LATE GLACIAL and EARLY HOLOCENE HISTORY OF THE GLACIAL LAKES AITKIN AND UPHAM BASIN, NORTH-CENTRAL MINNESOTA: IMPLICATIONS FOR THE TIMING OF POST GLACIAL EOLIAN ACTIVITY

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Abstract

Dune formation throughout Minnesota has been attributed to mid-Holocene aridity, however, the environmental setting of dunes in the Glacial Lake Aitkin and Upham basin suggest a Late Glacial and Early Holocene origin that is closely linked with exposure of source sands rather than aridity. Grigal et al. (1976) dated buried soils within a sequence of eolian sand near Lake Winnibigoshish, Minnesota, and identified several phases of eolian activity. These episodes were related to cycles of aridity and lake level fall during the Middle Holocene. Based on the work of Grigal et al.(1976), subsequent studies of eolian activity in Minnesota (Keen and Shane, 1990; Dean et al, 1996; Dean, 1997) concluded that all dunes in Minnesota are a result of Middle Holocene aridity.

Glacial Lakes Aitkin and Upham occupied a basin in north-central Minnesota bounded on the north by the Giants Range and to the east, south, and west by hummocky moraines of the Rainy and Superior Lobes, and the St. Louis sublobe. Dune clusters occur sporadically throughout the Glacial Lakes Aitkin and Upham basin and occur in areas with abundant sources of fine sand. Granulometry indicates a 4φ grain size signature characterizes most dunes in the basin. Maximum dune amplitude is ~5 meters and dune morphologies suggest northwesterly winds. A sediment core collected from Hay Lake (93°W,52°N), located within a dunefield at the edge of Glacial Lake Upham, records three prominent peaks in whole-core magnetic susceptibility between 10,100 and 6,600 yr BP. All dates referred to this paper are in uncalibrated ^{14}C years B.P.. No clastic input is evident after 6,600 yr B.P., suggesting dune stability. Eolian events recorded in the core are interpreted as eolian activity that resulted from episodic lake drainage and exposure of abundant source sediment during the late Glacial and Early Holocene. The timing of dune formation within the basin has important implications for other dune fields throughout Minnesota.

Using a Digital Elevation Model (DEM) the elevation of lake basin was adjusted for isostatic rebound based on the highest lake level, then tilted incrementally through several stages to assess relations among beaches, inlets, and outlets over time. The lake eventually drained, which led to the dune formation on littoral areas of exposed source sand.

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Acknowledgements

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Chapter 1 Introduction

The Glacial Lakes Aitkin and Upham basin in north-central Minnesota contains numerous dune colonies. These dunes colonies are but a few of the numerous eolian deposits throughout Minnesota and the Midwest (Fig. 1-2). Despite well over a century of research on Minnesota's glacial landscape these fields of fossil sand dunes have received relatively little attention. Although described by Hall and Sardeson (1898), Elftman(1898), Upham(1896), Winchell (1896), Leverett and Sardeson (1917,1919) and Leverett (1932), the first comprehensive work on Minnesota's eolian landscape was that of Cooper (1935, 1938). Cooper examined the dunes of the Anoka Sand Plain and those between Brainerd and St. Cloud along the Mississippi River and gave brief mention of those along the St. Croix River. He concluded that the aforementioned dunes are located on areas of fine to very fine sand that provided the sediment source. Cooper (1935) also suggested that proximity to a river functioned to draw down the water table to make source sands available for eolian transport. He initially suggested a post-glacial timing of dune formation similar to the other aforementioned researchers, however, later acknowledged the possibility of Holocene climate-driven eolian activity (Cooper, 1938).

In a more recent study, Grigal et al. (1976) examined a field of sand dunes along the southeast shore of Lake Winnibigoshish (Fig. 2) and described several buried soil horizons. Radiocarbon dates on two of these horizons suggested episodic dune formation from before 7910 to after 5040 BP during the Middle Holocene (Grigal et al., p. 1252). The formation of these dunes was related to lake-level lowering caused by Middle Holocene aridity and subsequent exposure of nearshore lacustrine sediments to eolian activity (Grigal et al., 1976). The mid-Holocene climate in the upper Midwest was considerably warmer and drier than the modern climate (Webb et al., 1983; Dean et al., 1984; Dean et al., 1996; Bartlein et al., 1984; COHAP, 1988) and Grigal et al. (1976, p.1254) speculated that the dunes of the Anoka Sand Plain may also be Mid-Holocene in age. This idea was explored in detail by Keen (1985) and Keen and Shane (1990) in their paleoenvironmental investigation of Ann Lake located on the Anoka Sand Plain (Fig. 2). From analysis of a lake sediment core they concluded that there was a relatively continuous record of eolian activity throughout the Middle Holocene; a reversal of the ideas of earlier workers (Cooper, 1935; Hall and Sardeson, 1899). A profile of whole-core magnetic susceptibility from the sedimentary record of Ann Lake records increased clastic sediment input from about 8000 until 4500 BP with discrete peaks in clastic input ca. 7400, 5800, and 4900 BP (Keen & Shane, 1990). The clastic sediment is primarily silt and very fine sand. Keen and Shane (1990) interpret this clastic input to be eolian, however, they discuss at length the difficulty distinguishing eolian sediment from that deposited by lacustrine processes. Despite those difficulties Keen and Shane (1990) conclude that the increased clastic input to Ann Lake records widespread dune formation on the Anoka Sand Plain during the Middle Holocene. The dunes were thought to have formed as a result of landscape destabilization due to Middle Holocene aridity during the eastward extension of the Prairie/Forest border (Fig. 2).

The difference between water table drawdown and landscape destabilization, both caused by aridity, may seem subtle but it is important. Although the Prairie ecotone shifted to the east in the Middle Holocene a prairie ecotone in itself does not constitute an environment conducive to widespread landscape destabilization. For example, in the central Dakotas today there is a prairie environment and ample sand source, but no widespread dune formation because of stabilizing vegetation. To get the conditions necessary for landscape destabilization and dune formation, an existing Prairie environment must be further dried as occurred in the Nebraska Sand Hills (Winspear and Pye, 1996). Furthermore, whereas Grigal et al. (1978) directly dated soils between eolian events, Keen and Shane's magnetic susceptibility record from Ann Lake could be a record of local eolian activity, but it could also be a record of dryer dustier conditions of the Middle Holocene without widespread dune formation.

Additionally, the strong link between location of dunes thoughout Minnesota in relation to source sediment imply dune formation coincident with exposure of source. Conceptually, it seems logical that the optimum time for dune formation would be immediately following the exposure of sand source, prior to the establishment of vegetation (Seppala, 1993; Windspear and Pye, 1996). Similarly for the dunes on glaciolacustrine sediment, the optimum time of dune formation would be immediately following lake drainage and exposure of source.

The distribution and grain-size of dunes in the Aitkin and Upham basin combined with a 10,000 year magnetic susceptibility record from Hay Lake, which is located downwind of a prominent dune cluster in Lake Upham, indicate that dunes formed during the late-glacial/early-Holocene upon exposure of source sand as the lakes drained episodically. This episodic draining of the lakes promoted dune formation on littoral areas of the lake where fine to very fine sand was exposed. The episodic draining is recorded in the Hay Lake core. Peaks in magnetic susceptibility at 9,800 and 9,300 yr B.P. represent the eolian activity following the drainage of Lake Upham while the later peak, 7,400 yr BP, represents eolian activity immediately following the final drainage of Lake Aitkin. The two earlier peaks (9,800 and 9,300 yr B.P.) predate the Middle Holocene hypsothermal period. In addition, Hay Lake is located far enough north and east to be isolated from the dusty conditions of the Prairie, and does not have a continuous magnetic susceptibility proxy of eolian activity throughout the Middle Holocene such as that seen at Ann Lake. Hay Lake is located 250 km from the Prairie/Forest Border (Fig.2), which decreases the possibility of impact on the magnetic susceptibility record by long-distant transport of eolian sediment.

The results of this investigation suggest that the formation of the dunes on the Glacial Lakes Aitkin and Upham basin began during late glacial time and occurred episodically as the lakes drained. Dune formation was intimately related to exposure of extensive source areas of fine to very fine sand. Dunes form colonies that are coincident with source areas. The clusters are elongate in a northwest-southeast orientation suggesting predominantly northwest winds.

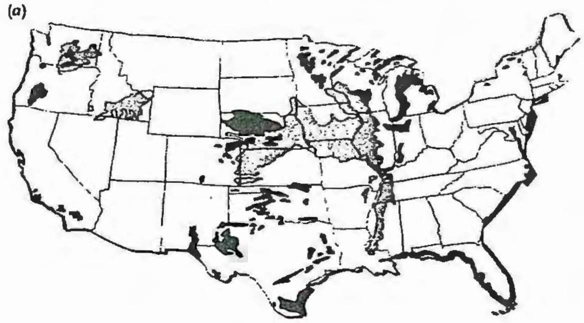


Figure 1. Dune and Loess Distribution (Brady and Weil, 2000 p. 42)
Black =Dune and Stippled = Loess.

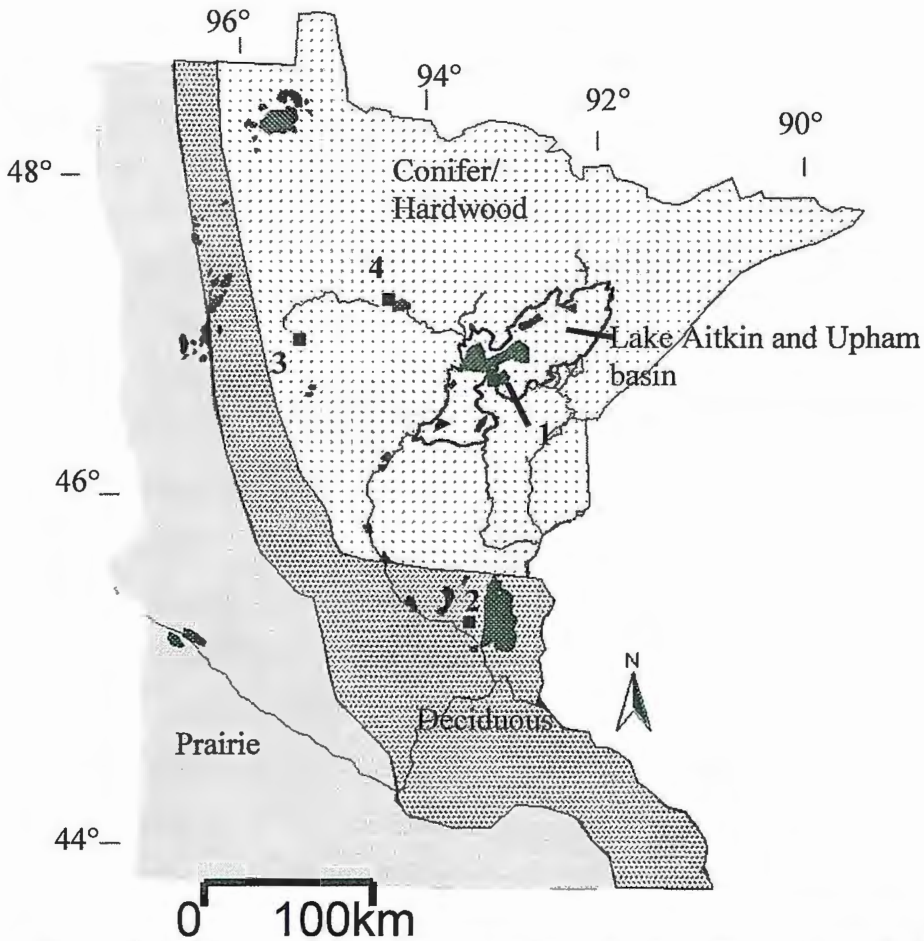


Figure 2. Minnesota vegetation map showing locations of dune fields and lakes referred to in this study. 1) Hay Lake 2) Lake Ann 3) Elk Lake 4) Lake Winnibigoshish. Filled areas are prominent dune colonies. (Modified from Keen and Shane,1990)

Chapter 2 Description of Study Area

2.0 Study Area: Lakes Aitkin and Upham

Glacial Lakes Aitkin and Upham occupied a basin of ~15,000 km^2 bounded by the Giants Range to the north and drift of the Superior and Rainy Lobe and St. Louis sublobe to the south, east, and west (Fig. 3). The glacial sediment varies in thickness up to 300 ft. with the greatest thicknesses within moraines that bound the lake. The basin lies within a bedrock low (305 m./ 1000 ft.) between the Giants Range on the north (518 m./ 1700 ft.) and the granites to the south (335 m./ 1100 ft.). The low resulted from the scouring of less resistant argillite and cretaceous sandstones and clays during repeated glacial advance throughout the Pleistocene (Wright, 1972; Hobbs and Goeble, 1982).

