12-1996

A Depositional Analysis of the Wymer Site, Berrien County, Michigan and Its Relationship to Post-Glacial Lake Levels in the Lake Michigan Basin

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by

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A Thesis
Submitted to the
Faculty of The Graduate College
in partial fulfillment of the
requirements for the
Degree of Master of Science
Department of Geology

Western Michigan University
Kalamazoo, Michigan
December 1996

ACKNOWLEDGMENTS

This thesis could not have been written without the aid of several people. My thanks to W. Thomas Straw, who gave me three chances to wrap this up, and seemed to think I had it in me to finish my degree, despite ample evidence to the contrary. His insights in pointing me in the right direction helped make this a better product. For good or ill, Tom's glacial geology class made me want to do this stuff for a living.

I owe Elizabeth Garland my thanks for asking me to work on the geologic problems at her archeology site. My thanks to Al Kehew, for bringing me "into the '90s."

Having friends like Mike Wireman and Nick Pogonchef, followers of Dr. H.S. Thompson, made graduate school all that it should be in combining scholarship with fellowship. Julie Stoneman's friendship and work as field assistant was invaluable, whether she was making meaningful notes from my babble, wielding a trowel, or digging me out of a cave-in. I hope that the late W. David Kuenzi knows the inspiration he was in making sedimentary processes, for me, a most compelling subject.

But most of all, I thank Valerie Haan, my partner in every sense of the word. She alone provided me with the love, respect and trust that allowed me to finish this work. I can't wait to spend the rest of my life with her, because together, all things are possible.

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The Wymer site is an archeological site on the St. Joseph River floodplain 1.5 miles north of Berrien Springs, Michigan. Early and Middle Archaic cultural phases were found at the site. These sites are unusual in that their occurrence corresponds to low lake levels associated with Lake Chippewa and were assumed to occupy the now submerged shoreline areas.

A site evaluation was undertaken to examine the types of sediments and depositional structures. Trenches, grain-size analysis, ^{14}C samples and valley topography were used to determine that the site was a point bar situated on a terrace above the modern floodplain. The site was probably associated with headward erosion during lower base level. Comparison of data to lake level changes over the last 14,000 years yielded an early Chippewa regression time for point-bar deposition and headward erosion. Dated samples indicated a subsequent period of aggradation that preceded the Nipissing high stand by 1900 years.

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INTRODUCTION

The Wymer site is a multi-component archeological site on the St. Joseph River flood plain, about one mile north of Andrews University in Berrien Springs, Michigan (Figure 1). The site was discovered during an archeological survey of the proposed U.S. 31 extension (Garland and Mangold, 1980). Preliminary excavations revealed the presence of Early, Middle, and Late Archaic, as well as Late Woodland cultural components at the site (Garland and Clark, 1981). The site is 21 km (13 miles) upstream along the St. Joseph River from Lake Michigan in township 6 south, range 18 west, section 2, in Berrien County.

The combination of the cultural phases listed above at the same site presented a unique opportunity to correlate post-glacial lake stages with an aboriginal occupation (Figure 2). This is especially true in regard to the Early and Middle Archaic. Very few Early or Middle Archaic sites have been discovered in Michigan (Fitting, 1975). It had been proposed that the extreme low post-glacial lake levels (Lake Chippewa-Stanley) were contemporaneous with the Early and Middle Archaic

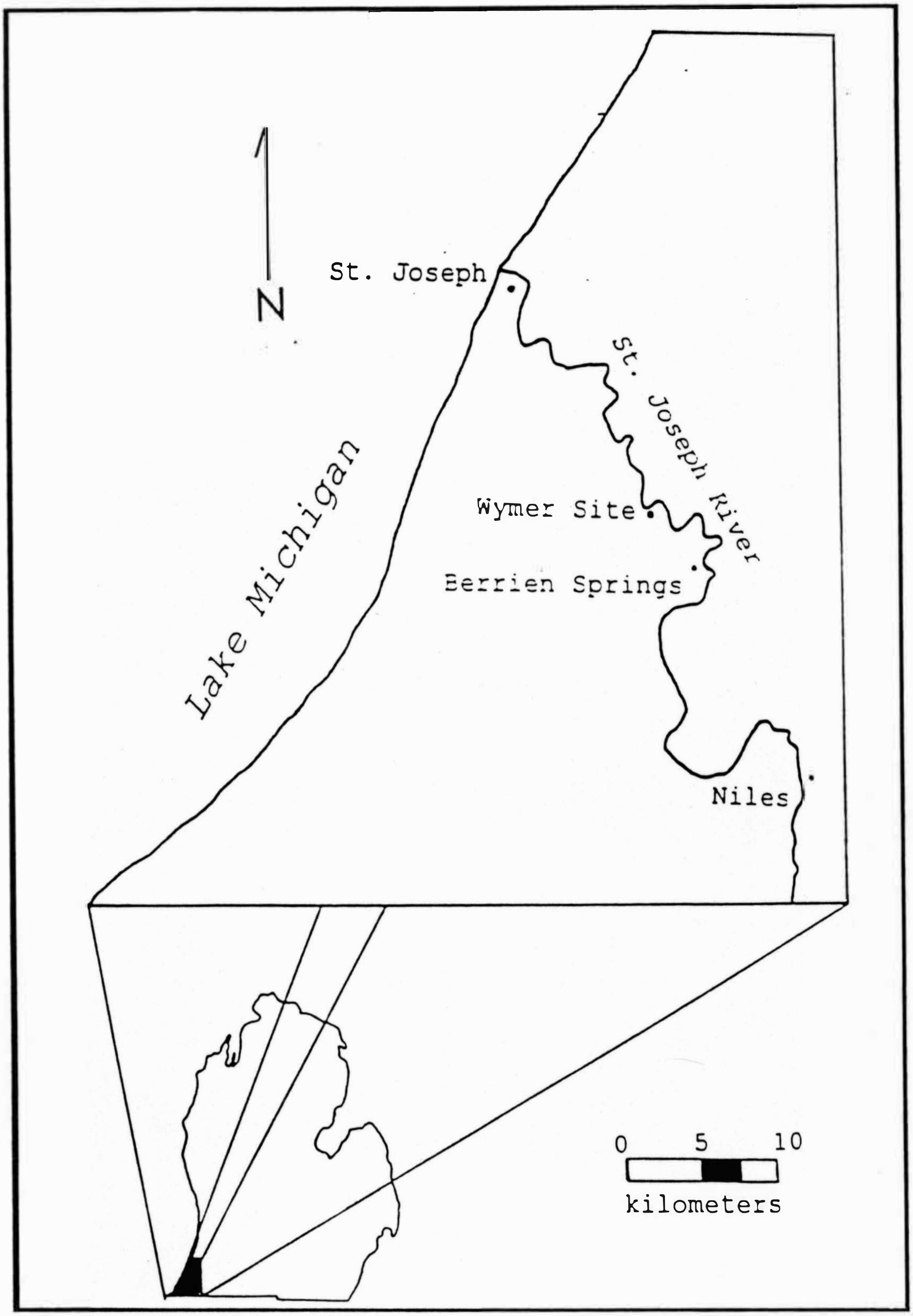


Figure 1. Location of the Wymer Site, Berrien County, Michigan.

Years Before Present	Lake Phase (mid-point)	Elevation (meters)	Cultural Phase
14000	Glenwood I	195	
	Mackinaw	?187-170	
13000			
	Glenwood II	195	
12000	Two Creeks	?173.7	
	Calumet	189	
11000	Algonquin	184-80	
	Algonquin ends		
10000	(North Bay open)		
	Chippewa	77-97	
9000			
			Paleo-Indian
	Olsen Forest Bed	153	
8000			
7000			Early Archaic
6000			
			Middle Archaic
5000	Nipissing	184	
4000			Late Archaic
	Algoma	181	
3000			Early Woodland
	Present	177	
2000			

compiled from Hansel et al., 1985; Colman et al., 1994; Fitting, 1975.

Figure 2. Post-Glacial Lake Levels and Cultural Phases With Time.

cultural period (Fitting, 1975; Griffin, 1965). Because people of these cultures occupied shoreline areas (Cleland, 1976), Early and Middle Archaic people were assumed to live in the lands which were exposed by the low lake levels (regression). Research also has indicated the possibility that populations were very low during those cultural phases (Fitting, 1975). When the lake level returned to its near present elevation, the previously exposed land was inundated (transgression), along with the majority of the Early and Middle Archaic sites. Independently derived geophysical and stratigraphic evidence have led some workers to believe that the rise from Lake Chippewa-Stanley (70.1 m) to Lake Nipissing (184.4 m) may have occurred much faster than originally believed (Larsen, 1985; Fillon, 1972; Brotchie and Silvester, 1969). Larsen (1985) stated that most of the Middle Archaic period existed between the time the water level of Lake Chippewa-Stanley had already risen back to present levels (176.8 m) and when it continued to rise to Lake Nipissing (184.4 m). This suggests that Middle Archaic sites may not all be underwater. They may lie on the present day land surface which had been flooded by the rise in water level to the Nipissing Lake Stage. In that case, the sites may have been buried by transgressive sediments or reworked by the transgression (Larsen, 1985).

The Middle Archaic artifacts found at the Wymer site consisted of only two points. Although typologically diagnostic, they were not associated with datable cultural features (e.g., charcoal from a fire pit). This leaves the origin of their final distribution at the site in question. Were they left directly by the manufacturers, deposited with transgressive sediments, or left by later occupants who had found and reused them? What was the possibility that the location contained a site buried by transgressive sediments?

PROBLEM STATEMENT

In order to answer these questions, the writer undertook a geologic evaluation of the Wymer site to examine the types of sediments and depositional structures. A depositional analysis could give an indication of the stream velocities and water depths with time at the site. This could indicate whether a buried archeological site would have survived intact or would have been reworked. A survey of the elevations of the depositional structures would be correlated with post-glacial lake stages. This could help determine when the site was available for occupation and possibly, when it was most likely to have been occupied.

Several separate tasks were required to bring the archeology and geology together. An understanding of stream characteristics, their adjustments to base-level changes and the depositional features associated with those adjustments was necessary for the investigation. An understanding of lake level changes in the southern Lake Michigan basin was required to compare depositional changes with known base-level changes. The profile and valley characteristics of the St. Joseph River were examined to observe how they were adjusted to the present lake level and whether they displayed any evidence of past changes. The Wymer site was trenched to reveal the stratigraphy of depositional features and relate them to base-level changes. A temporal correlation of aboriginal cultural phases to post-glacial lake levels was constructed. Finally, all the points mentioned above were correlated to determine when the Wymer site had been open to occupation.

METHODOLOGY

Field work for this project was conducted on July 7-8, 1980, August 21, 1981, through September 5, 1981, and October 20-23, 1981. The field work began with an examination of the Wymer site. Four trench locations were chosen to reveal the vertical and horizontal distribution of sediments (Figure 3). Trenches were dug with a backhoe to depths ranging from 2.01 to 3.30 meters. Trench walls were profiled using grain size and sedimentary structures as unit indicators. Soil samples were taken from units with recognizable sedimentary structures. Core samples were taken from the bottom of Trench 2 to provide samples of the fine grained deposits underlying the coarse deposits. After the cores were logged for geological features, samples of organic materials were separated for carbon-14 dating and pollen analysis. Top-of-trench elevations were established by the archeological field crew on November 9, 1981.

Soil samples were taken to the Western Michigan University Geology Department sedimentology lab for analysis. They were dried at $100^{\circ}F$ and split with a

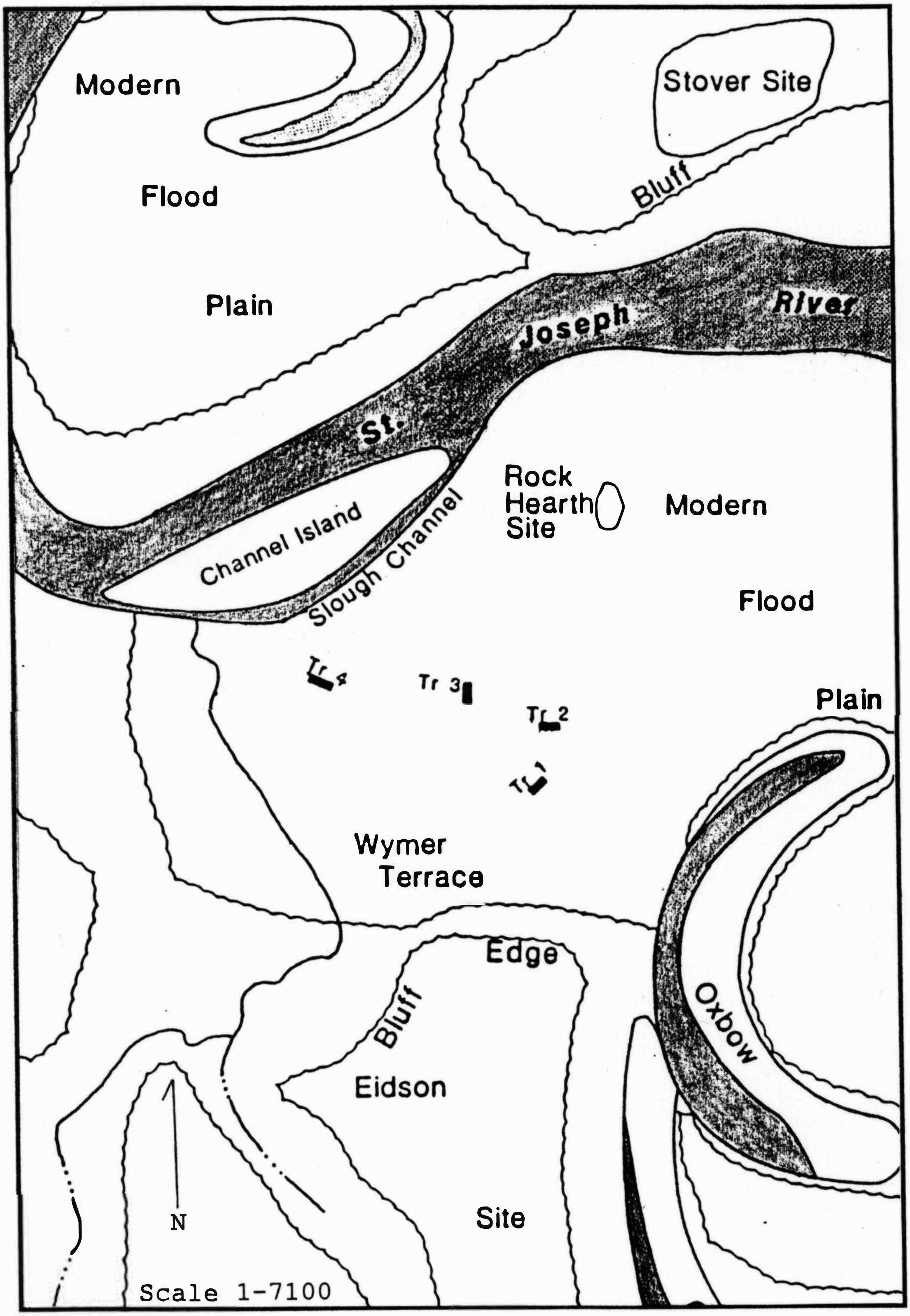


Figure 3. Wymer Site Map With Trench Locations.

random sample splitter. They were then screened with a Ro-Tap shaker. Each screen split was weighed and grain-size frequency distributions were developed (Appendix I). Cumulative curves with arithmetic and probability axes were constructed using these data. The curves were employed to statistically characterize the samples. Statistical parameters used included the following: Mode (Mo), Mean (Md), Graphic Mean (Mz), Inclusive Graphic Sorting (σi), Inclusive Graphic Skewness (Sk_i), and Graphic Kurtosis (K_G). Methods for these analyses are described in detail by Krumbien and Pettijohn (1938) and Folk (1974). Estimates of stream velocity, water depth and depositional environment were based, in large part, on discussions, charts and graphs presented by Reineck and Singh (1975), Simons and Richardson (1962), Guy et al. (1966), Rubin (1987) and Jopling (1966).

Topographic features and the longitudinal profile of the St. Joseph River valley (Figure 4) were examined by using elevation and distance data from USGS topographical quadrangles. River mileage was measured along the river channel. Elevation contours were correlated to total channel distance from Lake Michigan. Sinuosity was determined by measuring channel length versus valley

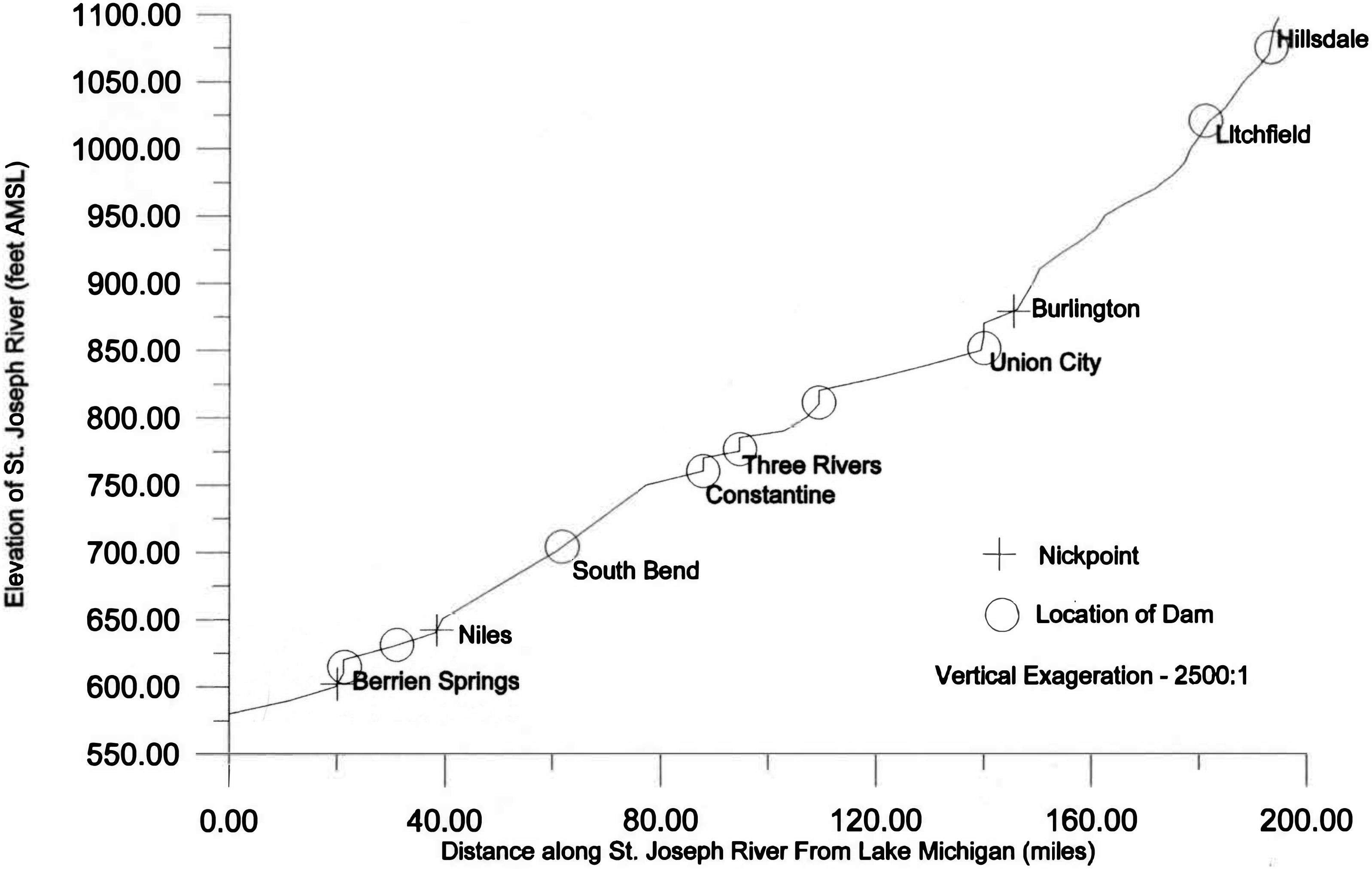


Figure 4. Longitudinal Profile of the St. Joseph River.

length from topographic maps. The following topographic maps were utilized:

1. 7.5 minute series: Baroda, Benton Harbor, Berrien Springs, Burlington, Colon, Constantine, Elkhart East, Elkhart West, Galien, Hillsdale, Litchfield, Lyon Lake, Mendon, Mosherville, Mottville, Niles East, Niles West, North Adams, Nottawa, Sodus, South Bend East, South Bend West, Southwest Albion, Tekonsha, Three Rivers East, Three Rivers West, Union City.

