan has also experienced historic fluctuations in landcover/landuse, making the state a potentially good place for historic channel change research. An example of extensive human impact during Michigan's historic period is the massive deforestation that occurred during the logging era, which lasted from the mid-1800s to the early 1900s. Large rivers, such as the

Muskegon River, were used to transport logs from the headwaters to mills. Both of these practices could have led to increased erosion and sedimentation rates during this time period. Shortly after the logging era, multiple earthen dams were placed on several Michigan streams to produce hydroelectric power. Many of these impoundments caused discharge and sedimentation rates to change, as seen on the Pine River in Lower Michigan, which filled with sediment during the 1950s (Hansen, 1971; Consumers Power Company, 1994).

Given that extensive human impacts have occurred within Michigan, it is conceivable that measureable alterations have also occurred to river systems. A few studies have been conducted on selected stream reaches that suggest human impacts have played a role in historic channel change in Michigan (e.g., Bowman, 1904; Hansen, 1971; Burroughs et al., 2009). Bowman (1904) for example, investigated the area near the confluence of Willow Run and the Huron River and demonstrated a case of stream capture that occurred in 1904. Later, Hansen (1971) examined a ~42 km reach of the Pine River and found that increased sedimentation had occurred in the stream after the removal of the Stronach Dam.

Although this work in Michigan has demonstrated that isolated stream segments have changed during the historic period, an entire stream system has not yet been analyzed. This study thus focuses on historic channel changes on the Muskegon River in north-central Lower Michigan. The Muskegon River is an excellent stream to assess the extent of historic channel change for multiple reasons. First, numerous oxbow lakes and meander scars have been observed in the upper part of the system, indicating that the stream has actively meandered. Arbogast et al. (2008) indicated that some ^{14}C dates from these oxbows and meander scars on the current floodplain were <1,000 years old, suggesting that stream migration has recently occurred in the system. Secondly, major changes in landcover/landuse have occurred in the region since the onset of the historic period in association with deforestation due to logging. Thirdly, several dams have been placed on the stream in the historic record (e.g., between 1906 and 1931), which may have affected stream discharge. These dams have likely caused sedimentation rates to change, as seen on the Pine River in northwestern Lower Michigan (Hansen, 1971; Consumers Power Company, 1994). Finally, the initial land survey in Michigan was conducted in the mid-1800s, providing a good record of channel dimensions and conditions before human settlement.

1.2 Goals of the Research

This study has several goals. The first is to characterize elements of the Muskegon River from its headwaters to its terminus at Lake Michigan. This will be done by assessing channel attributes, such as sinuosity, slope, radius of curvature, meander wavelength, amplitude, and width along section lines. A second goal is to identify changes in channel position that occurred in the historic period. A third goal is to calculate the extent of change in channel width in the system during the historic period. This will be calculated by collecting initial survey measurements and field data along the same section lines. The final goal of this study is to determine where historic channel changes have occurred and speculate on their causes.

CHAPTER 2

LITERATURE REVIEW

2.1 Fluvial Processes and Landforms

2.1.1 Introduction

A stream is a "self-formed" channel thats "morphology results from the entrainment, transportation, and deposition of unconsolidated sedimentary materials of the valley fill and floodplain deposits across which they flow" (Richards, 1982, pp. 1). Streams are also thought to be the "primary agent by which the surface of the earth is degraded" (Morisawa, 1985, pp. 11), suggesting that the geomorphology of most landscapes have at one time been shaped by a fluvial system. These systems are found in many different environmental settings and are thus diverse in form, such as channel pattern, discharge, slope, and many other variables (Charlton, 2008). This review is organized to discuss both channel patterns and the landforms that can be found in association with a fluvial system.

2.1.2 Channel Patterns

Stream systems exhibit a variety of drainage patterns. The spatial characteristics of these patterns provide valuable information about the environment in which a particular stream occurs (Gabler et al., 2007). The dominate influences on drainage patterns are bedrock structure and the topography of the surrounding landscape (Gabler et al., 2007). These factors can control a variety of stream characteristics, such as the number of tributaries, sinuosity, and drainage density.

The most common type of drainage pattern is dendritic, which has an irregular branching appearance with tributaries connecting to main channels at acute angles (e.g., less than 90^{o}; Figure 2.1; Gabler et al., 2007). This drainage pattern develops in terrain with a uniform bedrock structure, which normally does not influence the arrangement of the fluvial system (Gabler et al., 2007; Arbogast, 2011). In contrast, a trellis drainage pattern is commonly controlled by parallel outcrops of resistant ridges with more erodible bedrock structures in between (Gabler et al., 2007). For example, this scenario is common in the Ridge and Valley province of the Appalachian Mountains, which has a folded landscape due to faulting (Gabler et al., 2007). Trellis patterns have long parallel main streams with short tributary channels joining the main channels at right angles (Figure 2.1).

Like the trellised pattern, some drainage patterns are only found in areas with certain bedrock structures. This is the case for radial patterns, which develop outward from rounded uplands, such as a volcano or dome (Figure 2.1; Arbogast, 2011). The opposite pattern is centripetal, where drainage flows inward into a lake or basin (Figure 2.1). This pattern is common in the arid Basin and Range province of the American Southwest, frequently forming playas (Strahler and Strahler, 1992). Another bedrock controlled drainage pattern is rectangular. This pattern is typical in areas with jointing and faulting of shallow bedrock and exhibits straight streams flowing in fractures and commonly intersecting at right angles (Figure 2.1; Gabler et al., 2007).

The final drainage pattern is deranged, which is characterized as having an irregular direction of flow (Figure 2.1; Arbogast, 2011). Deranged drainage patterns typically have a low number of tributaries with many interconnected wetlands and lakes (Arbogast, 2011). This pattern commonly develops in areas that have been recently deglaciated (e.g., northern Canada; Gabler et al., 2007), with relatively low slopes that promote these stream systems to wander between marshes and small lakes (Gabler et al., 2007).

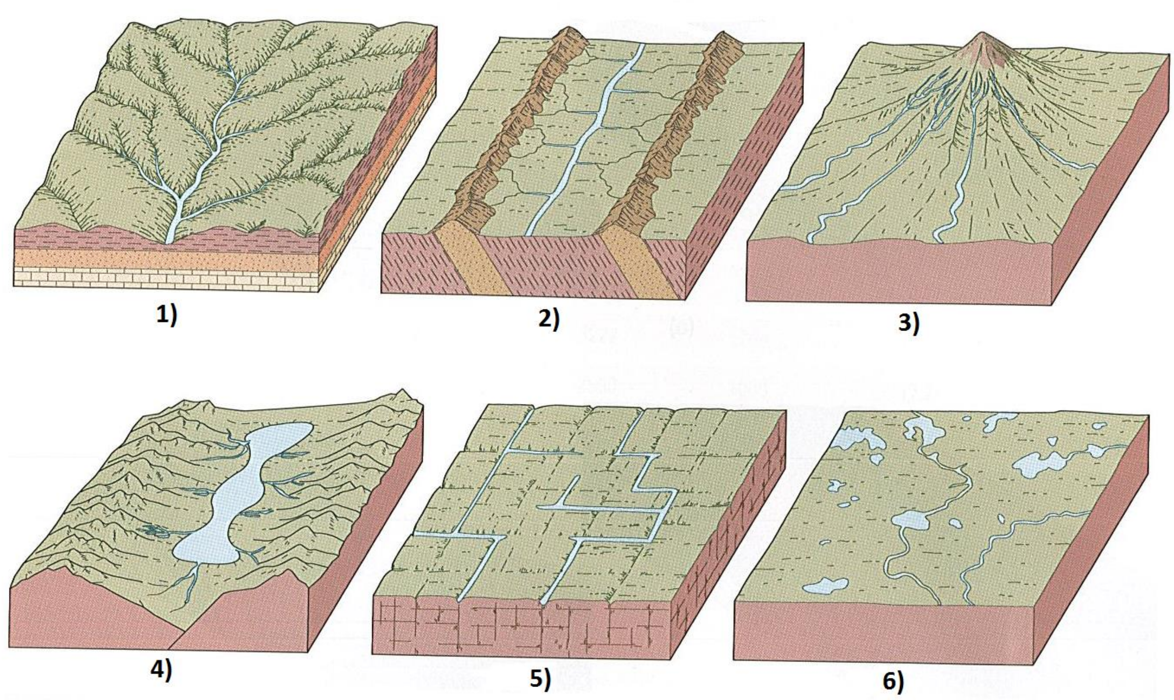

Figure 2.1. Common drainage patterns: 1) Dendritic Pattern, 2) Trellis Pattern, 3) Radial Pattern, 4) Centripetal Pattern, 5) Rectangular Pattern, and 6) Deranged Pattern. (modified from Gabler et al., 2007)

In addition to these drainage patterns, stream systems can be further classified based on the form of their channel, specifically whether they are single or multi-thread (Petts and Foster, 1985). Single-thread streams can have many forms, such as straight, sinuous, or meandering. While straight streams are rare in natural flowing systems, they can occur in some stream reaches. Bradshaw et al. (1978) suggested that straight stream segments on natural streams are typically less than ~100 m in length. Related work by Langbein and Leopold (1966) suggested that straight stream segments typically are not longer than about ten channel widths. This relationship suggests that wider streams generally have longer straight channel reaches than those of narrower streams.

Typically, most single-thread channels are sinuous or meandering (Charlton, 2008). The beginning of a meandering pattern in these streams is thought to occur due to a uniform energy loss (Leopold et al., 1964), which is exerted laterally across the floodplain. Richards (1982) suggested that meander formation is common in areas that have topographical and sedimentological constraints that disturb the directional uniformity of a low energy stream. The degree of meandering in these systems varies from relatively straight stream segments to elaborate meander bends (Charlton, 2008), which are common in streams that experience erosion on concave banks adjacent to scour pools (Knighton, 1998). Most of the sediment eroded along a channel bank is transported via helical flow (e.g., corkscrew type of circulation within the flowing channel) toward the next point bar deposit (Figure 2.2; Easterbrook, 1999). The continuation of this process accelerates the development of meandering in a river system.

There are several mechanisms behind the formation of a meandering channel. Perhaps one of the most important factors to consider about a meandering system is the streams tendency to distribute its energy across its floodplain. This distribution is generally achieved as a stream laterally migrates across the floodplain via individual meander bends (Charlton, 2008). The major force that causes a meander to migrate is channel circulation and flow patterns. As previously stated, a common type of flow associated with a meandering system is helicoidal flow. Commonly, helicoidal flows parallel both sides of the stream's thalweg (e.g., the deepest and fastest portion of a stream; Figure 2.2), which migrates and in time forms several cutbanks. Frequently, these cutbanks become undercut, providing additional sediment to the fluvial system, while also slightly changing the channel boundary. Eventually a stream's helicoidal flow becomes stronger as the system increases in velocity and depth. Hey and Thorne (1975) believe this model is over simplified and suggest that there are secondary helicoidal cells, which converge and diverge depending on channel flow. This research suggests that there are many moving parts within the stream, only adding to the complexity of meander formation and stream flow.

In addition to helicoidal flow, a stream's centrifugal force also plays a role in the movement of the channel (Wormleaton et al., 2005). The force of a channel's flow pushes water against the outer bend of a meander, causing the water surface to be slightly elevated. This leads to alterations to the channel bed and lateral flow within the stream, eventually causing the channel to slowly form a new morphology (Wormleaton et al., 2005). Together with helicoidal flow these forces aid in the migration of meander bends (e.g., translation, extension, and/or rotation) in a downstream trend (Chen and Duan, 2006).

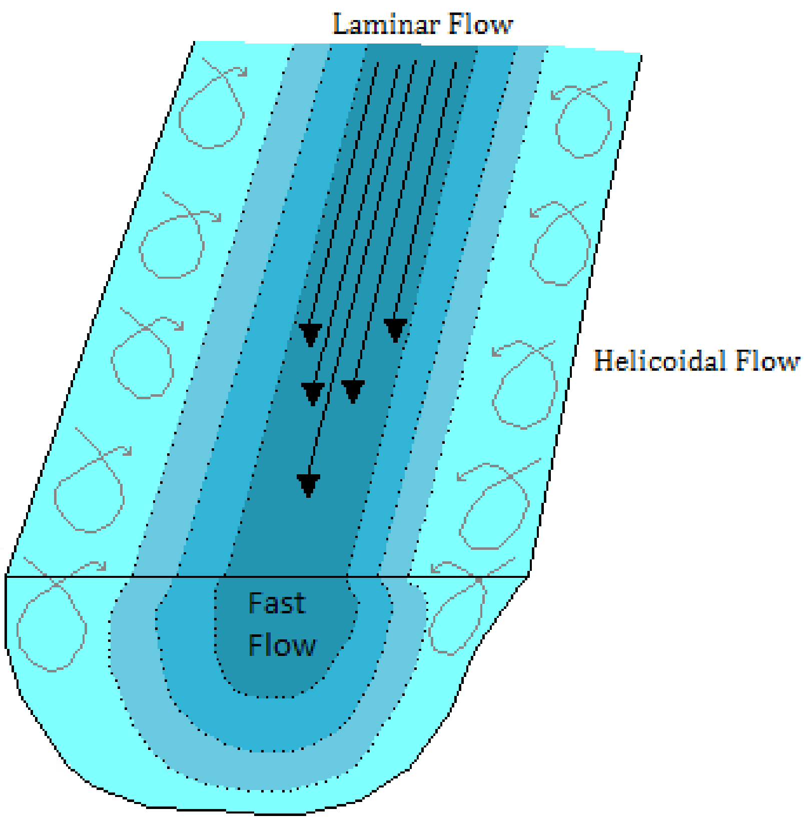

Figure 2.2. Diagram showing both helicoidal and laminar flow, as well as flow speed in a three-dimensional channel cross-section.

According to Knighton (1998), meanders can be analyzed using two main approaches. The traditional approach examines the meander geometry of individual channel bends, such as meander wavelength, meander amplitude, and its radius of curvature. Another approach for measuring the spacing between meanders in a given reach is done by determining meander wavelength. This variable is calculated by measuring the distance between two adjacent meander crests (Figure 2.3). Chorley et al. (1984) suggested that a relationship exists between channel width and meander wavelength, with meander wavelength typically about 10 to 14 times more than its bankfull width.

In contrast to the wavelength of a meandering stream, meander amplitude is the measurement of a river bend's lateral extension across the floodplain (Figure 2.3). Lateral extension of a meander can occur as a channel migrates horizontally across a floodplain, increasing the amplitude of the meander and length of the channel (Charlton, 2008). Finally, radius of curvature is used to determine the tightness of a single meander bend. This measurement is calculated by placing a circle along a meander bend, then calculating its radius (Figure 2.3). Dividing a stream's radius of curvature and width provides a function to assess streams of different sizes (Charlton, 2008). Small numbers are representative of tight meanders, whereas larger numbers indicate bends with subtle curvature.

A second approach to characterizing a meandering stream involves analyzing sinuosity. Sinuosity is calculated by dividing the length of a given channel reach by its river valley length (Figure 2.4, a; Charlton, 2008). Generally, streams with a sinuosity of ~1 are considered straight, while streams with sinuosity values between ~1 and 1.5 are categorized as sinuous (Figure 2.4, b). Streams with a sinuosity >1.5 are labeled as

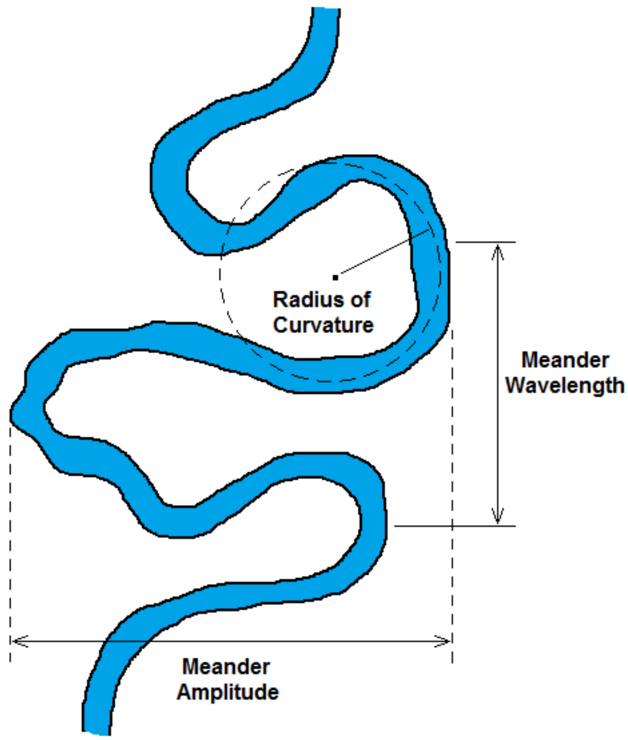

Figure 2.3. The meander geometry of a stream, including meander wavelength, amplitude, and radius of curvature. (adopted from Charlton, 2008)

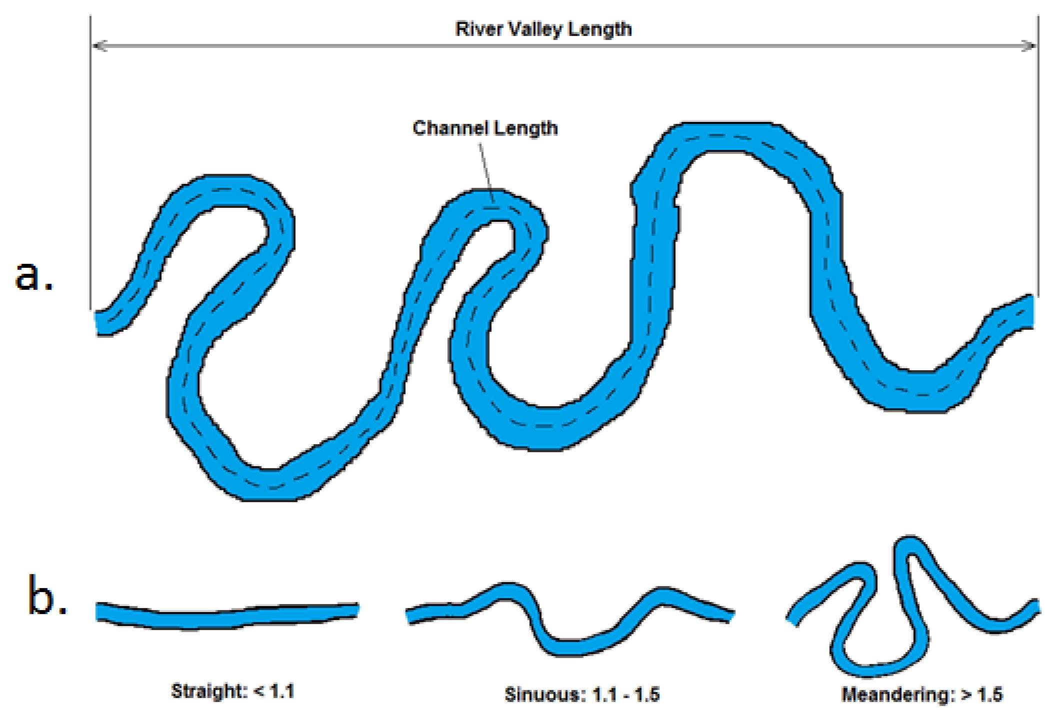

Figure 2.4. a.) Diagram showing how sinuosity is calculated. b.) The appearance of straight, sinuous, and meandering streams in regards to sinuosity. (modified from Charlton, 2008)

meandering (Figure 2.4, b; Leopold et al., 1964; Petts and Foster, 1985; Charlton, 2008). Schumm (1968) noted that sinuosity can vary among different rivers and stream segments and suggested that factors such as valley slope, channel gradient, and the percentage of silt-clay in channel banks can influence channel migration.

Multi-thread channels are also found in natural flowing systems and are generally categorized as being either braided or anastomosing. Braided streams are the primary multi-thread channel type with channels that are separated by bars or islands (Knighton, 1998). These types of streams are found in many different settings. Braided streams are common in areas with a coarse-textured sediment source, such as downstream from an alluvial fan or melting glacier.

Braided and meandering streams differ in several important ways. First of all, braided streams tend to have wide shallow channels because their banks are less cohesive (e.g., sands and gravels) than meandering channels (Figure 2.5; Schumm, 1960; Nanson and Croke, 1992). Secondly, braided systems commonly have higher and coarser sediment loads than meandering streams. In an experiment by Schumm et al. (1987), for example, stream power and the rate of sediment feed in braided channels was almost 5 times greater than that of a meandering stream. Lastly, braided systems tend to form in higher energy streams (Richards, 1982; Schumm et al., 1987; Nanson and Croke, 1992), which is associated with the medium to high gradient of the channel bed (Schumm, 1960; Mangelsdorf et al., 1990).

An anastomosing stream is another type of multi-thread channel that is less common and forms in flood-dominated areas near a body of water or marsh (Nanson and Knight, 1996). Bradshaw et al. (1978) described an anastomosing stream as an assemblage of channels that divide and recombine with hills between the individual branches (Figure 2.5). Schumm (1989) distinguished the differences between a braided channel and an anastomosing stream by noting an anastomosing channel's low gradient with multiple deep narrow channels. Bradshaw et al. (1978) attributed the formation of an anastomosing channel to a shift from an arid climate to humid climate, which increases stream discharge. Anastomosing channels can also be regarded as a subcategory of an anabranching river, which is associated with banks resistant to erosion in flood-dominated fluvial systems (Nanson and Knight, 1996).

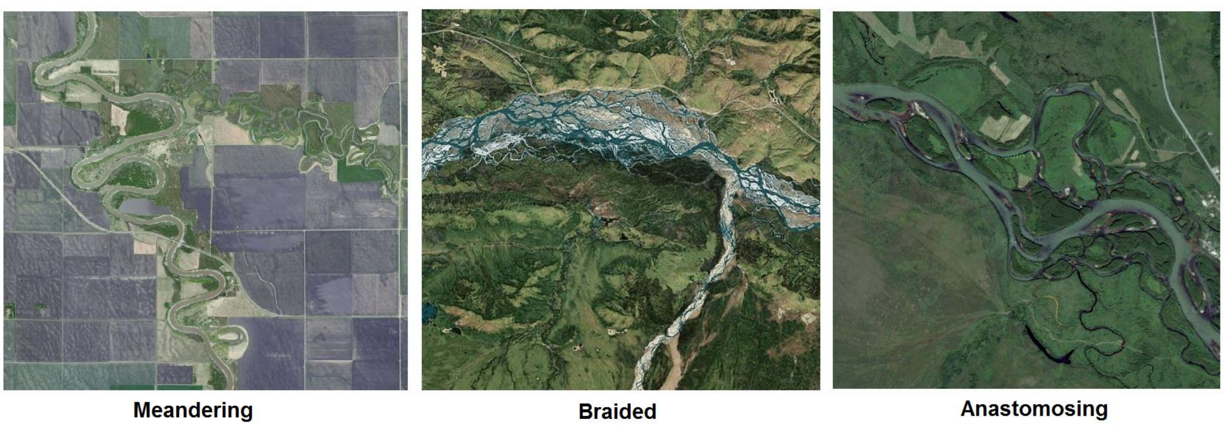

Figure 2.5. Aerial images of meandering, braided, and anastomosing channel patterns.

2.1.3 Stream Landforms

As previously discussed, stream channels can develop in an array of different patterns and settings. These patterns form over time as channelized water erodes and transports material from small tributaries to higher-order streams, eventually reaching the ocean. Stream valleys are generally regarded as erosional features which formed via flowing water. Stream valleys are commonly shaped differently depending on factors, such as bedrock geology, gradient, and time (Gabler et al., 2007). Commonly, stream valleys are V-shaped in the upper reaches of a system where streams typically are in contact with underlying bedrock and valley gradients are high (Gabler et al., 2007), which promotes downcutting. Valleys become increasingly broad in downstream reaches where gradients are low and channel migration is common. This allows alluvium to be deposited across the entire valley floor as the channel continually erodes and deposits sediments.

Of the fluvial surfaces in a stream valley, the floodplain lies closest to the elevation of the stream and is the most active surface. All streams have some form of a floodplain, which is generally defined as “a strip of relatively smooth land bordering a stream and overflowed at time of high water” (Leopold et al., 1964, pp 317). Dunne and Leopold (1978) describe a floodplain as a feature that is continually being shaped by a stream in its current condition and climate. The generally accepted view is that a floodplain is inundated by the associated stream about every 1 to 2 years during periods of high discharge (Wolman and Leopold, 1957; Leopold et al., 1964; Dury, 1973; Dunne and Leopold, 1978; Morisawa, 1985; Charlton, 2008).

Floodplains are thought to be produced by one of two processes. One way is by lateral accretion, which deposits sediments on the point bars via channel meandering. The source of such sediments is usually from upstream channel banks (e.g., cutbanks). Once eroded, these sediments are then later deposited in low-energy reaches of the stream (e.g., point bars; Mackin, 1937; Wolman and Leopold, 1957; Leopold et al, 1964; Bradshaw et al., 1978; Richards, 1982; Knighton, 1998; Charlton, 2008). Another process that contributes to floodplain development is vertical accretion, which occurs when a river overflows its banks and deposits sediments. These sediments typically are transported as fine suspended load and deposited on the adjacent floodplain during overbank events. These fine vertically accreted sediments generally comprise ~10% of a stream's floodplain (Morisawa, 1985). As a result, the morphology of a stream's floodplain is directly linked to the behavior and characteristics of the stream that formed it (Charlton, 2008).

Many landforms are produced by a migrating channel and the natural evolution of its floodplain. The most common features found on a floodplain are point bars, cutbanks, oxbows, meander scars, natural levees, splays, backswamp deposits, and sloughs (Figure 2.6; Leopold et al., 1964). In many cases these features are found in conjunction with one another on a floodplain. A good example of this relationship is the association of point bars and cutbanks. As previously stated, sediment eroded from high energy flows along a cutbank is transported downstream to the next point bar where energy is again low. Other important features are found in highly sinuous reaches in a river valley, where oxbows and meander scars typically form. These features form as a meander bend "short-circuits" two adjoining cutbanks (Charlton, 2008, pp. 136). Many forces can cause this short-circuiting, such as flood events with higher discharges, continued lateral erosion, and/or an actively migrating channel. While oxbow formation can be enhanced due to external forces, these features typically form throughout hundreds of years (Schumm, 1977). When two adjoining cutbanks have merged, the stream takes a straighter appearance, abandoning its old meander pattern. In time, these oxbows become drier and fill with sediment, commonly transitioning into marshes or wetlands.

A variety of additional features form on the floodplain adjacent to the channel boundary. These features are dependent on overbank flood events for their continued

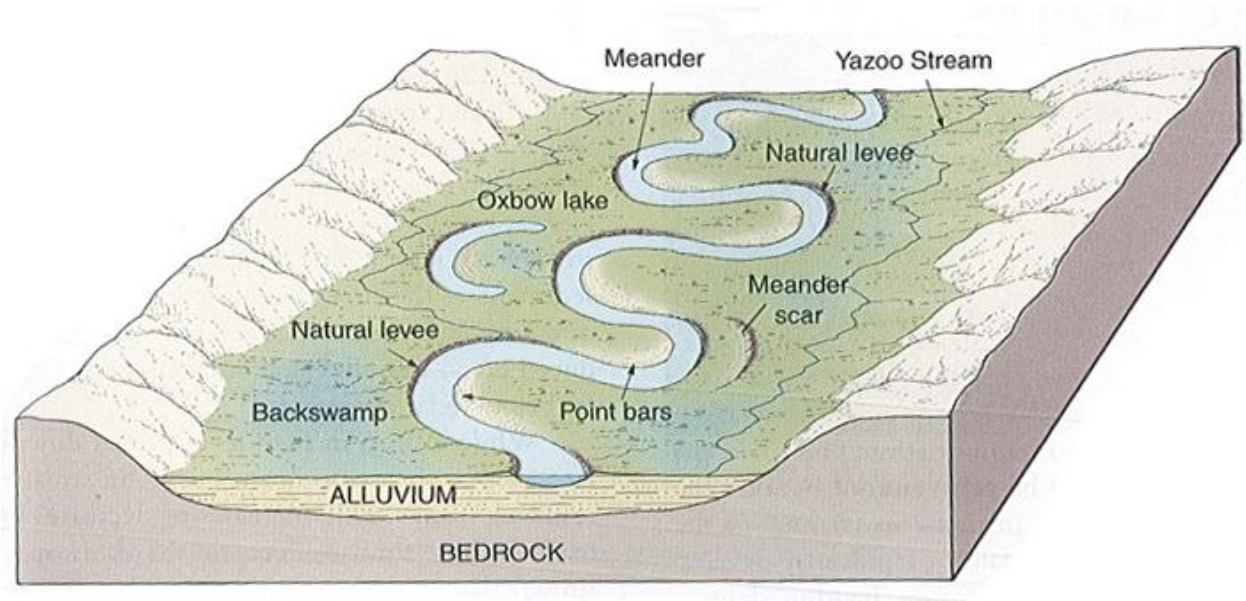

Figure 2.6. Features of a well-developed floodplain. (modified from Gabler et al., 2007)

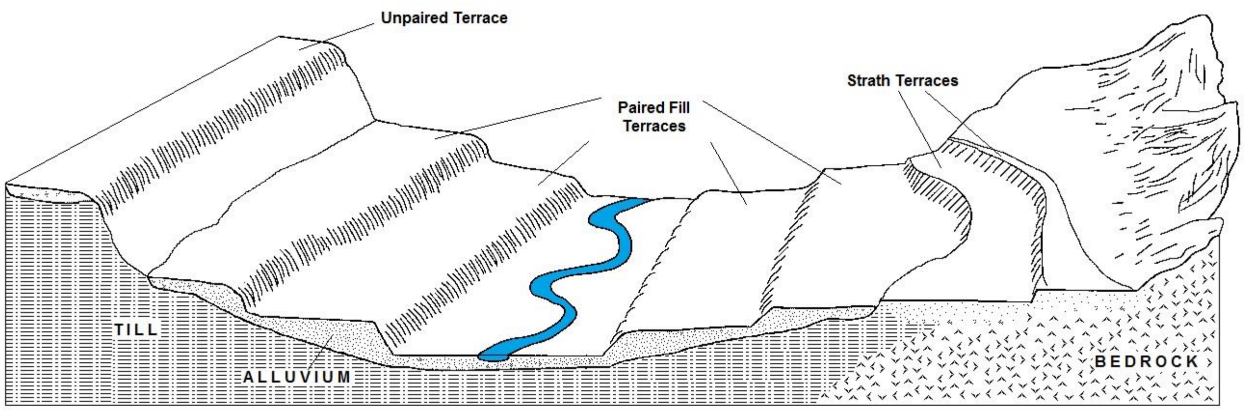

Figure 2.7. Examples of different types of terraces (e.g., paired vs. unpaired, fill vs. strath; modified from Charlton, 2008).

development. These events produce sediments and the flow needed to form and shape features, such as natural levees, crevasse splays, and backswamps (Figure 2.6). Natural levees are elongated, raised ridges that form at the channel-floodplain boundary during periods of overbank flow, when coarse sediments are deposited next to the stream (Charlton, 2008). Natural levees can subsequently be breached by floodwaters, especially when they have been weakened by previous flood events. Such breaches produce crevasse splays, which are fan-shaped lobes of sediments deposited on the floodplain (Charlton, 2008). The frequent breaching of natural levees may also produce backswamp deposits.

