# MAPPING AND CHARACTERIZING A RELICT LACUSTRINE DELTA IN CENTRAL LOWER MICHIGAN

By

Christopher B. Connallon

A THESIS

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

Geography – Master of Science

2015

**ABSTRACT**

MAPPING AND CHARACTERIZING A RELICT LACUSTRINE DELTA IN CENTRAL LOWER MICHIGAN

By

Christopher B. Connallon

This research focuses on, mapping and characterizing the Chippewa River delta - a sandy, relict delta of Glacial Lake Saginaw in central Lower Michigan. The delta was first identified in a GIS, using digital soil data, as the sandy soils of the delta stand in contrast to the loamier soils of the lake plain. I determined the textural properties of the delta sediment from 142 parent material samples at ≈1.5 m depth. The data were analyzed in a GIS to identify textural trends across the delta. Data from 3276 water well logs across the delta, and from 185 sites within two-storied soils on the delta margin, were used to estimate the thickness of delta sands and to refine the delta's boundary. The delta heads near Mount Pleasant, expanding east, onto the Lake Saginaw plain. It is ≈18 km wide and ≈38 km long and comprised almost entirely of sandy sediment. As expected, delta sands generally thin away from the head, where sediments are ≈4-7m thick. In the eastern, lower portion of the delta, sediments are considerably thinner (≈<1-2m). The texturally coarsest parts of the delta are generally coincident with former shorezones. The thick, upper delta portion is generally coincident with the relict shorelines of Lakes Saginaw and Arkona (≈17.1k to ≈ 16k years BP), whereas most of the thin, distal, lower delta is generally associated with Lake Warren (≈15k years BP). Delta sediments from neighboring drainages merge with lower Chippewa Delta sediments, obscuring landform boundaries. Together, these data suggest that the Chippewa Delta formed and prograded as lake levels episodically fell, producing a two-tiered delta complex and an overall, thin, sandy landform.

**ACKNOWLEDGEMENTS**

I would like to begin my acknowledgements by thanking my research advisor, Randy Schaetzl. Working with Randy has helped me develop my approach to research, and to improve as a scientist, a writer, and a teacher. I'm grateful for the time he invested in my education and development, his perpetually open door, and his insistence on holding my work to his high standards. Randy's willingness to push me to improve has enriched my experience here at MSU and for that I thank him.

I also thank the other members of my committee, Drs. Arbogast and Lusch, as well as everyone in the Department of Geography, for leading by example, and engendering a sense of community within the department. It's been a pleasure to work with such welcoming, engaged, passionate, and knowledgeable professionals. I owe a great deal of gratitude to the Department of Geography faculty, staff, and fellow graduate students for being so supportive, accommodating, and available. Because of them, my two years at MSU have been the most fulfilling stop in my long journey through higher education. I'm grateful for the opportunity.

Thanks also to my family, for all their support and encouragement over the years. I especially thank my brother Tim for his endless enthusiasm for scientific discovery, and for encouraging me to pursue science, despite my initial aversion. Tim facilitated my introduction to working in archaeology, which sparked my interested in pursuing a career in earth science, so for that I thank him. I would also like to thank my parents and my brother Matt for always being supportive and helping me reach my goals. I believe that having a strong family is very important to a person's ability to achieve, and that my family deserves recognition for their contributions to my success.

Lastly, I'd like to recognize individuals who contributed, directly or indirectly, to my personal or professional development, to this research, and to my experience at MSU. Thanks to Mike Luehmann for the assistance, advice, and data; his help was instrumental in my completion of this research. Thanks to the instructors whose classes I assisted; you helped me better understand what it is to be a teacher, and a leader. Thanks to Charles Bello, Richard Hunter, and John Stiteler, who afforded me opportunity and advice, as well as strong encouragement. Thanks to Stefan Kawecki and Debbie Cutchin, who helped me understand the importance of finding creative solutions. Those lessons were strongly influential in my personal development, and contributed strongly to my success. Thanks to Nugget, my 12 year old Australian Shepherd, who was my assistant in collecting field samples for this research. Thanks, finally, to everyone who helped me along the way, specifically Brian Maguire, Danny Kluxen, and James Caverly. All of your help has been much appreciated.

This research was funded by fellowships from the Department of Geography at Michigan State University.

## TABLE OF CONTENTS

**LIST OF TABLES**

## LIST OF FIGURES

## 1. Introduction

Soils have the ability to preserve the characteristics of sediment in which they are formed, and have been increasingly studied for their ability to help inform interpretations of the depositional systems which emplaced their parent sediment (Karathanasis and Golrick, 1991; Muhs et al., 1997; Schaetzl, 1998; Wysocki et al., 2005; Silva et al., 2008; Lusch et al., 2009; Stanley and Schaetzl, 2011; Luehmann et al., 2013; Schaetzl and Luehmann, 2013). Soil parent material, the stratum below the depth of soil development, often retains sedimentary characteristics, such as structure, texture, degree of particle size sorting, and sediment stratification. In landscapes where soils are young, and have not yet been subjected to intense weathering or erosion, these parent materials remain in generally unaltered form and hence, are sometimes easily identifiable as to type.

Michigan has such a landscape, being both geologically young at the surface and comprised primarily of glacially deposited sediments (Farrand and Eschman, 1974; Larson and Schaetzl, 2001; Krist and Lusch, 2004; Kincare and Larson, 2009). Soil maps can therefore serve as potential proxies for surficial geologic data in this type of setting, enabling researchers to study soil properties, observe patterns, and draw inferences about past sediment transport and depositional systems (Schaetzl and Weisenborn, 2004; Wysocki et al., 2005; Stanley and Schaetzl, 2011; Luehmann et al., 2013; Schaetzl and Luehmann, 2013; Wald et al., 2013). The United States Department of Agriculture, Natural Resources Conservation Service (USDA-NRCS) has made its soil maps available in digital form, enabling their use in geographic information systems (GIS) and empowering researchers with a wealth of geographic, and potentially geologic, data. These maps are periodically refined by the NRCS, to achieve even greater spatial and geologic accuracy.

By employing a geographic approach, researchers have reconstructed past depositional environments in the upper Midwest by interpreting soil patterns in a GIS (Stanley and Schaetzl, 2011; Luehmann et al., 2013). Schaetzl et al. (2000), for example, provided evidence for a then unidentified paleolake in northern Lower Michigan by analyzing soil properties and patterns across the landscape, and concluded that the sediments that had previously, and erroneously, been identified as glacial till were actually stratified lacustrine sediments. Another, slightly more general, example of a soil's ability to preserve the record of its sedimentary process is in soils formed in loess, or wind-blown silt. Loess transport and deposition typically conforms to predictable spatio-textural patterns; deposits are usually thickest and coarsest near their source, and become increasingly thinner and finer textured with distance (Smith, 1942; Olson and Ruhe, 1979; Mason, 2001; Bettis et al., 2003). Because these patterns can be preserved in soils, they can be examined spatially. This type of spatial analysis can then help to interpret details about the process by which sediment is transported and deposited, from where the source sediment originates, and may provide information about paleoenvironmental conditions during the period of sediment movement. In this thesis I employ a similar type of geographical approach, by identifying, mapping, and characterizing a late-Quaternary landform which may contain interpretable spatio-textural patterns: a relict Wisconsinan-age delta.

Proxy evidence for past environmental or sedimentary conditions, when preserved in deltaic deposits and morphologies, can be useful for interpreting the processes that were at work during periods of delta formation and progradation. Delta asymmetry, for example, provides proxy evidence for longshore drift; the severity of delta asymmetry is often correlative with the strength of longshore currents (Bhattacharya and Goisan, 2003; Weiguo et al., 2011). Vader et al. (2012) analyzed the characteristics of a Glacial Lake Algonquin-aged delta and spits to support the assertion that strong easterly winds in northern Lower Michigan drove westward flowing longshore currents during the Late Pleistocene. Leverett and Taylor (1915) observed relict deltas in the Erie basin with a coating of fine lacustrine sediment and surmised that the delta building event preceded a subsequent rise in lake level, which helped inform their interpretation of the sequence of lake level fluctuations within that basin. These examples illustrate the potential wealth of data, proxy or otherwise, that relict deltas can offer to researchers interested in reconstructing past environmental or depositional systems and sequences.

Deltas form in lacustrine and oceanic basins when sediment-bearing rivers discharge into standing water, depositing sediment, commonly in a lobate or cuspate form (Leeder, 1982; Boggs, 2001). The formation of a delta is dependent on a number of conditions; 1) there must be a sufficient supply of unconsolidated sediment on the landscape, 2) there must be a river with sufficient energy for sediment transport, and 3) there must be a standing water basin, the receiving body for sediment deposition (Leeder, 1982; Boggs, 2001). Delta sedimentary characteristics vary, depending on the energy of the system of transport and deposition, the nature of available sediment, and extent of post-depositional reworking and redistribution by waves and/or tides (Galloway, 1975; Wright et al., 1980; Bhattacharya and Giosan, 2003). Despite these variations, deltaic deposits often have morphologic and textural characteristics that allow their paleoenvironmental conditions to be identified and characterized, even after the lacustrine or marine setting in which they formed is no longer extant.

In mid-Michigan, these three conditions existed. Thus, the landscape was highly conducive to delta formation. Nearly all of the surface of Lower Michigan consists of thick deposits of unconsolidated glacial sediment, and during the period of proglacial lake formation, vegetation was minimal, providing a ready supply of sediment for erosion, transport by rivers, and subsequent deposition in lakes, as deltas. Michigan has a humid climate today, as it did in the Pleistocene, with ample rain and snowfall feeding into a vast network of streams and rivers capable of carrying large amounts of sediment. Finally, the area had several large proglacial lakes during the late-Pleistocene, which not only spanned vast parts of the landscape, but which persisted for thousands of years (Leverett and Taylor, 1915; Bretz, 1951; Hough, 1958; Calkin and Feenstra, 1985; Larson and Schaetzl, 2009). This research focuses on a sandy, relict delta associated with the Chippewa River, which is an underfit and incised river in central Lower Michigan. During the late Quaternary glacial retreat, this delta developed into a series of proglacial lakes in the Saginaw Lowlands: Lakes Saginaw, Arkona, and Warren (Leverett and Taylor, 1915; Bretz, 1951; Martin, 1955; Hough, 1958, Lusch et al., 2009). Although a small portion of this delta had previously been marked on a map, by Martin (1955), it had been incompletely mapped and has never previously studied in detail.

Prior to this study, the origin of the delta sands on the Saginaw Lake Plain has been misunderstood, misinterpreted, or simply unknown to researchers. Therefore, researchers have typically referred to these sands in very general terms. For example, Quaternary maps by Farrand and Bell (1982) refer to the Chippewa Delta sands as “Lacustrine sand and gravel”, and the map from Martin (1955) refers to large parts of the delta simply as “Lake beds, sand.” Indeed, even a more recent study by Schaetzl et al. (2013) which used a GIS to map physiographic regions in Michigan named to the sandy region which includes much of the delta surface, “Lake Saginaw dunes and plains”, and suggested that, “Sand may be shallow deltaic sediment from streams entering paleolakes”, but offer no definitive assertion as to its specific origins (Schaetzl et al., 2013, Appendix, 16). Studies of sand dunes in this region described the sands generally as “shallow-water lacustrine deposits” and differentiated the dunes formed in these sediments from dunes formed in outwash, but they did not specifically characterize the sands as deltaic (Arbogast et al., 1997, 70). Confirming these deposits as deltaic in origin may be of much interest to scientists studying the dunes formed from such parent material, as well as scientists conducting future geomorphic research within the Saginaw Lowlands. Therefore, in this thesis I present evidence that the sands in my study area that have heretofore been referred to only in general terms are actually deltaic in origin. In this study, I will analyze NRCS soil data, textural data from samples collected across the delta surface, water well data, and a digital elevation model in order to map and characterize the relict Chippewa River Delta. By studying the morphology and its spatio-textural characteristics of this delta, and tying these data into the remnant shorelines of the various proglacial lakes within the Saginaw Lowlands, I hope to make inferences about the relative timing of delta progradation, association of delta building phases with various lake stages, and the alteration of the delta sediments by fluctuating lake levels. Fortuitously, this landform persists and is readily accessible in the present day, allowing for this research into its characteristics and geomorphic origins which will culminate in an increased understanding of sediment emplacement, chronology and movement within the Saginaw Lowlands during the late-Wisconsinan.

With this information in mind, the goals of this research are 1) to document, characterize, and map the sandy delta of the Chippewa River in central Lower Michigan, 2) to determine if a sequence of delta progradation existed for this delta with respect to different lake levels that existed within the Saginaw Lowlands, and how fluctuating lake levels may have affected neighboring landforms, 3) and to demonstrate the efficacy of the spatial methodology used to conduct this research.

## 2. Literature Review

Michigan's landscape bears the imprint of past glacial episodes. The most recent glaciation, the Wisconsinan, left features at the surface which have been studied and interpreted by geomorphologists to make inferences about the processes and sequences of glacial movement and related sediment deposition (Fisher et al., 2005; Kozlowski et al., 2005; Kehew et al., 2012). In this section, I will review a) the sequence of glacial events that led to inundation of the Saginaw Lowlands by proglacial lakes, b) the various lake stages, names and elevations in this basin, and c) the final retreat of this proglacial lake from my study area. When available in the literature, I will report radiocarbon ages in their calibrated form, with the original uncalibrated values in parentheses. Radiocarbon ($^{14}C$) calibration employs a statistically derived calibration curve to convert radiocarbon ages to calendar years before present, and likewise converts the uncertainty values into calendar years (Fairbanks et al. 2005). All reported $^{14}C$ dates have been calibrated using the Fairbanks et al. (2005) calibration curve, but because different calibration methods yield slightly different results, I have also reported the original radiocarbon dates.

### 2.1 Retreat of the Saginaw Glacial Lobe

Much of central Lower Michigan was glaciated by the Saginaw Lobe, sometimes called the Saginaw sublobe, of the Laurentide Ice Sheet (Leverett and Taylor, 1915; Bretz, 1951; Hough, 1958; Farrand and Eschman, 1974; Krist and Lusch, 2004; Kincare and Larson, 2009; Kehew et al., 2012). Therefore, most of the glacial features and landforms within the Saginaw basin and surrounding areas are Saginaw Lobe features. During the advance to its maximum position in central Indiana and Illinois, the three dominant lobes in the central Great Lakes

Region, the Lake Michigan, Huron-Erie, and Saginaw lobes, merged in southwestern Michigan near the Indiana border. Similarly, during their retreat and subsequent episodes of deglaciation, the three lobes decoupled from one another in southwestern Michigan around 19,000 years B.P. as their ice margins retreated northward (Kincare and Larson, 2009). Once separated, the lobes were able to move independently from one another.

The Saginaw Lobe, likely the thinnest and weakest of the three lobes, retreated faster than the Lake Michigan and Erie-Huron, and underwent extensive stagnation as evidenced by the hummocky ice-stagnation landscapes and eskers it left behind (Kehew et al., 2012). Subglacial tunnel channels in southwestern and central Michigan suggest subglacial outburst floods from beneath the Saginaw Lobe (Fisher et al., 2005; Kozlowski et al., 2005; Kehew et al., 2012). In southwestern Michigan, Saginaw Lobe moraines are overridden on their edges by Lake Michigan and Huron-Erie lobe sediments, indicating that the timing of the retreat for each lobe was asynchronous. Additionally, drumlinized till plains in southwestern Michigan indicate periods of glacial surging for the Saginaw Lobe (Kehew et al., 2012).

As the Saginaw margin retreated toward present-day Saginaw Bay, tunnel channels and hummocky ice stagnation topography transitioned to till plains and a series of concentric, low-relief moraines which mark the locations of temporary standstills of the ice margin (Leverett and Taylor, 1915). These moraines are part of the West Branch-Gladwin group in central Lower Michigan, and mark the transition between subaerially deposited, till-cored moraines and the tills deposited later in the proglacial lakes that occupied the Saginaw Lowlands beginning around 17,100 years B.P. (Leverett and Taylor, 1915; Ward, 1979; Kincare and Larson, 2009). The Flint and Owosso moraines are the two last subaerially deposited moraines in this system, recording the position of the last ice marginal oscillation before ponded water began accumulating across the eastward sloping land surface in front of the retreating (and periodically re-advancing) ice margin. Over the next few thousand years, a series of proglacial lakes occupied the Saginaw Lowlands, before the region finally drained completely sometime after 13,000 years ago (Hough, 1958; Kincare and Larson, 2009). Parts of the Saginaw Lowlands were inundated again during the Nipissing Transgression, around 6,300 years ago, but this rise in lake level did not reach the study area and therefor will not be discussed further (Kincare and Larson, 2009).

Also noteworthy is that the Maple River-Grand River Valley, through which many of the lake phases of Glacial Lake Saginaw discharged. Water flowing through this outlet cut through and incised into the moraines and till plains left by the Saginaw Lobe. The various episodes of incision contributed to the fluctuating elevations for the various lakes that occupied the lowlands during the late-Wisconsinan period (Bretz, 1951). In the following section, I will provide sequential information about the proglacial lakes which inundated the study area within the Saginaw basin from approximately 17,100 years ago to roughly 13,000 years ago.

**2.2 Glacial Lakes in the Saginaw Lowlands**

**2.2.1 Early Lake Saginaw**

Early Lake Saginaw was the first proglacial lake to occupy the Saginaw Lowlands during the late-Wisconsinan glacial retreat (Leverett and Taylor, 1915; Bretz, 1951; Hough, 1958; Farrand and Eschman, 1974; Ward, 1979; Kincare and Larson, 2009; Fig. 2.1). After the Saginaw Lobe retreated from the Flint-Owosso Moraine, impounded water that pooled between the ice

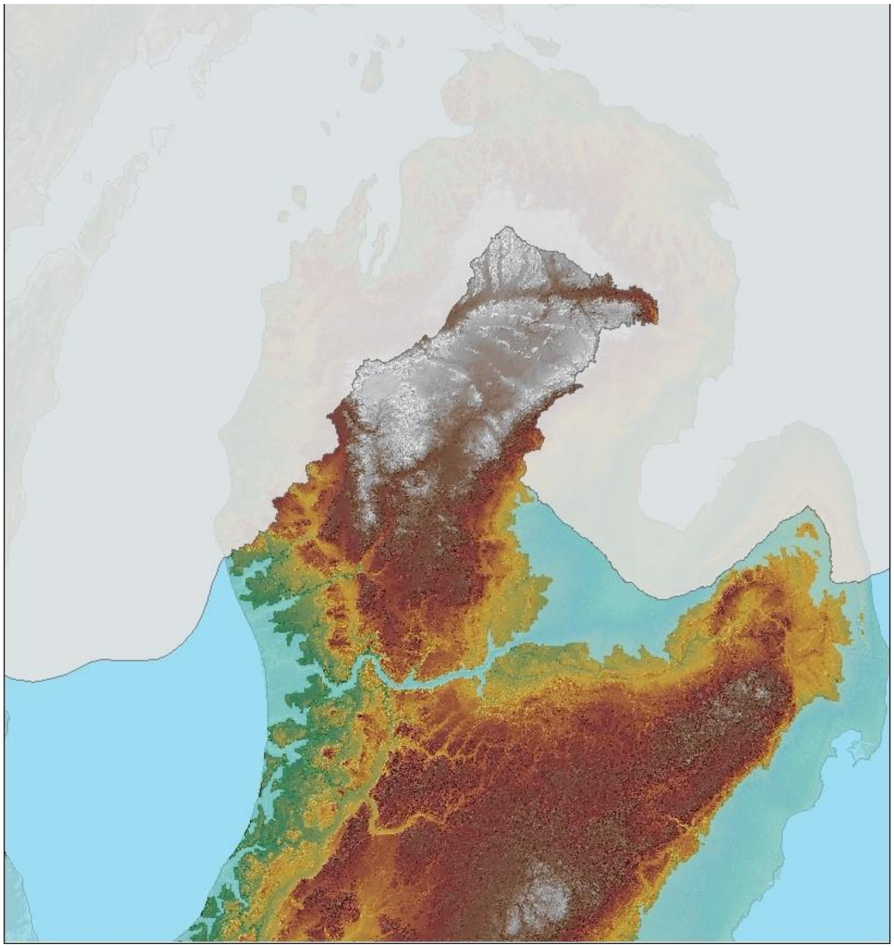


Figure 2.1 – Representation of the approximate configuration of glacial lobes (white) and lakes (blue) for Michigan during Early Lake Saginaw time.

margin and the moraine left a beach ridge to mark the elevation of the lake at roughly 224m (Bretz, 1951; Hough, 1958; Fig. 2.2). Leverett and Taylor identified an Early Lake Saginaw shoreline near Flushing in Genesee County, at an elevation of 720 to 725 feet (~219.5 to 221 meters) above sea level. Early Lake Saginaw drained westwardly through the Maple-Grand Valley (Krist and Lusch, 2004). Downcutting in this valley caused Early Lake Saginaw levels to fall as low as 216m, forming a lake phase which Hough (1958) termed Later Lake Saginaw. During this time, the Saginaw Lobe continued its retreat, allowing Later Lake Saginaw to expand until ice uncovered the "thumb." This allowed Saginaw basin waters to merge with water ponded in the Erie basin (then called Lake Maumee). The merged lake in the Saginaw and Erie basins was

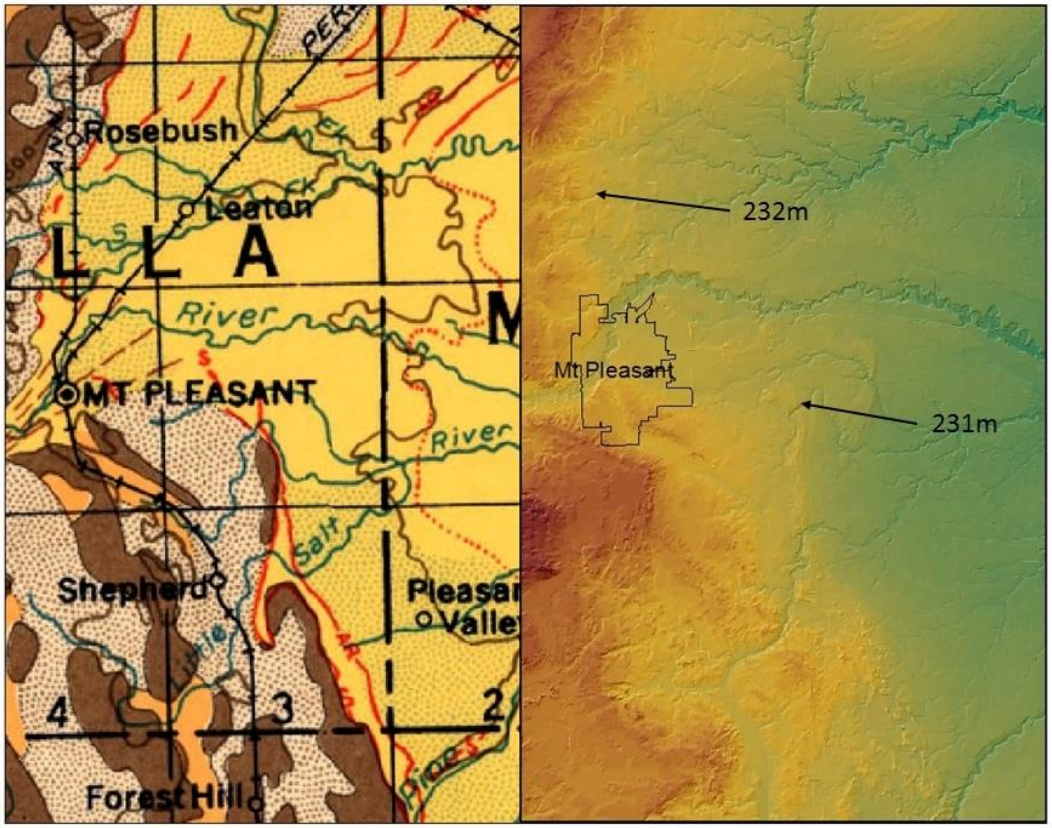



Figure 2.2 – Martin (1955) map showing Early Lake Saginaw shorelines (left) and the DEM hillshade with those beach ridge features in the study area, and their associated elevations.

called Lake Arkona. Leverett and Taylor (1915) also suggested that, due to lake level fluctuations and prolonged wave action, the majority of the beaches made by Early (and Later) Lake Saginaw were modified and incorporated into Lake Arkona beach ridges, which range from 215m to 225m (Leverett and Taylor, 1915; Bretz, 1951; Hough, 1958).

**2.2.2 Lake Arkona**

When parts of the “thumb” became uncovered by the retreating ice margin, water from the Erie basin and the Saginaw Lowlands merged to form Lake Arkona around 16,500 years ago (Kincare and Larson, 2009; Fig. 2.3). Lake Arkona was a single lake that occupied both the Saginaw and Erie basins. Leverett and Taylor (1915) identified three successive beach ridges

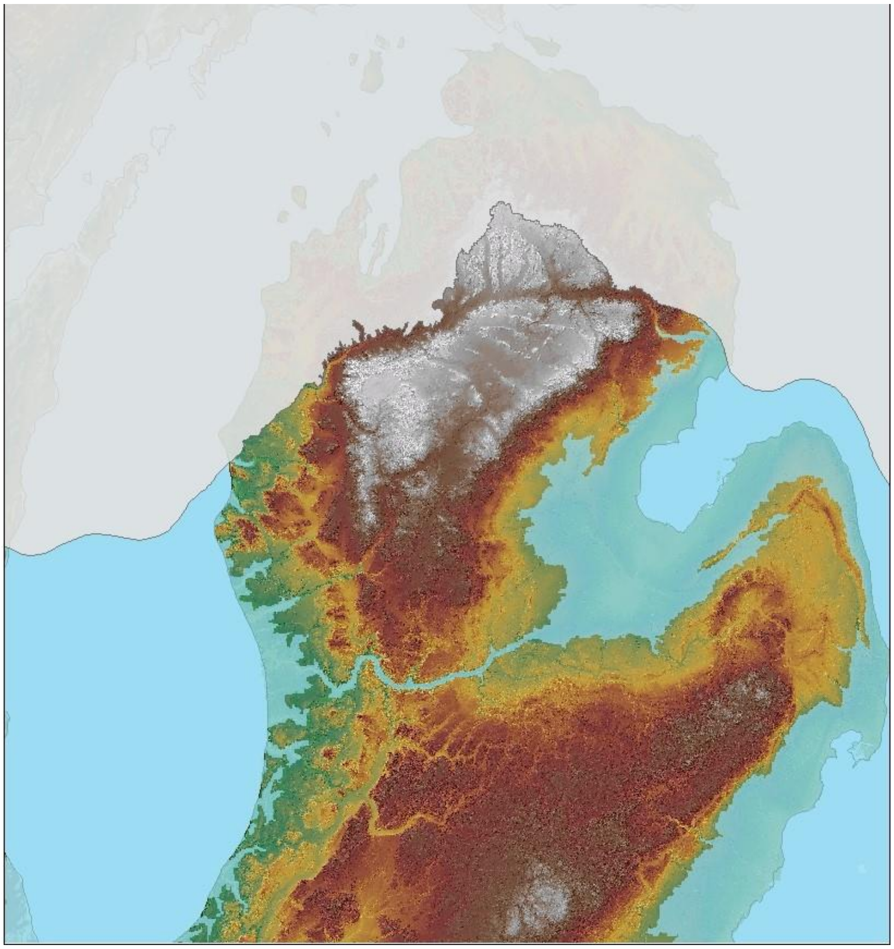


Figure 2.3 – Representation of the approximate configuration of glacial lobes (white) and lakes (blue) for Michigan during Lake Arkona time.

associated with Lake Arkona, and noted that the extent of their preservation or modification was dependent on location. Unmodified Arkona ridges in the Saginaw Lowlands, for example, show stronger expression than those in the Erie basin, where they had presumably been modified by waves and storm surges from the later-occurring Lake Whittlesey (Leverett and Taylor, 1915). Lake Arkona beach ridges mapped by Leverett and Taylor in northern Clinton County range in elevation range from 216m and 217m, but in northern Genesee County and southwestern Tuscola County, shoreline elevations range from 219m to 225m and show a strongly parallel, three-ridge pattern (Fig. 2.4). Similarly, in central Gratiot County weak Arkona shorelines are mapped between 216m and 218m in elevation, but in the northwest corner of

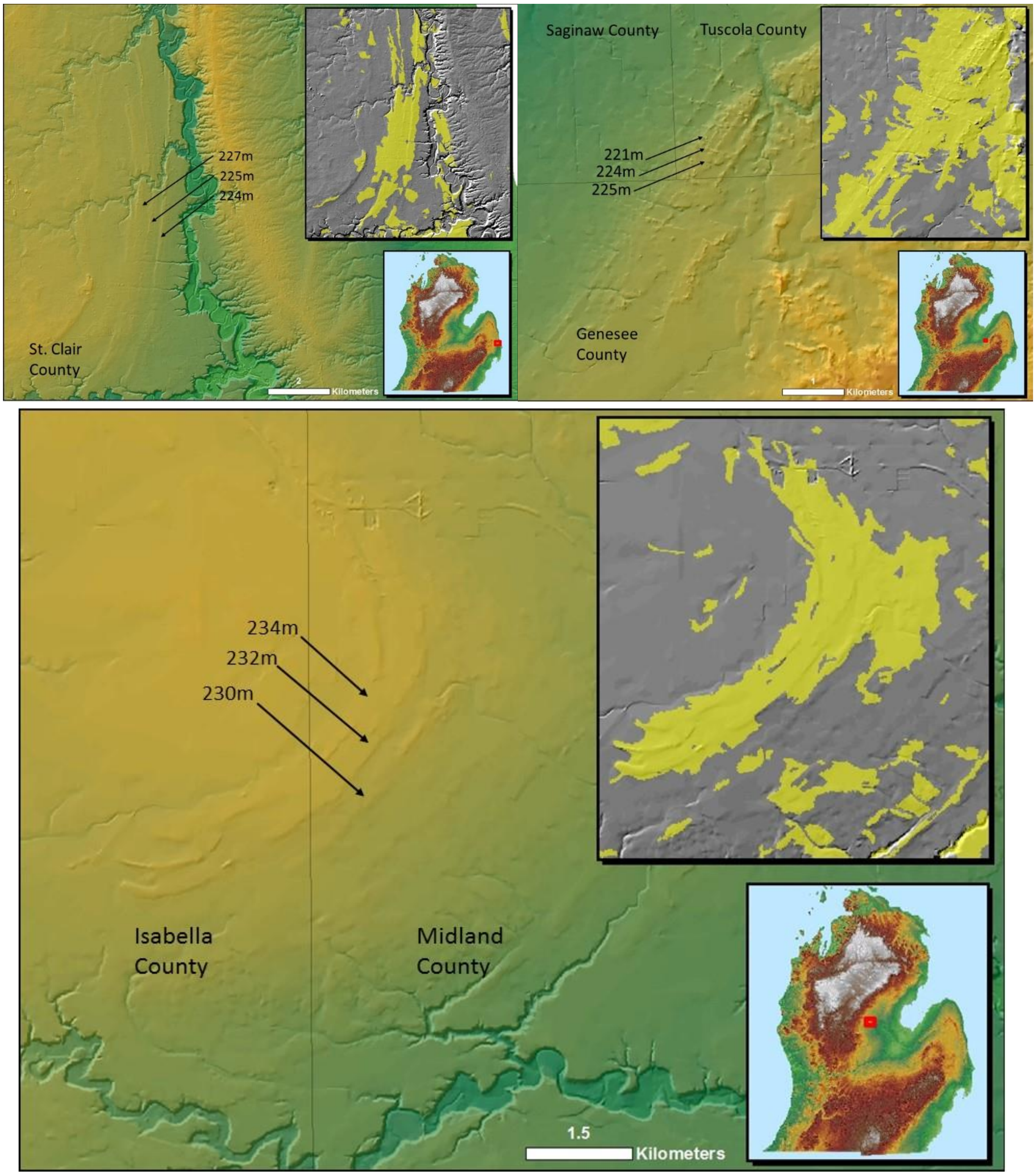



Figure 2.4 – DEM and hillshade showing Lake Arkona beach ridges (main map) and their associated elevations. The lower right side inset map shows location and extent within Michigan, the upper right side inset map shows sand textured subsoil areas on these Arkona ridges (data from the USDA-NRCS) Maps are within St. Clair County (upper left), the border of Genesee and Tuscola Counties (upper right), and northwestern Midland County (bottom) which is within the study area extent.

Midland County, mapped Arkona shorelines range between 215m and 225m. The differences in shoreline elevations is likely due to the effects of isostatic rebound (Lusch et al., 2009).

Leverett and Taylor (1915) also described an Arkona delta associated with the Huron River south of Ypsilanti, in Washtenaw and Wayne counties (Fig. 2.5). This delta is 6.5-8km wide, and consists of coarse gravels and sand in the upper reaches, which becomes increasingly finer textured toward the periphery. Leverett and Taylor (1915, p. 372) stated that “This delta, and in fact all of the Arkona deltas, protrude farther into the lake bed than do the deltas of any other state of the lake waters.”

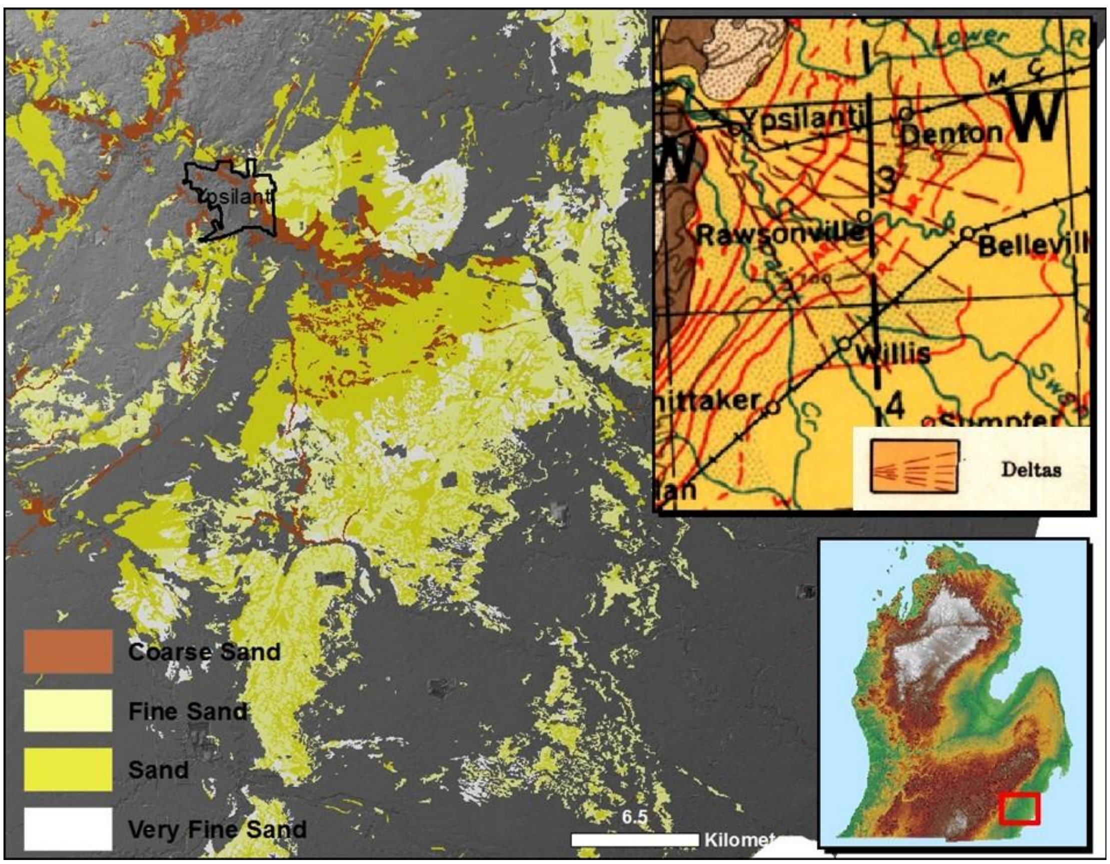



Figure 2.5 – Sand textured subsoil over a DEM hillshade in Washtenaw and Wayne Counties for the Arkona delta identified by Leverett and Taylor (1915). Lower right inset map shows location, upper inset map is from Martin (1955) showing these deposits mapped as deltaic.

| $^{14}$C Dates | Date (cal yrs B.P.) | Location | Event | Lake |
|---|---|---|---|---|
| Burgis (1970) | 16,043± 288 | Imlay Channel | Retreat of LIS ice from "thumb" | First Arkona |
| Goldthwait (1958) | 15,868 ± 658 | Cleveland, OH | Arkona low (212m) | Last Arkona |
| Monaghan and Hansel (1990) | 15681 ± 183 | Riverside, MI | Beginning of lacustrine sedimentation after Ypsilanti-era low stand in Lake Michigan basin | Intra-Glenwood low-water phase, correlative to Lake Ypsilanti in the Erie Basin |

Table 2.1 – Cited radiocarbon dates and locations.

Radiocarbon dates reported by Burgis (1970) on leaf material in the Imlay Channel, interpreted as marking the abandonment of the outlet through which Lake Maumee drained into Early/Later Lake Saginaw, indicate that the retreat of LIS ice from the tip of the "thumb" is estimated at 16,043 ± 288 cal. years B.P. (13,770 ± 210 B.P.). Radiocarbon dates reported by Goldthwait (1958) on woody material near Cleveland, Ohio are interpreted as the Lake Arkona low-stand elevation (212m) by 15,868 ± 658 cal. years B.P. (13,600 ± 500 B.P.) (Table 2.1; Monaghan and Hansel, 1990). In the Erie basin, Lake Arkona shorelines rise in elevation to 230-235m due to isostatic rebound (Barnett, 1985).

**2.2.3 Makinaw Interstade**

As the lobes of the Laurentide Ice Sheet continued to oscillate, a retreat event during the end of the Lake Arkona phase uncovered a drainageway to the east that was at a lower elevation than the Maple-Grand Valley. Hough (1958) called this the Cary-Port Huron interval, and presented evidence for a possible low lake stage within the Erie basin, called Lake Ypsilanti.

The exact elevations of this low lake stage are not known, as they are below the present day Lake Erie, but Kunkle (1963) indicates, citing proxy evidence, that they are between 166m and may be as low as 114m. Monaghan and Hansel (1990) provided correlative evidence of a low-water phase in the Lake Michigan basin and provided a $^{14}C$ date of 15681 ± 183 cal. years B.P. (13,470 ± 130 B.P.) for it (Table 2.1). This date represents the end of the low-water phase, and beginning of lacustrine sedimentation at an elevation of 178m at the point where the $^{14}C$ sample was collected, near the town of Riverside in Berrien County, Michigan. They suggested that incremental lowerings of Lake Arkona occurred as the ice sheet receded and uncovered new, lower outlets to the east, in Ontario and New York State (Monaghan and Hansel, 1990). One inference that can be made from these possible low lake stands in the Erie and Lake Michigan basins is that the Saginaw Lowlands would have been dry during this period. This low-water phase was likely very short lived, considering the age constraint of the radiocarbon date of last-Arkona stage (15,868 ± 658) and the timing of the Port Huron re-advance (~16,000); the Ypsilanti low stage may have lasted only a few hundred years. When ice began re-advancing during the Port Huron phase, these low elevation drainageways, as well as the Straits of Mackinaw which connect the Lake Michigan and Huron basins, would have been blocked, ponding meltwater in the basins and once again inundating the Saginaw Lowlands (Hough, 1958; Blewett and Winters, 1995; Krist and Lusch, 2004; Blewett et al., 2009).

### 2.2.4 Glacial Lake Saginaw

By around 16,000 cal. years ago, glacial ice had re-advanced to the Port Huron position, building the moraine of the same name, and once again closing off the drainageway at the tip of the “thumb”, which had previously connected the Saginaw and Erie basins (Monaghan and

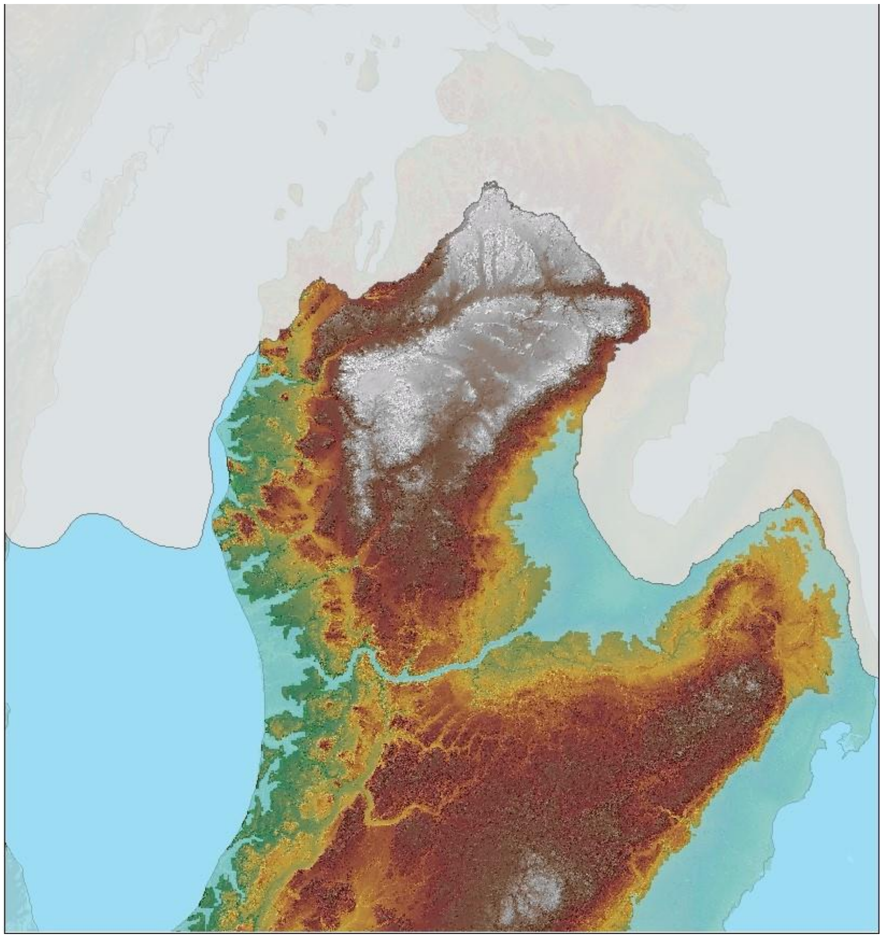


Figure 2.6 – Representation of the approximate configuration of glacial lobes (white) and lakes (blue) for Michigan during Glacial Lake Saginaw time.

Hansel, 1990; Blewett and Winters, 1995; Krist and Lusch, 2004; Blewett et al., 2009; Kincare and Larson, 2009; Fig. 2.6). Lake water ponded in both basins; Glacial Lake Saginaw (GLS) occupied the Saginaw Lowlands, and Lake Whittlesey occupied the Erie basin. Lake Whittlesey waters drained around the ice margin, through a spillway near Ubly, in Huron County, into GLS. GLS had by then resumed westward drainage through the Maple-Grand Valley (Bretz, 1951; Monaghan and Hansel, 1990). The Port Huron moraine marks the position of the ice margin at this time, although in places that were inundated with lake water, the Port Huron tills were waterlain and the moraine was wave-planed. In these places the landform does not stand in as much prominence as in places where tills were deposited subaerially. Monaghan and Hansel

(1990) indicated that the elevation of this lake is equivalent to that of the Arkona low stage (212m). Lusch (1983) mapped Glacial Lake Saginaw beach ridges throughout the Saginaw Lowlands, all of which are NE of the zero isobase for Lake Saginaw – Lake Whittlesey (Calkin and Feenstra, 1985) and, therefore, are deformed upward in the range of 218m to 228m.

**2.2.5 Lake Warren**

By around 15,000 years ago, the tip of the "thumb" once again became ice-free, allowing the formation of Lake Warren in the Erie and Saginaw basins (Figs. 2.7, 2.8). Much like Lake Arkona, Lake Warren existed as one lake connecting both basins. Lake Warren had two main stages, separated by about 3m of elevation (Bretz, 1951; Krist and Lusch, 2004; Kincare

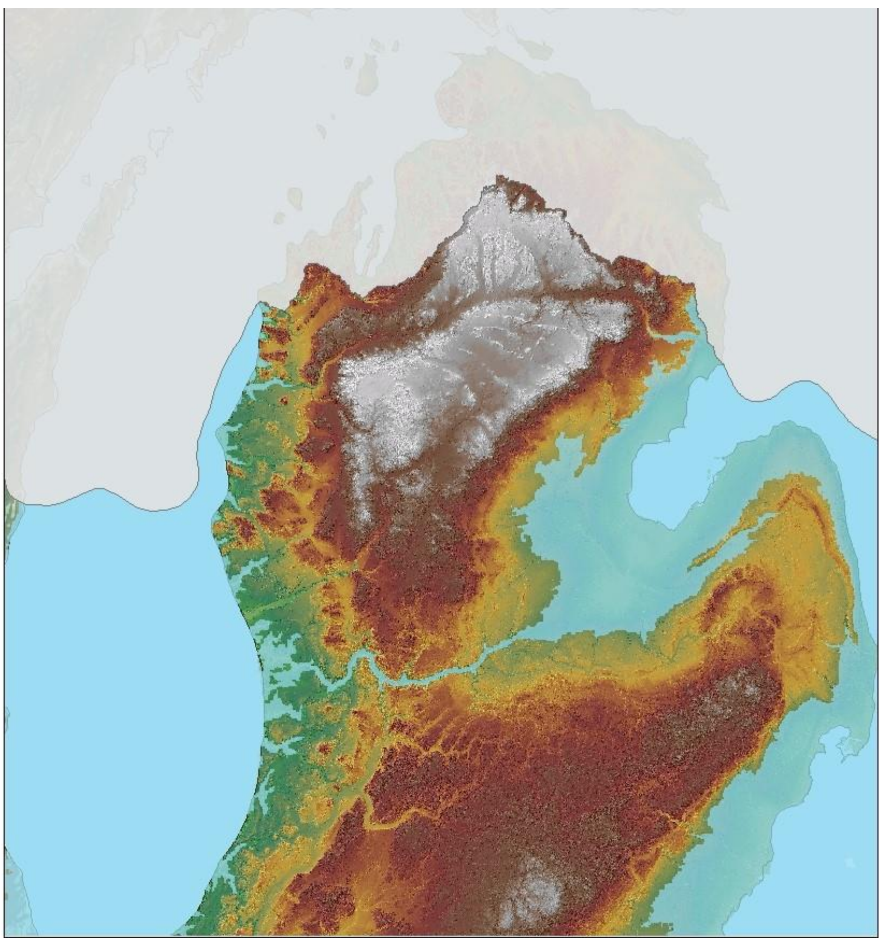


Figure 2.7 – Representation of the approximate configuration of glacial lobes (white) and lakes (blue) for Michigan during Lake Warren time.