Glacial Lake Aitkin was first mapped and named by Warren Upham (1896), and Glacial Lake Upham was named by Winchell (1901) after the aforementioned scientist who had previously mapped much of the geology in the region. Although the basin was described by early geologists (Leverett and Sardeson, 1932) little detailed work was done. Baker (1965) studied glacial landforms of the St. Louis sublobe and used lake sediments to study paleoclimate in the region surrounding Spider Creek (Fig.4). Farnum et al. (1964) examined soils in the Aitkin basin (Fig. 4), and Hobbs (1983) worked out the sequence of ice advance and retreat and the history of the inlet and outlet elevations.

The highest lake levels in the northeast portion of the basin are marked by smooth wave-washed topography with only scattered lakeforms. The lake in this area was short-lived because of rapid isostatic rebound across the basin relative to the position of the outlet (Hobbs, 1983). Lake sediments range from sand and gravel in nearshore to clay in offshore environments, including rhythmite sequences in Lake Upham (Ballantine, 1991) and in Lake Aitkin (Alan Knaeble, Minnestoa Geological Survey, Personal Communication) The lake reached a maximum depth of 25 meters, with the deepest portion located in Lake Upham.

The inlets and outlets to the lakes varied throughout the lake's evolution as a function of the position of glacier ice and isosatatic rebound (Hobbs, 1983). There were two major inlets that were active at different times; Embarrass Gap, which entered the northeast part of Lake Upham and the Prairie River, which resulted in a large underflow fan in Lake Aitkin (Fig.4) (Hobbs, 1983). Both inlets stemmed from lakes located north of the Giants Range. Embarrass Gap inflow came from Lake Norwood through the Pike River, whereas, the Prairie River inflow came from Lake Koochiching (Fig. 4) (Lehr and Hobbs, 1992; Hobbs, 1983). Throughout most of the lakes' history drainage was through Lake Upham. All of the major outlets to the lakes are in Lake Upham, with the exception of the Mississippi River and possibly the Snake River in Lake Aitkin. Of the seven outlets in southeastern Lake Upham only the Uskabwanka and Chicken channels were active in the earliest stage of the lake because of stagnant ice in the basin (Fig. 4) (Hobbs, 1983). The others (Hellwig, Birch, Spider, and St. Louis channels) became active during later stages of the lake.

Dunes in the Aitkin and Upham basin are sporadically distributed throughout the basin. (Fig. 4) The dunes generally occur in elongate clusters trending NW-SE. The dunes have definite areas of dense and sparse distribution. The densest dune clusters are concentrated near the Prairie River inlet and the shallow sill between the Aitkin basin and the Upham basin, herein named the Swan River Sill (Fig. 4).

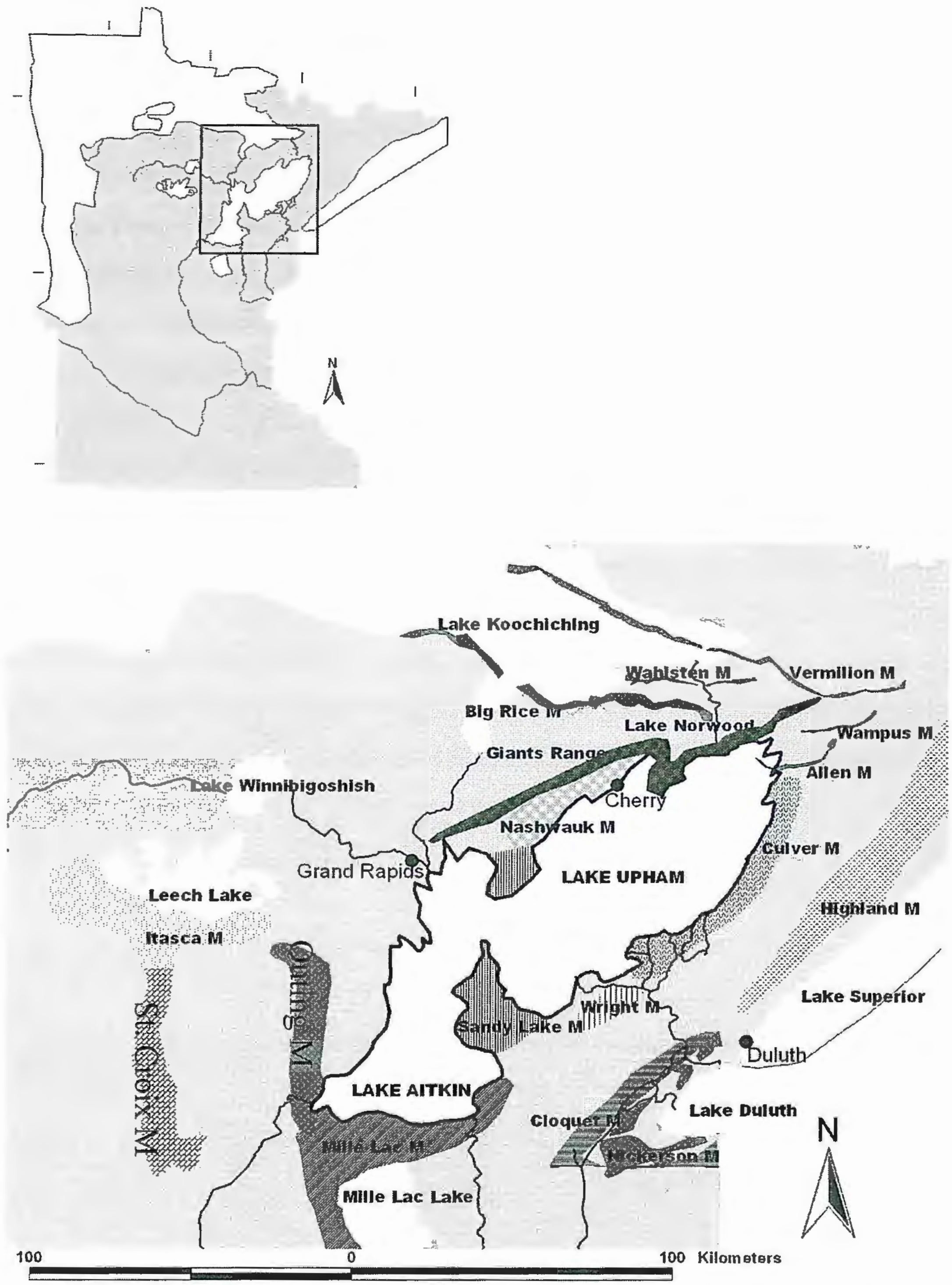


Figure 3. Study Area. Prominent moraines. Gray areas are undifferentiated. Black lines indicate inlets and outlets and modern day Mississippi River. Related paleolakes and modern lakes are outlined Modified from Mooers (1988) and Lehr and Hobbs (1992).

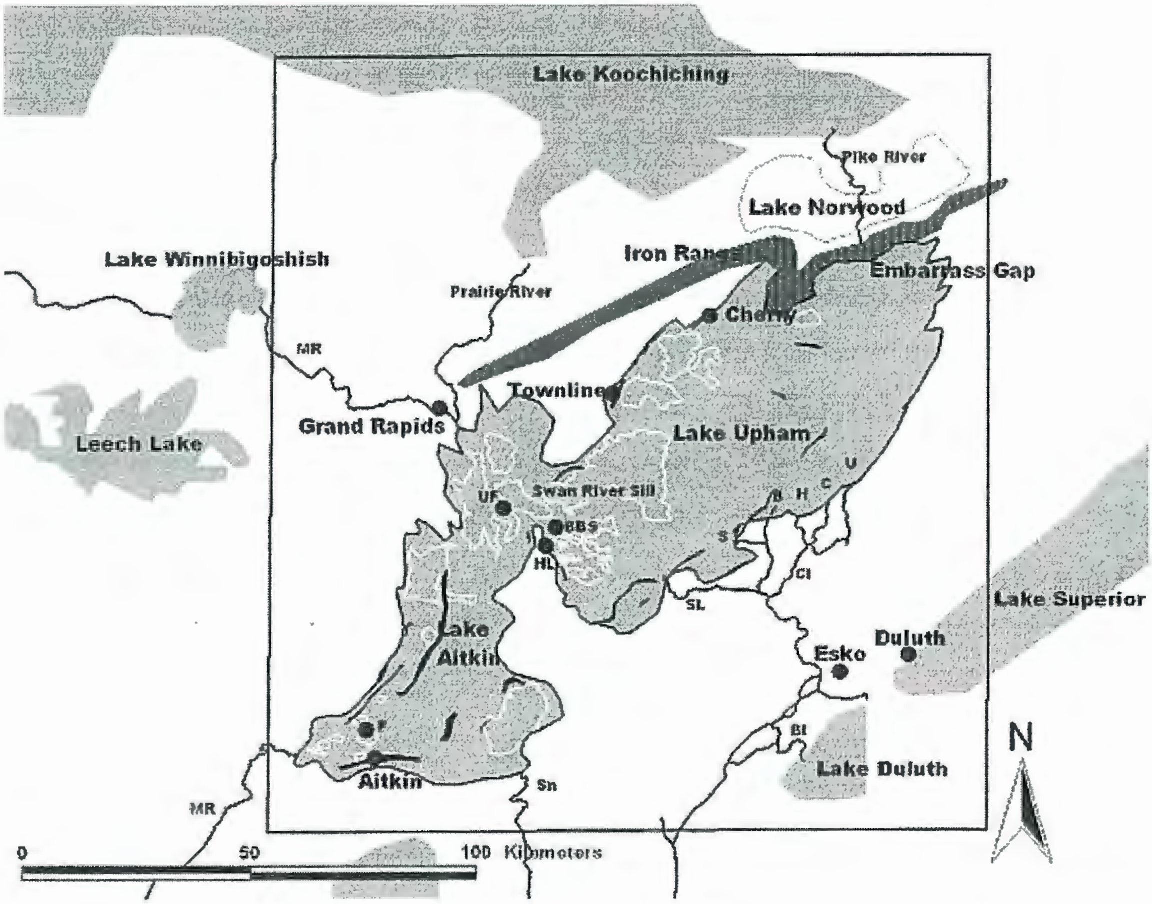


Figure 4. Major features of Lake Aitkin and Upham. Black solid polygons are beaches. White polygons are dune colonies. MR- Mississippi River; F-Farnum site; UF-Underflow fan site; BBS-Ball Bluff Spit; HL-Hay Lake. Outlet Channels are noted as follows (Modified from Hobbs, 1983): U-Uskabwanka; C-Chicken; H-Hellwig; B-Birch; S-Spider; SL- St. Louis River; Cl-Cloquet; Bl-Blackhoof; Sn-Snake.

2.1 Study Area: General Glacial History

Minnesota was glaciated repeatedly throughout the Pleistocene, however, the present geomorphology is largely a result of Late Wisconsinan glaciation. The glacial history and chronology of these deposits have been documented in numerous regional summaries (Wright, 1972; Bjork, 1988; Mooers and Lehr, 1997; Mooers, 1990; Mooers, 1988; Mooers and Dobbs, 1993; Clayton and Moran, 1982). During Late Wisconsinan glaciation the Rainy and Superior Lobes advanced from the northeast to their maximum at the St. Croix Moraine by about 16,000 yr B.P. (Clayton and Moran, 1982; Mooers and Lehr, 1997) (Fig.3). Other glacial landforms record the numerous ice-marginal positions that represent general ice recession that was periodically punctuated by ice readvances (Wright, 1972; Wright et al., 1973; Mooers 1990). Continued retreat of the Rainy Lobe was punctuated by readvances to the Outing and Sandy Lake Moraines (Mooers, 1988), while a readvance of the Superior Lobe resulted in the deposition of the Mille-Lacs Moraine. The Cromwell and Wright Moraines (Wright, 1972) (Fig. 3) are interlobate between the Superior and Rainy Lobe. As the Rainy Lobe retreated from those two prominent ice-marginal positions water ponded in front of the ice, which resulted in the first of two phases of Glacial Lake Aitkin and Upham (Wright, 1972).