2. 15 minute series: Benton Harbor, Leonidas, Niles, Three Rivers, Spring Arbor, Hillsdale, Union City, Vandalia.

3. 30 x 60 minute series: Adrian, Benton Harbor, Elkhart, Jackson, Kalamazoo, South Bend.

Streamflow data from gaging stations was gathered from the Environmental Protection Agency's STORET computerized data storage and retrieval system. This information is available from the Michigan Department of Natural Resources mainframe data repository. A summary of this information was published in the final report of the Great Lakes Basin Commission (1975). This information was used to evaluate the discharge of the St. Joseph River, which is one of the controlling factors of the shape of the present river channel.

A reconnaissance examination of the flood plain of the St. Joseph River was done from the dam at Berrien Springs to the river mouth at St. Joseph, Michigan. Flood-plain features were observed and hand-auger cored to determine sediment types. The features at the Wymer site were compared to reconnaissance data to observe whether depositional conditions could be correlated upstream or downstream from the site.

THE CONCEPT OF BASE LEVEL

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

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

THE CONCEPT OF THE GRADED STREAM

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

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

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

Many artificial flume experiments (e.g., Gilbert, 1914; Leopold and Wolman, 1957; Wolman and Brush, 1961) and field studies (e.g., Rubey, 1952; Mackin, 1948; Leopold and Bull; 1979) have been performed in efforts to quantify the factors that control stream morphology. Leopold and Wolman (1957) summarized these to include: discharge, sediment volume, grain size distribution, velocity, slope, channel width, depth and roughness. Said variables are either dependent or independent of stream action. For example, a stream cannot control the climate (and therefore its discharge) nor can it control the amount of sediment supplied by the drainage basin.

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

The compilation of experimental and empirical data allowed workers like Mackin (1948) and Rubey (1952) to constructively refine the concept of the graded stream. Mackin (1948) said that a graded condition is an equilibrium reached when the slope at all locations along a stream is just enough to carry the load being supplied, given the existing circumstances. A graded stream's "...

defining characteristic is that any change in any of the controlling factors will cause a displacement of the equilibrium in a direction that will tend to absorb the effects of the change" (Ibid., p. 471). Mackin used the valley of the Shoshone River, Wyoming, as an example of a stream that shows evidence of having alternately been at grade and downcut (degraded) during its development. The Shoshone has a slope of more than 30 ft per mile (both terraces and present channel) that does not vary when crossing boundaries between soft and hard bedrock. Yet Mackin (Ibid.) observed that even with its steep slope, the Shoshone was at grade during terrace construction and not eroding its channel. Otherwise, its slope would have varied between bedrock units with differing resistance to corrasion. The terraces were formed by lateral corrasion of the confining walls of the valley. The thin alluvium covering the bedrock floor was formed by lateral shifting of meanders (see discussion below).

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

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

ADJUSTMENTS TO GRADE BY A SHIFT OF BASE LEVEL

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

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

Changes that an alteration in energy level can bring about are numerous. An increase in energy can cause the onset of erosion along the affected stream channel. The process of downcutting can lead to the terrace formation, entrenchment of the stream in an incised channel, and loss of contact with other stream valley features (e.g., point bars, flood plains, etc.) of the previous flow regime (Ibid.). Increasing erosion and stream competence can increase its sediment load and may lead to deposition on those point bars that are still in contact with the stream or new point bars may be established. The depositional sequence may show an increase in grain size before returning to the typical fining upward sequence characteristic of point bars (Thornbury, 1969; Reineck and Singh, 1975).

An increase in base level has the obvious effect of the water level rising and drowning the lower portion of a stream. Of major importance, however, is the loss of stream competence in carrying its sediment load. Hence, the stream will deposit those particles it is no longer capable of transporting. The stream may adjust for the loss of competence by increasing its slope (Mackin, 1948). Deposition may take place on point bars, in the river channel (which increases slope), and in flooded reaches below the new base level. Hypothetical depositional sequences for these areas of the stream must consider the loss of competence. Point bars may experience increased deposition with particle size either fining upward or being of uniform size (Reineck and Singh, 1975).

A graded river flows in dynamic equilibrium and does not deposit significant amounts of sediment in its channel (Mackin, 1948; Rubey, 1952). The deposits within the channel are referred to as bedload sediment. Bedload is moved by a stream within the channel and transported mainly by rolling action (Bagnold, 1960; Reineck and Singh, 1975). These deposits account for only a small percentage of total stream deposits. A loss of competence, as in the waning phases of a flood, may form a fining upward sequence in the channel. Deposition within the channel raises its elevation and could be indicative of loss of competence. A rise in base level can cause loss of competence. Channel deposits, however, are particularly subject to erosion should competence subsequently increase, (e.g., lowered base level), and may not be preserved.

Areas below the new base level may receive a fining upward sequence (deltaic) of sediment deposited in the water. Deposits would include silt and clay-size particles which are now able to settle out in the still water. Flooded areas that were formerly floodplains will receive only silt and clay deposits as normally happens during floods. As such, this type of deposit may be indistinguishable from normal flood-plain deposits.

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

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

The St. Joseph River lacks continuous channel division around alluvial islands and clearly is not braided in any of its sections. Throughout its length it contains both straight and meandering reaches. The St. Joseph River at the Wymer site is a meandering reach with a sinuosity of 1.54 (excluding the estuary at Lake Michigan). Therefore, the following discussion will concentrate only on the depositional features of meandering streams.

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

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

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

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

Meandering streamflow adds more complexity in horizontal relationships. High-water stages can temporarily inundate the point bar and temporarily create a short-term high energy environment. During flood stage, the thalweg is wider; coarser grains and higher energy sedimentary structures can move over point bar. A two-tier point bar may develop (McGowen and Garner, 1970). The lower tier is formed by the mean flow velocity; the upper tier is formed by peak flow periods. Flood chutes may also develop in the low area along the rear of the point bar (McGowen and Garner, 1970; Reineck and Singh, 1975). These chutes funnel some of the overflow behind the more fully developed, topographically higher portion of the point bar. The low area initially develops from the beginning stages of point-bar formation. At that time, the current pattern that creates point bars is not fully developed and sedimentation is low. Subsequent high water flows cause chute bars to be deposited in the flood chute. Grain size and sedimentary structures in chute bars are dependent on the strength of the flood. McGowen and Garner (1970) indicated that sand waves, foreset ripple, and climbing ripple lamination are the most common chute bar structures. Waning flood strength creates a fining upward sequence, often with a mud drape and vegetation structures on top. Successive flood periods can create alternating fining upward sequences within a composite chute bar.

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

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

where:

F - Froude number

v - flow velocity

g - acceleration due to gravity

h - height of water column

A Froude number >1 is in the rapid flow regime. Because the St. Joseph River does not fit the rapid flow category by either velocity, gradient or Froude number, rapid flow will not be discussed.

Tranquil flow sedimentary structures are the dominant forms of fine-grained meander belt streams (McGowen and Garner, 1970; Reineck and Singh, 1975). Tranquil flow sedimentary structures (from lowest energy to highest energy) consist of small ripples, small trough sets, megaripples, and large trough sets. Grain-size distribution exerts a partial control on the type of structures that will form. When the mean grain size is greater than 0.6 mm (coarse sand), small ripples will not form. When the critical velocity is reached for particle motion on such a sand bed, the first structure that is formed will be megaripples (Simons and Richardson, 1962). Structures formed in the tranquil flow regime were found to be independent of the depth of water (Guy et al., 1966; Simons and Richardson, 1962). The controlling factors were the velocity of flow and, to the degree previously mentioned, grain size. Hence, depth of water in an ancient deposit laid under tranquil flow conditions, cannot be estimated with a reasonable degree of accuracy.

CHARACTERISTICS OF THE ST. JOSEPH RIVER

The St. Joseph River and its tributaries (excepting the Paw Paw River) drain approximately 11,187 km^2 (4,300 $miles^2$) in southwestern Michigan and northcentral Indiana. It originates from Baw Beese Lake, one kilometer southeast of Hillsdale, Michigan. The river flows a total of 316 kilometers (196 miles) and falls 158.8 meters (521 feet).

The surface material in the drainage basin is almost entirely unconsolidated glacial material. Glacially derived material provides a wide range of sediment sizes for erosion and transport. Moraines provide clay through boulder-size particles. Outwash plains contain mostly sand-size sediment. The only known bedrock outcrops along the course of the St. Joseph River are in the vicinity of Hillsdale, Michigan. Here, the Marshall Sandstone crops out in places from under a thin mantle of ground moraine. The Marshall is composed of medium to coarse sand grains in this area. It is compacted and partially cemented.

The St. Joseph River's flow is moderately controlled with a total of nine dams along its course (Great Lakes

Basin Commission, 1975). The river profile displays a progressively headward gradient increase common to river systems (Figure 4). In some reaches of the river, the gradient changes are gradual; elsewhere the changes are abrupt. The abrupt changes are nickpoints (Thornbury, 1969). Anthropogenic nickpoints can be accounted for and eliminated by interpolation as illustrated in Figure 4. Once these cultural nickpoints are removed, the true nickpoints remain.

Three natural nickpoints exist on the St. Joseph River. Located at 34 km, 62 km, and 226 km from Lake Michigan, respectively, these features effectively divide the St. Joseph River into four sections (Figure 4). The first section of river, from 0 to 34 km, has a narrow valley bordered by steep bluffs (Figure 5). Here, the river meanders across a maximum valley width of 1.6 km. River features include scroll type point bars, steep valley walls, a well-developed meander pattern, five visible meander cutoffs, and a well-drained flood plain. The Rock Hearth site, on the St. Joseph River's modern floodplain 300 meters northeast of the Wymer site, yielded a ^{14}C date of 3740 +/- 80 years B.P. (Garland, 1984). The material for this date was recovered at a depth of less than 0.5 meters. This showed that very

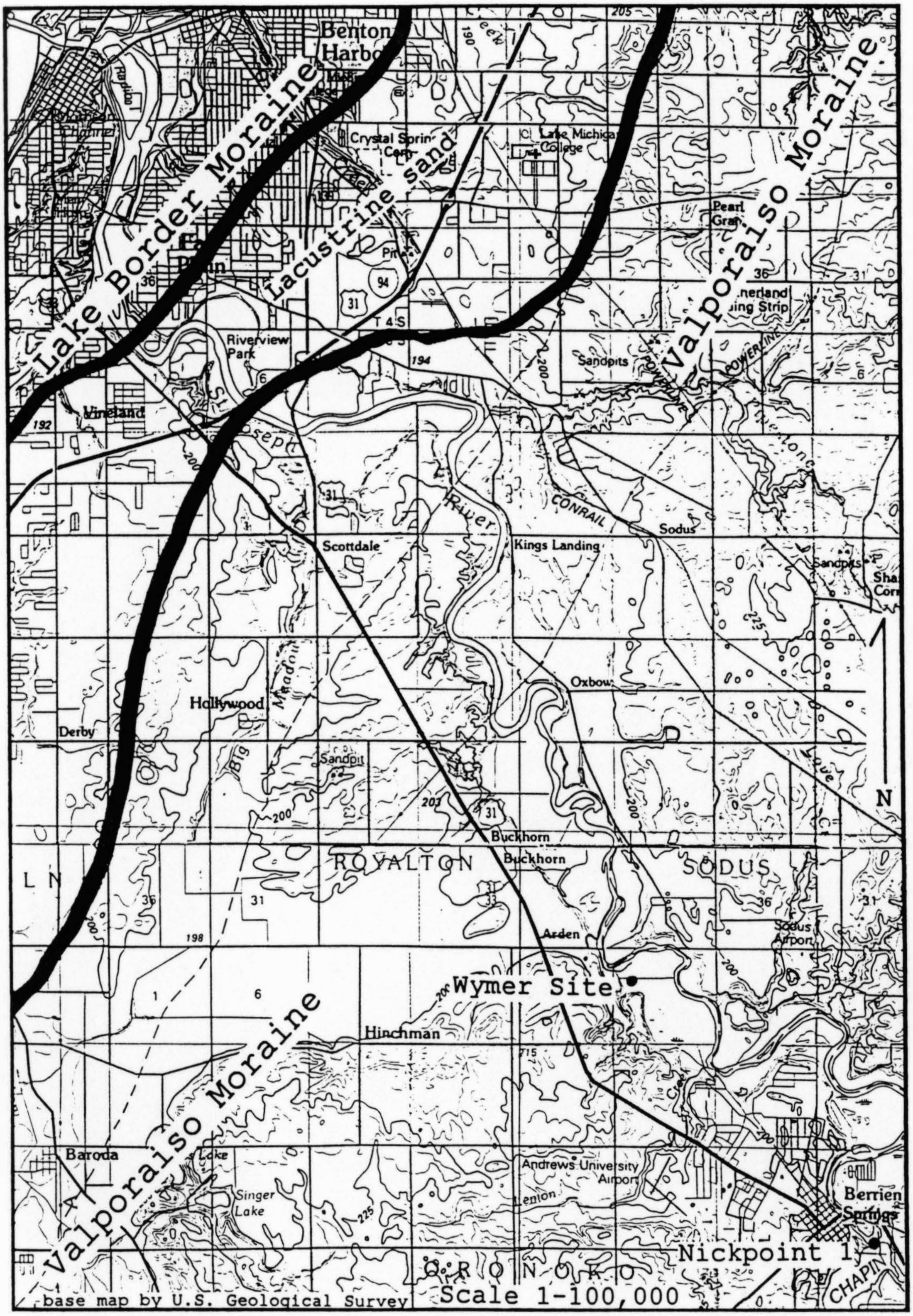


Figure 5. St. Joseph River-Plan View From Mouth to Nickpoint 1.

little alluviation has occurred since that time. The average gradient of this section is 0.19 m/km (1 foot/mile) and it cuts across the Lake Border Moraine and into the inner edge of the Valparaiso Moraine. Winters et al. (1986) stated that the valleys across moraines in West Michigan originated as a result of draping of the glacial debris over preexisting valleys in the bedrock. Accepting this, it would be not be genetically correct to say that the St. Joseph River "cuts" across or into the moraines. This theory seems plausible in that the river already had an outlet to the southwest at South Bend, Indiana when the ice blocked the present valley across the Valparaiso Moraine (Figure 6). There is also no field evidence of a delta from deposition of eroded material where the river exits the moraine, despite the depiction of a delta on maps by Martin (1955) and Leverett and Taylor (1915).