After a floodplain forms it can be abandoned due to processes such as stream incision or overbank deposition, creating a terrace (Petts and Foster, 1985). Multiple terraces can form in a river valley due to a series of cut and/or fill events, which can create a sequence of benches above the current floodplain (Charlton, 2008). One way a terrace forms is through the aggradation of sediment on a pre-existing floodplain. A second way is through channel incising, which is commonly caused by a change in sediment load or uplift. Both of these processes can occur in the same river valley throughout time. These alterations normally occur because of local and regional effects, such as baselevel adjustments and changes in climate or landuse (Charlton, 2008).

There are two primary types of terraces, specifically 1) fill and 2) strath. A fill terrace can be created through repeated cycles of aggradation in the valley and associated incision of the stream (Leopold et al., 1964; Morisawa, 1985). Valley wide aggradation is typically brought on by changes in sediment load, which can occur from multiple external changes (e.g., climate change, landuse/landcover change, mass wastage events, or a flux in glacial sediments). In contrast, a strath terrace usually forms during the lateral incision of the channel into an existing surface. Strath terraces are typically cut into bedrock with a thin alluvial cover (Figure 2.7; Morisawa, 1985).

In addition to the type of terrace (e.g., fill and strath), such surfaces can be further classified as being either paired or unpaired. Paired terraces are surfaces that occur at the same elevation of either side of the stream (Figure 2.7; Morisawa, 1985) and typically form during relatively quick downcutting events. Unpaired terraces, in contrast, form during an episode of slow incision that occurs in conjunction with lateral migration (Morisawa, 1985). As a result, an uneven terrace surface is produced during a single downcutting event (Figure 2.7).

2.2 Historic Channel Change Studies in North America

The following discussion reviews previous studies associated with historic channel change in North America. Such research has focused largely in the American Southwest, Central Great Plains, the Driftless Area in the Midwest, and along both the Atlantic and Pacific coasts. In these regions, changes in variables such as climate, riparian vegetation, landuse, discharge, and sedimentation during the historic period have caused measurable fluctuations in fluvial systems. This review is organized by regions, specifically (1) the American Southwest, (2) the Central Great Plains, (3) the Driftless Area of the Upper Midwest, (4) Atlantic Coast, and (5) Pacific Coast.

2.2.1 American Southwest

The majority of studies about historic changes in stream systems have been conducted in the American Southwest, where recent climatic and landcover adjustments have altered fluvial systems. Most of these studies have focused on arroyos, which are deep gullies or streams that have relatively flat channel beds. The first of these studies was conducted by Bryan (1928), who investigated channel change along the Rio Puerco, a tributary of the Rio Grande in eastern New Mexico. Bryan (1928) used multiple land surveys and personal accounts taken between 1855 and 1927 to estimate channel width. Results indicated that the stream became deeper and wider in some places, which Bryan (1928) linked to the introduction of livestock and overgrazing during this time. That being said, Bryan also indicated that cyclic changes in stream sedimentation and erosion occurred before settlement, suggesting that channel change and arroyo formation is a natural process. The introduction of cattle could have, however, been a threshold event that initiated a period of erosion and arroyo enhancement (Bryan, 1928).

Following the Bryan (1928) study, Burkham (1972) examined channel changes in the Gila River, a tributary of the Colorado River in southern Arizona. In this study a variety of historical data (e.g., diaries, journals, surveys, and aerial photography) were used to examine changes in stream width, depth, morphology, and riparian vegetation from 1875 to 1970. In contrast to Bryan's (1928) study, overgrazing was not considered to be an important factor because only a small amount of the land in the basin had been cleared (Brukham, 1972). Burkham (1972) reported that stream widening had occurred intermittently between 1905 and 1917 and was likely associated with major floods events.

Other research in the American Southwest has focused on fluvial adjustments associated with the spread of tamarisk north across the Colorado Plateau. Tamarisk is a deciduous shrub that is highly invasive in the American Southwest because it produces a prodigious amount of seeds and prefers moist sandy areas along streams. A study by Graf (1978), used historical ground photography and survey records to analyze the spread of tamarisk between 1914 and 1968. Results indicated that tamarisk advanced into the area at the rate of 20 km/yr. During this time, tamarisk stabilized on low terraces where there was increased water availability. The establishment of tamarisk in these areas promoted the deposition of alluvial sediments around plant rootlets, which in turn caused these streams to narrow (Graf, 1978). The reduction of channel width ranged from about 13% to 55% between 1890 and 1976 (Graf, 1978). In response, the fluvial system adjusted by developing enlarged stabilized islands and bars.

Work by Webb et al. (1991) built off prior arroyo research to examine the extent and causes of historic channel changes on Kanab Creek, a tributary of the Colorado River in southern Utah and northern Arizona. Prior work along Kanab Creek (e.g., Dutton, 1882; Davis, 1903; Dellenbaugh, 1908; Brandenberg, 1911; Gregory, 1950; Burkham, 1970; Robinson, 1970; Butler and Mundorff, 1972; Robinson, 1972; Webb, 1985) had been extensively documented in the past, allowing Webb et al. (1991) to determine the degree of arroyo formation in the valley. Results indicated that extensive arroyo formation had occurred during the last century, resulting in the incision of Kanab Creek and its tributaries. This incision began during the 1930s after a series of large flood events that eroded Kanab Creek's floodplain (Webb et al., 1991). Webb et al. (1991) associated these flood events with poor land use practices throughout the watershed, which increased the rate of runoff.

Research by Dominick and O'Neill (1998) examined changes in stream morphology and riparian vegetation cover along several tributaries of the upper Arkansas River basin in south-central, Colorado. Aerial photography was used to analyze these changes between 1939 and 1988. From this imagery, Dominick and O'Neill (1998) determined that changes in channel pattern had occurred in many of the study sites, specifically a shift from a highly sinuous channel to a less meandering and/or braided channel. This transition was attributed to an adjustment in channel shape in order to achieve a new equilibrium (Dominick and O'Neill, 1998). As a result, many streams experienced channel widening, which caused a ~10% loss in riparian vegetation in bottomland areas adjacent to the active channel (Dominick and O'Neill, 1998). This loss, in turn, accelerated channel widening, creating exposed gravel bars in previously vegetated reaches.

Related work conducted by Van Steeter and Pitlick (1998) along the neighboring upper Colorado River evaluated how changes in stream flow, sediment load, and channel morphology affected endangered fish habitats. Three contiguous reaches of the river were studied near Grand Junction, Colorado. Aerial photography taken between 1937 and 1993 was used to analyze changes to stream morphology. This assessment indicated that the main channel of the Colorado River narrowed an average of 20 m during this time (Van Steeter and Pitlick, 1998), which had a dramatic effect on side channels and backwater reaches. Channel narrowing was thought to occur through two main processes: lateral accretion along channel banks and vertical accretion in side channels (Van Steeter and Pitlick, 1998). Van Steeter and Pitlick indicated these areas were shallow and susceptible to lower velocities associated with dam construction that regulated normally high flows. These changes in streamflow and channel morphology negatively impacted squawfish populations throughout the study area by increasing the deposition of fine sediments.

Additional research on channel narrowing was done by Grams and Schmidt (2002) along the Green River in northeastern Utah. This study measured streamflow and changes in channel width downstream from the Flaming Gorge Dam. Grams and Schmidt (2002) obtained pre-dam photos from 1871 and matched them with photographs taken between 1993 and 1995. These photographs showed substantial changes in channel and floodplain characteristics, including an increase in riparian vegetation. Photos also indicated that the Green River had narrowed since 1871, primarily due to the aggradation of sediments on gravel bars. These bars have increased in size, allowing riparian vegetation such as tamarisk to stabilize channel banks (Grams and Schmidt, 2002). This study indicated that regulated flows coming from the Flaming Gorge Dam have increased the rate of aggradation, creating multiple post-dam surfaces.

Another study using repeat photography was conducted by Dean and Schmidt (2011), who focused on channel adjustments on the lower Rio Grande River in the Big Bend region of Texas. Multiple ground and aerial photographs were used to analyze channel changes. Ground images from the early 1900s were used to examine channel change before aerial photography was available. Dean and Schmidt (2011) determined from these photos that channel narrowing had occurred during the historic record. The authors attributed the narrowing to large flood events, which scoured channel banks and created large sediment deposits along channel margins. These flood events also produced multiple alluvial fills on the floodplain that were dated by counting tree rings from already established vegetation. This reconstruction allowed the authors to correlate the rate of filling with historically recorded flood events.

Related work was conducted on the upper Rio Grande River near Albuquerque, New Mexico by Swanson et al. (2010), who also focused on historical channel narrowing. By analyzing aerial photography acquired between 1935 and 2008, measurements of stream widths were calculated in GIS and compared to changes in annual peak discharge at gauging stations. This comparison allowed the authors to relate width adjustment to on ground measurements. Swanson et al. (2010) suggested that stream widths narrowed near the confluence of tributaries and in the upstream portions of the Rio Grande. They related the Rio Grande's change in width to increased sedimentation above the Cochiti Dam, which has caused degradation of downstream river segments. The narrowing at tributary confluences is likely caused by reductions in peak discharge, which are attributed to both short-term climatic fluctuations and dam construction (Swanson et al., 2010). Reductions in discharge have also aided the expansion of riparian vegetation near channel margins. As a result, a number of once vegetated islands have coalesced with the surrounding floodplain, changing the morphology of the stream (Swanson et al., 2010).

Skorko et al. (2012) built upon regional historic channel-change research with their work on adjustments to numerous tributaries of the Great Salt Lake in Utah. This study analyzed the response of Lee Creek and Goggin Drain to a change in base level (i.e., the Great Salt Lake). A series of satellite and aerial images from 1965 to 2005 were compiled to track fluvial adjustments on a decadal scale. The level of the Great Salt Lake fluctuated ~6 m in the period after 1965, from an historic low stand (<1279 m) to an historic high stand (>1284 m), to near another low stand in 2010 (Skorko et al., 2012). This adjustment has caused many tributaries in the basin to change their meander pattern, such as Goggin Drain, which had experienced three separate major avulsions (e.g., dramatic shifts in the meandering pattern) since 1965 (Skorko et al., 2012). These changes were linked to rapid lake-level fluctuations. In contrast, streams like Lee Creek had shown little sign of channel adjustment since 1965. Lee Creek is thought to have not been as influenced by these fluctuations because of its lower annual discharges.

2.2.2 Central Great Plains

In addition to historic channel change studies in the American Southwest, there have also been several such studies conducted in the Central Great Plains. The earliest of this work was conducted by Schumm and Lichty (1963), who studied channel widening along the Cimarron River, a tributary to the Arkansas River in southwestern Kansas.

Schumm and Lichty (1963) used survey notes, historical accounts, and aerial photography to estimate channel widths between 1874 and 1960. Schumm and Lichty (1963) claimed that channel widths dramatically increased from ~15.2 m to ~365.7 m between 1914 and 1942. This period of channel widening was thought to be initiated by the flood of 1914, which was followed by extensive drought conditions. Shortly after 1942, the Cimarron River went through a period of floodplain reconstruction that was predominately driven by the vertical accretion of sandy sediments in the system (Schumm and Lichty, 1963). After the reconstruction of the floodplain, channel widths narrowed to ~168 m (Schumm and Lichty, 1963).

Related work done in the Medicine Lodge River basin, a tributary to the Salt Fork of the Arkansas River in southwestern Kansas, was done by Martin and Johnson (1987), who identified historic changes in stream width on three streams in the basin. These adjustments were calculated by comparing current stream widths (along section lines) to measurements taken during the original survey. Results indicated that channel widths narrowed in most locations since the 1871 survey, likely due to decreased seasonality of precipitation (Martin and Johnson, 1987). This decrease resulted in less variable stream discharge and lower peak discharges. Their research also suggested that riparian vegetation had increased near channel margins since European settlement. This change may have been due to climate and land use changes, which influenced channel morphology and vegetation densities (Martin and Johnson, 1987).

A similar study by Johnson (1994) also indicated increased riparian vegetation in the Platte River system of Nebraska. Johnson examined historic woodland vegetation change using survey notes (1859 through 1881) and aerial photography. His calculations indicated that woodland vegetation encroached channel boundaries between the mid-1800s and early 1900s, narrowing stream width. Aerial photography indicated that these changes first occurred in upstream reaches, then advanced downstream (Johnson, 1994). Channel narrowing was thought to be associated with increased irrigation and dam construction, which decreased stream discharge (Johnson, 1994). According to Johnson (1994), the rate of narrowing declined around 1969 and widths in several reaches even increased during this time period as stream flow became stable.

Additional research in the Central Great Plains by Juracek (2002) examined historic channel change along Soldier Creek, a tributary to the Kansas River in northeast Kansas. Juracek analyzed streamflow and channel conditions at multiple gauging stations in the basin between 1936 and 2001. Channelization of the lower Solider Creek basin occurred around 1961 to improve the drainage of floodwaters entering the Kansas River following the 1951 flood. His results indicated widths in the lower basin were ~11 m greater than before 1956. The cause for channel change is thought to be related to the instability induced by channelization (Juracek, 2002). Similar channel adjustment occurred in the upper basin in the form of channel cutoffs, which increased stream gradient and shortened channel length. Changes in the upper basin may be linked to changes in land use and climate (Juracek, 2002).

Similar anthropogenic channel changes were investigated by VanLooy and Martin (2005), who examined channel and vegetation changes on the Cimarron River in the mid to late 1900s. This study re-examined Schumm and Lichty's (1963) interpretation of channel change in the basin. VanLooy and Martin suggested that the Cimarron River narrowed after the 1940s, which they attributed to a decrease in peak discharge. Lower discharge may be due to multiple factors, such as irrigation, earthen dams on tributaries, encroaching riparian vegetation, and the variability in long-term precipitation (VanLooy and Martin, 2005). Like Johnson (1994), VanLooy and Martin (2005) also suggested that an increase in riparian vegetation, irrigation, and dam construction could be associated with channel narrowing during the mid-1900s.

The most recent study of historic channel change on streams conducted in the Central Great Plains was conducted by Horn et al. (2012), who looked at adjustments in the Platte River Valley of central Nebraska. In the study, Horn et al. (2012) compiled aerial photography, survey maps, and discharge data to examine how the stream evolved during the historic period. They discovered that Platte River widths were on average 539% greater in 1858 than in 2006. This trend continued into the 1900s, where the channel decreased in area upwards of 46% between 1938 and 2006. Results from Horn et al. (2012) indicate that channel narrowing is occurring due to reduced discharge as a result of irrigation, which has caused mid-channel bars to attach to channel banks or disappear. In addition to this work, Horn et al. (2012) also identified that the Wood River, a tributary of the Platte River, experienced an increase in sinuosity after residing in the north-channel of the Platte River. Horn et al. (2012) suggested that this increase in sinuosity occurred because the Wood River has superimposed the meandering pattern of a prehistoric Platte River system.

2.2.3 Driftless Area of the Upper Midwest

This section describes historic channel change studies that have been done in the Driftless Area of southwestern Wisconsin. Early work in this area was conducted by Happ

(1944), who researched the effect of sedimentation on floods in the Kickapoo Valley. Resurveys of river cross sections were conducted to measure the average rate of aggradation. Happ (1944) suggested that ~0.3 m of alluvium was deposited within the floodplain. Causes for this rapid aggradation are thought to be associated with increases of agricultural development and associated deforestation (Happ, 1944).

Excessive sedimentation in stream valleys was also recorded along multiple other streams in the region. During an extensive study of channel change on the Platte River, a tributary of the Mississippi River in southwestern Wisconsin, Knox (1977) suggested that stream widths in the basin had increased since 1832-1833 in watersheds < ~60 mi^2 in size. In contrast, Knox (1977) indicated that larger, neighboring basins have recorded narrower stream widths during the same time period. These changes were closely associated with the impact of floods on sediment transport since settlement (Knox, 1977).

In a later study on the neighboring Galena River, Magilligan (1985) identified that anthropogenic effects of hill slope erosion had greatly exceeded the long-term geologic norm, producing high floodplain sedimentation rates. Magilligan (1985) proposed that the dramatic increase in the rate of sedimentation in the system was likely associated with land use changes in the mid-1800s.

Similarly, a study by Trimble (2009) discussed the fluvial processes, morphology, and sediment budget in the Coon Creek basin of southwestern Wisconsin. Historic documentation of channel adjustment in the basin has been collected since the mid-1800s to the 1990s, indicating that streams have been continually changing throughout time. Changes in land use/practices (e.g., deforestation, increases in agriculture, and pasturing) increased rates in the overall sediment budget (Trimble, 2009). A conservation effort in the early 1900s attempted to decrease the sediment yield of the system. Although these effects worked, they failed to decrease the rate of sediment that entered the Mississippi River (Trimble, 2009). Fluvial responses to ongoing changes in the basin have resulted in a wider more incised channel from that of pre-settlement.

A recent study by Lecce (2013) in the Blue River watershed, a tributary of the Wisconsin River, also recorded channel widening during the historic period. Lecce (2013) used both initial land survey notes (e.g., ~1830) and modern field measurements to monitor changes in stream width, power, and channel geometry. Results from Lecce (2013) indicated that both stream width and cross-sectional area increased substantially after settlement. This increase is associated with the channels enhanced stream power, which is due to logging and pasturing. As a result, stream power is ~3 times higher than prehistoric levels. Like previous studies done in the region (e.g., Happ, 1944; Knox, 1977; Magilligan, 1985; Trimble, 2009), Lecce (2013) also suggested that land disturbance was the main cause for these historic channel adjustments.

2.2.4 Atlantic Coast

In addition to historic channel change studies in the Driftless Area, similar studies have also been conducted near the Atlantic seaboard. The first of these studies was conducted by Ruhlman and Nutter (1999), who examined historic changes in stream channel morphology in the Oconee River basin of Georgia. Historic and modern discharge rates were calculated using Manning's equation and historical records. Once historic discharge rates were calculated, multiple resurveys were conducted to evaluate changes in channel morphology. Results indicated that channel expansion occurred throughout the basin (Ruhlman and Nutter, 1999). Extensive widening was found in the upper reaches of the study area where channel bank erosion has accelerated. This acceleration is in response to past land use practices, such as early land-clearing and cultivation practices (Ruhlman and Nutter, 1999).

A study by Galster et al. (2008) researched how changes in landuse (e.g., increased urbanization) can affect fluvial systems. The study first measured the impact of urbanization on channel widths using historic aerial photography and modern surveys. The study focused in the Lehigh Valley of Pennsylvania in the Little Lehigh and Sacony Creek watersheds. In these watersheds, land use has increasingly urbanized, which in turn, increased the percentage of impervious surfaces and the amount of runoff and peak discharge during storm events. Galster et al. (2008) associated this change in land use to the widening of stream width ~4 m in these watersheds since the 1940s.

Another study by McBride et al. (2010) analyzed riparian reforestation and channel change along several tributaries of the Sleepers River in northeastern Vermont. Historical channel and vegetation data from 1966 and 2004 were used to analyze changes to present channel dimensions. In many of these locations land use had transitioned from non-forested to forested. Comparisons showed that reforested stream reaches widened at a rate of ~4 cm/year between 1966 and 2004, while rapid widening occurred after 2004 in reforested areas at a rate of ~8.7 cm/year (McBride et al., 2010). The authors attribute this widening to a recovering and newly forested system that may continue to evolve in response to climate change or watershed changes (McBride et al., 2010).

A recent study conducted by Mossa (2013) examined historic channel adjustments on the Lower Old River, in central Louisiana. The Old River connects both the Atchafalaya and Red Rivers to the Mississippi River near the Louisiana and Mississippi border. Given the importance of these rivers to the regional economy, the Old River has been extensively documented since the 14th century (Mossa, 2013). The first documented channel change in the area was recorded in the 16th century, when the Mississippi River migrated across its floodplain, causing an episode of stream capture with the Red River (Mossa, 2013). This event caused the Atchafalaya River to change its course, now directly empting into the Mississippi River, instead of its prior debouching into the Red River. Later during the 1800s, humans altered the flow of the Old Man River (e.g., artificial cutoffs), in attempts to improve navigation through the river junction (Mossa, 2013). Results from Mossa (2013), suggested that together natural and anthropogenic changes have altered channel flow and sedimentation rates near the junction of these rivers (Mossa, 2013).

2.2.5 Pacific Coast

Historic channel change studies have also been conducted along the Pacific coast. These studies have mainly examined changes to the fluvial systems that were caused by mining operations during the late 1800s, which commonly induced a period of rapid sedimentation followed by channel incision. Some studies have demonstrated that channel incision is still occurring in many streams in the region. The first of these studies was conducted by White et al. (2010), who examined channel adjustments along the Timbuctoo Bend of the lower Yuba River in central California. The study used GIS to digitize a series of aerial images collected between 1937 and 2006 to identify channel changes. This method allowed the authors to investigate whether the locations of riffles and pools responded to changes caused by modern floods. Results indicated that the steam underwent a period of rapid incision between 1984 and 2002, with an estimated average incision rate of ~0.78 m (White et al. (2010). Lateral shifts in the channel also occurred with this incision. Despite these fluctuations in channel morphology (e.g., reservoir construction, incision, lateral migration), White et al. (2010) indicated that riffles persisted in the same locations.

Similar work on the lower Yuba River by Ghoshal et al. (2010) identified channel and floodplain changes over a 100 year period. This study used 1906 topographic maps and pre/post-flood aerial photography to document historic erosion and deposition from documented flood events. Ghoshal et al. (2010) identified similar observations to White et al. (2010) of lateral migration and channel incision. This incision in the lower Yuba River cut ~13 m into historic sediment, which in turn, increased the sediment load of the stream (Ghoshal et al., 2010). The associated rapid sedimentation of the system produced a braided appearance. After mining operations ceased during the early-1900s, the stream returned to a single-thread channel. Since this time, the system has tried to return to its pre-disturbance state through continued incision (Ghoshal et al., 2010). Although this study suggests the effects from mining are still prevalent in the system, a declining rate of incision may indicate the stream is nearing a quasi-equilibrium.

Recent work done by Wallick et al. (2011) investigated channel change and bedload transport in the Umpqua River basin in southwestern Oregon. Historic documents (e.g., aerial photographs, observations and accounts of channel conditions) from as far back as around 1900 were used to analyze changes in channel morphology and conditions. Like Ghoshal et al. (2010), mining during the mid-1800s increased rates of sedimentation in the basin, which produced numerous gravel point bars. Results from this study suggested that the overall platform of the system has remained stable, other than the variability of these gravel bars (Wallick et al., 2011). Between 1939 and 2005, the total reduction in gravel bar area was ~29% (Wallick et al., 2011). They suggest that these bars are naturally occurring features along bedrock rapids or large bends in the stream. The system uses the gravel bars as bed-material storage, contracting and increasing in size during flood events or periods of low flow.

2.2.6 Summary

These studies show that measurable channel changes have occurred in fluvial systems throughout much of the United States during the historic period. Several of these studies suggest that climate fluctuations have played a major role, specifically in the frequency of large flood events (e.g., Schumm and Lichty, 1963; Burkham, 1972; Knox, 1977; Martin and Johnson, 1987; Webb et al., 1991; Juracek, 2002; Dean and Schmidt, 2011; Swanson et al., 2010) that cause erosion and deposition in a river system.

Other research suggests that channel changes can occur because of modifications in the density of riparian vegetation (e.g., Graf, 1978; Johnson, 1994; Dominick and O'Neill, 1998; Grams and Schmidt, 2002; McBride et al., 2010). In the American Southwest and Central Great Plains, the invasion of tamarisk has stabilized channel banks, in turn narrowing many streams. Similarly, a number of studies (e.g., Bryan, 1928; Happ, 1944; Magilligan, 1985; Webb et al., 1991; Ruhlman and Nutter, 1999; Juracek, 2002; Galster et al., 2008; Trimble, 2009) have attributed changes in landcover and landuse to historic channel changes. These changes in landcover/landuse can affect sedimentation rates in the fluvial system, causing channel narrowing, incision, and/or sediment deposition.

A number of other studies (e.g., Van Steeter and Pitlick, 1998; VanLooy and Martin, 2005; Ghoshal et al., 2010; White et al., 2010; Wallick et al., 2011) suggest that anthropogenic effects, specifically the effects of mining and dams can cause channel change. Studies conducted by Ghoshal et al. (2010), White et al. (2010), and Wallick et al. (2011) indicated that mining has choked many streams in the western United States with sediment, causing changes in channel patterns. Other work conducted by Van Steeter and Pitlick (1998) and VanLooy and Martin (2005), claimed that dams have altered stream discharge and sediment loads.

2.3 Fluvial Studies in Michigan

The following discussion reviews the geomorphic studies that have been conducted on streams in Michigan. Although few of these studies focus specifically on fluvial geomorphology, they nevertheless have yielded some information about stream processes and landform evolution in the state. The following section outlines the nature of this research and is subdivided geographically by Michigan's Upper and Lower Peninsula.

2.3.1 Michigan's Upper Peninsula

Streams in the Upper Peninsula of Michigan are distinctive due to the regions highly variable surficial geology and geomorphology. The western Upper Peninsula is mainly dominated by Precambrian igneous and metamorphic bedrock with overlying glacial and lake-plain sediments in places. A study by Hack (1965) examined the postglacial drainage evolution and stream geometry of river systems near Ontonagon, Michigan. Rivers in this area flow on a steep glaciolacustrine surface, underlain by a bedrock-dominated landscape, which has resulted in an increase in local stream gradients and a parallel channel pattern. Hack (1965) indicated that the trellised pattern may be due to surface irregularities, such as glacial groves or flutes near the channel boundary. Hack (1965) further suggested that the current stream system most likely formed as Glacial Lake Duluth withdrew from the region ~9 ka ago. He also graphed valley depth to drainage area to infer where channel erosion is occurring. Results suggested that headward erosion of this stream from its baselevel (Lake Superior), is minimal compared to the gradual incision of the stream on the lake plain (Hack, 1965).

Following the Hack (1965) study, the next work conducted in the Upper Peninsula was by VanDusen et al. (2005) on the Otter River, a tributary of the Sturgeon River in northwestern Michigan. VanDusen et al. (2005) examined the role that logging had on brook trout and macroinvertebrate habitat. The study located nine first and second order stream segments that had been logged within the last 2-30 years. VanDusen et al. (2005) then measured total dissolved solids, pH, and temperature in these streams during the summer of 2000. Results indicated that parcels logged in the previous 10 years increased the fine sediment content of nearby stream segments. This change is due to an increase in runoff associated with newly logged surfaces (VanDusen et al., 2005). As a result, the number of brook trout is thought to be low in these river segments for about ten years after a logging event (VanDusen et al., 2005).

Related work by Morris et al. (2010) analyzed the distribution of wood jams on six streams (e.g., Scott Creek, and the Little Carp, Big Carp, Upper Carp, Union, and Little Iron Rivers) in the Porcupine Mountains of northwestern Michigan. The purpose of the study was to determine if stream valley geomorphology and forest age had an effect on the distribution of wood jams in a given stretch of a stream. Each river segment studied had its valley gradient, sinuosity, width, and channel bedding characterized to provide an overall regional assessment. From this examination the authors identified four major stream segment types; 1) low gradient, with a high sinuosity and low channel width, 2) high gradient, with a low sinuosity and low channel width, 3) medium gradient, with a high sinuosity and medium channel width, and 4) low gradient with a medium-high sinuosity and medium-high channel width (Morris et al., 2010). Results indicated that the distribution of wood jams were high in segment types 1 and 3, which suggested that river morphology (e.g., high sinuosity) and interactions between streams and riparian vegetation were important factors for the distribution of such jams in the river system.

The most recent study of a stream in the Upper Peninsula of Michigan was conducted by Goebel et al. (2012), who examined the influence that variations in flood frequency had on riparian vegetation and fluvial landforms. The research was conducted in the Little Carp River watershed, located in the Porcupine Mountains Wilderness State Park. Goebel et al. (2012) used three different stream reaches within the watershed to estimate the effect that floodwaters had on vegetation. These reaches represented stream segments with a high gradient that are deeply entrenched, high gradients that are moderately entrenched, and low gradients that are poorly entrenched. Results indicated that nearly half of the riparian vegetation and ground shifts of channel boundaries occur in relatively low gradient streams that are somewhat entrenched. Goebel et al. (2012) found that most of these shifts were mainly induced by infrequent flooding events. Similarly, high gradient stream sections also recorded vegetation shifts mostly due to infrequent flooding (Goebel et al., 2012). In contrast, frequent flooding events tended to effect riparian vegetation in streams that are deeply entrenched (Goebel et al., 2012). This research suggests the "balance between physical and ecological processes" is likely controlled by the physical character of a streams valley (Goebel et al., 2012, pp 690).

2.3.2 Michigan's Lower Peninsula

In contrast to many streams in the Upper Peninsula, stream systems in the Lower Peninsula of Michigan typically flow on uniform glacial sediment and are therefore commonly dendritic. This relationship occurs because these streams flow on thick glacial sediments and are less dominated by exposed bedrock structures, which is reflected in the drainage patterns between the areas. For example, Hack (1965) recognized that streams in the western Upper Peninsula have a parallel pattern, which he attributed to bedrock geology, sediment composition, and channel gradient.

Several streams in the Lower Peninsula have been studied to examine fluvial adjustments since European settlement. The first of these studies was conducted by Bowman (1904), who identified a case of stream capture of Oak Ravine, a tributary of Willow Run, which joined the Huron River in southeastern Michigan. The two small tributaries in the study area were less entrenched than the Huron River. Lateral migration of the Huron River on its floodplain undercut the bluff near the middle reach of Oak Ravine. In time, stream capture of Oak Ravine occurred, resulting in a steepened gradient for the stretch of Oak Ravine flowing down the bluff. In addition, Bowman (1904) attributed the stream capture of Oak Ravine from Willow Run to the Huron River to the persistent bank erosion of the Huron River against the perched bluff.

The next study in the Lower Peninsula was conducted by Hansen (1971), who analyzed stream sedimentation on the lower Pine River, a tributary of the Manistee River in northwestern Lower Michigan. Hansen analyzed a 41.8 km (26-mile) section of the lower Pine River above the Stronach Dam, which has since been removed. In the study, Hansen (1971) examined the rate of sedimentation between 1967 and 1969, while also identifying fluvial landforms (e.g., terraces). Results indicated that 55% of the streams sediment load came from channel bank erosion, delivering ~70,000 tons of sediment to the river during the three year study (Hansen, 1971). Around 70% of the sediment was sand sized or greater. The source for this sediment is thought to have come from 204 eroding banks that were identified throughout the meandering stretch of river. Hansen (1971) indicated that this was not the first time that an increase of sediment load was recorded on the Pine River. This first occurred when the Stronach Dam completely filled with sediment between 1912 and 1953, which forced Consumers Power Company to terminate power generation (Hansen, 1971; Consumers Power Company, 1994). This excessive sedimentation is suggested to be associated with increased runoff and bank erosion after Michigan's logging era. In addition, Hansen (1971) also identified multiple terraces (e.g., ~3.3 m., ~5.1 m, and ~12.1 m above the current floodplain) all of which had stream gravels overlying glacial drift at their surface.

A later study by Rieck and Winters (1979) was conducted in the south-central part of the Lower Peninsula. The purpose of the study was to determine the main factors that affected the location and spatial trends of stream courses (Rieck and Winters, 1979). To meet the study goals in this part of the stream, Rieck and Winters also analyzed well records to determine the thickness of glacial drift over bedrock and to map the bedrock surface topography. These records assisted in the interpretation of past drainage patterns before the last ice age.