The Rainy Lobe continued to retreat to a position south of the Giants Range in the northeast part of the Aitkin and Upham basin. At this time the St. Louis sublobe advanced across the Aitkin and Upham basin at least as far as Cherry, Minnesota (Fig. 3). This interpretation is based on an exposure in a gravel pit just north of Cherry, where subaqueously deposited Rainy Lobe outwash is overlain by St. Louis sublobe till that contains lenses of red Lake Upham I sediments (Fig. 5). The advance of the St. Louis sublobe from its parent Des Moines lobe from the northwest encountered the ice-cored topography of the Rainy Lobe, the Sugar Hills moraine. The ice was diverted southward into the Aitkin and Upham basin. Ice advanced to fill the basin and the limit is marked by the Culver moraine (Baker, 1965) on the periphery of the Aitkin and Upham basin. The advance of the St. Louis sublobe into the Aitkin and Upham basin was thought to have occurred at ~12,500 yr B.P. after the retreat of the Rainy Lobe to the Vermilion Moraine

(Wright, 1972; Clayton and Moran, 1982). However, the relationship at Cherry, and work by P.C. Larson (Personal Communication) indicate an earlier St. Louis sublobe advance.

The wastage of St. Louis sublobe ice in the basin combined with inflow from other glacial lakes to the north brought into existence the second phase of the lakes, Lake Aitkin and Upham II (Fig.4). As the Rainy Lobe retreated north of the Giants Range, Lake Norwood formed between the Giants Range and the retreating ice (Lehr and Hobbs, 1992; Hobbs, 1983; Bjork, 1988). Lake Norwood drained south through the Embarrass Gap and into Lake Upham (Hobbs, 1983). Meltwater flowing through the Embarrass Gap from Lake Norwood was diverted around the easternmost limits of the St. Louis sublobe and likely drained via the Uskabwanka and Chicken channels (Fig. 4). Stagnant ice in the basin affected the formation of outlets and water transport between lakes Aitkin and Upham. It is not known at what point the basin became completely ice-free. However, wastage of the ice of the St. Louis sublobe was likely fairly rapid, and lower outlets formed along the Birch, Spider, and Hellwig channels; all of these are tributary to the St. Louis River (Hobbs, 1983). As the St. Louis sublobe wasted further and isostatic rebound continued, the higher outlets were abandoned in favor of the lowest outlet along the St. Louis River (Fig. 4).

The Rainy Lobe and St. Louis sublobe continued to retreat north of the Giants Range. Lake Norwood grew in size still draining through the Embarrass Gap. The outlet cut down to an elevation of 1430 ft., which is referred to as Lake Koochiching. Lake Koochiching formed north of Lake Norwood and continued to drain south, along the Pike River, then through the Embarrass Gap. A lower outlet then opened for Lake Koochiching along the Prairie River, which carried water south and entered Lake Aitkin. The Prairie River was abandoned when lower outlets for Lake Koochiching opened further west into Lake Agassiz (Hobbs, 1983). Once the inflow along the Prairie River was abandoned, continued isostatic rebound resulted in the drainage of the lakes until they were eventually separated and drained through their respective outlets; Upham through the St. Louis River and Aitkin through the Mississippi River (Hobbs, 1983). As the lakes drained, littoral areas of the lake became exposed and resulted in dune formation, until vegetation stabilized the dunes.

Figure 5A. Subaqueously deposited Rainy outwash- Cherry, Minnesota.

Figure 5B. Glacial Lake Upham sediment in St. Louis sublobe till, overlies the Rainy Lobe outwash – Cherry, Minnesota.

2.2 Study Area: Post Glacial Vegetation Succession

As glacial ice retreated from the region the landscape was quickly colonized by vegetation. Although no pollen analysis was done as a part of this study, a clear understanding of the post glacial environment exists (Wright & Watt, 1969; Birks, 1981; Bjork, 1990). Generally an herb tundra existed until about 11,000 yr B.P. when spruce migrated into the area. Pine began to invade the area sometime around 10,000 yr B.P. followed quickly by deciduous species, primarily aspen and birch. Today the basin is covered with a veneer of peat, most of which is between 5,000 and 3,500 yr B.P. (Malterer, 1979) (Fig. 6). Other vegetation associated with the peat includes lowland conifers (black spruce, tamarack, and white cedar) and lowland hardwoods such as black ash. The dunes in the Aitkin and Upham basin in some locations form 'islands' in the peat ecosystem that support aspen, red pine, cedar and other mixed upland forest types. The shoreline and border material contains a range of aspen, birch and upland conifer forest and some white and red pine on the moraines of the border. Characteristic soil types that developed on this terrain include entisols and alfisols with the majority being peat. Alluvial soils are associated with major rivers. The region receives annual precipitation of 61cm in the northwest to 69 cm in the east. The modern drainage includes several major rivers which dissect the Aitkin and Upham basin and the Mississippi, St. Louis, Whiteface, East Swan River, Savanna River, Willow Rivers.

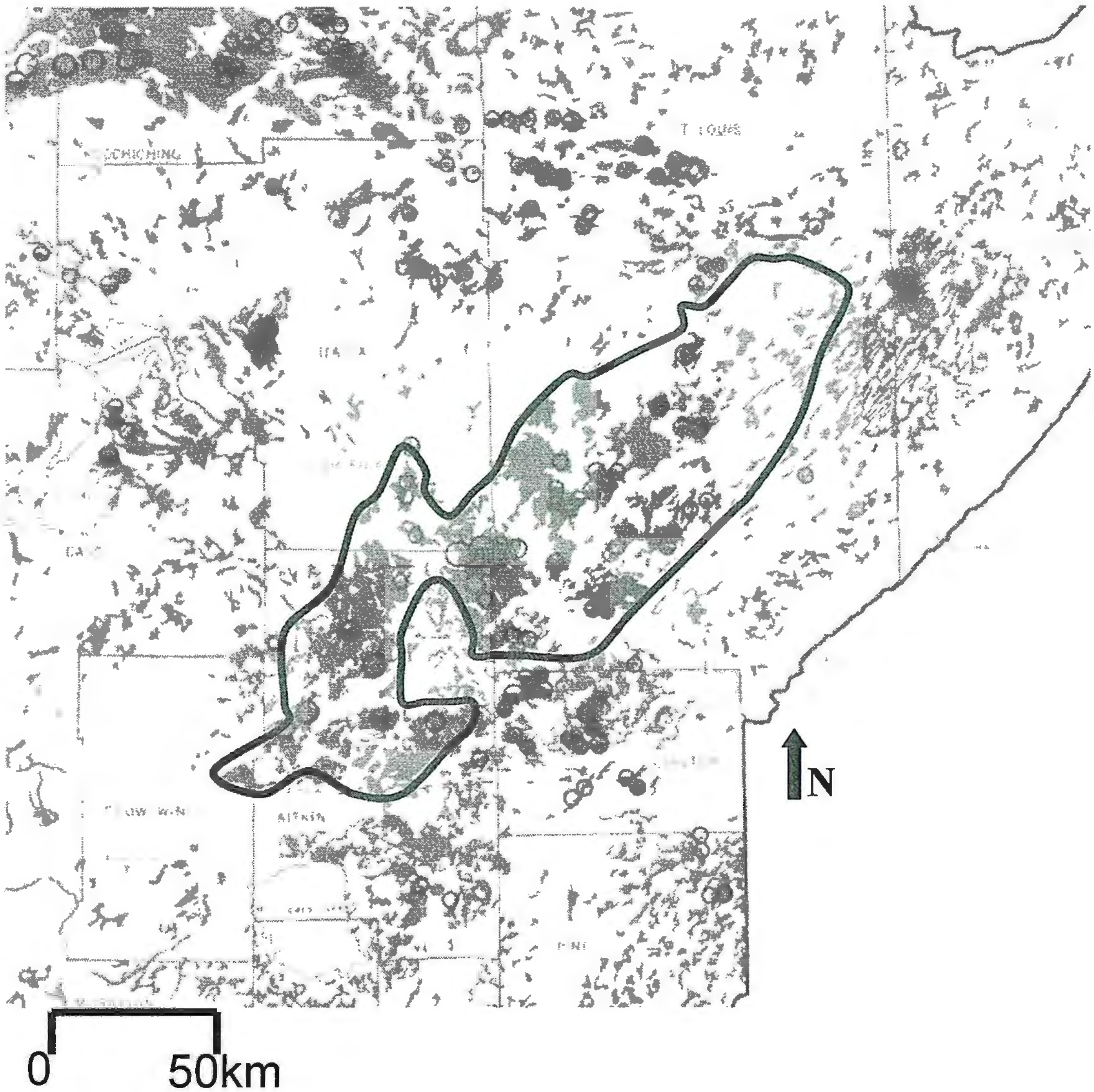


Figure 6. Shaded areas delineate peat deposits in Minnesota (Modified from Malterer, 1979) Lakes Aitkin and Upham delineated by bold line.

Chapter 3: Methods

3.0 Geomorphic Mapping and Field Work

The field area was outlined in summer 2001 by delineating geomorphic features such as dunes and shorelines from 1:24,000 and 1:250,000 scale topographic maps to gain a regional perspective of the basin and serve as the primary field map. Dunes were identified by locating recognizable common dune forms such as cresentic and longitudinal dunes. Shorelines were identified by a variety of different features including wave-cut scarps, wave-washed topography, and beaches. 291 field sites were visited and locations marked using GPS (Fig. 7). Descriptions were recorded at the field sites and landforms were interpreted as lake border, beaches, deep-water lake, and dunes. 118 samples collected for grain size analysis. Of the 118 samples were, 77 samples were of dune sand and 31 samples were till, beach sands or deep-water lacustrine clays. Where applicable, stratigraphic relationships, grain size, and sorting were described and photographs were taken. All samples were taken below the soil-forming horizon. The color of the tills was determined with a Munsell Color chart from moistened samples. Rock lithologies were noted at some localities. The sediments were tested for calcium carbonate by application of 10% HCl.

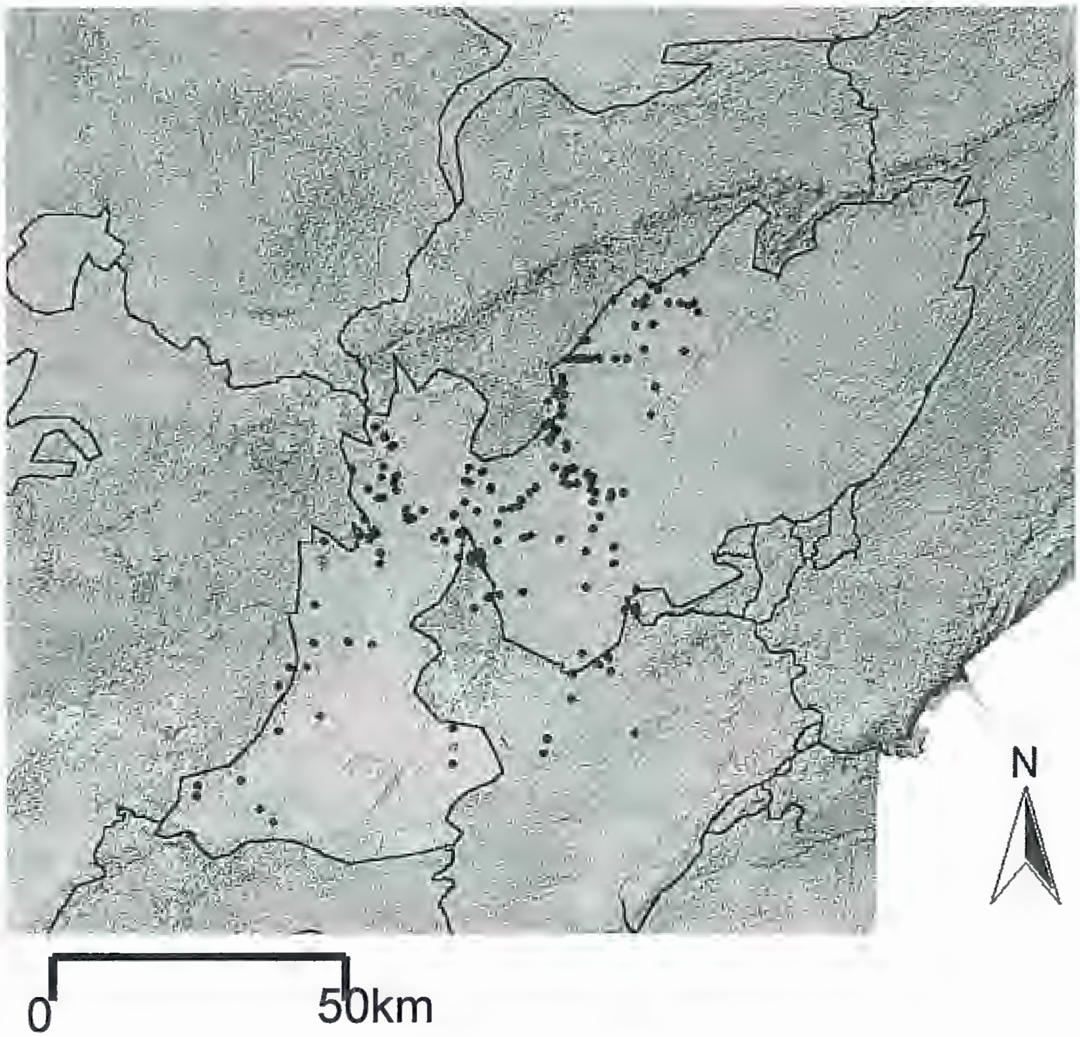


Figure 7. Field sites in the Aitkin and Upham basin. Base map is a 3x vertically exaggerated mosaiced 30-meter DEM (Minnesotat DNR).