The first nickpoint, 34 km from Lake Michigan, is located in Berrien Springs, Michigan (Figure 5). It is at an elevation of about 183.5 meters (602 ft). It is situated just below the artificial nickpoint of the Lake Chapin dam. Above the nickpoint, the St. Joseph River has a straight channel (sinuosity of 1.06). The flood plain is narrow to nonexistent in the section within the

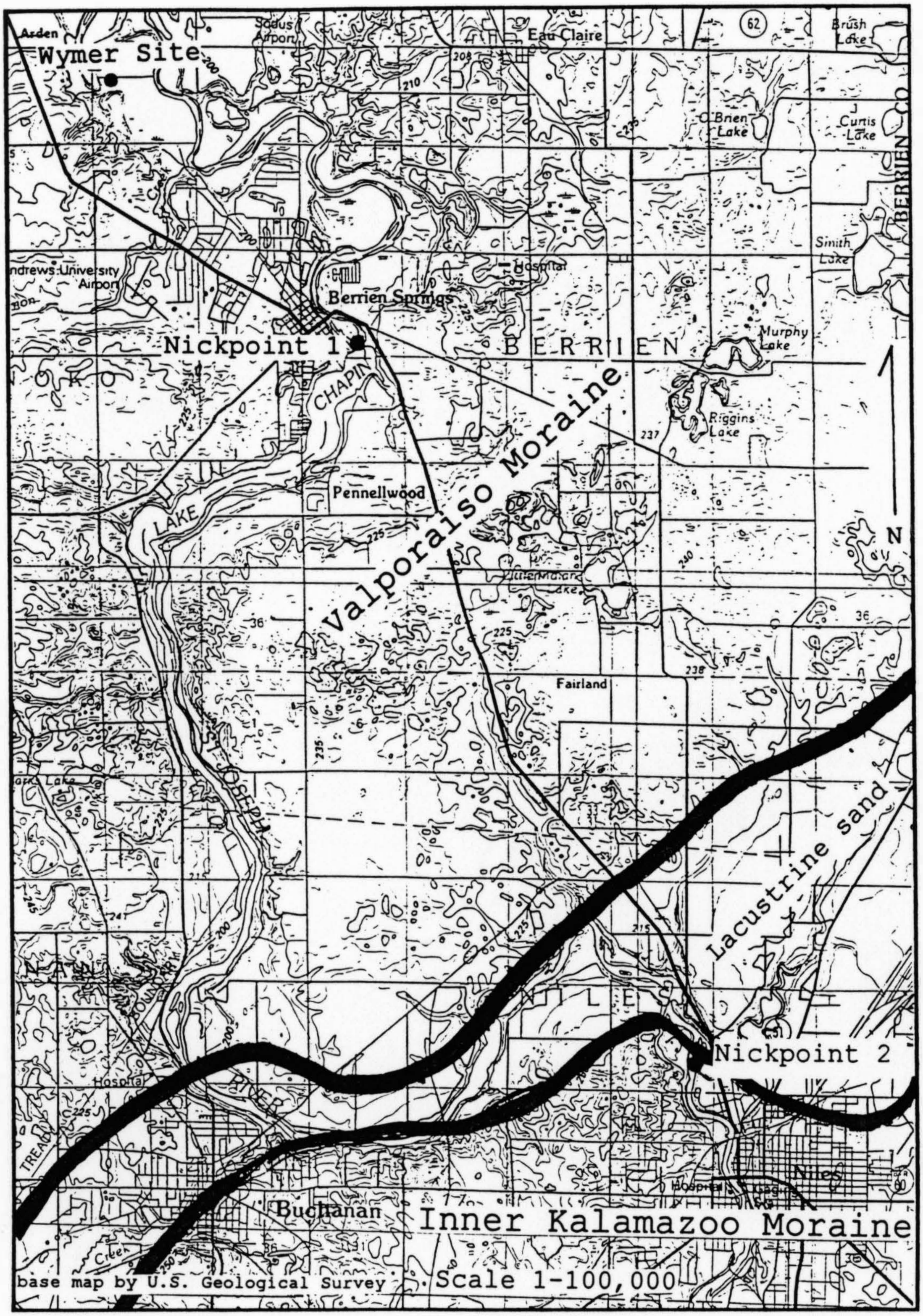


Figure 6. St. Joseph River-Plan View From Nickpoint 1 to Nickpoint 2.

Valparaiso Moraine. The valley walls, though not as steep as in the previous river section, encroach upon the river; valley width is no greater than 0.4 km. While it is true that part of this section is within the Chapin Dam pond, it is clear from Figure 6 that the valley would be barely wider than the river even without the dam. The gradient above the first nickpoint increases to 0.30 m/km (1.6 feet/miles). In this stretch the river has one major tributary and many short tributaries. The short tributaries cut sharp valleys into the surrounding uplands. The river enters the Valparaiso Moraine 8 km above the dam from a strip of outwash/lacustrine sand deposits that lay between the Valparaiso and Inner Kalamazoo Moraines (Leverett and Taylor, 1915; Farrand, 1982). Dowagiac Creek, the first significant tributary river of the St. Joseph River, joins just below the second nickpoint.

The second nickpoint is located at Niles, Michigan, 62 km above the mouth of the river (Figure 6) at an elevation of about 195.1 m. At this point, the gradient increases to 0.47 m/km (2.5 feet/miles). In the area of this nickpoint, there is a change in the surficial geology. The river goes into the valley across the Inner Kalamazoo Moraine. This third river section maintains the low sinuosity of the previous section to the vicinity of Elkhart, Indiana. At this location, the density of significant tributaries begins to increase and the sinuosity increases to about 1.3 without a change in slope.

The third and final nickpoint evident from the profile is near Burlington, Michigan, 226 km above the mouth at an elevation of about 280 meters (918 feet). The river again crosses between different glacial landforms. Sand and gravel outwash deposits and ground moraine exist below the nickpoint and above it the river crosses the Tekonsha Moraine. The gradient continues to rise smoothly to a maximum of 1.05 m/km before the river reaches its origin in Hillsdale. Below the nickpoint, the gradient remains uniform at a decline of 0.47 m/km. Given this, and because the third nickpoint is located nearly three-quarters the total length of the St. Joseph River from its mouth, it seems unlikely that it would be affected by base-level changes in the Lake Michigan basin.

All three nickpoints occur at a change in glacial terrain where the river debouches from a moraine. In addition, the second nickpoint is also at the elevation that marks the Glenwood level of Lake Chicago, the highest of the Lake Michigan Basin's post-glacial base levels (see discussion below). The first nickpoint is below the three highest post-glacial base levels.

BASE-LEVEL CHANGES IN THE LAKE MICHIGAN BASIN

One of the factors controlling the grade in the lower St. Joseph River is the base level in the Lake Michigan Basin. Glacial and post-glacial events in the basin controlled the lake level (Figure 2). The St. Joseph River has been subjected to those changes in post-glacial lake level. The base-level controls the energy available for erosional and depositional processes at the Wymer site. Timing of the base-level change events establishes when the sequence of events derived from the Wymer stratigraphy could have taken place. The average elevation of the Wymer site is about 183.2 meters (601 ft) with a maximum of 184.8 meters (606 ft).

Much of the general sequence of events established by Leverett and Taylor (1915) is still in use today (Figure 2). Controversies have centered on specific problems such as the duration of certain lake stages, the ice front position at specific times, and the correlation of stratigraphic units (e.g., Farrand and Eschman, 1974). Of particular importance to this study are recent developments indicating that paleoclimatic influences may also have had an effect on lake levels (Larsen, 1985; Lovis, 1981). These paleoclimatic variations would have been superimposed on earlier megascale fluctuations (i.e., crustal downwarping and subsequent rebound from glacial load), blending into the background of megascale changes. By 7,000 years B.P., however, megascale variation ceased to be a major influence (Fillon, 1972) and paleoclimatic variations may have become the dominant control over lake levels.

During the deglaciation of Michigan, the St. Joseph River basin was one of the first of its present day major drainage basins to be free of ice. Before the ice retreated from the Valparaiso Moraine, the St. Joseph River's outlet was south to the Kankakee River (Figure 7). Water from the Kalamazoo and Grand River basins joined with the St. Joseph at Niles, Michigan to drain to the Kankakee (Leverett and Taylor, 1915). These drainage waters ponded into a lake behind the height of land (around 223 meter elevation) at two locations: southwest of South Bend, Indiana and southwest of Buchanan, Michigan.

When the ice retreated to the Lake Border Moraine drainage entered the Lake Michigan Basin for the first time. When this occurred about 14,200 years B.P., the

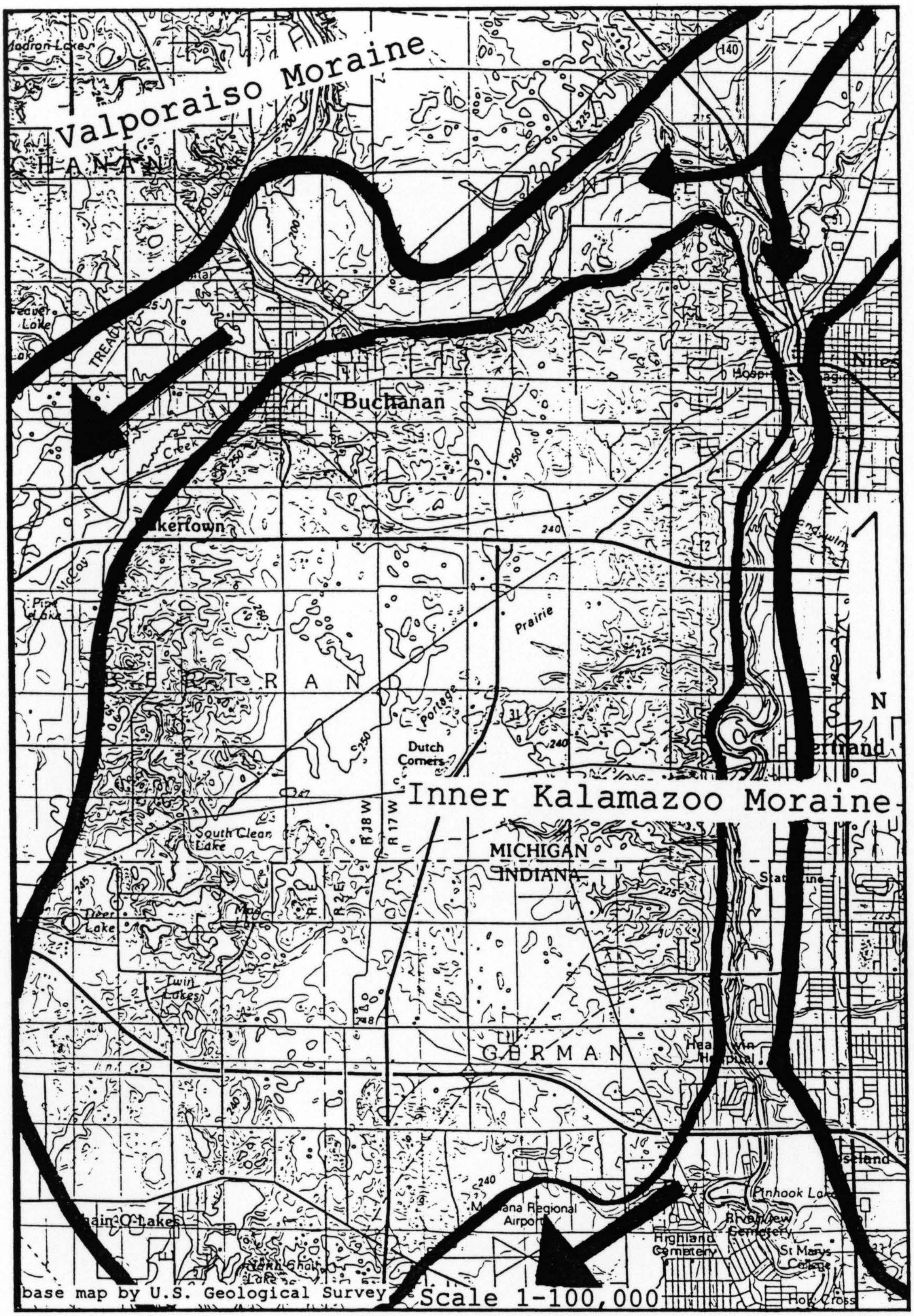


Figure 7. St. Joseph River-Drainage to Kankakee River During Valporaiso Glacial Substage.

river was subjected to the base-level changes of the basin. Drainage was to the south, ponded behind the Tinley Moraine (Farrand and Eschman, 1974). This is known as the Glenwood I phase of Lake Chicago which stood at 195 meters (640 feet) M.S.L. (Figure 8). This phase is believed to have lasted for about 200 to 400 years. Glenwood I would have been stable at the Wymer site from the time that the ice retreated to the Lake Border Moraine until ice front moved north of the Indian River lowland in Cheboygan County, Michigan. This also assumes that the channel across the Tinley Moraine in northern Illinois resisted erosion. The elevation of the Glenwood stage would have ponded water at the Wymer site allowing only lacustrine deposition. No delta would have formed near the site as the water was ponded all the way to the 195 meter contour in Niles, Michigan.

Ice recession from the Lake Border Moraine marks the beginning of the Mackinaw lake phase or Cary/Port Huron interstadial. Farrand et al. (1969) reported evidence of ice recession beyond Cheboygan County dated at 13,300 years B.P. The evidence was in the form of a tundra-type bryophyte bed found in the Indian River area. This implied ice recession north of this point with a low lake stage for the time. However, Larson (personal

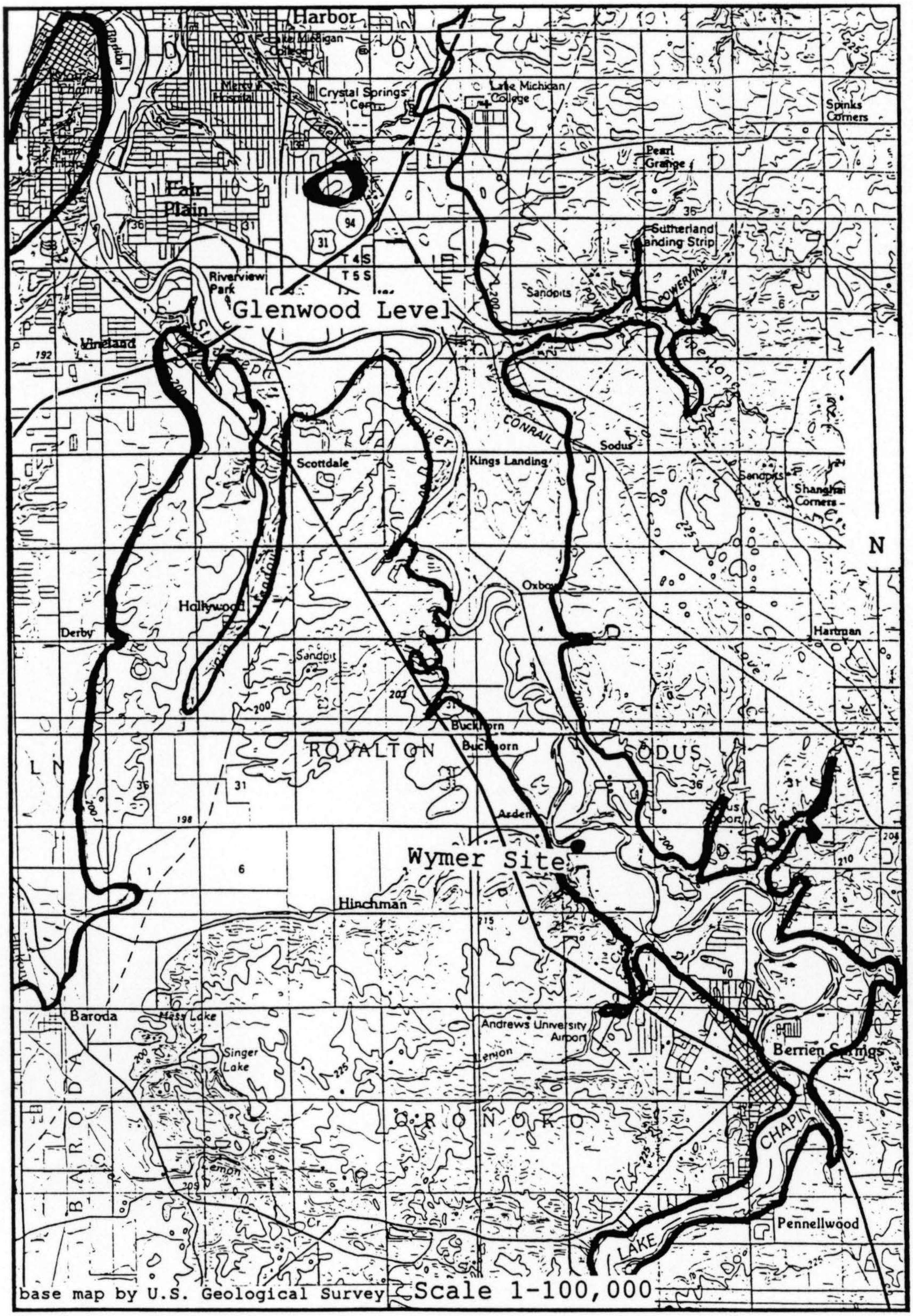


Figure 8. St. Joseph River-Wymer Site Ponded During the Glenwood Phase Lake Chicago.

communication) reexamined the bryophyte bed and, by separating different sources of organic carbon, refined the date to 11,800 years B.P. Beaches from this stage have not been identified or do not exist in the Michigan basin. Without these features, a height estimate for this stage would be speculation. Hough (1958) cited evidence that indicates a possible elevation of 187.5 meters (615 feet) for this stage. Hansel et al. (1985) cited the lack of evidence as reason to believe that the Mackinaw Phase was lower than present. Glacial rebound should have elevated such a beach above the present lake level had it been persistent. More convincing perhaps, is the 13,500 year B.P. date at the base of transgressive sediments at near present lake elevation by Riverside, Michigan (Monaghan and Hansel, 1990). This evidence suggests that flow conditions at the Wymer site would be similar to those found today. The Mackinaw Phase lasted about 300 years and workers have yet to find decisive evidence of its elevation and ice front location and, by inference, its ability to have effected lasting change. Flow conditions at the Wymer site during the Mackinaw Phase are speculative.

The well-dated Port Huron glacial advance followed the brief recession around 13,000 years B.P. (Taylor, 1990; Blewett, 1990). This advance once again sealed off the northeastern Michigan drainage routes (assuming the ice front had been that far north). Drainage returned to the Chicago outlet to begin the Glenwood II lake level at 195 meters (640 feet). This ponded the Wymer site again. Glenwood II has an average age of 12,400 years B.P. (Karrow et al., 1975). Hansel et al. (1985) cited dates on material in Glenwood II beach features at the Dyer spit between 12,200 and 12,700 years B.P. Hough (1966) believed that erosion of the Chicago initiated two lower lake stages. The first lower stage, Calumet at 189.0 meters (620 feet) and then the Toleston at 184.5 meters (605 feet). The later stage being the elevation of the bedrock sill at the Chicago outlet which prevented further erosion (Leverett and Taylor, 1915; Hough, 1963). The timing and duration of these event was questioned in Hansel et al. (1985) and Colman et al. (1994). The mechanism for renewed erosion of the Glenwood outlet to lower its elevation after 400 years of stability cannot be seriously explained without a change in the conditions that existed at the outlet (Kehew, 1993). Hansel and Mickelson (1988) argued that an outlet at one elevation can accommodate flow at more than one level. Colman et al. (1994) do not believe that there is evidence for a

Calumet or Toleston level until after the advance of the Two Rivers ice though Hansel and Mickelson (1988) think that the fall from Glenwood II to Calumet happened at the close of the Glenwood phase. Regardless, the return to the Glenwood II phase even to a Toleston level assumes a ponded condition at the Wymer site.

The recession of Port Huron ice eventually uncovered the Indian River lowland and Straits of Mackinac again and allowed lower outlets to act upon the lake level about 12,000 B.P. Known as the Twocreekan interstadial, forest beds of this time have been dated at 11,850 years B.P. +/- 100 years (Broecker and Farrand, 1963). These forest beds established that the lake level was somewhat lower than the present day level. The Twocreekan interstadial allowed for riverine conditions at the Wymer site.