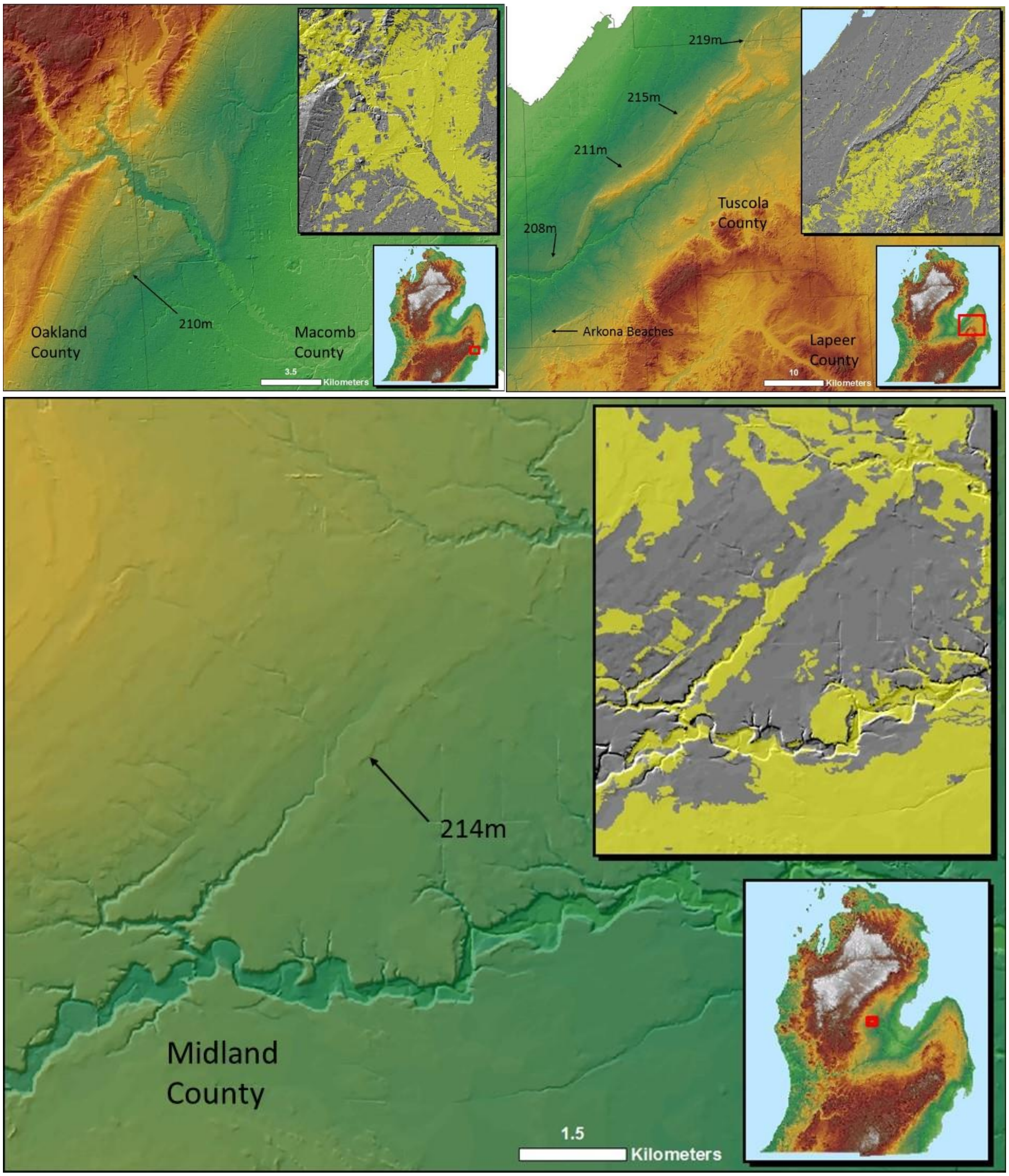



Figure 2.8 – DEM and hillshade showing Lake Warren beach ridges (main map) and their associated elevations. The lower right side inset map shows location and extent within Michigan, the upper right side inset map shows sand textured subsoil on these Warren ridges (data from the USDA-NRCS) Maps are within Tuscola County (upper right), the border of Macomb and Oakland Counties (upper left), and northwestern Midland County (bottom) which is within the study area extent.

and Larson, 2009). In Ontario, shorelines associated with Lake Warren were reported at isostatically uplifted elevations of 235m and 238m, higher than the Lake Warren shorelines further south in the Erie basin that were not subject to uplift, reported at 210m and 207m (Kunkle, 1963; Barnett, 1985; Fig. 2.8). Some sources cite three separate elevations for Lake Warren, two during the initial (and episodic) retreat of the ice margin from the Port Huron position, and a third after the Lake Wayne stand (see below). This triplet of beaches was likely due to ice marginal retreat and re-advance, lowering the water plain to Lake Wayne elevation and then raising it to (lowest) Lake Warren III elevation at 205m (Kunkle, 1963).

**2.2.6 Lake Wayne**

It is generally accepted that retreat of the ice margin from the Port Huron position again opened a drainageway to the east in Ontario or New York, causing a relatively short-lived, low water stand called Lake Wayne to form in both the Erie and Saginaw basins (Bretz, 1951; Larson and Schaetzl, 2001). Bretz (1951) suggested Lake Wayne was a precursor to Lake Warren, and that it occupied the Saginaw and Erie basins during maximum retreat before the Port Huron re-advance. Indeed, Leverett and Taylor (1915) indicated that the Lake Wayne beaches show evidence of alteration from submergence and suggested that the Wayne beach appears to be older than the Warren beaches. The evidence of beach submergence can be explained by a third Lake Warren stage (205m), at a higher elevation, after the Lake Wayne stage had formed (200m). Most sources place Lake Wayne either after Lake Warren or between the Lake Warren highest and lowest stand elevations, but there is some chronological uncertainty (Barnett, 1985; Hough, 1958). Lusch (1982) associated several weak shorezone features at 200m - 201m on top of the waterlain Port Huron Moraine in western Bay and eastern Midland counties with Lake Wayne.

**2.2.7 Subsequent Lake Stages**

In the Saginaw Lowlands, Lakes Grassmere (195m) and Elkton/Lundy (189m) chronologically follow the Lake Wayne and Lowest Lake Warren (third stage). However, their shorelines are found to the east, beyond the study area (Fig. 2.9). Leverett and Taylor (1915) claimed that both of these stages were a product of ice retreat and uncovering of sequentially lower outlets, draining the lakes to elevations commensurate with the elevation of the outlet's

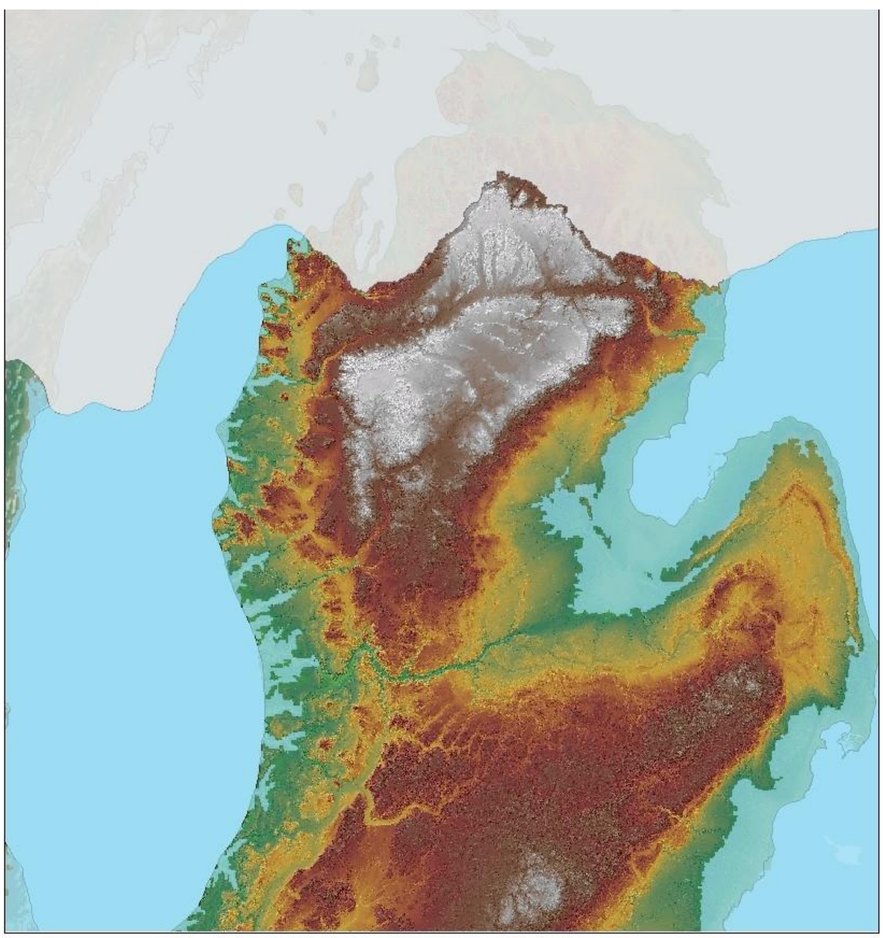


Figure 2.9 – Representation of the approximate configuration of glacial lobes (white) and lakes (blue) for Michigan after Lake Warren, when the study area would have been subaerial.

knickpoint. Hough (1958), however, believed that the Lake Grassmere and Lundy stages correlate with lake levels of equal elevation in the Lake Michigan basin, and therefore likely drained west through the Maple-Grand Valley. . However, the published elevations of Lakes Grassmere and Lundy are significantly below the threshold at the head of the Maple-Grand drainageway, making it unlikely to be the outlet for these proglacial lakes. Regardless, after the last stages of the lowest Warren elevation, the study area wasn't again inundated with proglacial lake waters, so for the purpose of this study I have focused primarily on those lake levels which would have affected delta progradation and modification therein.

**2.3 Summary**

Evidence of the complex history of fluctuating lake levels in the Saginaw Basin is preserved in the relict shorelines and wave-cut bluffs that the lakes left behind on the landscape. Many of the relict shorelines within the study area are discontinuous, and many more are faint or equivocal, likely because they were either not strongly formed, or were not well preserved because they are sandy, or a combination of these. Leverett and Taylor (1915) and the map from Martin (1955), which was largely derived from Leverett and Taylor's work, reported a wealth of information concerning the locations, elevations, and characteristics of relict shorelines within the Saginaw and Erie basins on the eastern side of Michigan. More recent mapping efforts like Farrand and Bell (1982) and Lusch et al. (2009) have produced GIS shapefiles of shorelines for the region (Fig. 2.10), but they can be improved to be more spatially accurate. The process by which I made these improvements is discussed in the methods section, and the dataset that I compiled will be reported and discussed in the results section. The refined shoreline dataset compiled for this study is useful for interpreting the relative timing of delta progradation, and the sequence of events in this region with respect to the various lake stages within the Saginaw Lowlands. These interpretations will be discussed further in the results and discussion section of this thesis.

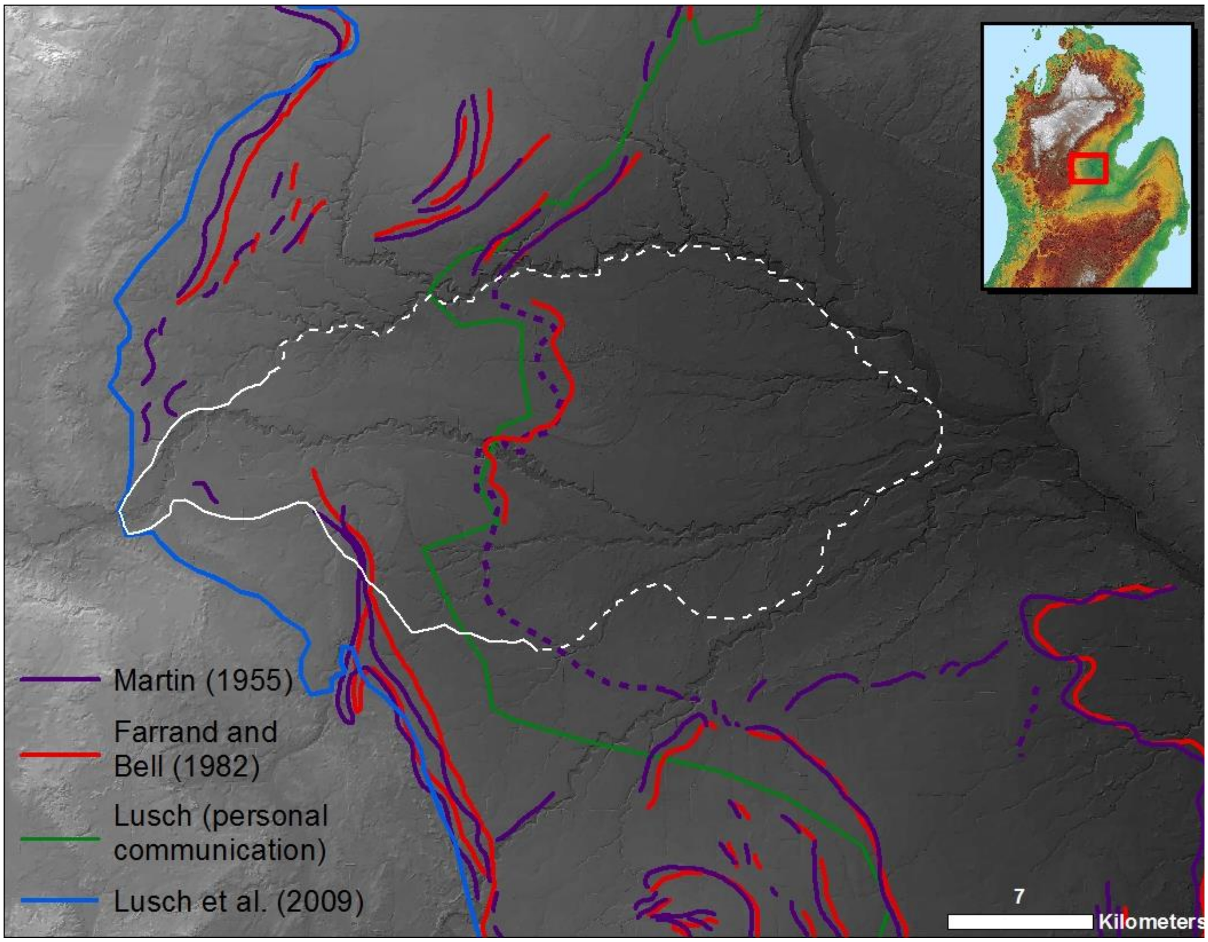



Figure 2.10 – Shorelines within the study area from previous mapping efforts, over a hillshade DEM with the outline of the Chippewa Delta (The delta outline is dashed where there is some uncertainty to the boundary's exact location, and solid where boundary location is more certain). Only the Martin (1955) shorelines are labeled, as Farrand and Bell did not label their shorelines with the associated lake stages. Lusch et al. (2009) is the highest Lake Saginaw shoreline, while Lusch (personal communication) is an unpublished Lake Warren shoreline.

## 3. Study Area

The study area extent is roughly coincident with Isabella and Midland Counties in central Lower Michigan. The terrain is made up of glacial deposits, consisting of loamy, till-cored morainic uplands in the western part, adjoining a comparatively low relief proglacial lake plain on the eastern edge (Fig. 3.1). The morainic upland adjacent to the lake plain is characterized by parallel, concentric, till-composed ridges formed by the Saginaw lobe, with inter-morainal till plains and outwash plains (Fig. 3.2). The till plain soils, like the morainic deposits, are loam textured, which is characteristic of Saginaw lobe drift. Where present, the outwash surfaces

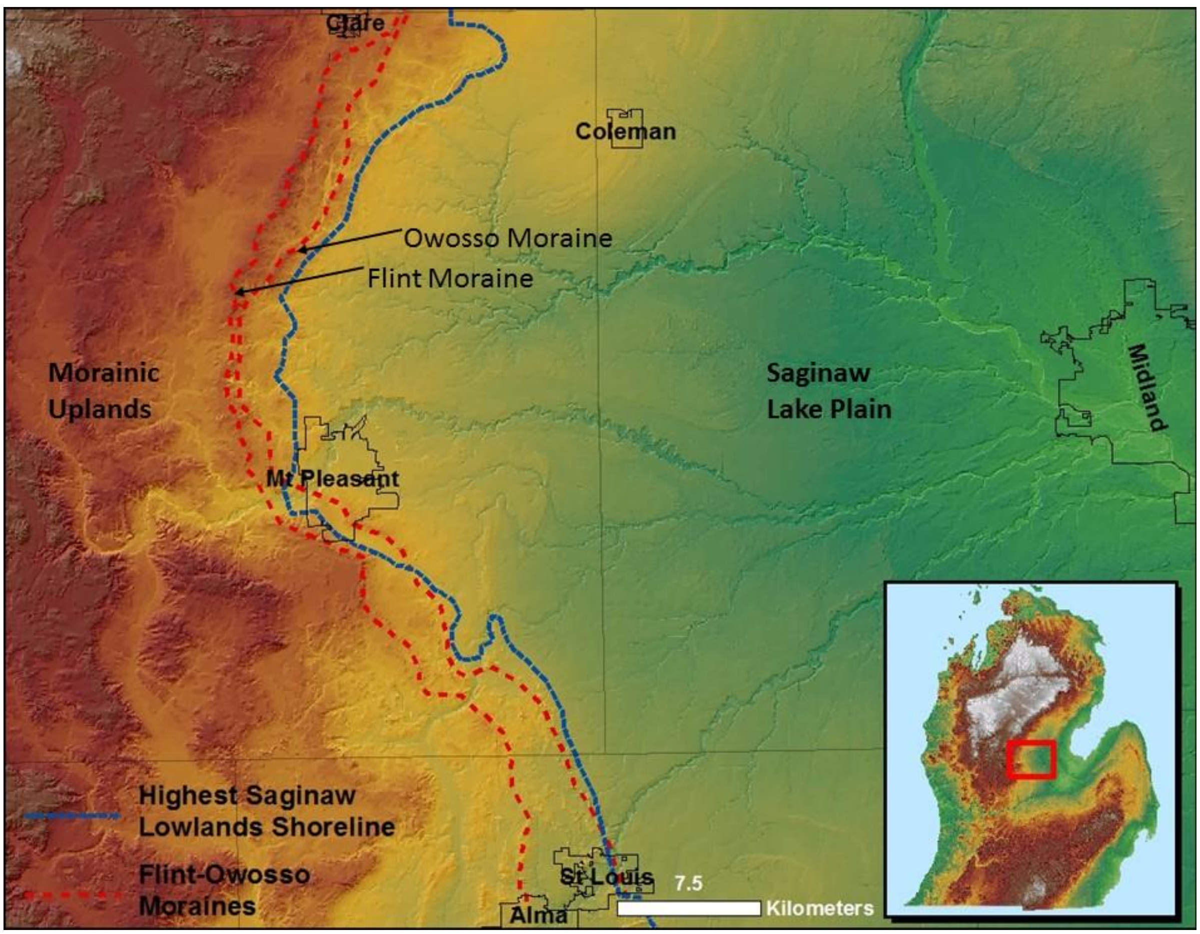



Figure 3.1 – DEM and hillshade showing the topography within the study area and the locations of major cities and villages, as well as the Flint-Owosso moraines and the highest shoreline in the study area. Inset map indicates the study area location within central lower Michigan (data from USGS).

between the morainic ridges are sandy. The Chippewa River drains from northwest to southeast, across the study area, cutting through the moraines and till plains as it makes its way to the Saginaw lake plain through the city of Mt. Pleasant. The transition between the morainic uplands and the lake plain is marked by the easternmost moraine, the Flint-Owosso moraine, which appears in some places as two distinct ridges and in others as a single ridge, with the Owosso moraine "piled on the back of the Flint moraine" (Bretz, 1951, p. 245). Between the morainic uplands and the lake plain is about 30m of elevational change. The lake plain itself is a low-relief landscape, typically consisting of slopes of <2%.

The delta was initially identified using USDA-NRCS soil data (Figs. 3.2, 3.3). The delta boundary was delimited by analyzing NRCS soil data, the delta's morphology and its association with neighboring landforms, as well as the thickness and spatio-textural data generated for this study. The delta extent, represented in figures throughout this thesis, is mapped with a dashed line where exact boundaries are uncertain, and with a solid line where the boundary is more certain.

The soils on the delta are mainly sandy, and stand in contrast with the loamy background soils on the lake plain, which are comprised of till and lacustrine sediments (Fig 3.2). Most of these sands are deep, comprising the main body of the delta, but where they are shallow they overlie the loamy substrate that is typical of the lake plain (Fig. 3.2). The distribution of shallow sandy soils helped delimit the delta edges, representing the periphery of the landform. Additionally, the sand textured sediments have a lobe shaped distribution, which is characteristic of deltas (Galloway, 1975; Leeder, 1982; Boggs, 2001; Fig 3.3). Indeed, the proximal part of the delta had been previously mapped by Martin (1955), but the lower portion

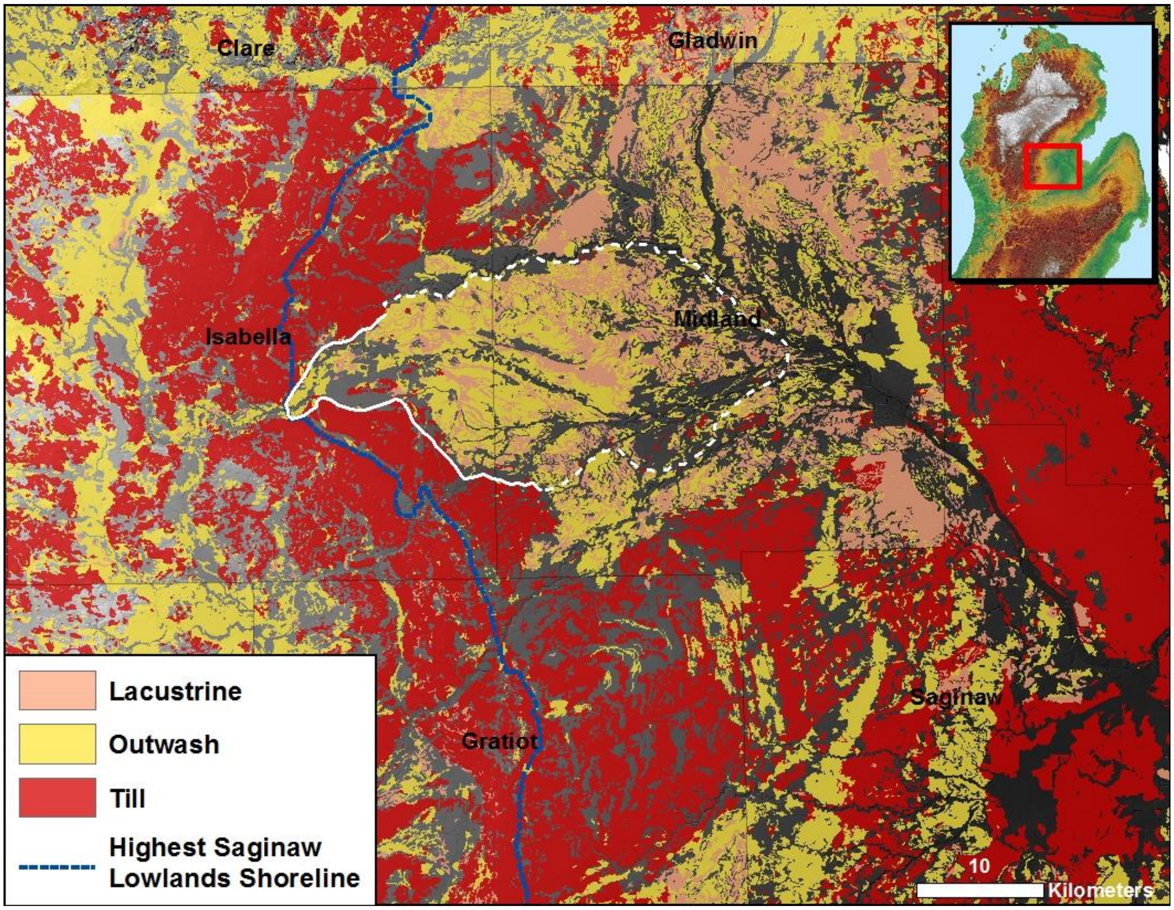



Figure 3.2 – Extent of the Chippewa Delta, shown on a map of soil parent material over DEM hillshade in an 11 county extent. Morainic uplands west of the lake plain consist of Saginaw Lobe moraines and inter-morainal outwash plains. The gray color of the hillshade is visible in places where a distinct parent material is not specified in the soil data (data from USDA-NRCS).

of the delta east of Mt. Pleasant was not mapped as deltaic, but as “Lake beds, sand”, which does not assign a depositional history to the sandy sediment (Fig. 3.4). More recent research conducted as part of a doctoral dissertation did identify the sediments in the distal portion of Martin’s Chippewa Delta as deltaic (Luehmann, personal communication), but did not characterize this area in detail.

The data from this study indicate that the delta is larger than previously mapped by Martin; it is roughly 18km wide and 38km long, and its sediments are around 5-6m thick in the

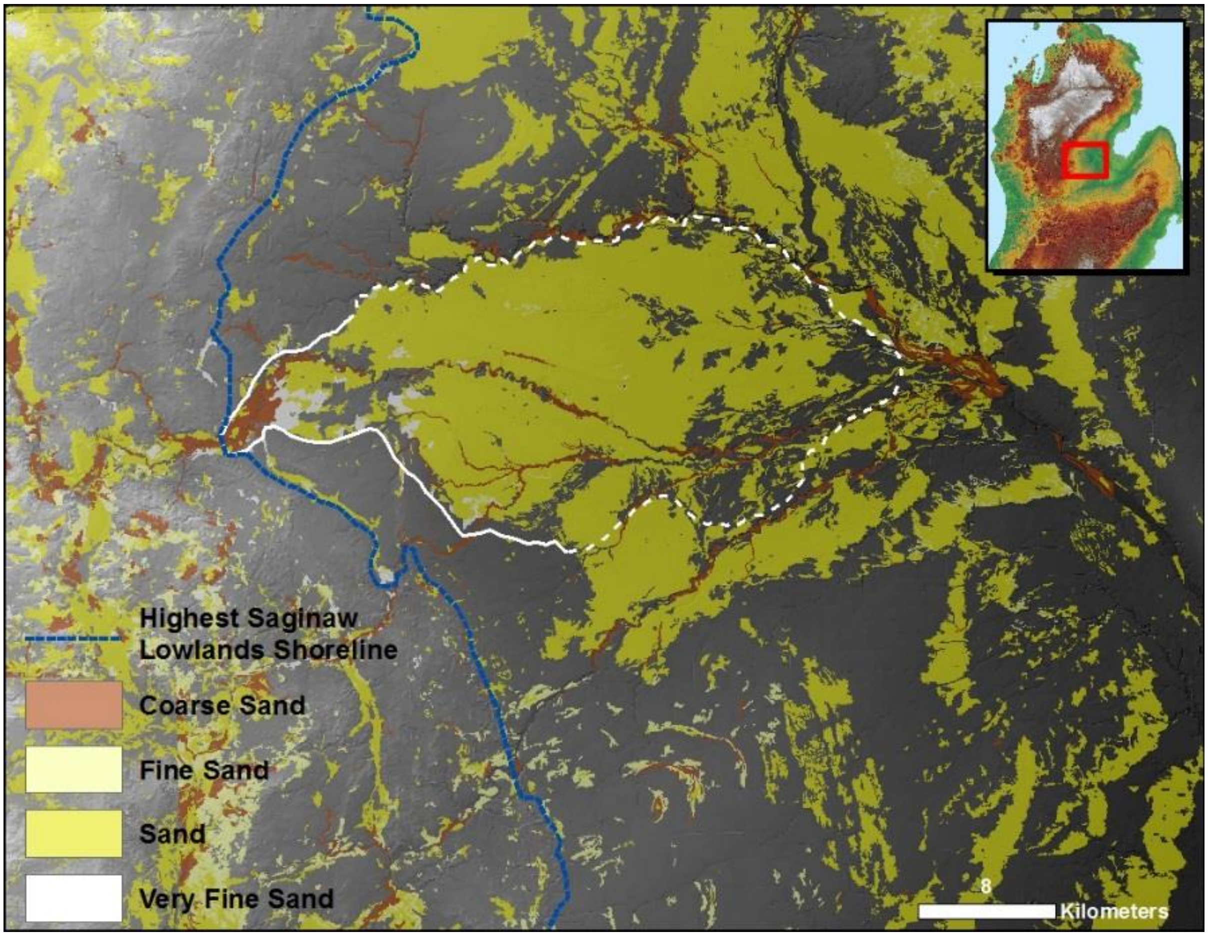


Figure 3.3 – Sandy subsoil texture and the delta extent over DEM hillshade (data from USDA-NRCS).

center, thinning with distance from the river. It is also worth noting that the present day Chippewa River is deeply incised, having eroded into the deltaic sediments after the landform had been emplaced.

Finally, I analyzed the characteristics of neighboring landforms, specifically looking for sand textured soils that could have been the source sediment for the delta. Directly to the west of the delta are loamy morainic uplands and their sandy inter-morainal outwash plains, through which the Chippewa River flows. A portion of the upland outwash plains are within the present-day Chippewa River watershed, which drains 1059 $km^2$ in Isabella, Clare, Osceola, and Mescosta Counties, indicating that there was, and still is, an ample supply of sand upstream of the delta.

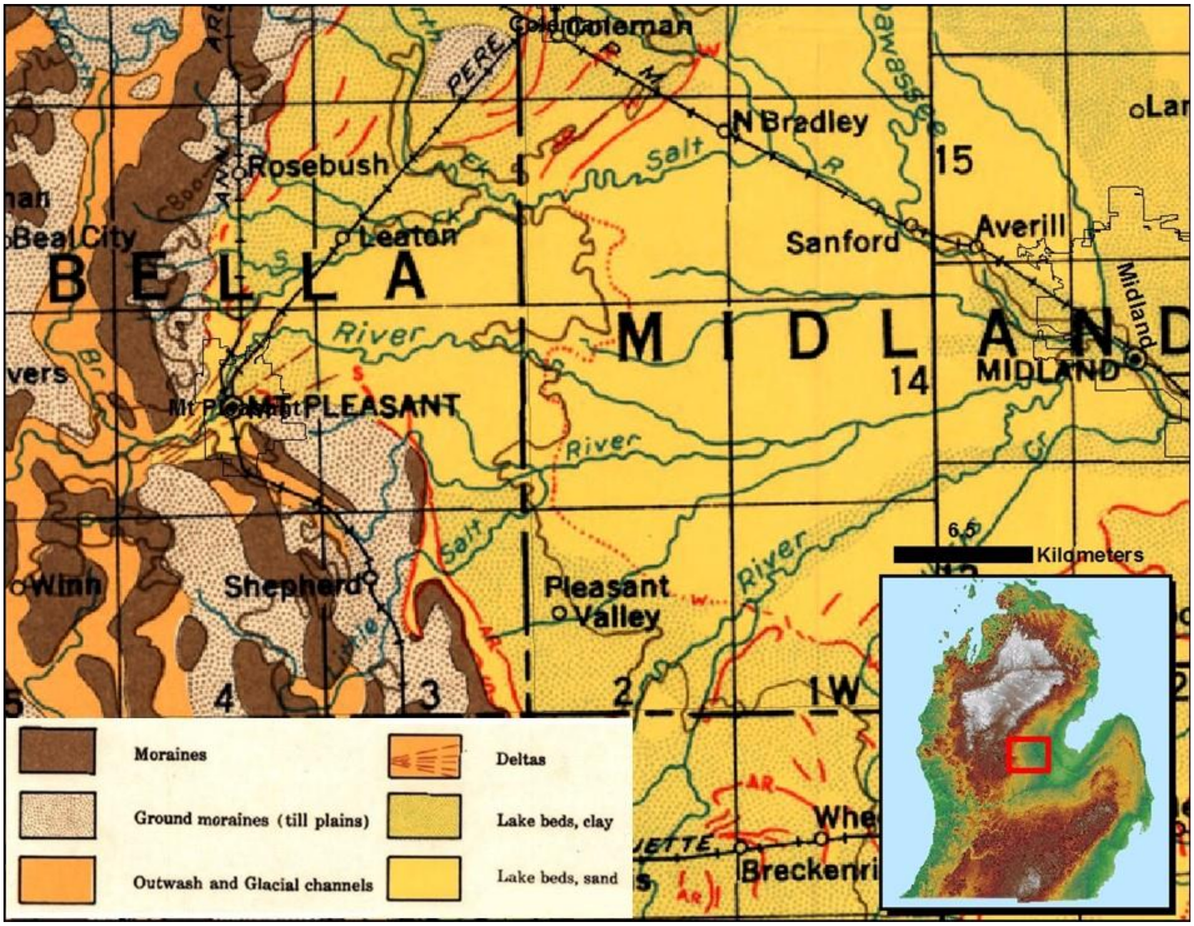


Figure 3.4 – Part of the 1955 surficial geology map by Helen Martin, “Map of the Surface Formations of the Southern Peninsula of Michigan”, showing the area around Mt. Pleasant mapped as deltaic.

These initial observations helped to direct the more detailed study into this delta. Because this research relied heavily on soil data, I will, in the following section, discuss the soils in the study area that were analyzed to draw inferences about the delta.

### 3.1 Soils

Michigan's landscapes are geologically young, and are almost entirely formed in glacial sediment (Hough, 1958; Farrand and Eschman, 1974; Kincare and Larson, 2009). Because of this relative youth, the soils that form in Michigan's glacial drift preserve signals of the processes which emplaced the sediment. The USDA-NRCS digital soil data are the most detailed, precise, and most frequently updated data available for the Michigan, and thus are useful for studying the characteristics of the landforms which make up this glacial landscape. In this section, I will discuss the sandy soils on the delta, the loamy and sandy soils in the morainic uplands and outwash plains, the loamy soils on the lake plain, and the two-storied soils which have a sandy upper solum and loamy lower horizons. Soil data in this section are from the USDA-NRCS official soil series descriptions (OSDs http://www.nrcs.usda.gov/wps/portal/nrcs/detailfull/soils/home/?cid=nrcs142p2_053587).

**3.1.1 Sandy Soils on the Delta**

Soils across the delta surface are sandy and deep, consisting of two major and six minor sandy soils. These sandy soils contrast with the typically loam textured soils in the lowlands on the lake plain; areas on the lake plain that have deep sandy soils can be readily identified because, unlike the loamy soils, they are agriculturally poor and are typically forested. In the following section I will discuss the major and minor sandy soils on the delta.

**3.1.1.1 Major Sandy Soils on the Delta**

Pipestone and Kingsville are the most extensive soils in the study area, consisting of 76% (Kingsville 32%, Pipestone 44%) of the deep sandy soils across the delta surface (Table 3.1, Fig. 3.5). Pipestone soils are described as forming in “sandy outwash deposits…on outwash plains, lake plains, beach ridges and water-worked till plains” (OSD). Although Pipestone soils are found on slopes ranging from 0-8%, their dominant slope range is 0-4%. Kingsville soils are described as having formed in “glaciolacustrine sediments...on deltas and offshore bars on Wisconsinan age lake plains” (OSD) (Table 3.1).

| Series | Taxonomic Subgroup | Drainage Class | Parent Material (from OSD) | Landform | DI* |
|---|---|---|---|---|---|
| Pipestone | Sandy, mixed, mesic Typic Endoaquods | Somewhat poorly drained | Sandy outwash deposits | Outwash plains, lake plains, beach ridges, and water-worked till plains | 66 |
| Kingsville | Mixed, mesic Mollic Psammaquents | Very poorly drained | Glaciolacustrine sediments | Wisconsinan age lake plains | 76 |
| Mescota | Sandy-skeletal, mixed, mesic Typic Udorthents | Somewhat excessively drained | Sand and gravel deposits | Outwash plains and moraines | 31 |
| Thetford | Sandy, mixed, mesic, Aquic Arenic Hapludalfs | Somewhat poorly drained | Sandy glacial till or glacial outwash | Ground and end moraines, lake plains, outwash plains, terraces and beach ridges | 62 |
| Covert | Sandy, mixed, mesic, Oxyaquic Haplorthods | Moderately well drained | Sandy glacial drift | Ground moraines, outwash plains, lake plains, and dunes | 51 |
| Plainfield | Mixed, mesic Typic Udipsamments | Excessively drained | Sandy drift | Outwash plains, valley trains, glacial lake basins, stream terraces and moraines and other upland areas | 14 |
| Oakville | Mixed, mesic Typic Udipsamments | Excessively drained | Dune sand | Dunes and beach ridges on outwash plains, lake plains and moraines | 44 |
| Kinross | Sandy, mixed, frigid Typic Endoaquods | Poorly drained and very poorly drained | Glaciofluvial material | Outwash plains, stream terraces, lake plains, kame, disintegration and ground moraines | 91 |

Table 3.1 – Characteristics of sandy soils on the delta (USDA-NRCS). *Drainage Index (DI).

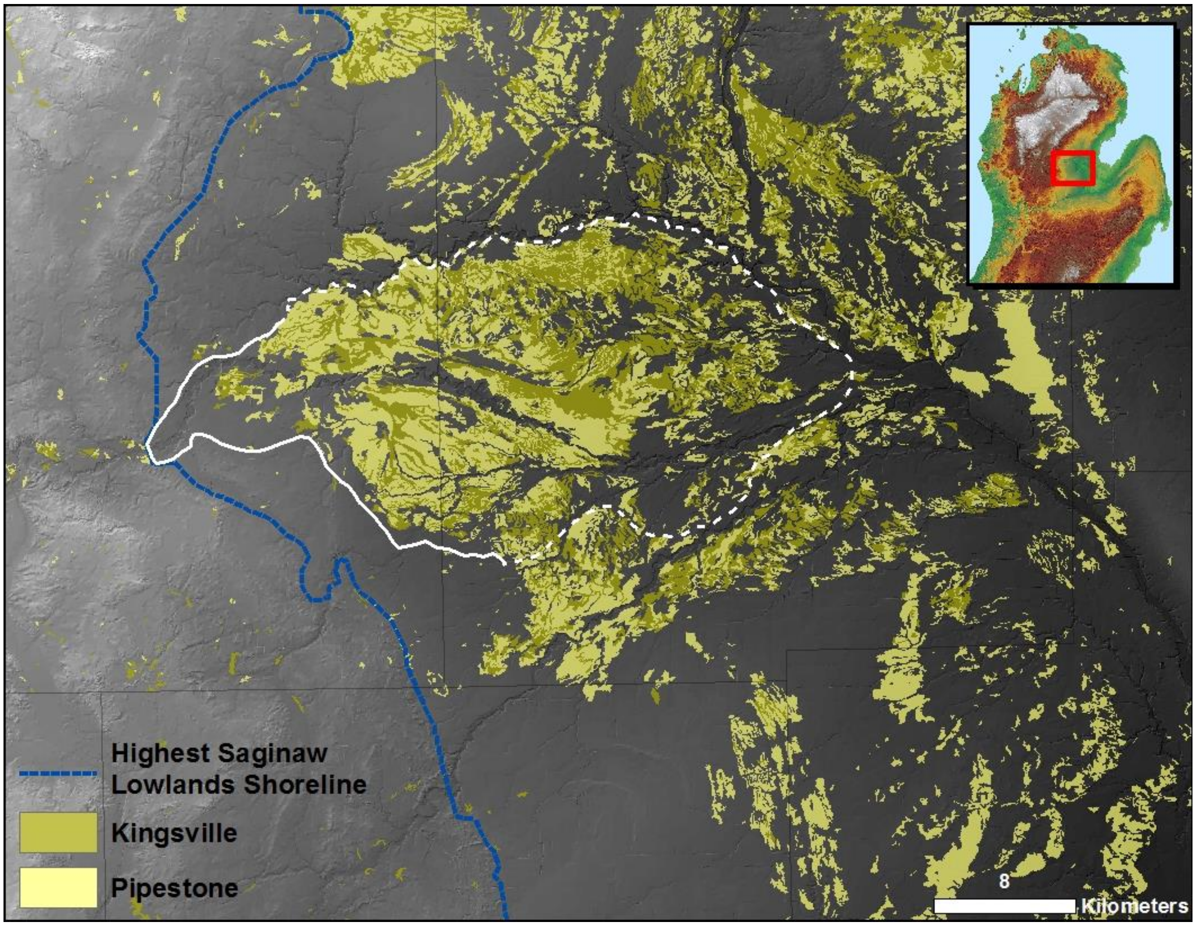


Figure 3.5 – The distribution of major sandy soils (Kingsville and Pipestone) in the study area. These two major soils comprise 76% of deep sandy soils on the delta surface (data from USDA-NRCS).

Pipestone and Kingsville soils are in a drainage association with one another, with Pipestone being somewhat poorly drained, and Kingsville being very poorly drained. Oakville, an excessively drained minor soil in the study area, is also in this drainage association, and is found mainly on isolated uplands and small sand dunes (Table 3.1). I interpret the Kingsville and Pipestone soils in the study area as having been formed in delta sediments because of 1) the lobate shape of their distribution within the study area, 2) their proximity to the Chippewa River and relict shorelines within the lowlands, 3) and the knowledge that proglacial lakes occupied this region in geologic history.

### 3.1.1.2 Minor Sandy Soils on the Delta

Several minor soils also occur on the delta, covering 24% of the total area of deep, sandy soils (Fig. 3.6). I will discuss these soils as they appear from the western edge of the delta, near Mt. Pleasant, to the eastern edge, near Midland.

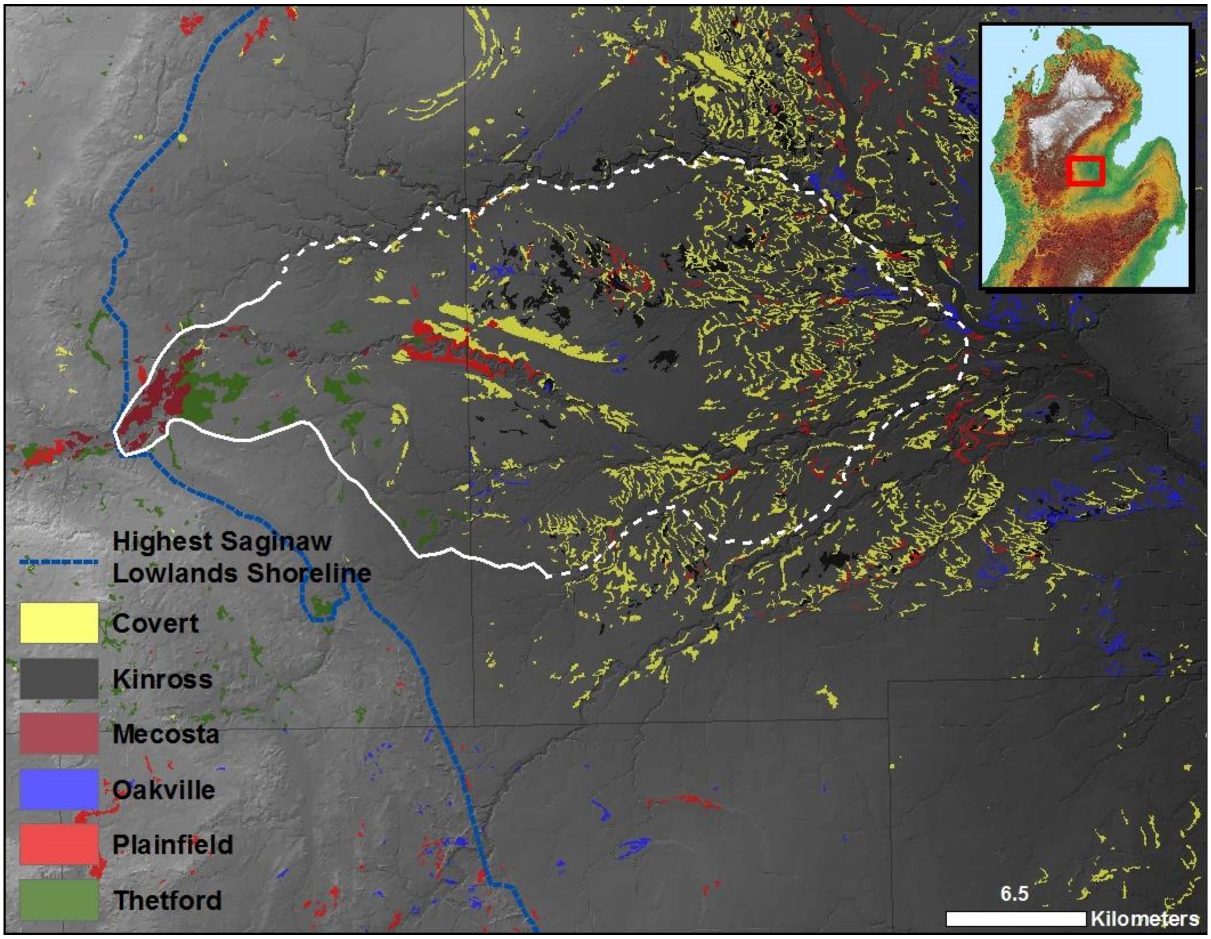



Figure 3.6 – The distribution of minor sandy soils (Covert, Kinross, Mescota, Oakville, Plainfield, and Thetford) in the study area. These six minor soils comprise 24% of deep sandy soils on the delta surface (data from USDA-NRCS).