3.1 Lab Analysis: Grain Size of the dunes

The grain size of dune sand is often used to link it to its source sediment (Cooper, 1935; Warren, 1976; Winspear and Pye, 1996; David, 1980). Determining the grain size distribution of sand is accomplished by mechanical analyses. Percentages of silt and clay are determined from the setting velocity of each size range. The grain size procedures followed Folk (1974) and are summarized below. A portion (~60 grams wet weight) of each sediment sample was placed in a settling column that was then filled to the 1 liter mark with dispersant (2.55g/l Calgon). Samples were agitated and subsamples were removed with a pipette at the depths and times shown in Table 1.

Ø	mm	depth	settling times (24°C)
<4	0.063	20cm	20 seconds
<5	0.031	10cm	1 min. 44 sec.
<6	0.016	10cm	6 min. 56 sec.
<7	0.008	10cm	28 min.
<8	0.004	10cm	1 hr. 51 min.
<9	0.002	5cm	3 hrs. 43 min.

Table 1. Depths and corresponding times of sample extraction from a settling tube at 24 degrees Celsius (Folk 1974).

Extracted subsamples were placed in beakers of known mass to be dried and weighed. Contents remaining in the column were rinsed through a .063mm screen to separate sand from silt and clay. The course size fractions were then dried and placed in a nest of screens to separate by size in 1 phi intervals (-1 Ø through 4 Ø) based on the Wentworth classification system (Table 2).

Grain Size (mm)	Grain Size (phi= -log2mm)	Wentworth Size Class
2.00 to 1.00	-1.0 to 0.00	very coarse sand
1.00 to .5	0.0 to 1.0	coarse sand
.50 to .25	1.0 to 2.0	medium sand
.25 to .125	2.0 to 3.0	fine sand
.125 to .0625	3.0 to 4.0	very fine sand
.0625 to .031	4.0 to 5.0	coarse silt
.031 to .0156	5.0 to 6.0	silt
.0156 to .0078	6.0 to 7.0	medium silt
.0078 to .0039	7.0 to 8.0	fine silt
.0039 to .002	8.0 to 9.0	very fine silt
< .002	9.0 to 14.0	clay

Cobbles 64-256mm

Pebbles 2-64mm

Coarse Sand .5-1mm

Table 2. Correlation of mean grain size, reported in mm and phi, and the Wentworth classification system (modified after Folk, 1974).

Individual sample statistics were calculated by the method of Folk (1974) using software developed by Mooers (unpublished).

3.2 Digital Elevation Model: Isostatic Rebound Correction of Aitkin and Upham basin--Basis for Boundary Conditions

The retreat of the Laurentide Ice Sheet during the last deglaciation resulted in extensive isostatic rebound across this region. Although the St. Louis sublobe, which advanced from the northwest, was the last lobe to cover the Aitkin and Upham basin it was not responsible for most of the isostatic rebound, as it was much thinner and short lived than the Rainy and Superior lobes which advance from the northeast (Wright, 1972; Mooers, 1990; Mooers, 1988). As a result inlets, outlets, beaches, wave-cut scarps, and spits associated with the lakes are now tilted up toward the northeast (Hobbs, 1983). To examine the relationships of these features the topography was corrected for isostatic rebound. Hobbs (1983) used elevations of beaches in Lake Koochiching to make inferences about the timing of inlets and outlets. Generally isobases, which are lines of equal uplift, trend NW-SE. A re-examination of the highest strandlines in the Aitkin and Upham basin was undertaken. By matching the highest shoreline features from the northeast part of lake Upham with the highest strandlines in the southwestern part of lake Aitkin the total amount of rebound across the basin could be estimated. As the isostatic rebound in the northeast corner of Upham was rapid well developed beaches did not have a chance to form so other features were used to estimate the maximum elevation of the lake in Upham. Wave-washed topography could be distinguished by its smooth appearance in contrast to hummocky terrain (Fig. 8). Figure 9 is an example of a typical wave-cut scarp and Figure 10 a beach found in the Aitkin and Upham basin.

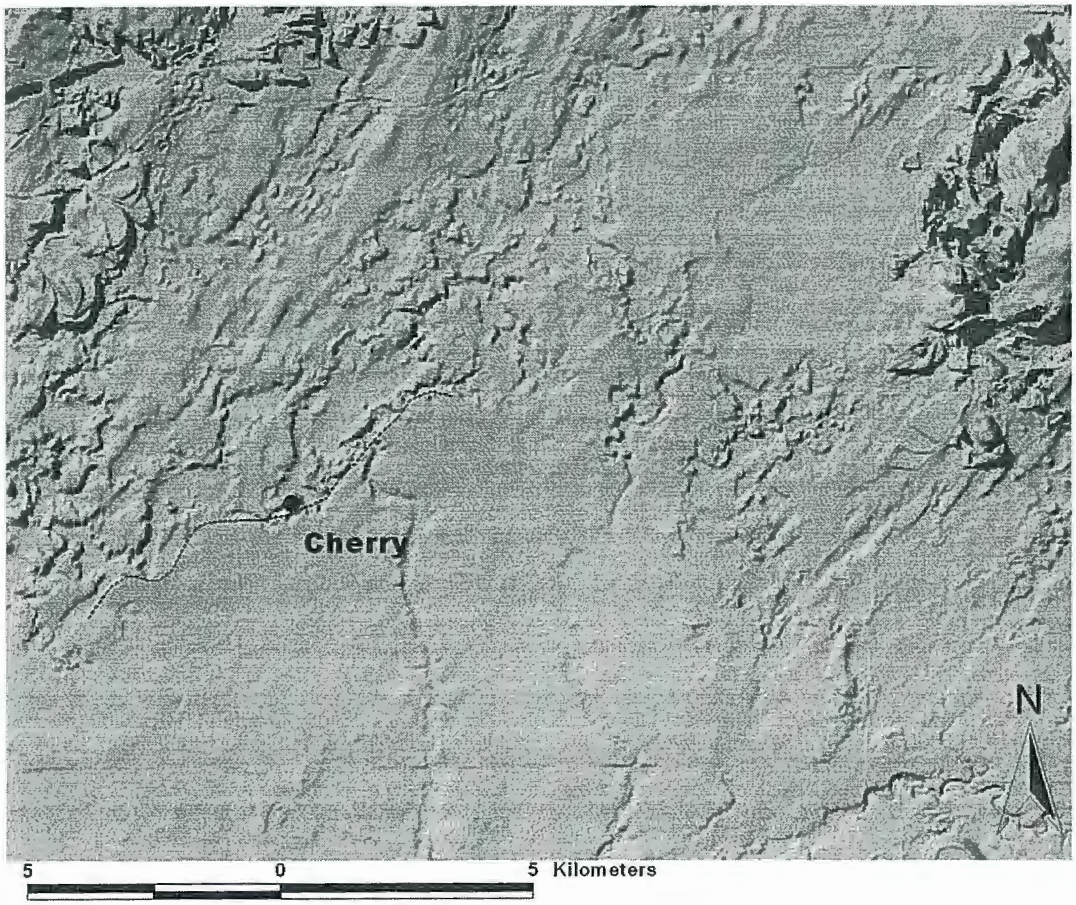


Figure 8. Wave-washed topography in northeast Lake Upham; Dashed line indicates a portion of the boundary that separates wave-washed from non-wave-washed topography. Image is from the same aforementioned base map.

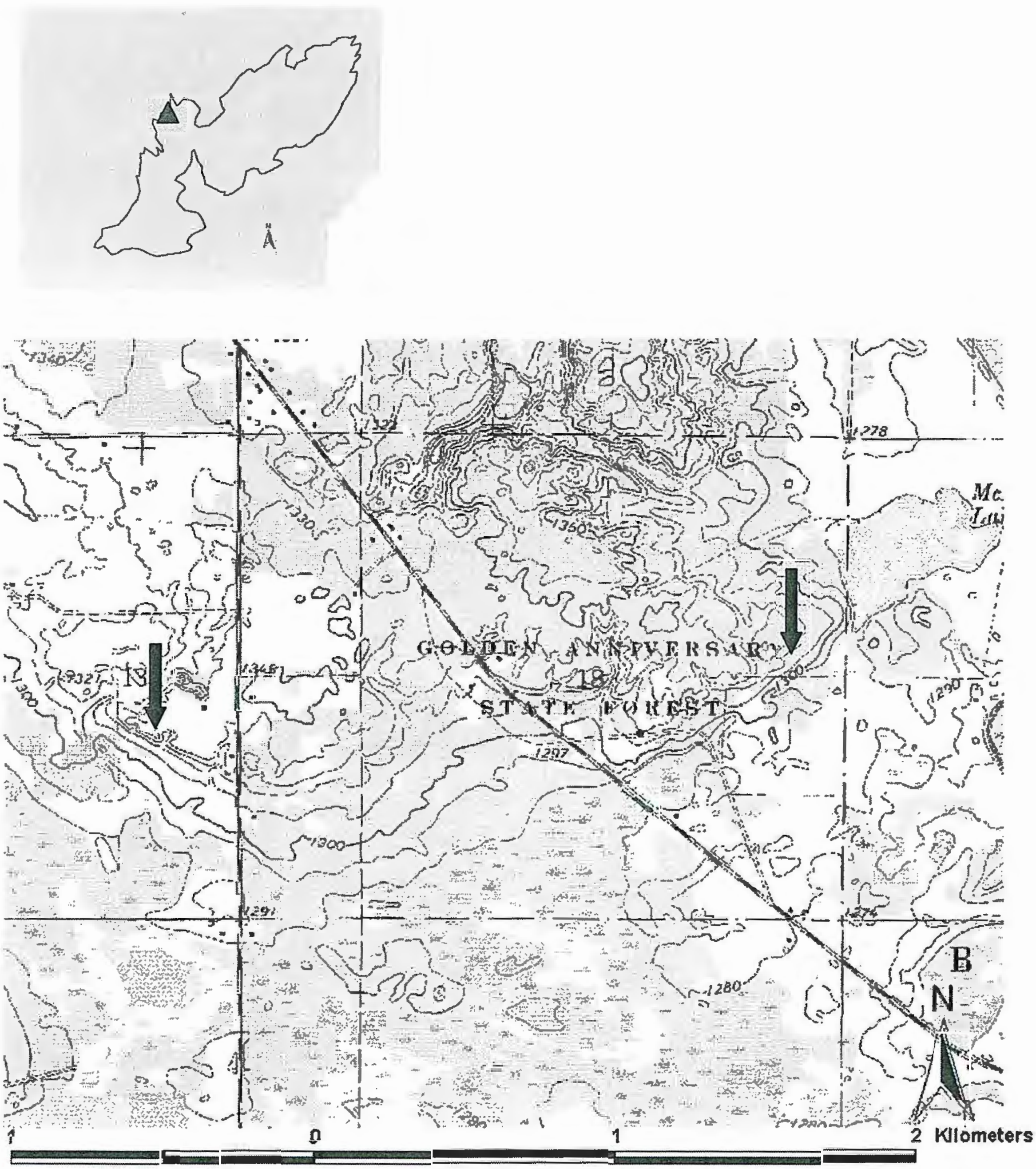


Figure 9. An example of a wave-cut scarp in Lake Aitkin (La Prairie 7.5 minute USGS quadrangle).