Glacial advance recovered the northern outlets during what is known as the Greatlakean glacial stage (Evenson et al., 1976). Formerly termed the Valders glacial stage, Farrand (1976) explained that, at its type section in Wisconsin, the Valders till lies below the Two Rivers till. The latter of which was the Wisconsin correlative of the Valders stage in Michigan. The confusing terms necessitated the renaming of the time stratigraphic unit. Nomenclature notwithstanding, the event glacial advance isolated the Lake Michigan Basin and drainage returned to the Chicago outlet. The elevation of this lake, known as the Calumet stage, was at 189 meters (620 feet). The furthest advance of the Greatlakean stage existed at about 11,800 years B.P. (Taylor, 1990). The advance was the last significant oscillation of ice in the lower peninsula of Michigan. The Calumet stage ponded the Wymer site.

Once the ice retreated beyond the Straits of Mackinac, waters of the Lake Michigan and Huron basins were joined. This was the beginning of the Main Algonquin stage at about 11,000 years B.P. (Hansel et al., 1985). The lake level of Main Algonquin was at 184.4 meters (605 feet). Drainage was originally thought to be through both the Chicago outlet and the Port Huron (St. Clair River) outlet (Leverett and Taylor, 1915; Hough, 1958). Hansel et al. (1985) and Larsen (1987) contended that uncovered outlets in Georgian Bay, Ontario are more likely drainage paths. Larsen (1987) argues that the Algonquin stage may have been well below the present lake surface in the southern part of the basin due to the downwarped northern part of the basin. The effect can be likened to that of a basin tilted to one end with the water collecting in the lowest portion. The Wymer site would be ponded at a 184.4 meter level of Lake Algonquin. However, if Larsen (1987) is correct, then the gradient at the Wymer site was significantly higher than today during the existence of Lake Algonquin.

Further recession of the ice uncovered successively lower outlets in the Georgian Bay, Ontario region. Several beach strands were formed when lake levels stabilized at one outlet before the next lower outlet was deglaciated. Hough (1958) published adjusted elevations for several of these beaches, referring to them as the "Upper Group" beaches. The lowest stage that occurred is referred to as Lake Chippewa, whose level may have been as low as 70.1 meters (230 feet). This level should be controlled by the level of the glacially depressed river channel which flowed through the bottom of the Straits of Mackinac to drain the Lake Michigan Basin into the Lake Huron Basin at that time. By the same reasoning, the level of the Lake Huron Basin would have been controlled by the height of the channel that led to Georgian Bay. In any case, the Algonquin level probably began to decline shortly after it began. The Chippewa low stage existed at approximately 10,300 years B.P. when the North Bay outlet was uncovered. The controlling factor in the rate of lake level decline was the rate at which ice recession uncovered lower outlets. Evidence as to what this rate may have been is tentative. From the dates given, it appears that about 1,000 years separated the end of the Algonquin stage from the Chippewa low stage. Larsen (1987) presented data that showed this period of time to be from 10,300 - 8150 years B.P. in the Lake Michigan basin. The transition from ponded to flowing stream began at the Wymer site when the Algonquin level dropped below the existing floodplain. The St. Joseph River adjusted its grade accordingly as the base level dropped to successively lower elevations.

The Chippewa stage began to rise when the glacially depressed outlets began to isostatically rebound from the ice load they had carried. Geophysical experimentation and modeling (Brotchie and Silvester, 1969; Fillon, 1972; McConnell, 1965; Clark et al., 1994) show that elastic rebound initially occurs at a rapid rate. The rate decreases geometrically with time. Fillon (1972) goes on to say that perturbations of the rate of rebound are more likely to occur early in the uplift sequence, rather than later, when their effects may have been distinguishable. With regard to uplift in the Great Lakes area, Fillon

(1972) stated that the rebound curve should be smooth (i.e., without perturbation) by 7,000 years B.P.

The Olson drowned forest site off Chicago yielded ^{14}C dates of around 8,300 years B.P. This shows that rebound has raised the lake level to 153 meters (502 feet) in about 1,700 years. By 5,000 years B.P. the lake level had returned to the 184.4 meters (605 feet) elevation (Hough, 1963). Larsen (1987) and Hansel et al. (1985) indicated that this event occurred by 6000 years B.P. This lake level is known as Lake Nipissing. Drainage at this time was through the Chicago outlet (on bedrock) and the Port Huron outlet (on unconsolidated glacial outwash). The Port Huron sediments were apparently susceptible to erosion because the lake level dropped from the Nipissing level. Another period of relative stability occurred long enough to build a weak beach system. Known as Lake Algoma, at 181.4 meters (595 feet), it existed 3,750 years B.P. Lake levels continued a general decline until the present stage of 176.8 meters (580 feet) was achieved. When the Nipissing transgression reached the level of the floodplain, the Wymer site was partially ponded, similar to conditions that may have existed during Lake Algonquin. The site remained ponded until after the regression from Lake Algoma around 3000 years B.P. (Hansel et al., 1985).

The events leading from Lake Chippewa to the present Lake Michigan, however, are not as straightforward as the description from the previous paragraph. Larsen (1985) and Lovis (1981) reported depositional and archaeological data that indicated a high degree of variability in lake levels from 7,000 years B.P. to the present day, particularly for the post-Nipissing era. They reported a pre-Nipissing peak of 178.5 meters (585.6 feet), a double Nipissing peak (4,750 years B.P. and 4,000 years B.P.), and several peaks and low stages after 2,000 years B.P. They also report apparent discrepancies between what should have been concurrent levels in the Huron and Michigan basins. These occurrences are in just the period that Fillon (1972) claimed rebound perturbations would have ended. One explanation appears to lie in paleoclimatic shifts. Larsen's data (1985) indicate that lake levels have not dropped below 175.0 meters (574 feet) from Nipissing time to the present day. The changes Larsen (1985) demonstrate appear to be short-lived geologically. Arguments such as these, lend further credence to the theory that the outlets were not eroding to drop water levels. Less water was moving out of the system and the water column was simply more shallow.

RESULTS OF TRENCH ANALYSIS

The Wymer site is a low, sandy ridge above the silty floodplain of the St. Joseph River (Figure 3). The site is on the west edge of the of the floodplain and has a convex northeast shape. The trenches at the Wymer site revealed a wide variety of grain sizes and sedimentary structures. Sediment found at the site ranged in size from silt and clay, indicative of low energy, to cobble gravel diagnostic of high energy. Samples for grain size analysis were taken from horizons with visible sedimentary structures. Otherwise, field estimates were used to determine grain size provenance. Statistical analysis of the grain size samples is provided in Table 1. The grain size distribution graphs are in Appendix A. Sedimentary structures ranged from straight ripple sets which indicate low energy currents up to trough sets indicating moderate velocities (Simons and Richardson, 1962; Reineck and Singh, 1975).

Trench 1 is located on the east knoll of the Wymer site, between the crest of the deposit and the valley wall (Figure 3). The base of Trench 1 (Figure 9) is at

Table 1

Grain-Size Frequency Distributions of Soil Samples From Trenches

Sample Location, Depth & Description	Mode (mm)	Sorting (phi)	Skewness (phi)	Kurtosis (phi)
Trench 1 157-164 cm silty fine sand, poorly sorted, fine skewed, leptokurtic	0.215	1.08	0.274	1.277
Trench 1 164 cm coarse sand, moderately well sorted, near symmetrical, leptokurtic	0.50	0.515	0.1	1.158
Trench 1 165-191 cm silty very fine sand, moderately sorted, strongly fine skewed, leptokurtic	0.155	0.772	0.322	1.475
Trench 1 192-209 cm medium sand, well sorted, coarse skewed, extremely leptokurtic	0.3	0.478	-0.198	6.22
Trench 1 213-272 cm medium sand, well sorted, symmetrical, mesokurtic	0.42	0.357	0.0	1.00
Trench 1 273-289 cm fine gravelly coarse sand, poorly sorted, strongly coarse skewed, platykurtic	0.42	1.12	-0.421	0.808
Trench 3 108-136 cm (upper sand unit) slightly gravelly medium sand, moderately sorted, strongly coarse skewed, very leptokurtic	0.42	0.788	-0.340	2.28
Trench 3 179 cm gravelly very coarse sand, poorly sorted, near symmetrical, platykurtic	0.42	1.36	-0.093	0.745

Table 1—Continued

Sample Location, Depth & Description	Mode (mm)	Sorting (phi)	Skewness (phi)	Kurtosis (phi)
Trench 4	0.42	1.088	-0.657	2.36
96-124 cm slightly gravelly medium sand, poorly sorted, strongly coarse skewed, very leptokutic				
Trench 4	0.30	0.477	0.045	1.155
125-135 cm (upper zone) medium sand, well sorted, near symetrical, slightly leptokurtic				
Trench 4	0.354	1.08	-0.439	1.732
125-135 cm (lower zone) Slightly gravelly medium sand, poorly sorted, strongly coarse skewed, very leptokurtic				
Trench 4	0.42	1.645	-0.499	1.056
136-164 cm gravelly coarse sand, poorly sorted, strongly coarse skewed, mesokurtic				

where:
- cm = centimeters
- mm = millimeters
- Phi = $-\log_2 d$
- d = diameter in mm
- mode = most frequently occurring grain-size diameter
- sorting = degree of spread of grain size distribution
- skewness = degree of asymmetry of grain-size distribution
- kurtosis = degree of departure from Gaussian distribution

Sorting:

<0.35	very well sorted	1.00-2.00	poorly sorted
0.35-0.50	well sorted	2.00-4.00	very poorly sorted
0.50-0.71	moderately well sorted	>4.0	extremely poorly sorted
0.71-1.00	moderately sorted		

Skewness:

>0.3	strongly fine skewed	(-0.1)-(-0.3)	coarse skewed
0.3-0.1	fine skewed	<(-0.3)	strongly coarse skewed
0.1-(-0.1)	near symmetrical		

Kurtosis:

<0.67	very platykurtic	1.11-1.50	leptokurtic
0.67-0.90	platykurtic	1.50-3.00	very leptokurtic
0.90-1.11	mesokurtic	>3.00	extremely leptokurtic

(after Folk, 1974; Krumbein and Pettijohn, 1938)

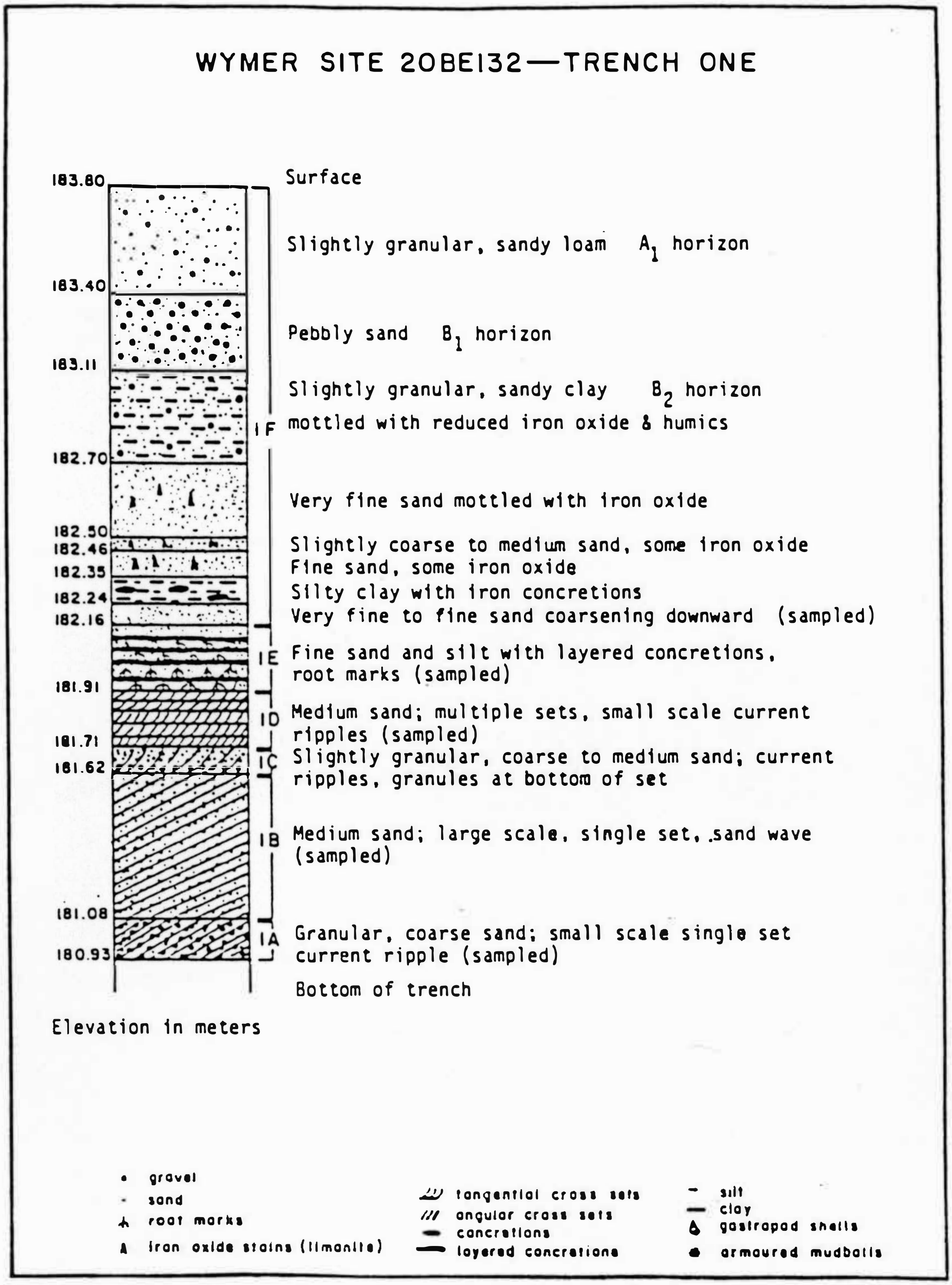


Figure 9. Wymer Site Trench 1 Cross Section.

180.93 meters M.S.L. (593.45 feet) in a granule gravel, coarse sand with current ripples (Unit 1A). This represents a moderate energy, relatively shallow water environment. The unit was poorly sorted, possibly indicating a variable current velocity. The next higher unit (1B) is a 0.54 meter thick, single set cross-bed of very well sorted, medium sand. It has a higher energy structure and its thickness implies deposition in deeper water than the unit beneath. The sorting infers a stable current. This unit conforms to the description of a chute bar (McGowen and Garner, 1970). Above lies a thin, slightly granule gravelly, coarse to medium sand layer of tangential set current ripples (1C). This indicates current strength similar to layer 1B, but in more shallow water. The poorer sorting also suggests less consistent flow. The next unit (1D) consists of multiple sets of thin current ripples in well sorted medium sand. The current had diminished appreciably by this time and the water depth was very shallow. Above this unit is a layer of organic rich silt and very fine sand (1E) with root marks and concretions. At this point in the depositional sequence vegetation had been established. No further sedimentary structures exist up section, but grain size increases and decreases alternately two more times. This indicates possible inundation and recession of moving waters on two more occasions. The evidence shows that this location experienced at least three separate episodes of flooding by water capable of transporting fine gravel at an elevation as high as 183.80 meters. It is the writer's opinion that Trench 1 exhibits chute-bar type deposition. By implication, the deposit is a point bar.

Trench 2 (Figure 10) is located on the floodplain side of the deposit, upstream of its axis. This trench did not reveal any sedimentary structures. Units that were visible consisted of two separate sequences. The lower sequence (Unit 2A) is mainly organic with identifiable grasses and coniferous seeds set in a matrix of silt, clay, and fine sand. These relatively thick layers alternated with thin beds of fine to medium sand. These deposits appear to be flood plain or channel cutoff type. The upper units (2B) at elevation 181.42 meters (595.06 feet) of medium to fine sands indicate a return to a somewhat higher energy, possibly from periodic flooding as the sand sizes appear to fluctuate and the sorting is poor. A piston coring device was used to collect materials from deeper in the trench. Two samples from the core were age dated by ^{14}C (Beta 6325 and 6326).

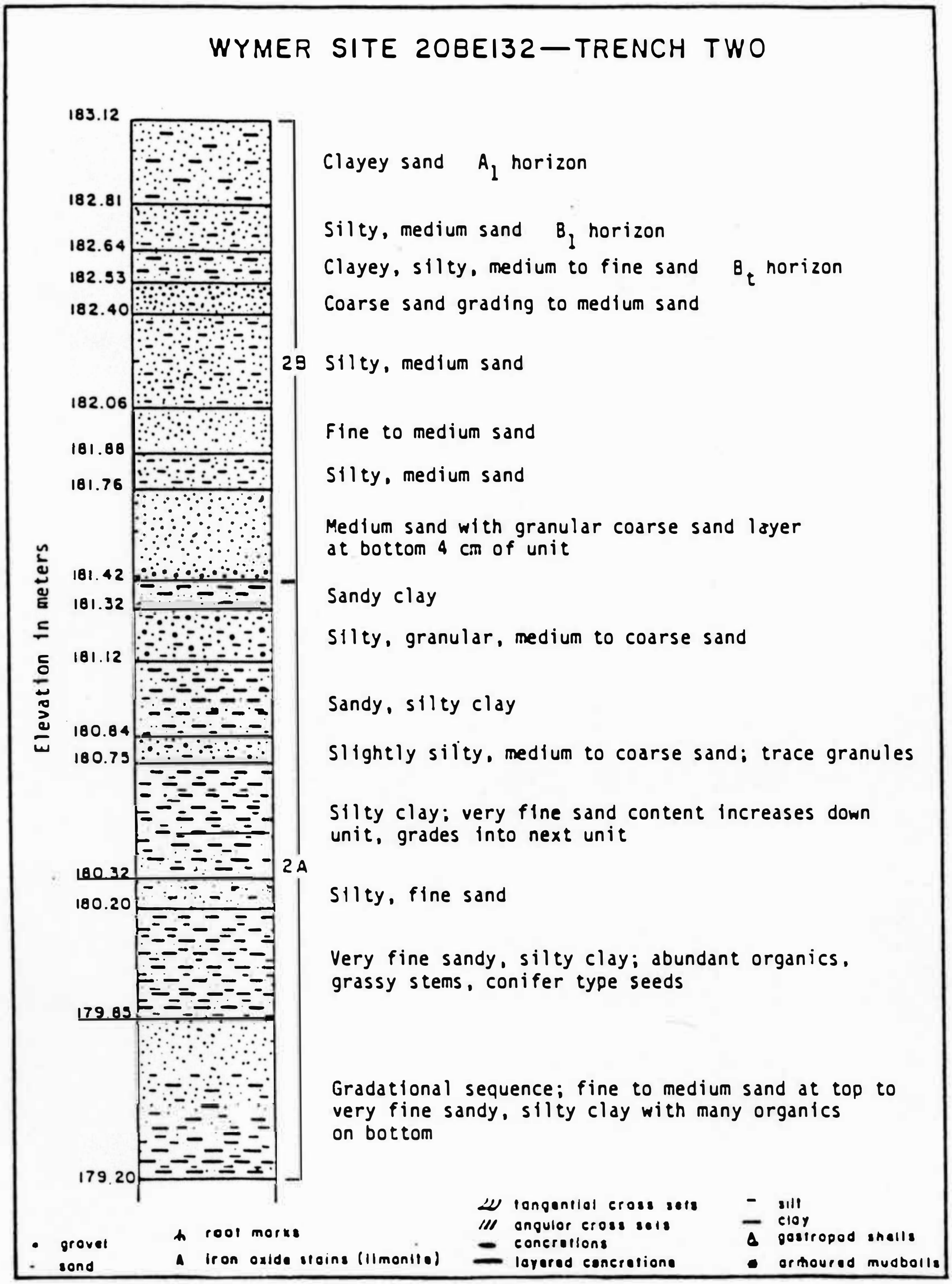


Figure 10. Wymer Site Trench 2 Cross Section.