During the retreat of the Laurentide Ice Sheet, the bedrock surface was buried by an average of ~19.8 m of drift in this part of Michigan (Rieck and Winters, 1979). This glacial drift filled depressions, which in turn, lowered the overall relief of Lower Michigan. Results from Rieck and Winters (1979) suggested that fluvial erosion alone since the last glaciation cannot account for current drainage patterns. Instead, they suggested that large river valleys, like those found in third-order streams and larger, were likely glacial spillways flowing in large bedrock valleys (Rieck and Winters, 1979). Interestingly, Rieck and

Winters revealed that many well records documented organic deposits (e.g., peat, muck, marl, and reeds) within or directly above bedrock valleys. These well locations today are in close proximity to topographic depressions or stream valleys, suggesting that current drainage patterns tend to mimic past drainage patterns and bedrock relief.

A study conducted a few years later by Strayer (1983) investigated the surface geology and size of streams in southeastern Michigan. The purpose of the study was to determine to what extent surface geology and stream size had on freshwater mussel (*unionid mussel)* distributions in four rivers (e.g., Clinton, Rouge, Huron, and Raisin Rivers). The headwaters of these rivers are in high morainic uplands that have high soil infiltration, allowing for little surface runoff and steady baseflow throughout the year (Stayer, 1983). The downstream margin of these streams flow upon a clay-rich lake plain of Glacial Lake Maumee. Results from the study indicated that stream size wasn't the dominant control on mussel populations, but rather complex effects on surface geology (e.g., hydrology, slope, and turbidity; Strayer, 1983).

A later study by Baker and Barnes (1998) investigated the diversity of landscape ecosystems of multiple floodplains in northern Lower Michigan. In the area, 22 river valley transects were taken across nine rivers (e.g., Cedar Creek, Baldwin, Big Sable, Little Manistee, Little Muskegon, Manistee, Pere Marquette, Pine, and White Rivers) in the Manistee National Forest. The authors used aerial photography, surfical geology maps, and field reconnaissance to analyze landscapes. Using this approach, Baker and Barnes (1998) reported that much of the region was comprised of non-pitted outwash, which is commonly found on or near the toe slope of moraines. Streams (e.g., segments of the Manistee, Pere Marquette, and Baldwin Rivers) that flow upon pitted outwash surfaces often have incised until encountering glacial till (Baker and Barnes, 1998). Baker and Barnes (1998) suggested that stream valleys in non-pitted outwash had a noticeably smaller mean depth and a higher width-to-depth ratios than stream valleys in moraines. In addition to these findings, Baker and Barnes (1998) also associated the presence of meander scars in northern Lower Michigan to a stream's valley wall morphology (Baker and Barnes, 1998).

Burroughs et al. (2009) also worked in northwestern Lower Michigan on the lower Pine River to investigate the effects of the Stronach Dam removal. Burroughs et al. (2009) surveyed 31 "permanent" cross-sectional stream transects that were assessed annually between 1996 and 2003. This assessment allowed them to analyze channel adjustments during and after dam removal. Results from Burroughs et al. (2009) suggested that the ongoing headward cutting of sediments had occurred in the former dammed pond after the dam was removed. This incision extended ~3.9 km upstream from the former location of the dam and is estimated to have removed ~92,000 m^3 of sediment in the 10 years after the dam was removed (Burroughs et al., 2009).

Another study in 2009 by Rachol and Boley-Morse attempted to delineate hydraulic geometry curves to estimate bankfull discharge for Michigan streams. Initially, 343 USGS gauging stations were analyzed to produce these curves. After further analysis, Rachol and Boley-Morse decided to reduce the number of gauges in the dataset to 44. Many gauges were ommited from the study because of anthropogenic features, such as dams and road crossings that effected stream flow. As a result, Rachol and Boley-Morse (2009) narrowed the study area to southern Lower Michigan, where the majority of unaffected gauge stations were located. Rachol and Boley-Morse (2009) then surveyed these gauge stations to estimate bankfull discharge and channel gradient. This information was used to calculate flood-reoccurrence intervals, which suggest that bankfull discharges occur more frequently than every two years (Rachol and Boley-Morse, 2009).

The most recent study in Michigan was conducted by Webb-Sullivan and Evans (2015) on the Ottawa River, a tributary of Lake Erie in northwestern Ohio and southeastern Michigan. In the study, the authors collected 14 vibracores and 52 push cores to examine the alluvial stratigraphy of the river valley. Webb-Sullivan and Evans (2015) also collected 4 ^{14}C and 6 OSL dates, allowing them to classify the geochronology of the stream into 5 stages: (1) a meandering point-bar sequence that formed ~5,000 years BP, (2) an organic rich layer dated between 5,000 and 200 years BP, (3) a mineral rich layer associated with land-clearing and agriculture between the early 1800s and the early-1960s, (4) a 1.7 m fill-terrace that formed as early as the 1950s due to rapid urbanization, and (5) a reduction of sediment input starting in the 1980s due to no-till agriculture and revegetation. These early river stages suggest that the Ottawa River was influenced by the higher Nipissing level of Lake Erie during the mid-Holocene (Webb-Sullivan and Evans, 2015). As a result, thick organic-rich sediments were deposited throughout the basin in the "Great Black Swamp" that formed due to higher lake levels associated with proglacial Lake Maumee, which formed ~14 ka ago (Webb-Sullivan and Evans, 2015). Later river stages indicate that human involvement (e.g., land-clearing, agriculture, and urbanization) within the basin had a dramatic effect on the sedimentation rates of the Ottawa River (Webb-Sullivan and Evans, 2015).

CHAPTER 3

STUDY AREA

This study focuses on the Muskegon River, which is located in the north-central part of Michigan and is the second largest river system in Michigan. It originates in the north-central region of Lower Michigan at Houghton Lake and drains an area of 7,057 km^2 as it flows southwest for ~220 km until it debouches into Muskegon Lake (Figure 3.1; Ray et al., 2012). Average annual discharge at Houghton Lake is about 54 m^3/sec, whereas it is ~380 m^3/sec at the city of Muskegon (Institute of Water Research, 2006). The basin is long and narrow, with a length of ~210 km and width as narrow as ~25 km, respectfully. Relief in the basin ranges from nearly level in the upper and lower portions, whereas the middle part consists of hilly terrain.

3.1 Glacial History and Geomorphology

The origin of the Muskegon River is likely associated with the retreat of the Lake Michigan and Saginaw Lobes, of the Laurentide Ice Sheet, from Michigan during the Late Pleistocene (Kehew et al., 2011). The melting of these lobes produced large quantities of meltwater, which transported sandy outwash up to tens of meters thick in front of the glacial margin (Schaetzl et al., 2013). The earliest ancestral stream related to the Muskegon River may have been a braided stream associated with the formation of the St. Helen outwash plain ~16 ka ago (Arbogast et al., 2008). This stream was likely fed by meltwater from surrounding ice margins, which would have continually added sediment to the system. The current course of the Muskegon River likely developed during this time as the

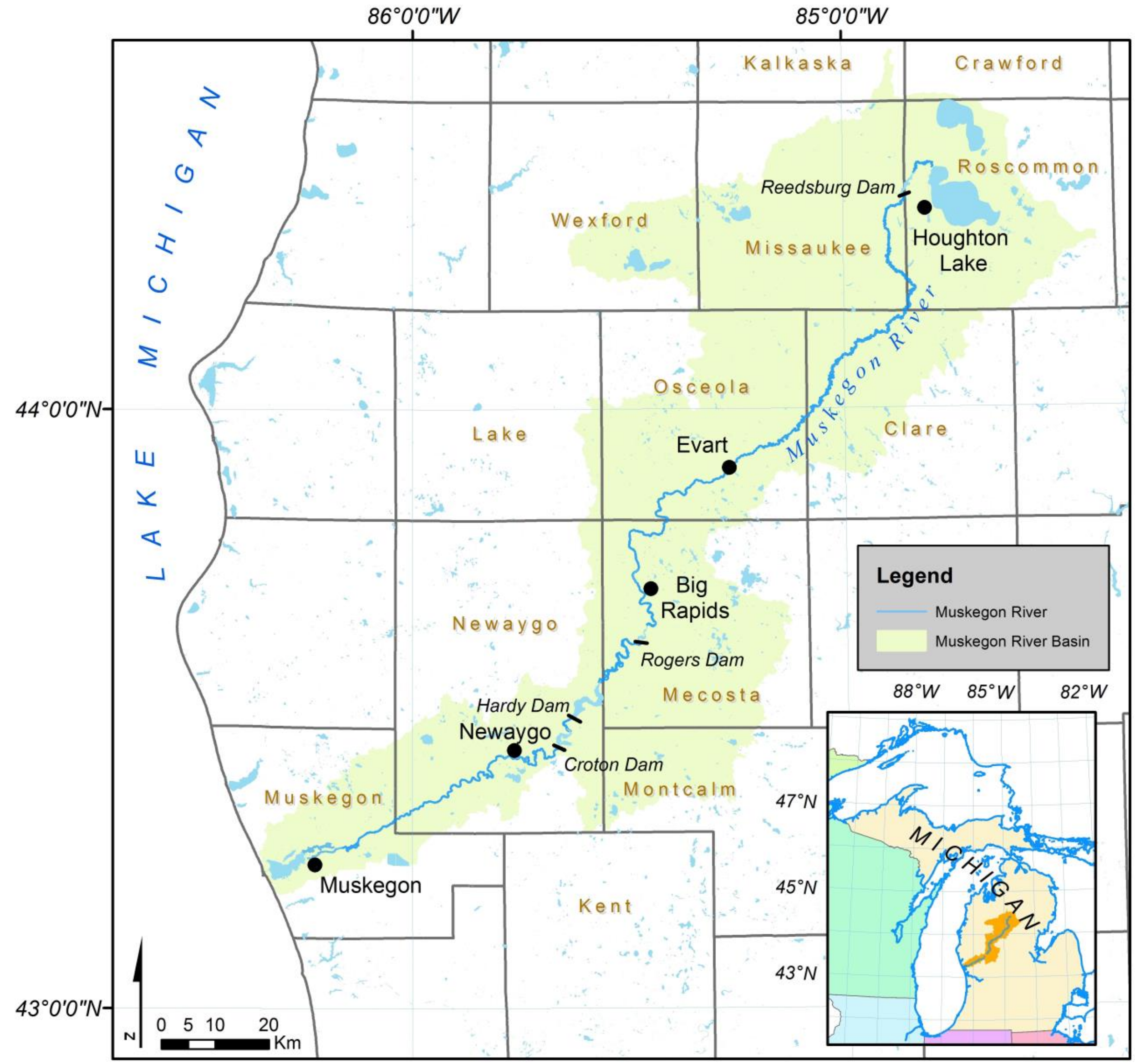

Figure 3.1. The Muskegon River basin in north-central Lower Michigan.

topographically high Lake Border and West Branch Moraines, which formed between 17.1 and 15 ka ago (Larson and Scheatzl, 2001; Blewett et al., 2009), influenced the draining and physical characteristics of the early meltwater channel (Figure 3.2).Arbogast et al. (2008) hypothesized that the ancestral Muskegon River transitioned from a braided to a meandering stream sometime between ~13 and 12.5 ka ago, as continued deglaciation caused sediment yields to decrease. This date supports the hypotheses of Blewett and Winters (1995) and Schaetzl and Forman (2008) that streams in northern Lower Michigan were likely braided near the glacial margin during the last ice age.

The first geomorphic study of the Muskegon River was conducted by Arbogast et al. (2008) who mapped landforms and collected ^{14}C ages from peat-filled paleomeanders in the upper half of the basin. Results from Arbogast et al. (2008) indicate that the Upper Muskegon River valley contains four alluvial terraces and numerous paleomeanders. In an effort to estimate the age of these landforms, Arbogast et al. (2008) collected basal peat samples in abandoned oxbows for ^{14}C dating. The reconstructed chronology suggests that a paired T-4 terrace formed on an outwash/lacustrine surface ~12 thousand cal yrs ago (Arbogast et al., 2008), suggesting the T-4 surface formed as the Muskegon River downcut ~2 m into the outwash/lacustrine surface in association with a large meandering channel. The dimensions of macromeanders in Arbogast et al. (2008) were compared to similar ancestral channels in Europe and the U.S. Atlantic Coastal Plain. Given the general geometric consistency of the macromeanders, Arbogast et al. (2008) estimated that the mean annual discharge of the ancestral Muskegon River was ~400 m^3/sec, which was ~8 times greater than the modern river (60-85 m^3/sec; Arbogast et al., 2008).

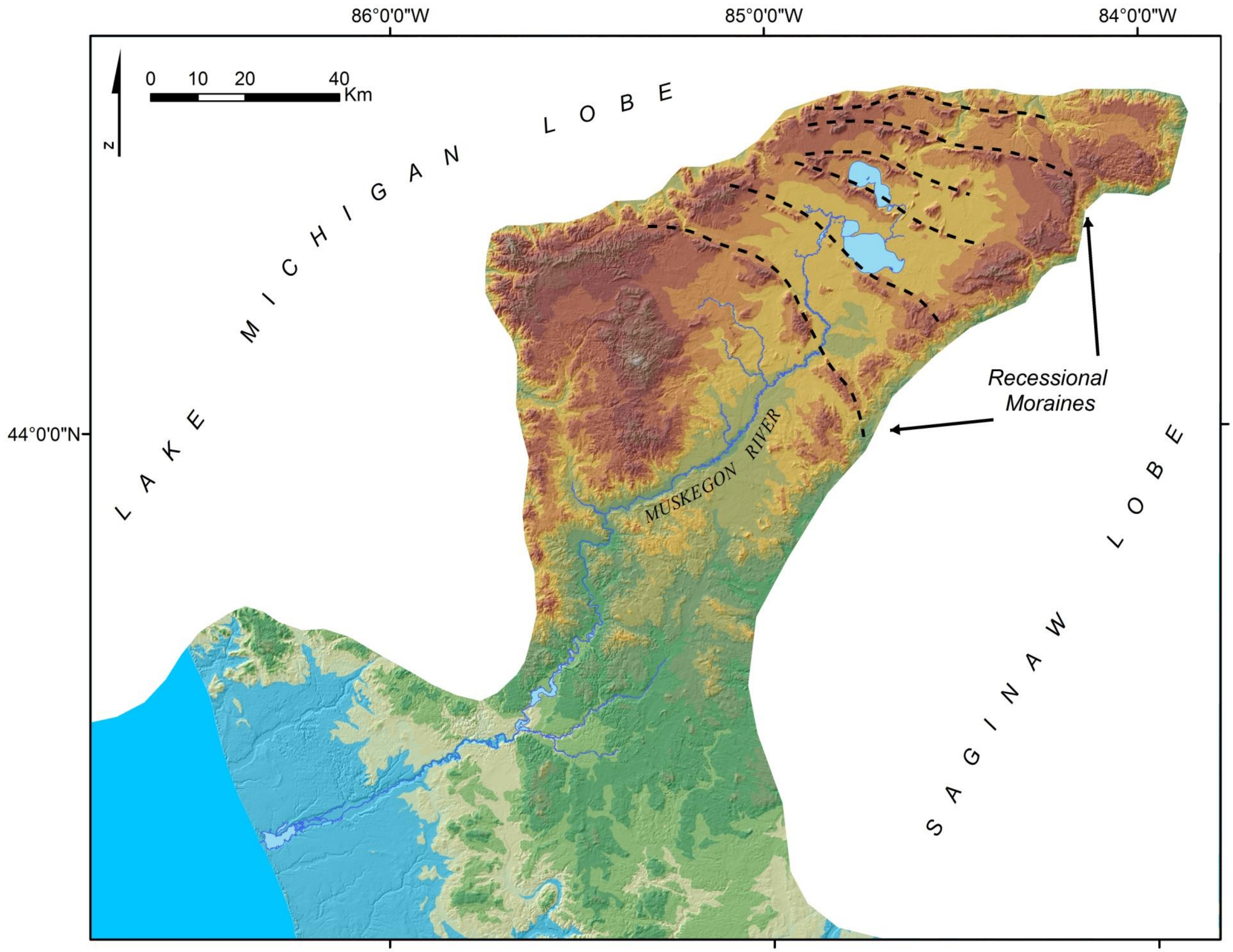

Figure 3.2. The Muskegon River basin as the Lake Michigan and Saginaw Lobes retreated from the region ~13,000 yrs ago (modified from Kehew et al., 2011).

During the Early Holocene, a warming trend likely caused the Muskegon River to reduce in size, likely reaching its current dimensions (Arbogast et al., 2008). During this period, the upper Muskegon River downcut ~6 m into the T-4 terrace forming an escarpment and the T-3 terrace, which is a fill-strath surface (Arbogast et al., 2008). Arbogast et al. (2008) suggests the T-3 terrace was cut between ~12 ka and possibly ~9.5 ka ago. The Muskegon River likely formed its T-3 surface during the Late Pleistocene-Holocene boundary, when the Great Lakes region experienced a warmer and wetter climate.

A cool and moist climate during the mid-to late-Holocene likely caused the Muskegon River to incise yet again, forming a T-2 terrace in its upper system (Arbogast et al., 2008). In contrast to the T-3 surface, numerous paleomeanders were found on the T-2 surface, allowing Arbogast et al. (2008) to measure several widths, radius of curvatures, and meander wavelengths. Results from the study suggest that discharge during this time was similar to the modern Muskegon River. According to ^{14}C dates from paleomanders in the inner valley, the T-2 surface likely formed sometime between 6.3 and 3.7 thousand cal yrs BP. Today the upper Muskegon River flows on the youngest surface of the T-1 terrace complex, which formed between ~2.5 ka and 500 years ago in an increasingly variable climate (Arbogast et al., 2008).

3.2 Climate

The climate of the study area is strongly affected by the close proximity of the upper Great Lakes. Winter usually lasts from December to March with an average temperature of -6.9^{o}C at Houghton Lake and -1.4^{o} C at Muskegon. The summer growing season typically occurs from April to September, during which time the average high temperature of Houghton Lake is 21.2° C and 21.9° C at Muskegon. Total annual precipitation at Houghton Lake is 73.2 cm and 85.6 cm in Muskegon. Most precipitation in the study area comes from the frontal boundaries of mid-latitude cyclones, which produces over half of the precipitation in Michigan (Heideman and Fritsch, 1988). Lake effect snow commonly occurs in the winter, with larger snow accumulations occurring closer to Lake Michigan than at inland locations. Monthly temperature and precipitation values for the cities of Muskegon, Big Rapids, and Houghton Lake, represents the conditions in the upper, central, and lower Muskegon River basin, respectively (Table 3.1).

3.3 Soils

Soils within the Muskegon River basin are classified as Spodosols, Entisols, Alfisols, and Histosols. The location of these soil orders typically varies depending on landscape position, climate, and parent material. Spodosols are commonly found in the upper basin formed in glacial outwash deposits (Corder et al., 1979; Frederick, 1985; Tardy et al., 2005). These soils are typically sandy and facilitate the infiltration of water throughout the soil profile (Schaetzl and Isard, 1996; Schaetzl and Anderson, 2005). Similarly, Entisols are also in the upper and central portions of the basin and are typically found in strongly sloping sandy soils (e.g., moraines, side slopes of river valleys) or areas where frequent fires have inhibited soil development (Mokma and Vance, 1989). Alfisols are commonly found in the central basin, forming in moderately sloping glacial outwash and till deposits (Mettert, 1969; Corder, 1984; Purkey, 1995). These soils tend to support more deciduous

Table 3.1. Temperature and Precipitation at Muskegon, Big Rapids, and Houghton Lake, MI.												
	J	**F**	**M**	**A**	**M**	**J**	**J**	**A**	**S**	**O**	**N**	**D**
Muskegon												
Mean Daily Max. (C^o)	-0.66	0.66	6.11	13.27	19.38	24.44	26.88	26.0	21.83	14.88	8.0	1.72
Mean Daily Min. (C^o)	-6.61	-6.05	-2.77	2.94	8.27	13.66	16.61	16.11	11.55	5.83	1.11	-3.72
Precipitation (mm)	51.56	46.48	57.15	73.91	82.55	64.77	60.19	86.1	98.8	78.99	85.34	64.77
Big Rapids												
Mean Daily Max. (C^o)	-2.11	-0.11	5.44	13.11	19.77	25.0	27.33	25.88	21.5	14.22	6.55	0.11
Mean Daily Min. (C^o)	-11.44	-11.11	-7.16	-0.66	5.94	11.27	13.83	12.5	7.77	1.77	-2.88	-8.05
Precipitation (mm)	53.59	43.68	61.72	83.31	87.63	82.8	85.09	102.61	100.33	86.36	81.28	61.21
Houghton Lake												
Mean Daily Max. (C^o)	-3.22	-1.44	4.27	12.27	19.05	24.16	26.5	25.05	20.5	13.16	5.7	-0.83
Mean Daily Min. (C^o)	-12.27	-11.88	-7.5	-0.55	5.27	10.11	12.5	11.55	7.5	2.16	-2.55	-8.16
Precipitation (mm)	38.35	30.73	46.73	62.99	71.62	78.74	70.1	86.36	78.74	65.02	58.92	41.91

Source: Midwestern Regional Climate Center vegetation, aiding in the thickening of organics within the profile (Schaetzl and Anderson, 2005). As a result, these soils are the most productive for agriculture in the basin.

Poorly drained Histic soils are located in the upper basin near the headwaters at Houghton Lake (Tardy et al., 2005). These soils are also found in the anastomosing portion of the Muskegon River before the Muskegon River debouches into Muskegon Lake (Pregitzer, 1968). Typically, these soils are common on the relatively flat surface of the current floodplain, where plant matter accumulates and decomposes, creating a thick organic horizon (Schaetzl and Anderson, 2005).

3.4 Pre-Settlement Vegetation

As the Laurentide Ice Sheet retreated from northern Lower Michigan ~14 ka, flora soon became established on the previously glaciated landscape (Yansa and Adams, 2012). Initially, this freshly deglaciated landscape was likely a tundra environment, with annual temperatures ~5-10^{o} C below modern day (Jackson et al., 2000). This tundra environment hugged the retreating glacial margin, allowing a tundra-wetland biome transition to occur as the Laurentide Ice Sheet retreated northward (Holman, 2001; Agenbroad, 2005; Saunders et al., 2010; Yansa and Adams, 2012).

During this transition, uplands became covered with grasses and scattered tree species, such as white spruce (Picea glauca) and black spruce (Fraxinus nigra; Yansa and Adams, 2012). In contrast, lowlands were inhabited by sedges (e.g., Carex spp., Cladium spp.) and aquatic species (Yansa and Adams, 2012). Like its neighboring tundra environment, this wetland biome was time-transgressive (Yansa, 2006) and likely lasted until ~10 ka BP in Lower Michigan (Hupy and Yansa, 2009).

In the early Holocene, broadleaf deciduous and coniferous forests began to dominate northern Lower Michigan (Hupy and Yansa, 2009; Lovis, 2012). During this time period the climate became warmer, allowing species such as birch (likely <u>Betula papyrifera</u>), ironwood (<u>Ostrya</u> or <u>Carpinus</u>), elm (<u>Ulmus</u>), jack pine (<u>Pinus banksiana</u>/<u>resinosa</u>), and white pine (<u>Pinus strobus</u>) to flourish (Hupy and Yansa, 2009). These forest communities slightly decreased in northern Lower Michigan ~7 ka ago, when a warmer and more arid period began (e.g., Altithermal; Kutzbach et al., 1998). This period encouraged mixed oak (<u>Quercus</u> spp.) and hickory (<u>Carya</u> spp.) forests to increase in numbers (Hupy and Yansa, 2009).

During the late Holocene (in northern Lower Michigan) new tree species, such as maple (<u>Acer</u> spp.), beech (<u>Fagus grandifolia</u>),white pine, tamarack (<u>Larix laricina</u>), and eastern hemlock (<u>Tsuga canadensis)</u> became established (Michigan Forests Forever, 2006; Hupy and Yansa, 2009), due to a cooler climate with increased precipitation (Webb et al., 2004). The extent and characteristics of these species were noted in the initial survey of northern Michigan.

<u>3.5 Cultural History</u>

The cultural history of the region began ~12 ka ago with early Paleoindians (i.e., Clovis people; Shott and Wright, 1999; Lovis, 2009), who were hunter-gatherers and lived in adverse tundra and boreal environments (Lovis, 2009). During the early Holocene, these indigenous peoples formed into several local cultures. Evidence for this separation is linked to the variability of projectile point styles found during this time period, each of which are thought to be associated with a separate cultural group (Haynes, 2002).

Archaic groups continued to hunt and gather (Robertson et al., 1999) while also adapting to modern forest conditions (Lovis, 2009). These groups continued to adapt in the early part of the Late Holocene through the introduction of horticulture practices after ~4 ka ago (e.g., farming maize, squash, and beans). This subsistence strategy allowed Archaic people to sustain larger villages (Lovis, 2009). Mound building and ceramic use began ~2 ka, adding to the complexity of these groups (i.e., the Woodland people; Lovis, 2009) and initiating the Woodland Period. The Woodland people continued to integrate and formed egalitarian social systems, becoming the Ojibway (Chippewa), Odawa (Ottawa), and Potawatomi (Cornell, 2009; Lovis, 2009) that were encountered by Europeans during the early historic period.

The main inhabitants of the Muskegon River basin during the 16th-17th century were the Odawa people, who used the waters to hunt, fish, and travel (Cornell, 2009). The French and British were the first European inhabitants to arrive in Michigan during the 16th century. The Odawa had close relations with the French, who allied with them during most of the fur trade (Cornell, 2009). During the late 16th century, the Iroquois of present day New York and Ontario, invaded Michigan in an effort to control the region's fur trade for British interests (Cornell, 2009). The British defeated the French in the French and Indian Wars during the mid-17th century, allowing the British to take control of the Great Lakes region.

Present day Michigan became part of the United States shortly after the British lost the American Revolutionary War. Michigan experienced increased settlement of Europeans/Americans after the passing of the Northwest Ordinance of 1787, which allowed war veterans from Massachusetts to buy land in present day Lower Michigan

(Cornell, 2009). Later, the Treaty at Washington in 1836 ceded Ojibway and Odawa lands in the northern Lower Peninsula and Upper Peninsula of Michigan to the Territory of Michigan (LeBeau, 2005). The treaty allocated a large portion of land to the territory, allowing the Michigan Territory to reach statehood in 1837. The Treaty of Washington in 1836 allowed the initial survey of the region to occur, while also permitting an increase of white inhabitants in the region.

One of the impacts of white settlement in Michigan was mass deforestation due to logging. The Muskegon River basin was one of the most heavily logged areas in Michigan and was extensively clear cut between 1837 and 1910 (Alexander, 2006). The rate of logging picked up dramatically in the late-1800s as the large white pine stands of northern Lower Michigan were cut to meet the lumber demand of Chicago, Illinois after the Great Chicago Fire in 1871 (Alexander, 2006). Lumber mills were built in towns such as Big Rapids, Cadillac, Muskegon, Manistee, and Newaygo, and processed most of the logs in western Lower Michigan (Alexander, 2006). Dates of logging and practices differed between the upper and lower Muskegon River in the late-1800s. In the mid-1800s, the lower Muskegon River basin was heavily logged between Big Rapids and Muskegon, where the wider river allowed the efficient transportation of logs to mills in the lower basin (Figure 3.3). In the 1880s, logging railroads became a versatile and an inexpensive way to transport logs. These rail systems allowed lumber located in the upper Muskegon River basin to be more easily transported to mills, such as Cadillac, which were closer than the mills along Lake Michigan (Allen and Titus, 1941; Alexander, 2006). This innovation accelerated the clearing of inland pine stands of Michigan.

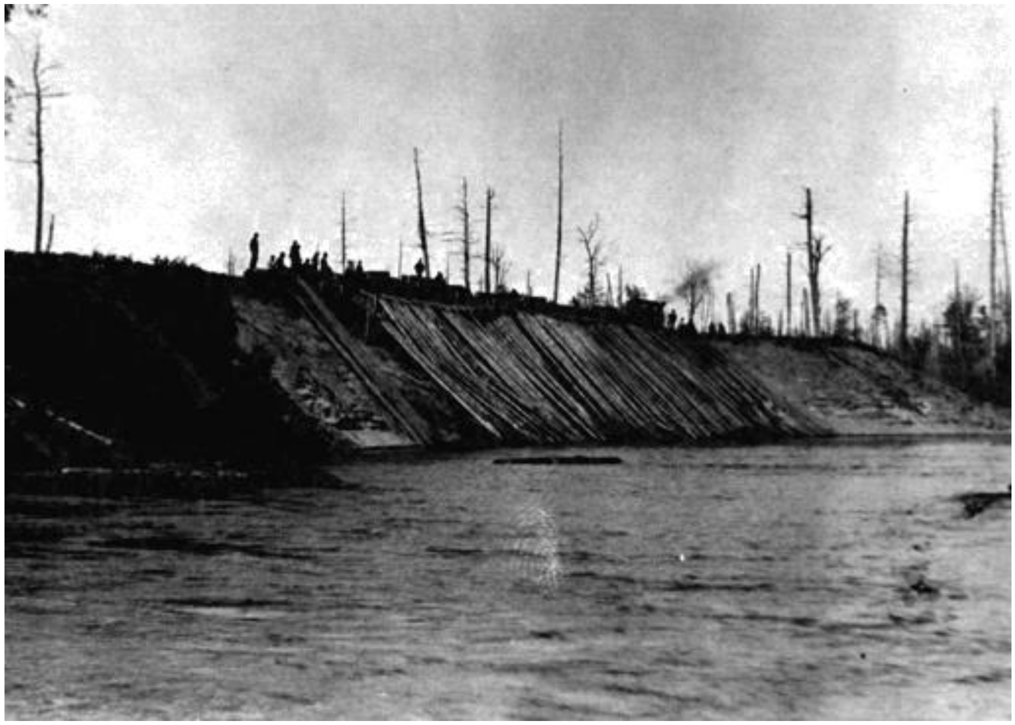

Figure 3.3. A logging rollway in the lower Muskegon River valley during Michigan's logging era.

The clearing of forested lands in the Muskegon basin lead to several surges of agriculture in the early 1900s, 1940s, and 1960s (Ray et al., 2012). Today, the Muskegon River basin is primarily forested in the upper basin, whereas agriculture dominates in the lower basin (Ray et al., 2012). Current land-use data for the Muskegon River basin indicates that most of the basin is occupied by forests (over 50%) and agriculture (over 20%), while the remainder of the basin is composed of wetlands, open water, scrublands, and urban uses (Ray et al., 2012). Tree species such as American aspen (Populus grandidentata), red maple (Acer rubrum), and paper birch (Betula papyrifera) have increased in numbers since Michigan's logging era, likely due to the reduction of coniferous vegetation (Michigan Forests Forever, 2006).

Along with landcover/landuse changes during the early historic period, a number of earthen dams were also constructed on large river systems throughout northern Lower Michigan, such as the Muskegon, Manistee, Pine, Au Sable, and Grand Rivers to produce hydroelectric power. Although this technology provided the area with a renewable power source, these structures extensively altered the fluvial system with respect to channel width, flow, and sedimentation rates (Hansen, 1971; Consumers Power Company, 1994). Many of these dams are still in use today (Figure 3.1).