Mescota soils (1%), the coarsest and most gravelly of the sandy soils, are mostly found at the transition between morainic upland and the delta plain, near the city of Mt. Pleasant, but also occur sparingly further east on the delta. Here, they are mainly found within the incised Chippewa River channel. The Mescota series is described in the OSD as having "formed in sand

and gravel deposits...on outwash plains and moraines” (OSD) (Table 3.1). It is noteworthy that the majority Mescota soils, the coarsest and gravelliest sandy soil in the study area, is mapped near the highest Saginaw Lowlands shoreline, where the Chippewa River would have discharged into the proglacial lake. The somewhat excessively drained Mescota soils are in a drainage association with the moderately well drained Covert series.

The Covert series is the most extensive of the minor sandy soils on the delta (14%). Covert soils have formed in “sandy drift on ground moraines, outwash plains, lake plains, and dunes” (OSD), and are in an association with the somewhat poorly drained Pipestone soils, and the very poorly drained Kingsville soils. Within the study area, Covert soils occupy three main landscape positions: abandoned beach ridges (Arkona beaches in northwestern Midland County), dunes along the eastern delta surface, and on an abandoned Chippewa River floodplain near the border between Isabella and Midland Counties.

Thetford soils (3%) have formed in “sandy till or outwash on ground moraines, end moraines, lake plains, outwash plains, terraces, and beach ridges” (OSD). The Thetford series is a somewhat poorly drained member in a drainage association with the well drained Oakville soils. Within the study area, Thetford soils are generally located just east of most of the Mescota soils, near the city of Mt. Pleasant. They do not appear further east on the delta surface beyond the border of Isabella and Midland Counties.

Plainfield soils (2%) have formed in “sandy drift on outwash plains, valley trains, glacial lake basins, stream terraces, and moraines and other upland areas” (OSD). Indeed, much of the extent of Plainfield soils within the two county extent occurs on outwash surfaces in the morainic uplands to the west of the delta. On the delta, Plainfield soils appear on alluvial terraces and sparingly within dune fields.

Oakville soils (1%) have formed in “sandy eolian deposits on dunes and beach ridges on outwash plains, lake plains, and moraines” (OSD). Oakville soils within the study area are located at the far eastern periphery of the delta surface, close to the city of Midland, although areas mapped as Oakville also occur near abandoned shorelines, and on sand dunes and stream terraces. The majority of Oakville soils in the study area form on dunes.

Kinross soils (3%) are the sandy soils of low, swampy areas. They are characterized as muck over sand and have formed in “glaciofluvial material on outwash plains, stream terraces, lake plains, kame, disintegration and ground moraines” (OSD). Within the study area, Kinross soils occur mainly near the eastern periphery of the delta and in interdunal areas; generally, they have a discontinuous distribution.

### 3.1.2 Sandy Soils on Uplands

Within the moranic uplands there are a number of sandy soils on outwash plains between moraines, and in small fluvial channels. Among these soils are Coloma, Graycalm, Grayling, Kalkaska, Montcalm, Spinks and Vestaberg (Fig. 3.7, Table 3.2). Each of these soils has formed in outwash deposits, sandy drift or sandy deposits on outwash plains, valley trains, terraces, moraines, and kames (OSD). Grayling soils are also described as forming in outwash deposits on deltas, but they are not mapped on the Chippewa delta surface. Additionally, Plainfield and Mecosta soils occur on the morainic uplands, as well as on the delta surface, as discussed in the previous section. Mecosta soils are adjacent to fluvial channels, and Plainfield

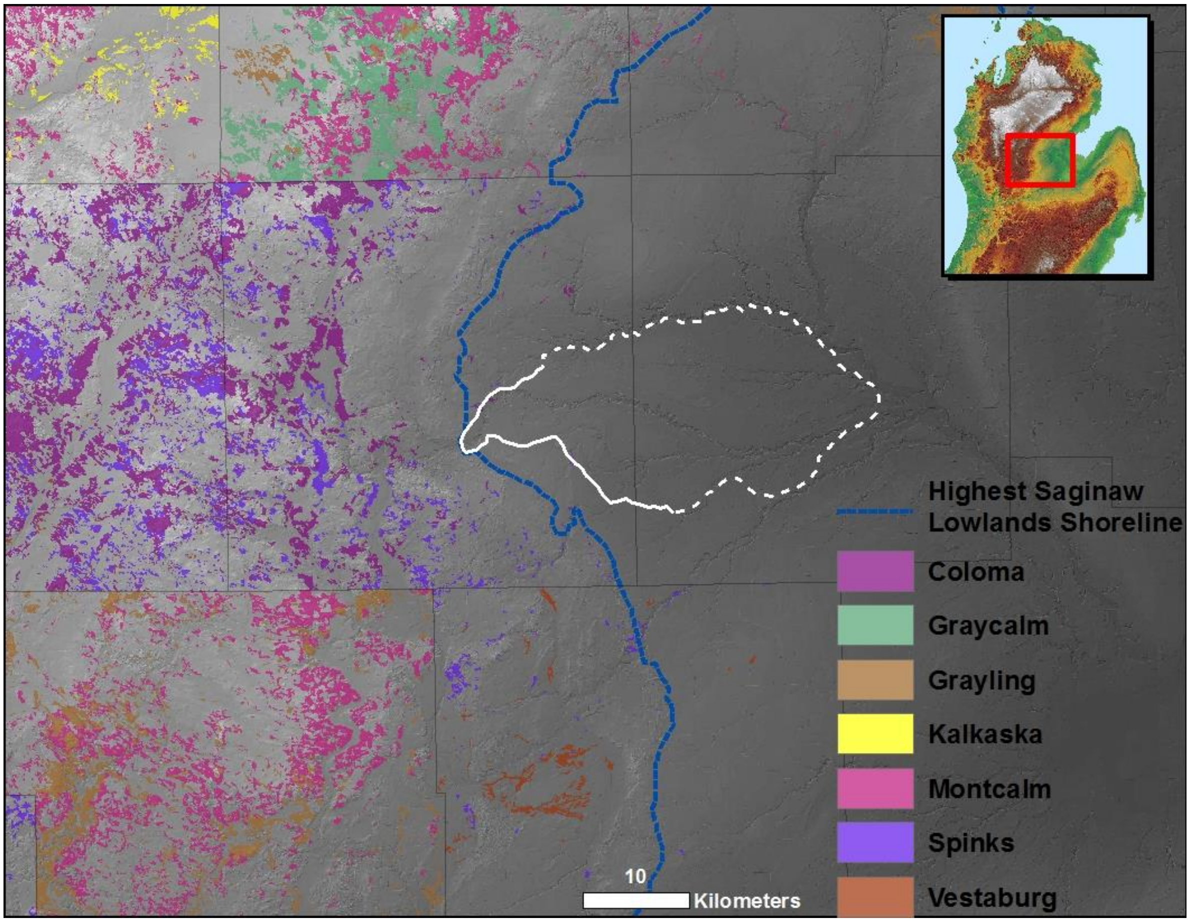


Figure 3.7 – The distribution of sandy upland soils (Coloma, Graycalm, Grayling, Kalkaska, Montcalm, Spinks, and Vestaburg) in this study area (data from USDA-NRCS).

soils are on outwash surfaces, adjacent to Coloma soils, which are loamier in the upper solum than are Plainfield soils.

Graycalm and Montcalm soils are found mainly north of the delta, in neighboring Clare County. It should be noted that there is a sharp boundary on the county line; surveyors in Isabella County mapped Spinks and Coloma, whereas surveyors in Clare County mapped Montcalm and Graycalm on roughly similar landscapes.

| Series | Taxonomic Subgroup | Drainage Class | Parent Material (from OSD) | Landform | DI |
|---|---|---|---|---|---|
| Coloma | Mixed, mesic Lamellic Udipsamments | Somewhat excessively drained | Sandy drift | Moraines, outwash plains, deltas and stream terraces | 17 |
| Graycalm | Isotic, frigid, Lamellic Udipsamments | Somewhat excessively drained | Sandy glaciofluvial deposits | Moraines, kames, stream terraces, outwash plains and glacial drainage channels | 27 |
| Grayling | Isotic, frigid, Typic Udipsamments | Excessively drained | Sandy glaciofluvial deposits | Outwash plains, deltas, kames, kame moraines, stream terraces, disintegration moraines and lake plains | 14 |
| Kalkaska | Sandy, isotic, frigid Typic Haplorthods | Somewhat excessively drained | Sandy deposits | Outwash plains, valley trains, moraines and stream terraces | 29 |
| Montcalm | Coarse-loamy, mixed, semiactive, frigid Alfic Haplorthods | Well drained | Sandy and loamy drift | Ground moraines, end moraines and outwash plains | 42 |
| Spinks | Sandy, mixed, mesic Lamellic Hapludalfs | Well drained | Sandy eolian and outwash material | Dunes, moraines, till plains, outwash plains, beach ridges and lake plains | 35 |
| Vestaburg | Mixed, mesic Mollic Psammaquents | Poorly drained or very poorly drained | Sandy deposits | Outwash plains and outwash valley trains | 76 |

Table 3.2 – Characteristics of sandy soils in the study area (USDA-NRCS).

### 3.1.3 Loamy Soils

In this section I will briefly discuss the loamy soils of the study area. These soils dominate the morainic uplands, where they are formed in till, and the lake plain, where they formed in waterlain till and lacustrine sediment.

#### 3.1.3.1 Loamy Soils on Uplands

The major loamy soils in the uplands are Ithaca, Perrinton, and Remus (Fig 3.8; Table 3.3). All of these soils have formed in till, and are found mainly on till plains and moraines. Remus soils are mapped in the uplands to the west of the delta. Ithaca and Perrinton soils form dominantly in the uplands to the northwest of the delta.

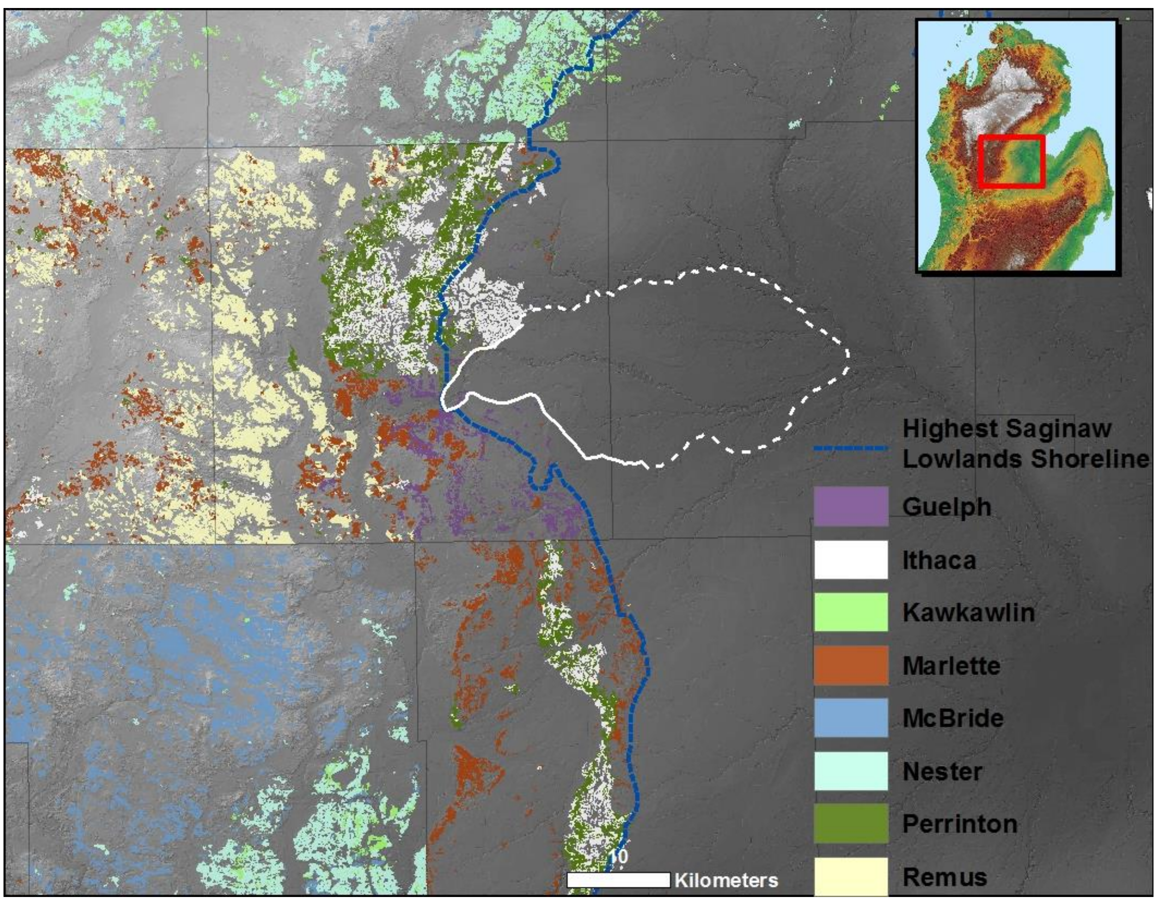



Figure 3.8 – The distribution of loamy upland soils (Guelph, Ithaca, Kawkawlin, Marlette, McBride, Nester, Perrinton, and Remus) in the study area (data from USDA-NRCS).

Guelph, Kawkawlin, Marlette, McBride, and Nester are minor soils formed in the loamy uplands, in till. Guelph and Marlette soils have formed on moraines and till plains, and are found to the southwest of the delta. Kawkawlin, McBride and Nester soils form on moraines to the north of the delta in neighboring Clare County. Like with the soils of the sandy uplands, surveyors in neighboring counties mapped similar soils differently across county borders; surveyors in Clare County mapped Kawkawlin and Nester, while surveyors in Isabella County mapped Ithaca and Perrinton on similar landscape positions.

**3.1.3.2 Loamy Soils on the Lake Plain**

Londo and Parkhill are major soils on the loamy lake plain. Parkhill soils are described as having formed in “ablation till overlying dense till...wave-worked till plains and moraines” (OSD). Londo soils have formed in “loamy glacial till on moraines and till plains” (OSD). Londo and Parkhill are found on the lake plain mostly to the north and south of the delta. Londo soils are also mapped east of the delta, on the east side of the city of Midland, in Bay County. Parkhill soils have a large distribution to the south and southeast of the delta in Gratiot and Saginaw Counties. The minor loamy soils on the lake plain are Bowers, Corunna, Ingersoll, Lenawee, Metamora, Pella, Wauseon, and Ziegenfuss (Fig 3.9; Table 3.4).

Most of the minor loamy soils have formed in till or lacustrine sediment associated with the lake plain. Corunna, Metamora, and Ziegenfuss soils form mostly in the transition between uplands and lowlands, along the western edge of the lake plain and in low-lying areas between moraines. Bowers, Ingersoll, Lenawee, Pella soils are mainly found in the eastern part of the study area, near the city of Midland.

| Series | Taxonomic Subgroup | Drainage Class | Parent Material (from OSD) | Landform | DI |
|---|---|---|---|---|---|
| Ithaca | Fine, mixed, semiactive, mesic Aquic Glossudalfs | Somewhat poory drained | Glacial till | Ground and end moraines | 69 |
| Perrinton | Fine, mixed, active, mesic Oxyaquic Glossudalfs | Moderately well drained | Glacial till | Ground and end moraines | 43 |
| Remus | Fine-loamy, mixed, semiactive, mesic Haplic Glossudalfs | Well Drained | Loamy glacial till | Ground and end moraines | 43 |
| Guelph | Fine-loamy, mixed, active, mesic Haplic Glossudalfs | Well drained or moderately well drained | Loamy till | Ground moraines and end moraines | 43 |
| Kawkawlin | Fine, mixed, semiactive, frigid Aquic Glossudalfs | Somewhat poorly drained | Moderately fine and fine textured glacial till | Moraines | 68 |
| Marlette | Fine-loamy, mixed, semiactive, mesic Oxyaquic Glossudalfs | Well drained | Glacial till | Till plains, ground moraines and end moraines | 43 |
| McBride | Coarse-loamy, mixed, semiactive, frigid Alfic Fragiorthods | Well drained | Glacial till | Ground and end moraines | 43 |
| Nester | Fine, mixed, semiactive, frigid Oxyaquic Glossudalfs | Well drained | Glacial till | Moraines | 43 |

Table 3.3 – Characteristics of loamy soils in the study area (USDA-NRCS).

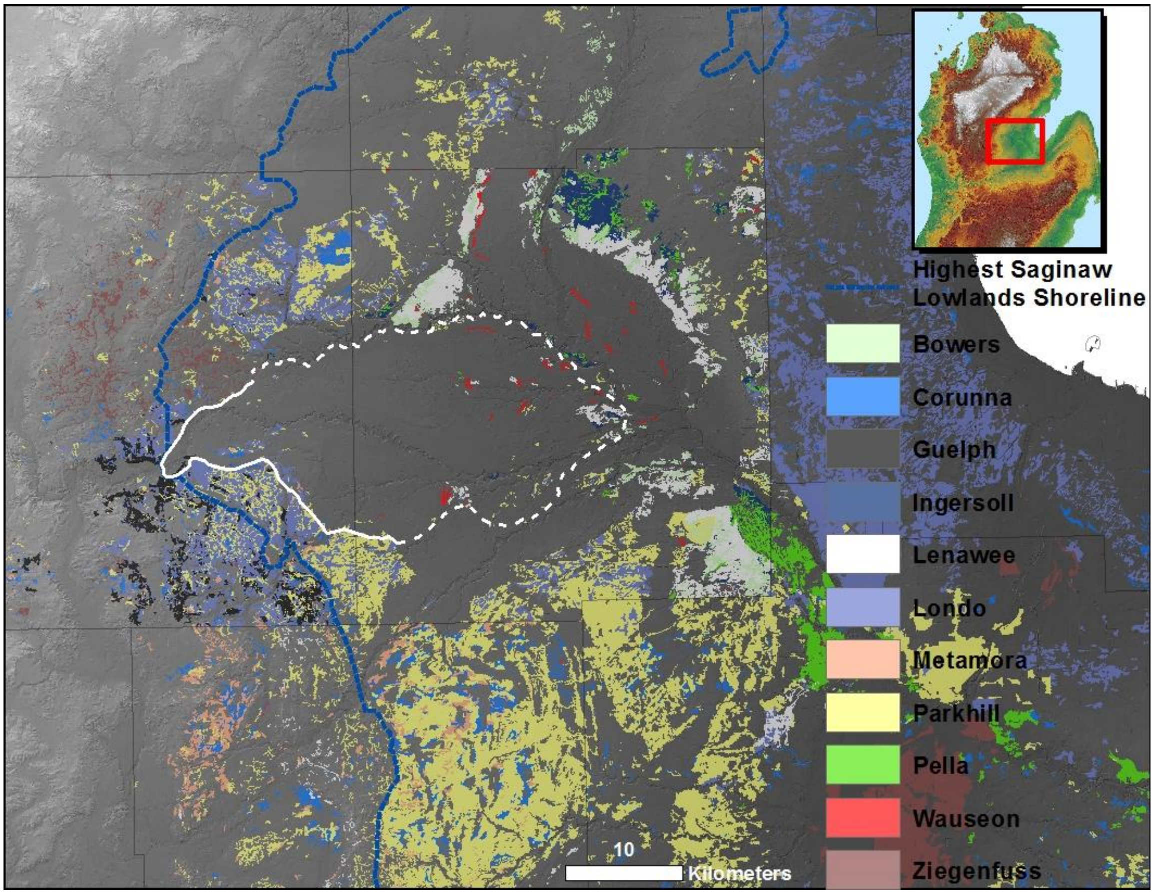



Figure 3.9 – The distribution of loamy soils (Bowers, Corunna, Guelph, Ingersoll, Lenawee, Londo, Metamora, Parkhill, Pella, Wauseon, and Ziegenfuss) in the study area (data from USDA-NRCS).

Wauseon soils have a very sparse distribution, but appear on the delta surface in interdunal areas and near fluvial channels. Wauseon soils are sandy loam textured in the upper solum, overlying clay loam; the contact between them is a lithologic discontinuity.

It is likely that Wauseon soils within the study area are somewhat similar to the sandy-over-loamy textured two-storied soils that will be discussed in the following section. Additionally, a number of the other minor loamy soils have a lithologic discontinuity, including Corunna, Metamora, Pella, and Ziegenfuss soils, but these lithologic discontinuities are either loamy lacustrine sediments over till, or the reverse (Table 3.5).

| Series | Taxonomic Subgroup | Drainage Class | Parent Material (from OSD) | Landform | DI |
|---|---|---|---|---|---|
| Parkhill | Fine-loamy, mixed, semiactive, nonacid, mesic Mollic Epiaquepts | Poorly drained and very poorly drained | Loamy ablation till overlying loamy dense till | Wave-worked till plains, till plains, and moraines | 82 |
| Londo | Fine-loamy, mixed, semiactive, mesic Aeric Glossudalfs | Somewhat poorly drained | Loamy glacial till | Moraines and till plains | 69 |
| Bowers | Fine, mixed, semiactive, frigid Aquic Glossudalfs | Somewhat poorly drained | Stratified loamy lacustrine deposits | Lake plains | |
| Corunna | Coarse-loamy, mixed, semiactive, mesic Typic Endoaquolls | Poorly drained | Loamy till and the underlying lacustrine deposits | Lake plains and till plains | 82 |
| Ingersoll | Fine-silty, mixed, semiactive, mesic AquicHapludalfs | Somewhat poorly drained | Stratified silty deposits | Lake plains, outwash plains, and deltas | 69 |
| Lenawee | Fine, mixed, semiactive, nonacid, mesic Mollic Epiaquepts | Poorly drained and very poorly drained | Lacustrine deposits | Lake plains and in depressional areas on moraines, outwash plains, and glacial drainageways | 80 |
| Metamora | Fine-loamy, mixed, semiactive, mesic Udollic Epiaqualfs | Somewhat poorly drained | Loamy glaciofluvial or lacustrine deposits and the underlying loamy till | Lake plains, near-shore zones (relict), till plains and low moraines | 68 |
| Pella | Fine-silty, mixed, superactive, mesic Typic Endoaquolls | Poorly drained | Loamy sediments and the underlying stratified loamy glacial sediments | Outwash plains and till plains | 82 |

Table 3.4 – Characteristics of loamy soils in the study area (USDA-NRCS).

Table 3.4 (cont'd)

| Series | Taxonomic Subgroup | Drainage Class | Parent Material (from OSD) | Landform | DI |
|---|---|---|---|---|---|
| Wauseon | Coarse-loamy over clayey, mixed over illitic, superactive, mesic Typic Epiaquolls | Poorly drained and very poorly drained | Fine sandy loam, sandy clay loam, clay loam | Clay loam | 80 |
| Ziegenfuss | Fine, mixed, semiactive, nonacid, mesic Mollic Epiaquepts | Poorly drained | Loamy and clayey till | Ground moraines and depressional areas of steeper end moraines | 83 |

| Series | Upper Solum Texture | Lower Solum Texture | Parent Material (from OSD) | Depth to LD |
|---|---|---|---|---|
| Corunna | Sandy loam | Silty clay loam | Loamy till and underlying lacustrine deposits | 91cm |
| Metamora | Sandy loam | Clay loam, loam | Loamy glaciofluvial or lacustrine deposits and the underlying loamy till | 71cm |
| Pella | Clay loam, silty clay loam | Stratified loam, sandy loam, silt loam, and clay loam | Loamy or silty sediments and the underlying stratified loamy glacial sediments | 79cm |
| Wauseon | Fine sandy loam, sandy clay loam | Clay loam | Loamy and sandy glaciolacustrine sediments and the underlying till | 81cm |
| Ziegenfuss | Clay, clay loam, silty clay | Loam | Clayey ablation till overlying dense basal till | 178cm |

Table 3.5 – Characteristics of loamy soils with a lithologic discontinuity in the study area. (USDA-NRCS). See also Table 3.4.

### 3.1.4 Two-Storied Soils

Soils with a lithologic discontinuity at a shallow depth, often called two-storied soils, are formed in sediment with different textural properties in the upper and lower sola. In most cases, these kinds of soils indicate a change in depositional environment (Foss and Rust, 1968; Raad and Protz, 1971; Schaetzl, 1998; Schaetzl and Anderson, 2005). In this section I will discuss the importance to this research of two-storied soils, and will report and discuss the coarse-over-fine two-storied soils within the study area.

The sandy-over-loamy, two-storied soils within the study area generally occur in areas near the distal margins of the delta, and thereby represent two distinct depositional environments: thin sandy delta sediments over waterlain till or lacustrine sediments (Fig. 3.10). These soils are important to identify and discuss because they mark the outer margins of the delta, where the sandy sediment is present, but thin. In places where two-storied soils appear far from the delta, they are generally coincident with former shorelines, and are therefore also useful in identifying shoreline locations (Table 3.6, Fig. 3.11).

The two-storied soils in the study area include the Belleville, Menominee, Metea, Poseyville, Selfridge, and Wixom series. They are formed in either sandy material or sandy glaciofluvial deposits over till or lacustrine sediments. The thickness of sandy material over the loamy substrate ranges from 51-102cm for all series except Poseyville, which has <61cm of sand. Selfridge and Metea are found mainly in the western extent of the study area, in near-shore zones, along drainageways and outwash surfaces in the morainic uplands. Wixom and Poseyville are mostly in the eastern portion of the study area, in western Midland County, but

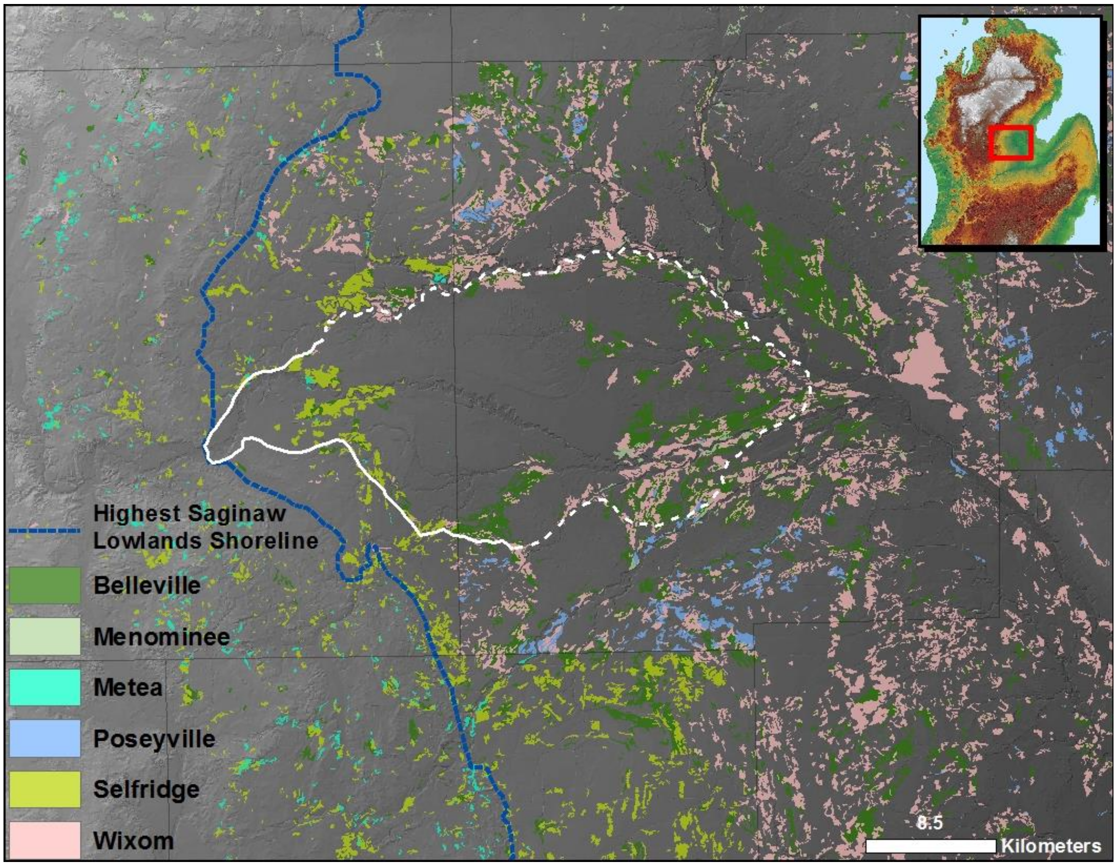



Figure 3.10 – The distribution of two-storied soils (Belleville, Menominee, Metea, Poseyville, Selfridge, and Wixom) in the study area (data from USDA-NRCS).

small bodies of Wixom soils also appear in the morainic upland to the west. Wixom and Poseyville soils are found mostly on the lake plain but are also in near-shore zones, such as near remnant Arkona and Warren beaches in northwestern Midland County. Belleville and Menominee soils form on the lake plain and are located mostly around the periphery of the delta. Menominee has an extremely sparse distribution and is found mostly near drainageways, suggesting that there may be a fluvial contribution to the Menominee's sandy upper solum. For much of the distal part of the delta Belleville and Wixom soils are good indicators of the location of thin deltaic deposits. In the western part of the delta, in areas of higher landform position, the thin deltaic sediments are mapped as Selfridge soils.

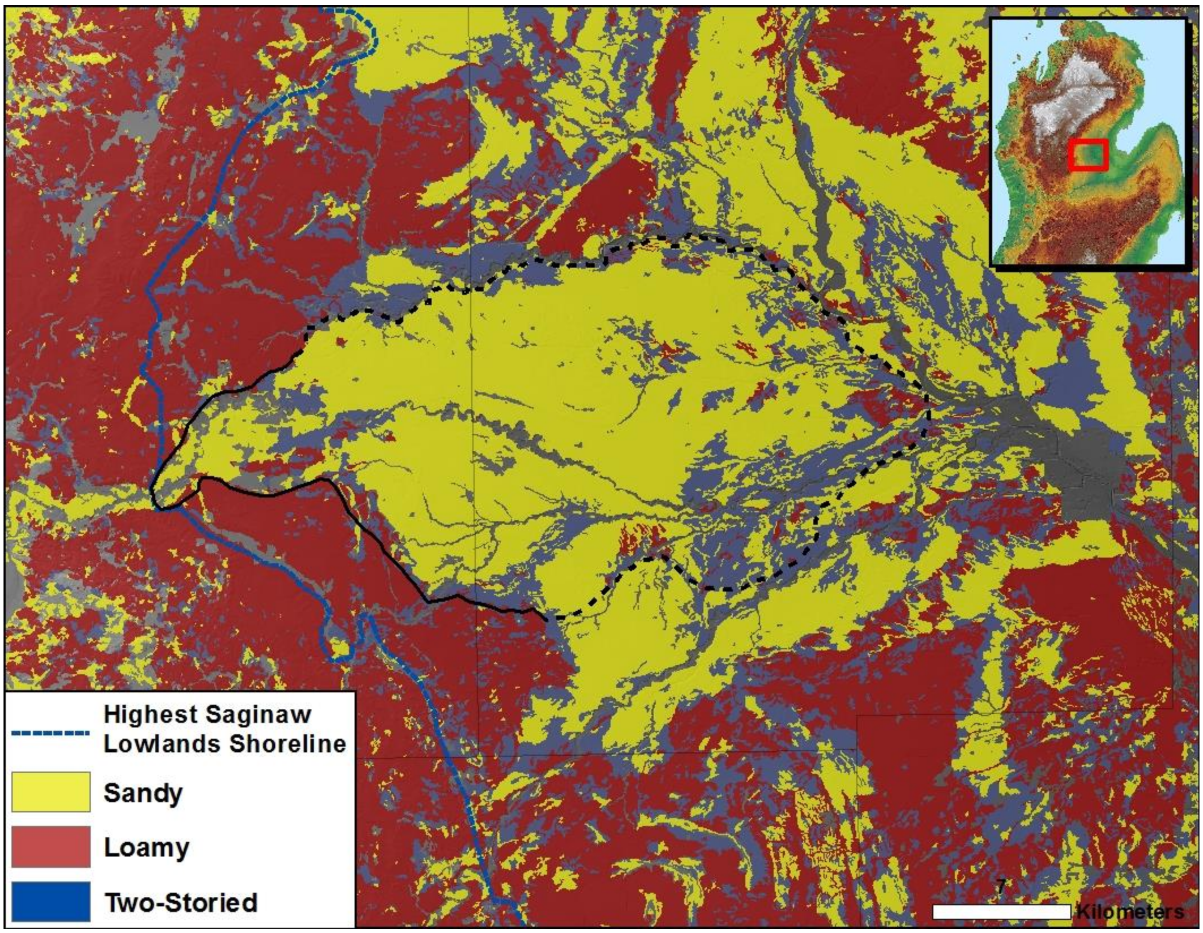


Figure 3.11 – The distribution of loamy, sandy, and two-storied soils within the study area, and the outline of the Chippewa Delta and Saginaw (data from USDA-NRCS).

| Series | Taxonomic Subgroup | Drainage Class | Upper Solum Texture | Lower Solum Texture | DI | Depth to LD |
|---|---|---|---|---|---|---|
| Belleville | Sandy over loamy, mixed, active, mesic Typic Endoaquolls | Poorly drained and very poorly drained | Loamy fine sand, fine sand | Heavy clay loam, silty clay loam | 82 | 76cm |
| Menominee | Sandy over loamy, mixed, active, frigid Alfic Haplorthods | Well drained | Sand | Clay loam, loam | 53 | 58cm |
| Metea | Loamy, mixed, active, mesic Arenic Hapludalfs | Well drained | Loamy fine sand, fine sand, sandy loam | Clay loam | 39 | 81cm |
| Poseyville | Coarse-loamy, mixed, active, mesic Aquollic Hapludalfs | Somewhat poorly drained | Loamy sand, sand, sandy loam | Loam | 63 | 58cm |
| Selfridge | Loamy, mixed, active, mesic Aquic Arenic Hapludalfs | Somewhat poorly drained | Sand, sandy loam | Clay loam | 65 | 64cm |
| Wixom | Sandy over loamy, mixed, semiactive, mesic Alfic Epiaquods | Somewhat poorly drained | Loamy sand, fine sand | Sandy clay loam, silty clay loam | 67 | 74cm |

Table 3.6 – Characteristics of two-storied (sandy over loamy textured) soils in the study area (USDA-NRCS).

### 3.2 Vegetation and Climate

In the early 19th century, the United States General Land Office (GLO) surveyed the territory that would become Michigan (Thomas, 2009). Surveyors took detailed field notes during the surveying process, including descriptions of vegetation communities at section corners and along the township and section lines (Thomas, 2009). The survey notes and associated plat maps were used by plant ecologists at the Michigan Natural Features Inventory (MNFI) to reconstruct pre-settlement vegetation (Comer et al., 1995). The resulting pre-settlement vegetation data were digitized for use in a GIS.

Pre-settlement vegetation data from the GLO notes show that nearly all of the study area was forested by northern hardwoods and swamp forest during the period of time in the 1800s when the land was surveyed. The dominant forest types in the study area were beech-sugar maple and hemlock-white pine forest. Hemlock-white pine forests were generally found on the sandy delta surface, while the loamy morainic uplands and lake plain were forested with beech-sugar maple or beech-sugar maple-hemlock forest. Toward the eastern margin of the delta, there was an increase in aspen-birch forest, and hardwood and conifer swamps. In the southwestern part of the study area, in the morainic uplands, the pre-settlement vegetation was dominated by white pine-red pine-white oak mixed hardwood forest (Fig 3.12). Presently, the delta surface remains largely forested, with deciduous, evergreen, and mixed forests as well as wooded swamps, shrublands, and non-forested wetlands. Indeed, most areas that are mapped as having deep sandy soils, the delta and upland outwash surfaces, are generally not used for farmland. Agriculture across the study area is primarily found in areas of loamy soils: the till-cored morainic uplands and lake plain (Fig. 3.13). The majority of agricultural land is

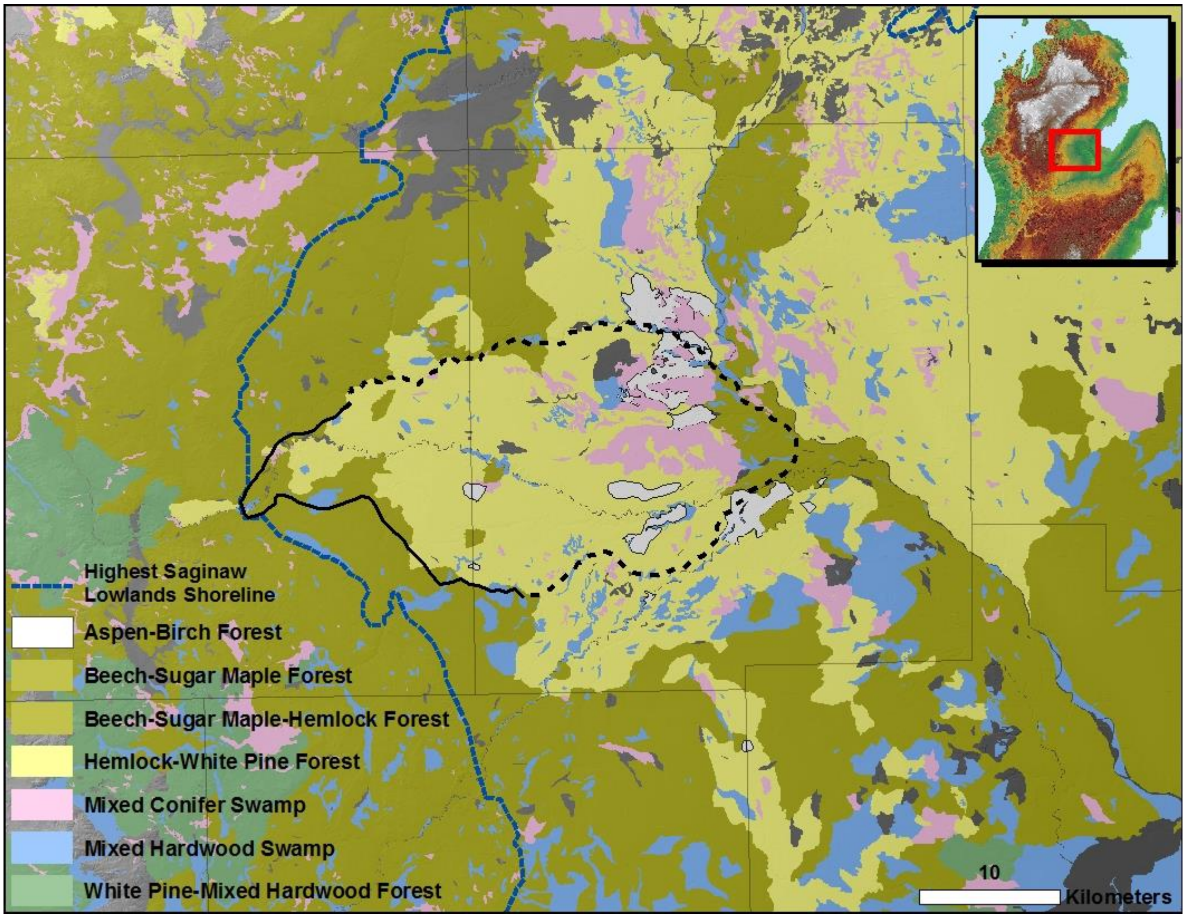


Figure 3.12 – Circa 1800 pre-settlement vegetation of the study area (data from General Land Office).

dedicated to row crops, typically corn and soybeans, but other crops include alfalfa, beans, oats, potatoes, sugar beets, winter wheat, and miscellaneous fruits and vegetables (Fig. 3.14). Central Lower Michigan has a cool, humid continental climate.

The mean annual temperature for Isabella and Midland Counties is 47.3 °F (8.5 °C); the summer average is 69.2 °F (20.7 °C) during June, July, and August, while the winter average is 24.2 °F (-4.3 °C) during December, January, and February. Mt Pleasant has a total average precipitation of 30.3 inches (76.8cm) and total average snowfall of 36.1 inches (91.7cm) per year, while Isabella County experiences 29.3 inches (74.3cm) of precipitation and 38.1 inches (96.8cm) of snowfall per year (Fig. 3.14; McLeese and Tardy, 1985; Hutchinson, 1979).

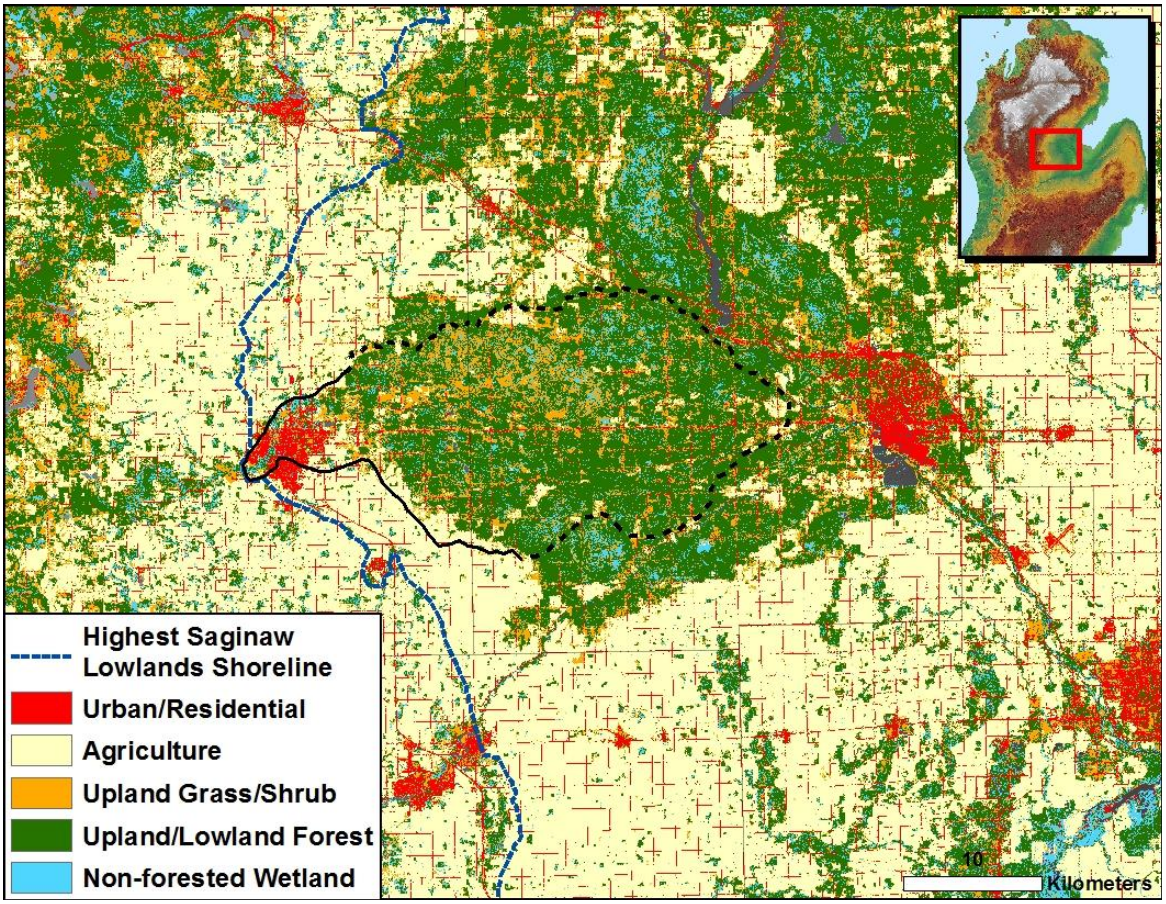


Figure 3.13 – Land cover circa 2000 in the study area, from the Michigan Department of Natural Resources (DNR).