The dates returned were 6,920 +/- 100 years B.P. and 5,550 +/- 160 years B.P. 4.5 and 3.5 meters land below surface, respectively. Without sedimentary structures, interpretation of this trench is difficult. However, its position places it where the river channel would have been during deposition of the point bar. It can thus be interpreted as channel fill after abandonment of the meander defined by the point bar. The materials in this trench must, therefore, post-date the point-bar deposits. The dates indicate an early pre-Nipissing timeframe.

Trench 3 (Figure 11) is located in the topographically high area on the leading edge of the point bar and along its axis. From its base at 183.60 meters (602.2 feet) up to 185.83 meters (609.21 feet) (Unit 3A), the trench displays moderately high energy sedimentary structures with pebble to granule gravel and very coarse to medium sand. These are typical deposits for this position on a point bar (Reineck and Singh, 1976). Several units within this zone contain large amounts of gastropod shells. A unit towards the top of this zone contained armored mudballs, indicative of erosional conditions (Folk, 1974). They are produced when clay banks are eroded creating clay balls that roll downstream picking up a coating of sand and/or pebbles that

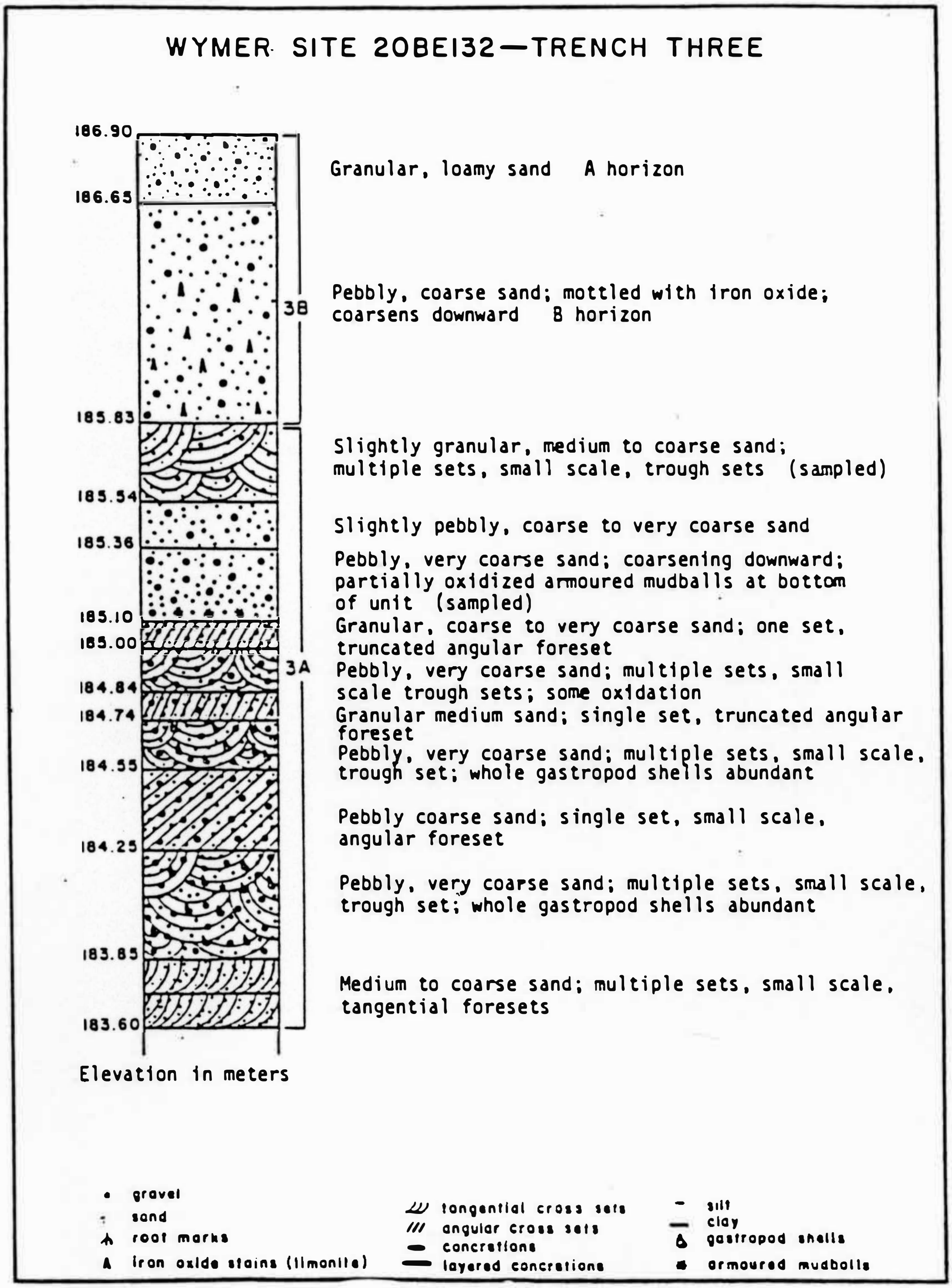


Figure 11. Wymer Site Trench 3 Cross Section.

protects them from disintegration. The trench was abandoned at a depth of 3.6 meters because of caving, and as a consequence, the underlying deposits were not sampled. Above 181.83 meters (609.52 feet) (Unit 3B) lay more coarse sand deposits without structures. These would also have required moderately high energy for deposition. This trench confirms that normal point-bar processes were operating at these elevations, over 2.5 meters above the present river elevation, at some time in the past. The previous discussion indicated that the Cary-Port Huron interstade, Two Creekan interstade, and the Lake Chippewa are the only periods when the base level was low enough to allow riverine processes to occur at the Wymer Site.

Trench 4 (Figure 12) is located on the downstream side of the deposit, toward the far end of the meander curve. The base of this trench is in channel lag deposits (Unit 4A) at elevation 183.40 meters (601.55 feet). The lag is composed of a very coarse sand, granule and cobble gravel. The maximum grain size observed was a cobble weighing 1320 grams (2.91 pounds). The lag deposit is higher in elevation than lower energy deposits in Trenches 1 and 3. The trench shows a single sequence of fining upward sediments (Unit 4B) above the

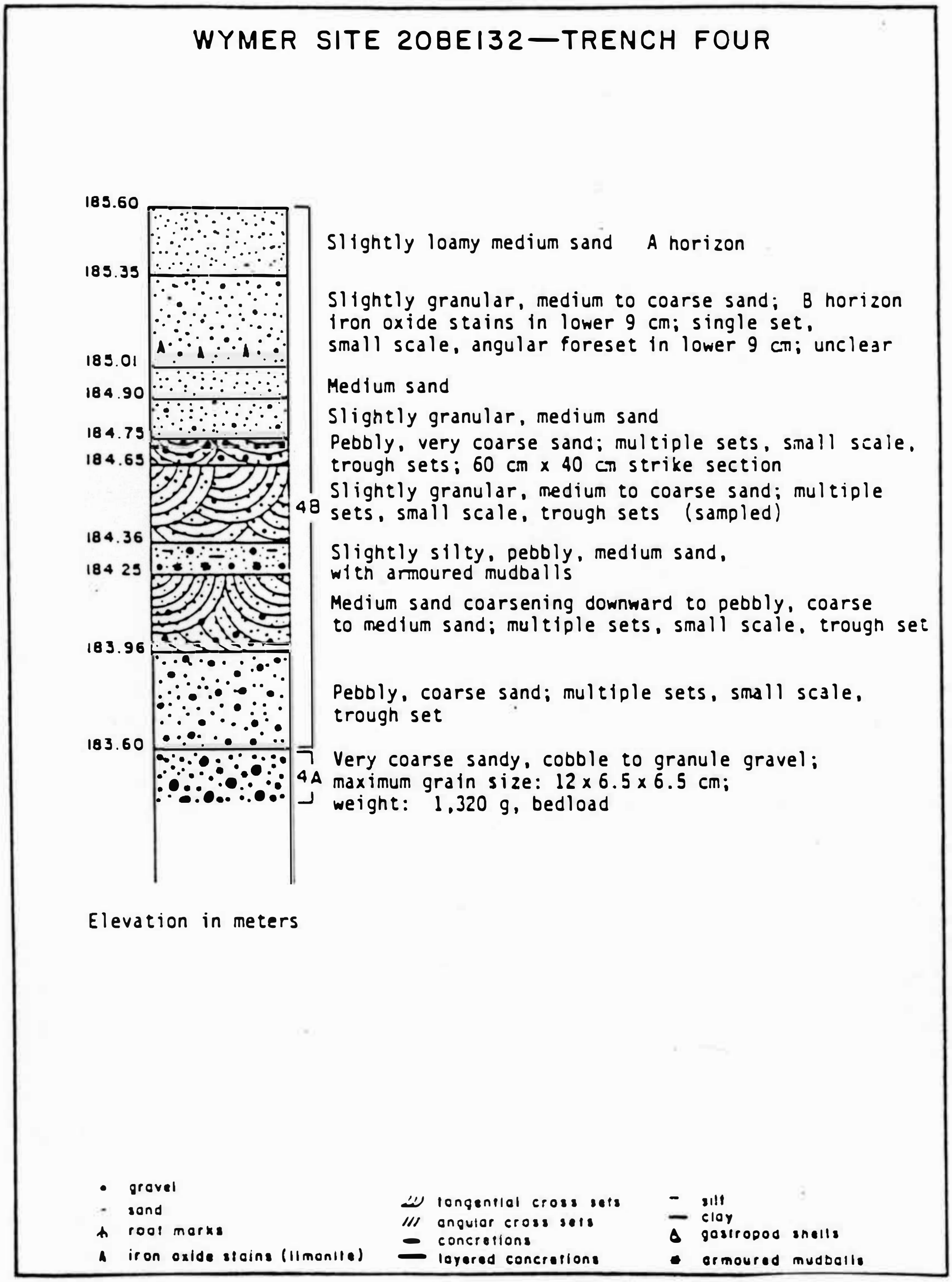


Figure 12. Wymer Site Trench 4 Cross Section.

lag deposit. The deposits in Trench 4 appear to grade into the general point-bar deposits. Due to their elevation, they could not have been laid concurrently. They were probably deposited late in the point-bar sequence when the fast flow vectors were beginning to return to this side of the channel. This is also supported by their position on the same terrace. A secondary or chute channel off the main channel may also have been responsible for the Trench 4 sediments. The present river maintains a channel island at this location (Figure 3).

CONCLUSIONS

The preceding discussions described the fluctuations in base level that occurred in the lower St. Joseph River and the types of erosional/depositional features associated with base-level change. The last section described the sedimentary features that were observed in trenches at the Wymer site. Figure 13 relates the various base levels over time to the streamflow conditions at the Wymer site. The average elevation of the point bar at the Wymer site is 183.5 meters (602 feet). Many of the glacial lake stages were at elevations of 184.4 meters (605 feet) or higher. A base level at this elevation would have ponded water at the site. By implication, a case could be made that the Wymer site is a delta where the St. Joseph River dumped debris into a high level lake. Only the Toleston/Nipissing lake level was close to the site (about 1.5 miles upstream). However, this analysis falls short considering that the deposit contains rip-up clasts, current ripple and trough set structures, and is not more coarse in the upstream direction. Another alternative depositional environment is that of meltwater discharge. The existence of gastro-

Years Before Present	Lake Phase (mid-point)	Elevation (meters)	Cultural Phase	Flow Conditions at Wymer Site
14000	Glenwood I	195		Ponded
	Mackinaw	?187-170		Uncertain
13000				
	Glenwood II	195		Ponded
12000	Two Creeks	?173.7		?Low gradient?
	Calumet	189		Ponded
11000	Algonquin	184-80		Ponded or High gradient
10000	Algonquin ends (North Bay open)			Gradient increases
	Chippewa	77-97		High gradient
9000				
			Paleo-Indian	
	Olsen Forest Bed	153		Gradient decreases
8000				
7000			Early Archaic	
6000				
			Middle Archaic	
5000	Nipissing	184		Ponded
4000			Late Archaic	
	Algoma	181		Very low gradient
3000			Early Woodland	
	Present	177		gradient 0.00019
2000				

compiled from Hansel et al., 1985; Colman et al., 1994; Fitting, 1975.

Figure 13. Post-Glacial Lake Levels, Cultural Phases and Wymer Site Flow Conditions With Time.

pods within the deposit, though, indicates a milder climate than would exist in close proximity to a glacier. Also, the deposit would have been under 11 meters of water when the glacier was in the Lake Border position.

Analysis of the trench profiles at the Wymer point bar has shown that the sediments were deposited by a stream capable of moving cobble size (264 mm) sediments. From this fact alone, we can conclude that the point bar was deposited during a lake stage at or lower than the present day level. This includes the Cary-Port Huron interstadial, Twocreekan interstadial, and the majority of the Algonquin-Nipissing interval.

Evidence from Trench 2 indicates that the meander bend that formed the Wymer point bar was abandoned at some time after the point bar's development. River basin morphology in the vicinity of the Wymer site shows other terraces that appear to relate to the Wymer terrace on the basis of elevation (Figure 5). The terraces indicate that a period of downcutting must post-date deposition of the Wymer point bar. The river was downcutting to establish a new equilibrium at a higher energy. Local uplift in this vicinity has not had a significant influence on the river/base level relationship in Holocene time. Therefore, it can be assumed that the downcutting described above was solely a response to the basin-wide influence of lower base level. The increase in the slope of the river without a concurrent increase in load would require increased erosion and lateral shifting of the channel (Mackin, 1948, page 487).

The profile of the St. Joseph River (Figure 4) shows that all the nickpoints are influenced by a change in glacial terrain. The nickpoint at Berrien Springs however, also shows evidence of a period of downcutting and headward erosion by a stream adjusting to a new, lower base level. The river has carved a 1.6 kilometer wide floodplain 12 kilometers into the inner flank of the Valparaiso Moraine. Above the nickpoint there is no floodplain. Because the load and flow of the St. Joseph River is controlled by dams at several locations, it is difficult to tell how it is adjusted to its present base level. Due to the Lake Chapin dam the lowest river segment most certainly has a reduced sediment load. Erosion does not appear to be a dominant process today. Point-bar deposits on present day meander bends are poorly developed. Floods are not uncommon. A post-flood inspection of the floodplain after the spring flood in 1984 showed minor alluviation rather than scour. At the peak of the flood, the Wymer site remained above water.

According to local fishermen, cutbanks are stable as are the channel islands. The river channel and islands are in the same position on the 1988 edition 7.5 minute quadrangles as they are on the 1927 edition 15 minute quadrangles. This suggests that the present river channel is adjusted to its base level.

The evidence thus far has indicated the following points:

1. A point-bar building episode occurred at the Wymer site by a stream of higher strength and sediment transport capacity than the present stream. The coarse sediment and rip-up clasts indicate active erosion was taking place at that time.

2. The Wymer site meander bend (Figure 14) was abandoned and organic, slackwater type sediments are present in what had been the river channel.

3. The St. Joseph River was either rejuvenated after the Wymer site point bar was formed or the point bar is a remnant of the early part of the rejuvenation. The river adjusted to its new conditions by eroding the floodplain and river channel by at least 2.5 meters. In order to downcut and remove that much sediment, a considerable input of energy would have been required (i.e., a much lowered base level).

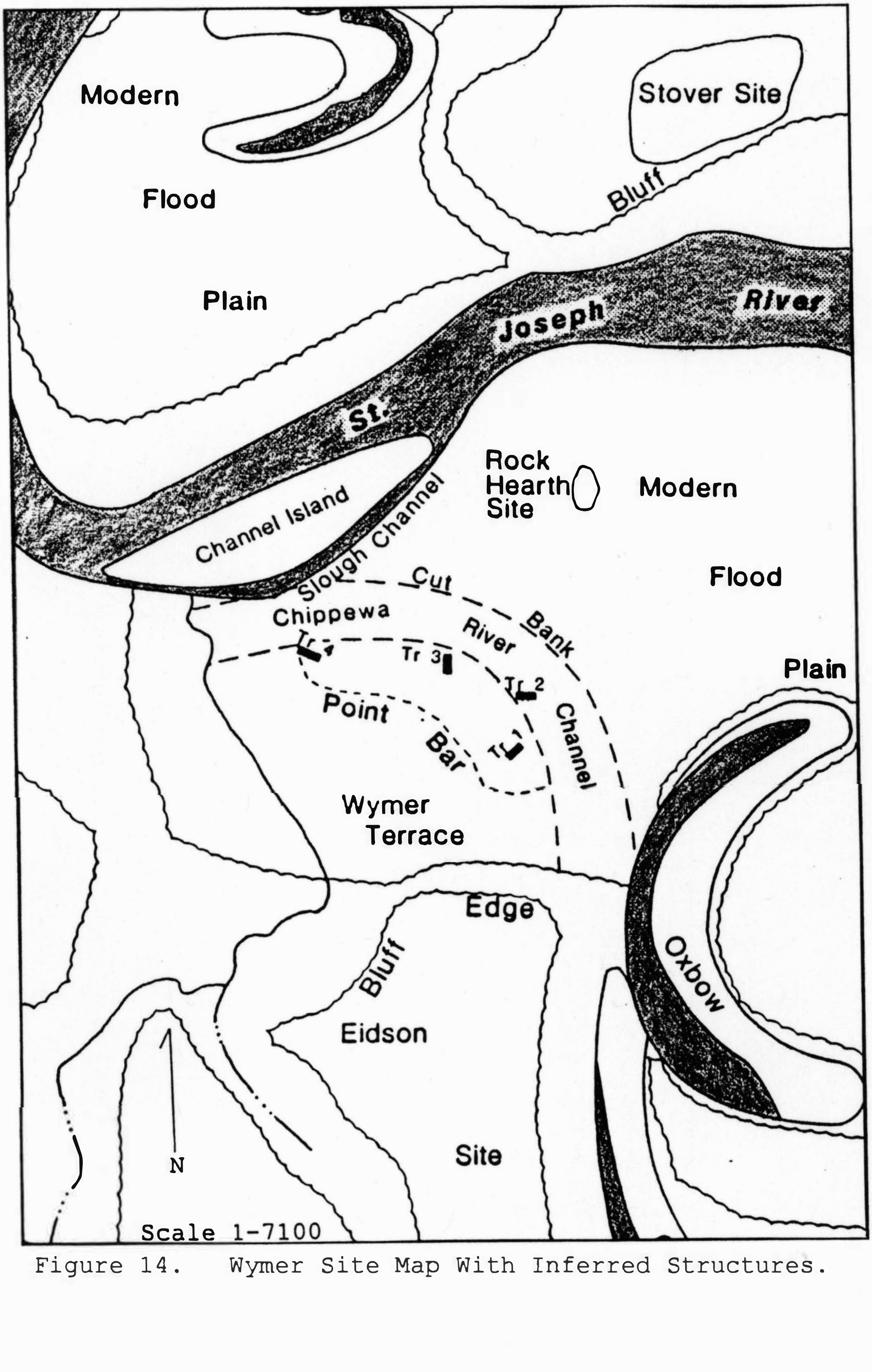


Figure 14. Wymer Site Map With Inferred Structures.