CHAPTER 4

METHODOLOGY

A generally accepted methodology for assessing historic channel change has evolved from past studies in the United States, which includes a variety of preparatory and fieldwork procedures. For example, many studies (e.g., Schumm and Lichty, 1963; Burkham, 1972; Knox, 1977; Graf, 1978; Martin and Johnson, 1987; Johnson, 1994; Van Steeter and Pitlick, 1998; VanLooy and Martin, 2005; Swanson et al., 2010; Horn et al., 2012; Lecce, 2013) have used survey notes to examine pre-settlement stream conditions. In some circumstances, these locations were found and resurveyed to detect changes in channel width. Since the initial surveys, other studies (e.g., VanLooy and Martin, 2005; Hughes et al., 2006; Buckingham and Whitney, 2007; Galster et al., 2008; Ghoshal et al., 2010, White et al., 2010; Swanson et al., 2010; Martin and Pavlowsky, 2011; Skorko et al., 2011; Wallick et al., 2011; Horn et al., 2012; James et al., 2012; Lichter and Klein, 2012) have used aerial imagery and GIS to calculate changes in channel width, examine changes in morphology, or interpret changes in landuse/landcover. The following discussion is a summary of the procedures that were used in this study.

4.1 Preparatory Work

Early in the project, original survey maps/notes and aerial photography of the study area were obtained. Original survey information was acquired from the State Archives of Michigan and the Michigan Department of Natural Resources. These documents represent ground conditions and measurements of the stream between 1836 and 1857. Aerial photography from 1938 was also acquired from the Remote Sensing & GIS Aerial Imagery Achieve in the Department of Geography at Michigan State University.

The survey maps and aerial imagery were georectified using ESRI - ArcGIS 10 software to match the orthorectified 2012 aerial imagery obtained from the USDA's National Agriculture Imagery Program (NAIP). During the rectification process, the Root Mean Squared Error (RMSE) was kept as low as possible (in all cases < 3.0 m). Because this study did not calculate stream widths using a rectified image, such as VanLooy and Martin (2005), total RMSE was not recorded.

Once the survey maps and aerial photography were georectified, image interpretation was conducted to identify changes in channel position, such as abandoned meanders, extended point bars, or changes in meander pattern. After such changes were digitized, meander geometry (e.g., sinuosity, gradient, radius of curvature, amplitude, and meander wavelength) was calculated for the entire length of the Muskegon River. To accomplish this task, a channel center line was generated by hand in ArcMap and divided into 142 regularly spaced segments for every river valley mile. For each river segment, sinuosity was calculated by dividing stream miles by river valley miles. Slope was also calculated for each river segment by using the USGS NED 10 m Digital Elevation Model (DEM) of Lower Michigan. Other attributes, such as radius of curvature, meander wavelength, and amplitude were measured for each meander bend, providing valuable information about the environment and conditions in which meanders have formed.

In addition to characterizing the channel pattern of the Muskegon River, changes in channel width were also calculated along section lines. The use of section lines at regularly spaced intervals has been a common approach for measuring historic changes in width

(Knox, 1977; VanLooy and Martin, 2005). This method produces multiple data points and effectively analyzes a river system. Section line locations are commonly located in the field by using roads and property lines (Knox, 1977). The section line sampling method was chosen for this study to obtain regularly spaced data and also to simplify the recognition of sampling locations while in the field. Specifically, a total of 151 sampling locations were selected for the study from the Muskegon River's headwaters to Muskegon Lake.

The first measurement of channel widths along the Muskegon River was conducted in the initial survey of the basin in the 1830s (some resurveyed in the 1850s). This survey produces the baseline that was used to compare subsequent changes in channel width. Stream widths on the Muskegon River were measured by the government surveyors along section lines with chains, with one chain being equivalent to 20.1 m. Measurements at the time were made from the section line intersection of one bank and then again taken at the opposing bank, still along the section line (Figure 4.1). These measurements provide an accurate idea of what pre-settlement channel widths were during the 1830/50s. Width

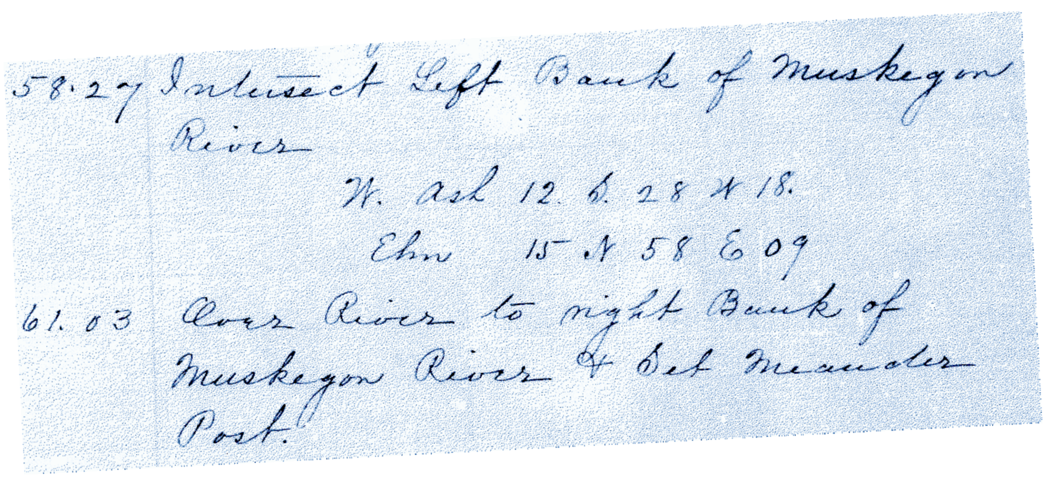
58.27 Intersect Left Bank of Muskegon River
W. ash 12 S. 28 N 18.
Elm 15 N 58 E 09
61.03 Over River to right Bank of Muskegon River & Set Meander Post.

Figure 4.1. A typical survey description of the Muskegon River (Archives of Michigan, 2014). This description shows the channel width in this location was 2 chains and 76 links, which equals ~55.5 m.

measurements in this study were used to calculate 1800s widths on the Muskegon River for comparative purposes with 2013 widths. Note that many section lines cross the river at an oblique angle relative to the channel. At several of these intersections the degree of this angle is such that it results in exaggerated channel width measurements. In an effort to avoid this problem, channel-width measurements were included only at intersections where the angle of the stream fell within a range that was 80^{o} to 100^{o} relative to the intersecting section line.

4.2 Field Work

The comparison of initial survey and modern channel width measurements has been conducted in several studies (i.e., Knox, 1977; Martin and Johnson, 1987; VanLooy and Martin, 2005). These studies shed light on the potential complications of using this methodology because some uncertainty exists with respect to the exact width measurement initially calculated in the survey. According to Knox, (1977), surveyors most likely measured the water width at river crossings. On the other hand, Martin and Johnson (1987) noted that it is unclear whether the initial survey measured bankfull width or water width. Both Martin and Johnson (1987) and VanLooy and Martin (2005) choose to measure bankfull width because of its reliability and the relative uncertainty of the initial surveys measurement procedures. Given the apparent greater reliability of the bankfull measuring method, this study uses this approach in an effort to minimize errors.

Between June and August 2013, a total of 106 width measurements were collected on the Muskegon River. These width measurements were compared to reliable land survey measurements to see if the set of data were distinguishable. After the initial survey measurements were obtained, the same locations were used to collect width measurements in the field for comparative purposes. Floating the river with a kayak allowed access to otherwise inaccessible sample locations. Predefined waypoints were created on a Global Positioning System (GPS) to identify the relevant section lines. Michigan Department of Natural Resources survey markers were often present at these locations, providing ground truth for the waypoint (Figure 4.2). At each of these sample locations, a bankfull-width measurement was taken using a Simmons LRF600 TI laser rangefinder, which has a range of up to ~550 m and a +/- 0.91 m accuracy. Width measurements were taken from on top of one bank, and the digital rangefinder sent a laser to the opposing bank (Figure 4.3). The bankfull widths at sampling locations were identified by the flat surface near the channel boundary (e.g., adjacent floodplain) and denser vegetation (e.g., tree line). All measurements and field notes were recorded on waterproof paper, to assure that data would not be erased by water.

Figure 4.2. A survey marker alongside the Muskegon River in Missaukee County (photographed by author).

Figure 4.3. Collecting bankfull width measurements using a digital rangefinder (photographed by Daniel Kowalski).

CHAPTER 5

RESULTS AND DISCUSSION

This chapter describes the character of the Muskegon River valley and associated channel. It begins with a discussion of the stream's longitudinal profile in relationship to the surrounding geomorphology. Secondly, channel geometry for the entire length of the Muskegon River is described. Thirdly, changes in the stream's historic channel position is analyzed. Fourthly, stream width data taken along section lines throughout the entire expanse of the Muskegon River is analyzed. Finally, the possible causes of historic channel changes in meander pattern and width in the Muskegon River are considered. Channel characteristics (e.g., meander geometry and channel width) and the degree of historic change along the river are presented in Table 5.1.

5.1 Longitudinal Profile

The Muskegon River flows through a number of different geomorphic environments from its outflow at Houghton Lake to its inflow at Muskegon Lake (Figure 5.1), producing several variations in channel gradient throughout the system (Table 5.1; Figure 5.2). In the upper reach, the stream mainly flows on a low gradient floodplain averaging ~0.6 m per km. These low gradients are likely associated with a past proglacial lake (i.e., Glacial Lake Roscommon), which created a lake-plain during the last glaciation (Schaetzl, 2014). As a result, the upper portion of the stream has some of the lowest gradients in the entire system, giving it an unusual longitudinal profile. For example, low gradients are not typical

TABLE 5.1. Channel characteristics of the Muskegon River for the Upper, Middle, and Lower Reaches.

	UPPER REACH (Houghton Lake to Evart)	**MIDDLE REACH** (Evart to Hardy Dam)	**LOWER REACH** (Hardy Dam to Muskegon Lake)
Average Channel Slope	~0.6 m/per km	~1.1 m/per km	~0.8 m/per km
Average Sinuosity	1.94	1.33	1.23
Average Radius of Curvature	58.05 m	408.87 m	237.83 m
Average Meander Wave-Length	0.36 km	1.80 km	0.72 km
Average Meander Amplitude	0.30 km	1.11 km	0.71 km
Intial Survey - Average Channel Width	33.73 m	84.92 m	86.13 m
Summer 2013 - Average Channel Width	36.48 m	61.73 m	63.19 m
Abandoned Meanders - 1830s to 1938	20	0	3
Abandoned Meanders - 1938 to 2012	10	0	0
Stream Bottom	Sand/Muck	Sandy/Gravel	Sand/Muck

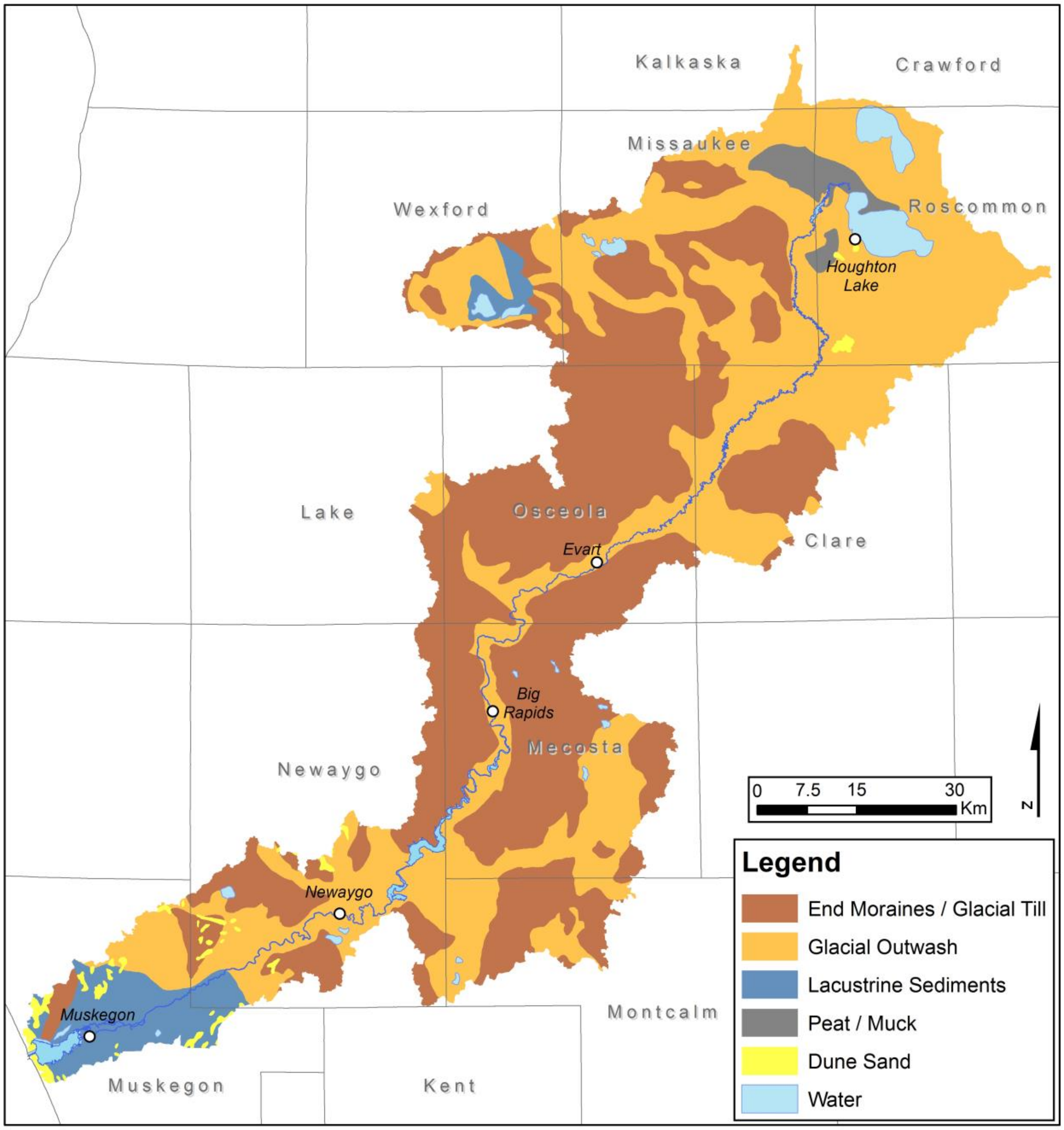

Figure 5.1. A Quaternary geology map of the Muskegon River Basin. Note that the stream flows largely on glacial outwash until it reaches the lower reach, where the stream flows on lacustrine sediments. Also note that the stream is confined by several end moraines in the middle reach.

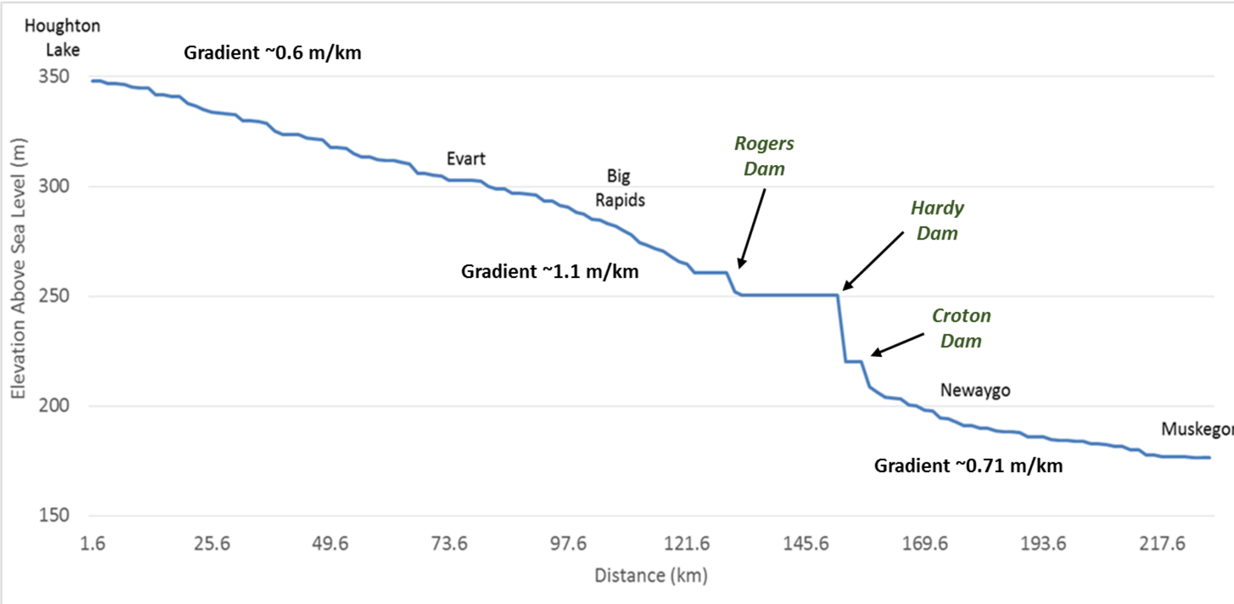

Figure 5.2. Longitudinal profile of the Muskegon River.

of most headwater reaches, which commonly have a concave profile, making the reach less sinuous (Morisawa, 1985)

Much of the upper reach has a relatively low gradient compared to the lower reaches of the Muskegon River. As a result, the stream has formed a meandering pattern throughout the upper reach. The stream in this part of the system incised into the surrounding St. Helen and Grayling Outwash Plains producing a variety of distinctive cutbanks that range in height from a few meters to over ten meters above the stream (Figure 5.3). The surfaces associated with the cutbanks are level (Figure 5.3), indicating that the stream likely eroded into the various terraces identified by Arbogast et al. (2008).

As the Muskegon River flows into Osceola and Mecosta Counties channel gradients increase (Table 5.1; Figure 5.2), averaging 1.1 m per km. Higher gradients in this portion of the basin are related to the surrounding Cadillac Morainic Uplands (Schaetzl et al., 2013),

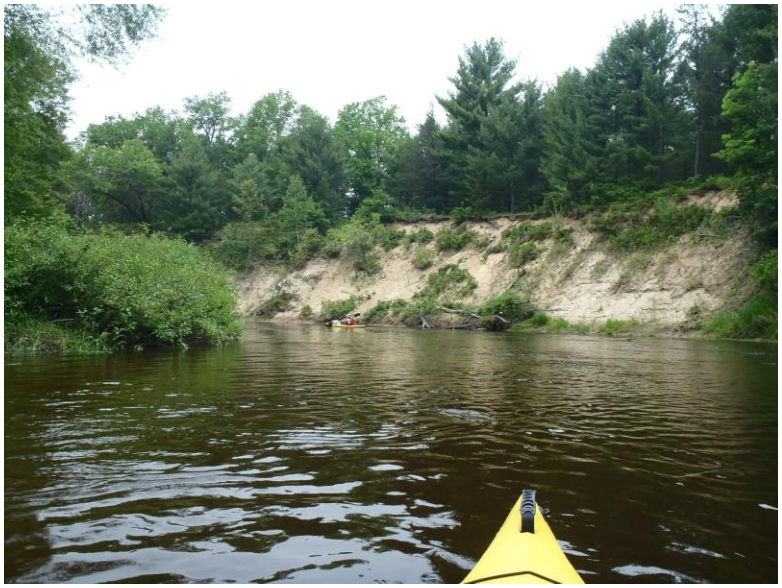

Figure 5.3. A high cutbank in the upper reach of the Muskegon River (photographed by author).

which encompasses much of the Muskegon River between Evart and the Hardy Dam (Figure 5.1). These uplands supply the middle reach with sandy alluvium and glaciofluvial sediments, changing the composition of the stream bottom from predominantly sandy in the upper Muskegon River to sandy/gravel with several large boulders in the middle reach (Figure 5.4). Like the upper reach, this portion of the stream has incised into the surrounding landscape, which rises ~90 m above the current floodplain in some locations.

The Muskegon River below the Croton Dam has a lower gradient [0.71 m per km] than in the middle reach (Figure 5.2). This decrease in slope is likely associated with the stream flowing out of the Cadillac Morainic Uplands (Figure 5.1) into a portion of the basin with more moderate relief (i.e., the Fruit Ridge Terrain; Schaetzl et al., 2013). In this reach, several high (i.e., ~15 m) sandy cutbanks are apparent, indicating that the channel has incised considerably into the surrounding Newaygo Outwash Plain (Figure 5.5). Several large hills (i.e., likely remnants of the Lake Border Moraine) also surround the Muskegon

Figure 5.4. A typical view of the middle reach of the Muskegon River (photographed by author).

Figure 5.5. A high cutbank on a meandering stretch of the lower Muskegon River, near Croton (photographed by author).

River in Newaygo County, supplying the stream with fine sand and gravels (Figure 5.1).

As the stream flows toward southwestern Newaygo and Muskegon Counties, channel slope continues to drop, averaging ~0.3 m per km (Figure 5.2). These low channel gradients are associated with the Glacial Lake Chicago lake-plain, which formed during the Late Pleistocene (Larson and Schaetzl, 2001). Sediments on the lake-plain are predominantly poorly drained (due to a high water table), sandy, lacustrine soils (Figure 5.1), which are also found within the stream channel. As a result, the river is sluggish before entering Muskegon Lake.

5.2 Channel Geometry

The Muskegon River is a meandering stream over much of its length. At the point where the stream flows out of Houghton Lake it begins to meander almost immediately, with sinuosity values that range from ~1.3 to 2.5 as the stream flows southwesterly into the Dead Stream Flooding (Figure 5.6). In this headwater reach, meanders are small and numerous (Figure 5.7 A.), with radius of curvature values that average only ~54 m. As a result, meander wavelength ranges between 0.21 and 0.55 km and amplitudes vary between 0.14 and 0.32 km, respectively.

As the Muskegon River flows out of the Reedsburg Dam, ~3.5 km northeast of the M-55 overpass, the stream becomes increasingly sinuous. This trend continues well into Clare County (Figure 5.6), with sinuosity values ranging between 1.3 and 3.2. The highest sinuosities occur in southwestern Roscommon and southeastern Missaukee Counties. Radius of curvature values are generally consistent with those found above the Reedsburg Dam, ranging between 47 and 75 m. Meander wavelength and amplitude in this reach

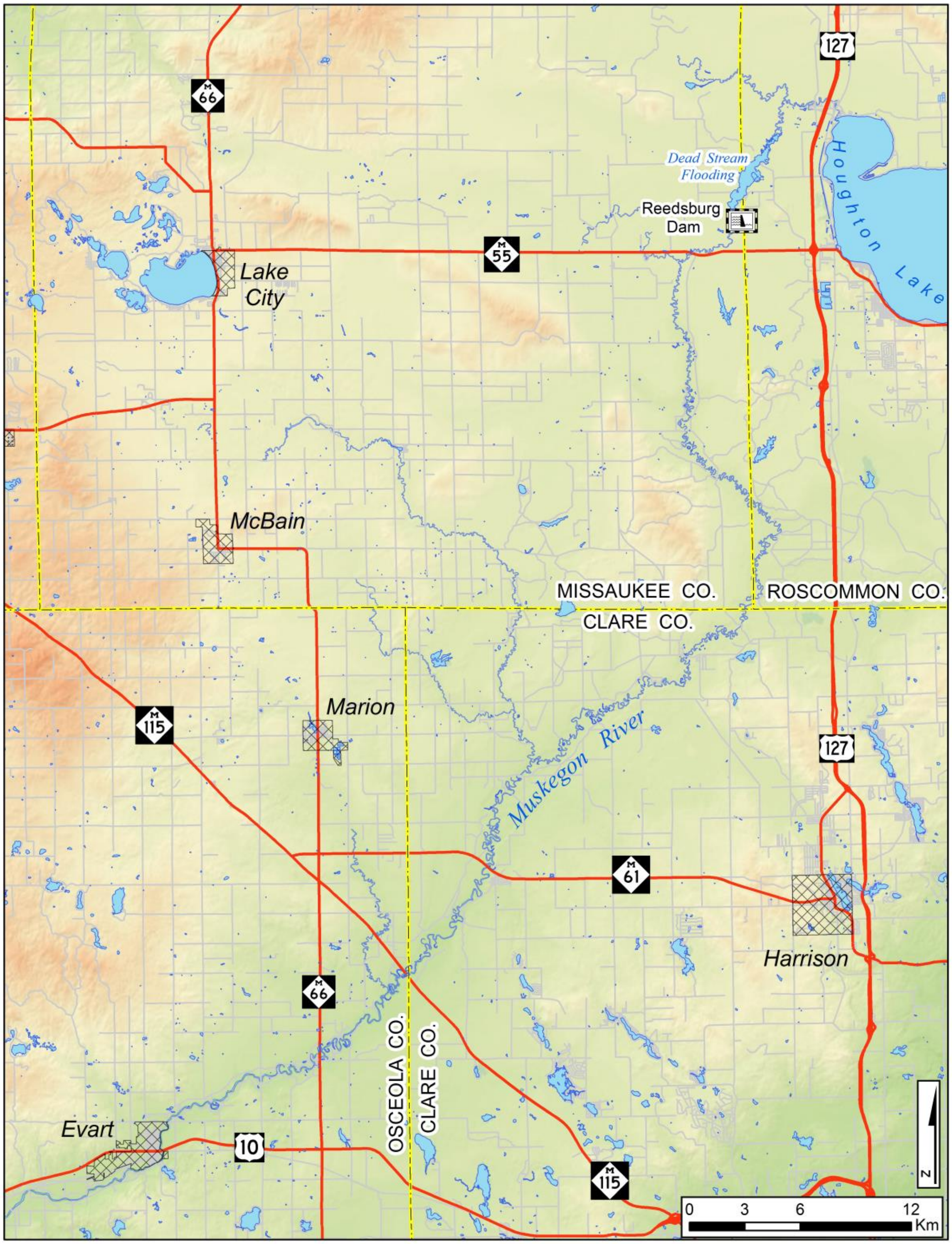

Figure 5.6. Map of the upper Muskegon River. In this reach the stream has a distinct meandering pattern, with sinuosity values up to ~3.2.

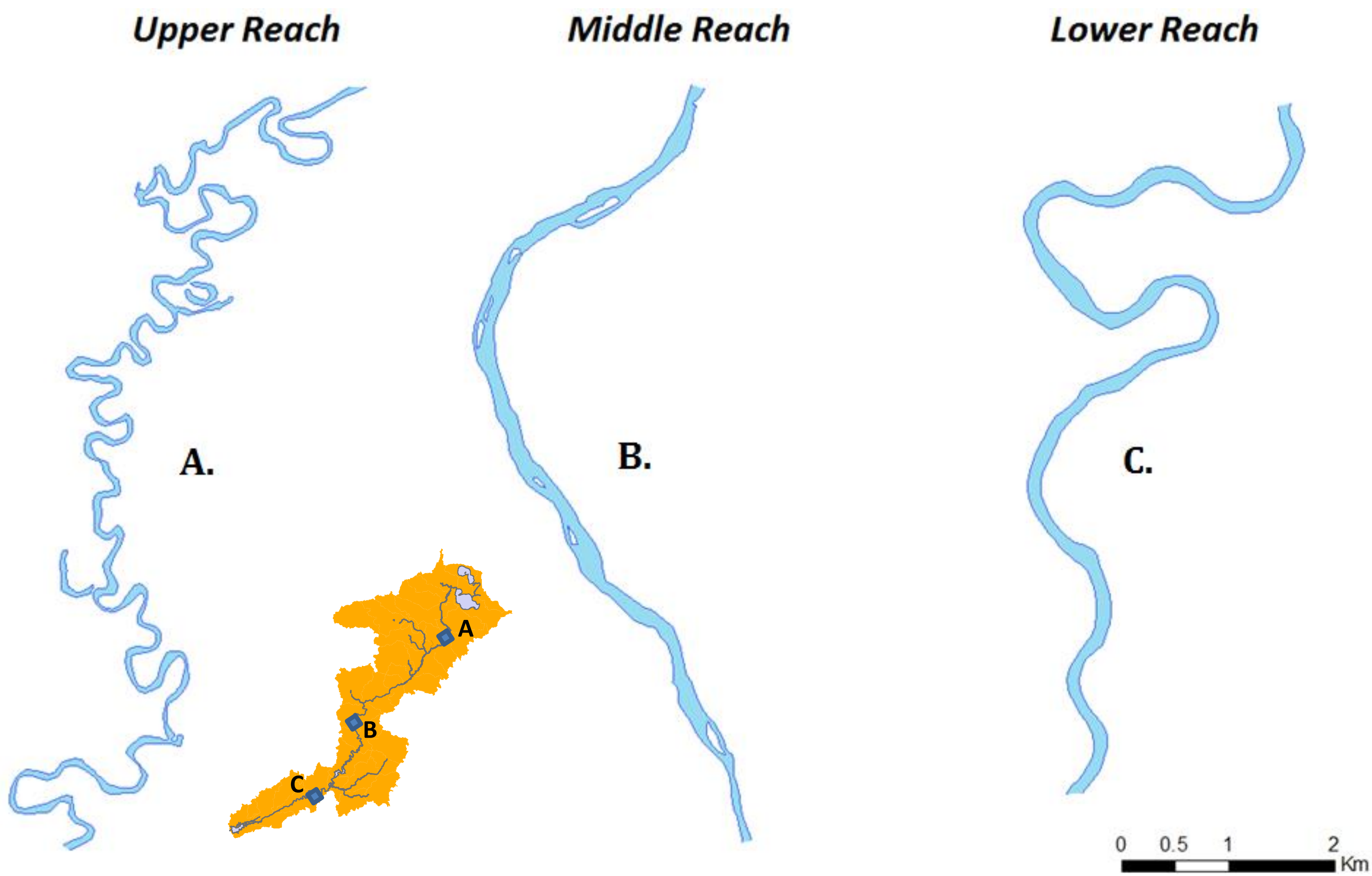

Figure 5.7. Representative meandering patterns for the distinct reaches of the Muskegon River. A.) the tortuously meandering upper reach, B.) the less sinuous to almost straight middle reach, and C.) the moderately meandering lower reach. Note the differences in channel sinuosity, width, and pattern between the reaches.

increases slightly from those found above Reedsburg Dam, with meander wavelength ranging from 0.21 to 0.61 km and amplitude values between 0.16 and 0.66 km. The meander pattern of the upper reach is associated with the lower stream gradient in this area (Figure 5.2), which averages ~0.6 m per km.

The channel geometry appears to change in the vicinity of the M-66 overpass, about 8 km upstream of Evart. In this reach, meander patterns are less sinuous, ranging between 1.1 and ~2.0. Meanders in this reach are broad and less compact compared to those upstream (Figure 5.6). This change in meander pattern is also reflected by the larger radius of curvature values in this portion of the Muskegon River, which range from 142 to 304 m compared to values averaging only ~66 m in the rest of the upper reach. In addition, meander wavelength in this reach ranges between 0.51 to 2.6 km, while amplitude ranges between 0.34 to 1.2 km. These values are substantially larger than those found in the rest of the upper reach (Table 5.1).