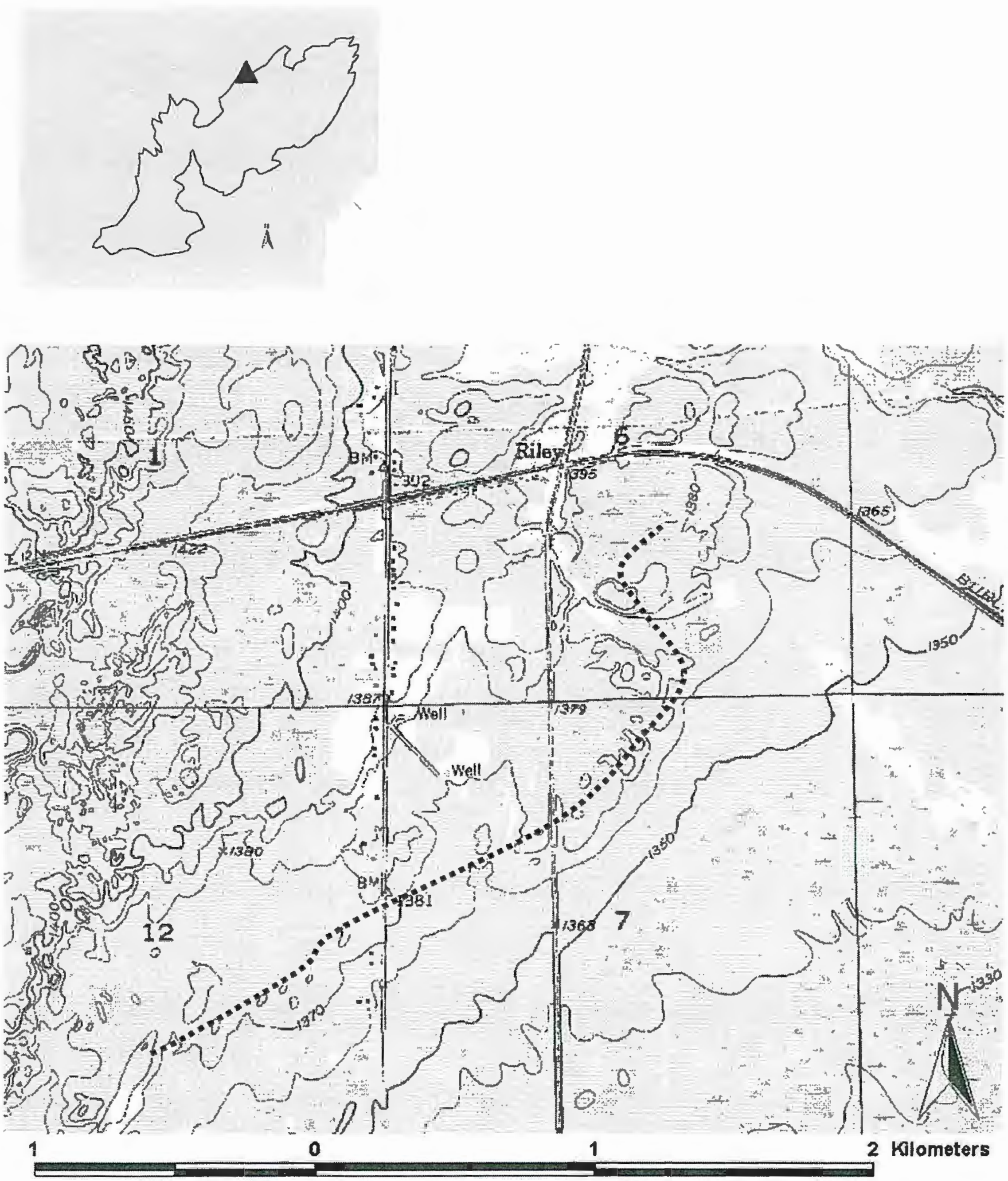


Figure 10. An example of a beach strandline (dashed line) in northern Lake Upham (Riley 7.5 minute USGS quadrangle).

Figure 11 shows the most well defined beaches of the Aitkin and Upham basin. These beaches indicate somewhat stable lake levels. However, more useful for the isostatic rebound correction is the plot of all of the strandlines throughout the basin, including wave-washed topography and wave-cut scarps (Table 3). The highest elevation of wave-washed topography is at 1430 ft. (436 m.) in northeastern Lake Upham and the highest beach elevation in southwest Lake Aitkin, 1210 ft. (369 m.) The total isostatic rebound across the basin from northeast to southwest over a distance of 160 km is 220 ft. (67 m.), which yields ~ .42 m/km of rebound.

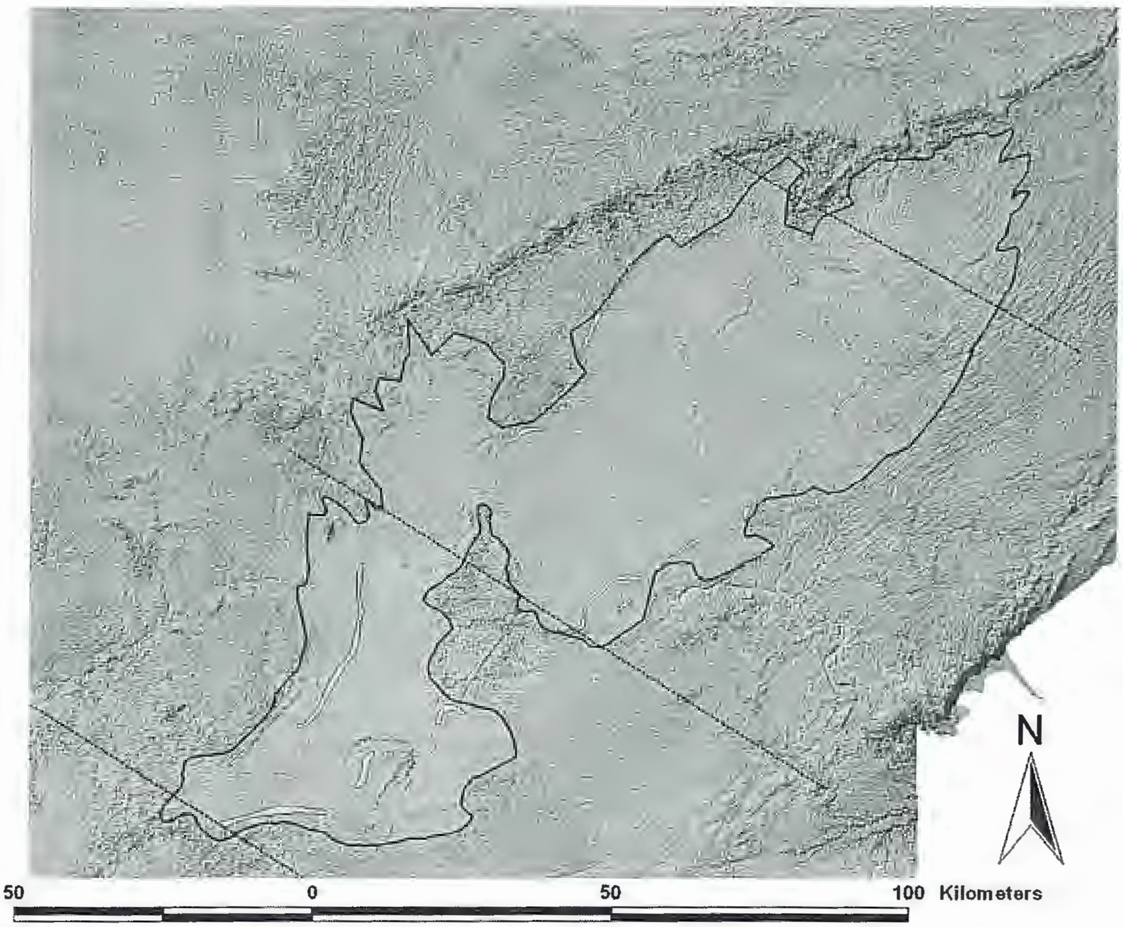


Figure 11. Beach Distribution: Dashed lines are suggested Isobase lines, lines of equal uplift, across the basin. Modified after Hobbs (1983).

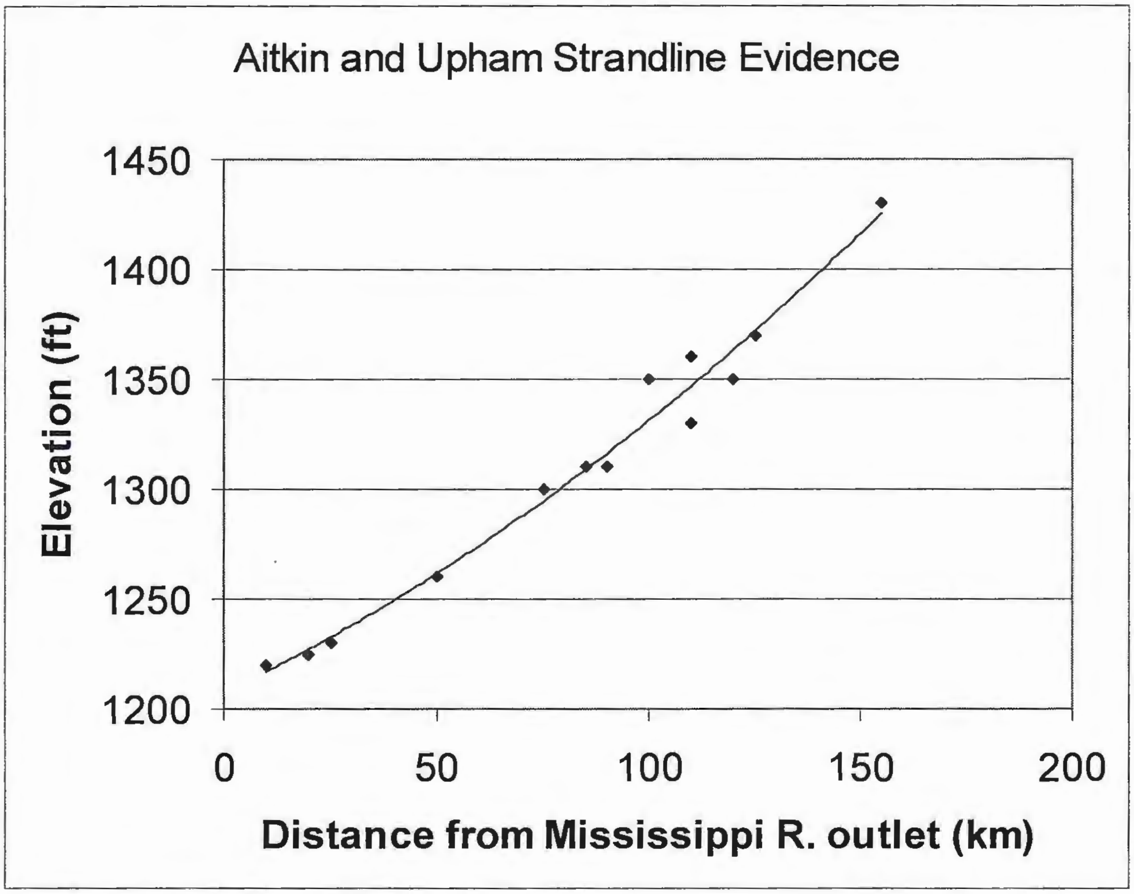


Table 3. Strandline elevation vs. distance along the axis of rebound from the Mississippi outlet (the southwestern most part of the basin).

The estimated amount of rebound of 220 ft. (67 m.) across the 161 km Aitkin and Upham basin corresponds well with the estimated 215 ft. across the same interval using Agassiz beaches by Hobbs (1983). The rebound curve and the equation of a best-fit polynomial to the highest strandline is $y=0.003x^2 + 0.9369x + 1207.3$. Using this relation for isostatic rebound the DEM of the Aitkin and Upham basin was completely adjusted for rebound. Then the basin was incrementally uplifted from a calculation (Table 4) based on the rebound curve. There were a total of 11 increments, which correspond to ~20 ft. of uplift at each step (Figure 12) without the consideration of time as a factor. The result is a record of the extent, inlets, outlets and beaches of the lake through different stages of uplift.

Y Axis is Distance in Meters	X Axis is Elevation in Feet										
0	1210	1210	1210	1210	1210	1210	1210	1210	1210	1210	1210
70958	1285	1277.5	1270	1262.5	1255	1247.5	1240	1232.5	1225	1217.5	1210
160746	1430	1408	1386	1364	1342	1320	1298	1276	1254	1232	1210
	Calc 1(Max)	Calc 2	Calc 3	Calc 4	Calc 5	Calc 6	Calc 7	Calc 8	Calc 9	Calc 10	Calc 11

Distance is calculated from the lower end of the basin (near Aitkin, MN)

The elevation at the different isostatic adjustment increments is calculated based on its distance from the base level (Aitkin), perpendicular to isobase (lines of equal isostatic rebound) that run roughly NW-SE across the basin.

Table 4. Isostatic Rebound Correction Calculation: 1430 ft. Upper Elevation, 1210 ft. Lower Elevation.