4. The organic sediments from Trench 2 were ^{14}C dated to 6,920 years B.P. and 5,500 years B.P. at 4.5 and 3.5 meters below surface, respectively. These dates place material 3.5 meters below the present land surface at just before the Nipissing transgression and material 4.5 meters deep at midway between the Olson forest bed inundation and the Nipissing stage. This further implies that a period of aggradation followed the downcutting episode in order to deposit the fill seen in trench 2.

5. The present St. Joseph River, at its current base level of 177 meters (580 feet), appears to be in a graded condition. The 3740 years B.P. ^{14}C date from the Rock Hearth site showed that little deposition has occurred on the St. Joseph River's modern floodplain since the Algoma lake stage. Hence, the present grade and base level appear incapable of significant erosion or deposition.

According to presently accepted data, the period between the Main Algonquin and Nipissing stages (which includes the Chippewa low stage) lasted almost 5000 years. Of the low lake stages prior to Chippewa, the Cary-Port Huron and Twocreekan substages were less than a fifth as long-lived and may not have been significantly lower than the present. All of the other previous lake stages were higher than the present (Figure 13). Three possibilities exist for the period when the Wymer point bar could have been deposited. The point-bar building episode could have occurred during the Algonquin-Chippewa transition, Cary-Port Huron interstade or the Twocreekan interstade.

The older ^{14}C date of the cutoff sediments from Trench 2 was about midway between the Olson forest bed and the Nipissing stage. Sediments from the bottom of the Wymer channel were not found during sampling. The relation of the channel bottom to the ^{14}C date is unknown. Oxbow channels fill over a period of time that is dependent on the rate of accumulation of organic detritus and whether it receives overbank flood deposits. It can be said with certainty only that the deposit must pre-date 6920 years B.P.

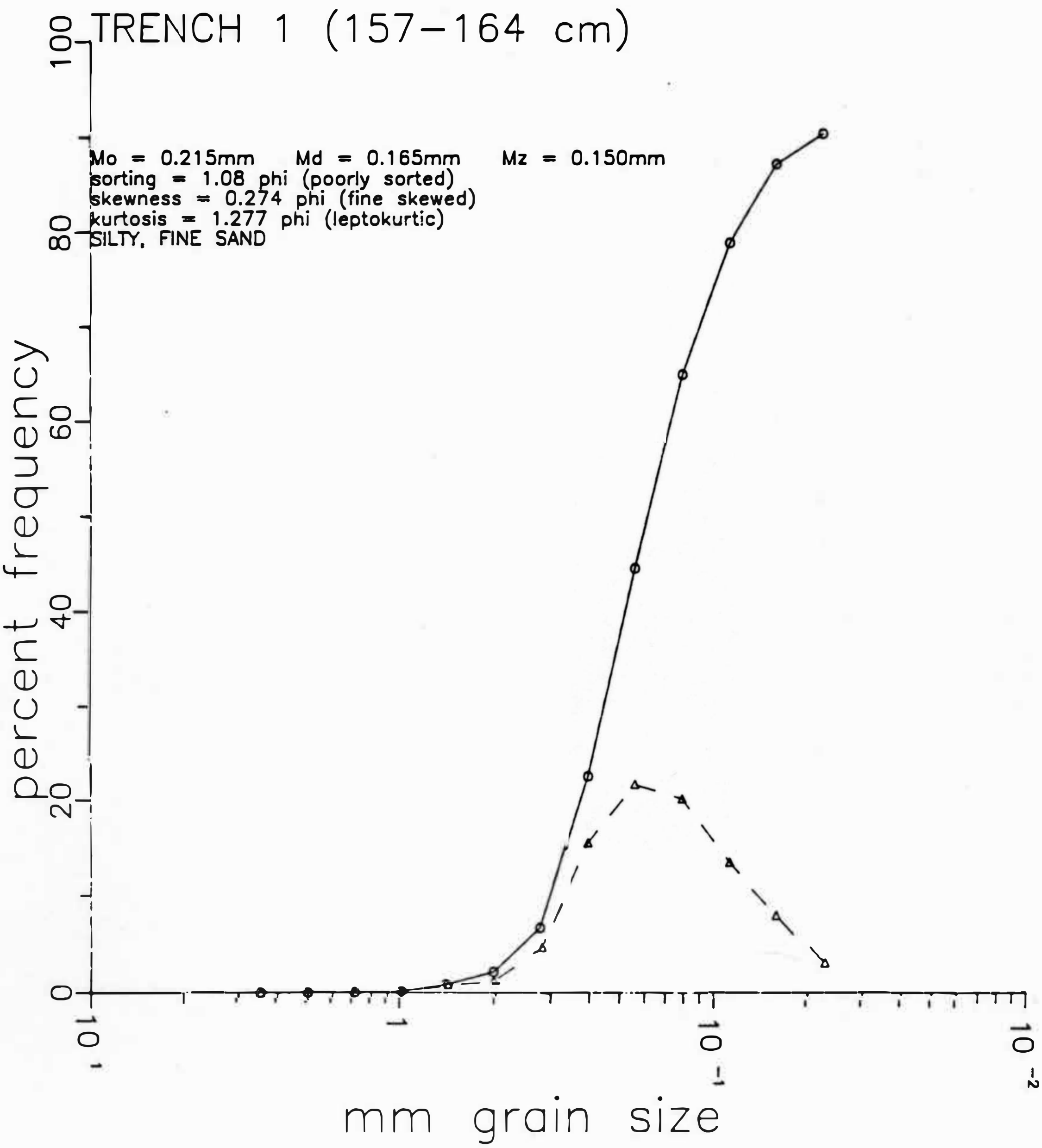
Assuming this, the Wymer site has existed since early Archaic time and therefore, was available for occupation throughout the entire Archaic cultural period. Because point bars are high energy depositional environments, any site that existed before the point bar was deposited (Paleo-Indian) would have been unrecognizably reworked by the river.

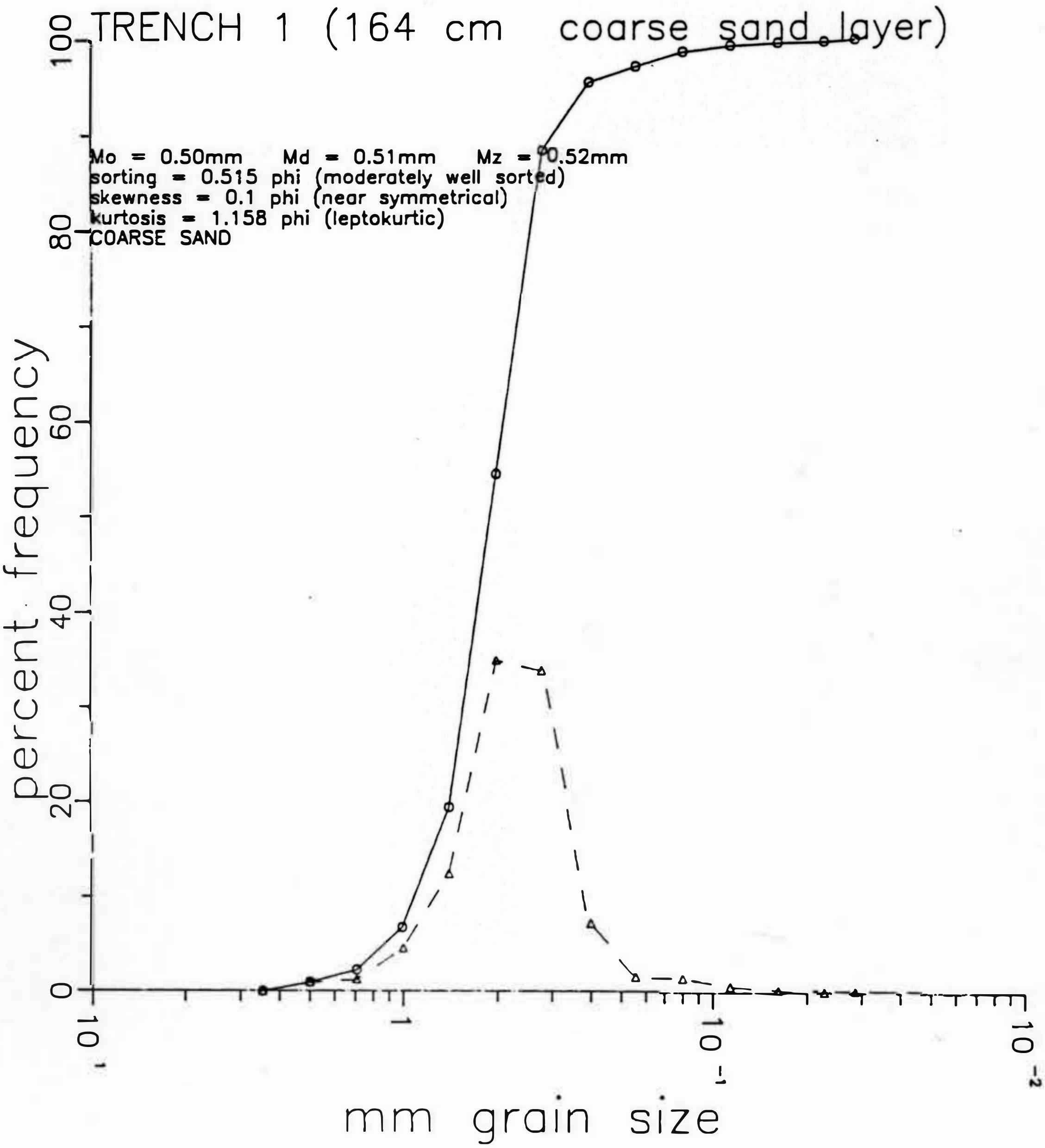
In the writer's opinion however, it would be an error to assume that the Wymer point bar was still active during early Archaic time. The first segment of the St. Joseph River valley was interpreted above as the product of headward erosion into the inside edge of the Valparaiso Moraine. Given that the present river appears to be graded, the energy conditions required to produce the erosion documented in this study must have ended before the Nipissing transgression reached the 177 meter elevation. In fact, the ^{14}C dates tell us that the river had returned to a state of aggradation by 6920 years B.P. An elevation versus time line from the Olson site to the Nipissing stage shows a rise of about 1 meter/106 years. This yields an elevation of 166 meters for the base level at 6920 years B.P. Whether the river could aggrade at that base level is unclear given that our present understanding of the river is clouded by both its dam-controlled state and a lack of any other studies of the river valley stratigraphy.

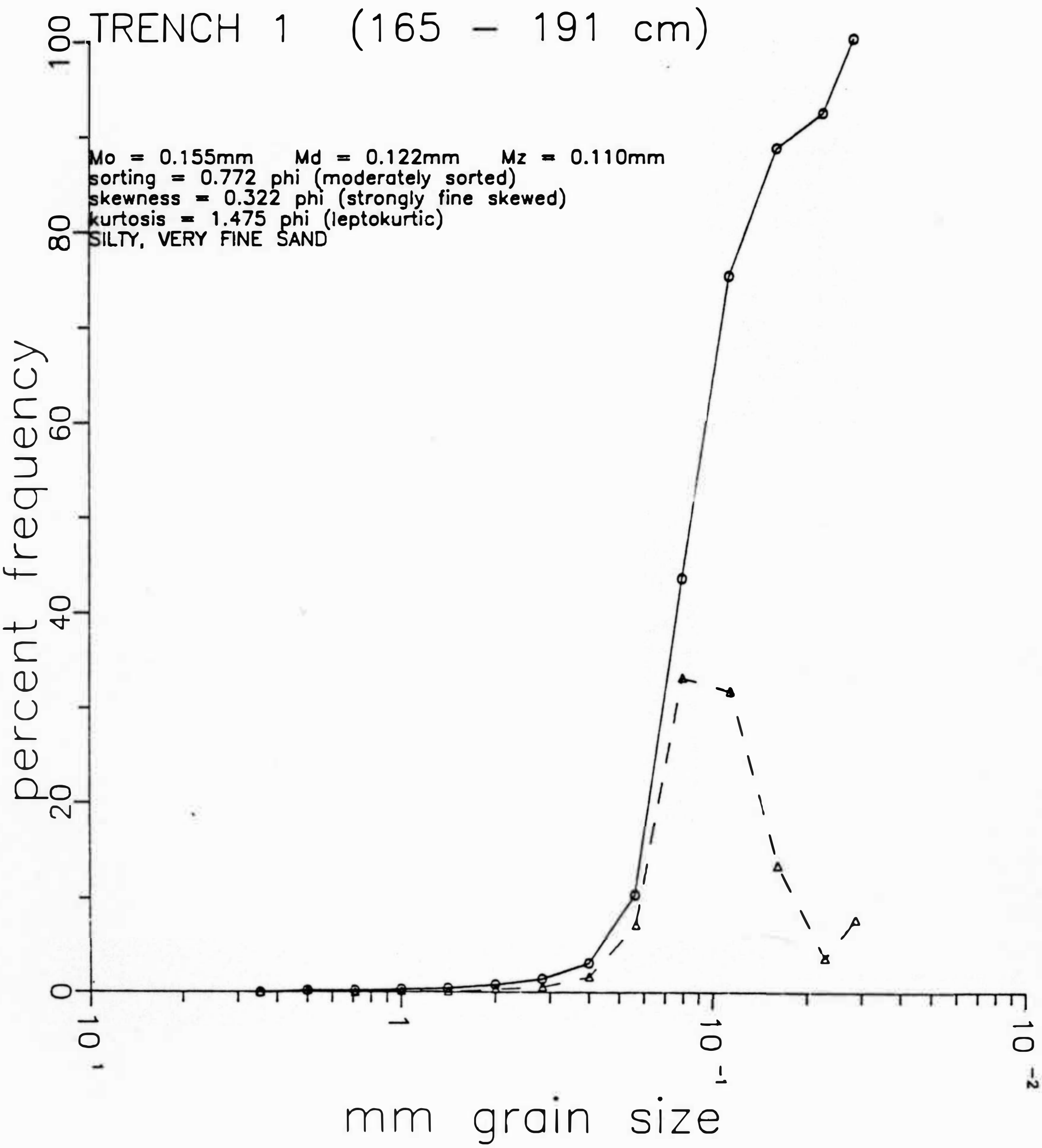
It may be that each of the three low lake stages of the last 14,000 years contributed to the headward erosion. However, the Chippewa low stage was certainly the dominant influence given its duration and its very low elevation. It is far more likely that the Wymer terrace was built during the early part of the Chippewa regression. The present floor of the valley was attained by the early post-Algoma stage and had been aggrading for at least 1,900 years before the Nipissing stage.