As the Muskegon River flows into Mecosta County south of Hersey (Figure 5.8), the stream becomes much straighter, with large/broad meanders (Table 5.1; Figure 5.7 B.). Channel sinuosity in this middle reach drops substantially, ranging between 1.5 to 1.1. This change is also reflected by large radius of curvature values that increase to between 231 m and 876 m. Meander wavelength and amplitude in this reach are similar to those in Osceola County, ranging from ~1.8 to 2.6 km, respectively. While this reach is similar to the upstream portion of the Muskegon River in Osceola County, some important differences occur, such as distinct riffle-and-pool sequences, large rapids, and several mid-channel bars (Figure 5.7 B.). These features are likely a product of local alternating energy losses/gains and sediment loads due to a higher gradient. The gradient of the middle reach is

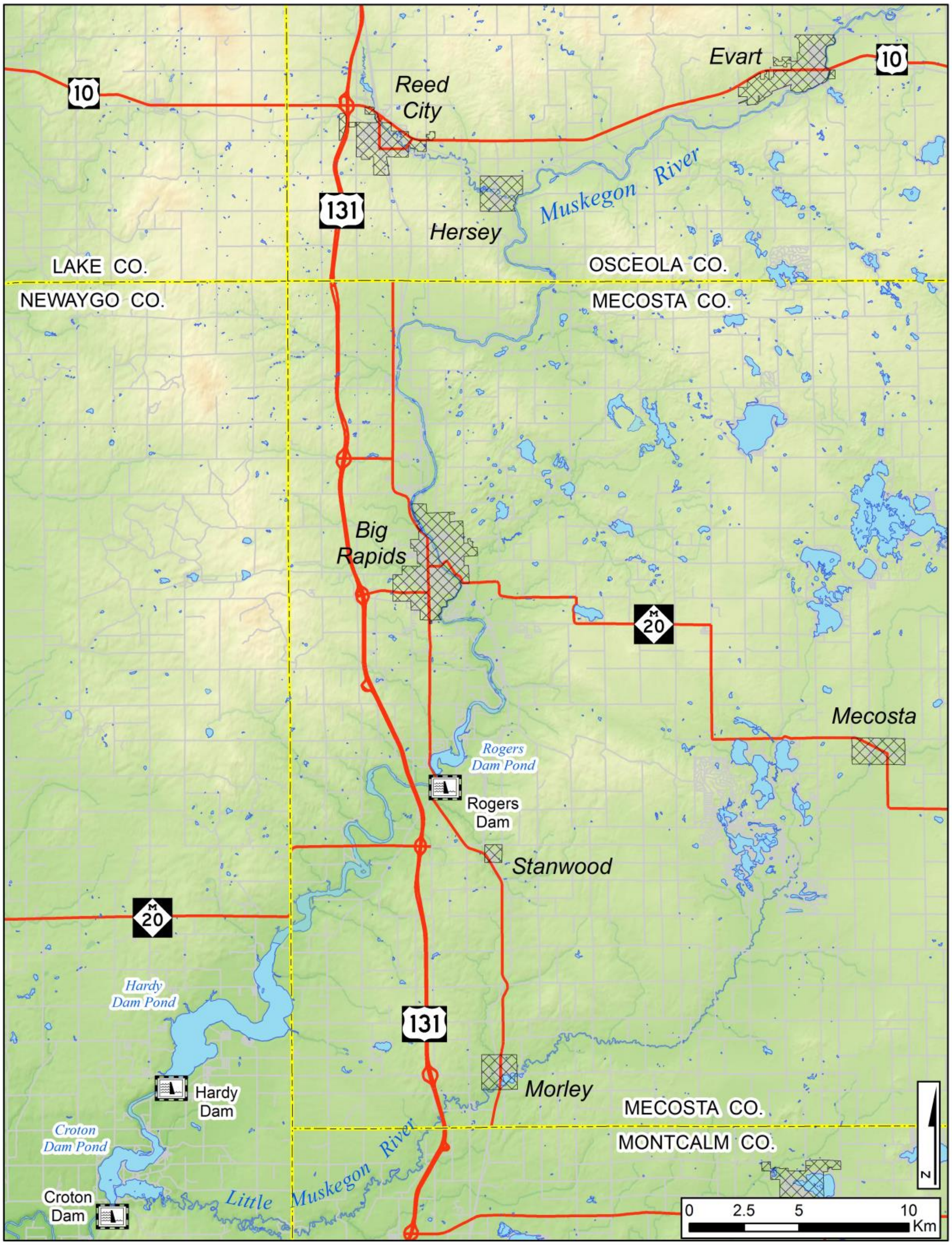

Figure 5.8. The middle reach of the Muskegon River between Evart and the Hardy Dam. Note the broad, sweeping meanders in southern Osceola and northern Mecosta Counties, as well as Rogers, Hardy, and Croton Dams.

substantially greater than that of the upper reach because the stream flows off an upland confined by a recessional moraine on both sides (e.g., Lake Border Moraine; Blewett et al., 2009), resulting in a drop in bed elevation of ~57 m in ~100 km. Gradients in this reach average ~0.98 m/km, but get as high as ~3.8 m/km.

In southwestern Mecosta County and southeastern Newaygo County, the Muskegon is dammed at three places (i.e., Rogers, Hardy, and Croton Dams), forming large impoundments behind each structure. As a result of these reservoirs, calculating the natural channel geometry is very difficult for much of the reach between Rogers and Croton Dam. Nevertheless, some values were obtained between the dams where the active channel has only been minimally altered (Figure 5.8). In this reach, sinuosity values range between 1.1 and 2.6, making this portion of the Muskegon River the most sinuous in the middle part of the stream. In addition to higher channel sinuosity, radius of curvature values in this portion of the stream are also very high (908 m), which could be due to stream damming (e.g., the flooded river valley produces larger radiuses than those formed naturally).

Below Croton Dam the Muskegon River appears to have a similar channel geometry to the portion of the stream below Rogers Dam. For example, the channel sinuosity of the Muskegon River in Newaygo County ranges from 1.1 to 2.0, which is more sinuous than much of the middle reach of the stream. This part of the channel also has a distinctive radius of curvature compared to the rest of the system, ranging from 182 to 455 m, with wide meander wavelengths [0.48 to 1.2 km] and amplitudes [0.46 to 1.35 km]. As a result, this portion of the stream exhibits large broad meanders (Figure 5.7 C.).

Further downstream in Muskegon County, the channel becomes less sinuous (Figure 5.9), with a maximum sinuosity of ~1.4. The decrease in sinuosity also has a direct effect on radius of curvature values, which range between 114 and 197 m. In this reach, meander wavelength ranges from 0.2 to 0.45 km and amplitudes range from 0.32 to 0.51 km, both of which decrease as the stream becomes straighter. As the Muskegon River continues to flow downstream it takes the form of an anastomosing channel near Muskegon Lake (Figure 5.9). This pattern likely formed because the channel floods more frequently near Muskegon Lake, due to a relatively low valley gradient and high stream discharge (Nanson and Knighton, 1996).

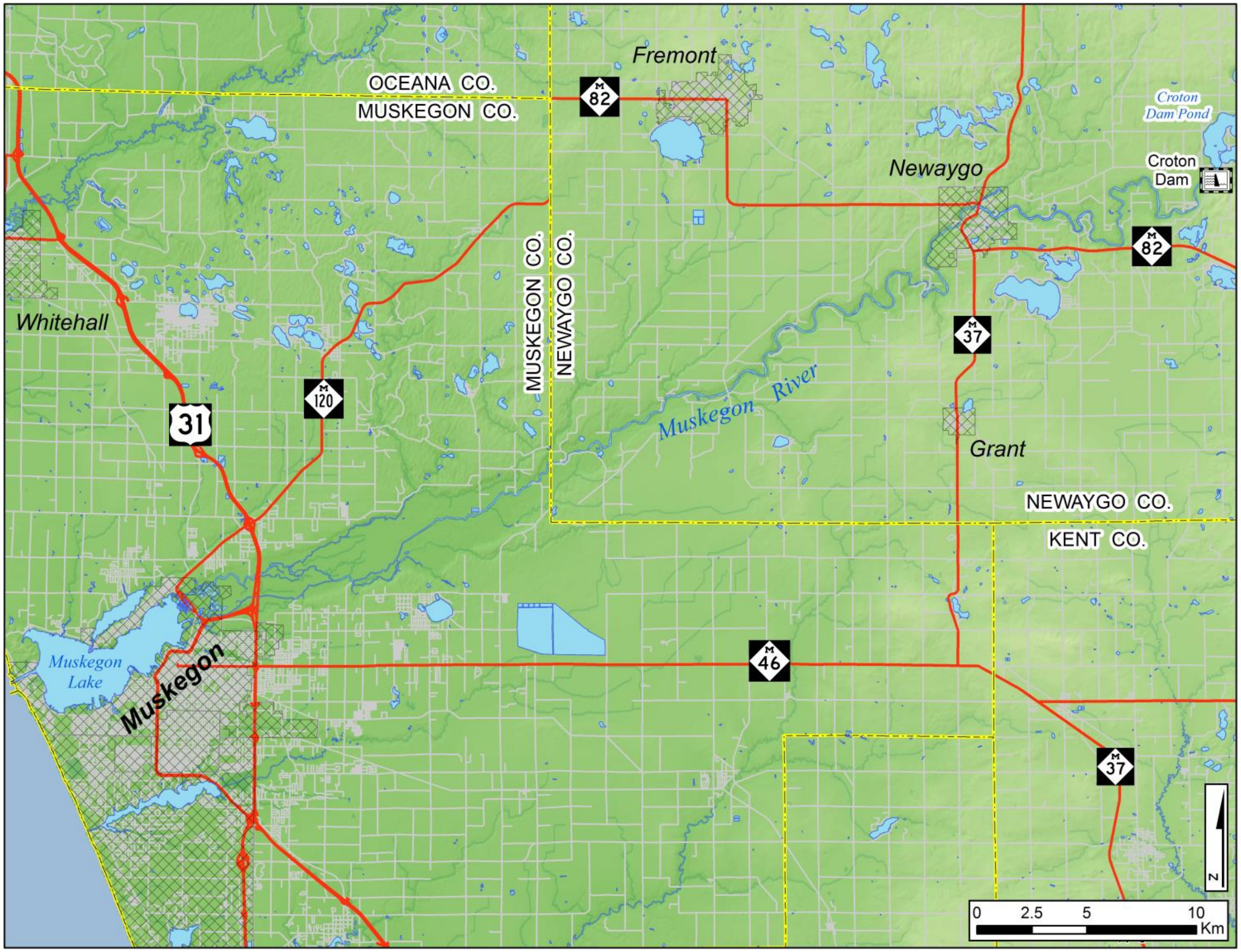

Figure 5.9. The lower reach of the Muskegon River between the Croton Dam and Muskegon Lake. Note the moderately meandering portion of the stream in Newaygo County, as well as the anastomosing pattern in central Muskegon County.

5.3 Historic Changes in Channel Position

Given the general meandering nature of the Muskegon River, a second goal of this thesis is to analyze historic changes in channel position. Several oxbows (Figure 5.10) and meander scars are present alongside the Muskegon River, indicating that the stream has been actively migrating on its floodplain in the recent past, especially in the upper reach (Arbogast et al., 2008). This section describes different types of historic channel adjustments that have taken place throughout the Muskegon River. In addition, the extent of these changes will be discussed to analyze the rate/span of changes both geographically and through time.

The most obvious location where historic channel adjustments have occurred in the Muskegon River is in its upper reach between Houghton Lake and Evart, where the stream

Figure 5.10. An oxbow from the upper Muskegon River. This portion of the stream was found to be abandoned in the 1960s using repeat photography (photographed by author).

meanders a great deal. Further analysis of the stream in the upper reach indicates several abandoned meanders (i.e., oxbows) have formed during the historic record (Figure 5.11). These cutoffs in the upper reach are 'neck cutoffs', in that a thin strip of land separating two meander bends has been breached, allowing the stream to continue flow in a less sinuous pattern (Charlton, 2008; Figure 5.12).

When the data were further analyzed it became apparent that more abandoned meanders occurred between the mid-1800s and 1938 (i.e., 20) than between 1938 and 2012 (i.e., 10; Table 5.1; Figure 5.11). The cause for this difference may be related to the time span between the two periods (i.e., 1845/1856 -1938 = 93 to 82 years, while 1938-2012 = 74 years), which may have allowed more abandoned meanders to form during the early historic record. Nonetheless, results indicate that an abandoned meander formed every 4.65 years between 1845 and 1938, whereas such change occurred every ~7.4 years between 1938 and 2012. In addition, the spatial extent of abandoned meanders in the upper reach has contracted during the late historic record (Figure 5.11), with the areas of most recent change occurring well within the zone affected between the mid-1800s and 1938.

Analysis of a representative reach provides a good illustration of change seen in the late historic period. Using repeat photography, local adjustments to channel slope and sinuosity can be calculated, potentially identifying trends and changes in the fluvial system. A good location to analyze local channel change is in northern Clare County, where the periodicity of meander cutoffs in the late historic period has been consistent. In this reach, the comparison of imagery indicates that the rate of cutoff meanders occurred at the rate of

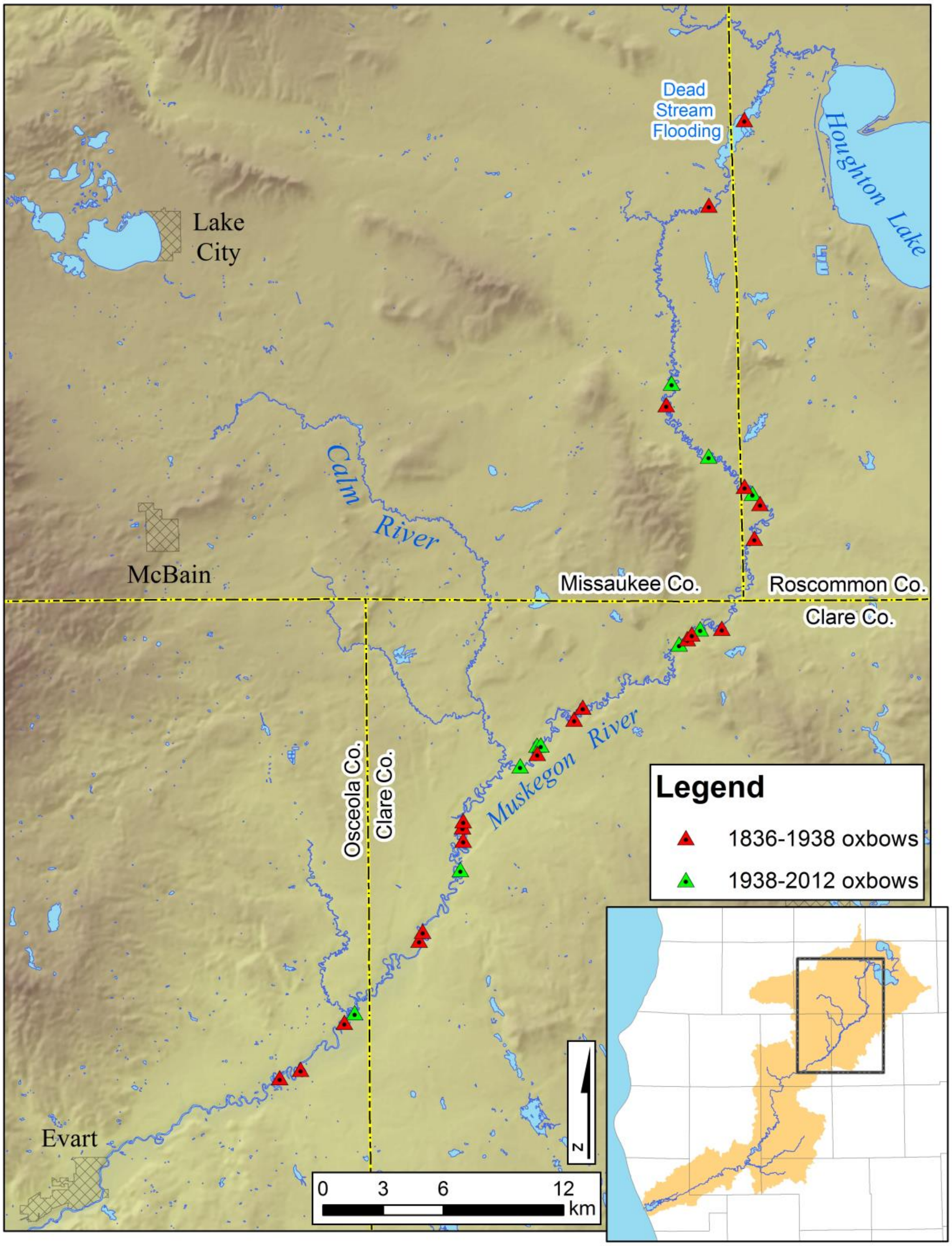

Figure 5.11. The location of abandoned meanders in the upper system during the historic record. Red triangles represent oxbows that formed between 1836 and 1938, whereas orange triangles represent oxbows forming between 1938 and 2012. Note the spatial extent of channel change between the two periods.

about one cutoff per decade (Figure 5.13). As a result, sinuosity in this reach decreased from; ~2.25 in 1938 to ~1.8 in 2012.

Although the previous channel alterations are a product of natural channel migration, some channel changes in the upper reach appear to be a result of human modification. Such changes in channel position occur where highways cross the stream. In these cases, the channel was likely straightened to prevent structural damage. For example, both the M-55 and M-115 overpasses have altered the natural flow of the Muskegon River (Figure 5.14).

As the Muskegon River flows into Osceola County, the number of abandoned meanders progressively decreases (Figure 5.11) until none occur downstream of the M-66

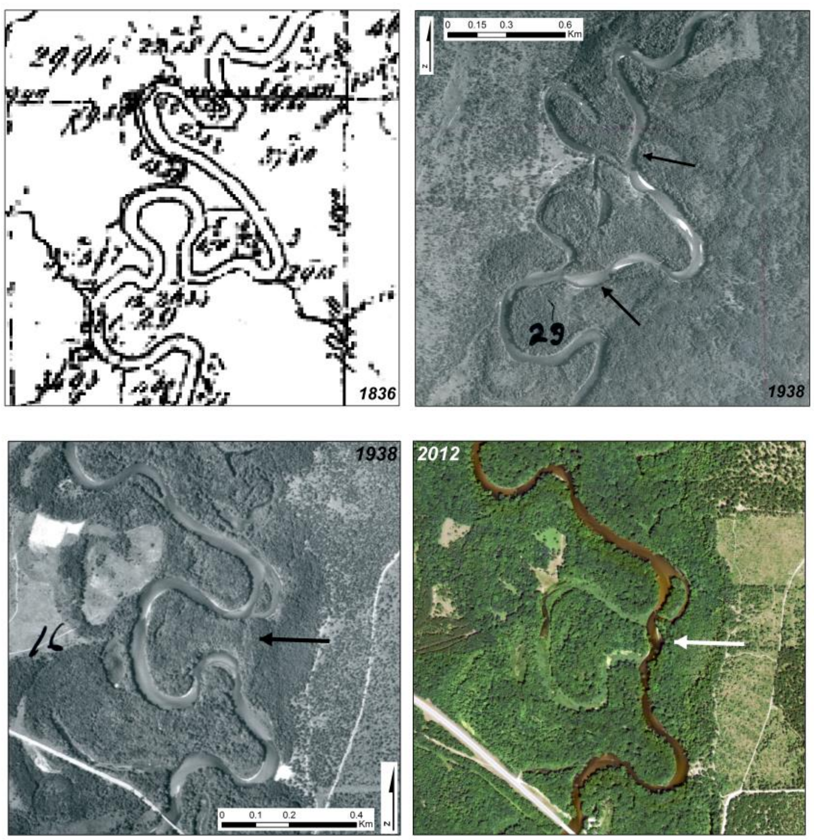

Figure 5.12. Historic meander cutoffs (arrows) from two separate areas in the upper part of the Muskegon River system.

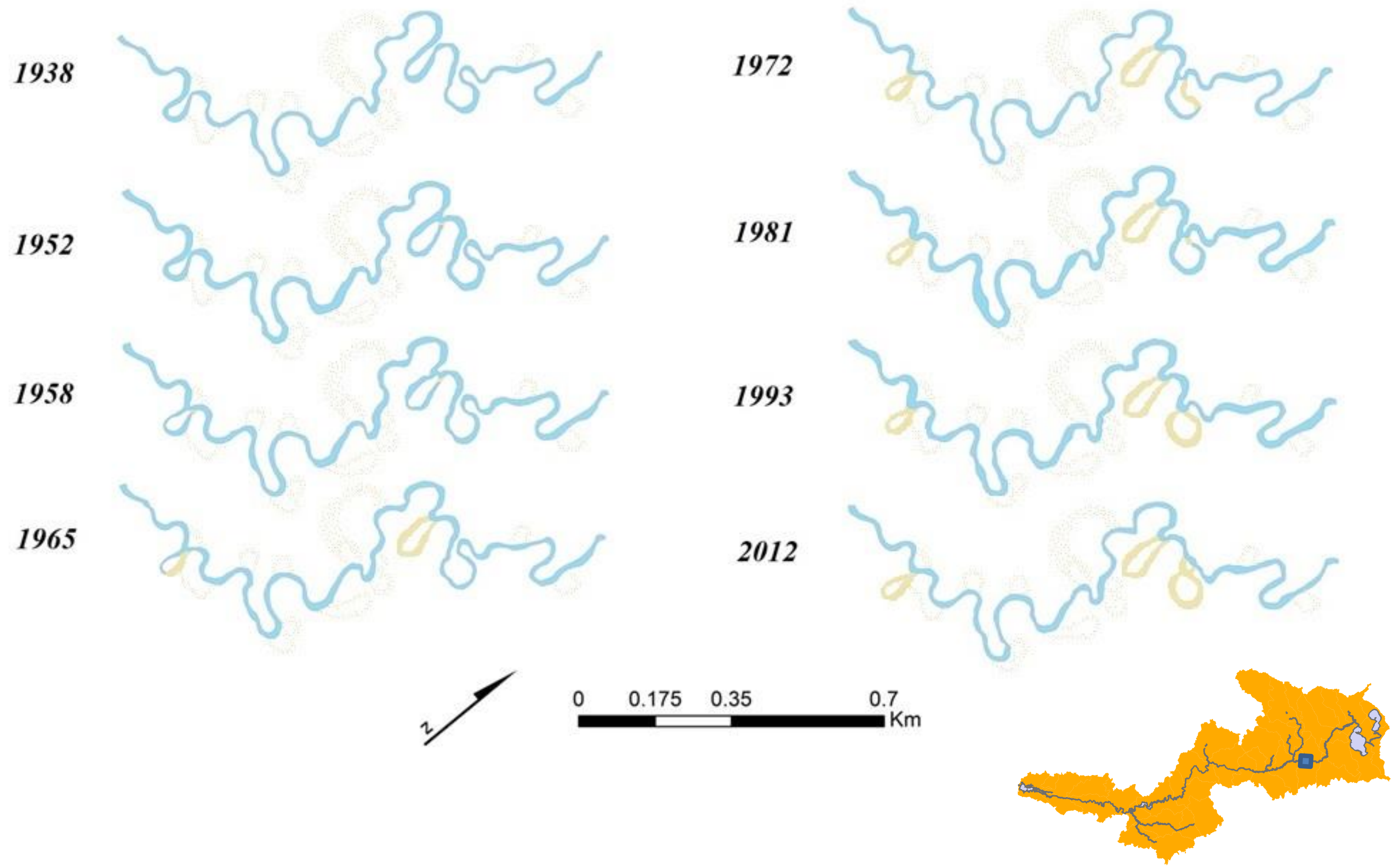

Figure 5.13. The progression of channel changes on a highly sinuous reach of the upper Muskegon River between 1938 and 2012. Note the series of abandoned meanders (i.e., light green) that have formed during the time period.

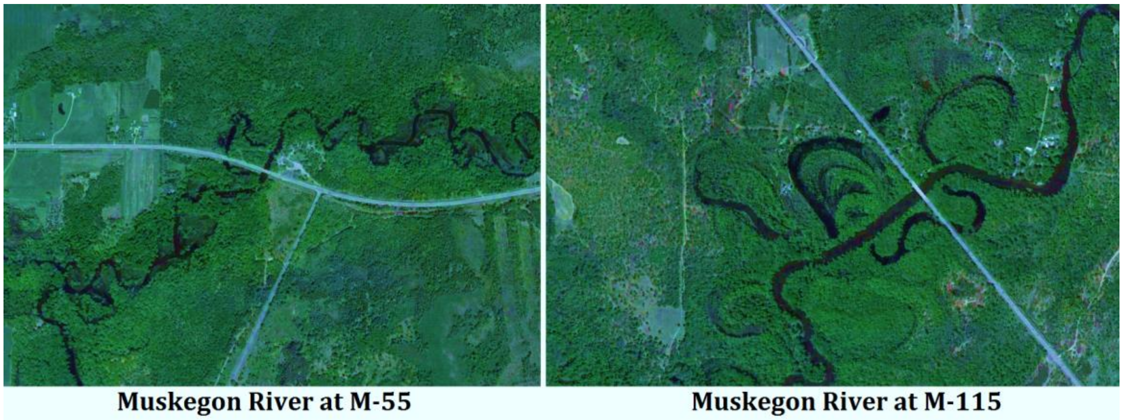

Figure 5.14. Aerial images showing the effects of channel dredging. Note the straightened reaches at the M-55 and M-115 Muskegon River overpasses.

overpass. As a result, the Muskegon River transitions into a relatively straight stream. When analyzing the floodplain in this reach of the stream, no evidence of meander scars or oxbows is evident, indicating that the stream position has not dramatically changed in the historic period. While the stream does not appear to have migrated in this area, some changes in channel bank morphology were found during the historic record. Examples of these changes are in the form of 1) increasingly eroded cutbanks 2) prograding point bars, and 3) the formation/deformation of mid-channel bars (Figure 5.15).

Perhaps the most apparent channel change in the middle reach is the formation, deformation, and/or alteration of sandy point bar and mid-channel bar deposits (Figure 5.16). These features alter their shape and size on a regular basis in the middle reach of the Muskegon River. As a result, the stream temporarily changes its morphology, even developing distributary channels (Figure 5.7 B.; Figure 5.15). Often these mid-channel bars attach themselves to the adjoining bank as vegetation stabilizes the features, which aids in changing flow patterns. This often results in the active channel altering its shape

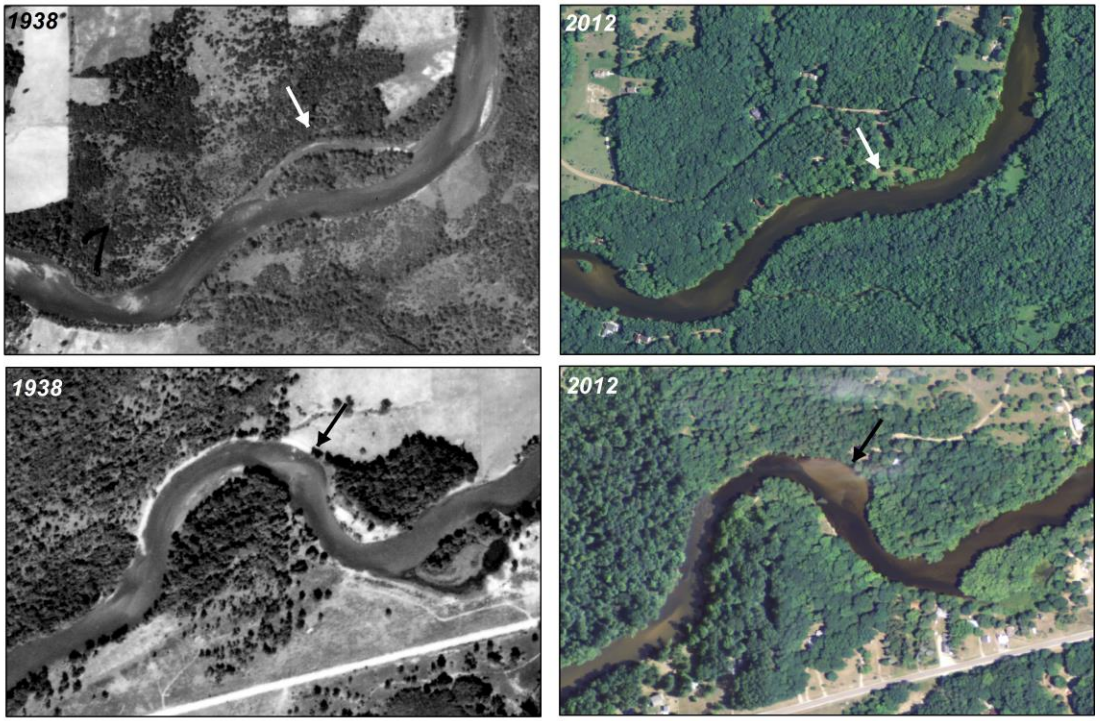

Figure 5.15. Aerial images showing representative locations of channel change on the middle reach of the Muskegon River. Note that arrows indicate an area of channel change.

Figure 5.16. A mid-channel bar downstream from Evart (photographed by author).

and course (Figure 5.15). These elongated bars typically form in the middle of the stream in relatively wide locations (Charlton, 2008) and form as faster flows divert against channel banks, allowing finer alluvial sediments to settle in the center of the stream (Church and Jones, 1982).

Conversely, the Muskegon River has generally retained its channel morphology in the middle reach (e.g., in Osceola and Mecosta Counties) during the historic record. This is likely associated with an increase in channel slope in conjunction with a confined river valley (e.g., steep moranic side walls; Figure 5.1). These factors cause the channel to straighten in this portion of the stream, not encouraging meander extension or abandonment. Increased sinuosity is not noted again until further downstream below the Rogers Dam. In this portion of the stream the channel cannot be properly analyzed due to the flooded river valley (e.g., Hardy Dam Pond and Croton Dam Pond). Consequently, channel changes could not be analyzed until below Croton Dam.