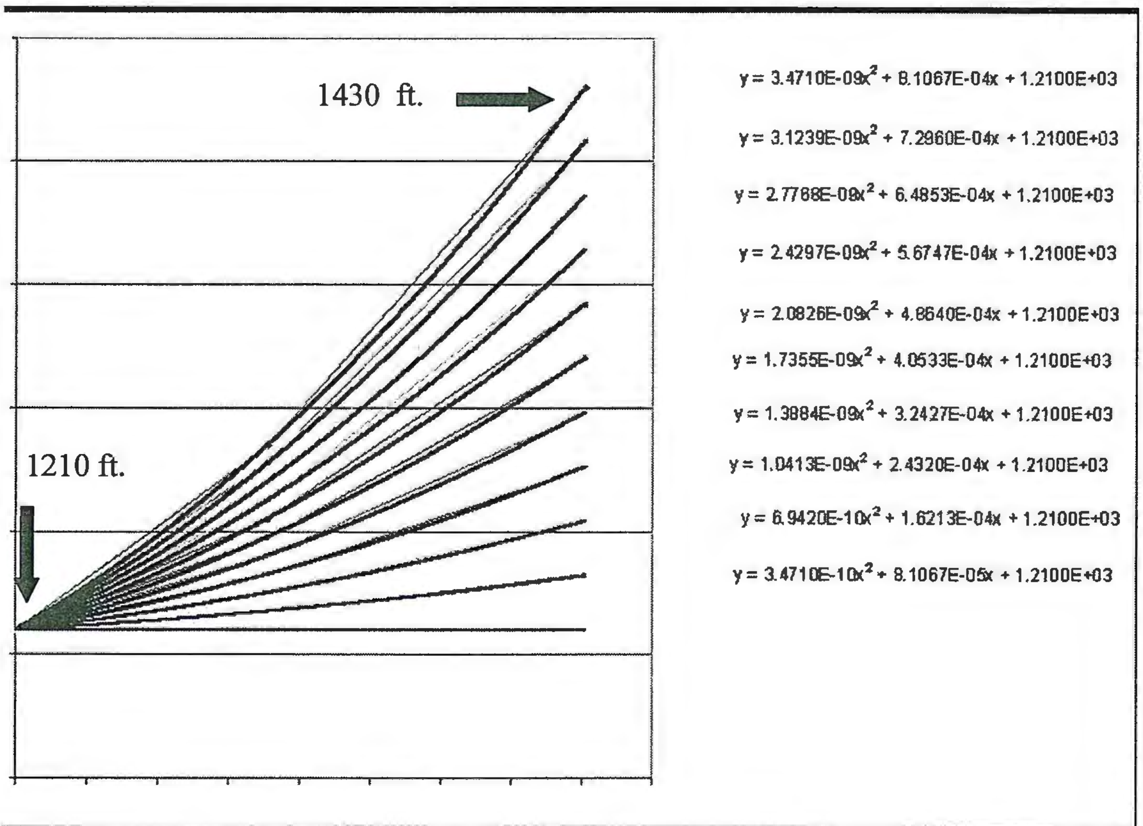


Figure 12. Isostatic Rebound Correction.

3.30 Lake Core: Retrieval

An 8.57 meter sediment core was retrieved from the deepest part of Hay Lake (38.2 ft.)(Figure 13). Hay Lake has an area of 106 acres. Hay Lake was primarily chosen for a core site because of its position in Lake Upham and its location within a dune colony, which increased the likelihood of an eolian signature in the lake sediments (Fig. 14 and 4). Hay Lake (93°30' W, 46°55'N: gps coordinates 483875; 5200233; elevation absl 1268 ft.) is a kettle lake on the edge of the moraine-outwash complex of the Rainy Lobe overridden by the St. Louis sublobe.

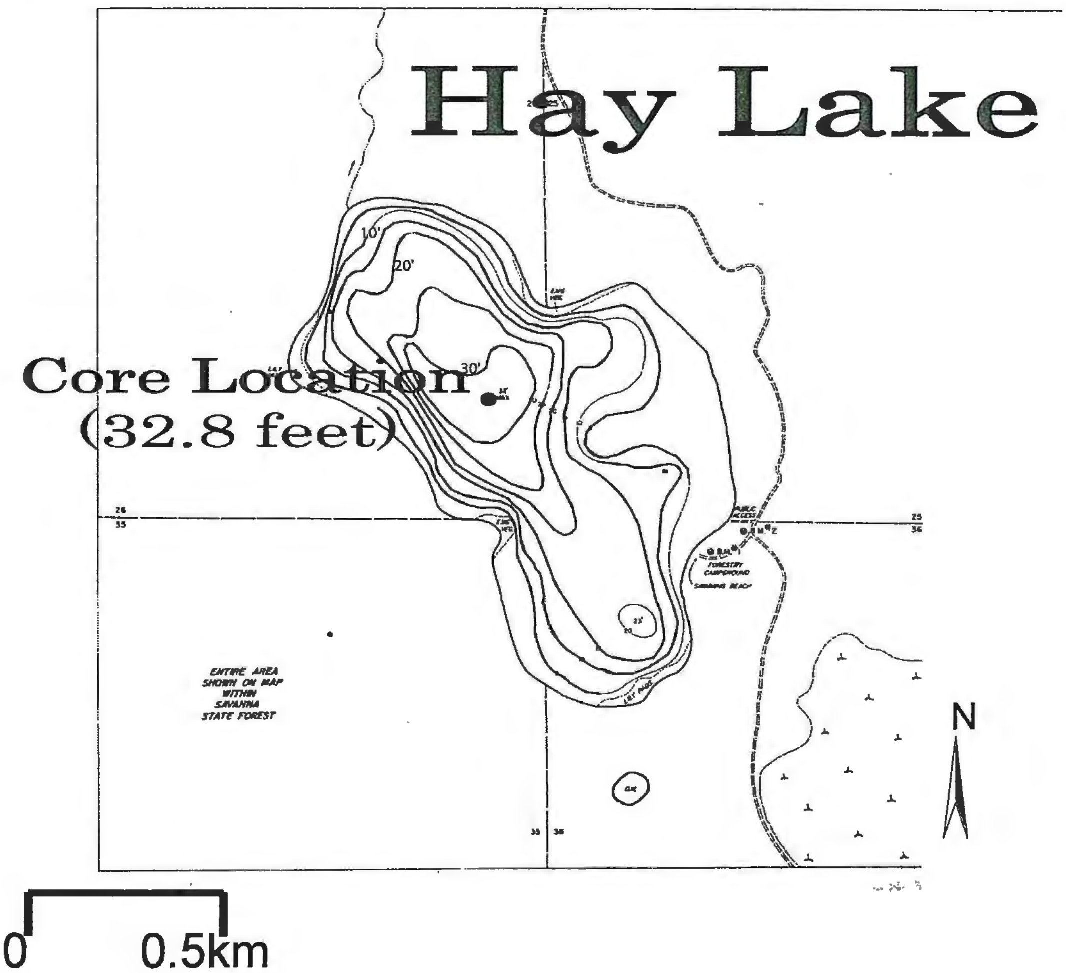


Figure 13. Bathymetry map of Hay Lake; Location of core marked with a circle. Modified from lake map found at http://www.dnr.state.mn.us.

Figure 14. Aerial photo of Hay Lake. Horizontal lined areas indicate Lake Upham border; White polygons are dunes (Vanduse 7.5 minute USGS quadrangle).

All procedures related to the coring and subsequent analysis were done in accordance with the University of Minnesota Limnological Research Center (LRC) Handbook (http://lrc.geo.umn.edu/services/handbook/overview.htm) A Bolivia coring system, which is based on the Livingstone square-rod piston push corer was used. The 8.57 m. of core were retrieved in 9 sections. Sections were labeled by LRC specifications.

3.31 Lake Core: Lab Analysis

Multisensor logging was performed on all sections of the core at 1 cm intervals using the GEOTEK multisensor logging track. The GEOTEK multisensor logging track performs magnetic susceptibility measurements using a Barington loop magnetic susceptibility, which is a measure of the abundance of magnetic minerals. Gamma ray densitomotry was also measured by the GEOTEK, providing a high-resolution record of water content and internal structure. Peaks in magnetic susceptibility represent increases in magnetic minerals (Sprowl et al., 1993) and is sometimes used as a proxy for influx of eolian terrigenous sediments (Keen et al., 1990; Bloemendal et al., 1989). The cores were split, visually described, and photographed. Since this was not intended to be a detailed study of the sedimentation patterns in Hay Lake, the sediments were described in general terms and thicknesses. Samples were taken at 20 cm intervals for bulk density. Smear slides were made at similar intervals, with additional slides at areas of interest. Smear slides were examined and described for their various components (mineral, biogenic, and authigenic).

3.32 Lake Core: AMS Dates

Terrigenous plant fragments were picked from the core at 6.9 m. and 8.2 m. The dated samples at 8.2 m. were a conifer tree needle and a twig, and the upper 6.9 m. samples were twig and leaf fragments. Samples were then given to the LRC AMS Radiocarbon Target Preparation Unit for preparation for accelorator mass spectrometry at the University of Arizona

3.33 Lake Core: Carbon Analysis

Total organic and inorganic carbon were determined at 20 cm intervals by Loss on Ignition. Values were then used to calculate the water content, percent organic material, percent inorganic material, and percent $CaCO_3$. The following temperatures and times were used to measure the water content, organic carbon, and inorganic carbon, respectively: 110° C , 1hour; 550° C, 1 hour; 1000° C, 1 hour. Total Inorganic Carbon -

TIC, Total Organic Carbon-TOC, Linear Sedimentation Rate –LSR, Mass Accumulation Rate –MAR, porosity, were then calculated by the formulas outlined in Table 5.

Formulas Used in Coulometry Calculations at 20 cm Intervals	
% porosity =	(Mass water/ ρ_w)/ (mass water/ ρ_w) +(mass dry sediment without crucible/ρ_s)
% organics (% lost @550) =	(crucible weight with sediments - 550 weight with crucible) / dry weight (cruc wt with sed - cruc wt) * 100
% loss @1000 =	[(550 wt - 1000 wt.)/ dry weight (Cruc wt with sed - cruc weight)] *100
% Carbonate =	% lost @1000 / 0.44
% minerals =	100 - %Carbonate -% Organics
mar organic carbon (g/cm2-yr) =	0.1 * (1- porosity)* 2.65 * (%organics/100)
mar caco3 (g/cm2-yr) =	0.1 * (1- porosity)* 2.65 * (% carbonate/100)
mar minerals (g/cm2-yr) =	0.1 * (1- porosity)* 2.65 * (% minerals/100)

depth (cm)	age BP	diff. Depth	diff. Age	sed rate (cm/yr)
10	0			
690	6635	680	6635	0.102487
820	10108	130	3473	0.037432

sedimentation rate =	differential depth -differential age

Table 5. Organic and Inorganic Carbon and Mineral Calculations.

3.40 Ground Penetrating Radar: Lake Upham Shoreline

In this study GPR was chosen to image the subsurface structure at a shoreline on the western side of Lake Upham at Townline Road (Fig. 15). This technique was chosen because internal reflections in the GPR images might record multiple lake levels. A 400 foot (122 m). GPR transect trended E-W perpendicular to the shoreline was recorded. The now vegetated western shore of Upham is marked by a sharp change in elevation by the transition from till to the shoreline deposit, 1400 to 1350 ft. (427-411 m.) respectively, to basin elevations from1320 to 1300 ft. (402-396 m.) over the distance of 1 to 4 miles (~2-4 km), respectively (Fig. 15).