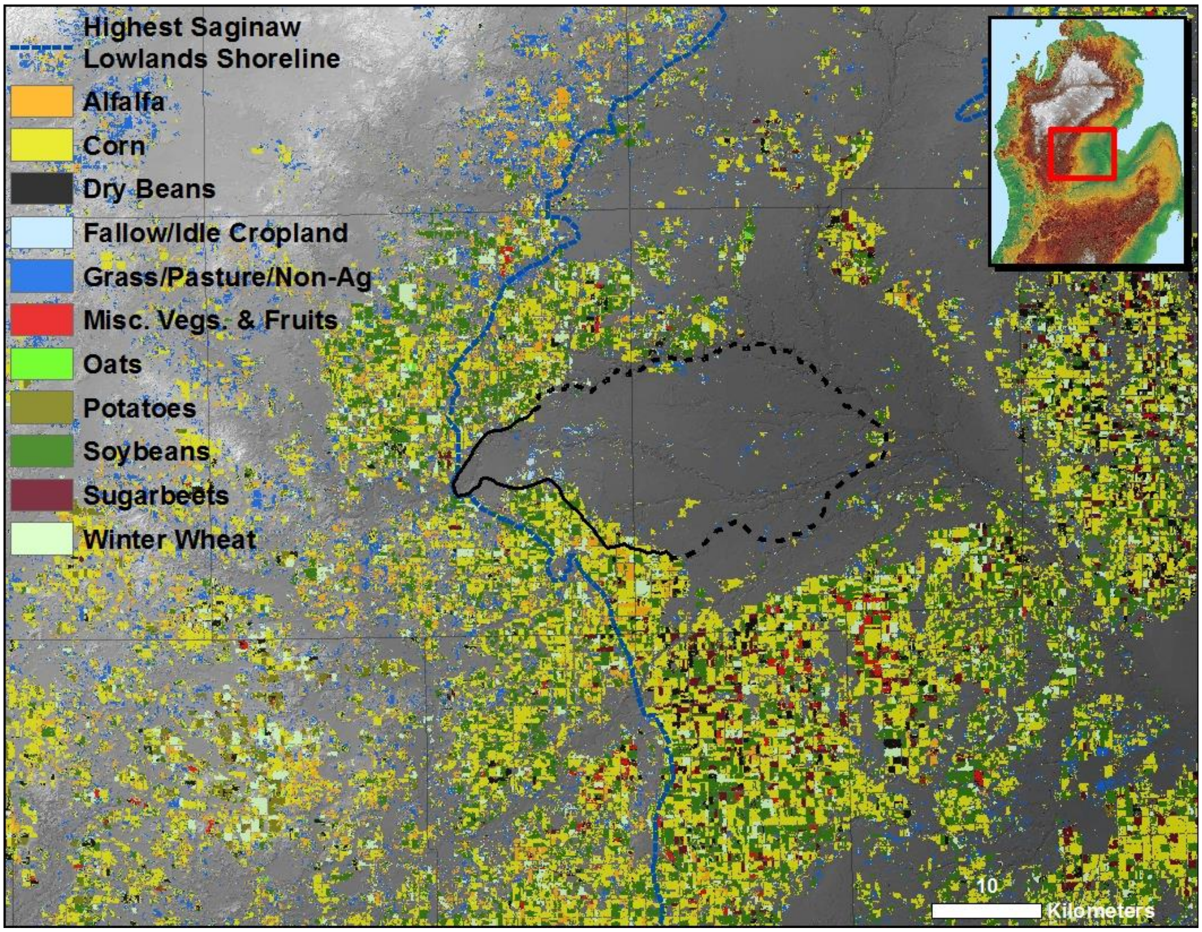
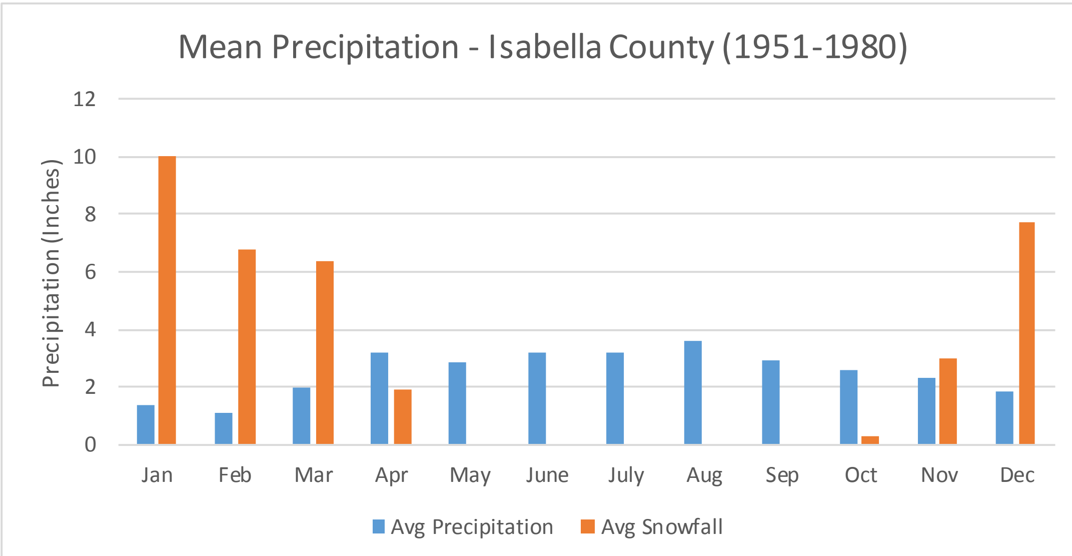



Figure 3.14 – Agricultural land use in the study area, from 2007 USDA National Agriculture Statistics Service Cropland Data Layer, the gray hillshade can be seen in places that are not used for agriculture (NASS-CDL).



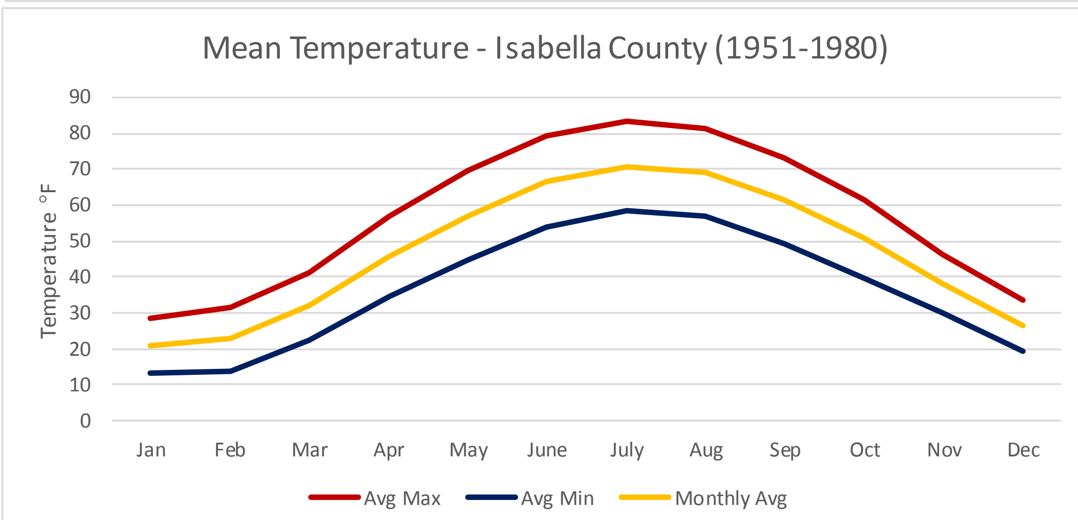




Figure 3.15 – Temperature, precipitation, and snowfall data from Mt. Pleasant from 1951-1980 (data from USDA-NRCS).

## 4. Methods

In this research I seek to characterize, map, and draw inferences about the formation of a Wisconsinin-age relict delta (the Chippewa Delta) and surrounding landscapes, by analyzing the spatio-textural characteristics of sandy C horizon materials, which I interpret as being the original sediment of the delta. To that end, I collected samples across and outside (but close to) the delta at evenly spaced locations to facilitate a spatial analysis of parent material textures across the delta surface. Additionally, within the Chippewa River watershed, till samples were collected for comparison, and to rule out tills in this area as the sole source of the deltaic sand.

### 4.1 Field Methods

Before entering the field, I identified potential sample sites using ArcGIS, by viewing SSURGO soil data from the United States Department of Agriculture Natural Resources Conservation Service (USDA-NRCS), overlain onto a 10-meter digital elevation model (DEM) hillshade from the United States Geological Survey (USGS). The SSURGO dataset was upgraded to include a number of additional variables, e.g., upper (A horizon) and subsoil (lowest described horizon) textures, and parent material type, as well as other data obtained from the NRCS official soil series descriptions (OSDs). The drainage index (DI) of Schaetzl et al. (2009) was also included in this dataset. After these soil data were variously overlain onto a hillshaded DEM, spatial patterns became apparent. For example, I used these data to display those raster cells that had sandy-textures within the C horizon. This display revealed an obvious fan-shaped distribution of sandy textured soils in an area where loamy textured soils would otherwise be expected, i.e., the Saginaw Lake Plain. To identify whether these sandy soils are part of a relict delta, I employed a spatial approach to investigate the sedimentology, topography and extent of this landform.

Soil parent material usually represents the sedimentary characteristics, e.g., degree of sorting, particle size modality, of the source sediment from which a soil has formed. Therefore, spatio-textural analysis of soil parent materials may reveal spatial trends in sediment character, thereby indicating patterns in paleo-depositional processes. For this investigation, the subsoil (lowest described horizon) texture variable in the SSURGO soil dataset served as a proxy for sedimentologic data, on which the field sampling strategy was based.

Sites were deemed suitable for field sampling if they were sandy in both the upper solum (A horizon texture) and parent material (texture of lowest described horizon). Potential sample sites were located approximately 1.5 km apart and regularly spaced, using a GIS; most were near roads and trails to ensure ready access. Sites near areas of human disturbance, erosion, or secondary deposition, e.g., a floodplain or a sand dune, were avoided. At each sampling location, samples of about 200-250 g were collected from the C horizon, in presumably unaltered parent material, typically at a depth between 1 and 1.5 meters. The depth at which each sample was collected was recorded in a field notebook. A standard three-inch bucket auger was used for sample collection. The location of each sample site was recorded in ArcMap 10.0, in a point shapefile, on a Panasonic Toughbook CF-19 laptop with built-in GPS.

A total of 174 samples were collected for textural analysis: 154 samples of sandy textured C horizon material, and 20 samples of loamy-textured till from the adjacent highland in central Isabella County (Fig 4.1). Care was taken to avoid taking samples from horizons with

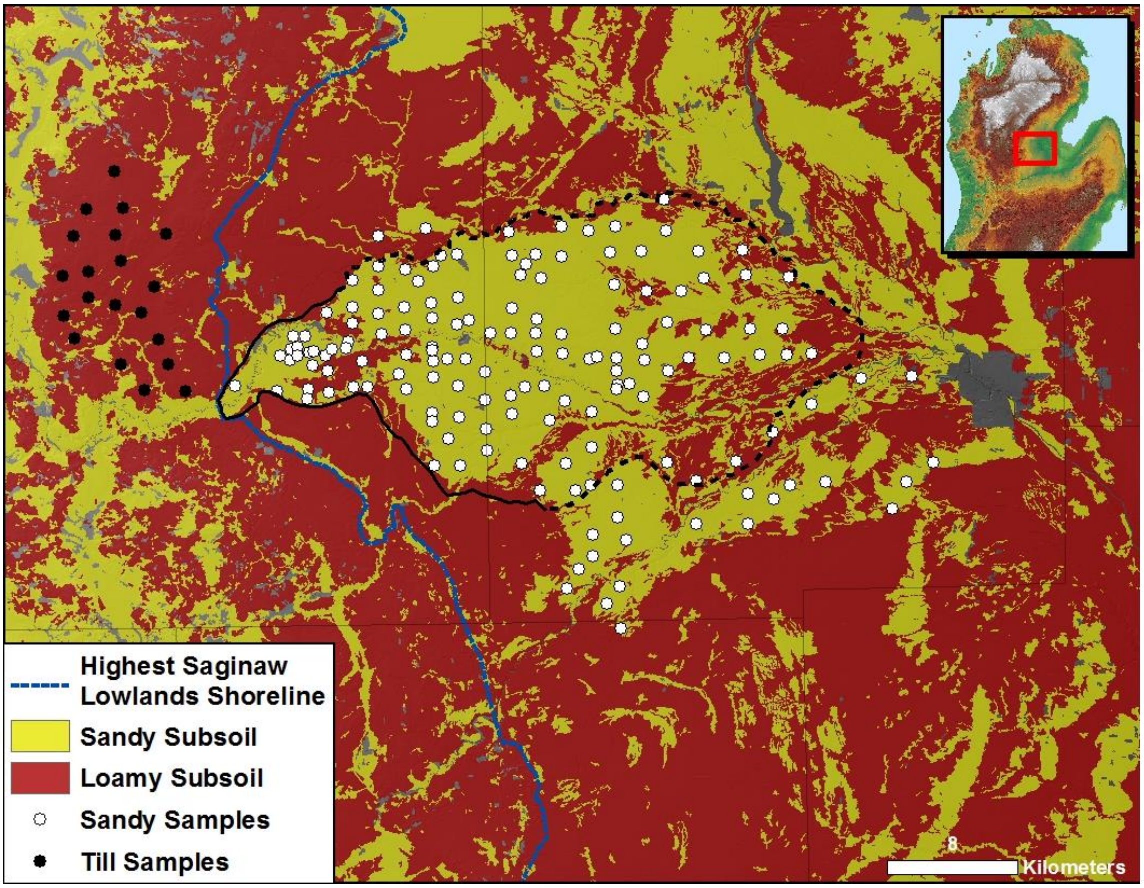


Figure 4.1 – The distribution of samples collected in the field in both sandy (white points) and loamy till (black points) C horizon. Till samples were collected from within the Chippewa River watershed.

abundant redox features, but in some instances samples were necessarily collected near or below the water table.

Till samples were also taken from the C horizon, after being exposed to dilute HCl to ensure that they were calcareous and had not been altered by soil forming processes. I collected till samples from within the Chippewa River watershed, for the purpose of comparing the textural characteristics of loamy uplands soils with the sandy soils on the delta surface.

**4.2 Lab Methods**

Samples were oven-dried overnight at ~30°C and lightly ground with a mortar and pestle. Gravels (> 2mm) were removed from the sample by sieving, and weighed. The remaining sample was weighed, and the percent (by mass) of gravels was calculated and recorded in a Microsoft Excel spreadsheet. Each sample was then passed through a sample splitter three times to homogenize it before being re-bagged. Before particle size analysis, each sample was next passed through a 1mm sieve. The very coarse sand fraction (between 1mm and 2mm) was then weighed. The mass of the remaining sample was weighed, and the percent by mass of the very coarse sand fraction was manually calculated, because our laser diffractometer does not reliably measure sediment with grain sizes >1000 µm. These values were also recorded in the Excel spreadsheet.

Samples were prepared for laser diffraction using ~5g of the < 1 mm sized sediment for the sandy textured samples, mixed with 3 ml of a sodium hexametaphosphate $(NaPO_3)_{13}\ Na_2O$ solution and approximately 20 ml of distilled water in a 25 ml vial. Loamy textured (till) samples required only ~1g of sample but were otherwise prepared similarly. After preparation, the samples were oscillated for a minimum of two hours and then measured using a Mastersizer 2000E laser particle size analyzer (Malvern Instruments Ltd, Worcestershire, UK). The data output from the Mastersizer were uploaded into an Excel spreadsheet and adjusted to include the manually calculated data for the very coarse sand fraction. The particle size data were then joined in a GIS for each sample location.

**4.3 GIS Methods**

**4.3.1 Sand Thickness**

The thickness of sandy delta sediment can be estimated using proxy data e.g. groundwater well data and soil data. These data, imported to a GIS, are useful for interpolating a trend surface which represents the changes in thickness of the delta sediments through space. Groundwater (water well) data obtained from the Michigan Department of Environmental Quality Statewide Ground Water Database (DEQ-SGWD) for both Isabella and Midland counties were used to analyze the thickness of sandy sediment across the study area. The well-log data contain information on lithology, including thickness of lithologic strata, measured in feet. The extent of the sandy sediment examined with the well-log data had been initially delineated in the GIS, using the SSURGO soil data, by displaying cells that contained sandy textured subsoil. However, the SSURGO data are only useful for sand thicknesses up to five feet. Well-log data provided more detailed information for areas where the sand was thicker, facilitating analysis of changes in thickness across the entire deltaic landscape. Wells were chosen for analysis based on their proximity to soils with sandy parent materials. Within the sandy textured area, data from 3276 wells were analyzed and thickness of sandy textured sediment recorded for each well in the attribute table of a shapefile (Fig 4.2). Generating a point shapefile with a column dedicated to sand thickness enabled me to use interpolation methods to analyze spatial trends in the point data.

Measurements of the thickness of deltaic deposits, characterized by sandy textured sediments, were recorded for wells that featured sand, gravel, or sand and gravel at the surface. These measurements record the thickness from the surface to the contact with an underlying fine textured stratum (Fig 4.3). Wells that had “topsoil” or “muck” recorded at the surface but which were directly underlain by sand, gravel, or sand and gravel were also included. The “topsoil” description has no textural implication, so I assumed that it shared the textural characteristics of the underlying strata. The “muck” description was interpreted as organic material that had formed after the sediment that underlies it was emplaced. Thus, it is

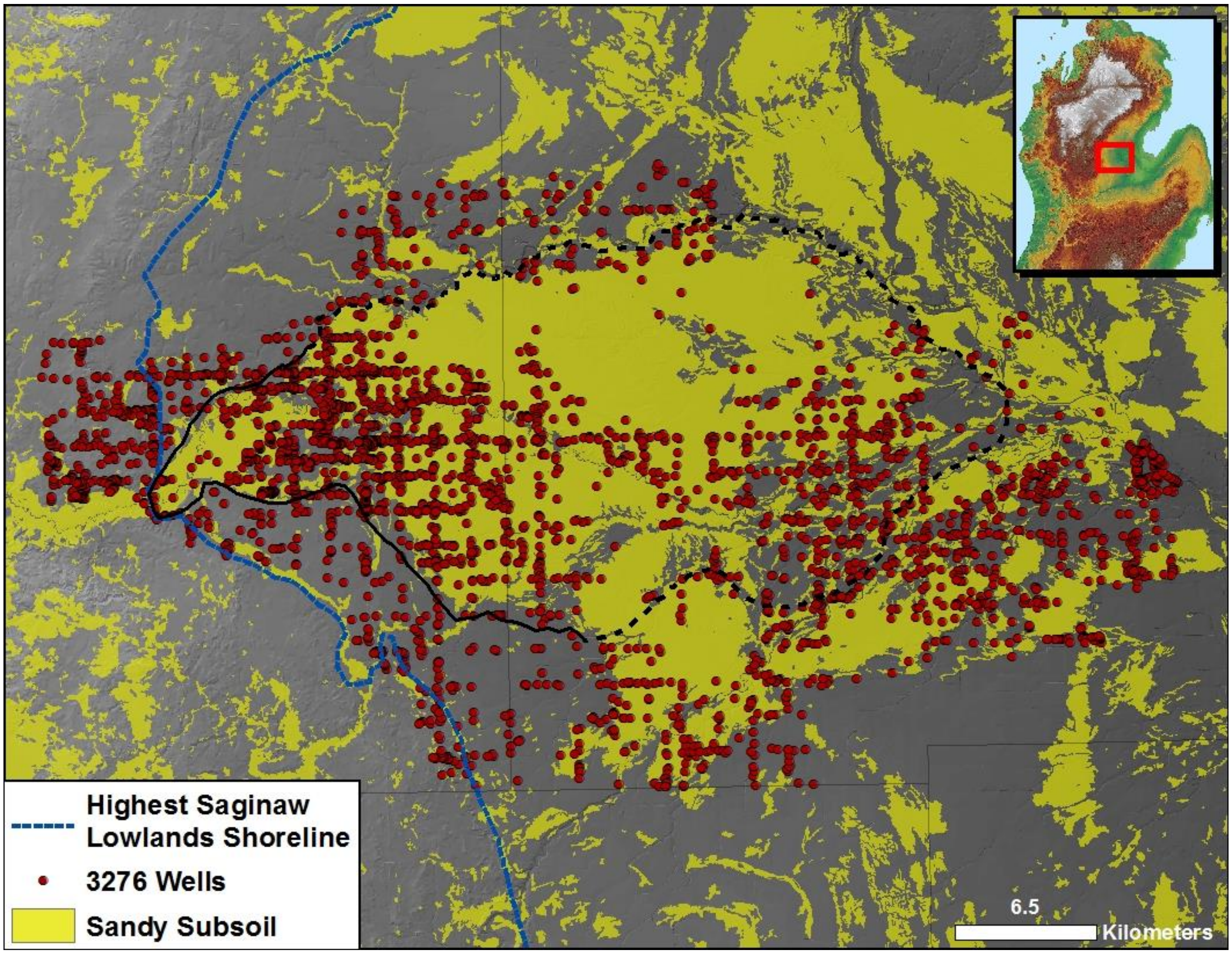


Figure 4.2 – The distribution of (1) sandy textured subsoil materials, from the NRCS, and (2) well-log locations. Both types of data were used to analyze the thickness of sandy sediment in the study area.

not indicative of the geo-sedimentary history of the underlying mineral sediment. The surface "muck" layer was not included in the thickness measurement, rather only the sandy textured sediment underlying it. Wells which featured clay or loam texture at the surface were given a thickness of zero. Mixed strata featuring "clay" in the description, e.g., "sand and clay" or "sand, clay and gravel", also were given a thickness value of zero. The presence of clay was interpreted as either indicative of a low energy depositional environment (lacustrine), or till, and hence was not a focus.

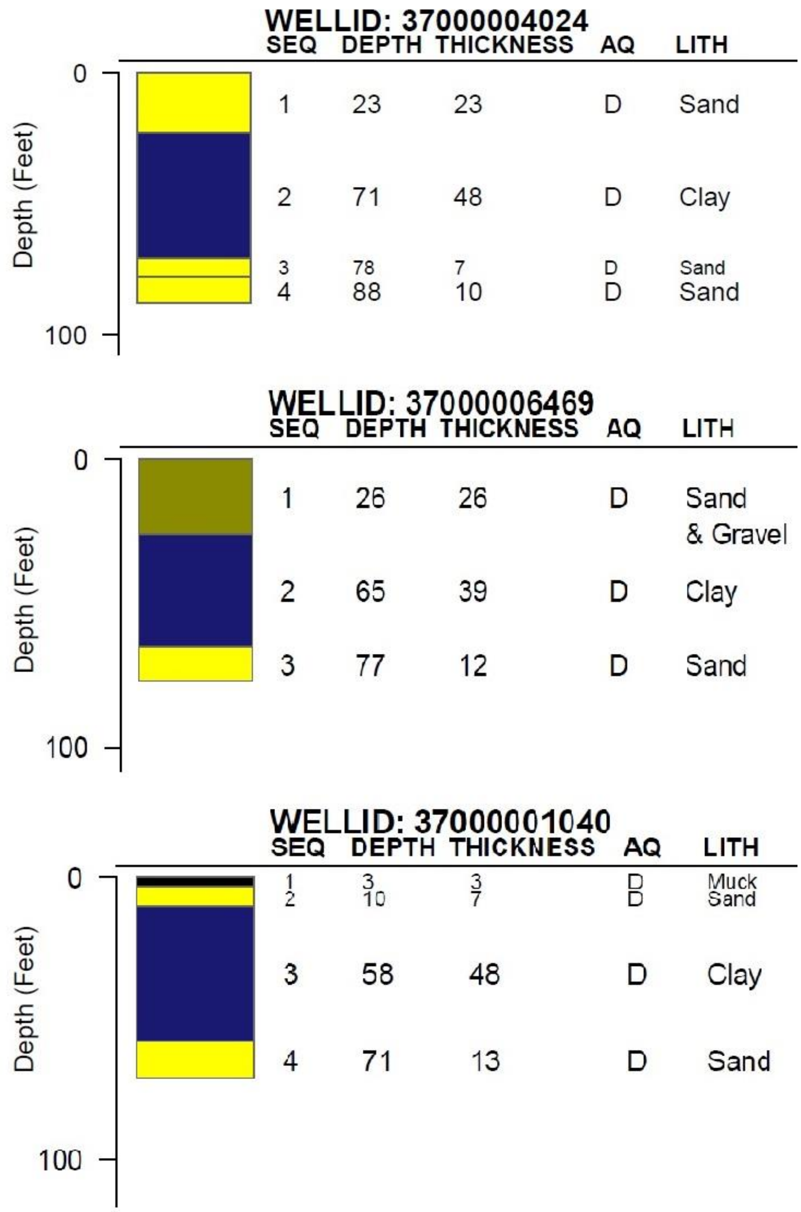



Figure 4.3 – Examples of well log data used to calculate sandy sediment thickness. From top to bottom, sand thickness values are 23', 26' and 7', (data from DEQ-SGWD).

For each well with sand recorded at the surface, the terminal depth (sand body thickness) was determined based on the presence of a thick (> 3 feet) intervening stratigraphic unit of composition other than sand, e.g., clay, clay loam, loam. Thinner layers were considered to represent temporary depositional changes, but not long-term sedimentary regime changes.

Layers >3' thick were considered to represent more long-term depositional trends, and those layers were interpreted as having been deposited by a different sedimentary process.

NRCS soil data were used to further refine the estimate of delta thickness. Within the study area, there is a fairly significant distribution of two-storied soils which have a sandy upper solum over a loamy or clayey substrate. The OSD for each of these two-storied soils (Belleville, Menominee, Metea, Poseyville, Selfridge, and Wixom) details the typical thickness of the sandy upper solum. A total of 185 additional points were manually placed, roughly evenly spaced, on cells with mapped two-storied soils, and a sand thickness value was recorded in the shapefile from the measurements provided in the OSD (Fig. 4.4). Supplementing the water well data with these soil data will yield a more accurate estimate for sand thickness across the delta.

**4.3.2 Particle Size Data**

The laser diffractometer particle size data were joined to the sample location layer in ArcGIS, resulting in a point shapefile with provenienced particle size data for each sample. Preliminary analysis of the particle size data show that the sandy samples are often unimodal, with all samples show strong modality in the sand fraction. An ordinary kriging routine was employed to determine spatial trends in both particle size fining and sand thickness thinning across the deltaic surface. Default settings were used to generate a semivarogram model that did not require any special customization to the data, so they were used for interpolation. The resulting interpolations were analyzed and interpreted, and will be referenced further in the discussion section.

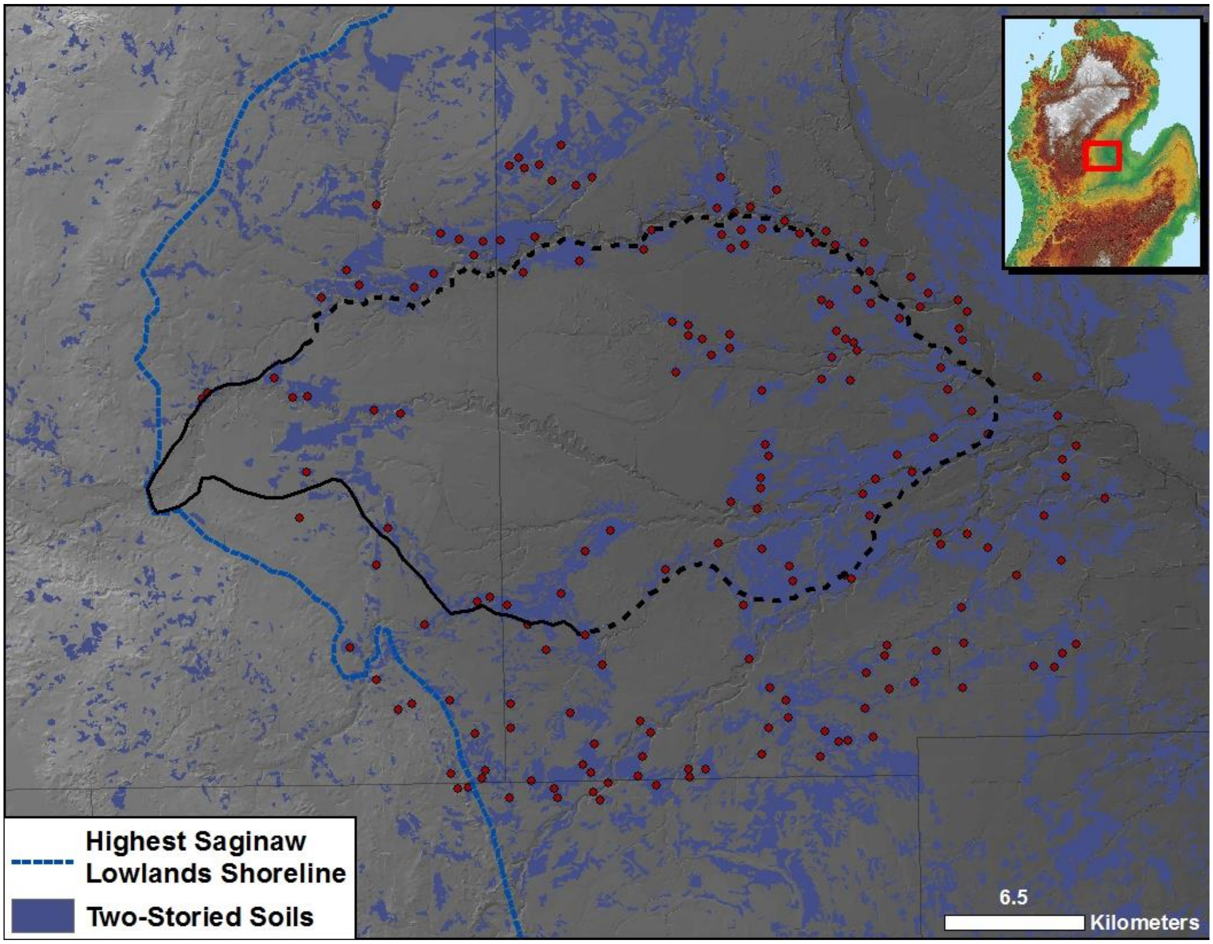


Figure 4.4 – The distribution of (1) two-storied soils (sandy upper solum over loamy substrate), from the NRCS, and (2) 185 points with sand thickness values recorded from the soil series OSD.

### 4.3.3 DEM and Shoreline Analysis

Records of former lake shorelines and their associated elevations were collected and analyzed using a combination of published maps and literature review, primarily Leverett and Taylor's (1915) USGS monograph, together with analysis and interpretation of the DEM and DRGs.

Leverett and Taylor (1915) recorded detailed descriptions of shoreline features, including their elevations and their locations relative to towns and villages. I identified these features (all associated with former shorelines) in the GIS and recorded their locations and elevations as point features in a shapefile. Additionally, I used a digitized and georectified scan of the 1955 Martin map, “Map of the Surface Formations of the Southern Peninsula of Michigan” as a reference. This map is especially useful because Martin marked some of the shoreline features on this map with symbols indicating the lake stage associated with the feature, e.g. Lakes Arkona, Warren, and Saginaw; this information was omitted from the later Farrand and Bell (1982) map. Shorelines from the Martin (1955) map had to be converted to a polyline shapefile from the scanned and georectified raster file, which I did manually in ArcMap. After I converted the shorelines from Martin (1955) to a polyline shapefile, I used the “Smooth Line” tool in ArcMap with a 100m smoothing tolerance in order to improve the aesthetic quality of the layer. Smoothing did little to affect the accuracy of the shoreline feature conversion. I also imported the shapefile of Quaternary Features from the Farrand and Bell (1982) “Quaternary Geology of Michigan” map, digitized by the Michigan Natural Features Inventory, and available from the Michigan Geographic Data Library. Both of these maps contain information on shorelines within the study area. Finally, a shapefile of the highest Lake Saginaw shoreline associated with the Lusch et al. (2009) study, and an unpublished shapefile of a Lake Warren shoreline, were also included in my data set (Lusch, personal communication).

After I compiled the shoreline datasets from the sources mentioned above, I evaluated them for spatial accuracy by comparing the location of the mapped shorelines with topographic features and with other linear or curvilinear changes in elevation on the DEM, all of which are commonly associated with shorelines. Within the study area, shoreline features vary between strongly expressed beach ridges, to subtle wave-cut bluffs. Both of these types of topographic features show distinct linearity, which helps to identify them as coastal landforms. These topographic features sometimes also coincide with soil patterns, especially regarding texture, parent material, and wetness, as indicated by the DI. Locations of relict shorelines often correspond with narrow, linear soil map units with sandy surface and/or subsurface horizons. These narrow bands of sandy soil contrast with the typically loamy subsoils of the lake plain. These patterns also appear in the drainage index (DI) data from Schaetzl et al. (2009), where areas of sandy shoreline appear drier than the surrounding loamy lake plain. By analyzing and comparing the shoreline maps with the DEM and soil data, I was able to create a new shoreline shapefile which refines data from the previous mapping efforts for the study area. The results of the shoreline mapping will be reported and discussed further in the results section.

## 5. Results and Discussion

In this section, I will first provide some background information about processes that may help put into context my interpretations of the Chippewa Delta. Second, I will present and briefly discuss the shorelines of Glacial Lake Saginaw that are pertinent to this study. The locations of these shorelines are important to my interpretations, and therefore must be discussed before other data are presented. Next, I will report data and interpretations from the GIS analysis of NRCS soil data, water well data, and field (soil) samples. Because this research relies heavily on data from the USDA-NRCS soil surveys, I will report interpretations of the distribution of soils on the delta surface as mapped by the NRCS before I report the water well or field soil sample data. Next I will report data on the delta thickness from GIS analysis of water well and two-storied soil data. I will also report and interpret the textural data from samples taken from soil parent material across the delta surface, and discuss their spatio-textural trends. Lastly, I will offer a synthesis of the stages of the Chippewa Delta's evolution through time and space, draw inferences about interactions between the Chippewa Delta and neighboring landforms, and interpret how the various stages relate to paleolake shorelines/elevations.

### 5.1 Background on Delta and Basin Processes

Delta characteristics and morphology, generally, are the product of two primary variables: (1) inputs of sediment from river mouths or eroded off of headlands, as influenced by (2) sediment reworking by waves, tides or currents in the basin (Wright and Coleman, 1972; Galloway, 1975; Boggs, 1982; Leeder, 2001). The energy and strength of each of these variables has a great influence on delta morphology. Fluvially dominated deltas, where sediment inputs outpace reworking efforts by waves, tides or currents, tend to be lobate shaped, often with a muddy distal fringe (Wright and Coleman, 1972; Galloway, 1975; Boggs, 1982). Wave-dominated deltas, by comparison, tend to have smoother, arcuate outlines, with a sandy composition, and with associated landforms like beach ridges, off-shore bars, and shoals (Galloway, 1975; Bhattacharya and Goisan, 2003). These competing energies within a deltaic system, however, are just some of a number of variables that influence delta morphology. In this section I will discuss a number of processes that I believe were at work during the formation of the Chippewa Delta.

One fundamental result of delta progradation, relied upon by geomorphologists for informing their interpretations of deltaic deposits, is fluvially driven sediment sorting during delta formation. In a "Gilbert-type" delta, for example, there are three distinct sedimentologic zones, topset, foreset, and bottomset, each of which is differentiated based on sedimentary characteristics, bed angle, and location (Galloway, 1975; Boggs, 2001). Topset beds are flat-lying, proximal to the point at which a stream discharges into a standing water body (discharge point), and usually consist of coarser sediment such as sand and gravel. Foreset beds are gently dipping beds, commonly of stratified sand, and are located in the basin, beyond the topset beds. Bottomset beds consist of fine grained sediments, deposited at the farthest fringe of the delta, in deeper water (Elliot, 1978; Boggs, 2001). Deltaic sediments are sorted in this way, from coarse grained deposits proximal to the discharge point to fine grained deposits far from the discharge point, due to loss of transportational energy; as the river discharges into the basin, the flow becomes unchannelized, spreads laterally, and slows (Galloway, 1975; Vader et al. 2012). The sediment that can no longer be carried in suspension or transported as bed load when the river becomes unchannelized is deposited, forming the delta (Galloway, 1975; Elliot, 1978; Boggs, 2001).

Because of these depositional patterns, evidence of landward transgressions, or basinward regressions, can be preserved in deltaic deposits, enabling researchers to reconstruct past changes in the basin. Geologists can interpret these patterns from lithified deltaic deposits in the rock record, either in outcrops, seismic soundings, or in drill cores (Bellotti et al., 1994; Brooks et al., 1994; Adams et al., 2001; Evans and Clark, 2011). Transgressions and regressions within the deltaic system are usually a product of base level fluctuations. Base level rise often results in “stacked” delta complexes; the discharge point moves upstream, and up-elevation, forcing the topset, foreset, and bottomset bed series to move landward. Hence, sediment gets deposited on top of the delta which had previously been prograding before base level rose, thus “stacking” the fining-outward series, one on top of another. Conversely, a drop in base level can result in river incision, erosion and redistribution of delta sediments, and/or forced regression, i.e., the delta complex’s outer boundary and the “coarse to fine” sequence moves further out, into the basin (Winsemann et al., 2011; Carrivick and Tweed, 2013; Nanson, et al., 2013; Santra et al., 2013). These two concepts - the “fining outward” of deltaic sediments, and the delta’s sedimentary response to base level fluctuation - are fundamental to the interpretation of the delta sedimentary and textural data.

**5.2 Paleolake Shorelines in the Study Area**

During Michigan’s recent glacial history, a number of lakes have been identified which had inundated parts of the state and left relict shorelines to mark their extent and elevation. Many of these paleolakes have been studied in some detail (Eschman and Karrow, 1985;

Farrand and Drexler, 1985; Larson and Schaetzl, 2001; Kincare and Larson, 2009; Drzyzga et al., 2012; Lewis et al., 2012), others barely at all, and others still have yet to be formally identified and studied. Of all the remnant paleolake shorelines in Michigan, the shorelines within the Saginaw Lowlands have been studied in less detail than most. Knowledge of paleolake levels, e.g., as indicated by the shorelines within the Saginaw Lowlands, can help inform the interpretations and timing of delta development within the paleolake.

My research on the shorelines within the study area included reviewing data from previous studies (Martin, 1955; Farrand and Bell, 1982), compiling maps that include shorelines, and creating a more complete and accurate shoreline dataset by using data/maps from previous studies, including  topographic and soil maps (Fig. 5.1). Previous efforts to map shorelines in this area did not have the benefit of digital data, such as DEMs, the Drainage Index of Schaetzl et al. (2009), detailed NRCS soil data, or a GIS environment within which to conduct the mapping effort. Despite this, the published maps I consulted were fairly accurate and required only moderate refinement. Among the most accurate was that of Martin (1955), and thus it served as a particularly helpful resource. As discussed in the methods section, the refined shoreline map generated in this study (Fig. 5.1) was created by identifying linear shoreline features/beach ridges on the DEM and linearly shaped, sand textured soil map units. I also referred to mapped shorelines from previous studies (Fig 5.2). Hopefully, in the future, this dataset can be further refined after more research is conducted in this region, or after LiDAR data become available.

Regardless of the existing data, there remains some ambiguity as to which shorelines are associated with the various lake stages in the Saginaw Basin, specifically Lakes Saginaw,

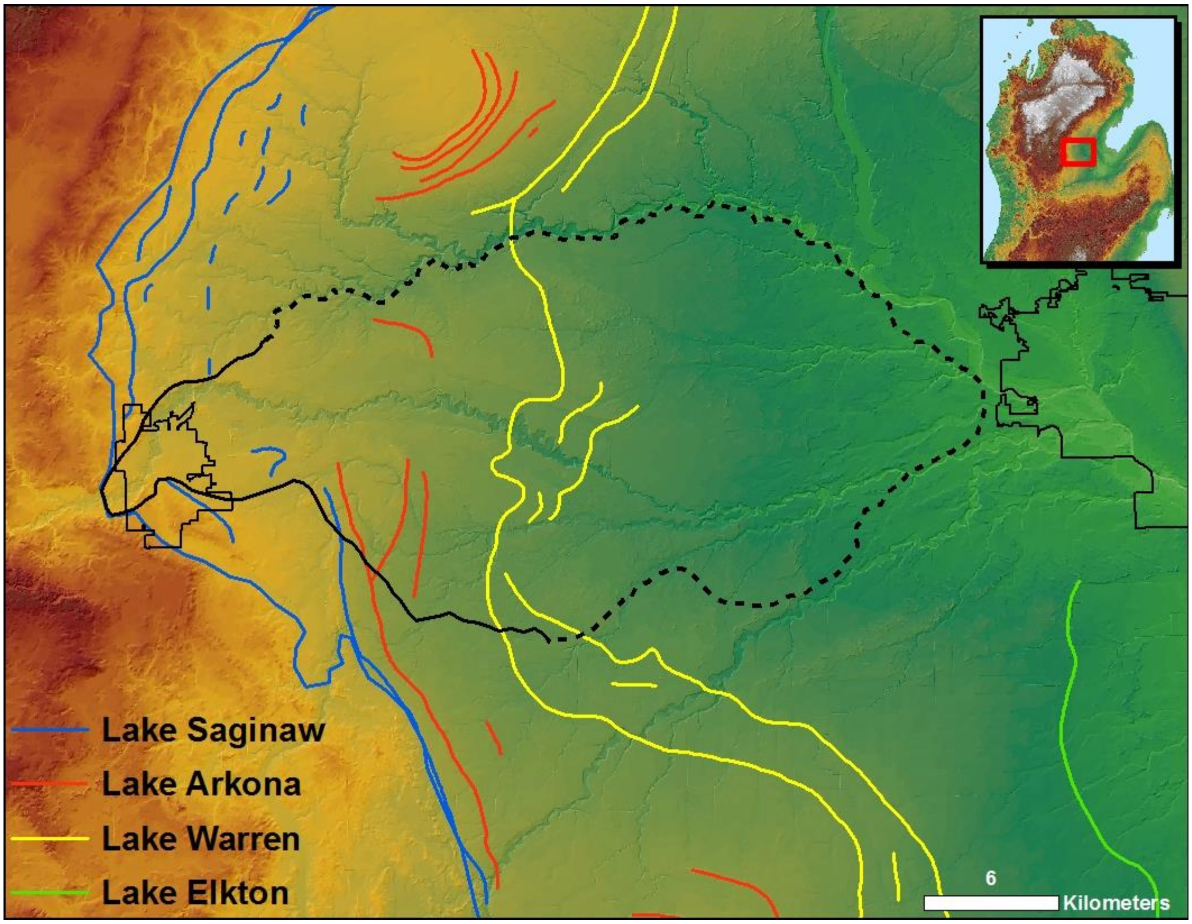


Figure 5.1 – Paleolake shoreline features in the study area, as modified from Martin (1955), Farrand and Bell (1982), and Lusch et al (2009). Also shown are the outline of the Chippewa Delta and the cities of Mt. Pleasant (west) and Midland (east), on a DEM hillshade.

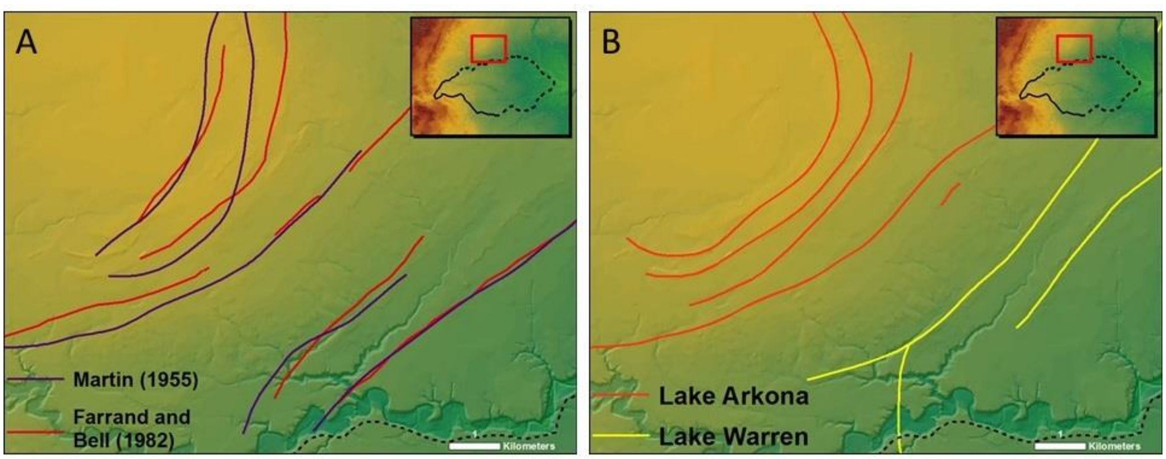


Figure 5.2 – An example of previously mapped shorelines in the study area (A) by Martin (1955) and Farrand and Bell (1982), and (B) the refined and expanded map generated in this study.

Arkona, and Warren. As mentioned by Leverett and Taylor (1915) and Bretz (1951), some of the lake stages within the Saginaw Basin probably had similar elevations, i.e., the lowest Early/Later Lake Saginaw with the highest Lake Arkona, as well as lowest Lake Arkona with the highest Glacial Lake Saginaw. The elevations of these lakes were controlled by the Maple-Grand River outlet channel, which had only small amounts of elevational fluctuation overtime, implying that lake stages at different times were elevationally similar. Because of this ambiguity, I chose to take a conservative approach in this mapping exercise, by refining and expanding upon what has been previously done so that it is more spatially accurate and complete, and using the lake stage names attributed to shorelines, following Martin (1955).

The shorelines in the study area are only weakly expressed, likely because they were either weakly formed or were not well preserved. The exceptions are the Lake Arkona shorelines in the northern part of the study area, which are fairly well expressed, though discontinuous. The oldest shorelines, those of Lake Saginaw, are cut into loamy textured till which may be fairly resistant to wave erosion; this characteristic may account for their weak expression (Fig. 5.3). Lake Arkona, which chronologically followed Lake Saginaw, had shorelines that are sandier and likely less resistant to wave erosion, which may have enhanced their expression; they appear on the DEM as distinctly linear, sandy beach ridges and small wave-cut bluffs (Figs. 5.2, 5.3). Lake Warren shorelines are also sandy, but are not as strong as the Lake Arkona shorelines.