Like the upper reaches of the Muskegon River, the lower reach from Croton Dam to Muskegon Lake is sinuous to meandering, with numerous meander scars and oxbow lakes identified between Croton and Newaygo (e.g., Brooks Township). Analysis of the historic air photos indicates that several meanders were abandoned between the 1920s and 1938 (Figures 5.17 and 5.18). Unlike the upper Muskegon River, the majority of channel changes in this reach occurred several decades to a century later, indicating that these parts of the system act independently from one another. In addition to these abandoned meanders, several extended meanders have formed downstream in Cedar Creek Township during the late historic record (Figure 5.19), suggesting that the lower Muskegon River continues to migrate on its floodplain well into Muskegon County. Unlike the portion of the

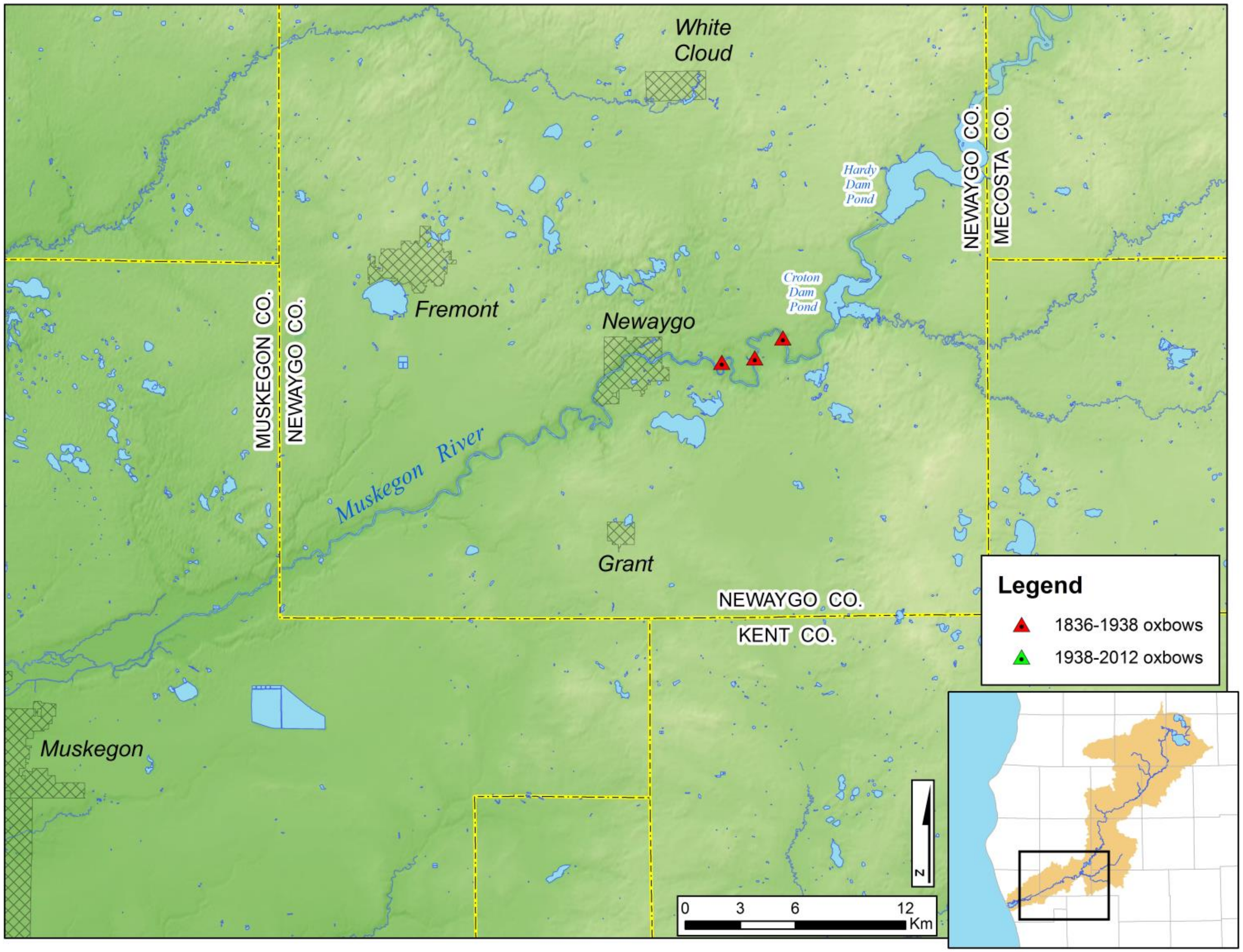

Figure 5.17. The location of abandoned meanders in the lower system during the historic record. Red triangles represent oxbows that formed between 1836 and 1938.

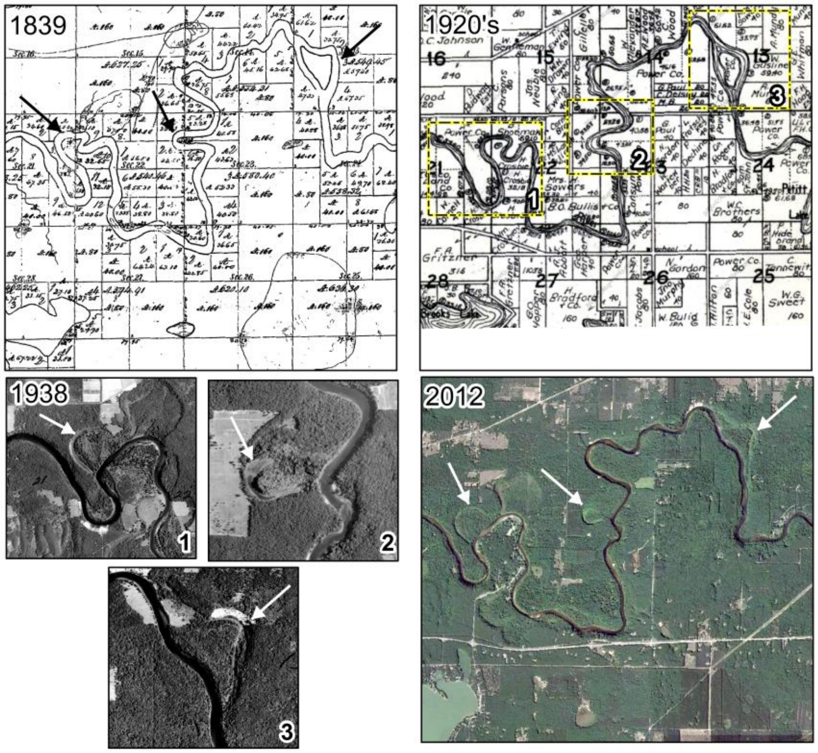

Figure 5.18. Survey map (1839), plat map (1920s), and aerial imagery (1938 and 2012) showing channel changes that occurred in the lower reach of the Muskegon River. Note that arrows indicate the location of abandoned meanders.

Figure 5.19. Aerial images showing locations of channel change on the lower reach of the Muskegon River. Note that arrows indicate an area of channel change.

stream in Brooks Township, meanders in this reach have larger wavelengths and amplitudes, deterring cutoffs from forming.

Further downstream in Muskegon County, the stream again straightens to a less sinuous pattern. As a result, there is no evidence of historic meander extension or abandonment. Here the stream separates into several distributary channels, which branch off from the main channel producing an anastomosing pattern (Figure 5.9). This pattern starts near Maple Island Road, where the Maple River splits off from the Muskegon River for ~9 km. The Maple River then returns to the Muskegon River just upstream from Holton Duck Lake Road. Further downstream the main channel of the Muskegon River (e.g., in the center of the floodplain) appears to have been dredged during the early historic record, producing a straight reach about 3 km long. In this reach, several distributary channels break off from the main channel. As these channels near Muskegon Lake several converge into a main stream, whereas others parallel the main channel and debouch into Muskegon Lake separately.

5.4 Historic Changes in Channel Width

As previously discussed, analysis of aerial photos and maps indicate several alterations in channel position have occurred in the Muskegon River during the historic record. These changes indicate that the stream has migrated on its floodplain during the study interval, creating features such as meander scars, oxbows, and mid-channel bars. Changes in channel width, can also indicate alterations to a fluvial system. However, calculating changes in channel width is difficult to do accurately using such tools. Therefore this study uses survey notes and field data to analyze historical changes in stream width throughout the full extent of Muskegon River. This section will first describe the level of accuracy and differences between initial survey and field width data. It concludes with a discussion of historical changes in channel width from the headwaters to Muskegon Lake.

5.4.1 Accuracy and Complications of Calculating Channel Widths

Michigan was first surveyed as part of the United States Public Land Survey, which established a grid before initial settlement. Although some of these surveys occurred as early as 1815 in parts of a southeastern Michigan (Thomas, 2009), most were conducted in the 1830s, including those in the Muskegon River basin. During these surveys, multiple surveyors entered the field to layout townships and subdivide them into 36 sections (e.g., 1 square mile, 640 acres). In the course of establishing this grid, notes were taken, which provide pre-settlement conditions, such as soils, vegetation, and stream widths (Thomas, 2009).

Many of these surveys in northern Lower Michigan were fraudulent or deficient, causing a resurvey to be conducted between 1845 and 1856 (Thomas, 2009). This area included much of Roscommon, Missaukee, and Clare Counties, which encompass the upper Muskegon River. In many cases, width measurements in this part of the basin were resurveyed and represent widths two decades after the original survey. As a result, this thesis regards these measurements as mid-1800s widths.

Stream widths were typically described where a section line crossed a large stream (i.e., at north-south and east-west compass bearings). At such locations, the surveyors set a post at one channel bank and then set an opposing post at the adjacent bank. The surveyors then recorded a width measurement of the stream (Figure 4.1). As a result, channel width measurements were made throughout a stream's basin. In the case of the Muskegon River, 134 width measurements were calculated throughout the entire extent of the stream.

While these measurements provide a great way to analyze past stream conditions, they also may be problematic. When using survey notes for historic channel widths, it is difficult to ensure the validity of the measurements. For example, exaggerated widths are recorded when the orientation of the stream is not perpendicular to the section line, causing the width measurements to be variable (Figure 5.20). These width measurements appear much wider than they actually are (Figure 5.21). For this reason, these measurements were omitted when compared to summer 2013 channel widths.

Bank full width measurements recorded during the summer of 2013 are different from mid-1800s survey measurements in several ways. First, all of the 2013 width measurements were taken with a digital range finder perpendicular to the stream. As previously stated, initial survey widths were taken using links and chains. These measurements were not always taken perpendicular to the stream (Figure 5.20), sometimes causing width measurements to be unusually high (Figure 5.21). Finally, it is unclear whether measurements taken during the initial survey were done using bankfull or water width, potentially adding error when compared to 2013 bankfull widths.

Secondly, initial survey measurements were taken before any human infrastructure (e.g., dams, rip-rap, and roads) was placed in or alongside the Muskegon River. This allowed the widths of the natural flowing Muskegon River to be calculated from the headwaters to its terminus at Muskegon Lake. In contrast, measurements during the summer of 2013 were not taken in reservoirs above dams (e.g., Rogers Dam Pond, Hardy Dam Pond, and Croton Dam Pond) or in previously dredged portions of the lower

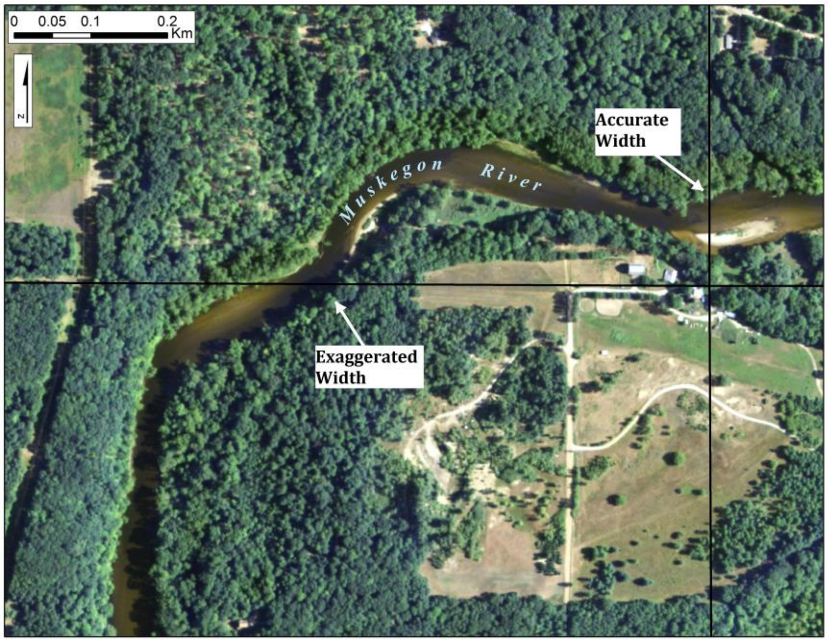

Figure 5.20. Examples of accurate and exaggerated channel widths when using the section line methodology.

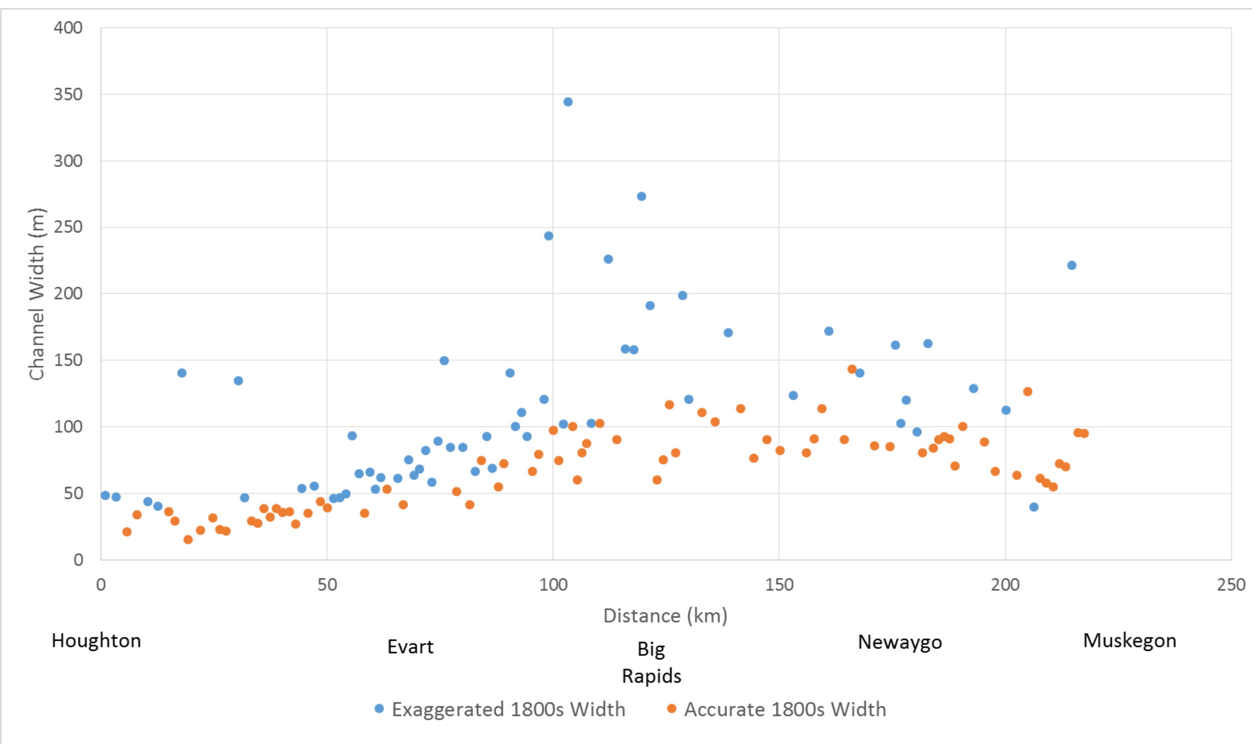

Figure 5.21. Initial survey measurements of channel width on the Muskegon River. Blue dots represent widths that are considered exaggerated, whereas orange dots represent accurate widths.

Muskegon River (e.g., above Muskegon Lake) were stream widths would not be representative of normal stream conditions. As a result, the spatial distribution of data points collected during the initial survey (Figure 5.22) does not match those taken during the summer of 2013 (Figure 5.23). Nevertheless changes in channel width can still be observed in reaches of the Muskegon River which have not been impounded or dredged.

5.4.2 Historic Changes in Channel Width

A main goal of this study was to analyze historical differences in channel width throughout the full extent of the Muskegon River. Width data calculated from the initial survey and during the summer of 2013 made this analysis possible. Data in the upper reach of the Muskegon River indicates that the stream typically ranges from ~20 to ~40 m wide (Figures 5.22, 5.23, and 5.24). Widths in this area stay relatively constant for around 135 km (e.g., near the Osceola and Clare County line) until increasing in Osceola County. Channel widths in the upper portion of the Muskegon River have stayed generally consistent throughout the historic record (Figure 5.25).

Further downstream in Osceola County, channel widths gradually widen as the Muskegon River receives additional water from major tributaries, such as the Clam River. In this reach, channel widths measure between ~40 and 60 m for a distance of around 55 km (i.e., to about Evart). The southwest orientation of the stream in this reach created several exaggerated mid-1800s widths, because the channel crossed area section lines at an angle. As a result, only a select few survey measurements were labeled as accurate (Figure 5.22), providing a limited amount of data points for comparison. Nonetheless the data correlates well within the dataset. Like in the upper reach, channel widths have

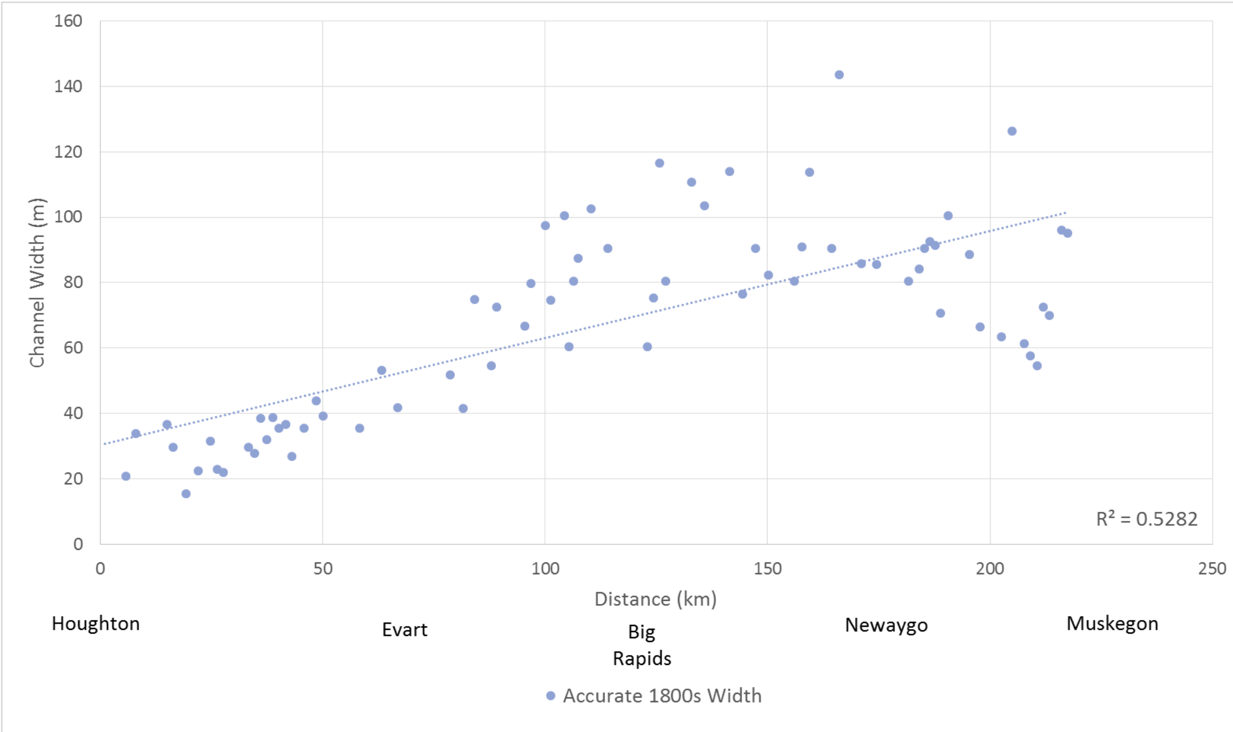

Figure 5.22. Accurate Muskegon River channel widths from the initial survey.

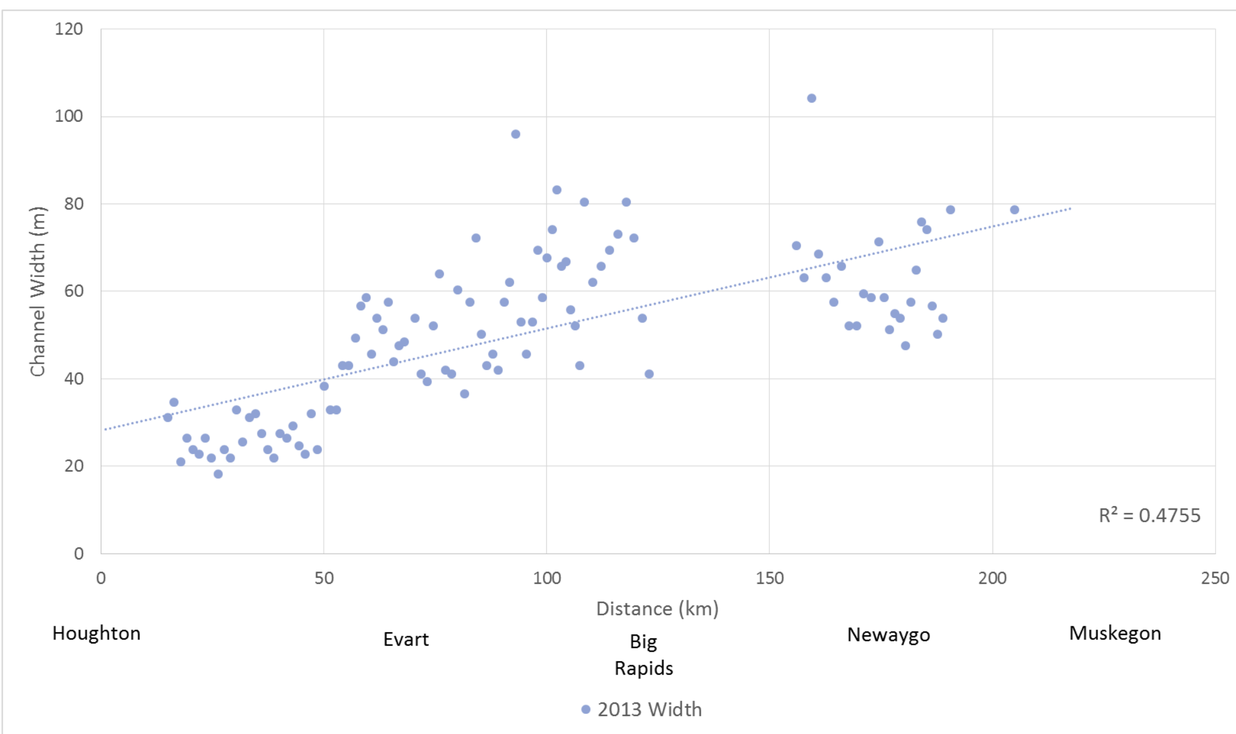

Figure 5.23. Muskegon River channel widths from the summer of 2013.

Figure 5.24. A photograph showing the typical stream width of the Muskegon River between Houghton Lake and Clare County (photographed by author).

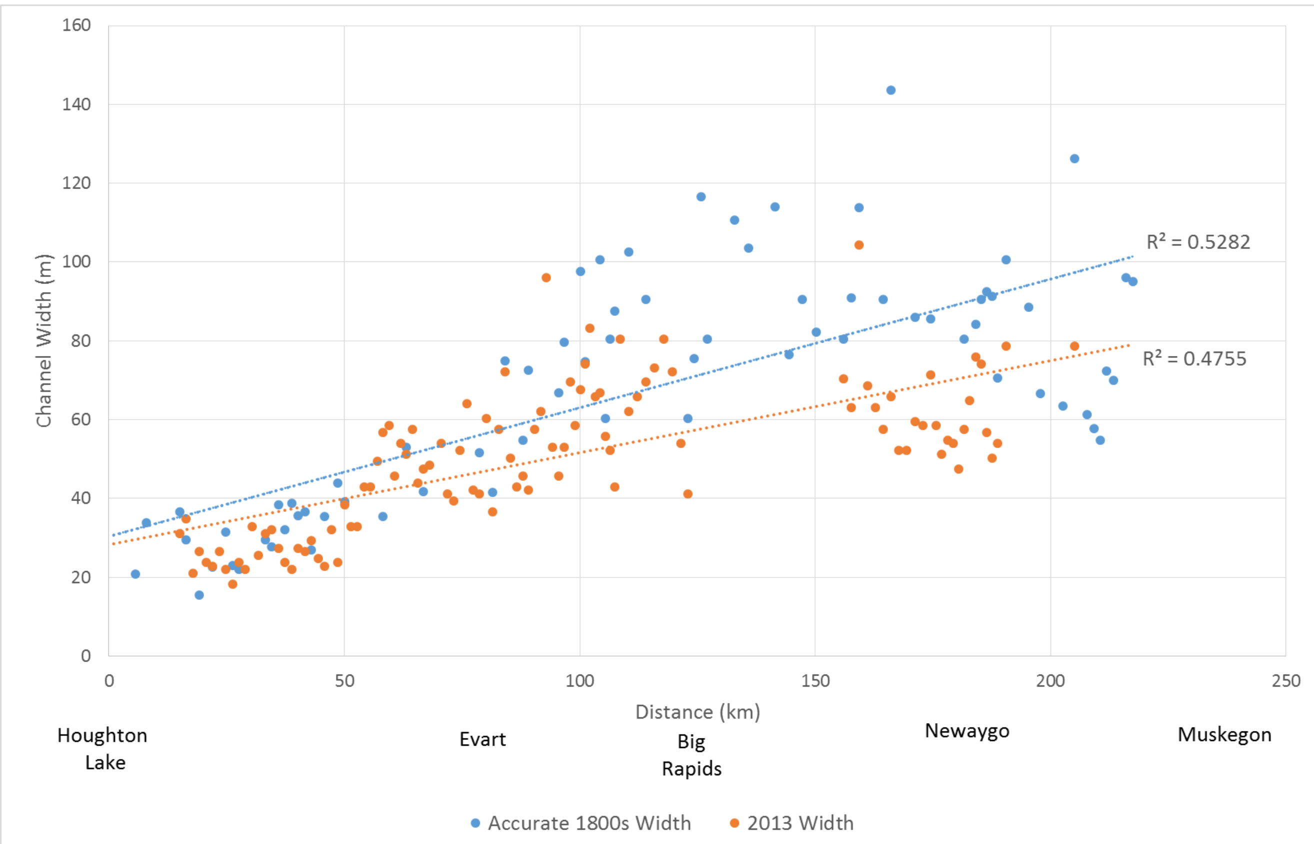

Figure 5.25. Channel widths recorded on the Muskegon River during the initial survey and the summer of 2013.

remained relatively consistent in this portion of the stream during the historic record (Figure 5.25).

As the Muskegon River flows past Evart and into Mecosta County, the width of the channel begins to fluctuate across space. This is likely due to the presence of several riffle-and-pool sequences, which causes the channel width to fluctuate between ~40 and 100 m (Figures 5.22 and 5.23). In addition, a number of other transitions occur in this reach. As previously stated, the channel becomes less sinuous in the middle reach than in the opposing reaches. This reach is also where the first apparent change in channel width occurs. For instance, widths collected in the summer of 2013 appear to be narrower to those taken in the mid-1800s (Figure 5.25). This narrowing trend continues throughout Mecosta County well past Big Rapids.

Some of the greatest widths recorded throughout the extent of the Muskegon River are between the towns of Big Rapids and Newaygo. Here initial survey widths indicate that the channel ranged between ~75 and 140 m wide (Figure 5.22). Conversely, width data collected during the summer of 2013 could not be measured in this reach until after the Croton Dam. Below Croton Dam, the Muskegon River returns to a meandering system and widespread channel narrowing was detected. Widths in this reach have apparently narrowed by up to ~30 to 20 m in some locations (Figure 5.25). The degree of channel narrowing appears to decrease as the stream flows toward Bridgeton (e.g., Warner Avenue), about 35 km downstream of Croton Dam (Figure 5.25). Here channel widths have remained stable during the historic record.

As the Muskegon River flows into Muskegon County, widths remain highly variable, ranging between ~55 to 125 m wide (Figure 5.25). While some channel widths remain relatively high, the Muskegon River appears to decrease in channel width near its terminus at Muskegon Lake. This is likely due to a change in channel pattern, from a sinuous stream to anastomosing, which allows the stream to distribute its overall flow into several distributary channels. This portion of the stream was not surveyed in the summer of 2013 because the main channel had been straightened and dredged, affecting the natural character of the stream. Also, survey records only measured one channel in this reach, making it difficult to recognize which branch of the stream they surveyed. Data collected at Maple Island Road during the summer of 2013 suggests that the Muskegon River has remained relatively stable in this location during the historic record.

5.5 Discussion

5.5.1 Introduction

Study results indicate that a variety of channel changes have occurred along the Muskegon River in the historic period. These changes include abandoned meanders, adjustments to channel pattern and meander shape, and channel narrowing. The cause of these channel changes could be due to a number of different variables which have or are still occurring in the stream system. This section is outlined to first discuss 1) the possible causes for channel change during the historic record, followed by 2) a discussion about the likely cause for channel narrowing in the lower reach.

5.5.2 Causes for Historic Channel Change in the Muskegon River

Previous research regarding historic channel changes in the U.S. has shed light on the possible causes for alterations within many fluvial systems. Studies conducted by

Schumm and Lichty (1963), Burkham (1972), Knox (1977), Martin and Johnson (1987), Webb et al. (1991), Juracek (2002), Dean and Schmidt (2011), and Swanson et al. (2010) suggested that stream systems are sensitive to climatic fluctuations, such as large precipitation events and drought. Similar work by Bryan (1928), Happ (1944), Magilligan (1985), Webb et al. (1991), Ruhlman and Nutter (1999), Juracek (2002), Galster et al. (2008), and Trimble (2009) suggest that changes in landcover and landuse have also caused channel alterations by changing sedimentation rates. During the historic record north-central Michigan has also experienced similar modifications in both landcover/landuse and climate. For example, data from the Midwestern Regional Climate Center (2014) dating back to 1895 shows that annual precipitation for the center of the Muskegon River basin (i.e., Evart, MI) has been increasing since around the 1930s (Figure 5.26). While this may have slightly added to the water budget of the Muskegon River, it is likely that this increase in precipitation has been offset by irrigation throughout the basin (e.g., as the basin became revegetated and agriculture increased; He, 1999).

In addition to the increase of precipitation, the Muskegon River basin also experienced several dramatic alterations in landcover/landuse during the historic record. The first of these alterations occurred during the late 1800s and early 1900s when the basin was intensively logged. This devegetated the surrounding landscape and channel banks allowing sediment to become easily eroded. Studies conducted by Happ (1944), Knox (1977), Magilligan (1985), and Trimble (2009) in the Driftless Area, indicated a period of rapid sedimentation likely occurred within stream valleys during this time period. Because the Muskegon River basin is mainly composed of sand, this filling episode may have been less evident than in the Driftless Area where the predominate soil texture is silt sized (e.g., the potential for sediment transport decreases as grain size increases; Knighton, 1998). Nevertheless, the Muskegon Rivers sandy channel banks would have likely collapsed or fell into the stream as bank-holding vegetation was clear-cut, like that identified on the lower Pine River by Hansen (1971).