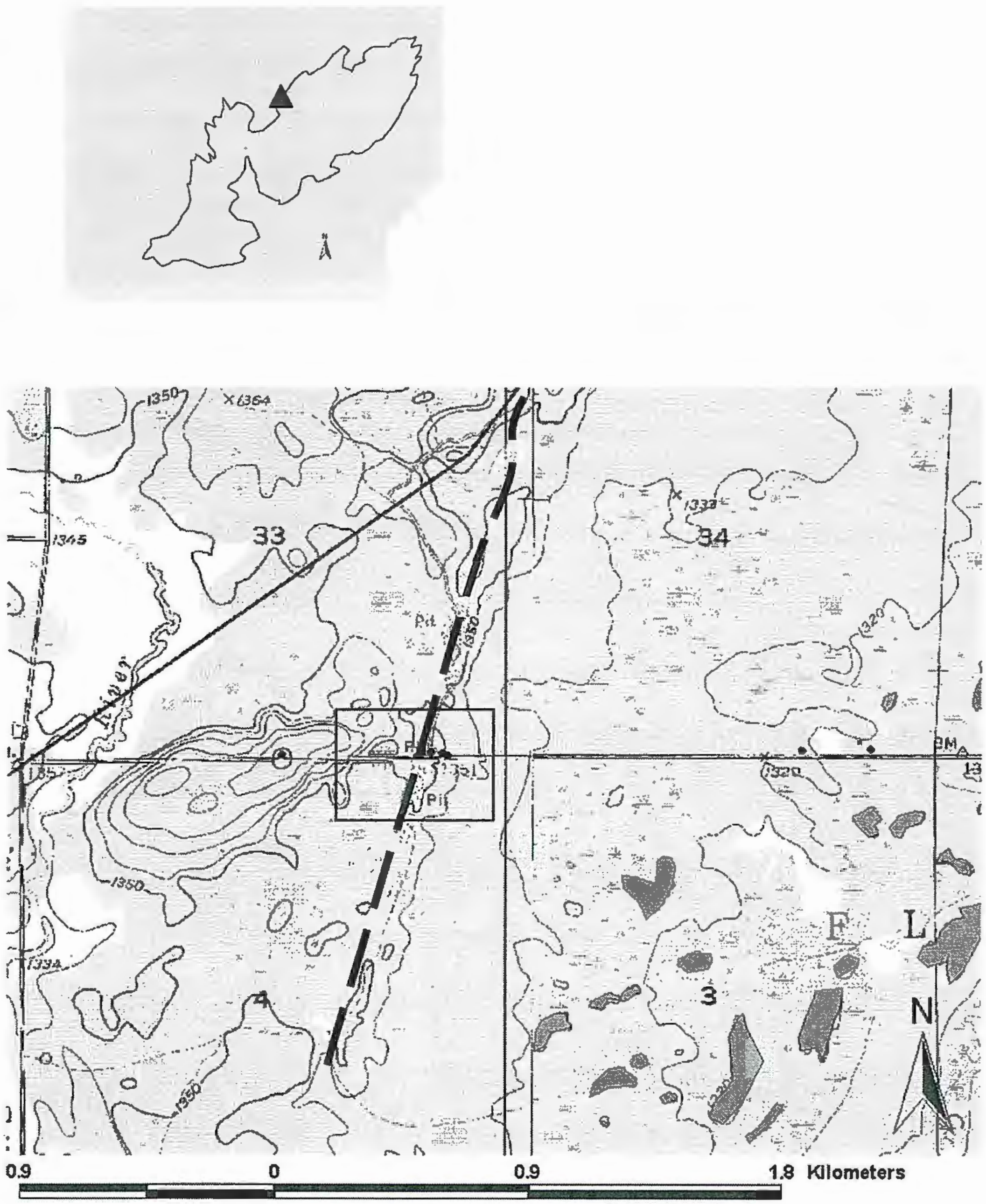


Figure 15. GPR survey along Lake Upham shoreline, marked by dots in the box. Dashed line is the border of Lake Upham. Polygons in southeast indicate dunes outlined in a colony (Silica and Riley 7.5 minute USGS quadrangle).

A *PulseEKKO 100* Ground Penetrating Radar, (*Sensors & Software Ltd., Mississagua, Ontario, Canada*) was used for the surveys. This system can be operated in a variety of modes, and for these surveys it was configured to collect zero-offset reflection data. In this configuration, the transmitting and receiving coils are located next to each other and the system records the 2-way reflection time for signals off subsurface interfaces (Nigel Wattrus, Personal Communication). The following discussion is largely from a written narrative by Nigel Wattrus.

> Like seismic sources, the frequency at which a GPR is operated influences the resolution of the data produced and the penetration of the radar pulse. Lower frequencies travel further into the subsurface than do high frequency signals of similar strength. Conversely the resolution of the system (i.e. how closely spaced two reflectors can be and still produce two distinct radar returns) increases with increasing signal frequency. The theoretical vertical resolution limit of a radar system is a quarter wavelength (wavelength = radar velocity/frequency), it may be poorer in the presence of noise. The GPR operator therefore has to select a frequency that provides the signal penetration required at the highest possible resolution. For these surveys the system was operated at 100 MHz. In the dry sand found near the surface at this location, radar pulses will travel at ca. 0.1m/ns. This means that the wavelength of the 100 MHz pulse is 1m and the maximum resolution of the system is 0.25m.
>
> The nature of the subsurface sediments controls the radar propagation velocity and the attenuation of the radar pulse. The electromagnetic properties of the sediments are related to their composition and water content. Some materials, such as dry sand, transmit radio waves very well whereas other materials, which are relatively good electrical conductors, such as clay or brine-saturated sediments, absorb them very effectively, making these types of sediments virtually opaque to radio waves. Radar reflections are produced by changes in the relative dielectric constant of the subsurface sediments. The electrical conductivity of the subsurface determines how deep a radar pulse penetrates. The more conductive the subsurface is, the quicker a radar pulse will be attenuated. Dry sands and sands saturated with freshwater have relatively low conductivities and a GPR operating at 100 MHz can achieve sub-surface penetration of greater than 5 m.
>
> The *PulseEKKO 100* uses two bistatic antennae one to transmit and a second to receive, spaced at 1 m and orientated perpendicular to the survey line (Figure 15). To avoid spatial aliasing, which is essentially undersampled data that can lead to incorrect interpretation of seismic reflectors, the data were acquired at 25 cm intervals along transects established across the target areas. The "classic" example of aliasing is the apparent reversed rotation of stage coach wheels in the old western movies. This occurs because the movie is actually composed of a series of still images that are recorded at a rate too slow to capture the true motion of the wheels. As a result they appear to go backwards. The survey geometry was obtained by recording GPS locations at regular intervals along the transect. (Nigel Wattrus, Personal Communication).

Figure 16. GPR shown on Lake Upham Shoreline: bistatic antennae perpendicular to survey line.

Post-survey processing was performed by Nigel Wattrus (UMD-LL0). Processing was done using *EKKOTools*, a package of processing tools provided by *Sensors and Software,* the manufacturer of the system, and consisted of cleaning the data, applying filters to reduce noise, application of gain to boost weak signals, deconvolution to simplify the radar wavelet (i.e. simplifying the appearance of the reflections), and applying time shifts to the data to correct for topographic relief. Topographic information was derived from 7.5 minute quadrangles with 10 ft contour intervals and adjustments were made according to relative height retrieved from a transit. After processing, the data were converted to the industry standard SEGY data format and loaded on a seismic interpretation system for display. The data were displayed using a wiggle trace display which enhances reflector continuity. The processed data transects are shown in later sections, and are compressed horizontally to display prominent reflectors. The data are plotted in two-way time. By choosing an appropriate radar velocity the reflection times can be interpreted in depth. For this study an average radar velocity of 0.1 m/ns has been used. To compute various thicknesses the time thickness was divided by two (to convert to one-way time), then multiplied by 1000 (conversion from seismic) then multiplied by 0.1m/ns to get meters of depth. The radargrams shown in later sections represent approximations of the geologic profiles of the study sites. Since the variation in radar velocity is not known the profiles may contain imaging artifacts that are produced by unknown changes in the velocity profile that do not reflect the true structure.

Chapter 4: Results

4.0 Dunes: Geographic distribution of dune sands

Dunes are located sporadically throughout the Aitkin and Upham basins with concentrations in certain locations. The most abundant dune distribution is in the region of the Prairie River inlet in northern Lake Aitkin and continue through the Swan River Sill and extend north and south of the sill into southwest Lake Upham (Fig. 4). There are minor dune concentrations on the western shoreline of Aitkin and Upham. The surface area of the colonies ranges from 2 km^2 to 300 km^2. The colonies are generally elongate in a NW/SE direction. (Fig. 4). According to Bagnold (1941) "Dunes tend to occur in belts or chains, whose direction coincides with that of the resultant long-period sand vector Q " (Bagnold, 1941, p. 218) where Q is the sum of the direction of the strong and gentle winds responsible for sand movement. Within the colonies there are regions of dense and sparse dune distribution. The position of colonies exhibit a strong relationship with the transition to different basin levels, which is marked by gradation of contour lines (Fig. 17).

The dune colonies that occupy the western shoreline of Aitkin and Upham extend basinward no more than 6 km. However, the colonies that originate from the Prairie River inlet extend completely across the basin, and through the Swan River Sill. The large colonies in the vicinity of the Prairie River inlet and Swan River Sill are where there was water exchange from Aitkin to Upham towards the outlets in southeast Upham (Hobbs, 1983). These dune colonies are elongate in a direction toward the outlets of Upham (Fig. 18).

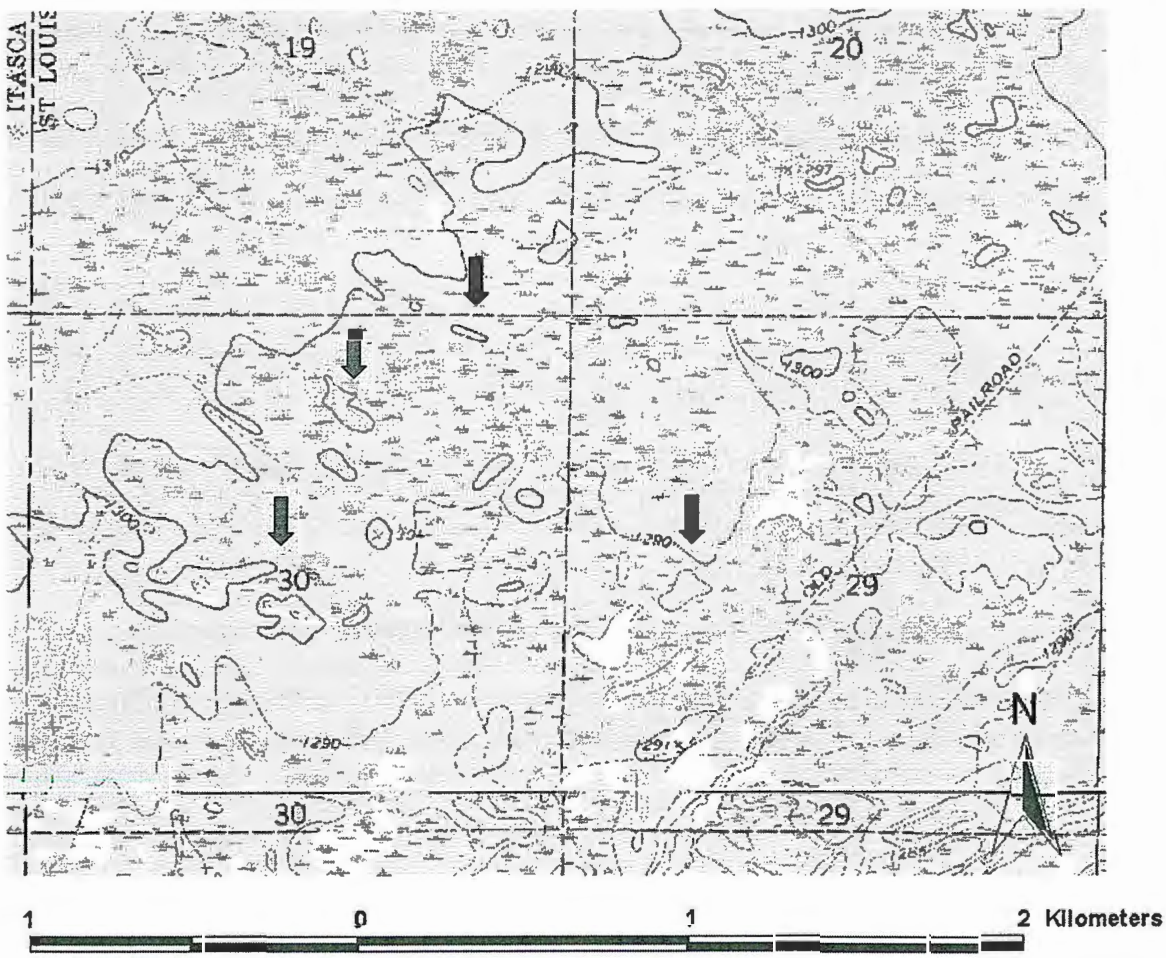


Figure 17. Dunes with respect to shore to basin contour lines (1300 and 1290 ft./ 396 m. and 393 m.), this is just northeast of the Swan River Sill in the Upham basin. Arrows are pointing to the dunes that are basinward of the 1300' and 1290' shoreline contour. (Floodwood 7.5 minute USGS quadrangle).