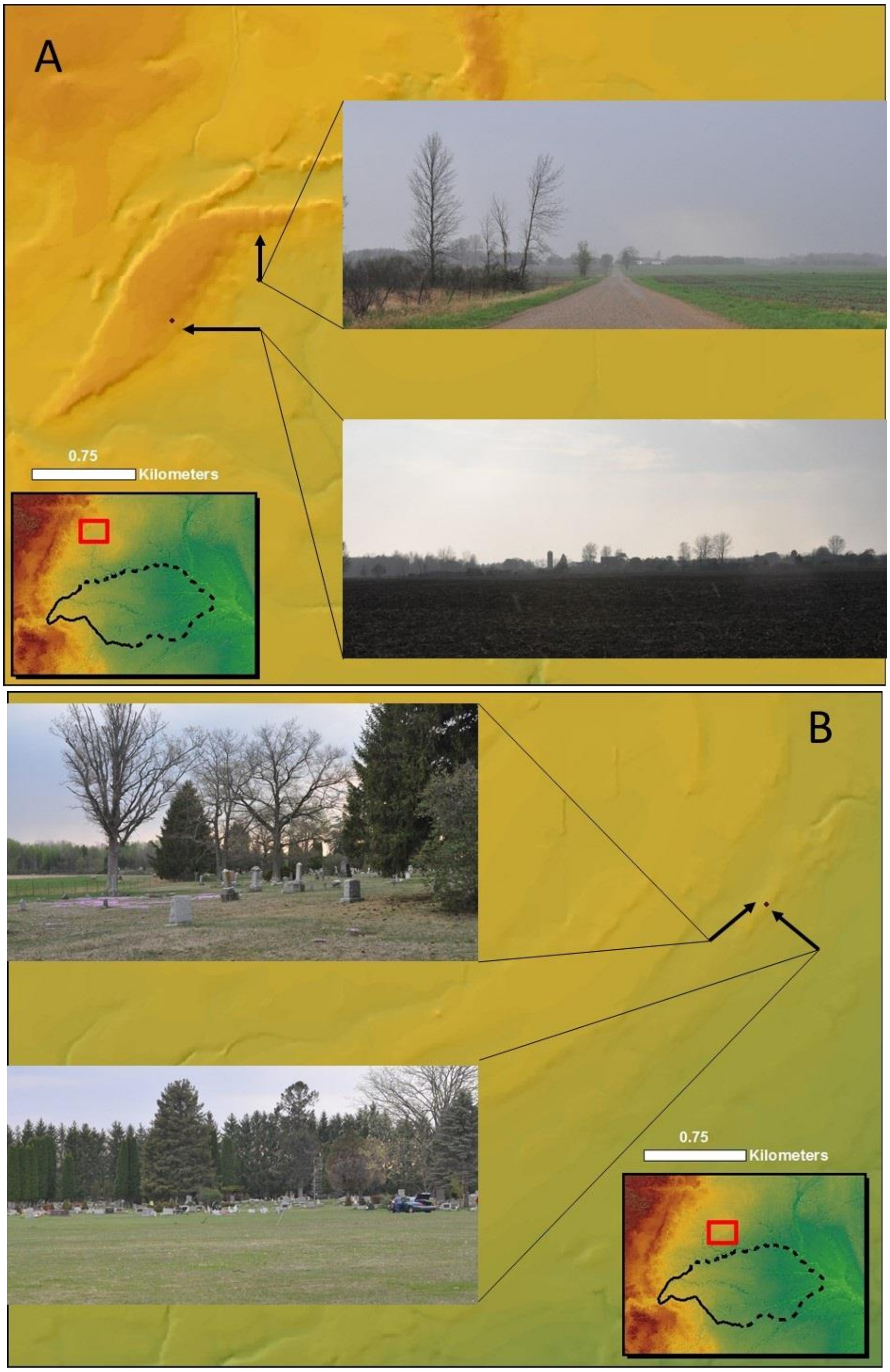



Figure 5.3 – Photos and locations of (A) Lake Saginaw and (B) Lake Arkona shoreline features. Photos show a subtle bluff for the Saginaw shoreline (A), this section of Lake Saginaw shoreline is discontinuous, and among the best expressed in the study area. The Lake Arkona beach ridge feature in (B) has gentle slopes on either side of the ridge, and is overall a subtle feature. It can only be seen on the ground because it is not entirely forested.

### 5.3 General Characteristics of the Chippewa Delta

The extent/boundaries of the delta were determined using a combination of NRCS soil data, stratigraphic logs from water well data, and textural data from field collected samples, all of which were examined in a GIS. As previously discussed, the boundary line presented here is dashed where an exact boundary cannot be determined, and thus the dashed line serves as an estimated boundary based on interpretation of the data in this study. The delta is approximately 18km wide and 38km long along its maximum axes. It is roughly lobate shaped (Fig. 5.4), and in many places has a diffuse, gradual boundary, rather than a sharp edge, or delta front. A broad but shallow erosional channel occurs on the surface of the delta, associated with the modern Chippewa River. As mentioned, soils on the delta are sandy, and contrast with the

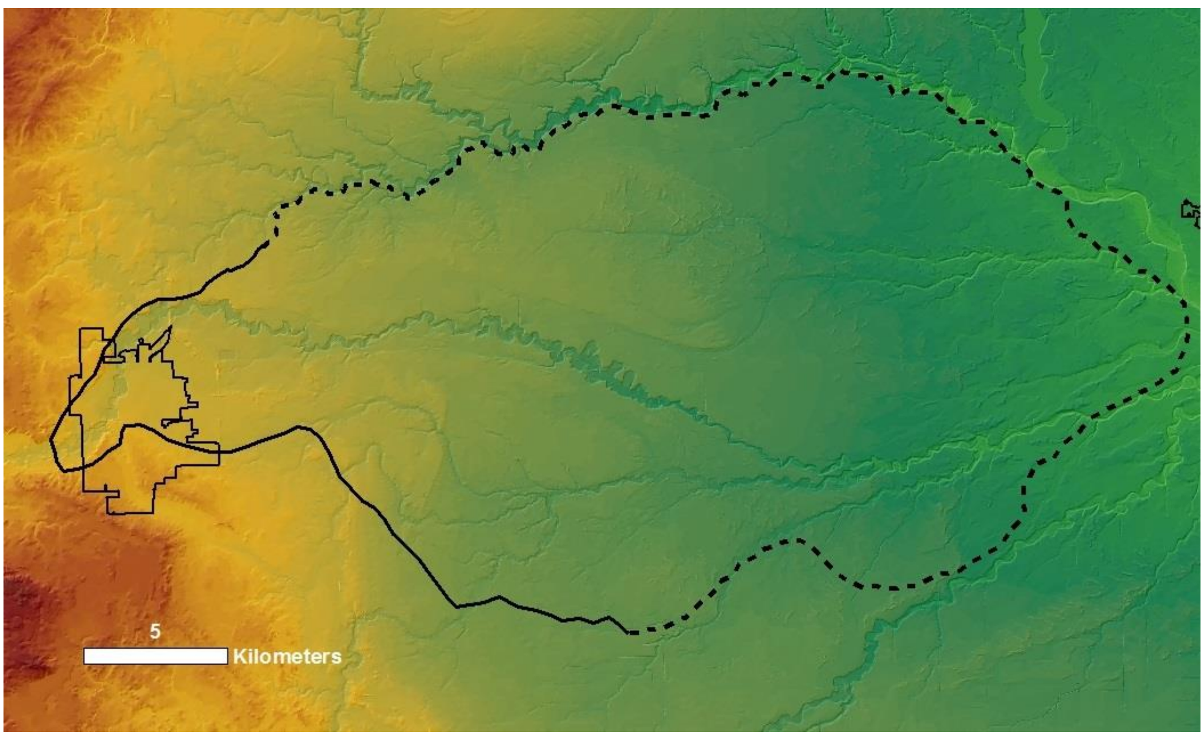



Figure 5.4 – DEM hillshade showing surface features of the Chippewa Delta, as well as the delta boundary, and the outline of the city of Mt. Pleasant.

loamy lake plain soils beyond the margins of the delta (Hutchinson, 1979; McLeese and Tardy, 1985; Schaetzl et al., 2013).

The Chippewa delta surface is a broad, relatively flat and sandy plain. It slopes gently (generally <1%) from the western uplands to the east. In the north-central part of the delta lies a fairly large (~5x8km) dune field, with a number of broad (~700m) and low lying (~1m in height), but closely spaced, parabolic dunes (Fig. 5.5). To the east and southeast of this dune field, dunes appear more sporadically, and are more widely spaced. Most of the delta serves as residential land or state forest; the Chippewa River State Forest and Tittabawassee River State Forest account for much of the northwestern, north-central and south-central parts of the delta, which include the previously mentioned dune fields (Fig 5.6). Large portions of the delta

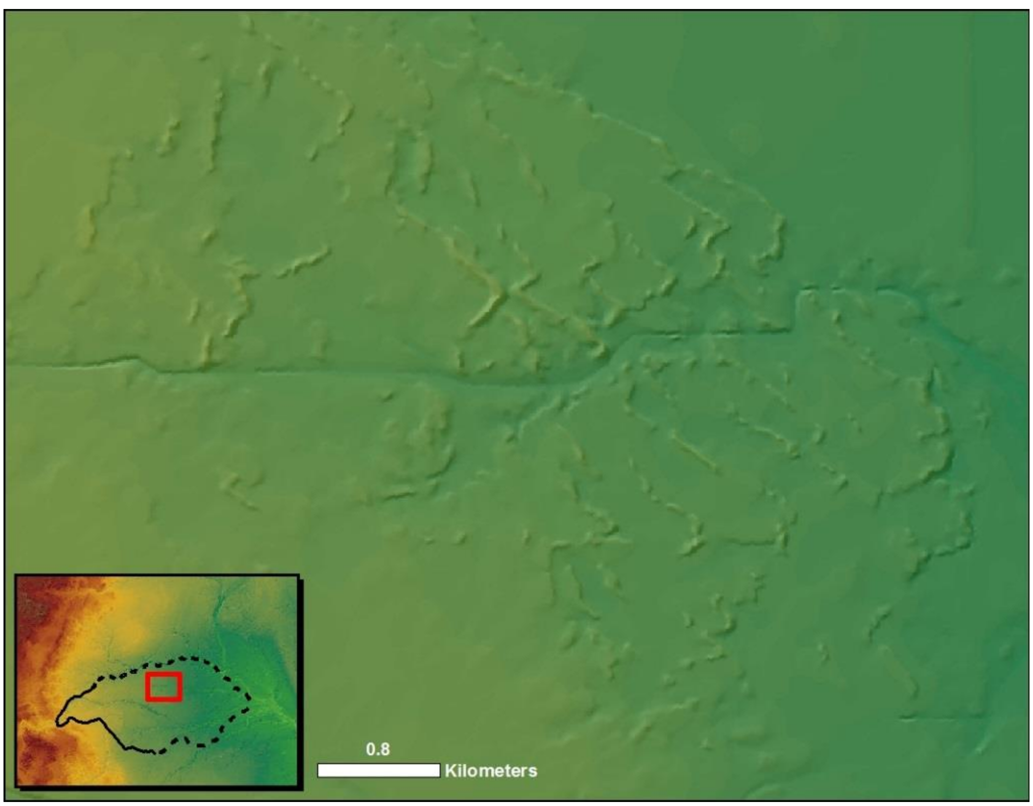



Figure 5.5 – Hillshaded DEM showing the dune field on the north-central portion of the Chippewa Delta. Inset map includes the outline of the delta, and extent of the main map.

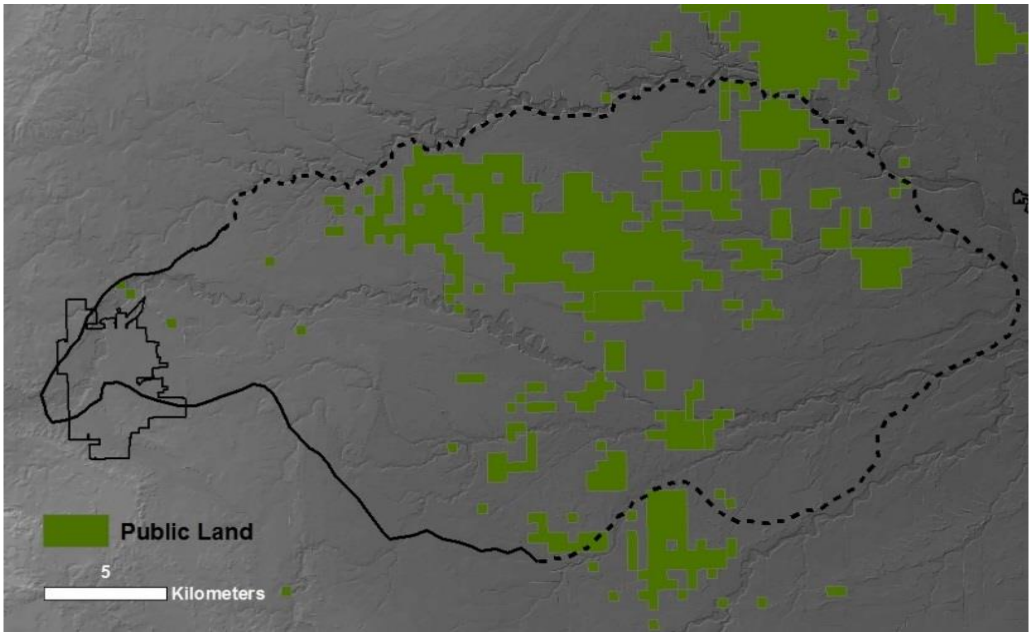



Figure 5.6 – Public land parcels on the Chippewa Delta, including Chippewa River and Tittabawassee River State Forest land (data from Michigan Geographic Data Library).

are swampy, and a number of drainages dissect the southern part of the delta, e.g., Black Creek, Onion Creek, Roller Creek, Salt Creek, Little Salt Creek, and Thrasher Creek. The northern delta boundary roughly follows the highest elevation terraces of the Salt River, which flows eastward where it eventually discharges into the Titabawassee River. The southern margin of the delta appears to overlap with deltaic sediments from another, smaller delta associated with the Pine River. Very little of the delta surface is used for farming, implying that it is comprised of poor quality agricultural land, i.e., it is sandy and wet.

**5.4 Soils on the Delta and Surrounding Landscapes**

NRCS soils data were critical to this research, as they provide a continuous and generally accurate spatial dataset that is a proxy for surficial geology. The hypothesis that the lobate sandy area at the junction of the Chippewa River and the Glacial Lake Saginaw plain was a delta was initially based on soil texture patterns, but required more detailed study to confirm. Often, geologists studying paleo-deltas will note and observe characteristics such as bedding planes and sedimentary structures to help inform their interpretations. In this research, I was unable to directly observe any outcrops of delta sediments, and so I tailored a methodology that focused on spatio-textural trends in sands, recovered from below the soil profile. Soil parent material data like these can be used to characterize the spatio-textural patterns across the delta surface, as initially demonstrated in a study of the Black River Delta in northern Michigan (Vader et al., 2012). Similar data were likewise helpful in outlining relict deltas from the literature. For example, Leverett and Taylor (1915) provided written descriptions of a number of relict deltas associated with paleolake shorelines in the Erie basin in southeastern Michigan (Fig. 5.7). I was able to locate these features by displaying the subsurface texture variable in a GIS, which showed the lobate/cuspate distribution of sandy soils associated with a former river mouth/junction with a known paleolake. These soils stood in contrast to loamier soils of the lake plain. Similar patterns in the soil data within my study area helped me to map and interpret the Chippewa Delta. Because this study relied on observing patterns and relationships in soil data, I will begin reporting results of GIS analysis of the NRCS soil data.

### 5.4.1 Patterns in Subsoil Texture on the Delta Plain

Soils on the Chippewa Delta primarily have sandy surface and subsurface textures; roughly 74% of the soils on the delta have sandy textures in their subsurface horizons (Table 5.1). The areal pattern made by these sand textured soils (based on their subsoil materials) is lobate, as is common in sandy, Gilbert-type deltas. The sands become more discontinuous and interspersed with areas of soils with loam textured parent materials near the delta periphery (Fig. 5.8). This pattern helped delimit the margins of the delta. As previously discussed, for this

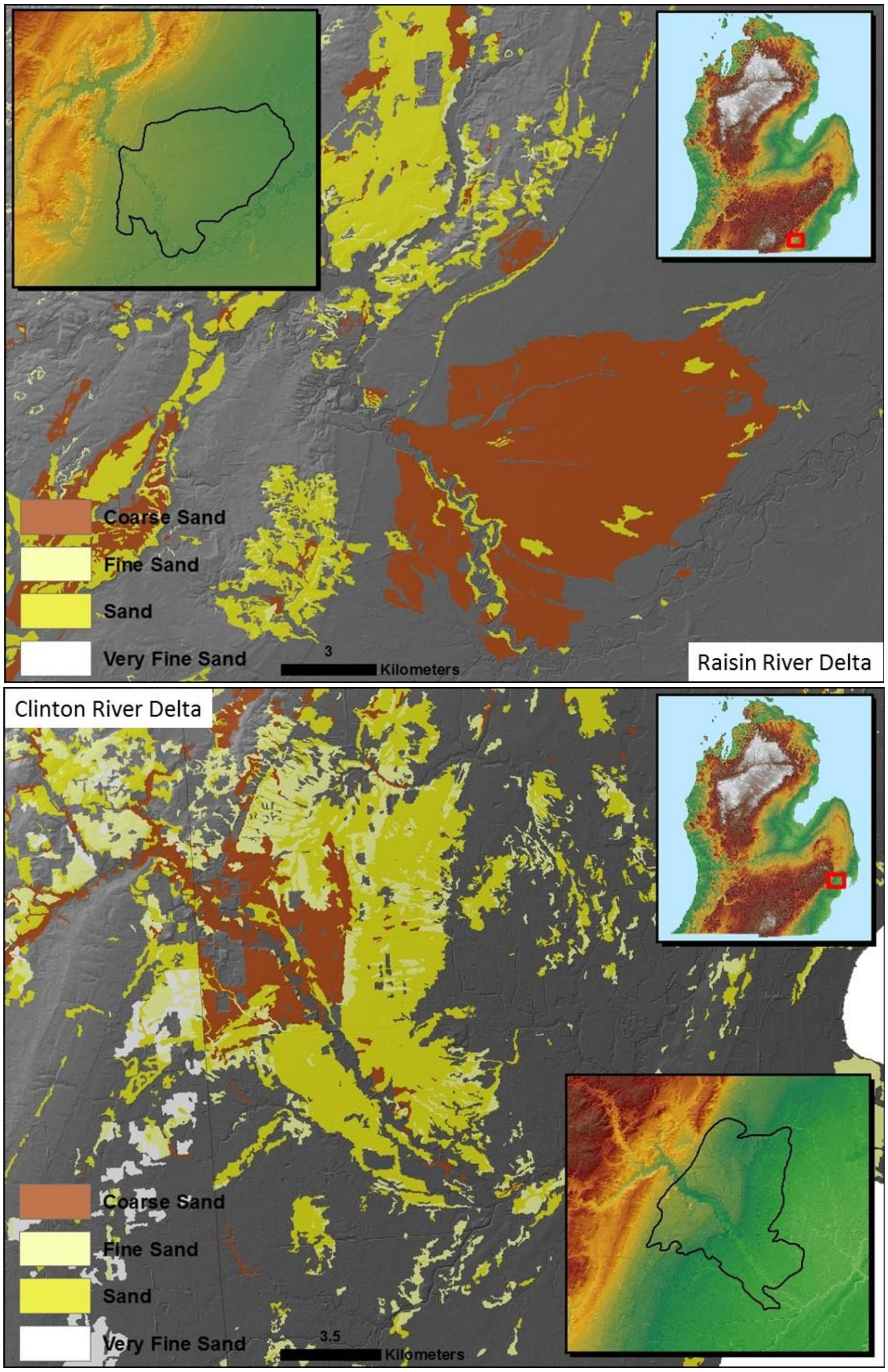



Figure 5.7 – Maps showing the subsurface textures (sand classes) for two relict deltas identified by Leverett and Taylor (1915): the Raisin River delta (top) and the Clinton River delta (bottom). Areas where the gray hillshade is visible are loamy. Inset map (top right corner) shows the location and extent, the other inset map shows the DEM and hillshade, and an outline of the sand textured subsoils that are displayed in the main maps (data from USDA-NRCS).

research, soil subsurface textural characteristics have been considered more useful as proxy geologic data than are surface horizons, because soil forming processes can alter the characteristics of the upper solum.

Where subsurface textures on the delta are sandy, they dominantly fall within the sand texture class (91.1%), followed by coarse sand (5.8%), very fine sand (2.2%), and fine sand (0.75%) (Table 5.1). Soils with coarse sand subsurface textures are mostly located in the

| Sand Textural Class | NRCS Description | Percent Within Sand Types On The Delta |
|---|---|---|
| **Sands** | More than 85% sand. Percent silt plus 1.5 times percent clay <15% of total. | |
| Coarse Sand | =>25% VCS and CS and <50% any other grade of sand | 5.3% |
| Sand | =>25% VCS, CS, and MS, <25% VCS and CS and <50% FS and <50% VFS | 91.2% |
| Fine Sand | =>50% FS, <25% VCS, CS, and MS, <50% VFS | 1.0% |
| Very Fine Sand | =>50% VFS | 2.5% |

Table 5.1 – Table of USDA-NRCS description of sands, and the various fractions of sands, as well as the percent representation of subsurface texture among sandy soils on the delta.

western, upper part of the delta, near the city of Mount Pleasant, and also in active stream channels (Fig. 5.8). Soils with very fine sand subsurface textures are also generally located in the western, upper part of the delta, near Mount Pleasant, and to the east of the city. Also noteworthy is an area to the east of Mount Pleasant which has soils with loamy parent materials. In general, patterns in subsurface textures show that they are coarsest in the upper, western reaches of the delta, are finer elsewhere on the delta, and are loamy near the distal margins, where they are overlain by thin sandy upper sola, in two-storied soils. Only general inferences can be drawn from the NRCS data, however, and thus a more detailed and quantitative analysis of soil texture from field collected samples will be discussed later.

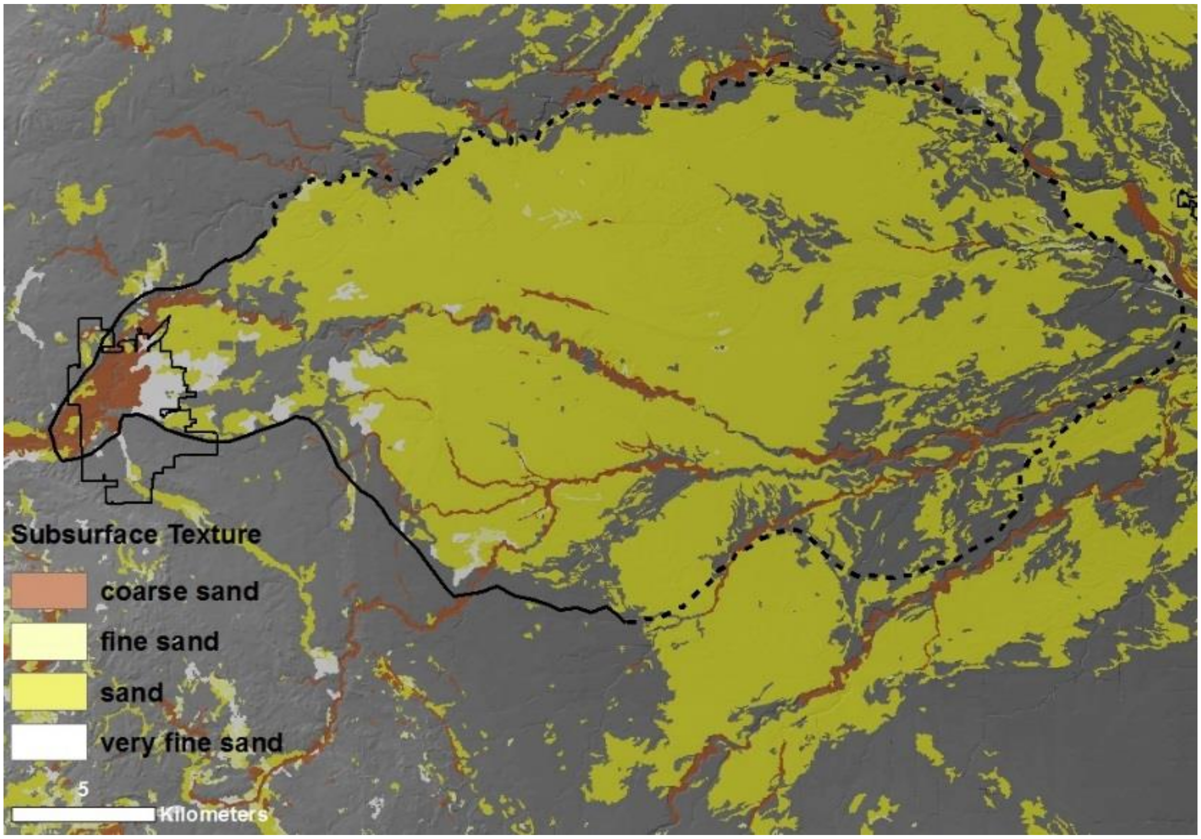



Figure 5.8 – Map of the distribution of soils with sand textured parent materials. Areas where the gray hillshade are visible are loamy (data from USDA-NRCS).

**5.4.2 Patterns in Upper Solum Texture on the Delta Plain**

Although upper solum texture data were thought to be less useful in this research than were subsurface texture data, they did exhibit general patterns that helped characterize the delta. As with the subsurface data, surface textures of soils across the delta plain are dominantly within sand and fine sand texture classes (Fig. 5.9). Near the fringes of the delta, surface textures usually fall within loamy sand, loamy fine sand, and loamy very fine sand texture classes (Figs. 5.9, 5.10). The loamy sands are generally located near the distal fringes of the delta, where they form the upper sola of many of the two-storied soils, implying that their distribution can also be used to infer overall sand thickness in these areas (Fig. 5.10). Loamy sands, by definition, have more silt and/or clay content than do sands, and so the loamy sands near the margins of the delta suggest that these sediments are not as well-sorted as are the soils that comprise the inner portion of the delta, typified by sand and fine sand textures. The increase in finer textures at the margins may also generally conform to the "fining-outward" sedimentary pattern typical of deltas.

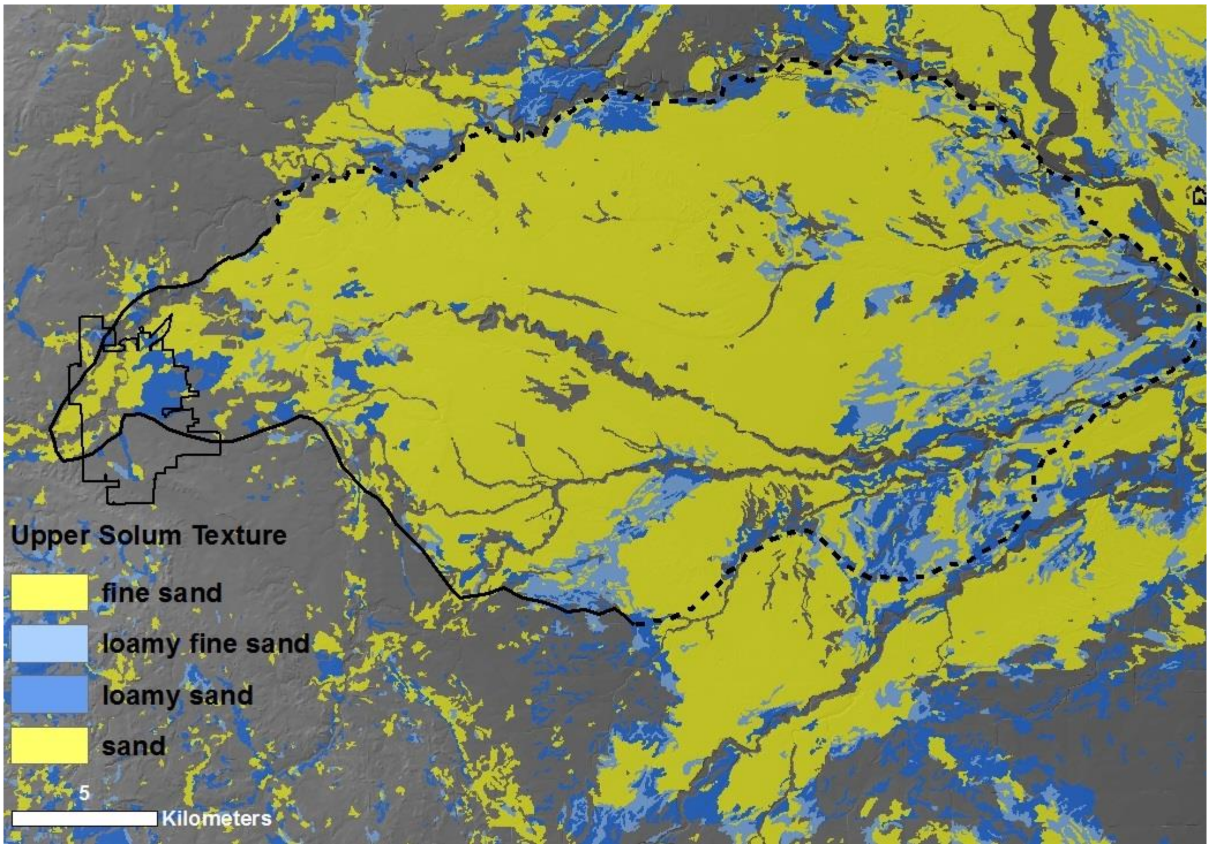



Figure 5.9 – Map of upper solum textures for the sandy soils on the delta and surrounding landscape, as well as the delta boundary and DEM hillshade. Soils with loamy sand textures are mainly found near the fringes of the delta, and generally coincide with two-storied soils (data from USDA-NRCS).

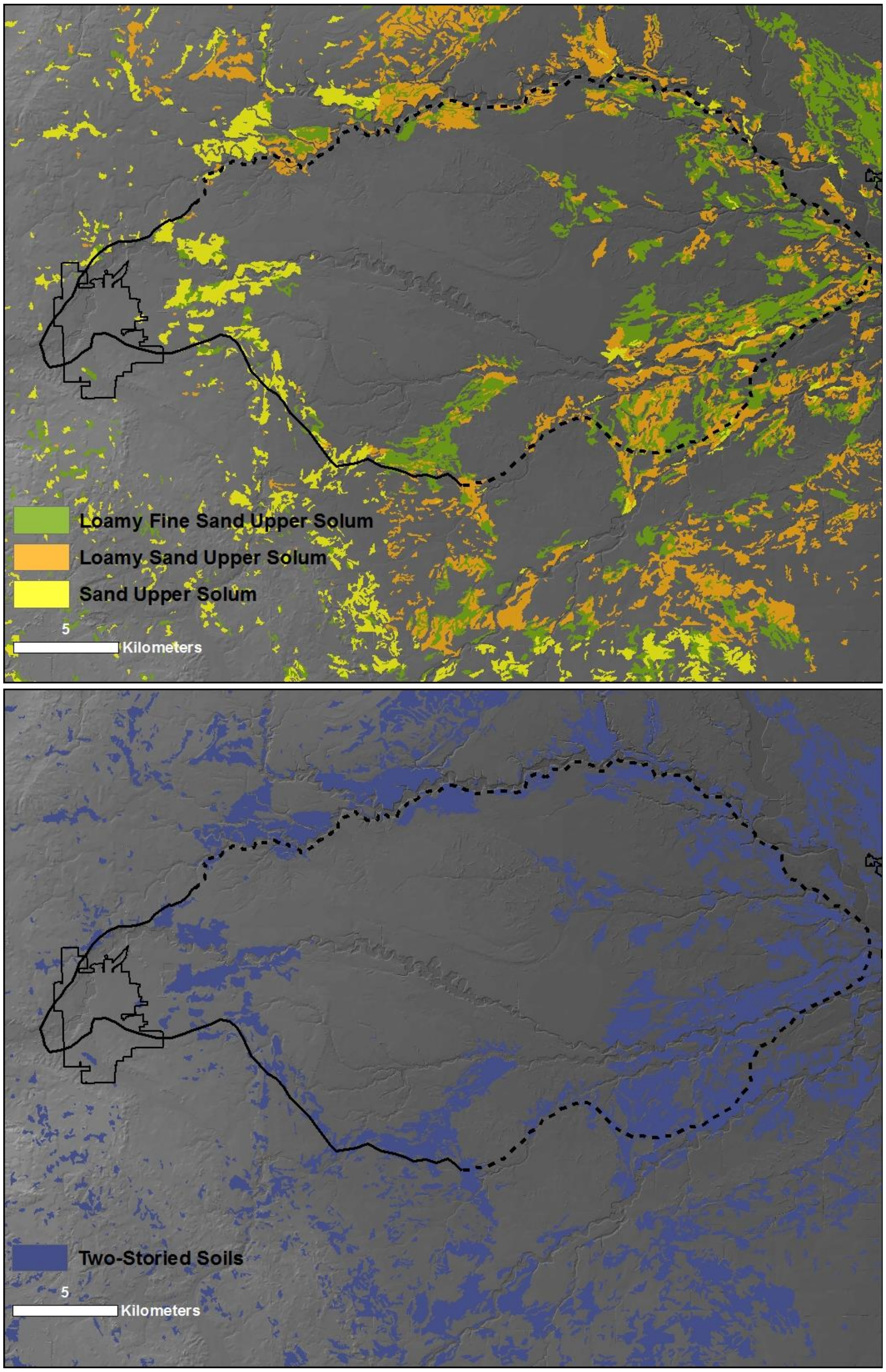


Figure 5.10 – Maps showing the (A) upper solum textures of two-storied soils on the delta. The inset figure shows “loamy sand” on the USDA-NRCS soil textural triangle, and (B) map of the distribution of two-storied, sandy over loamy, soils on the delta and in the surrounding landscape. Both maps include DEM hillshade, the city of Mt. Pleasant. and the outline of the delta.

### 5.4.3 Soils in the Chippewa River Watershed

Upstream of the city of Mt. Pleasant, the Chippewa River drains a watershed of roughly 1000 $km^2$ including parts of Clare, Isabella, Mecosta, and Osceola Counties (Fig. 5.11). The landscape within the watershed, west of the Chippewa Delta, consists of "morainic uplands" with loamy, till-cored moraines and sandy outwash plains. Outwash surfaces comprise a

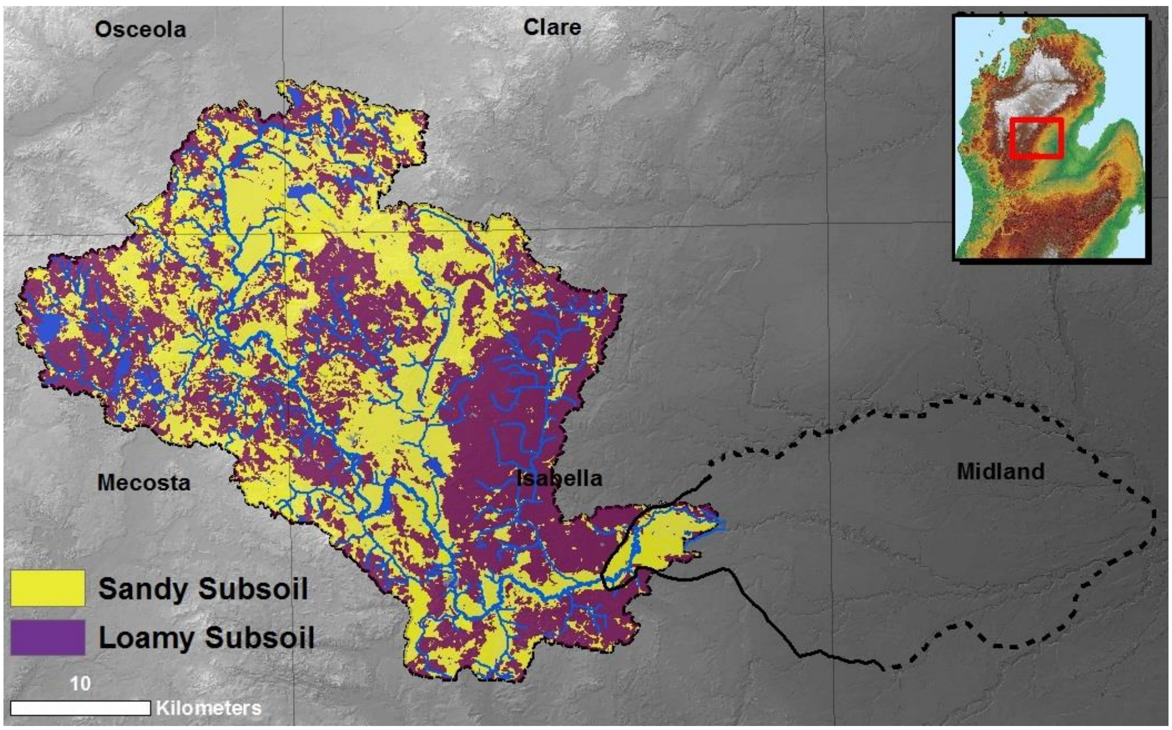



Figure 5.11 – The distribution of soils within the Chippewa River watershed grouped into two subsurface textural groupings, also showing streams and lakes within the watershed, and counties (data from the Michigan Geographic Data Library).

significant portion of the watershed; sandy textured soils (subsurface) comprise ≈49% of all soils within the watershed (Fig. 5.11; Table 5.2). The outwash surfaces within the watershed would have been potential sources of sand for transport and deposition on the delta.

| Texture | Percent Ratio Within Watershed | Percent Ratio Within Sand Types |
|---|---|---|
| **All Sand Types** | **49.2%** | |
| Sand | 33.8% | 68.6% |
| Fine Sand | 6.8% | 13.9% |
| Coarse Sand | 4.5% | 9.1% |
| Loamy Sand | 2.4% | 4.9% |
| Very Fine Sand | 1.5% | 3.1% |
| Loamy Fine Sand | 0.2% | 0.5% |

Table 5.2 – Table of sand textures (subsurface) from soils within the Chippewa River watershed. See Table 5.1 for NRCS descriptions of sand fractions.

### 5.4.4 Summary of Soil Data Results

NRCS soils data revealed a number of patterns that I interpret to be consistent with delta sedimentary systems, and which also helped delineate the extent of the delta (Fig. 5.12). Subsurface texture data show a general fining-outward pattern, from the coarse sand upper portion of the delta, to the sand textured main body (Fig. 5.8). This pattern was also present in the upper solum textures, where sand and fine sand were the dominant textures in the main body of the delta, and various loamy sand textures were located near the margins, as the upper part of two-storied soils (Figs. 5.9, 5.10). The two-storied soils were usually loamy sand textured in the upper solum, and overlaid loamy-textured, lake plain materials; they were interpreted as thin deltaic sediments associated with the delta margin (Figs. 5.9, 5.10). Later in this section, quantitative data on the sands of the delta, collected in the field, will be used to supplement the general inferences derived from the NRCS soils data, allowing more detailed inferences to be made concerning textural changes across the delta surface.

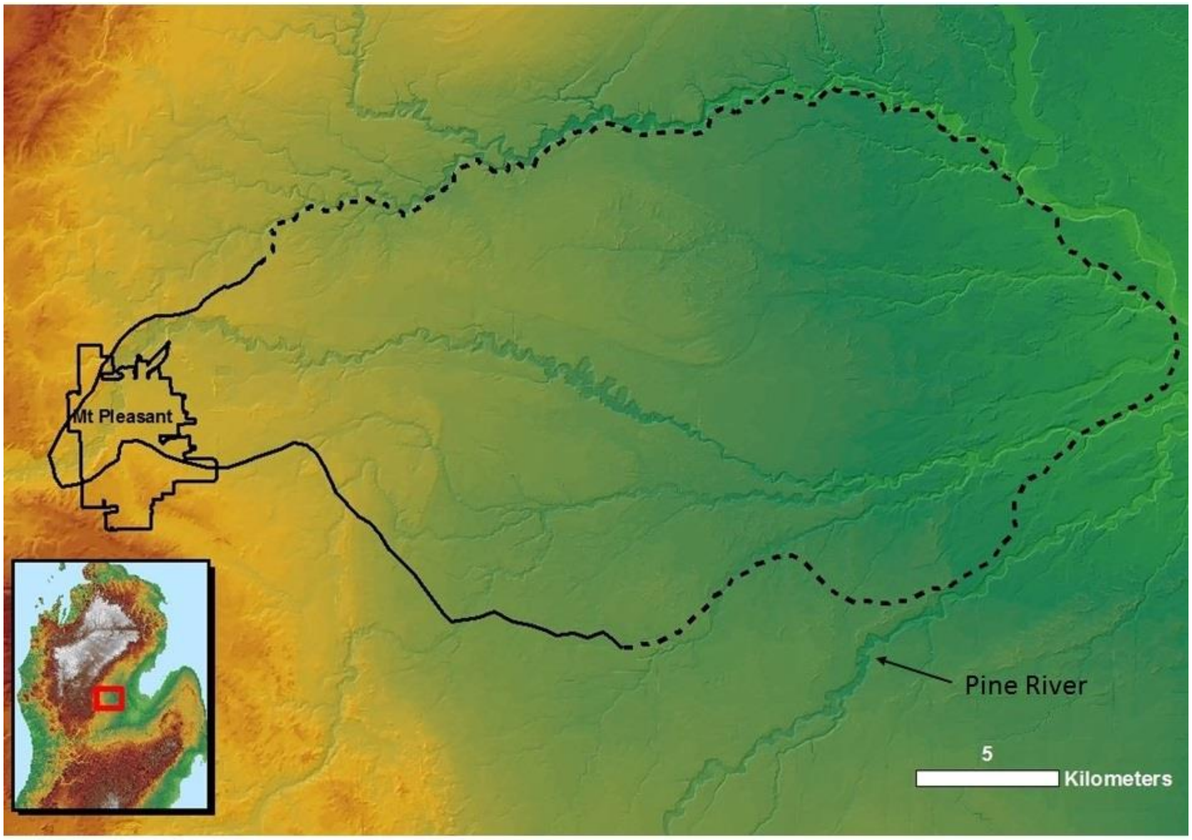


Figure 5.12 – The boundary of the Chippewa Delta as delineated in this research, over DEM hillshade, inset map indicates location and extent of the main map.

### 5.5 Delta Thickness

Data from 3276 water well logs and 185 two-storied soils were used to estimate the thickness of the sandy sediments on the Chippewa delta. As discussed in the methods section, 185 points were randomly placed on NRCS soil map units where two-storied soils were mapped, and the sand thickness value at each point was derived from the OSD for each soil series. These values represent the thickness of thin sandy upper sola of two-storied soils, nterpreted as being marginal deltaic sediments. The conversion of well log data to points representing sand thickness was conducted before the boundary of the delta was determined, and as a result, many points fell outside of the final delta boundary outline. This analysis assumes that the sands within the delta boundary are mainly deltaic in origin. The thickness data were interpolated using an ordinary kriging routine (Fig. 5.13).

Delta sediments (sands) are thickest in the upper, western half of the delta, and thin regularly toward the periphery (Fig. 5.13). In the thickest area, bisected by the present day Chippewa River, the sands are often 6-7m thick. In the lower, eastern part of the delta, the sands are generally thinner than the sediments to the west, ranging between roughly 3m in the

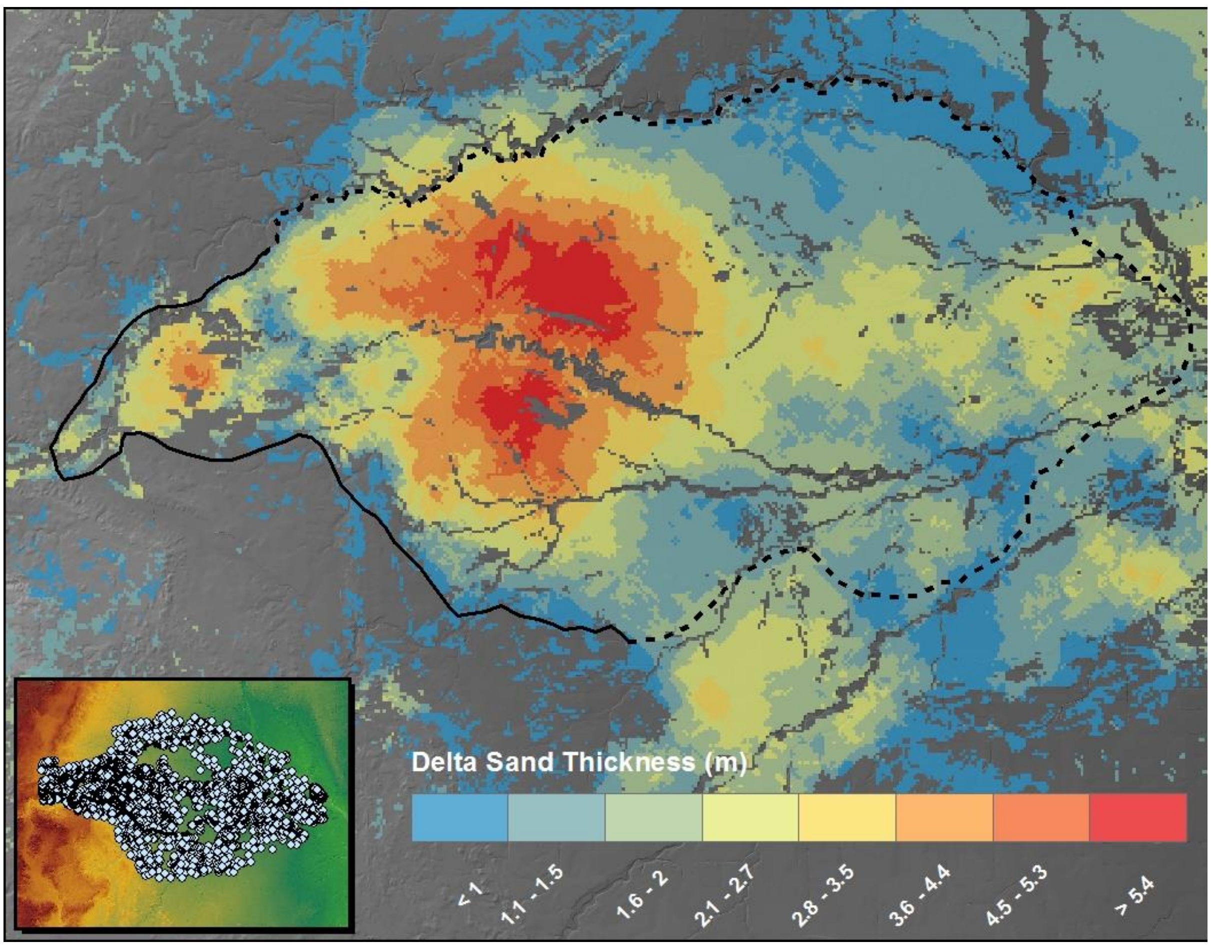



Figure 5.13 – Interpolated map of delta sand thickness, clipped to areas with sand textured upper solum. Inset map shows the distribution of the 3461 total points used to generate the interpolation.

thickest areas, to less than 1m in thinner areas. A relatively small area of thin sands northeast of Mt. Pleasant lies between two areas of comparatively thicker sands, displayed in red/pink in Figure 5.13. This thin area generally coincides with the only two-storied soils, as mapped by the NRCS, in the upper part of the delta (Fig. 5.10). To the east of this zone of thin sediments, the delta widens, and delta sediments thicken (Fig. 5.13).