Major logging operations ended in much of Michigan around 1910 when timber stands became largely depleted (Alexander, 2006). This cessation of logging allowed much of the Muskegon River basin to become revegetated during the early to mid-1900s (Alexander, 2006; Ray et al., 2012; Figure 5.27). Channel banks likely became increasingly stable during this time as they too slowly came to be revegetated (Figure 5.27). Since this time period, landuse changes started to diversify as agriculture and urbanization (e.g.,

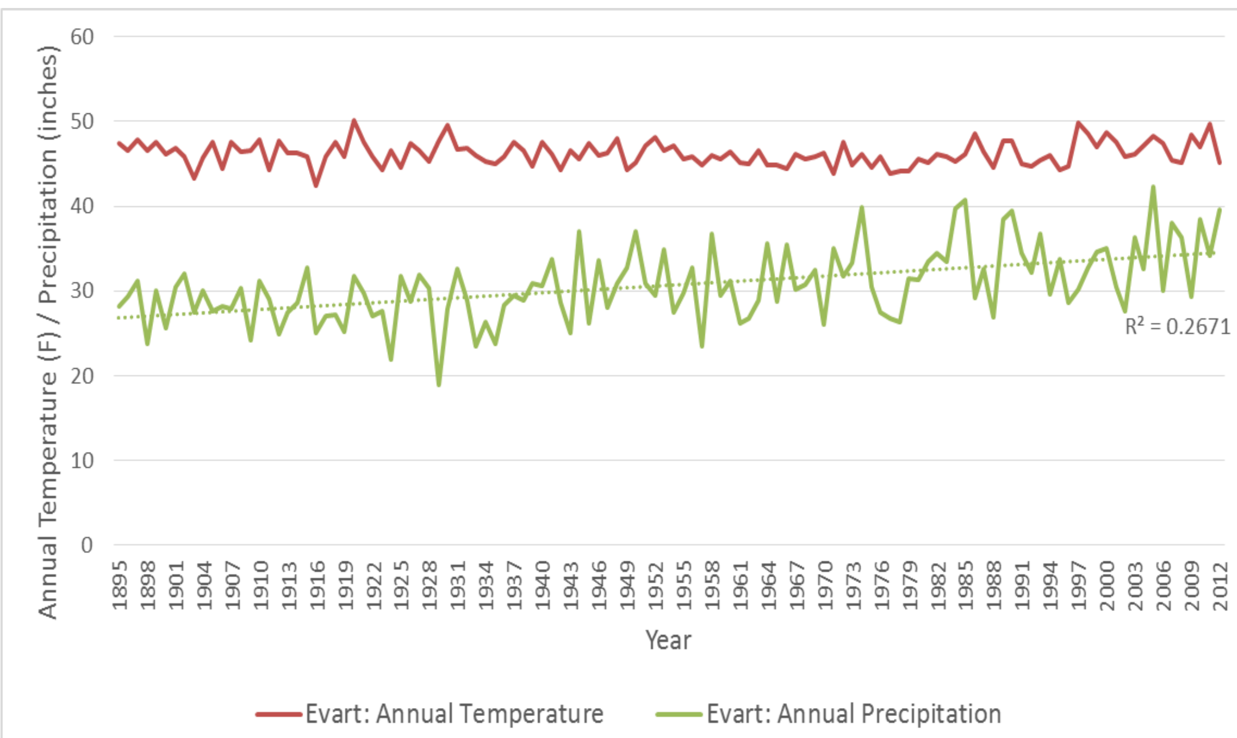

Figure 5.26. Annual temperature and precipitation at Evart since 1895. (adopted from Midwestern Regional Climate Center, 2014)

Figure 5.27. The revegetation of a channel reach in Evart from 1919 (left: unknown) to 2013 (right; photographed by author).

growing cities, roadways, and infrastructure) took hold within the basin (Ray et al., 2012). Today, the Muskegon River basin is predominantly clustered with forested and agricultural landuses (Ray et al., 2012).

The specific impacts that these rapid changes in landcover/landuse had on the Muskegon River are difficult to determine. It is conceivable that during Michigan's logging era these changes accelerated sedimentation rates within the fluvial system, in turn potentially causing channel narrowing, deposition and/or sediment reworking (Webb-Sullivan and Evans, 2015). As in several studies conducted in the Central Great Plains and American Southwest (e.g., Graf, 1978; Johnson, 1994; Dominick and O'Neill, 1998; Grams and Schmidt, 2002; McBride et al., 2010), the reintroduction of riparian vegetation to channel banks during the early to mid-1900s may have caused the Muskegon River to narrow as the channel boundary became increasingly stable. Similarly, Webb-Sullivan and Evans (2015) noticed rapid sedimentation within the Ottawa River valley of Michigan-Ohio between the early-1950s and the 1980s, which they associated with the streams sediment supply greatly exceeding the sediment conveyance capacity (Webb-Sullivan and Evans, 2015).

Another factor to consider when analyzing historic channel changes is stream power, which is a product of both stream discharge and slope (i.e., $P = Q * S$; Charlton, 2008). A reduction of stream power in some reaches may initiate channel migration in a system, in time creating several abandoned or extended meanders. For example, a study by Chang (1979) suggested that meander development in sand-bed channels is commonly associated with a minimum stream power per unit channel length in the system. Stream power was calculated for reaches throughout the entire Muskegon River and varied considerably. The highest stream power estimates were recorded between Paris and Newaygo (Figure 5.28) where the stream exhibits some of its lowest sinuosity values. As previously discussed, this reach of the stream has a high gradient compared to the rest of the system. A high gradient coupled with an increasing discharge allows the Muskegon River to become quite powerful in this reach of the system. In contrast, the upper (i.e., Houghton Lake to Evart) and lower reaches (i.e., Maple Island Rd. to Muskegon) of the Muskegon River record relatively low estimated stream power values (Figure 5.28). As previously noted, these portions of the Muskegon River have experienced the most historic channel adjustment, suggesting that low stream power values are associated with channel meandering and the formation of abandoned meanders.

Since discharge in an active stream system does not flow at a constant rate, it is possible for stream power values to fluctuate. For example, the annual mean discharge records for both Evart (from 1936-2013) and Croton (from 1995 to 2013) show that year to year flows are variable (Figure 5.29). While this variability would cause stream power to fluctuate annually, there is only a small increase in long-term mean annual discharge. For instance, discharge at Evart has slowly risen from ~950 cfs in 1936 to~1,200 cfs

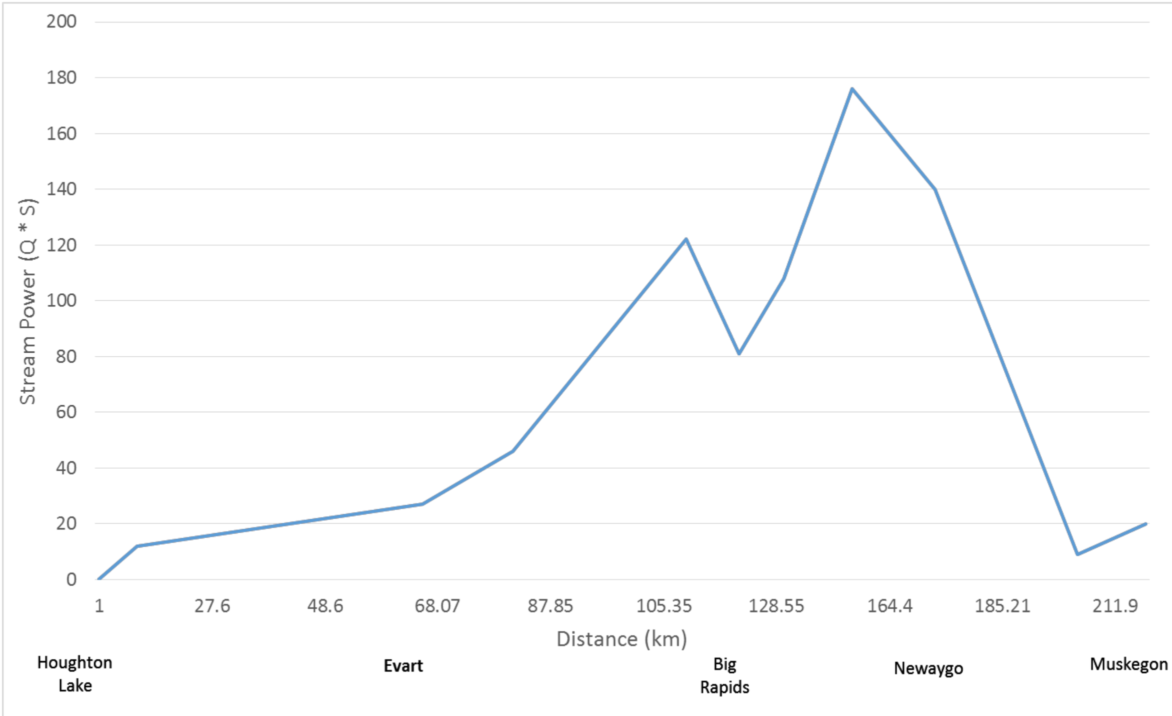

Figure 5.28. Present-day estimated stream power for the entire Muskegon River.

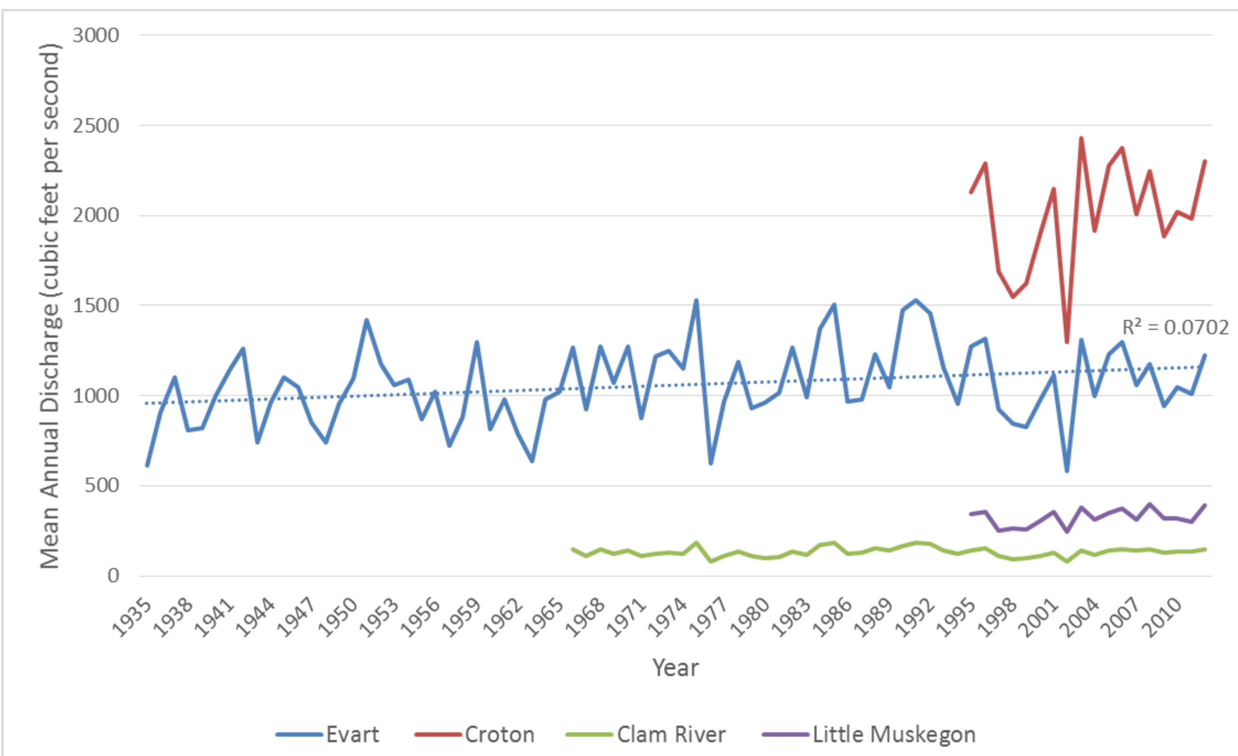

Figure 5.29. Annual mean discharge records of gauges in the Muskegon River basin. (adopted from USGS, National Water Information System: Web Interface)

in 2013 (Figure 5.29), indicating that stream power has also increased slightly. Although stream power has increased, estimated stream power values are still relatively low, suggesting that it likely is not the sole cause of historic channel adjustment in the upper and lower reaches of the stream.

The final and likely cause for historic changes in the Muskegon River, especially in the upper system, is related to channel slope and sinuosity. As previously discussed by Arbogast (2008), the upper reach of the Muskegon River has been actively meandering for thousands of years, suggesting that channel alterations have been occurring before the historic record. This history illustrates that the main factor causing the stream to regularly change its morphology is a variable that has remained fairly constant, such as slope. When looking at the longitudinal profile of the Muskegon River one can see that the upper and lower portions of the stream have a relatively low gradient compared to the middle reach of the stream (Figure 5.2). These areas of low slope can be directly linked with rises in channel sinuosity (a dependent variable of slope) in the upper and lower system (Figure 5.30).

When the location of cutoff meanders were plotted in conjunction with slope and sinuosity a general pattern was noticed. Cutoffs commonly occurred in stream reaches with a sinuosity above ~1.8 and a slope averaging ~0.3 m per km or less (Figure 5.30). These conditions only exist in the upper system between Houghton Lake and Evart and upstream from Newaygo (Figure 5.30). This finding likely explains why abandoned meanders recorded during the historic record are only found in these parts of the system. While values are occasionally greater than ~1.8 in the middle and lower reaches of the stream, sinuosity channel slope is too high to allow more active meandering to occur. Conversely, it

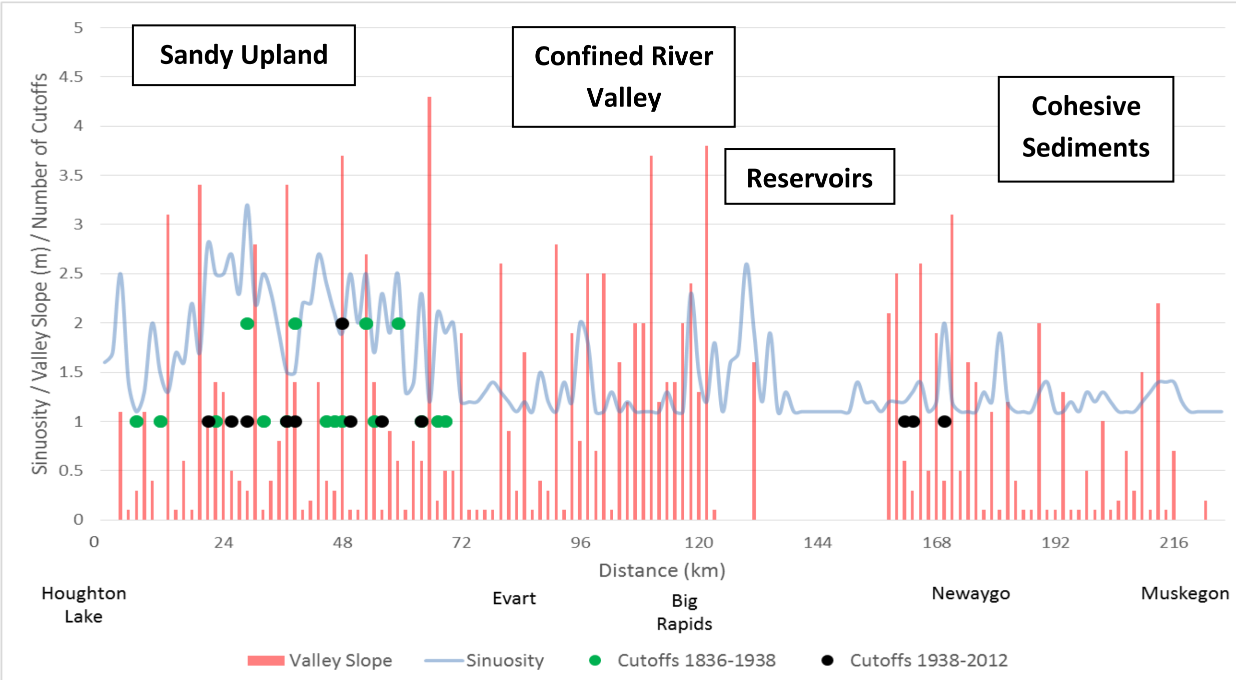

Figure 5.30. Average of sinuosity and channel slope measurements for the entire reach of the Muskegon River. Black dots represent meanders cutoffs that occurred between the initial survey and 1938, whereas the red dots represent cutoffs forming between 1938 and 2012. Note the character of the stream in the sandy upper reach, confined middle section, and on the silty lake-plain sediments of the lower reach.

is conceivable that meander translation, extension, and rotation has occurred in these reaches as the stream attempts to distribute its energy across the floodplain (Figure 5.19; Chen and Duan, 2006).

5.5.3 Potential Channel Narrowing Linked to Damming

Dams have been shown to impact channel flow and sediment transport in a system. Such modifications impact channel morphology and the sediment transport within a system (Van Steeter and Pitlick, 1998; VanLooy and Martin, 2005; Ibisate et al., 2013). Sediments in the sand fraction tend to be deposited behind the dam as bedload (Figure

5.31), whereas finer silts and clays, may pass through the obstruction as suspended load (Childs, 2010).

Downstream channel degradation commonly occurs in systems that have had their upstream bedload load trapped behind a dam. When regulated flow is released from the dam, the flow is relatively free of coarse sediment. This allows the now “clean water” to transport sediments below the dam (Childs, 2010), potentially widening or deepening the channel. The amount of degradation is dependent on the amount of regulated flow and sediment that leaves the dam. Degradation aims to reduce channel slope as the stream attempts to reduce the velocity of its flow (Knighton, 1998). Once the system reaches a quasi-equilibrium, degradation comes to an end (Knighton, 1998; Julien, 2002).

In contrast, aggradation below dams may happen through a number of circumstances. While degradation commonly occurs directly below dams, aggradation typically occurs further downstream until the stream reaches equilibrium (Petts, 1979; Petts, 1980; Kondolf, 1997; Childs, 2010; Ibisate et al., 2013; Figure 5.32). An entering tributary or another dam/obstruction downstream can cause further aggradation, causing prolonged channel narrowing (Petts, 1980; Figure 5.32). Dams on large river systems may also have an influence on annual or periodic flood peaks, which actively erode stream banks. As a result, riparian vegetation commonly encroaches these banks in response to a reduction in flood scour (Wilcock et al., 1996; Kondolf, 1997; Ibisate et al., 2013). Assuming that this processes continues through time, it may result in a narrower and/or shallower channel beneath the dam (Wilcock et al., 1996; Kondolf, 1997; Ibisate et al., 2013).

Four dams are currently in use on the Muskegon River (Figure 3.1). Three of the four dams are owned and operated my Consumers Power Company in the lower basin

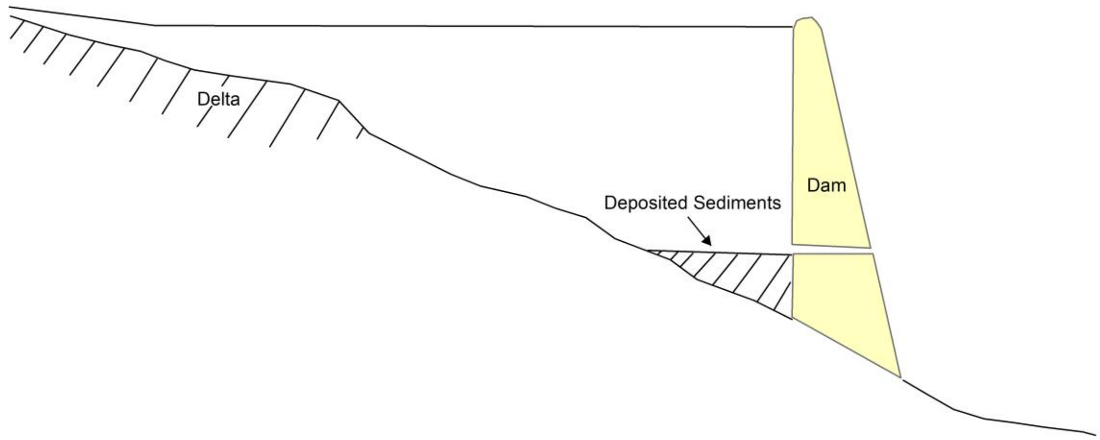

Figure 5.31. Diagram illustrating coarse bedload being deposited behind a dam as dead storage, while suspended fines are released as regulated flow. (modified from Childs, 2010)

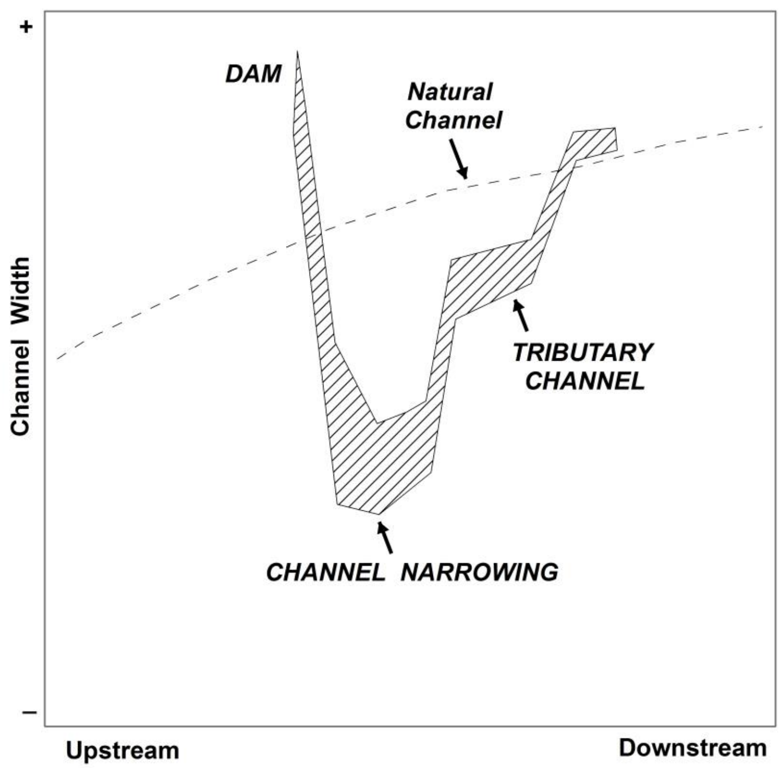

Figure 5.32. Channel responses to stream damming. Note how the channel narrows in response to stream damming. Prolonged narrowing may occur below dams at the confluence of tributary channels, as additional sediment enters the system. (modified from Petts, 1980)

between the towns of Big Rapids and Newaygo, Michigan. All of these earthen hydroelectric dams have large retaining reservoirs upstream. The most headword of the dams is Rogers Dam (constructed in 1906) in Mecosta County. Rogers Dam is strictly a "water in – water out" facility releasing all the incoming water downstream (Consumers Power Company, 2014). Rogers Dam is maintained normally at a "full pool stage", using all excess water flow passed as spill flow for power generation (Consumers Power Company, 2014).

The next dam downstream of the Rogers Dam is the Hardy Dam (constructed in 1931), which is a peaking facility. Hardy Dam is only used for power generation during the morning and afternoon, while flow is shut down for the evening and overnight hours (Consumers Power Company, 2014). The daily total flow from Hardy Dam is allowed to fluctuate up to one foot over a 24 hour period (Consumers Power Company, 2014). The lowermost dam on the Muskegon River is the Croton Dam (constructed in 1907; Figure 5.33) in Newaygo County. Croton Dam is operated to re-regulate the on/off flow from the Hardy Dam upstream to river type conditions (Consumers Power Company, 2014). The pond is brought down during the evening and nighttime hours to provide room for daily Muskegon and Little Muskegon River flows (Consumers Power Company, 2014). In the case the river flow at Croton Dam exceeds the need for power generation, the excess water is allowed downstream (Consumers Power Company, 2014).

The fourth and final dam on the Muskegon River is located in the upper basin in Missaukee County (Figure 3.1). Reedsburg Dam (constructed in 1941; Figure 5.34) is owned and operated by the Michigan Department of Natural Resources for recreational and wildlife restoration purposes. The dam floods a marshy area with around 30,000 acres of water coming out of nearby Houghton Lake. Reedsburg Dam is an earthen dam with a

Figure 5.33. Croton Dam in central Newaygo County. (photographed by author)

Figure 5.34. Reedsburg Dam in northern Missaukee County. (photographed by author)

Flow regulation facility constructed from concrete and wood. It functions as a “water in-water out” structure (Michigan Department of Natural Resources, 2014).

Several other dams have been placed in the Muskegon River during the historic record, but were later decommissioned and removed. These dams were originally used during the logging era to allow companies to stamp and sort logs heading to neighboring mills. Dams, such as the Tioga Dam (existed in Big Rapids from 1866 to 1907), were commonly built out of wood and were “water in – water out” structures (Alexander, 2006). The latest dam to be deconstructed on the Muskegon River was the Big Rapids Dam. While the Big Rapids Dam was only in use from 1912 to 1966 (Alexander, 2006), it was not fully removed from the river channel until 2001.

Because the Muskegon River has been extensively dammed during the historic record it is conceivable that alterations in channel morphology have occurred. When comparing channel widths from the initial land survey (mid-1800s) to the summer of 2013 a pattern is noticeable. Results indicate that channel widths appear to have narrowed considerably below both Big Rapids (i.e., the past Tioga and Big Rapids Dams) and the Croton Dam, suggesting that stream aggradation has occurred in these reaches during the historic record (Figure 5.35). A similar pattern was discussed by Petts (1980), who noticed channel narrowing below a dam in a study conducted on the Camps Reservoir in Scotland (Figure 5.32). Petts (1980) noticed that channel erosion (e.g., degradation) occurred within 250 m directly below the dam. This area of erosion quickly transitioned into a zone of aggradation, which Petts (1980) attributes to a highly sinuous channel and the nearby impoundment. This trend then slowly returned to normal as the channel became increasingly stable (Petts, 1980).

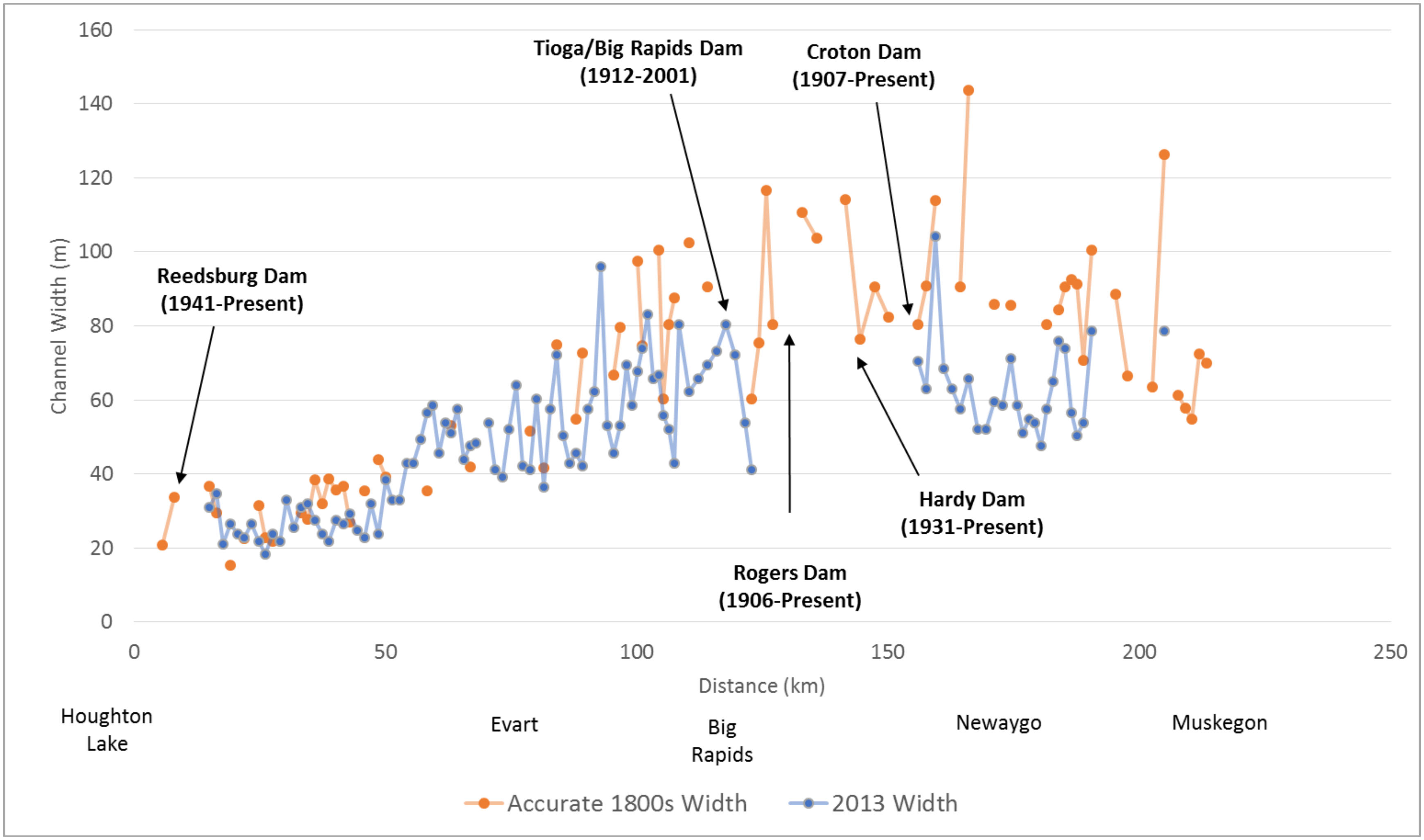

Figure 5.35. Diagram showing 1800s and 2013 channel widths and channel narrowing below past and present dams.

Similarly, widths on the Muskegon River seemed to return to the historical average ~35 km downstream from the last impoundment (Figure 5.35), suggesting that the dams may be playing a role in channel narrowing. In addition, the lower Muskegon River is a meandering stream, which may add to the local instability of the system below these impoundments. The timing of dam placement may have also have played a role in the local instability of these stream reaches. As previously discussed, the large earthen dams in the lower basin were constructed between 1906 and 1931, which was during the last part of Michigan's logging era. Channel banks during this time period were likely devegetated, allowing sediment to fill a channel that was experiencing lower flows due to damming. In addition to devegetated banks slopes, several large tributary channels (e.g., Little Muskegon River, Hersey River, and Clam River) may have also potentially increased the sediment yields of the Muskegon River during this time.

Conversely, channel narrowing is absent below Reedsburg Dam in the upper system (Figure 5.35). This lack of change may be due to a couple of factors. First, the Reedsburg Dam is a "flow in – flow out" structure, suggesting that there has been little change in stream discharge. Secondly, the structure was not constructed until 1941. This late construction date would have allowed the surrounding landscape to become revegetated, likely causing less bank erosion and channel aggradation. Thirdly, the upper system has fewer dams than the mid/lower reaches of the Muskegon River. The added effect of several of these structures may have accelerated or caused further impacts in the lower reach. Fourthly, the upper reach has very few tributary channels compared to the lower system, which may have potentially lessened the rate of sedimentation during the early-1900s, in turn minimizing channel narrowing. Finally, the upper Muskegon River was resurveyed between 1845 and 1856 (Thomas, 2009), around 10 to 20 years after the initial land survey, which potentially allowed sufficient time for the stream to change from its pre-settlement conditions. This scenario suggests that basin wide alterations would have occurred in a relatively short time period (e.g., between the mid-1830s and 1856) during initial settlement and has remained relatively stable since.

CHAPTER 6

CONCLUSIONS

Historic channel changes have been documented throughout much of the United States in areas such as the American Southwest, central Great Plains, Driftless Area of the Upper Midwest, Atlantic Coast, and Pacific Coast (e.g., Bryan, 1928; Happ, 1944; Schumm and Litchty, 1963; Burkham, 1972; Knox, 1977; Graf, 1978; Magilligan, 1985; Martin and Johnson, 1987; Webb et al., 1991; Johnson, 1994; Dominick and O'Neill, 1998; Van Steeter and Pitlick, 1998; Ruhlman and Nutter, 1999; Grams and Schmidt, 2002; Juracek, 2002; VanLooy and Martin, 2005; Galster et al., 2008; Ghoshal et al., 2010; McBride et al., 2010; White et al., 2010; Swanson et al., 2010; Dean and Schmidt, 2011; Skorko et al., 2011; Wallick et al., 2011; Horn et al., 2012; Lecce, 2013; Mossa, 2013). These studies have suggested that several variables have led to channel change during the historic record, such as fluctuations in climate, riparian vegetation, landcover/landuse, discharge, and sedimentation rates. Many of these alterations of the landscape, such as deforestation, pasturing/agriculture, stream damming, and irrigation have been initiated by human modifications to a landscape and/or fluvial system.