Appendix A

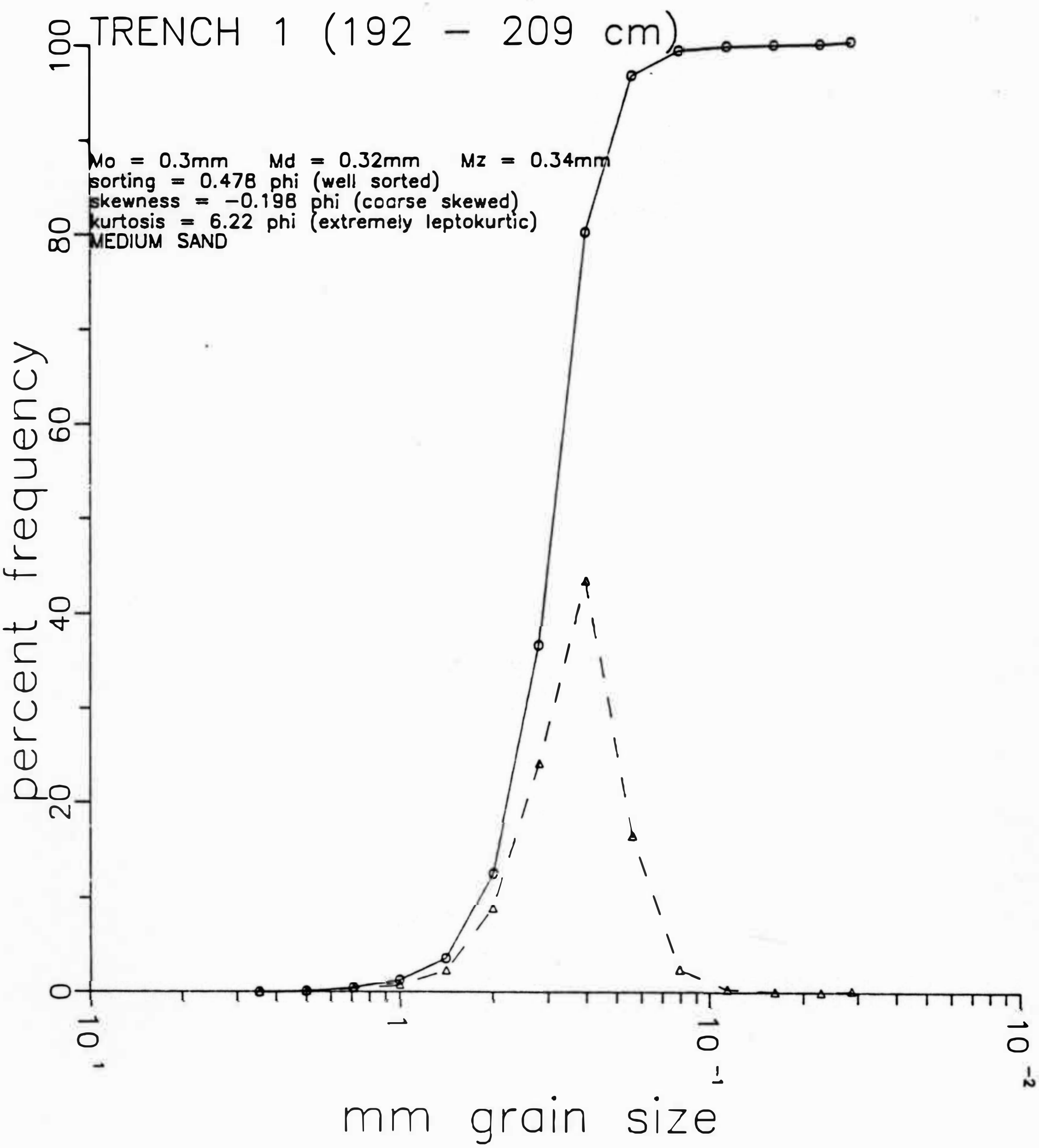
Grain-Size Distribution Curves of Soil Samples

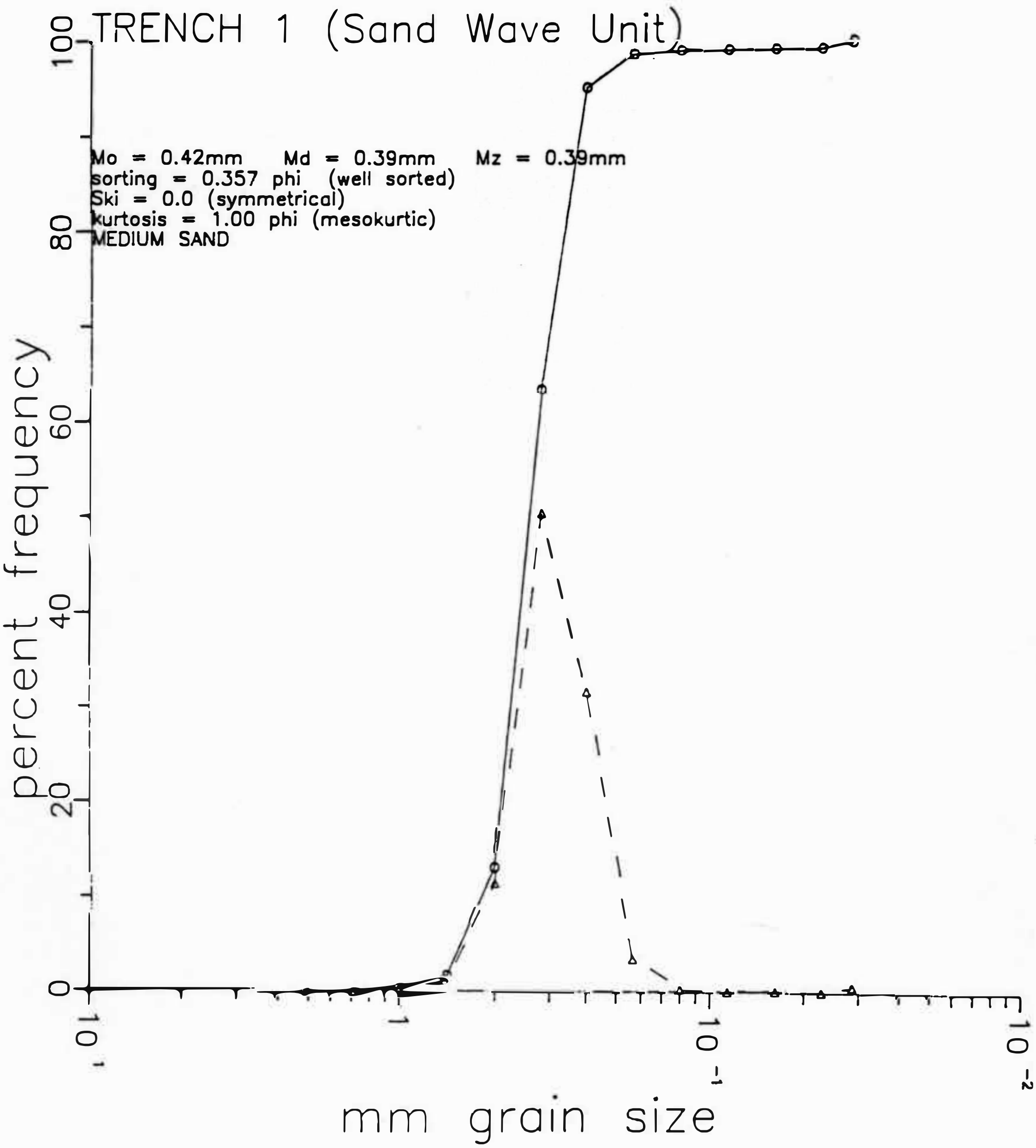

TRENCH 1 (157–164 cm)
Mo = 0.215mm Md = 0.165mm Mz = 0.150mm
sorting = 1.08 phi (poorly sorted)
skewness = 0.274 phi (fine skewed)
kurtosis = 1.277 phi (leptokurtic)
SILTY, FINE SAND
percent frequency
100
80
60
40
20
0
10^1
1
10^{-1}
10^{-2}
mm grain size

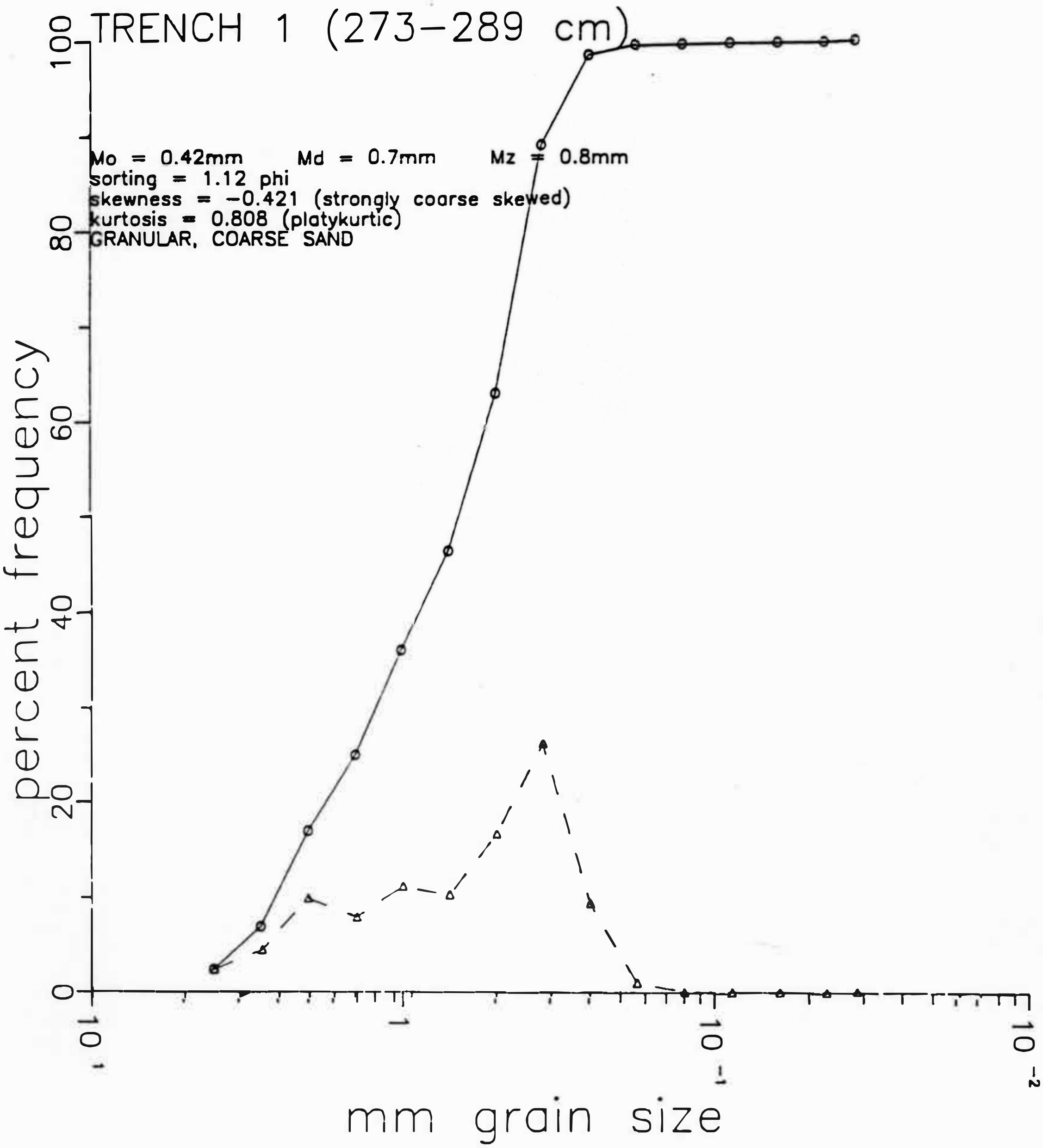

TRENCH 1 (164 cm coarse sand layer)
Mo = 0.50mm Md = 0.51mm Mz = 0.52mm
sorting = 0.515 phi (moderately well sorted)
skewness = 0.1 phi (near symmetrical)
kurtosis = 1.158 phi (leptokurtic)
COARSE SAND
percent frequency
0
20
40
60
80
100
10 1
1
10 -1
10 -2
mm grain size



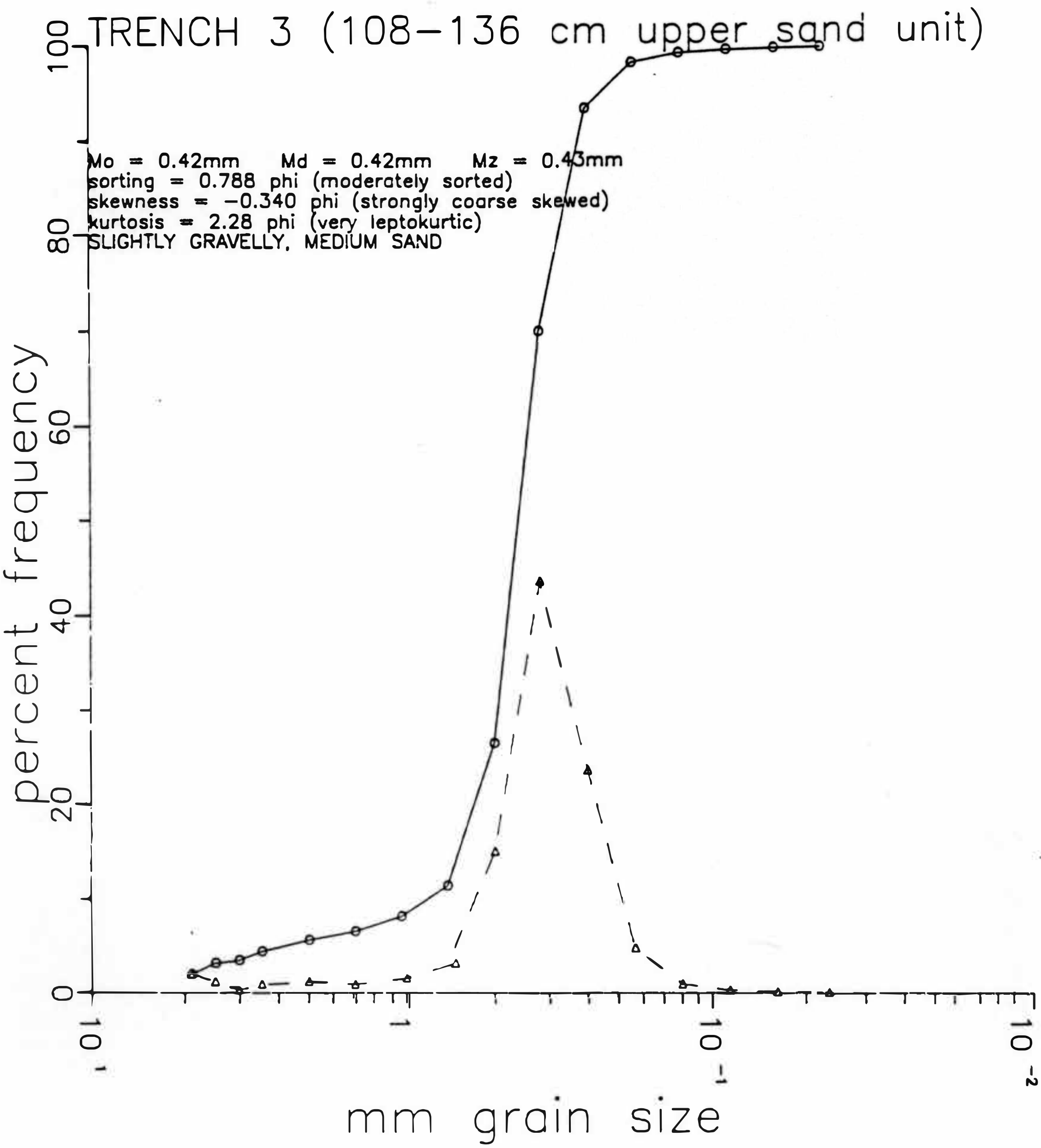
TRENCH 1 (165 – 191 cm)
Mo = 0.155mm Md = 0.122mm Mz = 0.110mm
sorting = 0.772 phi (moderately sorted)
skewness = 0.322 phi (strongly fine skewed)
kurtosis = 1.475 phi (leptokurtic)
SILTY, VERY FINE SAND
percent frequency
0
20
40
60
80
100
10 1
1
10 -1
10 -2
mm grain size

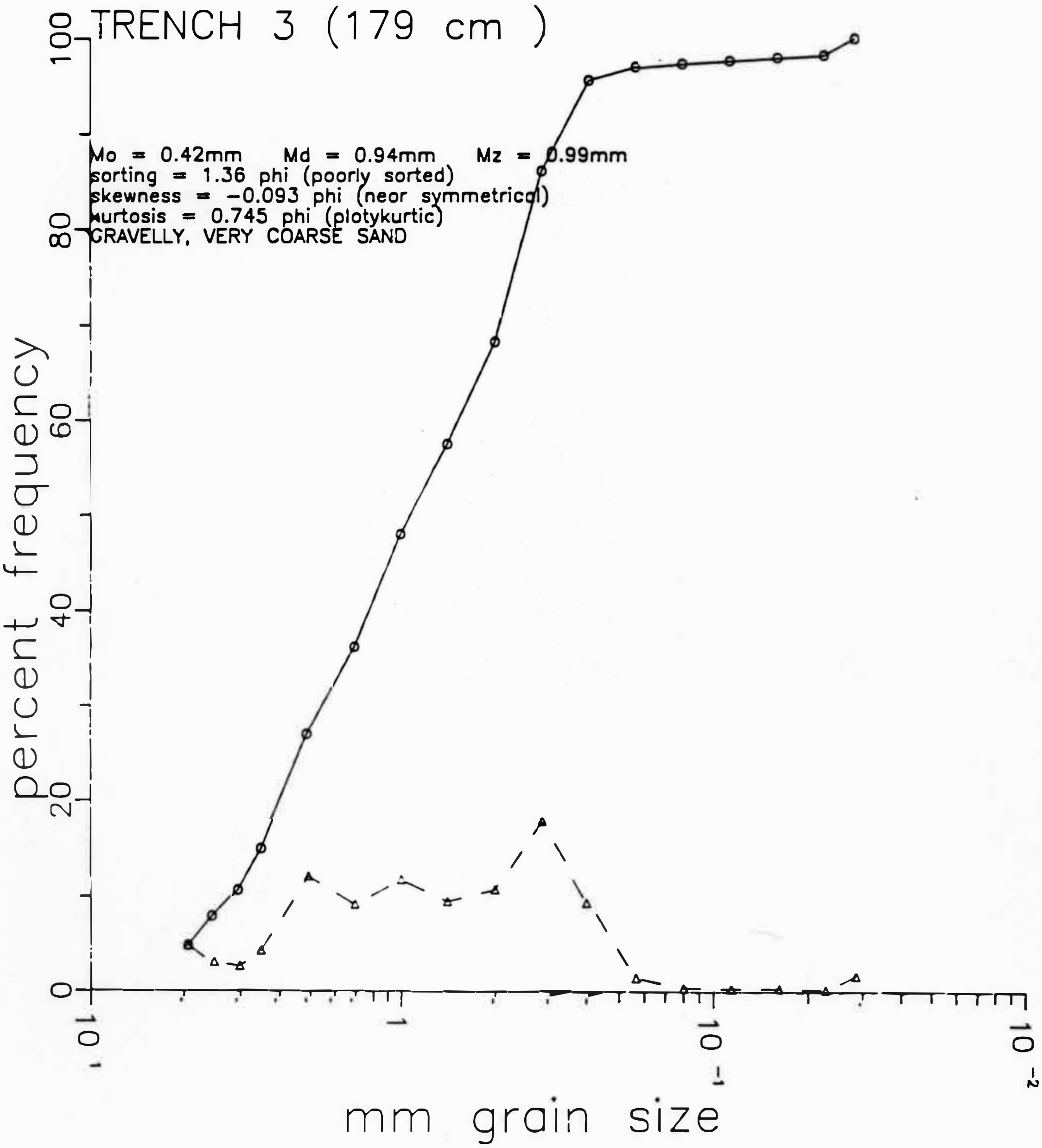

TRENCH 1 (192 - 209 cm)
Mo = 0.3mm Md = 0.32mm Mz = 0.34mm
sorting = 0.478 phi (well sorted)
skewness = -0.198 phi (coarse skewed)
kurtosis = 6.22 phi (extremely leptokurtic)
MEDIUM SAND
percent frequency
0
20
40
60
80
100
10^{1}
1
10^{-1}
10^{-2}
mm grain size