A random sample (*n*=200) of the krig of sand thicknesses on the delta show that delta sands are generally thin, averaging 2.3 m, with a median thickness of only 2.0 m (Fig. 5.14). The majority of sample points fell on locations where sand thicknesses were between one and two meters (Fig. 5.14). These data illustrate how generally thin the delta sands are, despite having a few areas of thicker sediment, as shown in the A-A' profile in Figure 5.15.

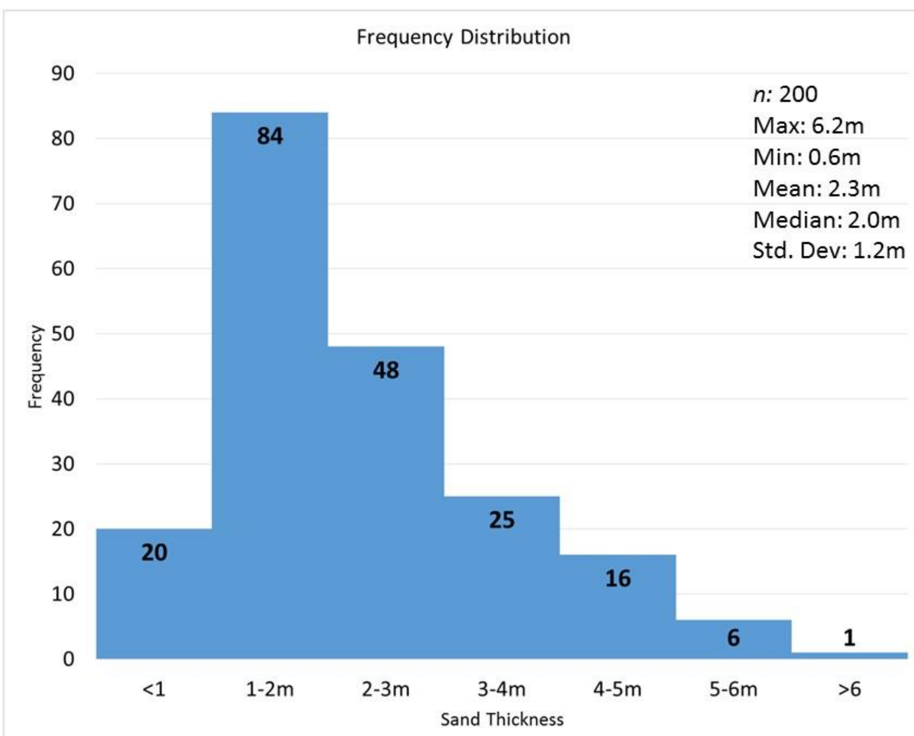



Figure 5.14 – Histogram of sand thicknesses across the Chippewa Delta, derived from 200 points randomly placed on the delta. The data values were extracted from the kriged thickness layer.

The thickness and textural data suggest that a two-tiered pattern of sandy sediment exists on the delta. The delta is sandy throughout, and lacks the distal muddy fringe that characterizes

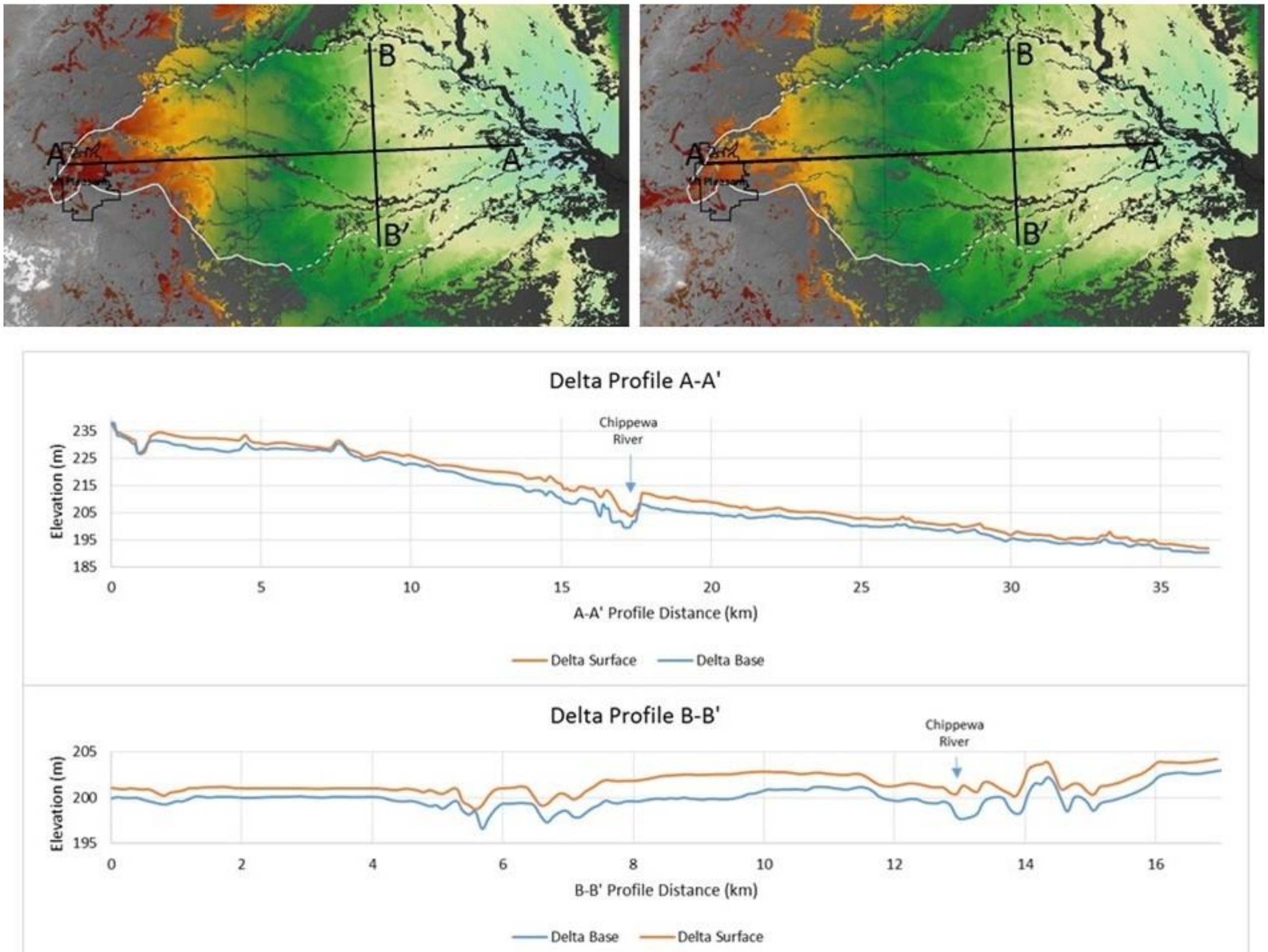



Figure 5.15 – Profiles of the delta surface (top left), determined from a DEM, and an estimate of the base of the delta (top right), estimated using a DEM with the kriged sand thickness interpolation subtracted using raster calculator. A-A' profile is ~west to east, B-B' profile is ~north to south.

the bottomset beds of the typical Gilbert-type delta. Therefore, the thick upper portion of the delta, and the thin distal portion of the delta, both texturally sandy, may actually be two coarse grained deltas, i.e., a two-part delta complex representing two distinct periods of delta building. This hypothesis will be discussed later in this section.

## 5.6 Particle Size Characteristics of the Field Samples

Although 154 parent material samples were collected from across the delta, some were not utilized for spatio-textural analysis. Twelve samples were removed after it was discovered that they had been collected from within a wide, but shallow, inset channel on the delta (Fig 5.16). This channel was likely formed by fluvial incision into delta sediments as a response to lake level lowering. It is possible that the sediment within the channel, while originally deltaic in origin, may have been subsequently mobilized by the stream system; some of the sediment here may be Chippewa River alluvium, deposited when the channel was active. A few (23)

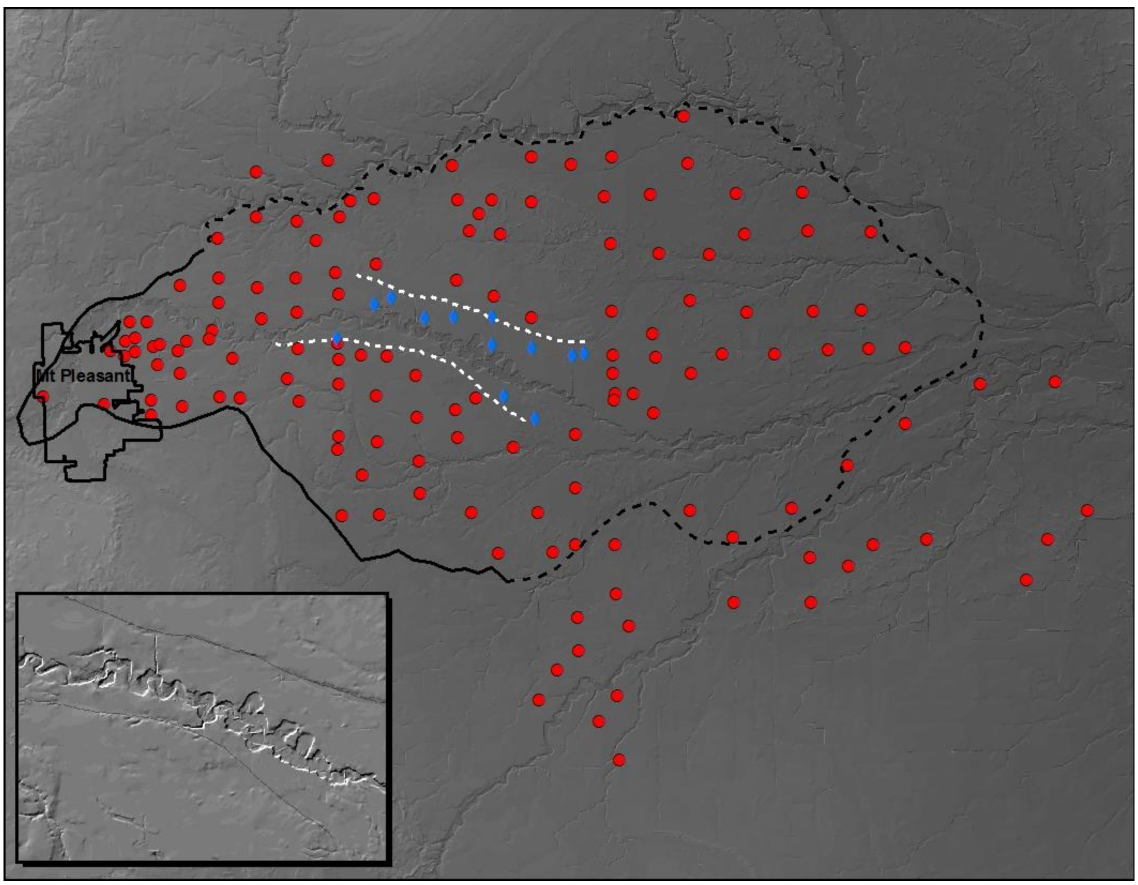



Figure 5.16 – Map showing the erosional channel on the delta surface, the samples within it (blue) that were removed from the dataset, and the samples (red) that were retained in the dataset. Inset map shows the channel as it appears on the DEM hillshade.

additional samples were collected from areas that were later determined to be outside of the delta boundary; those samples nonetheless were retained in the final dataset, because their data did prove helpful in defining the delta extent. The final dataset included 142 points in sandy sediments, and an additional 20 samples collected from till within the Chippewa River watershed, in the morainic uplands.

**5.6.1 Sandy Samples on the Delta and Lake Plain**

Texturally, the majority of field samples, 102 of 142, fall within the sand textural class (Soil Survey Division Staff, 1993). Of the remaining 41 samples, 21 were coarse sand, 10 were fine sand, three were very fine sand, and two were loamy fine sand. Only one sample each had textures of fine sandy loam, loamy coarse sand, very fine sandy loam, and sandy loam (Fig. 5.17).

Patterns in soil textural classes from the field data show similar trends to the NRCS data; most samples classify as sand. However, there are more coarse sand textures in the field data than are represented in the NRCS data. In the field data, the majority of coarse sand textured samples are in the upper portion of the delta, with one group near the city of Mt. Pleasant, and another, smaller group, farther east, where the delta becomes significantly wider. A third, smaller group is in the lower delta, where the sediments are thinner. These three general zones of coarse textured sediment appear consistently in the field sample data. Many of the fine and very fine sand textures are near the margins of the delta, which is a trend that also appears consistently in the field data.

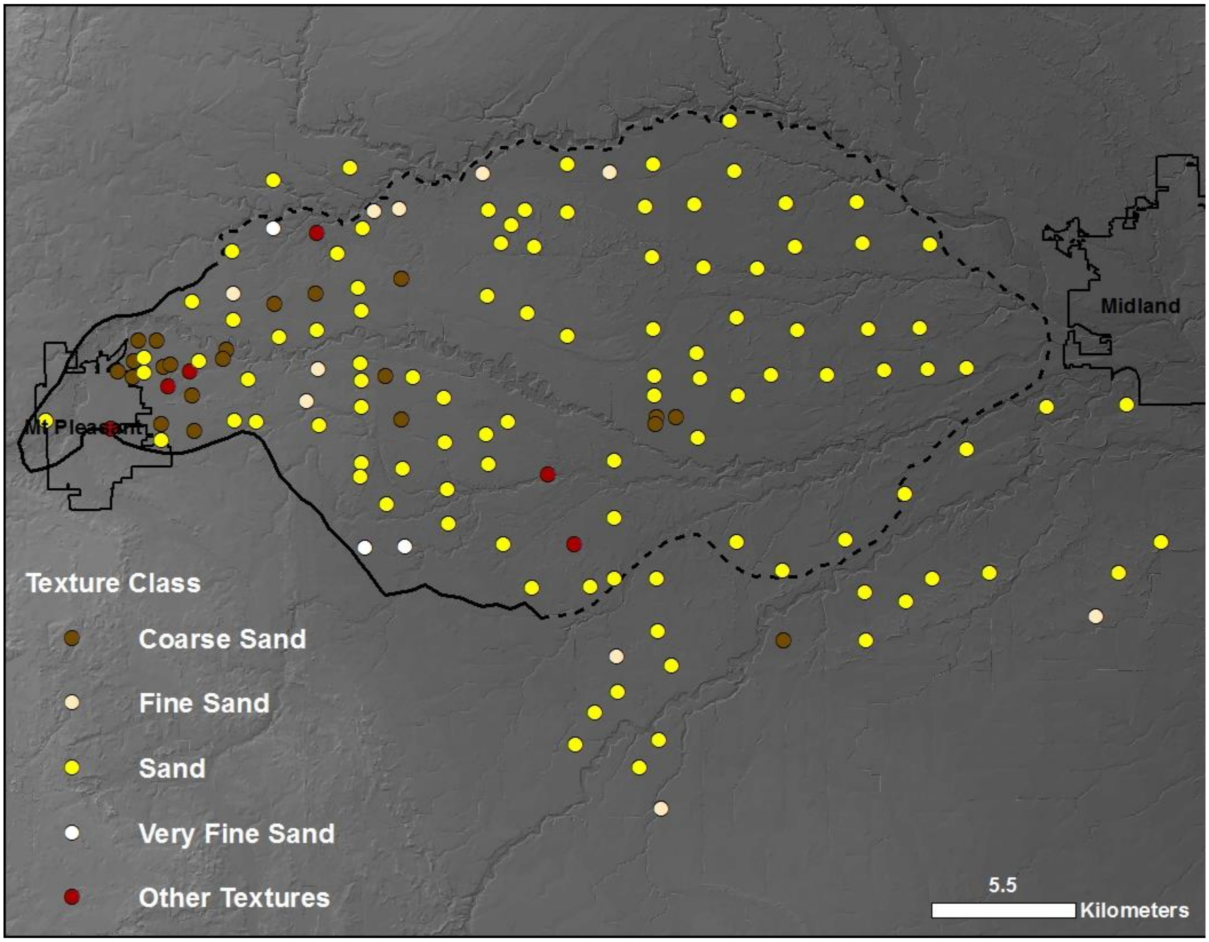


Figure 5.17 – The distribution of samples by texture class.

#### 5.6.1.1 Particle Size Mode

All of the samples collected from the delta surface had particle size modes in the sand fraction (>50µm). Most of the samples, 112 of 142, had particle size modes that peak within the medium sand (250-500µm) fraction. Of the remaining samples, seven peaked in the coarse sand fraction (500-1000µm), 16 peaked in fine sand (125-250µm), and seven peaked in very fine sand (50-125µm) (Figs. 5.18, 5.19, 5.20). For the purpose of visualizing the modal changes across the delta, these data were interpolated and clipped, using the distribution of sandy subsurface textures per the NRCS data (Fig. 5.19).

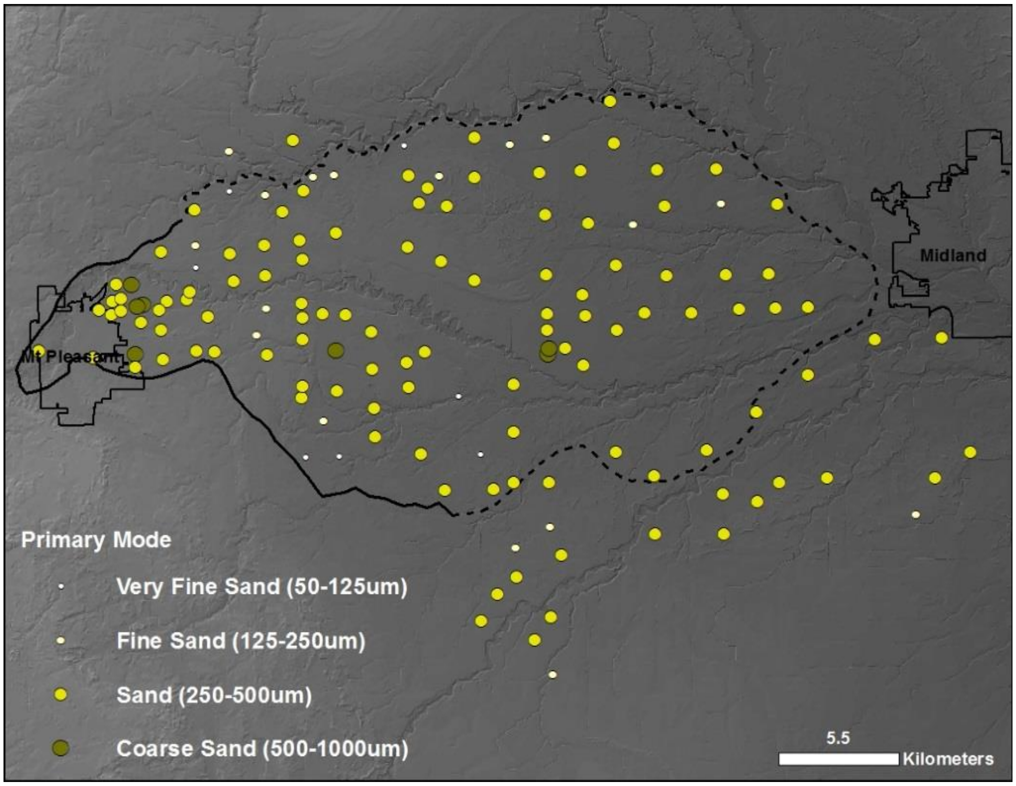



Figure 5.18 – The distribution of primary mode by sand fraction on the delta and in the surrounding landscape.

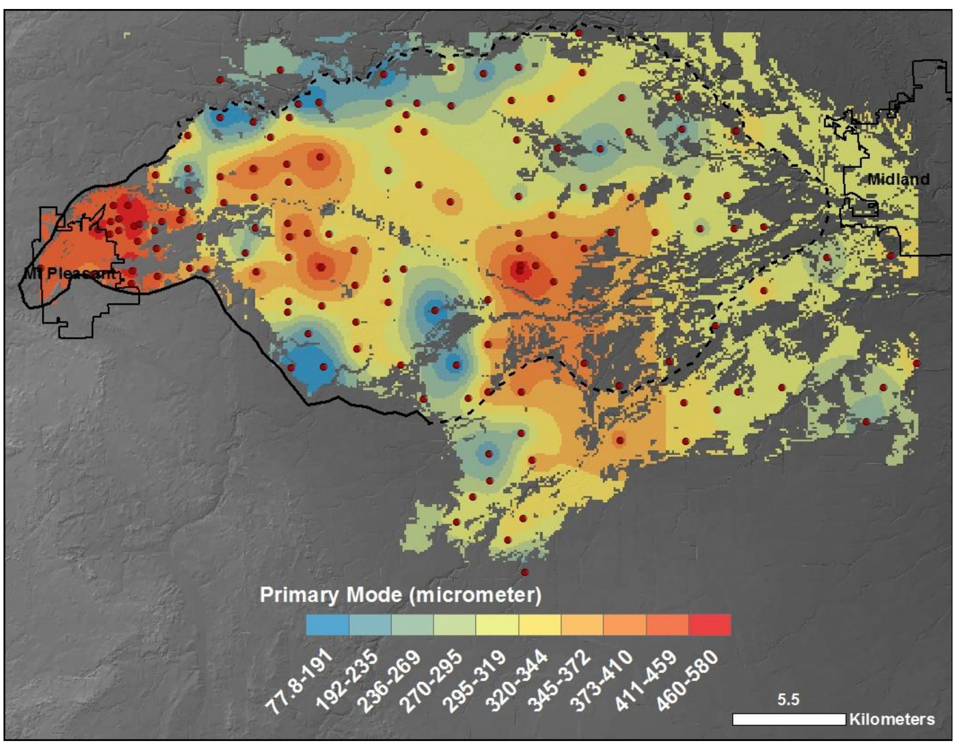



Figure 5.19 – Interpolated map of primary mode, clipped to areas with sand textured subsoil.

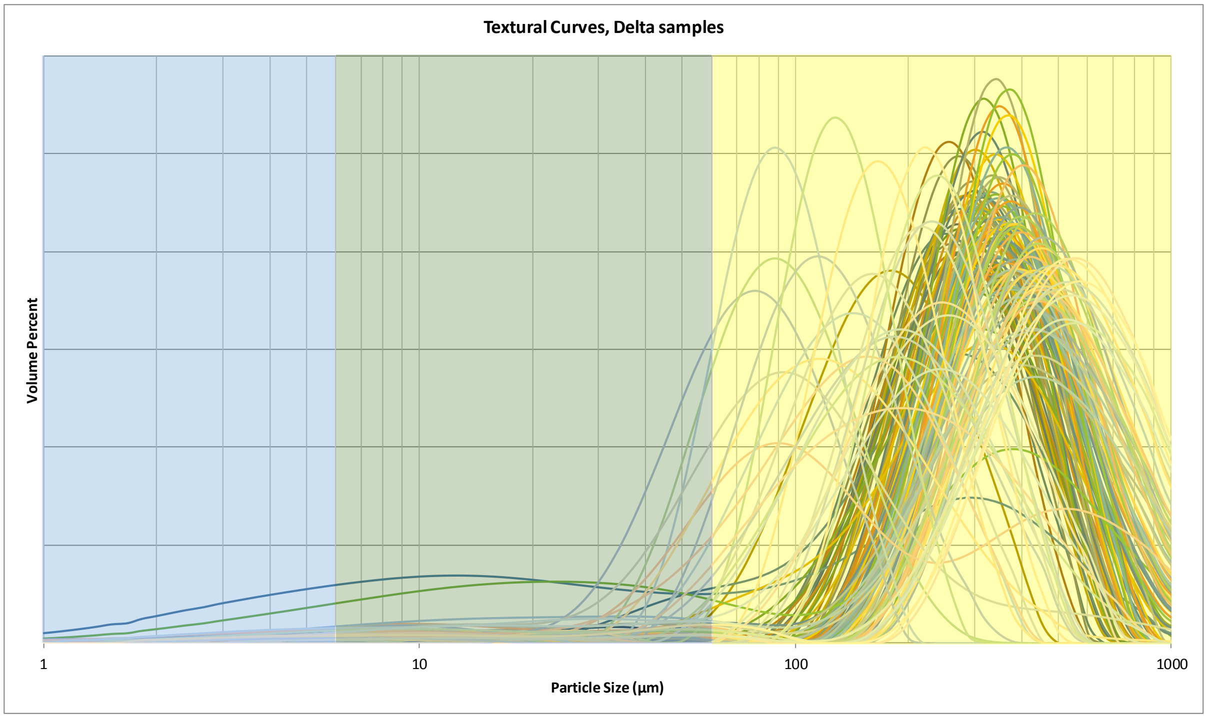


Figure 5.20 – Textural curves, the graphical representation of the binned data output from laser diffractometry, from field samples collected from the delta and sandy surrounding landscape. Particle size fraction are color coded on the chart: clay (blue), silt (green) and sand (yellow) fractions. All of the samples collected from the delta have sand modes.

### 5.6.1.2 Gravels

Gravel (>2000µm) contents in the samples were calculated by mass, as discussed in the methods section. Only 45 of 142 samples contain >1% gravel. These samples are mainly located in the upper part of the delta, near Mt. Pleasant (Figs. 5.21, 5.22). Other samples that have larger gravel percentages are generally located near the other texturally coarse areas, as shown in data from primary mode, and texture class (Figs. 5.17, 5.18, 5.19).

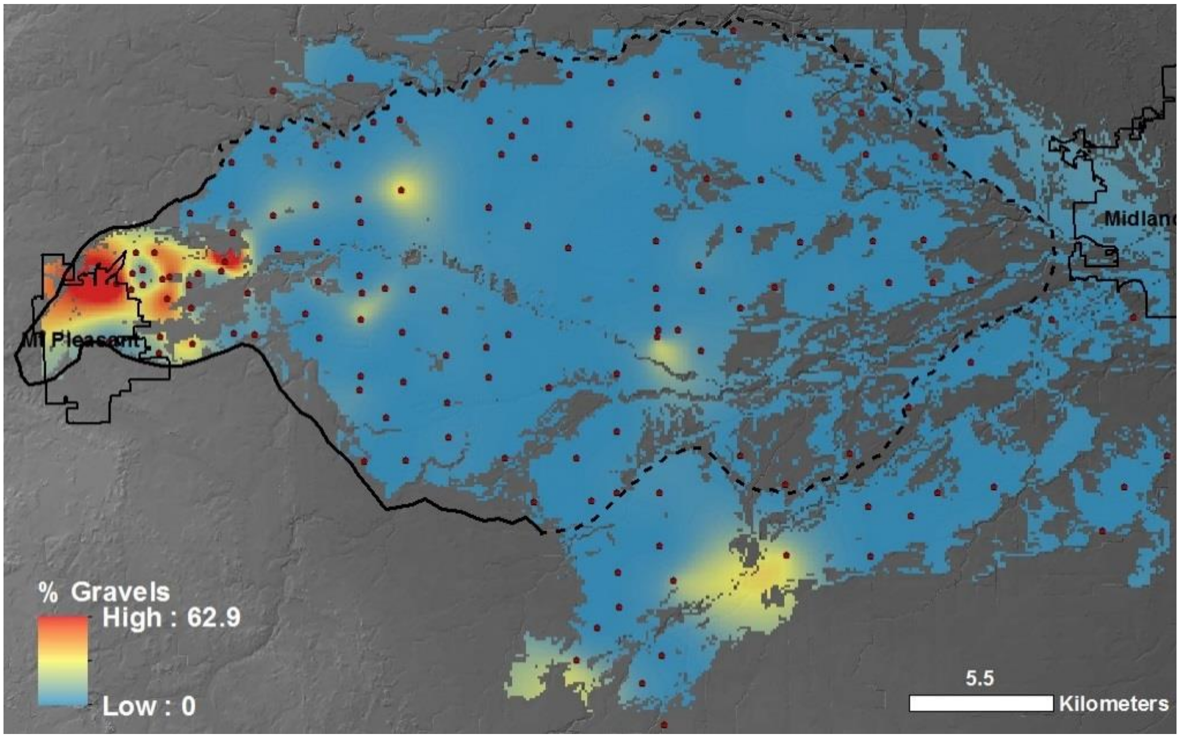


Figure 5.21 – Interpolated map of gravel (>2000µm) content, clipped to areas with sand textured subsoil.

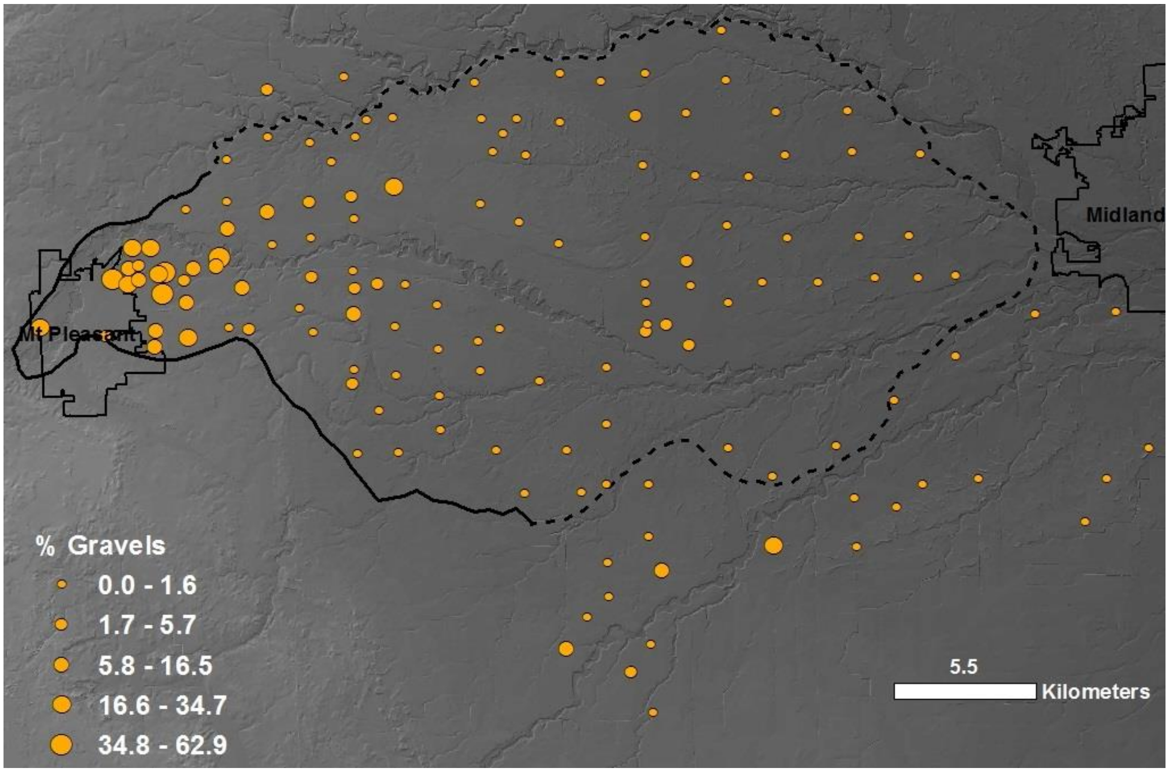


Figure 5.22 – Graduated circle map of gravel (>2000µm) content for the 142 soil samples used in analysis/interpolation.

#### 5.6.1.3 Very Coarse Sand

Like gravel content, the content of very coarse sand (VCS, 1000-2000µm) in the soil samples was also manually calculated. In general the amount of VCS within the delta sediment is relatively low, ranging from zero to ~12% (Figs. 5.23, 5.24). Samples with the largest percentages of VCS generally occur near the coarsest areas according to texture class, primary mode, and gravel content maps (Figs. 5.17, 5.18, 5.19, 5.21, 5.22).

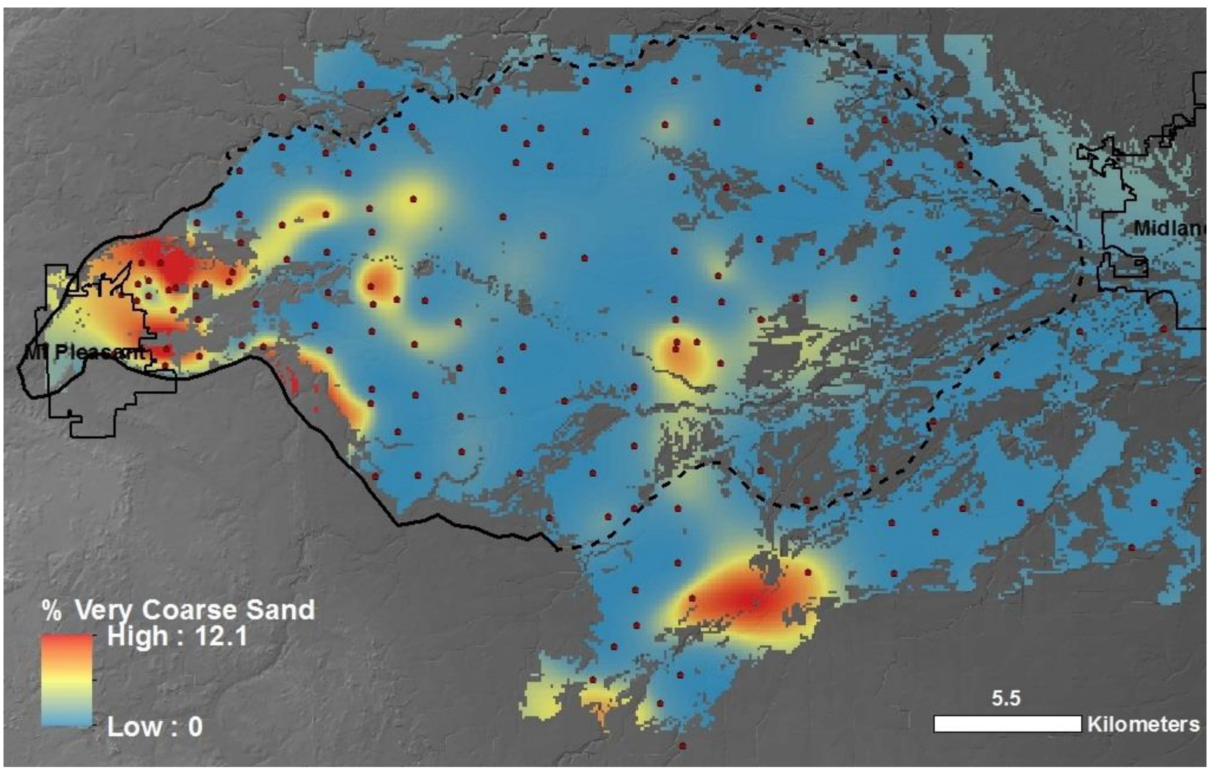



Figure 5.23 – Interpolated map of very coarse sand (1000-2000µm) content, clipped to areas with sand textured subsoil.

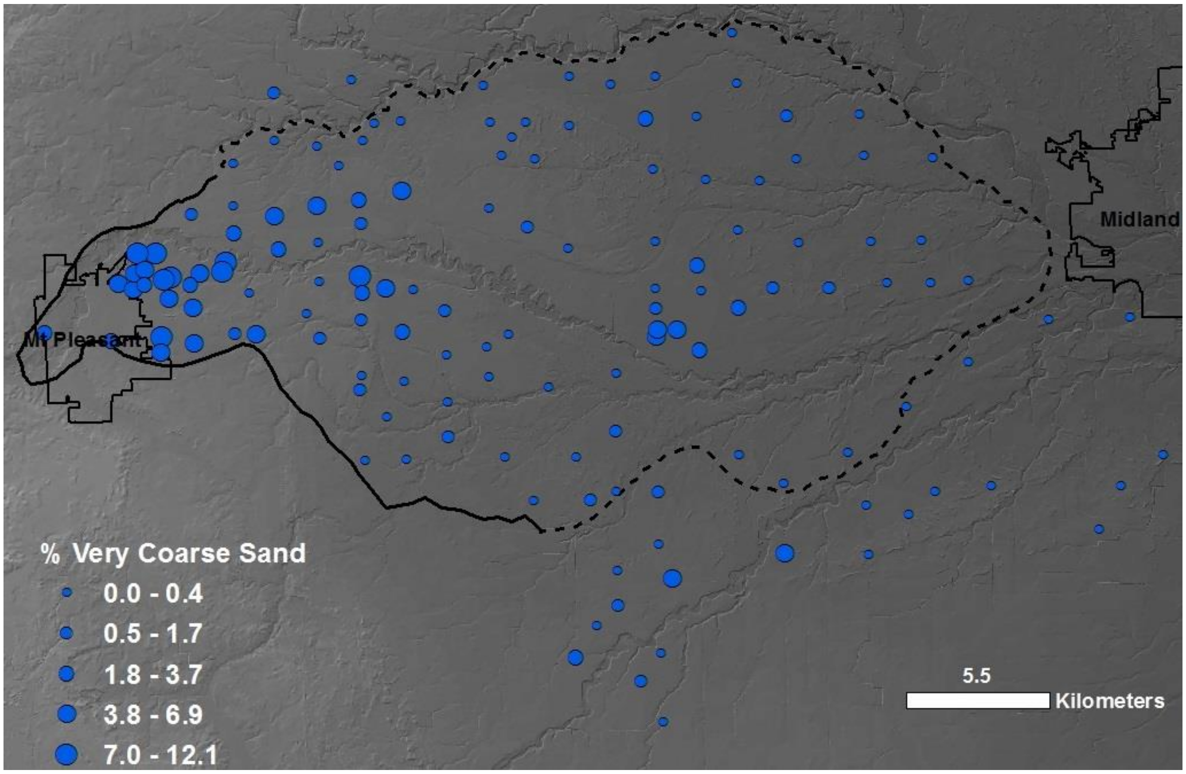


Figure 5.24 – Graduated circle map of very coarse sand (1000-2000μm) content for the 142 soil samples used in analysis/interpolation.

**5.6.1.4 Coarse Sand**

Sandy samples from the delta generally have more coarse sand than very coarse sand (Figs. 5.23, 5.24, 5.25, 5.26). Samples containing relatively high percentages (~30% or more) of coarse sand (CS, 500-1000μm) are located near the coarsest-textured areas, as previously indicated on maps of the texture class, primary mode, and percent gravel and VCS. Near the margins of the delta, coarse sand contents are generally lower (Figs. 5.25, 5.26).

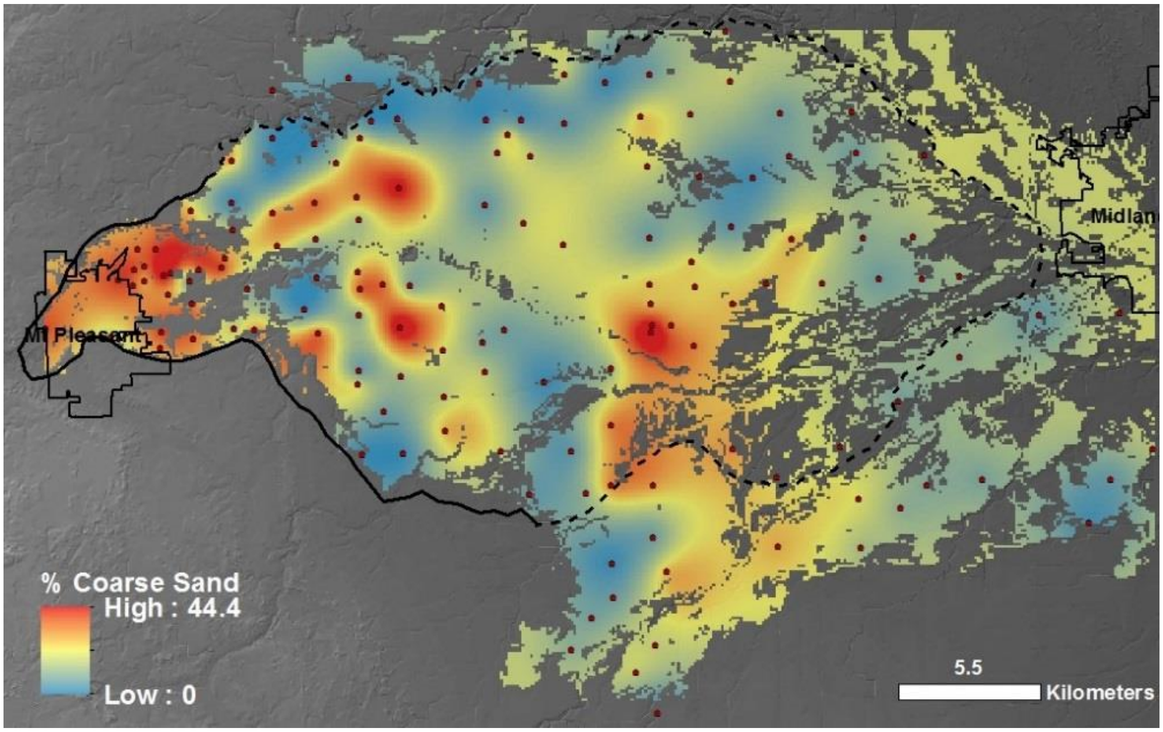


Figure 5.25 – Interpolated map of coarse sand (500-1000µm) content, clipped to areas with sand textured subsoil.

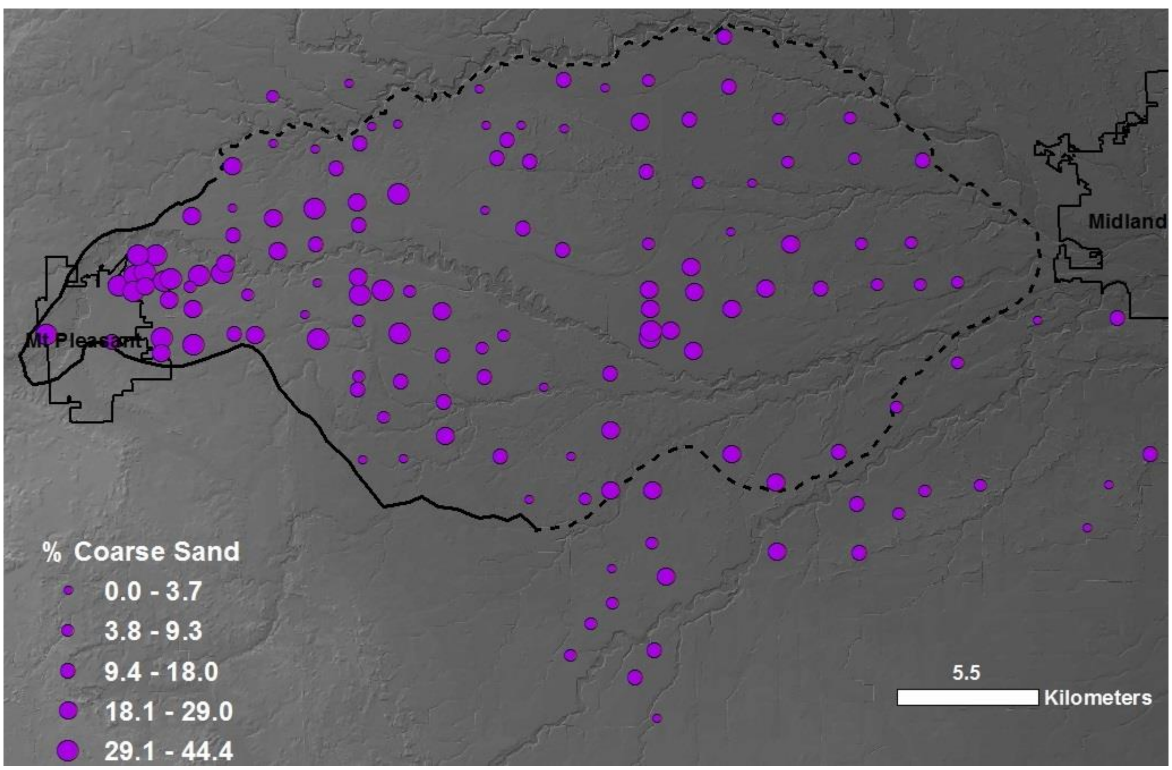


Figure 5.26 – Graduated circle map of coarse sand (500-1000µm) content for the 142 soil samples used in analysis/interpolation.