In contrast to the work done elsewhere, relatively few studies (e.g., Bowman, 1904; Hansen, 1971; Burroughs et al., 2009; Webb-Sullivan and Evans, 2015) have been conducted in Michigan focusing specifically on fluvial geomorphology. These studies have only targeted a few isolated stream segments within the state, instead of an entire stream system. One such example is Hansen (1971), who noted excessive sedimentation in the Pine River during the early part of the 1900s, which he suggested was due to increased runoff and bank erosion after Michigan's logging era. A later study by Burroughs et al. (2009) indicated that dam removal on the lower Pine River had caused ongoing headward erosion upstream of the former dam location. Similar research by Arbogast et al. (2008) in the upper reach of the Muskegon River, in the north-central part of Lower Michigan, has documented several fluvial adjustments in the system since the Late Pleistocene. Results from Arbogast et al. (2008) indicate that the system has been actively meandering on its current T-1 surface, suggesting that channel alterations may still be occurring during the historic record.

This research is aimed at answering that question and is the first to document historic channel adjustment throughout the entire expanse of a stream system in Michigan. Results from this study indicate that the Muskegon River has three main stream reaches. The upper reach flows on a fairly low gradient, producing a highly meandering system with sinuosities reaching as high as ~3.2 in some locations. As a result, several abandoned meanders have formed in the upper reach between Houghton Lake and Evart during the historic record (e.g., between the 1840s and 2012). These cutoffs formed in reaches of the stream that have a sinuosity > ~1.8 and a continually low gradient of ~0.3 m per km. In addition, channel width, radius of curvature, meander wavelength, and amplitude values are much smaller than those found in the middle and lower reaches. This is likely associated to the relationship between channel width and meander wavelength, which is around 10 to 14 times bankfull width (Chorley et al., 1984).

Downstream of the upper reach the channel gradients are greater (e.g., averaging ~0.98 m per km), causing the Muskegon River to become less sinuous. As a result, several riffle and pool sequences form, due to alternating energy losses/gains. Larger channel widths are also found in conjunction with this riffle and pool sequence in the middle reach. Wider stream segments aid in the formation of several longitudinal bars, which form in the middle of the channel as fine sediments deposit due to low flow. Similar to channel width, radius of curvature, meander wavelength, and amplitude values are also larger in the middle reach. This is likely related to the increase in channel width, which in turn increased the meander wavelength.

In the lower reach of the Muskegon River channel gradients drop as the stream flows on a relatively low sloping lake-plain, producing a rise in channel sinuosity, which reaches up to ~2.0 in some locations. In turn, this drop in gradient has formed a number of abandoned meanders, while also causing several meanders to extend and change shape during the historic record. Like the middle reach, channel widths in the lower reach mostly range between ~45 and 80 m. Conversely, radius of curvature, meander wavelength, and amplitude values slightly decrease in the lower Muskegon River system. This is likely related to lower channel gradients and higher sinuosity values. In addition to these adjustments in meander geometry, channel pattern also changes in the lower reach near Muskegon Lake. In this part of the stream the Muskegon River exhibits an anastomosing pattern, which is likely due to finer sediments found in this flood-dominated area. These fine sediments aid in the stabilization of channel banks, in turn allowing the channel boundary to become more resistant to erosion. As a result of the anastomosing channel pattern, widths decrease in this portion of the stream.

During the historic record the Muskegon River basin experienced several human induced modifications which likely caused the stream to change. The first of these modifications occurred during Michigan's logging era between 1837 and 1910, when widespread clear-cutting occurred in the basin. During this time, the river was used to transport large logs downstream to lumber mills. These practices likely increased the rates of sedimentation and channel bank erosion along the expanse of the Muskegon River and its tributaries.

Shortly after Michigan's logging era, several large dams were placed throughout the lower basin (e.g., 1906-1931). These dams formed several large ponds and also affected natural stream discharge and flow rates. During this time, high erosion rates paired with lower stream flows may have caused widespread channel aggradation in the lower system, like that seen in the Driftless Area (e.g., Happ, 1944; Knox, 1977; Magilligan, 1985; Trimble, 2009). This potentially caused the lower Muskegon River to narrow downstream from these dam locations where discharge rates were most effected. These dams also likely lessened the long-term effect of annual flood scour upon channel banks. This reduction in flood scour likely allowed encroaching riparian vegetation to stabilize near the stream boundary, in turn narrowing channel banks (Wilcock et al., 1996; Kondolf, 1997; Ibisate et al., 2013). As a result, channel narrowing has been recorded in the lower reach of the Muskegon River where stream damming has effected natural channel flows.

6.1 Contributions of this Research

This research has provided three main contributions. First, this is the first study to characterize the hydraulic geometry of a stream in Michigan. Secondly, this work has analyzed historic channel change throughout an entire stream system, identifying areas where historic channel change has occurred. For example, the meander pattern of the upper reach of the Muskegon River has changed frequently, while the appearance of the middle and lower reaches have remained relatively unchanged during the historic record, with the exception of channel changes downstream of Croton Dam. This difference is largely due to variations in channel slope and sinuosity, which is related to the geomorphology of the underlying surface. The third contribution of this study is the identification of widespread channel narrowing in the lower Muskegon River basin. Prior work by Hansen (1971) on the Pine River demonstrated that dams can act as a sediment-trap, however no study in Michigan has suggested that dams could be linked with channel narrowing. In contrast, this study suggests that historic channel narrowing in the lower Muskegon River basin is linked to dam placement and/or devegetated bank slopes and tributary channels that have potentially increased sediment yields.

APPENDIX

Table A.1: Average Radius of Curvature, Wavelength, Amplitude, and Stream Order data for the Muskegon River.

Section Boundary	Township	County	Average Radius of Curvature (m)	Wave-Length (km.)	Amplitude (km.)	Stream Order
Sec 3 - Sec 10	Lake	Roscommon	-	-	-	1
Sec 3 - Sec 4	Lake	Roscommon	54.45	0.55	0.14	1
Sec 9 - Sec 8	Lake	Roscommon	54.45	0.35	0.32	2
Sec 8 - Sec 17	Lake	Roscommon	54.45	0.21	0.31	3
Sec 25 - Sec 36	Enterprise	Missaukee	75.35	-	-	3
Sec 2 - Sec 3	Butterfield	Missaukee	53.13	0.21	0.21	3
Sec 3 - Sec 10	Butterfield	Missaukee	53.13	0.27	0.16	4
Sec 10 - Sec 15	Butterfield	Missaukee	46.85	0.31	0.21	4
Sec 22 - Sec 27	Butterfield	Missaukee	46.85	0.21	0.19	4
Sec 27 - Sec 34	Butterfield	Missaukee	46.85	0.26	0.27	4
Sec 34 - Sec 3	Butterfield-Holland	Missaukee	57.9	0.27	0.31	4
Sec 2 - Sec 11	Holland	Missaukee	57.9	0.24	0.21	4
Sec 18 - Sec 19	Roscommon	Roscommon	47.89	0.29	0.23	4
Sec 19 - Sec 30	Roscommon	Roscommon	47.89	0.27	0.24	4
Sec 30 - Sec 31	Roscommon	Roscommon	50.82	0.32	0.29	4
Sec 1 - Sec 2	Summerfield	Clare	50.82	0.39	0.42	4
Sec 10 - Sec 15	Summerfield	Clare	52.06	0.34	0.24	4
Sec 17 - Sec 20	Summerfield	Clare	58.45	0.37	0.35	4
Sec 34 - Sec 3	Winterfield-Redding	Clare	72.92	0.61	0.66	5
Sec 9 - Sec 10	Redding	Clare	89.38	0.56	0.64	5
Sec 16 -Sec 21	Redding	Clare	89.38	0.92	0.29	5
Sec 1 - Sec 12	Sylvan	Osceola	103.2	0.53	0.34	5
Sec 21 - Sec 20	Sylvan	Osceola	142.29	0.51	0.4	5
Sec 31 - Sec 32	Hersey	Osceola	304.09	2.66	1.22	5
Sec 12 - Sec 11	Green	Mecosta	231.28	-	-	5
Sec 15 - Sec 22	Green	Mecosta	472.16	-	-	5
Sec 34-Sec 35	Green	Mecosta	876.41	-	-	5
Sec 23 - Sec 24	Big Rapids	Mecosta	313.17	2.14	1.66	5
Sec 36 - Sec 1	Big Rapids - Mecosta	Mecosta	455.84	-	-	5
Sec 11 - Sec 14	Mecosta	Mecosta	392.16	1.83	1.21	5
Sec 16 - Sec 21	Mecosta	Mecosta	392.16	2.58	1.93	5
Sec 20 - Sec 29	Mecosta	Mecosta	315.11	2.32	1.04	5

Table A.1: (cont'd)

Section Boundary	Township	County	Average Radius of Curvature (m)	Wave-Length (km.)	Amplitude (km.)	Stream Order
Sec 6 - Sec 1	Aetna-Big Prairie	Mec.-New.	908.61	-	-	5
Sec 32 - Sec 5	Big Prairie - Croton	Newaygo	455.86	-	-	5
Sec 14 - Sec 23	Brooks	Newaygo	182.13	0.76	0.51	6
Sec 22 - Sec 21	Brooks	Newaygo	182.13	1.2	0.55	6
Sec 19 - Sec 24	Brooks-Garfield	Newaygo	237.81	0.48	1.35	6
Sec 35 - Sec 36	Garfield	Newaygo	199.1	0.62	0.46	6
Sec 34 - Sec 33	Garfield	Newaygo	199.1	1.16	0.7	6
Sec 5 - Sec 6	Ashland	Newaygo	316.56	0.92	1.19	6
Sec 7 - Sec 12	Ashland-Bridgeton	Newaygo	316.56	0.87	0.76	6
Sec 21 - Sec 20	Bridgeton	Newaygo	389.59	0.54	0.71	6
Sec 30 - Sec 25	Bridgeton-Cedar Creek	New.-Musk.	173.93	-	-	6
Sec 35 - Sec 34	Cedar Creek	Muskegon	197.62	0.2	0.32	6
Sec 10 - Sec 9	Muskegon	Muskegon	114.61	0.45	0.51	6
Sec 8 - Sec 17	Muskegon	Muskegon	126.76	-	-	6

Table A.2: Slope, Sinuosity, and Meander Cutoff data for the Muskegon River.

Location	Distance (km.)	Elevation (m)	Slope (m/km)	Sinuosity	Cutoffs (1840s-1938)	Cutoffs (1938-2012)
Houghton Lake	0	348.9	-	-	-	-
	1.6	347.9	1	1.6	0	0
	3.2	347.9	0	1.7	0	0
	4.8	346.8	1.1	2.5	0	0
	6.4	346.7	0.1	1.4	0	0
	8	346.4	0.3	1.1	1	0
Reedsburg Dam	9.6	345.3	1.1	1.3	0	0
	11.2	344.9	0.4	2	0	0
	12.8	344.9	0	1.5	1	0
M-55	14.4	341.8	3.1	1.3	0	0
	16	341.7	0.1	1.7	0	0
	17.6	341.1	0.6	1.6	0	0
	19.2	341	0.1	2.2	0	0
	20.8	37.6	3.4	1.7	0	0
	22.4	336.6	1	2.8	0	1
	24	35.2	1.4	2.5	1	0
	25.6	333.9	1.3	2.5	0	0
	27.2	333.4	0.5	2.7	0	1
	28.8	333	0.4	2.3	0	0
	30.4	332.7	0.3	3.2	2	1
	32	329.9	2.8	2.2	0	0
	33.6	329.8	0.1	2.5	1	0
	35.2	329.4	0.4	2.3	0	0
	36.8	328.6	0.8	1.9	0	0
	38.4	325.2	3.4	1.5	1	1
	40	323.8	1.4	1.5	2	1
	41.6	323.7	0.1	2.2	0	0
	43.2	323.5	0.2	2.2	0	0
	44.8	322.1	1.4	2.7	0	0
	46.4	321.7	0.4	2.4	1	0
	48	321.4	0.3	2.1	1	0
	49.6	317.7	3.7	1.9	1	2
	51.2	317.6	0.1	2.5	0	1
	52.8	317.5	0.1	2	0	0

Table A.2: (cont'd)

Location	Distance (km.)	Elevation (m)	Slope (m/km)	Sinuosity	Cutoffs (1840s-1938)	Cutoffs (1938-2012)
	54.4	314.8	2.7	2.5	2	0
	56	313.4	1.4	1.7	1	0
M-61	57.6	313.3	0.1	2.3	0	1
	59.2	312.4	0.9	1.9	0	0
	60.8	311.8	0.6	2.5	2	0
	62.4	311.7	0.1	1.3	0	0
	64	310.9	0.8	1.4	0	0
M-115	65.6	310.3	0.6	2.3	1	1
	67.2	306	4.3	1.2	0	0
	68.8	305.8	0.2	2.1	1	0
M-66	70.4	305.3	0.5	1.9	1	0
	72	304.8	0.5	2	0	0
	73.6	302.9	1.9	1.2	0	0
	75.2	302.8	0.1	1.2	0	0
	76.8	302.7	0.1	1.2	0	0
	78.4	302.6	0.1	1.3	0	0
	80	302.5	0.1	1.4	0	0
Evart / US-10	81.6	299.9	2.6	1.3	0	0
	83.2	299	0.9	1.2	0	0
	84.8	298.7	0.3	1.1	0	0
	86.4	297	1.7	1.2	0	0
	88	296.9	0.1	1.1	0	0
	89.6	296.5	0.4	1.5	0	0
	91.2	296.2	0.3	1.2	0	0
	92.8	293.4	2.8	1.1	0	0
	94.4	293.3	0.1	1.4	0	0
	96	291.4	1.9	1.2	0	0
Hersey	97.6	290.6	0.8	2	0	0
	99.2	288.1	2.5	1.8	0	0
	100.8	287.4	0.7	1.1	0	0
	102.4	284.9	2.5	1.1	0	0
	104	284.8	0.1	1.3	0	0
	105.6	283.2	1.6	1.1	0	0
	107.2	282	1.2	1.2	0	0

Table A.2: (cont'd)

Location	Distance (km.)	Elevation (m)	Slope (m/km)	Sinuosity	Cutoffs (1840s-1938)	Cutoffs (1938-2012)
Paris	108.8	280	2	1.1	0	0
	110.4	278	2	1.1	0	0
	112	274.3	3.7	1.1	0	0
	113.6	273.1	1.2	1.1	0	0
	115.2	271.7	1.4	1.3	0	0
	116.8	270.3	1.4	1.1	0	0
	118.4	268.3	2	1.1	0	0
	120	265.9	2.4	2.3	0	0
Big Rapids / M-20	121.6	264.6	1.3	1.5	0	0
	123.2	260.8	3.8	1.2	0	0
	124.8	260.7	0.1	1.8	0	0
	126.4	260.7	0	1.1	0	0
	128	260.7	0	1.6	0	0
Rogers Dam	129.6	260.7	0	1.7	0	0
US-131	131.2	252.1	8.6	2.6	0	0
	132.8	250.5	1.6	1.9	0	0
	134.4	250.5	0	1.2	0	0
	136	250.5	0	1.9	0	0
	137.6	250.5	0	1.1	0	0
	139.2	250.5	0	1.3	0	0
	140.8	250.5	0	1.1	0	0
	142.4	250.5	0	1.1	0	0
	144	250.5	0	1.1	0	0
	145.6	250.5	0	1.1	0	0
	147.2	250.5	0	1.1	0	0
	148.8	250.5	0	1.1	0	0
	150.4	250.5	0	1.1	0	0
Hardy Dam	152	250.5	0	1.1	0	0
	153.6	220	30.5	1.4	0	0
	155.2	220	0	1.2	0	0
Croton Dam	156.8	220	0	1.2	0	0
	158.4	208.6	11.4	1.1	0	0
	160	206.5	2.1	1.2	0	0
	161.6	204	2.5	1.2	0	0

Table A.2: (cont'd)

Location	Distance (km.)	Elevation (m)	Slope (m/km)	Sinuosity	Cutoffs (1840s-1938)	Cutoffs (1938-2012)
	163.2	203.4	0.6	1.2	1	0
	164.8	203.1	0.3	1.3	1	0
	166.4	200.5	2.6	1.4	0	0
	168	200	0.5	1.1	0	0
	169.6	198.1	1.9	1.2	0	0
	171.2	197.7	0.4	2	1	0
	172.8	194.6	3.1	1.2	0	0
Newaygo / M-37	174.4	194.1	0.5	1.1	0	0
	176	192.5	1.6	1.1	0	0
	177.6	191.1	1.4	1.1	0	0
	179.2	191	0.1	1.3	0	0
	180.8	189.9	1.1	1.2	0	0
	182.4	189.8	0.1	1.9	0	0
	184	188.6	1.2	1.2	0	0
	185.6	188.2	0.4	1.1	0	0
	187.2	188.1	0.1	1.1	0	0
	188.8	188	0.1	1.1	0	0
Bridgeton	190.4	186	2	1.3	0	0
	192	185.9	0.1	1.4	0	0
	193.6	185.8	0.1	1.1	0	0
	195.2	184.5	1.3	1.1	0	0
	196.8	184.4	0.1	1.2	0	0
	198.4	184.3	0.1	1.1	0	0
	200	183.8	0.5	1.3	0	0
	201.6	183.7	0.1	1.2	0	0
	203.2	182.7	1	1.3	0	0
Maple Island Rd.	204.8	182.6	0.1	1.2	0	0
	206.4	182.4	0.2	1.1	0	0
	208	181.7	0.7	1.1	0	0
	209.6	181.4	0.3	1.1	0	0
	211.2	179.9	1.5	1.2	0	0
	212.8	179.8	0.1	1.3	0	0
	214.4	177.6	2.2	1.4	0	0
	216	177.5	0.1	1.4	0	0

Table A.2: (cont'd)

Location	Distance (km.)	Elevation (m)	Slope (m/km)	Sinuosity	Cutoffs (1840s-1938)	Cutoffs (1938-2012)
US - 31	217.6	176.8	0.7	1.4	0	0
	219.2	176.8	0	1.2	0	0
	220.8	176.8	0	1.1	0	0
	222.4	176.8	0	1.1	0	0
Muskegon	224	176.6	0.2	1.1	0	0
	225.6	176.6	0	1.1	0	0
	227.2	176.6	0	1.1	0	0
Muskegon Lake	228.8	174.9	-	-	-	-

Table A.3: 1800s and 2013 Channel Width data for the Muskegon River.

Section Boundary	Township	County	1800s Width (m)	2013 Width (m)	Difference (m)
Sec 3 - Sec 10	Lake	Roscommon	*48.28*	-	-
Sec 3 - Sec 4	Lake	Roscommon	*47.67*	-	-
Sec 9 - Sec 8	Lake	Roscommon	20.92	-	-
Sec 5 - Sec 8	Lake	Roscommon	33.79	-	-
Sec 8 - Sec 17	Lake	Roscommon	*43.65*	-	-
Sec 36 - Sec 35	Enterprise	Missaukee	*40.63*	-	-
Sec 2 - Sec 3	Butterfield	Missaukee	36.61	31.08	-5.53
Sec 3 - Sec 10	Butterfield	Missaukee	29.57	34.74	+5.17
Sec 10 - Sec 15	Butterfield	Missaukee	*140.81*	21.03	-
Sec 15 - Sec - 22	Butterfield	Missaukee	15.48	26.51	+11.03
Sec 22 - Sec 27	Butterfield	Missaukee	-	23.77	-
Sec 27 - Sec 34	Butterfield	Missaukee	22.53	22.86	+0.33
Sec 34 - Sec 3	Butterfield-Holland	Missaukee	-	26.51	-
Sec 2 - Sec 11	Holland	Missaukee	31.58	21.94	-9.64
Sec 11 - Sec 12	Holland	Missaukee	22.9	18.28	-4.62
Sec 12 - Sec 13	Holland	Missaukee	21.92	23.77	+1.85
Sec 18 - Sec 19	Roscommon	Roscommon	-	21.94	-
Sec 19 - Sec 30	Roscommon	Roscommon	*134.78*	32.91	-
Sec 30 - Sec 31	Roscommon	Roscommon	*46.67*	25.6	-
Sec 31 - Sec 36	Roscommon-Holland	Roscommon-Missaukee	29.57	31.08	+1.51
Sec 36 - Sec 1	Holland-Summerfield	Missaukee-Clare	27.76	32	+4.24
Sec 1 - Sec 2	Summerfield	Clare	38.42	27.43	-10.99
Sec 2 - Sec 11	Summerfield	Clare	31.98	23.77	-8.21
Sec 10 - Sec 11	Summerfield	Clare	38.82	21.94	-16.88
Sec 10 - Sec 15	Summerfield	Clare	35.6	27.43	-8.17
Sec 15 - Sec 16	Summerfield	Clare	36.61	26.51	-10.1
Sec 16 - Sec 17	Summerfield	Clare	26.95	29.26	+2.31
Sec 17 - Sec 20	Summerfield	Clare	*53.51*	24.68	-
Sec 20 - Sec 19	Summerfield	Clare	35.4	22.86	-12.54
Sec 19 - Sec 24	Summerfield-Winterfield	Clare	*55.52*	32	-
Sec 24 - Sec 25	Winterfield	Clare	43.85	23.77	-20.08
Sec 25 - Sec 26	Winterfield	Clare	39.22	38.4	-0.82

151.48: inaccurate channel width

Table A.3: (cont'd)

Section Boundary	Township	County	1800s Width (m)	2013 Width (m)	Difference (m)
Sec 26 - Sec 35	Winterfield	Clare	33.39	-	-
Sec 35 - Sec 34	Winterfield	Clare	*46.46*	32.91	-
Sec 34 - Sec 3	Winterfield-Redding	Clare	*47.07*	32.91	-
Sec 3 - Sec 4	Redding	Clare	*49.88*	42.97	-
Sec 9 - Sec 10	Redding	Clare	*93.14*	42.97	-
Sec 9 - Sec 16	Redding	Clare	*64.97*	49.37	-
Sec 16 -Sec 21	Redding	Clare	35.4	56.69	+21.29
Sec 21 - Sec 20	Redding	Clare	*65.98*	58.52	-
Sec 20 - Sec 29	Redding	Clare	*53.1*	45.72	-
Sec 29 - Sec 30	Redding	Clare	*62.16*	53.94	-
Sec 30 - Sec 31	Redding	Clare	53.1	51.2	+1.9
Sec 31 - Sec 6	Redding-Freeman	Clare	-	57.6	-
Sec 1 - Sec 12	Freeman	Osceola	*61.35*	43.89	-
Sec 11-Sec 12	Freeman	Osceola	41.84	47.54	+5.7
Sec 11 - Sec 14	Freeman	Osceola	*75.23*	48.46	-
Sec 14-Sec 15	Sylvan	Osceola	*63.97*	-	-
Sec 15 - Sec 16	Sylvan	Osceola	*68.39*	53.94	-
Sec 16-Sec 21	Sylvan	Osceola	*82.07*	41.14	-
Sec 21 - Sec 20	Sylvan	Osceola	*58.33*	39.31	-
Sec 20 - Sec 19	Sylvan	Osceola	*89.31*	52.12	-
Sec 19 - Sec 24	Sylvan-Osceola	Osceola	*149.87*	64	-
Sec 25 - Sec 26	Osceola	Osceola	*84.69*	42.06	-
Sec 26 - Sec 27	Osceola	Osceola	51.7	41.14	-10.56
Sec 27-Sec 34	Osceola	Osceola	*84.69*	60.35	-
Sec 34 - Sec 3	Osceola-Evart	Osceola	41.64	36.57	-5.07
Sec 3 - Sec 4	Evart	Osceola	*66.38*	57.6	-
Sec 4 - Sec 5	Evart	Osceola	74.83	72.23	-2.6
Sec 5 - Sec 8	Evart	Osceola	*92.53*	50.29	-
Sec 8 - Sec 7	Evart	Osceola	*69*	42.97	-
Sec 7 - Sec 12	Evart-Hersey	Osceola	54.71	45.72	-8.99
Sec 12 - Sec 11	Hersey	Osceola	72.62	42.06	-30.56
Sec 11 - Sec 14	Hersey	Osceola	*140.81*	57.6	-

151.48: inaccurate channel width

Table A.3: (cont'd)

Section Boundary	Township	County	1800s Width (m)	2013 Width (m)	Difference (m)
Sec 15 - Sec 22	Hersey	Osceola	*100.58*	62.17	-
Sec 15 - Sec 16	Hersey	Osceola	*110.84*	96.01	-
Sec 16 - Sec 17	Hersey	Osceola	*92.53*	53.03	-
Sec 17 - Sec 20	Hersey	Osceola	66.78	45.72	-21.06
Sec 20 - Sec 19	Hersey	Osceola	79.66	53.03	-26.63
Sec 19 - Sec 30	Hersey	Osceola	*120.9*	69.49	-
Sec 30 - Sec 31	Hersey	Osceola	*243.41*	58.52	-
Sec 31 - Sec 32	Hersey	Osceola	97.56	67.66	-29.9
Sec 32 - Sec 5	Hersey-Grant	Osceola-Mecosta	74.63	74.06	-0.57
Sec 5 - Sec 8	Grant	Mecosta	*102.39*	83.21	-
Sec 7-Sec 8	Grant	Mecosta	*344.39*	65.83	-
Sec 7 - Sec 12	Grant-Green	Mecosta	100.58	66.75	-33.83
Sec 12 - Sec 11	Green	Mecosta	60.35	55.77	-4.58
Sec 11 - Sec 10	Green	Mecosta	80.46	52.12	-28.34
Sec 10 - Sec 15	Green	Mecosta	87.5	42.97	-44.53
Sec 15 - Sec 22	Green	Mecosta	*102.79*	80.46	-
Sec 22 - Sec 27	Green	Mecosta	102.59	62.17	-40.42
Sec 26-Sec 27	Green	Mecosta	*226.11*	65.83	-
Sec 26 - Sec 35	Green	Mecosta	90.52	69.49	-21.03
Sec 34-Sec 35	Green	Mecosta	*158.78*	73.15	-
Sec 34 - Sec 3	Green-Big Rapids	Mecosta	*157.91*	80.46	-
Sec 3 - Sec 10	Big Rapids	Mecosta	*273.18*	72.23	-
Sec 11 - Sec 14	Big Rapids	Mecosta	*191.11*	53.94	-
Sec 14 - Sec 23	Big Rapids	Mecosta	60.35	41.14	-19.21
Sec 23 - Sec 24	Big Rapids	Mecosta	75.43	-	-
Sec 24 - Sec 25	Big Rapids	Mecosta	116.67	-	-
Sec 25 - Sec 36	Big Rapids	Mecosta	80.46	-	-
Sec 36 - Sec 1	Big Rapids - Mecosta	Mecosta	*198.95*	-	-
Sec 11 - Sec 14	Mecosta	Mecosta	*120.7*	-	-
Sec 14-Sec 15	Mecosta	Mecosta	110.64	-	-
Sec 9-Sec 16	Mecosta	Mecosta	103.6	-	-
Sec 16 - Sec 21	Mecosta	Mecosta	*170.99*	-	-

151.48: inaccurate channel width

Table A.3: (cont'd)

Section Boundary	Township	County	1800s Width (m)	2013 Width (m)	Difference (m)
Sec 20 - Sec 29	Mecosta	Mecosta	114.06	-	-
Sec 29 - Sec 32	Mecosta	Mecosta	76.44	-	-
Sec 31-Sec 32	Mecosta	Mecosta	90.52	-	-
Sec 31 - Sec 6	Mecosta-Aetna	Mecosta	82.27	-	-
Sec 6 - Sec 1	Aetna-Big Prairie	Mecosta-Newaygo	*123.71*	-	-
Sec 18 - Sec 19	Croton	Newaygo	80.46	70.4	-10.06
Sec 19 - Sec 24	Croton-Brooks	Newaygo	90.92	63.09	-27.83
Sec 24 - Sec 13	Brooks	Newaygo	113.86	104.24	-9.62
Sec 13 - Sec 14	Brooks	Newaygo	*171.99*	68.58	-
Sec 14 - Sec 15	Brooks	Newaygo	-	63.09	-
Sec 14 - Sec 23	Brooks	Newaygo	90.52	57.6	-32.92
Sec 23 - Sec 26	Brooks	Newaygo	143.63	65.83	-77.8
Sec 27 - Sec 22	Brooks	Newaygo	*140.81*	52.12	-
Sec 22 - Sec 21	Brooks	Newaygo	-	52.12	-
Sec 21 - Sec 20	Brooks	Newaygo	85.89	59.43	-26.46
Sec 20 - Sec 19	Brooks	Newaygo	-	58.52	-
Sec 19 - Sec 24	Brooks-Garfield	Newaygo	85.49	71.32	-14.17
Sec 24 - Sec 23	Garfield	Newaygo	*161.53*	58.52	-
Sec 26 - Sec 25	Garfield	Newaygo	*102.59*	51.2	-
Sec 23 - Sec 26	Garfield	Newaygo	*120.09*	54.86	-
Sec 35 - Sec 36	Garfield	Newaygo	-	53.94	-
Sec 27 - Sec 34	Garfield	Newaygo	*96.56*	47.54	-
Sec 34 - Sec 33	Garfield	Newaygo	80.46	57.6	-22.86
Sec 33 - Sec 4	Garfield-Ashland	Newaygo	*162.94*	64.92	-
Sec 33 - Sec 32	Garfield	Newaygo	84.28	75.89	-8.39
Sec 32 - Sec 5	Ashland	Newaygo	90.52	74.06	-16.46
Sec 5 - Sec 6	Ashland	Newaygo	92.53	56.69	-35.84
Sec 6 - Sec 7	Ashland	Newaygo	91.33	50.29	-41.04
Sec 7 - Sec 12	Ashland-Bridgeton	Newaygo	70.6	53.94	-16.66
Sec 12 - Sec 13	Bridgeton	Newaygo	100.58	78.63	-21.95
Sec 13 - Sec 14	Bridgeton	Newaygo	*128.94*	-	-
Sec 14 - Sec 15	Bridgeton	Newaygo	88.51	-	-

151.48: inaccurate channel width

Table A.3: (cont'd)

Section Boundary	Township	County	1800s Width (m)	2013 Width (m)	Difference (m)
Sec 15 - Sec 16	Bridgeton	Newaygo	66.58	-	-
Sec 16 - Sec 21	Bridgeton	Newaygo	*112.65*	-	-
Sec 21 - Sec 20	Bridgeton	Newaygo	63.56	-	-
Sec 30 - Sec 25	Bridgeton-Cedar Creek	Newaygo-Muskegon	126.33	78.63	-47.7
Sec 25 - Sec 36	Cedar Creek	Muskegon	*39.63*	-	-
Sec 35 - Sec 36	Cedar Creek	Muskegon	61.35	-	-
Sec 35 - Sec 34	Cedar Creek	Muskegon	57.73	-	-
Sec 34 - Sec 33	Cedar Creek	Muskegon	54.71	-	-
Sec 4 - Sec 5	Egelston	Muskegon	72.42	-	-
Sec 7 - Sec 12	Egelston-Muskegon	Muskegon	70	-	-
Sec 12 - Sec 11	Muskegon	Muskegon	*221.28*	-	-
Sec 11 - Sec 10	Muskegon	Muskegon	95.95	-	-
Sec 10 - Sec 9	Muskegon	Muskegon	95.15	-	-

151.48: inaccurate channel width

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