TRENCH 1 (Sand Wave Unit)
Mo = 0.42mm Md = 0.39mm Mz = 0.39mm
sorting = 0.357 phi (well sorted)
Ski = 0.0 (symmetrical)
kurtosis = 1.00 phi (mesokurtic)
MEDIUM SAND
percent frequency
0
20
40
60
80
100
10 1
1
10 -1
10 -2
mm grain size

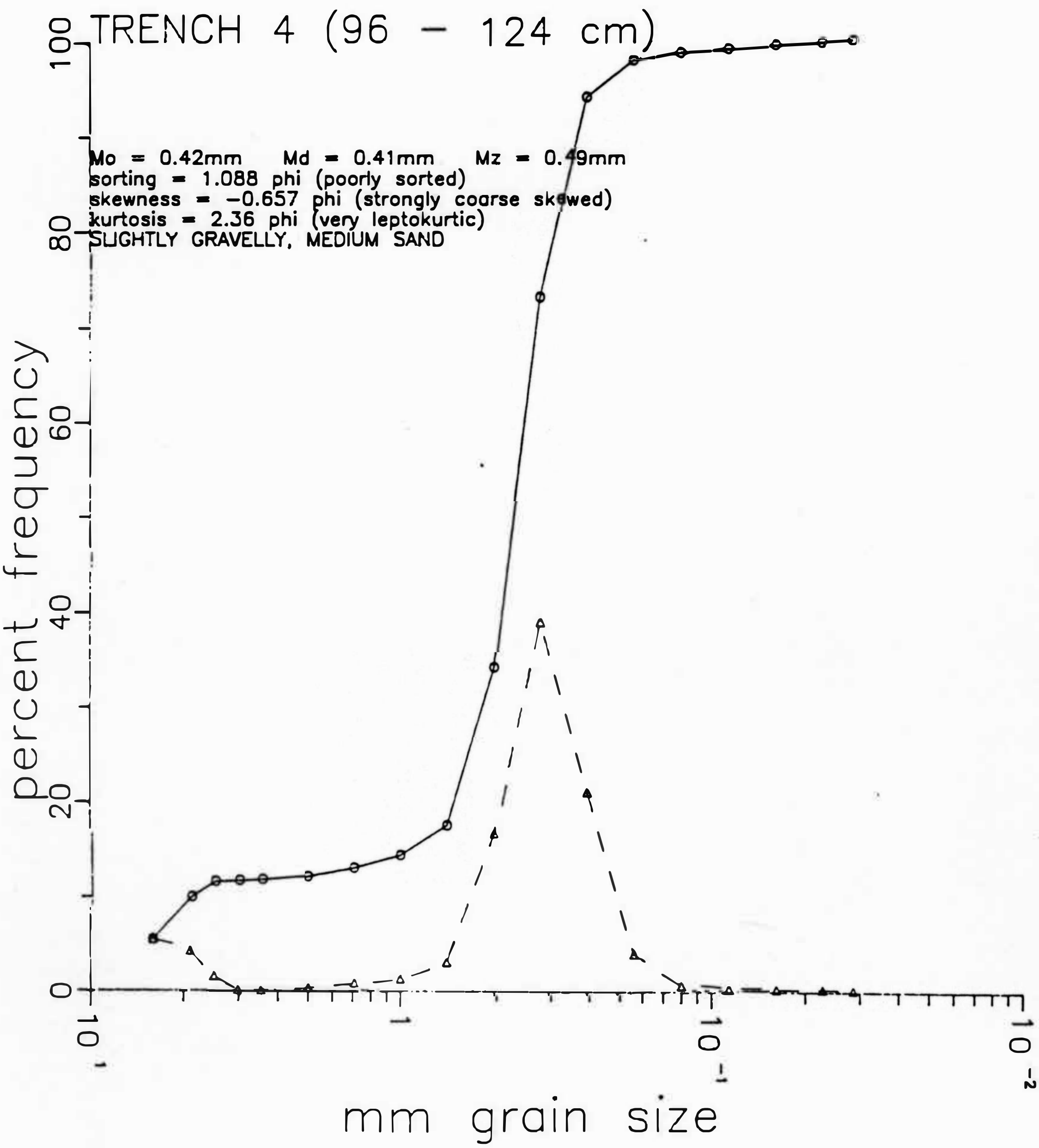


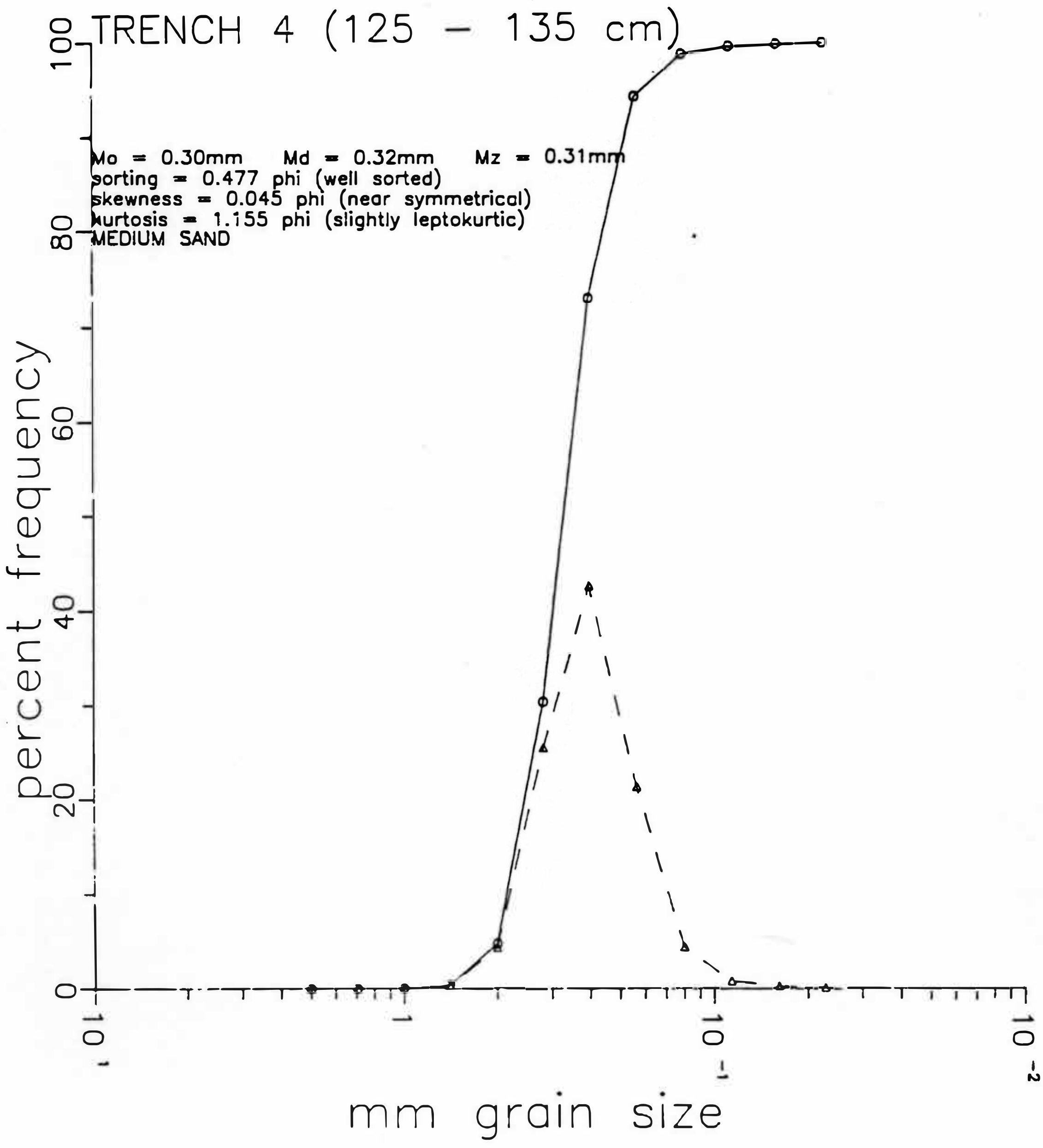
TRENCH 1 (273–289 cm)
Mo = 0.42mm
Md = 0.7mm
Mz = 0.8mm
sorting = 1.12 phi
skewness = −0.421 (strongly coarse skewed)
kurtosis = 0.808 (platykurtic)
GRANULAR, COARSE SAND
percent frequency
0
20
40
60
80
100
10 1
1
10 -1
10 -2
mm grain size

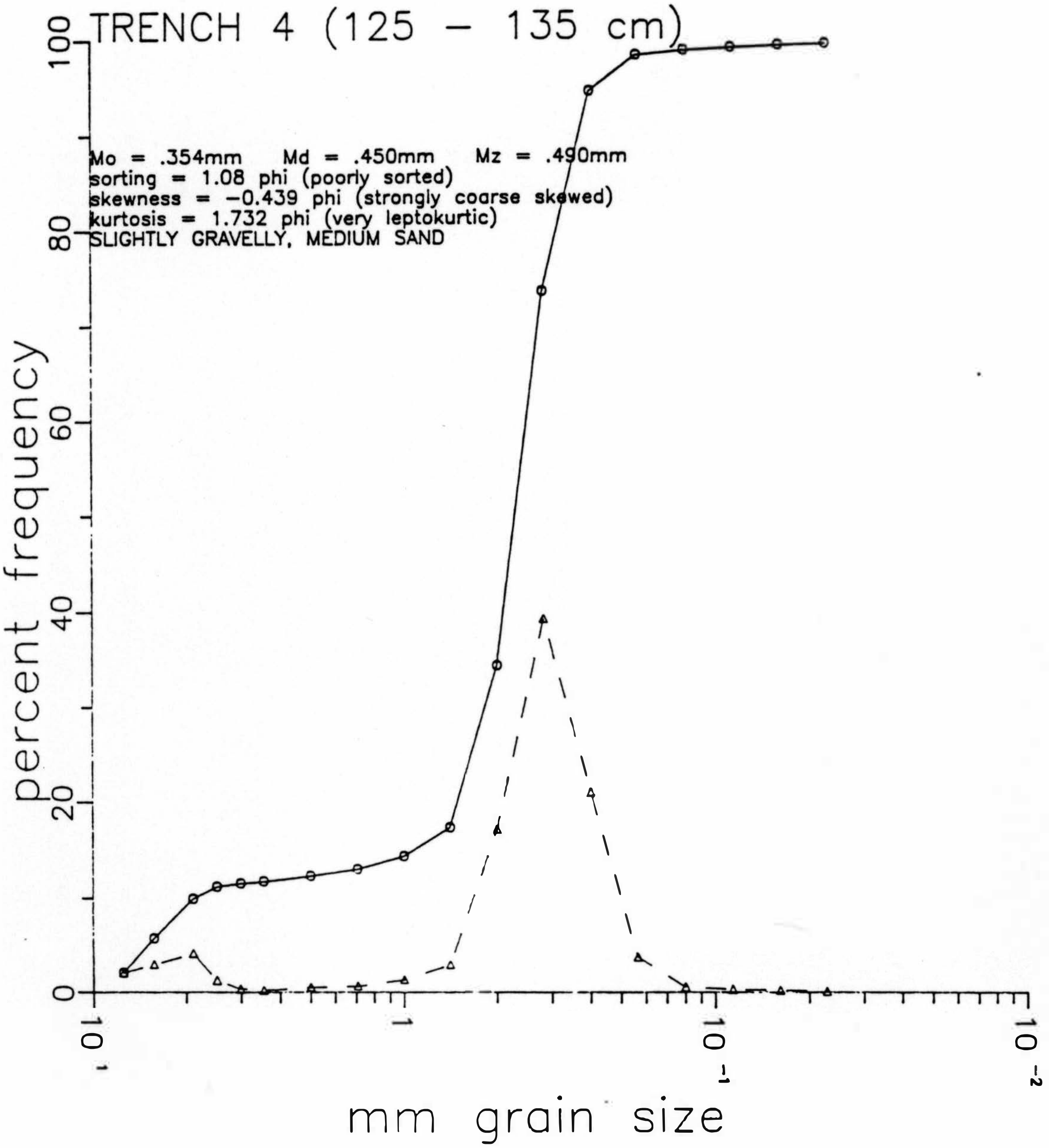

TRENCH 3 (108–136 cm upper sand unit)
Mo = 0.42mm Md = 0.42mm Mz = 0.43mm
sorting = 0.788 phi (moderately sorted)
skewness = −0.340 phi (strongly coarse skewed)
kurtosis = 2.28 phi (very leptokurtic)
SLIGHTLY GRAVELLY, MEDIUM SAND
percent frequency
0
20
40
60
80
100
10 ¹
1
10 ⁻¹
10 ⁻²
mm grain size

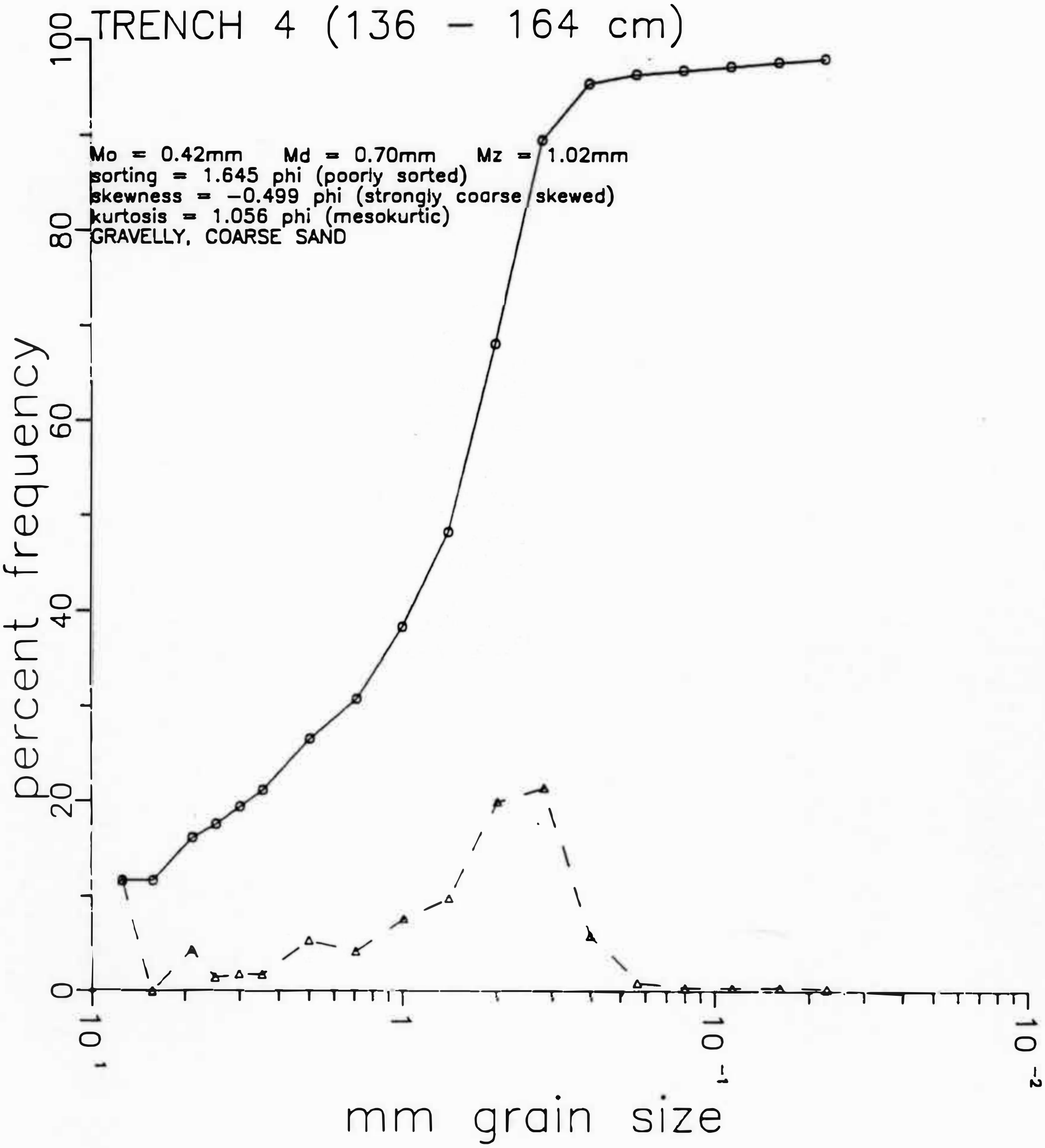


TRENCH 3 (179 cm)
Mo = 0.42mm Md = 0.94mm Mz = 0.99mm
sorting = 1.36 phi (poorly sorted)
skewness = -0.093 phi (neor symmetrical)
kurtosis = 0.745 phi (plotykurtic)
GRAVELLY, VERY COARSE SAND
percent frequency
100
80
60
40
20
0
10 1
1
10 -1
10 -2
mm grain size


TRENCH 4 (96 – 124 cm)
Mo = 0.42mm Md = 0.41mm Mz = 0.49mm
sorting = 1.088 phi (poorly sorted)
skewness = –0.657 phi (strongly coarse skewed)
kurtosis = 2.36 phi (very leptokurtic)
SLIGHTLY GRAVELLY, MEDIUM SAND
percent frequency
0
20
40
60
80
100
10^1
1
10^{-1}
10^{-2}
mm grain size


TRENCH 4 (125 – 135 cm)
Mo = 0.30mm Md = 0.32mm Mz = 0.31mm
sorting = 0.477 phi (well sorted)
skewness = 0.045 phi (near symmetrical)
kurtosis = 1.155 phi (slightly leptokurtic)
MEDIUM SAND
percent frequency
0
20
40
60
80
100
10 1
1
10 -1
10 -2
mm grain size


TRENCH 4 (125 – 135 cm)
Mo = .354mm Md = .450mm Mz = .490mm
sorting = 1.08 phi (poorly sorted)
skewness = –0.439 phi (strongly coarse skewed)
kurtosis = 1.732 phi (very leptokurtic)
SLIGHTLY GRAVELLY, MEDIUM SAND
percent frequency
0
20
40
60
80
100
10 1
1
10 -1
10 -2
mm grain size


TRENCH 4 (136 – 164 cm)
Mo = 0.42mm Md = 0.70mm Mz = 1.02mm
sorting = 1.645 phi (poorly sorted)
skewness = –0.499 phi (strongly coarse skewed)
kurtosis = 1.056 phi (mesokurtic)
GRAVELLY, COARSE SAND
percent frequency
0
20
40
60
80
100
10 1
1
10 -1
10 -2
mm grain size

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