#### 5.6.1.5 Very Fine and Fine Sand

As a combined variable, the percentages of very fine and fine sand (FS and VFS, 50-250μm) are generally the highest around the northern and southern margins of the delta (Figs. 5.27, 5.28). Unsurprisingly, percentages of VF and FS are lowest in the previously mentioned coarse zones, and increase with distance from these areas. These general patterns conform to the “fining-outward” trend common to deltas (seen also in Figs. 5.17, 5.18, 5.19).

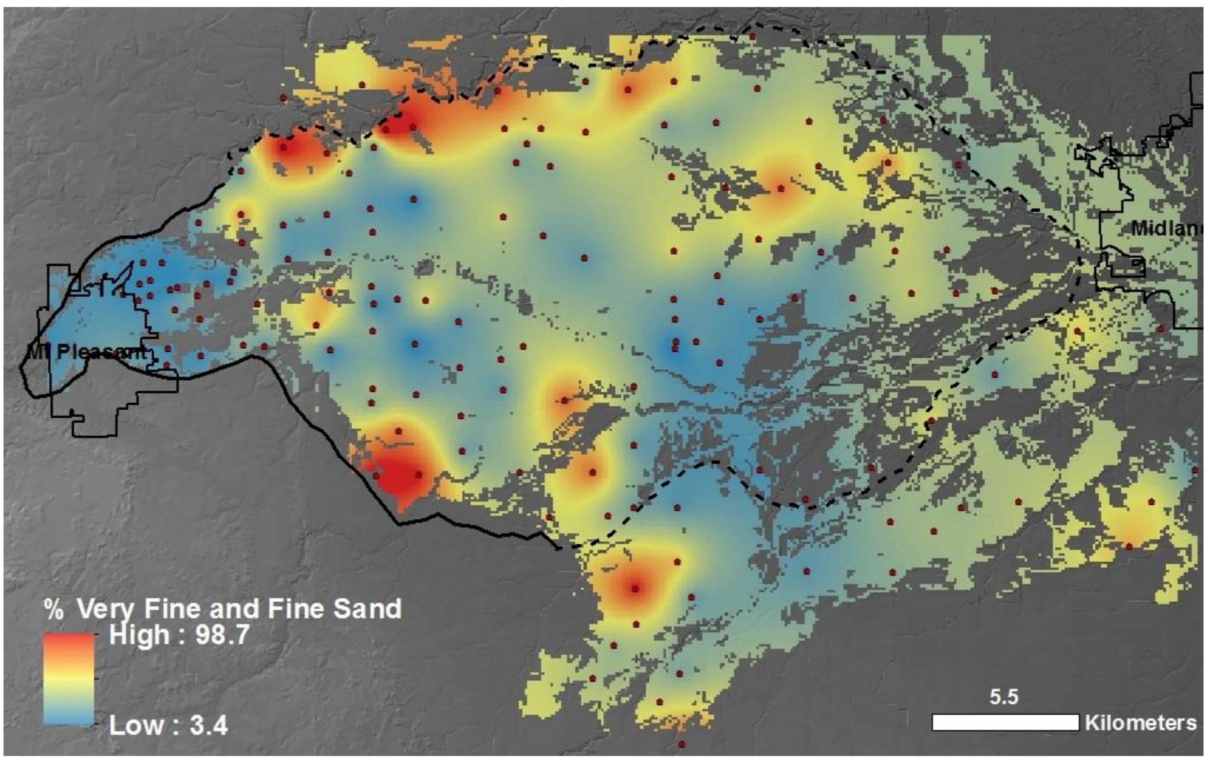



Figure 5.27 – Interpolated map of very fine and fine sand sand (50-250μm) content, clipped to areas with sand textured subsoil.

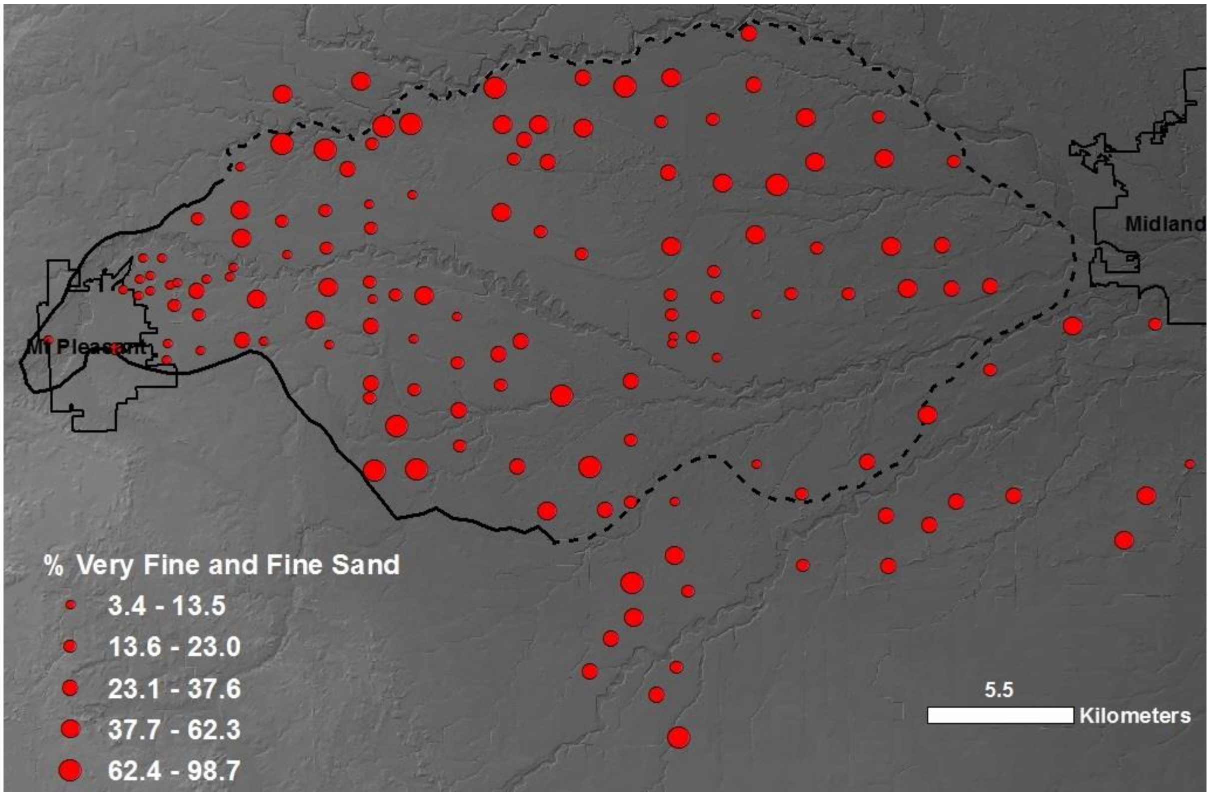



Figure 5.28 – Graduated circle map of fine and very fine sand (50-250µm) content for the 142 soil samples used in analysis/interpolation.

### 5.6.2 Loamy Samples From Till in the Chippewa River Watershed

**S**amples collected from till within the watershed are texturally variable, ranging from sand to clay texture classes, though the majority of samples are within loam classes (Fig 5.29). Most samples have bi-modal textural curves, with the primary mode in the silt fraction (Fig. 5.30). Sands deflated from till are not considered to have been a significant component of the delta sediments.

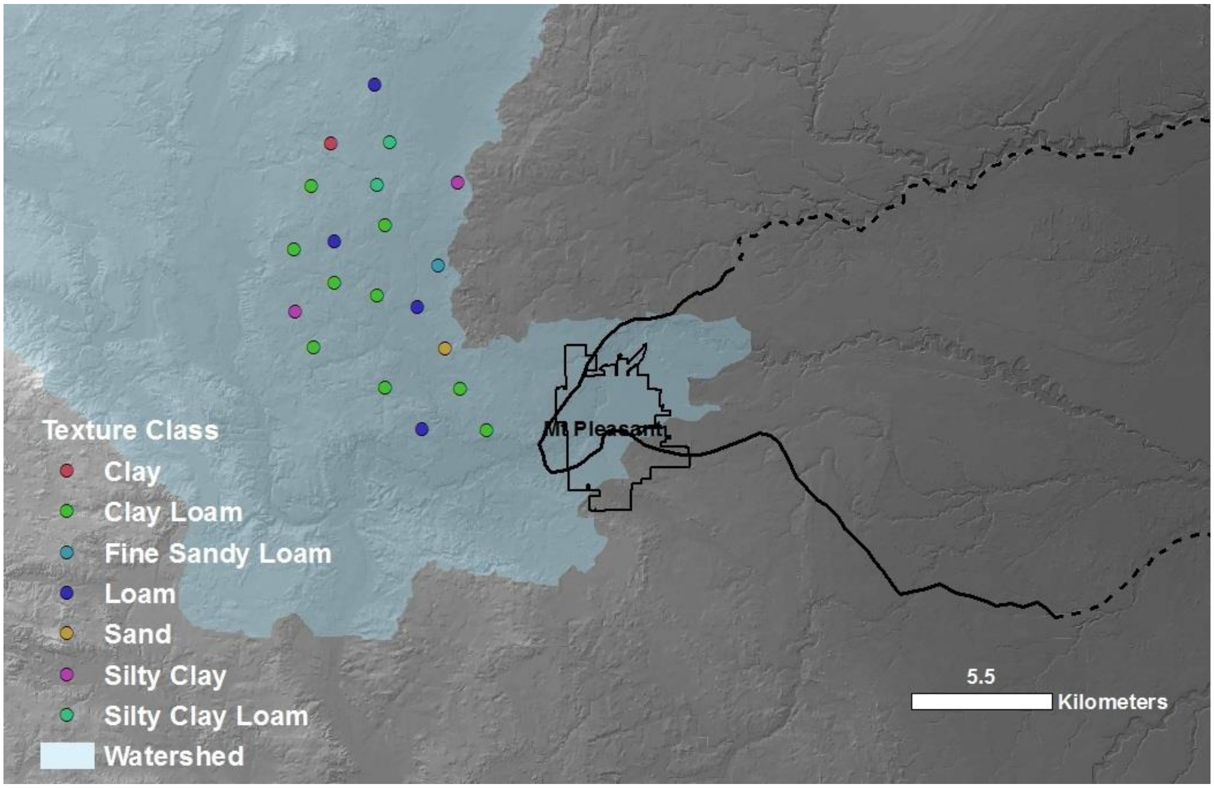



Figure 5.29 – The distribution of samples by texture class.

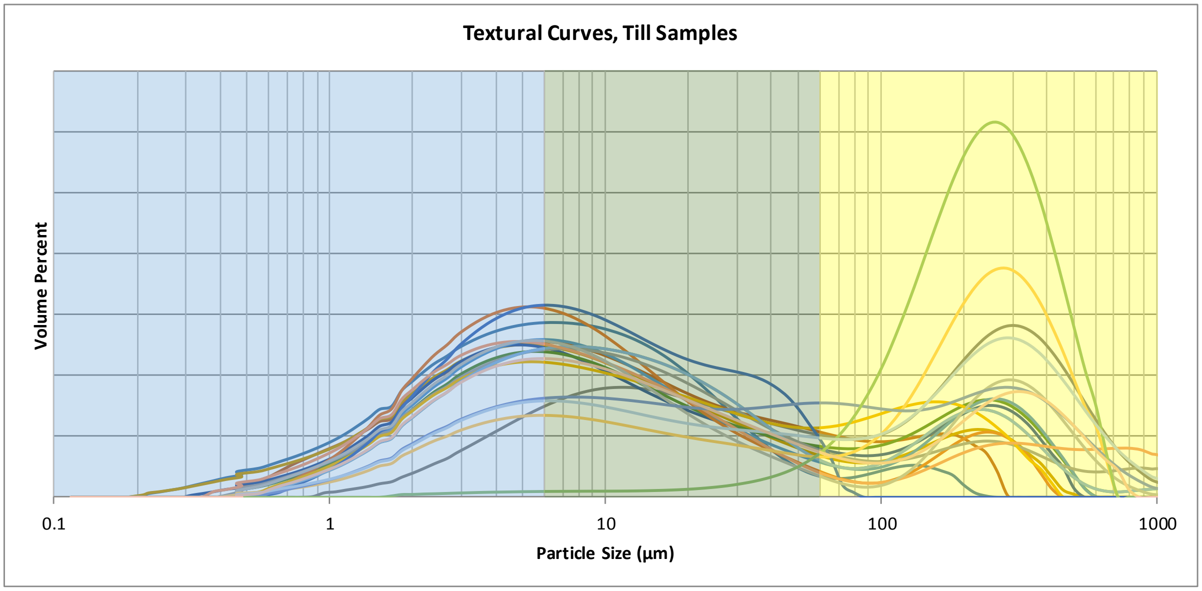



Figure 5.30 – Textural curves from field samples collected from till within the Chippewa River watershed.

### 5.7 Coarse Textured Zones on the Delta and Surrounding Landscape

After observing textural patterns in the field samples, I identified four distinct coarse textured “zones” on the delta and surrounding landscape (Fig. 5.31). These zones are areas of particularly coarse sediment, away from which sediment textures generally become finer.  As discussed, the “fining-outward” pattern, common in deltas (Vader et al., 2012), is an important spatio-textural trend to the interpretation of these data. Because the delta is a contiguous sand

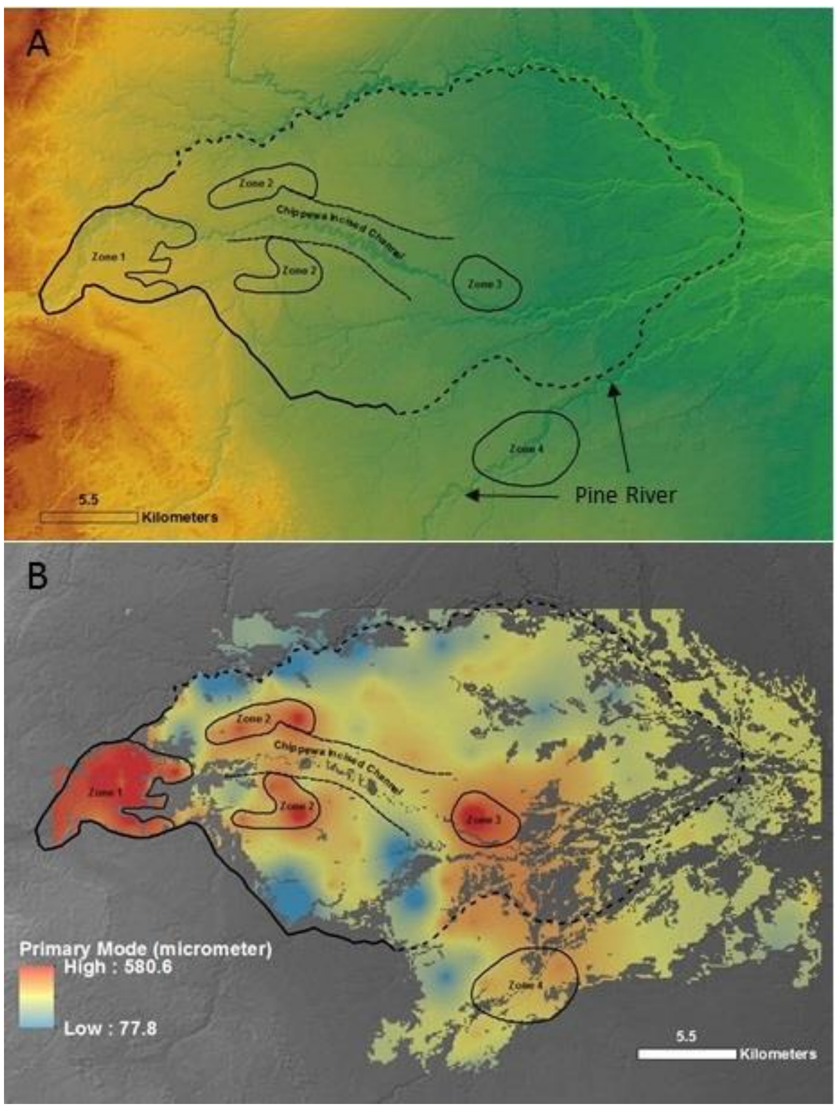



Figure 5.31 – A) Mapped coarse textured zones (from particle size analysis) over a hillshade DEM. B) Mapped coarse textured zones with the kriged map of particle size mode. Both maps also show the Chippewa Incised Channel and the delta outline.

body, identifying and mapping these textural sub-regions may inform interpretations of sedimentary processes that would have influenced delta development. Zones 1-3 are coincident with the Chippewa River, whereas Zone 4, outside of the Chippewa Delta boundary, is coincident with the Pine River which flows near the southern margin of the delta.

**5.7.1 Zone 1**

Zone 1 is in the uppermost portion of the delta, near the city of Mt. Pleasant, where the delta is comparatively narrow (Fig. 5.31). Sediments in Zone 1 are very coarse textured, with high amounts of gravel, VCS, and CS, and low amounts of FS and VFS. Modes in this zone are dominantly within the coarse sand fraction (Figs. 5.31, 5.32). In the NRCS soil data, Thetford soils, with coarse sand subsurface texture, are widely mapped here. Although Zone 1 is in the narrowest part of the delta, it is the largest (and coarsest textured) of the coarse zones.

**5.7.2 Zone 2**

Zone 2 is roughly 8 km to the east of Zone 1, where the delta width is much wider, and where delta sands are thicker (Fig. 5.31). Zone 2 sediments are generally coarse textured, but have markedly less gravel and VCS than do the sediments in Zone 1. Modes from sediments in Zone 2 are within the coarse or medium sand fraction, and there are high amounts of CS, and low amounts of FS and VFS (Figs. 5.31, 5.32). The Chippewa Incised Channel bisects Zone 2, separating it into two parts on opposite sides of the channel (Fig. 31). Because of the textural similarity within Zone 2 (low FS and VFS contents, relatively low gravel and VCS contents, high CS contents, and with textural modes of ~400-500μm) it is interpreted as representing a single textural sub-region. The channel later incised into, separating it into two parts.

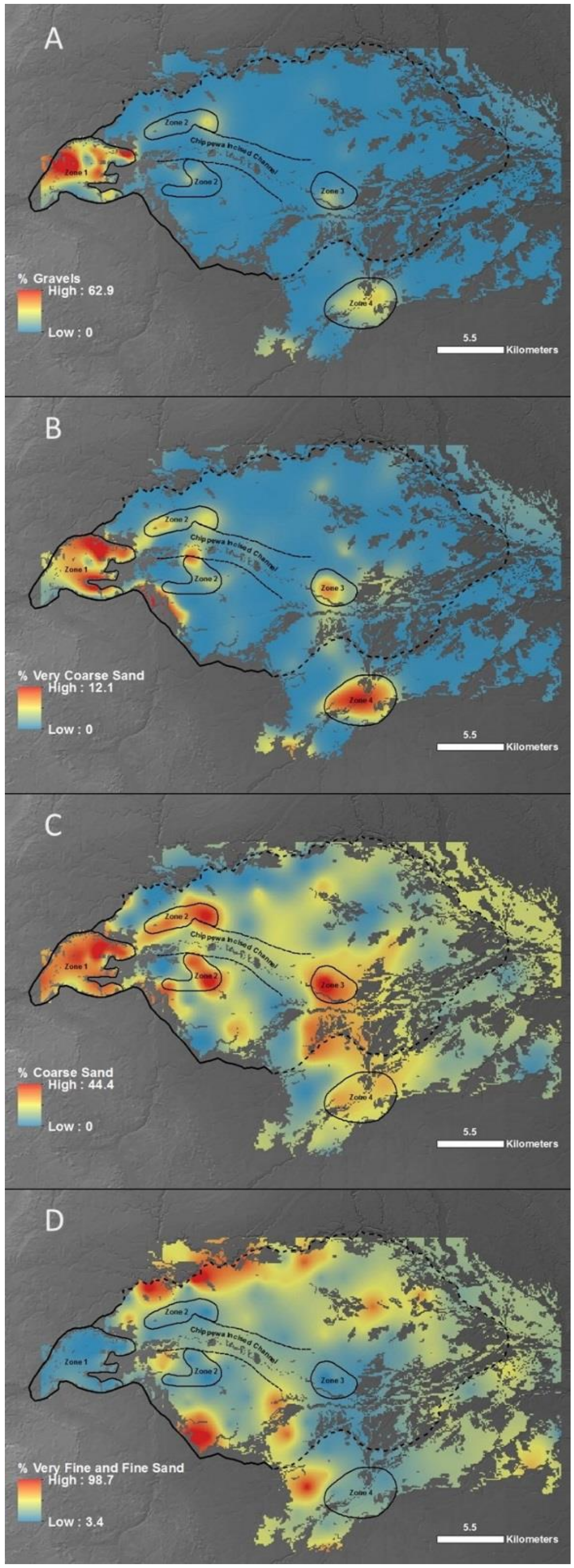



Figure 5.32 – Coarse Zones with krigs of A) gravels B) very coarse sand C) coarse sand and, D) very fine and fine sand.

**5.7.3 Zone 3**

Zone 3 is in the lower part of the delta, where delta sediments are generally thin, approximately 20km east of Zone 1 and 12km east of Zone 2 (Fig. 5.31). Zone 3 sediments (like Zone 2 sediments) have lower amounts of gravel and VCS than the sediments in Zone 1. Sediments in Zone 3 also have high amounts of CS, and low FS and VFS; modes from sediments in Zone 3 are within the coarse sand fraction (~400-500µm) (Figs. 5.31, 5.32). The Chippewa Incised Channel terminates ~2.5-3km west of this coarse zone.

**5.7.4 Zone 4**

Zone 4 is outside of the Chippewa Delta boundary, and is coincident with the northeastward flowing Pine River, not the Chippewa River (Fig. 5.31). The Pine River is roughly 166 km long, and drains a watershed of 800 $km^2$ in Mescota, Isabella, Montcalm and Gratiot Counties. Soils in the Pine River watershed, like those in the Chippewa River watershed, are texturally sandy and loamy. The mean discharge of the modern day Pine River ($9.1m^3/s$) is roughly equal to that of the modern day Chippewa River ($9.3m^3/s$). The Pine River does not flow across the Chippewa Delta proper; rather, it flows near, and roughly parallel, to the southern boundary of the delta (Fig. 5.31).

Sediments in Zone 4 are similar to those in Zone 1 in that they has high contents of gravel, VCS, and CS, and low FS and VFS contents; textural modes from sediments in Zone 4 are ~400µm (Figs. 5.31, 5.32). Because this zone is located adjacent to the Pine River, far from the modern day Chippewa River, it is interpreted as consisting of deltaic sediment transported and deposited into a separate delta, which I hereby name the Pine River Delta. Sediments from this delta merge with deltaic sediments associated with the Chippewa River. Sediments associated with the Pine River Delta were initially included in the Chippewa Delta, however upon further analysis, were determined to exhibit sufficient morphologic and spatio-textural characteristics to justify classifying them as part of a separate delta. Because the Pine River Delta was not a focus of this study, but was a feature discovered during the process of conducting this research, I have sufficient data to identify it, but not to completely map or characterize it.

**5.7.5 Summary of Coarse Textured Zones**

Overall, the coarsest parts of the delta, as measured by percentages of gravel, very coarse sand, and coarse sand, are in Zone 1. Zone 4, within the Pine River Delta, is a similar zone of coarse sand. Zone 1 is in the narrow “head” of the delta, while Zone 4 is in the southernmost part of the delta is coincident with the Pine River, removed from the present-day Chippewa River. Zones 2 and 3 sediments are coarse textured, but have lower percentages of gravel and very coarse sand than do sediments in Zones 1 and 4. Zone 2 is bisected by the Chippewa Incised Channel in the central part of the delta; Zone 2 is in the thick western part of the delta, whereas Zone 3 is in the thinner, eastern part of the delta.

**5.8 Coarse Textured Zones and Shorelines in the Study Area**

As discussed in the study area section, a number of paleolake stages occupied the Saginaw Lowlands. Between roughly 17,100 and 13,000 years ago, Lakes Saginaw, Arkona, and Warren inundated the study area, all of which had fluctuating lake levels, as indicated by their relict shorelines (Leverett and Taylor, 1915; Bretz, 1951; Kincare and Larson, 2009). Lake levels were controlled by downcutting in the Maple-Grand River valley, oscillations at the ice margin, and by the uncovering (and re-covering) of lower elevation outlets. Fluctuating lake levels may have had a significant impact on delta progradation, as discussed above, by moving the point at which the sediment-bearing Chippewa River would have discharged into the lake (discharge point). The coarsest sediments are deposited near the discharge point, while the finer sediments are transported and deposited farther out, onto the delta plain. The four coarse zones discussed in the previous section are coincident with the shorelines associated with some of the past lake stages (Fig. 5.33). The pattern of coarse textured loci coincident with paleoshoreline elevations implies that the delta progradation was tiered, prograding into the various lake stages over an extended period. To aid in visualizing the data with respect to the various shoreline elevations, I used the mapped shoreline features presented earlier in this

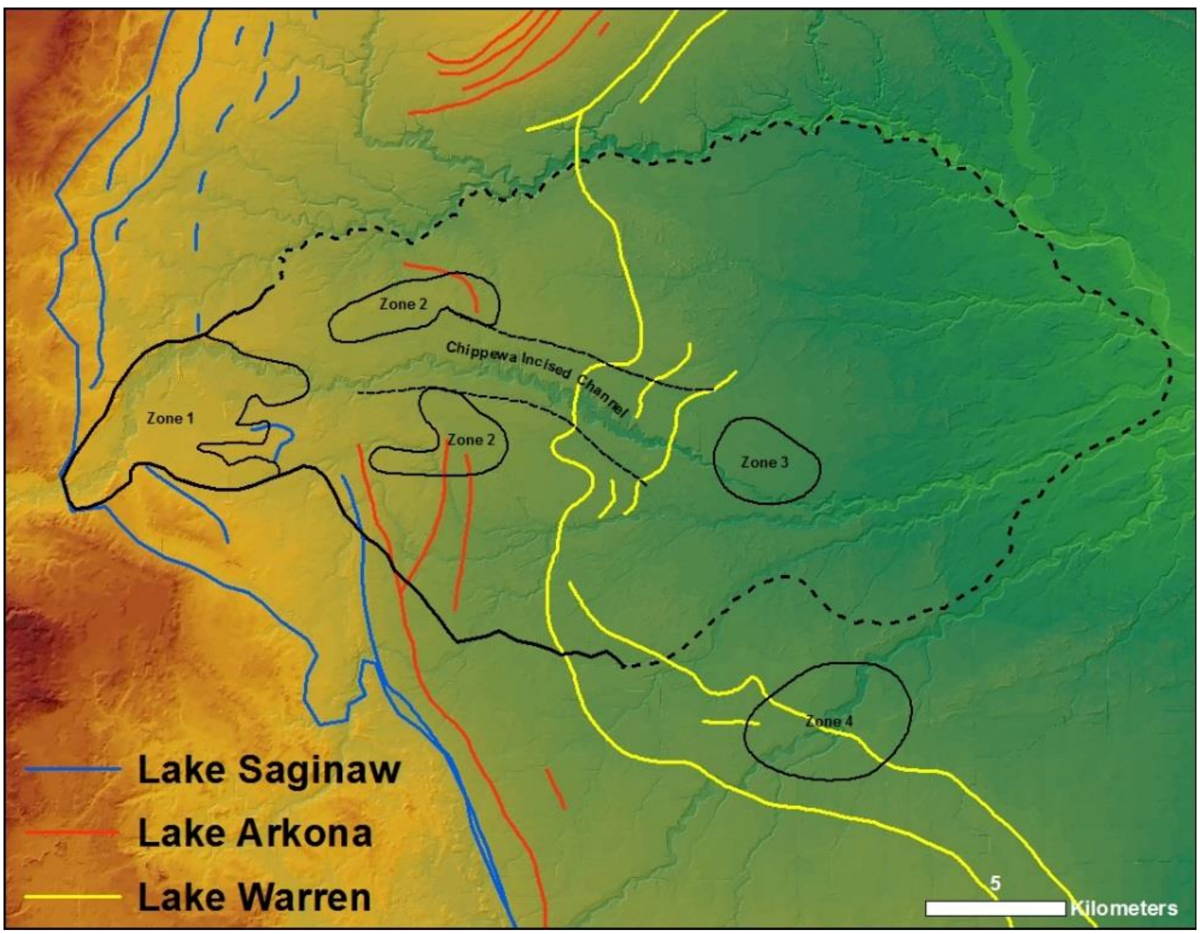



Figure 5.33 – Coarse textural zones, the Chippewa Incised Channel, and paleolake shoreline features. Also shown is the outline of the Chippewa Delta, over a hillshade DEM.

section (Fig. 5.1, 5.33) and connected the shoreline features which have roughly the same elevations (Fig. 5.34). The dashed lines roughly follow areas of equal elevation. This type of approach aids in visualizing which of the discontinuous shoreline features were formed at similar elevations, and thereby would likely have been formed by the same lake level(s).

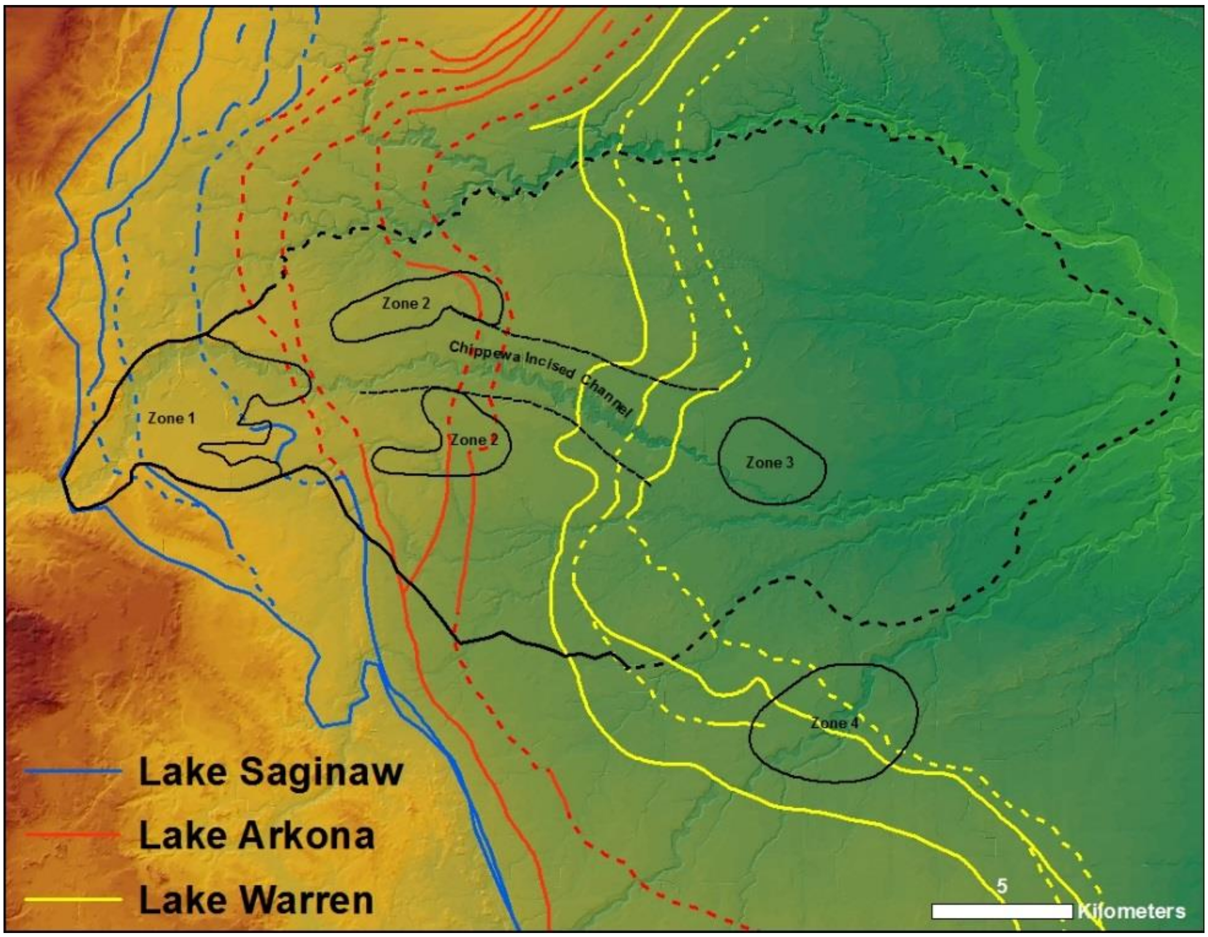



Figure 5.34 – Coarse textural zones, the Chippewa Incised Channel, and paleolake shoreline features with dashed lines connecting shorelines of roughly similar elevations. Also shown is the outline of the Chippewa Delta, over a hillshade DEM.

### 5.8.1 Lake Saginaw Shorelines and Zone 1

Zone 1 is generally coincident with Lake Saginaw shorelines, the oldest and highest elevation shorelines in the study area. In Zone 1, the delta is narrow, only about 4.5 km wide, moderately thick (~2.4m), and coarse textured, having the highest percentages of gravel and VCS. The Lake Saginaw shorelines represent both Early (and Later) Lake Saginaw, as discussed above. Multiple Lake Saginaw shoreline elevations are a result of incrementally lowering lake levels (pre-Lake Arkona) from downcutting in the westward draining outlet channel (Maple-Grand River). Therefore, I interpret Zone 1 sediments as coarse textured deltaic sediments deposited into Early (and/or) Later Lake Saginaw. The lowest elevation Lake Saginaw shoreline was also likely the highest Lake Arkona shoreline elevation.

**5.8.2 Lake Arkona Shorelines and Zone 2**

Zone 2 is generally coincident with Lake Arkona shorelines (Figs 5.35, 5.36), during which time the delta appears to have undergone significant widening. Zone 2 sediments are coarse textured, although not as coarse as are Zone 1 sediments. The delta reaches its maximum thickness in Zone 2, (mean ~4.2m), and is significantly thicker than other zones on the delta (Table 5.3). Lake Arkona (~16.5k YBP) sequentially followed (Early/Later) Lake Saginaw, and drained through the same outlet. As discussed in the study area section, the multiple levels of Lake Arkona are presumed the result of continued downcutting in the outlet channel, as well by as oscillations of the ice margin. Therefore, I interpret Zone 2 sediments as near-shore deltaic deposits associated with a major pulse in delta progradation during Lake Arkona time. At this time, the delta widened and thickened, likely incorporating sediments from the pre-existing delta that had originally been emplaced during Early/Later Lake Saginaw time.

Following Lake Arkona is a potential low lake level, Lake Ypsilanti, which would have drained the Saginaw Lowlands of proglacial lake waters and caused rapid incision into the delta. Following this possible low stand, the ice margin readvanced, establishing Glacial Lake Saginaw,

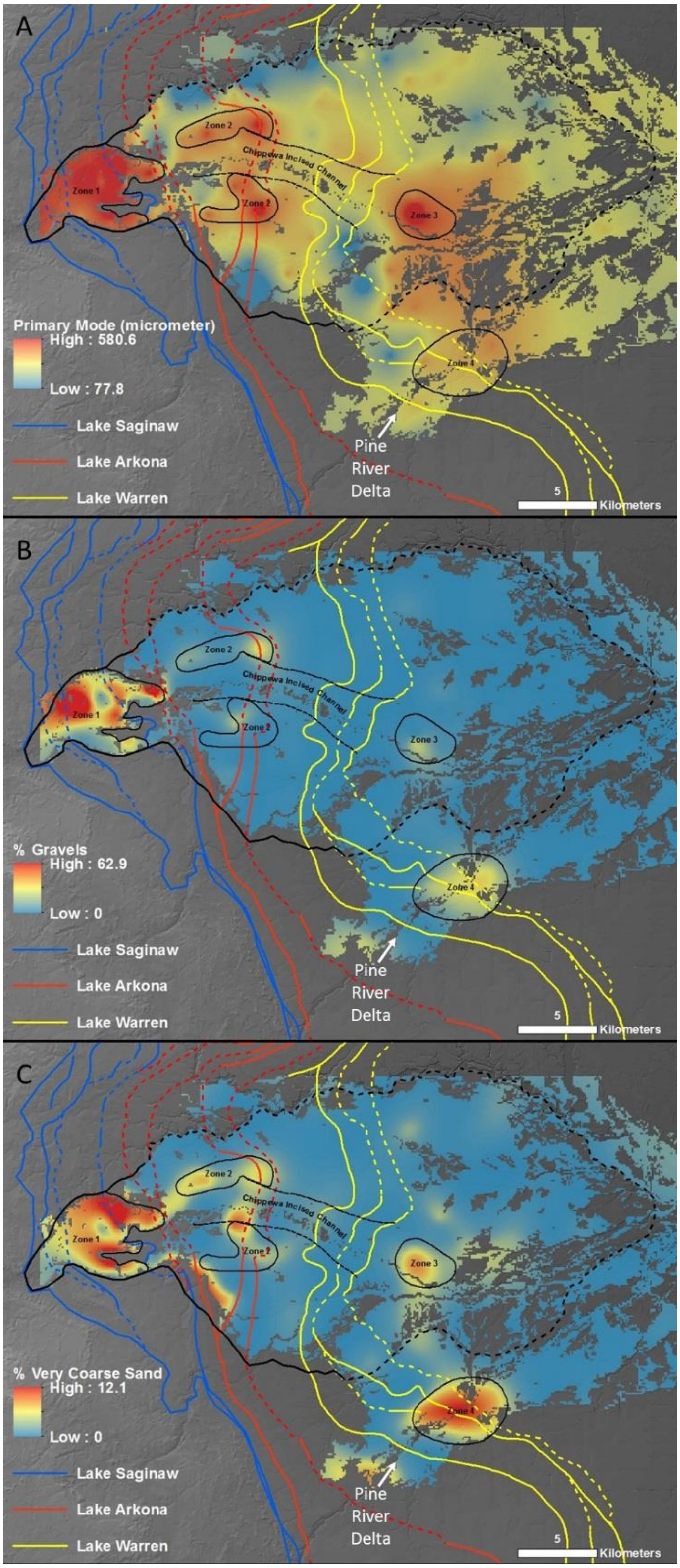



Figure 5.35 – Coarse textural zones overlain onto krigs of A) particle size mode B) gravels and, C) very coarse sand, with mapped, connected paleolake shorelines.

which likely would have been at roughly the same elevation of the lowest Lake Arkona level (Bretz, 1951; Monaghan and Hansel, 1990; Blewett and Winters, 1995; Krist and Lusch, 2004; Blewett et al., 2009; Kincare and Larson, 2009).

### 5.8.3 Lake Warren Shorelines and Zones 3 and 4

Zones 3 (on the Chippewa Delta) and 4 (on the Pine River Delta) are generally coincident with Lake Warren shorelines (Figs 5.35, 5.36). Zone 3 is coincident with the Chippewa River, whereas Zone 4 is near the Pine River, far from the Chippewa River, and outside of the Chippewa Delta boundary. Sediments in Zone 3 are coarse textured, generally resembling the textural properties associated with Zone 2 sediments, while Zone 4 sediments are texturally similar to those in Zone 1 (Table 5.3). Sediments in Zones 3 and 4 are relatively thin (mean ~1.7-1.8m), thinner than sediments in Zones 1 and 2, and thinner than the mean sediment thickness across the delta (~2.3m) (Table 5.3).

| | | **% Gravels** | **% VCS** | **%CS** | **%VF and FS** | **Thickness** |
|---|---|---|---|---|---|---|
| Zone 1 | Mean | 22.3 | 6.1 | 30.4 | 11.1 | 2.4m |
| | Median | 18.4 | 5.6 | 31.7 | 9.0 | |
| Zone 2 | Mean | 5.7 | 4.1 | 31.2 | 11.9 | 4.2m |
| | Median | 3.9 | 4.3 | 30.8 | 13.5 | |
| Zone 3 | Mean | 2.3 | 4.4 | 33.1 | 11.1 | 1.8m |
| | Median | 2.4 | 4.7 | 29.2 | 11.1 | |
| Zone 4 | Mean | 13.6 | 6.3 | 24.4 | 15.1 | 1.7m |
| | Median | 13.6 | 6.3 | 24.4 | 15.1 | |
| Background | Mean | 0.7 | 0.4 | 11.0 | 36.1 | 2.3m |
| | Median | 0 | 0.1 | 11.2 | 27.3 | |

Table 5.3 – Table of textural and thickness properties of sediments within the four zones, and the “background” textural and thickness properties on the delta that are outside of the zonal boundaries.

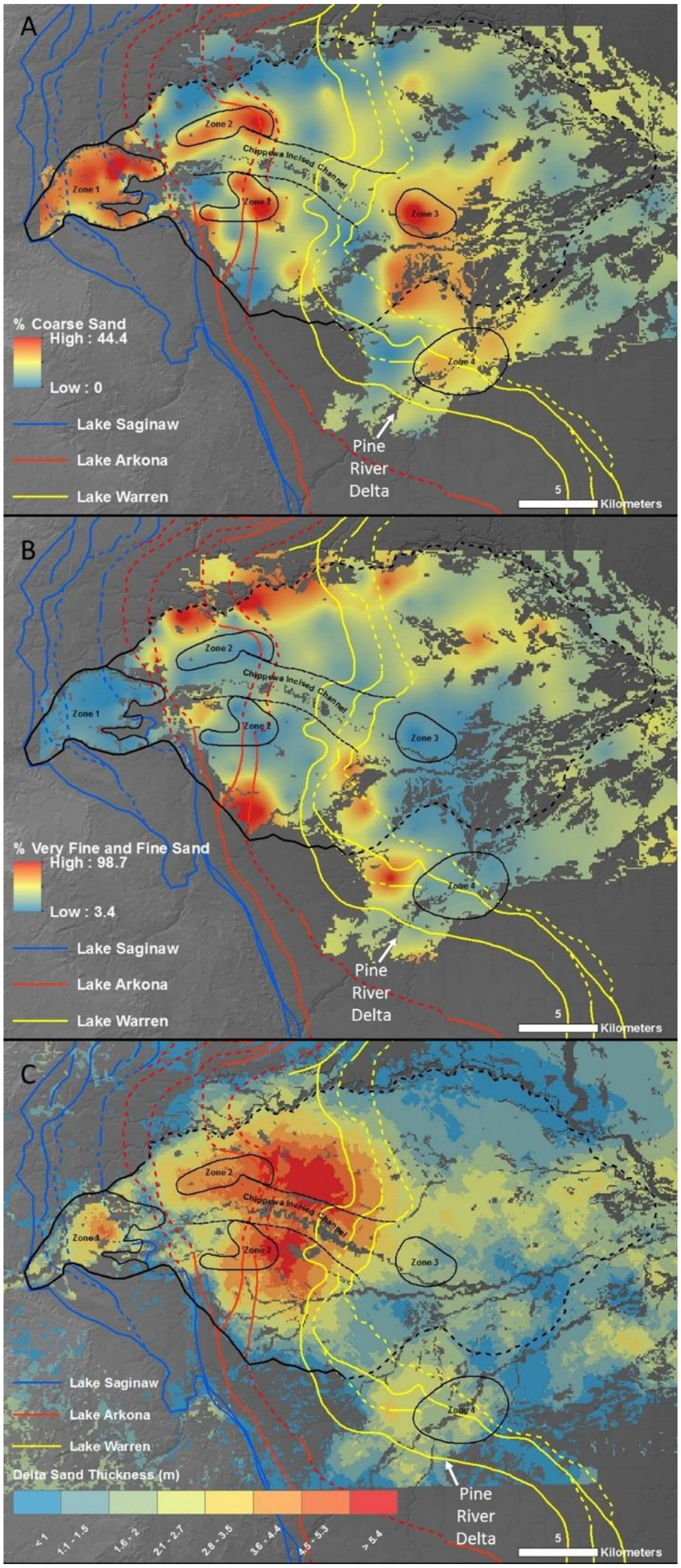



Figure 5.36 – Coarse textural zones overlain onto krigs of A) coarse sand B) very fine and fine sand and, C) thickness, with mapped, connected paleolake shorelines.

Lake Warren (~15k YBP), which sequentially followed Glacial Lake Saginaw, drained through the westward flowing Maple-Grand River Valley outlet. As discussed in the study area section, most sources cite at least two or three Lake Warren shoreline elevations (Bretz, 1951; Kunkle, 1963; Barnett, 1985, Kincare and Larson, 2009); in my study area I identified three (Fig. 5.33). Sediments in Zones 3 and 4 probably represent near-shore deltaic sediments deposited into Lake Warren. Zone 3 deposits were formed in association with the Chippewa River, whereas Zone 4 sediments are associated with the Pine River, and are a part of the Pine River Delta, which presumably formed in Lake Warren.

Between Zones 2 and Zone 3, adjacent to the Chippewa River, is the shallow Chippewa Incised Channel; it bisects Zone 2 sediments. This channel was likely cut as lake levels fell from Lake Arkona (and Glacial Lake Saginaw) elevations to Lake Warren, causing the upper part of the delta to become subaerial. Sediments in the upper part of the delta, eroded during channel incision at this time, could have been redeposited into Lake Warren. Sediments from the distal part of the upper delta were also likely eroded by wave action as lake levels fell.

An alternate hypothesis exists for the Chippewa Incised Channel: it is a channel-fill feature associated with the Lake Ypsilanti low stage. Evidence supporting the Ypsilanti low stage was first identified in the Lake Erie basin by Kunkle (1963), but had been previously suggested by Hough (1958) for the basins of Lakes Michigan and Huron. Evidence for the Ypsilanti low stage in the Saginaw Basin has never, however, been documented. During the Lake Ypsilanti low stage, the Saginaw Lowlands would have presumably been entirely subaerial, with the Lake Ypsilanti (and correlative) shorelines many meters below the elevation of the modern lakes. The low water levels would have caused rivers in the basin to incise, the rate and extent of which would have depended on stream power and erodibility of the sediment. Thus, it is possible that the Chippewa Incised Channel is a relict feature of such an incision into the sandy Chippewa Delta sediments. In this mode, the channel would have been initially cut into the delta by the Chippewa River, as a response to base level drop during the Ypsilanti low. The deeply incised channel may then have aggraded when lake levels rose back to the Glacial Lake Saginaw elevation, after the Port Huron advance. Following the Lake Saginaw stage, lake level in the Saginaw Basin dropped to the Lake Warren stage and the Chippewa River incised its channel into the aggradational deposits, making the floor of the Chippewa Incised Channel a fill terrace. The width of the Incised Channel may bear witness to the extreme base level fall from Lake Arkona to the Ypsilanti low stage.

Several mechanisms exist to determine, or infer, whether the Chippewa Incised Channel is a channel-fill feature, associated with the Ypsilanti low stage, or a surficial erosional feature associated with the transition from Lake Arkona to Lake Warren. If the channel is a channel-fill feature associated with the Ypsilanti low, then the Chippewa River deeply incised into the delta as a hydrologic response to base level drop, and the resulting channel aggraded when lake levels subsequently rose following the Port Huron advance. In this scenario, the spatio-textural trends within the channel would likely reflect the characteristics of sediments deposited during a transgression, or lake level rise, rather than a series of periodic regressions, as inferred for the overall spatio-textural trends on the Chippewa Delta. The textural trends within the channel, therefore, would likely exhibit a “fining-upward” character, also seen in stacked delta sequences. It is also possible that a channel cut feature may have been preserved below the channel-fill deposits.

If the channel is an erosional feature associated with incremental lake level drop from Arkona to Warren elevations, then the Chippewa River only shallowly incised into deltaic sediments, likely in a wide braided stream system. In this scenario, there would be no deep channel cut, nor fill deposits in the channel. Spatio-textural trends within the channel would likely be consistent with sandy braided stream sediments, and evidence of paleochannels may have been preserved. In this thesis, I did not obtain any of these kinds of data, limiting my ability to make specific inferences about the origin of the Chippewa Incised Channel.

**5.9 Saginaw-Arkona Transition Zone**

Near the lowest Lake Saginaw and highest Lake Arkona shorelines, directly east of textural Zone 1, is an area of thin (Fig. 5.13), comparatively fine textured sediments (Fig. 5.32). The soils in this area are largely two-storied (Fig. 5.10); they have a thin sandy upper sola over a loamy subsoil. These thin, sandy sediments lie adjacent to the texturally coarsest zone on the delta and in between the thickest areas of the delta, implying that they were subjected to erosion, or sediment reworking. Erosion or reworking of these sediments likely occurred after the Lake Saginaw portion of the delta was emplaced, and before or concurrent with the major pulse of progradation into Lake Arkona. If these sediments were originally deltaic, they would be expected to gradually thin and texturally "fine" with distance from the former shorezone, instead they abruptly "fine" and thin, suggesting that they have been reworked. Sediments eroded from this zone, hereafter referred to as the Saginaw-Arkona Transition Zone (SATZ, Fig. 5.37), would likely have been redeposited into the delta during Lake Arkona time. The SATZ therefore represents a shorezone during the transition between lowest Lake Saginaw and highest Arkona, and the coarse sediments in Zone 1 may represent erosion resistant lag deposits in the upper delta.

**5.10 Upper and Lower Chippewa Delta**

Based on textural and thickness patterns, and associated lake stages, the Chippewa Delta can be divided into upper and lower delta deposits (Fig. 5.38). The Upper Chippewa Delta, partially mapped by Martin (1955), includes Zones 1 and 2, the Chippewa Incised Channel and SATZ; sediments in the upper delta are generally coarse textured and comparatively thick (Table 5.4). The Upper Chippewa Delta formed within Lakes Saginaw and Arkona. The Lower

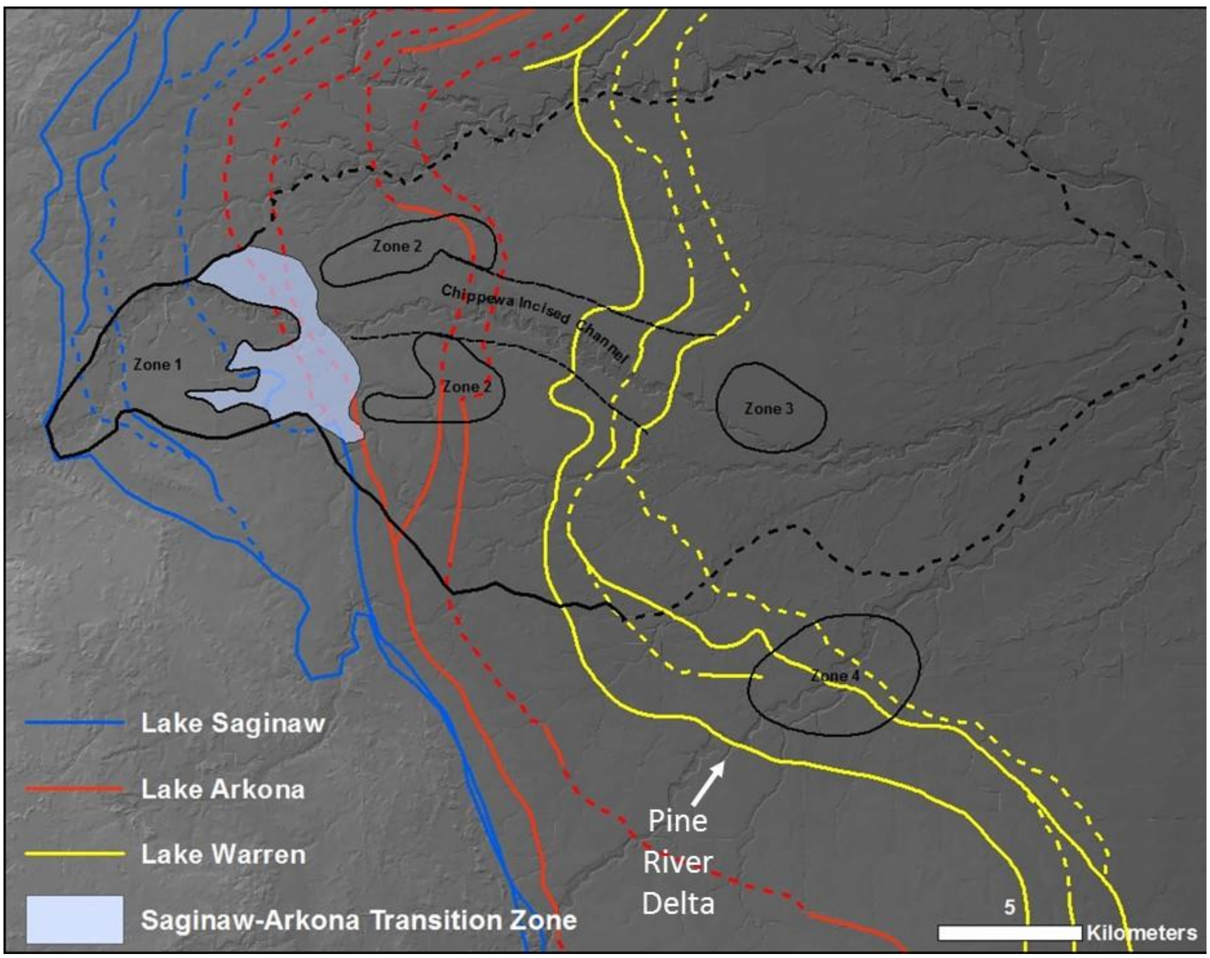



Figure 5.37 – The Saginaw-Arkona Transition Zone (SATZ), with coarse zones, paleolake shorelines, and the outline of the delta. The SATZ is generally coincident with lowest Lake Saginaw and highest Lake Arkona shoreline elevations.

Chippewa Delta includes Zone 3, and consists of delta sediments to the east of the Upper Delta. The Lower Chippewa Delta deposits are thinner than those in the upper delta, and textures, although still sandy, are generally finer (Table 5.4). Much of the sediment in the lower delta was likely eroded from the upper delta and redeposited into Lake Warren. Alternatively, deltaic sediments may have originally been emplaced during higher lake levels (Saginaw/Arkona), and were subsequently reworked by wave action as lake levels lowered. Sediments on the southern margin of the Lower Chippewa Delta merge with those of the Pine River Delta, obscuring the boundary between the two deltas.

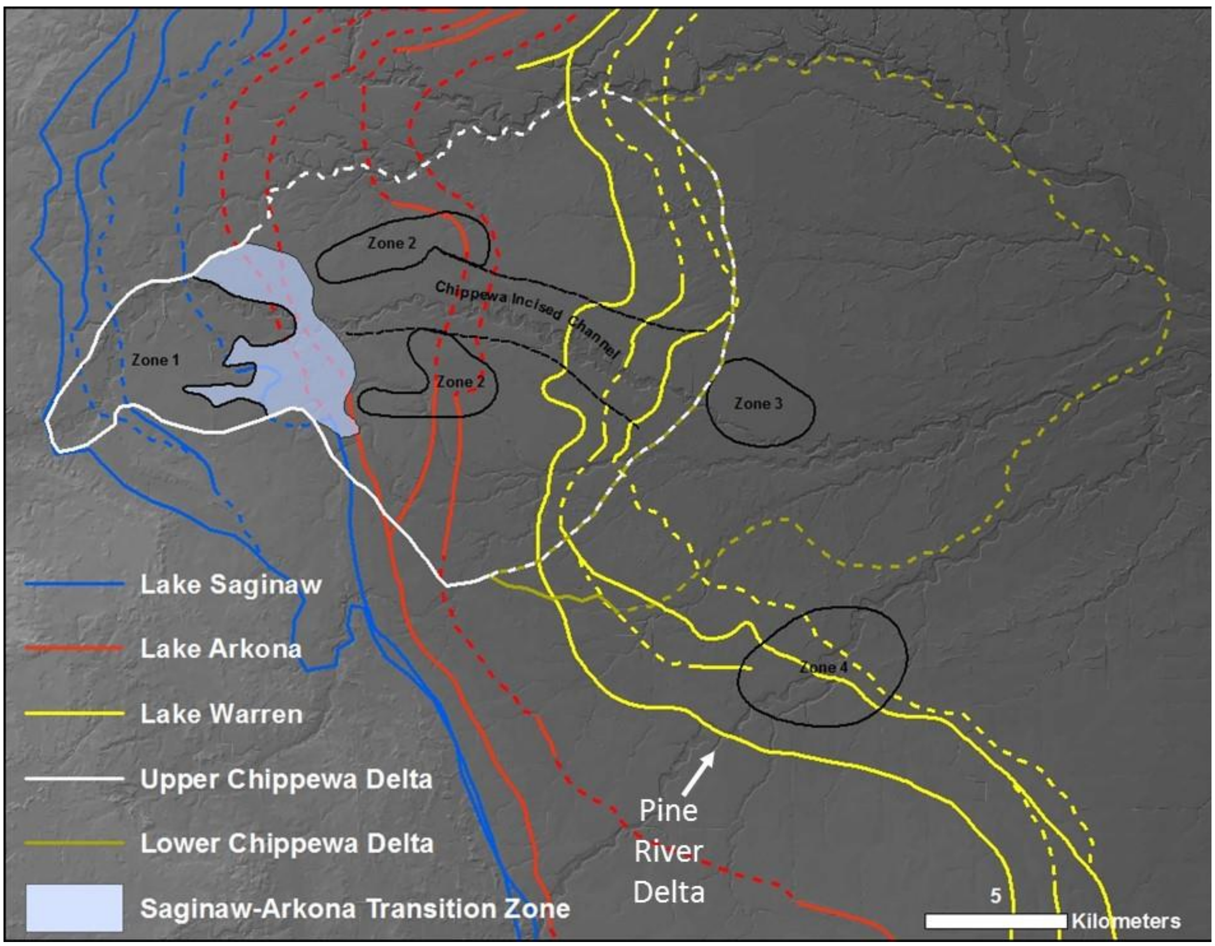


Figure 5.38 – Outlines of the upper and lower delta, the Saginaw-Arkona Transition Zone, with coarse zones, and paleolake shorelines.

| | Upper Delta | Lower Delta |
|---|---|---|
| Thickness (m) | Mean: 3.2, Std. Dev: 1.3 | Mean: 1.5, Std. Dev: 0.48 |
| % Gravels | Mean: 3.8, Std. Dev: 9.2 | Mean: 0.58, Std. Dev: 1.5 |
| % VCS | Mean: 1.4, Std. Dev: 2.3 | Mean: 0.65, Std. Dev: 1.1 |
| % CS | Mean: 14.7, Std. Dev: 9.5 | Mean: 13.6, Std. Dev: 7.7 |
| % VF & FS | Mean: 31.1, Std. Dev: 16.2 | Mean: 31.1, Std. Dev: 12.2 |
| Particle Size Mode (µm) | Mean: 326, Std. Dev: 69.6 | Mean: 319, Std. Dev: 45.0 |

Table 5.4 – Table of textural and thickness values for the Upper and Lower Chippewa Delta.

**5.11 Summary of Delta Progradation**

The Chippewa Delta began prograding into Early Lake Saginaw (17.1k – 16.5k YBP), and continued as lake levels fell due to downcutting in the outlet channel (Fig 5.39). Sediments in Textural Zone 1 are deposits from this phase of progradation. The distal margin of this phase of progradation is not known, as sediments were likely reworked and incorporated into subsequent pulses of deltaic deposition.

A major pulse of progradation began as Early/Later Lake Saginaw transitioned into Lake Arkona (16.5k YBP; Fig. 5.40), contemporaneous with a significant phase of ice retreat, which uncovered the “thumb” and merged the lakes in the Saginaw and Erie basins. Sediments in Textural Zone 1 were subaerial at this time, while sediments from Textural Zone 2 represent shorezone deposits from this phase of progradation. The Chippewa Incised Channel may have formed in response to falling lake levels, specifically as lake levels fell from Arkona elevations to Warren elevations. Alternately, it’s possible that the Chippewa Incised Channel represents a relict channel-fill feature associated with a possible Ypsilanti low stage erosional event, which was subsequently filled when lake levels rose to the Glacial Lake Saginaw elevation.

By the time that Lake Warren was established (15k YBP), the delta was in its terminal phase of progradation. At this time, water would have been routed through the Chippewa Incised Channel, eroding sediment from the subaerial upper delta, and depositing sediment into Lake Warren (Textural Zone 3; Fig. 5.41). Wave action was also likely deflating sediment from the upper delta, before it fully stabilized, while planing and reworking sediments in the lower delta plain. Lake levels dropped eventually leaving the delta almost entirely subaerial by the time that Lake Wayne (and/or Elkton) was established (13k YBP; Fig. 5.41).

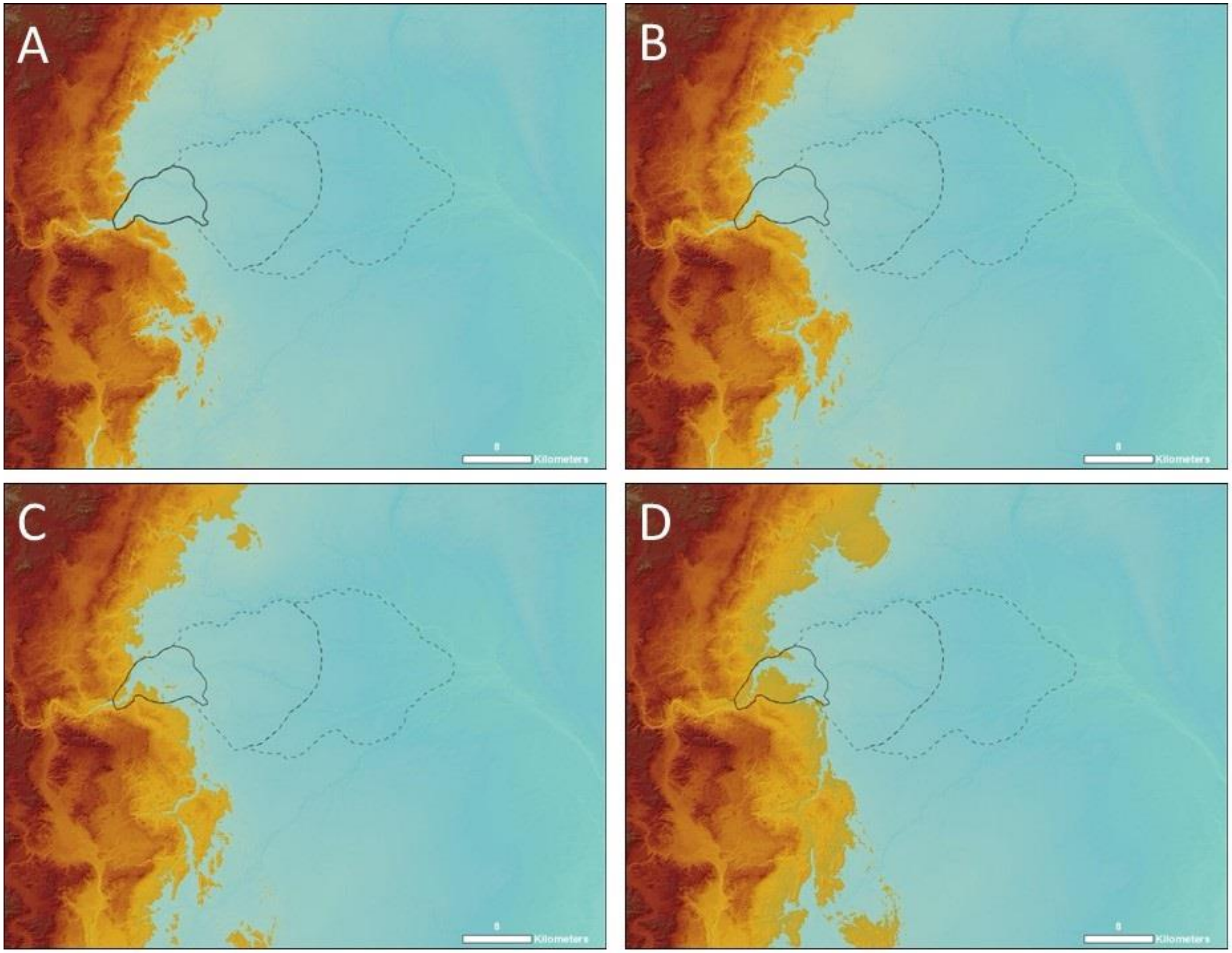



Figure 5.39 – Flooded DEM graded to Early and Later Lake Saginaw shorelines. Lake levels fell due to downcutting in the outlet channel. The active part of the delta at this time is depicted with a solid line. The rest of the upper and lower delta, not active at this time, are depicted with a dashed line.

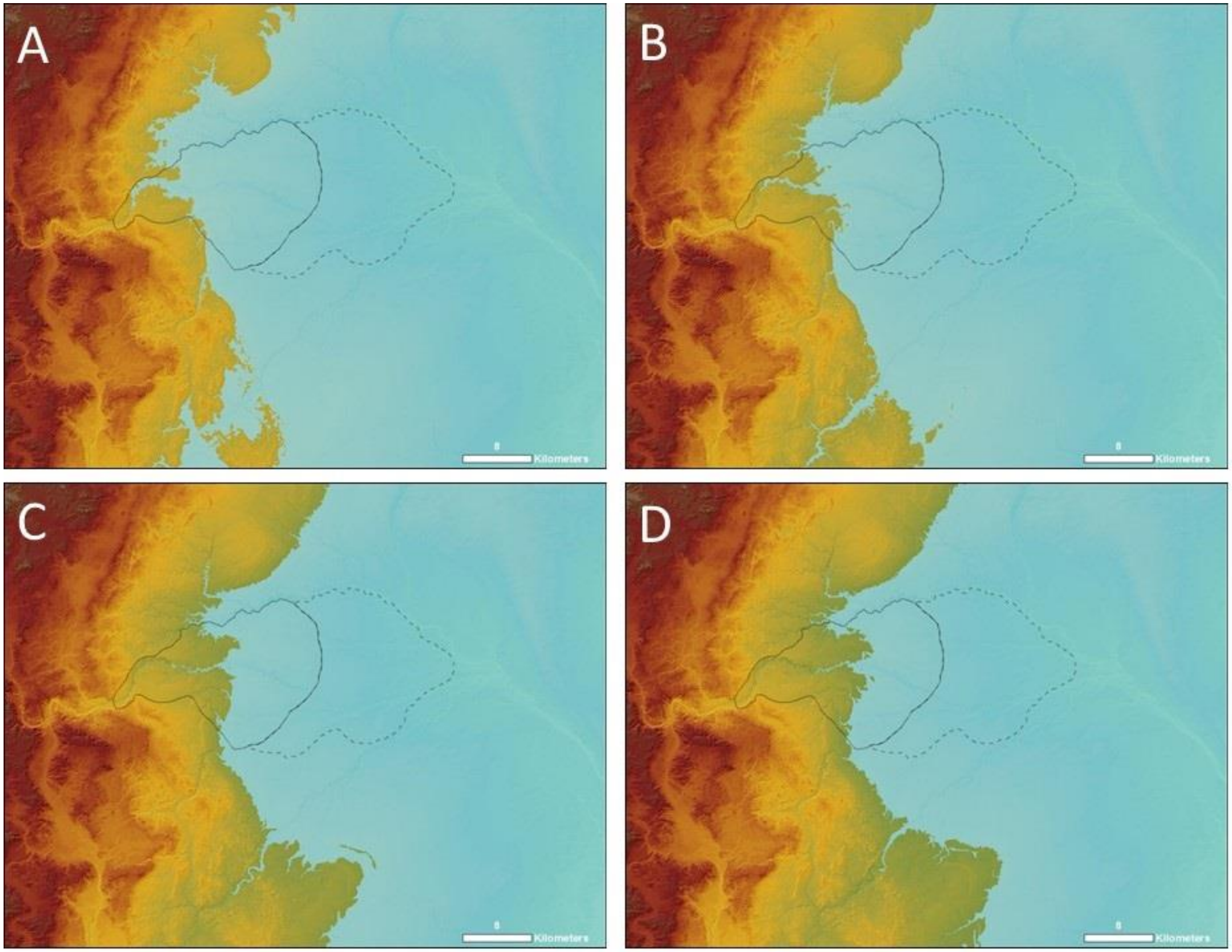



Figure 5.40 – Flooded DEM graded to (A) Lake Saginaw/Arkona transition and (B-D) Lake Arkona shorelines. Lake levels fell due to downcutting in the outlet channel and oscillations at the ice margin. The active part of the delta (upper delta) at this time are depicted with a solid line. The lower delta, not active at this time, is depicted with a dashed line. Glacial Lake Saginaw would also have been at the lowest Lake Arkona elevation (D).

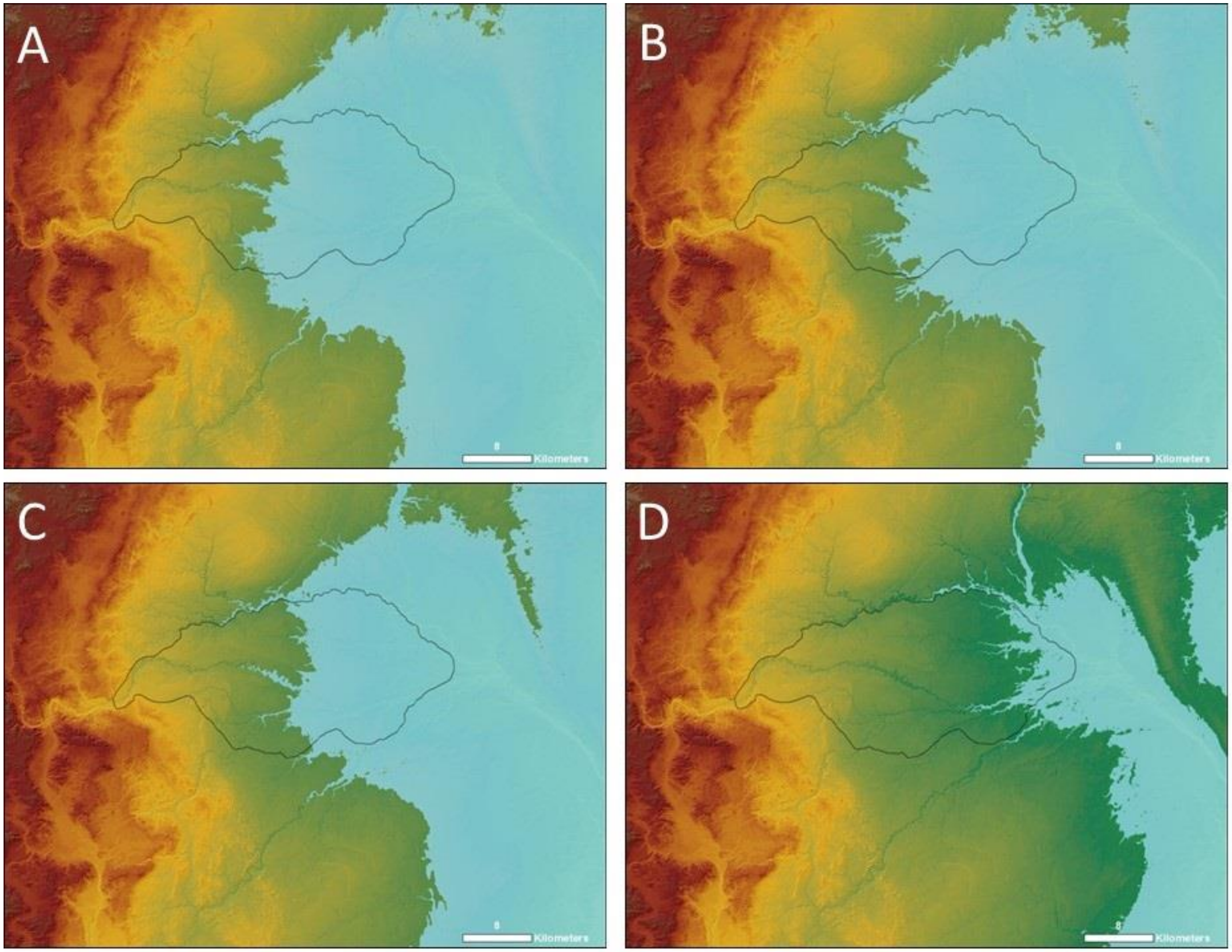



Figure 5.41 – Flooded DEM graded to (A-C) Lake Warren (D) Lake Wayne/Elkton shorelines. During Lake Warren time, the upper delta was subaerial, and the lower delta was active. The delta became inactive and nearly completely subaerial by the time of Lake Wayne/Elkton.

## 6. Conclusions

Many relict lacustrine deltas in Michigan can be mapped from patterns in NRCS soil data. However, very few of the NRCS OSDs cite “delta sediments” as soil parent materials. The language that mappers typically use for these soils is general and often generic, and does not imply a genetic origin for the sediments that comprise the parent materials of the soils. Texture data within the NRCS dataset, however, have proven useful for identifying and mapping relict lacustrine deltas, because patterns in soil texture can inform interpretations of sedimentary origin, as demonstrated in this research. Similarly, well log stratigraphic data, although not quality checked for use in geologic research, can nonetheless serve as proxy data for approximating the thickness trends across landforms and landscapes.

Textural data from field samples can add an important level of detail to the more general textural trends observed in the NRCS dataset. My field sample data allowed me to interpret the various stages of delta progradation with respect to the sequence of lake levels in the basin; without these data such an interpretation would have been impossible. Overall, the combination of these datasets, and the specifically tailored methodology focusing on spatial data, allowed for a robust and detailed characterization of the Chippewa Delta in southern Michigan. This research represents the first use of these methods for the geomorphic study of a relict delta.

A major unanswered question, accentuated by the results of this study, is the specific source of the large, rapidly flowing volumes of water that would have been necessary to move the sediment to the Chippewa Delta. In general, it can be assumed that glacial meltwater transported and emplaced the delta sediment, but because so little research into the glacial origins of the central part of the Lower Peninsula has been conducted, only inferences about meltwater dynamics in that region can be made.

Additional targets for future research in the region include the Pine River Delta, discovered during the course of this research, investigating the relationship between the multiple drainages that converge in the study area (Pine, Salt, and Tittabawasee Rivers) and the expansive distribution of sand in the region that fell outside of the Chippewa Delta boundary. It seems likely that, like with the Chippewa and Pine River Deltas, multiple streams deposited deltaic sediment in the basin over similar periods of time, and the sediments merged, obscuring the boundaries between each landform. Further refinement of the sources for this widespread area of sand may help our understanding of paleoenvironmental conditions across the Saginaw Lowlands in general.

Further study of the Chippewa Incised Channel could also yield the first evidence of the effect of the Ypsilanti low stage within the Saginaw Basin. If the Chippewa Incised Channel is a channel-fill feature, cut during the Ypsilanti low, the channel cut may have been preserved in the sedimentary record after it was subsequently filled upon lake level rise to Glacial Lake Saginaw. The contact between the channel wall (consisting of deltaic sediments) and the channel fill sediments (likely dominated by arcuate cut-and-fill structures) could be located using ground penetrating radar, or seismic surveys. Definitively identifying the Chippewa Incised Channel as a Lake Ypsilanti channel-fill feature would be a significant contribution to our understanding of the glacial history of the Saginaw Lowlands, and would confirm the widespread existence of this low stage.

Finally, this research has three main conclusions/contributions. First, it shows that the Chippewa Delta exists and was emplaced over an extended period of time, into multiple lake stages in the Saginaw Basin. The deltaic sediments preserved textural evidence of basinward forced regression, as well as evidence of a major pulse of progradation during Lake Arkona time. This study also demonstrates the efficacy of the methods used in this research to conduct detailed studies of glacial landforms using soil, spatio-textural, and well log data. Lastly, this study contributes a refined paleolake shoreline map for the study area.

# REFERENCES

## REFERENCES


Adams, E.W., Schlager, W., and F.S. Anselmetti. Morphology and curvature of delta slopes in Swiss lakes: lessons for the interpretation of clinoforms in seismic data. Sedimentology 48, pp. 661-679.

Arbogast, A.F., Scull, P., Schaetzl, R.J., Harrison, J., Jameson, T.P., and S. Crozier. 1997. Concurrent Stabilization of Some Interior Dune Fields in Michigan. Physical Geography 18, pp. 63-79.

Barnett, P.J. 1985. Glacial retreat and lake levels, north central Lake Erie Basin, Ontario, in Karrow, P.F. and P.E. Calkin (Eds.), Quaternary Evolution of the Great Lakes, Geological Association of Canada Special Paper 30, pp. 185-194.

Bellotti, P., Chiocci, F.L., Milli, S., Tortora, P., and P. Valeri. Sequence stratigraphy and depositional settings of the Tiber Delta: Integration of high-resolution seismics, well logs, and archeological data. Journal of Sedimentary Research 64, pp. 416-432.

Bettis, E.A., Muhs, D.R., Roberts, H.M., and A.G. Wintle. 2003. Last Glacial loess in the conterminous USA. Quaternary Science Reviews 22, pp. 1907-1946.

Bhattacharya, J.P. and L. Goisan. 2003. Wave-influenced deltas: geomorphological implications for facies reconstruction. Sedimentology 50, pp. 187-210.

Blewett, W.L., and H.A. Winters. 1995. The importance of glaciofluvial features within Michigan's Port Huron moraine. Annals of the Association of American Geographers 85, pp. 306-319.

Blewett, W.L., Lusch, D.P. and R.J. Schaetzl. 2009. The Physical Landscape: A Glacial Legacy. In Schaetzl, R.J, Darden, J. and D. Brandt (Eds.), Michigan Geography and Geology, Pearson Custom Publishing, New York, NY, pp. 249-260.

Boggs, Sam Jr. 2001. Principles of Sedimentology and Stratigraphy, Third Edition. Prentice-Hall Upper Saddle River, NJ.

Bretz, J.H. 1951. Causes of the Glacial Lake Stages in Saginaw Basin, Michigan. The Journal of Geology 59 pp. 224-258.

Brooks, G.R., Kindinger, J.L., Penland, S., Williams, S.J., and R.A. McBride. 1994. East Louisiana Continental Shelf Sediments: A Product of Delta Reworking. Journal of Coastal Research 11, pp. 1026-1036.

Burgis, W.A. 1970. The Imlay Outlet of Glacial Lake Maumee, Imlay City, Michigan. M.S. Thesis. University of Michigan, Ann Arbor, MI.

Calkin, P.E. and B.H. Feenstra. 1985. Evolution of the Erie-Basin Great Lakes. In P.F. Karrow and P.E. Calkin (Eds.), Quaternary evolution of the Great Lakes. Geological Survey of Canada Special Paper 30, pp. 149-170.

Carrivick, J.L., Tweed, F.S. 2013. Proglacial lakes: character, behaviour and geological importance. Quaternary Science Reviews 78, pp. 34-52.

Comer, P.J., Albert, D.A., Wells, H.A., Hart, B.L., Raab, J.B., Price, D.L., Kashian, D.M., Corner, R.A. and D.W. Schuen (map interpretation); Leibfreid, T.R., Austin, M.B., DeLain, C.J., Prange-Gregory, L., Scrimger, L.J., Korroch, K.M., and J.G. Spitzley (digital map production). 1995. Michigan's Presettlement Vegetation, as Interpreted from the General Land Office Surveys 1816-1856. Michigan Natural Features Inventory, Lansing, MI.

Drzyzga, S.A., Shortridge, A.M., and R.J. Schaetzl. Mapping the phases of Glacial Lake Algonquin in the upper Great Lakes region, Canada and USA, using a geostatistical isostatic rebound model. Journal of Paleolimnology 47, pp. 357-371.

Elliot, T. 1978. Deltas. In H.G. Reading (Ed.) Sedimentary Environments and Facies. Blackwell Scientific, Oxford, pp. 97-142.

Evans, J.E. and A. Clark. 2011. Re-interpreting Great Lakes shorelines as components of wave-influenced deltas: An example from the Portage River delta (Lake Erie). Journal of Great Lakes Research 37, pp. 64-77.

Fairbanks, R.G., Mortlock, R.A., Chiu, T., Cao, L., Kaplan, A., Guilderson, T.P., Fairbanks, T.W., Bloom, A.L., Grootes, P.M. and M. Nadeau. 2005. Radiocarbon calibration curve spanning - to 50,000 years BP based on paired 230Th/234U/238U and $^{14}C$ dates on pristine corals. Quaternary Science Reviews 24, pp. 1781-1796.

Farrand, W.R. and D.F. Eschman. 1974. Glaciation of the southern peninsula of Michigan: a review. Michigan Academician 7, pp. 31-56.

Farrand, W.R., and D.L. Bell. 1982. Quaternary Geology of Southern Michigan. 1:500,000-scale. Department of Natural Resources, Geological Survey Division, Lansing, Michigan.

Farrand W.R., and C.W. Drexler. 1985. Late Wisconsinan and Holocene history of the Lake Superior Basin. In P.F. Karrow and P.E. Calkin (Eds.) Quaternary Evolution of the Great Lakes. Geological Association of Canada Special Paper 30, pp. 17-32.

Fisher, T.G., Jol, H.M., and A.M. Boudreau. 2005. Saginaw Lobe tunnel channels (Laurentide Ice Sheet) and their significance in south-central Michigan, USA. Quaternary Science Reviews 24, pp. 2375-2391.

Foss, J.E. and R.H. Rust. 1968. Soil Genesis Stufy of a Lithologic Discontinuity in Glacial Drift in Western Wisconsin. Soil Science Society of America Journal 32, pp. 393-398.

Galloway, W.E. 1975. Process framework for describing the morphologic and stratigraphic evolution of deltaic depositional systems. Deltas, Models for Exploration, Houston Geological Survey, pp. 87-98.

Hough, J.L. 1958. Geology of the Great Lakes. University of Illinois Press, Ubrana, IL.

Hutchinson, D.E. 1979. Soil Survey of Midland County, Michigan. United States Department of Agriculture Natural Resource Conservation Service, Lincoln. NE.

Karathanasis, A.D., and P.A. Golrick. 1991. Soil Formation on Loess/Sandstone Toposequences in West-Central Kentucky: I. Morphology and Physiochemical Properties. Soil Science 152, pp. 14-24.

Kehew, A.E., Esch, J.M., Kozlowski, A.L., and S.K. Ewald. 2012. Glacial landsystems and dynamics of the Saginaw Lobe of the Laurentide Ice Sheet, Michigan, USA. Quaternary International 260, pp. 21-31.

Kincare, K. and G.J. Larson. 2009. Evolution of the Great Lakes. In Schaetzl, R., Darden, J. and D. Brandt (Eds.), Michigan Geography and Geology, Pearson Custom Publishing, New York, NY, pp. 174-190.

Kozlowski, A.L., Kehew, A.E., and B.C. Bird. 2005. Outburst flood origin of the Central Kalamazoo River Valley, Michigan, USA. Quaternary Science Reviews 24, pp. 2354-2374.

Krist and Lusch, 2004. Glacial history of Michigan, U.S.A.: A regional perspective. In J. Ehlers and P.L. Gibbard (Eds.) Quaternary Glaciations - Extent and Chronology Part II, Amsterdam, Netherlands, pp. 111-117.

Kunkle, G.R. 1963. Lake Ypsilanti: A probable Lake Pleistocene Low-Lake Stage in the Erie Basin. The Journal of Geology 71, No. 1, pp. 72-75.

Larson, G. and R.J. Schaetzl. 2001. Origin and Evolution of the Great Lakes. Journal of Great Lakes Research 27, pp. 518-546.

Leeder, M. R. 1982. Sedimentology Process and Product. George Allan & Unwin, London UK.

Leverett, F. and F.B. Taylor. 1915. The Pleistocene of Indiana and Michigan and the history of the Great Lakes. U.S. Geological Survey Monograph 53.

Lewis, C.F.M., Cameron, G.D.M., Anderson, T.W., Heil Jr., C.W., and P.L. Gareau. 2012. Lake levels in the Erie Basin of the Laurentian Great Lakes. Journal of Paleolimnology 47, pp. 493-511.
Luehmann, M.D., Schaetzl, R.J., Miller, B.A., and M.E. Bigsby. 2013. Thin, Pedoturbated and Locally sourced loess in the western Upper Peninsula of Michigan. Aeolian Research 8, pp. 85-100.

Lusch, D.P. 1982. The origin and morphogenetic significance of patterned ground in the Saginaw Lowland of Michigan. PhD Dissertation, Michigan State University, East Lansing, Michigan.

Lusch, D.P., Stanley, K.E., Schaetzl, R.J., Kendall, A.D., Van Dam, R.L., Nielsen, A., Blumer, B.E., Hobbs, T.C., Archer, J.K., Holmstadt, J.L.F. and C.L. May. 2009. Characterization and Mapping of Patterned Ground in the Saginaw Lowlands, Michigan: Possible Evidence for Late-Wisconsin Permafrost. Annals of the Association of American Geographers 99, pp. 1-22.

Mason, J.A. 2001. Transport Direction of Peoria Loess in Nebraska and Implications for Loess Sources on the Central Great Plains. Quaternary Research 56, pp. 79-86.

McLeese, R.L. and S.W. Tardy. 1985. Soil Survey of Isabella County, Michigan. United States Department of Agriculture Natural Resource Conservation Service, Lincoln. NE.

Muhs, D.R., Been, J., Mahan, S.A., Burdett, J., Skipp, G., Rowland, Z.M., and T.W. Stafford Jr. 1997. Holocene eolian activity in the Minot dune field, North Dakota. Canadian Journal of Earth Sciences 34, pp. 1442-1459.

Nanson, R.A., Vakarelov, B.K., Ainsworth, R.B., Williams, F.M., and D.M. Price. 2013. Evolution of a Holocene, mixed-process, forced regressive shoreline: The Mitchell River delta, Queensland, Australia. Marine Geology 339, pp. 22-43.

Olson, C.G., and R.V. Ruhe. 1979. Loess Disperson Model, Southwest Indiana, U.S.A. Acta Geologica Academiae Scientarum Hungaricae, Tomus 22, pp. 205-227.

Raad, A.T. and R. Protz. 1971. A New Method for the Identification of Sediment Stratification in Soils of the Blue Springs Basin, Ontario. Geoderma 6, pp. 23-41.

Santra, M., Goff, J.A., Steel, R.J., and J.A. Austin Jr. 2013. Forced regressive and lowstand Hudson paleo-Delta system: Latest Pliocene growth of the outer New Jersey shelf. Marine Geology 339, pp. 57-70.

Schaetzl, R.J. 1998. Lithologic discontinuities in some soils on drumlins: Theory, detection, and application. Soil Science 163, pp. 570-590.

Schaetzl, R.J., Krist, F., Rindfleisch, P., Liebens, J., and T. Williams. 2000. Postglacial landscape evolution of northeastern Lower Michigan, interpreted from soils and sediments. Annals of the Association of American Geographers 90, pp. 443-466.

Schaetzl, R.J., and B.N. Weisenborn. 2004. The Grayling Fingers region of Michigan: Soils, sedimentology, stratigraphy and geomorphic development. Geomorphology 61, pp. 251-274.

Schaetzl, R.J. and S. Anderson. 2005. Soils: Genesis and Geomorphology. Cambridge University Press, New York, NY.

Schaetzl, R.J., Krist Jr., F.J., Stanley, K., and C.M. Hupy. 2009. The Natural Soil Drainage Index: An Ordinal Estimate of Long-Term Soil Wetness. Physical Geography 30, Vol. 5, pp. 383-409.

Schaetzl, R.J., Enander, H., Luehmann, M.D., Lusch, D.P., Fish, C., Bigsby, M., Steigmeyer, M., Guasco, J., Forgacs, C., and A. Pollyea. 2013. Mapping the Physiography of Michigan with GIS. Physical Geography 34, pp. 2-39.

Schaetzl, R. J., and M.D. Luehmann. 2013. Coarse-textured basal zones in thin loess deposits: Products of sediment mixing and/or paleoenvironmental change? Geoderma 192, pp. 277-285.

Silva, P.G., Bardaji, T., Calmel-Avila, M., Goy, J.L., and C. Zazo. 2009. Transition from alluvial to fluvial systems in the Guadelentin Depression (SE Spain) during the Holocene: Lorca Fan versus Guadelentin River. Geomorphology 100, pp. 140-153.

Smith, G.D. 1942. Illinois Loess: Variations in its Properties and Distribution, a Pedologic Interpretation. University of Illinois, Agricultural Experiment Station.

Soil Survey Division Staff. 1993. Soil Survey Manual. USDA Handbook No. 18. US Government Printing Office, Washington, DC. pp. 60-64.

Stanley, K.E., Schaetzl, R.J. 2011. Characteristics and paleoenvironmental significance of a thin, dual-sourced loess sheet, north-central Wisconsin. Aeolian Research 2, pp.241-251.

Thomas, M.O. 2009. The United States Public Land Survey System. In Schaetzl, R., Darden, J. and D. Brandt (Eds.), Michigan Geography and Geology, Pearson Custom Publishing, New York, NY, pp. 430-445.

Vader, M.J., Zeman, B.K., Schaetzl, R.J., Anderson, K.L., Walquist, R.W., Freiberger, K.M., and J.A. Emmendorfer. 2012. Proxy Evidence for Easterly Winds in Glacial Lake Algonquin, from the Black River Delta in Northern Lower Michigan. Physical Geography 33, pp. 252-268.

Walt, J.A., Graham, R.C., and P.J. Schoeneberger. 2013. Distribution and properties of soft weathered bedrock at 1m depth in the contiguous United States. Earth Surface Processes and Landforms 38, pp. 614-626.

Ward, M.J. 1979. Glacial History of Early Lake Saginaw, M.S. Thesis, Michigan State University, East Lansing, Michigan.

Weiguo, L., Bhattacharya, J. P., Yingmin, W. 2011. Delta asymmetry: Concepts, characteristics, and depositional models. Petroleum Science Vol. 8, pp. 278-289.

Winsemann, J., Brandes, C., and U. Polom. 2011. Response of a proglacial delta to rapid high-amplitude lake-level change: an integration of outcrop data and high-resolution shear wave seismics. Basin Research 23, pp. 22-52.

Wright, L.D., and J.M. Coleman. 1972. River Delta Morphology: Wave Climate and the Role of the Subaqueous Profile. Science Vol. 176, pp. 282-284

Wright, L.D., Thom, B.G., and R.J. Higgins. 1980. Wave Influences on River-mouth Depositional Process: Examples from Australia and Papua New Guinea. Estuarine and Coastal Marine Science 2, pp. 263-277.

Wysocki, D.A., Schoeneberger, P.J., and H.E. LaGarry. 2005. Soil Surveys: a window to the subsurface. Geoderma 126, pp. 167-180.