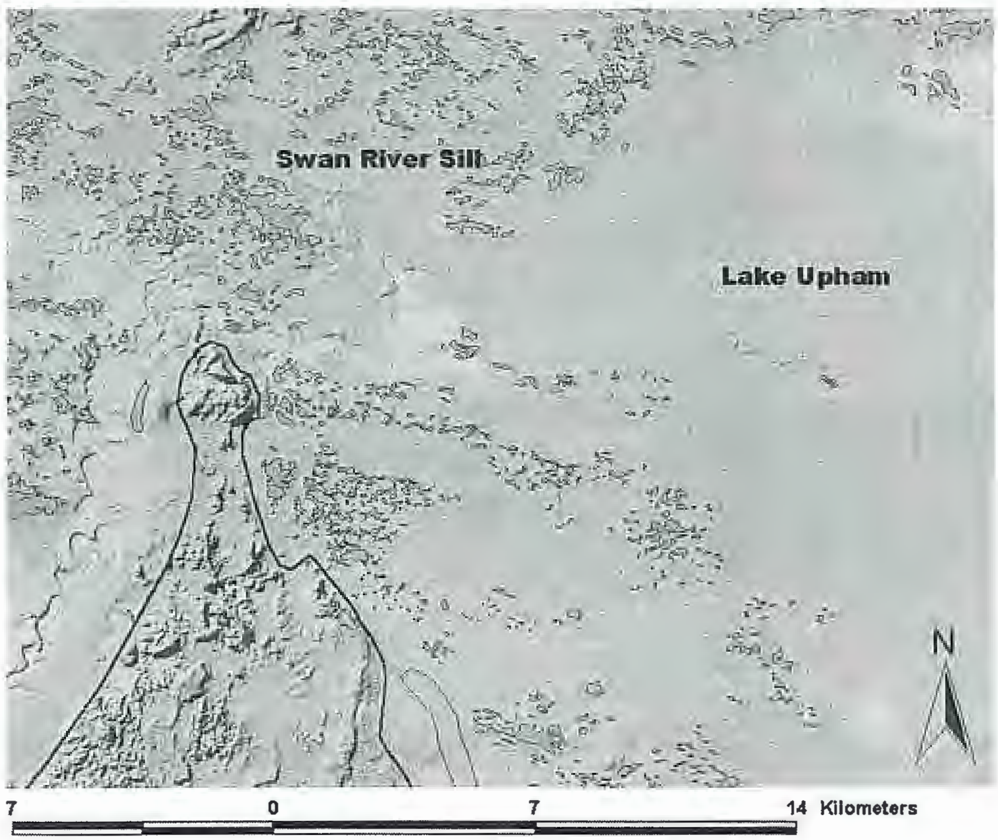


Figure 18. Colonies with respect to the Swan River Sill. Solid line indicates lake border; Solid lines with polygons are dunes; The two elongate polygons next to the border are beaches.

The parent material in the regions of the most prominent dune distribution, the vicinity of the Prairie River inlet and Swan River Sill has been mapped by the Aitkin and Itasca Soil Survey (Nyberg, 1999 and 1987) as Zimmerman, Cowhorn and Wawina soils. They are a fine to very fine sand of glacial lacustrine origin. Hobbs (1983) described this sediment as an underflow fan that extends for 50 km from its apex, the Prairie River inlet. It is seen in outcrop along the Mississippi River, where approximately 9 meters (27 ft.) is exposed from surface to the clay basement (Fig. 19 and 4).

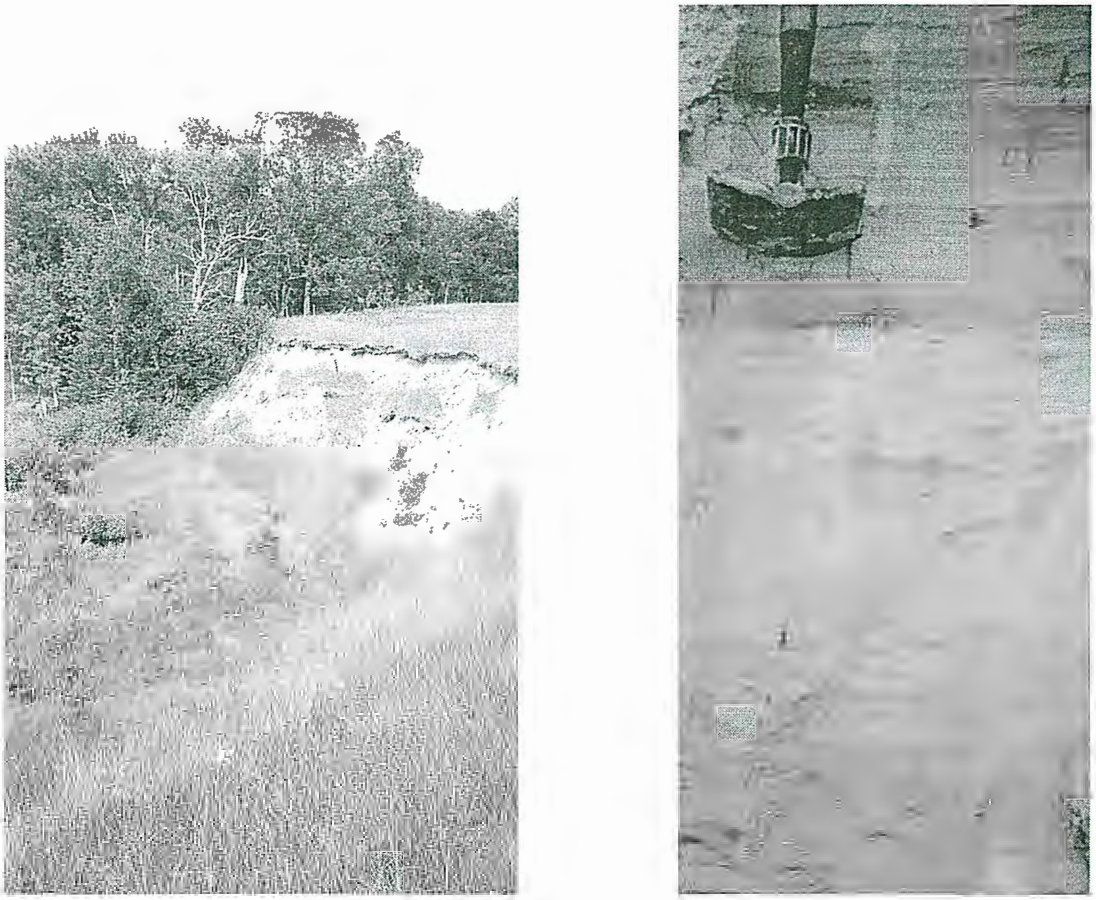
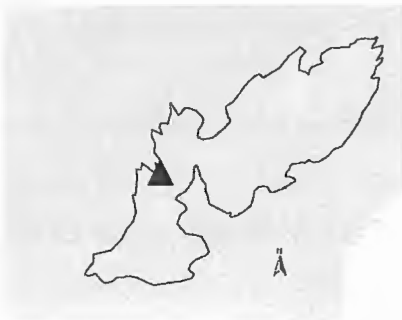


Figure 19. Mississippi River outcrop: Prairie River Underflow fan. Located ~10 miles (16 km) south of the Prairie River inlet in the Aitkin basin.

4.01 Dunes: Morphologies of dunes

The dunes in the Aitkin and Upham basin occur as various shapes and sizes. The form that a dune takes is dependent on the relative proportions of strong and gentle winds (Bagnold, 1941). The simplest forms of barchan, longitudinal or transverse dunes are typically only found in desert environments (Bagnold, 1941) where factors such as moisture, vegetation, and sand conditions do not hinder dune development. In regions where those factors are present it is common to have dunes with distorted shapes. This is generally the case in the Aitkin and Upham basin. There are several elongate dunes that may be examples of the longitudinal form, however they are often distorted in some way by sand accumulation off to one side, in which case they could possibly be groups of parabolic dunes (Fig. 20). Longitudinal or 'seif' (Arabic for sword) dunes occur when the strong winds of a region make up more than one quarter of the overall wind (Bagnold, 1941). There is a predominant NW-SE orientation of the long–axis of the longitudinal dunes in the Aitkin and Upham basin.



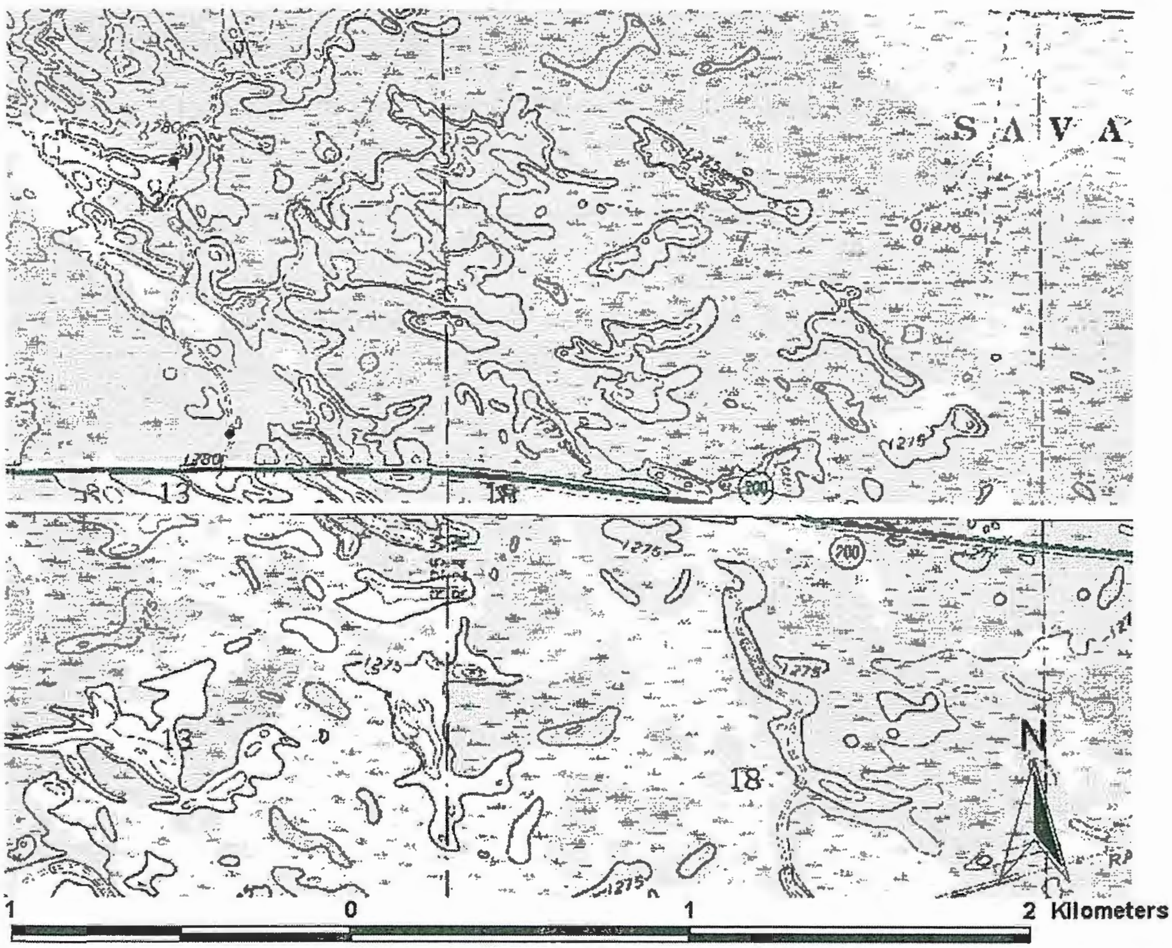


Figure 20. Longitudinal dune colony in Lake Aitkin (Split Hand and Rabey 7.5 minute USGS quadrangles).

Crescentic dune forms are also present throughout the basin as a minor component. Although these dunes exhibit the crescentic form, with the absence of a clear slip face, it is difficult to label the dunes barchan or parabolic (Fig. 21). Dunes without slipfaces are immature and have not had the conditions or time required to achieve mature dune form; with no slipface, dunes are supplied with new sand but are unable to trap sand moving over the crest of the dune (Bagnold, 1941, p.203). Although the crescentic dune forms are evident in the dune colonies throughout the basin, the most common morphology of dunes in the Aitkin and Upham basin is a distorted combination of dune forms. (Fig. 22) This distorted dune form can be a result of the collision of migrating dunes, which occurs in regions of variable hydrologic or vegetative conditions leading to disparity in relative dune migration rates (p. 214 Bagnold, 1941). The maximum dune crest is 5 m./ 15 ft.. The highest dunes occur near the Prairie River inlet and just to the south in the Aitkin basin. Elsewhere throughout the basin, dunes are generally a single contour line, and are ~1.5 m./5 ft. high. Dunes are often exposed in the field. When dunes are visible in cross-section (Fig. 23) sedimentary depostional structure is usually absent, likely a result of bioturbation, tree-fall and animal movement. At some locations laminations are present, though often discontinuous.

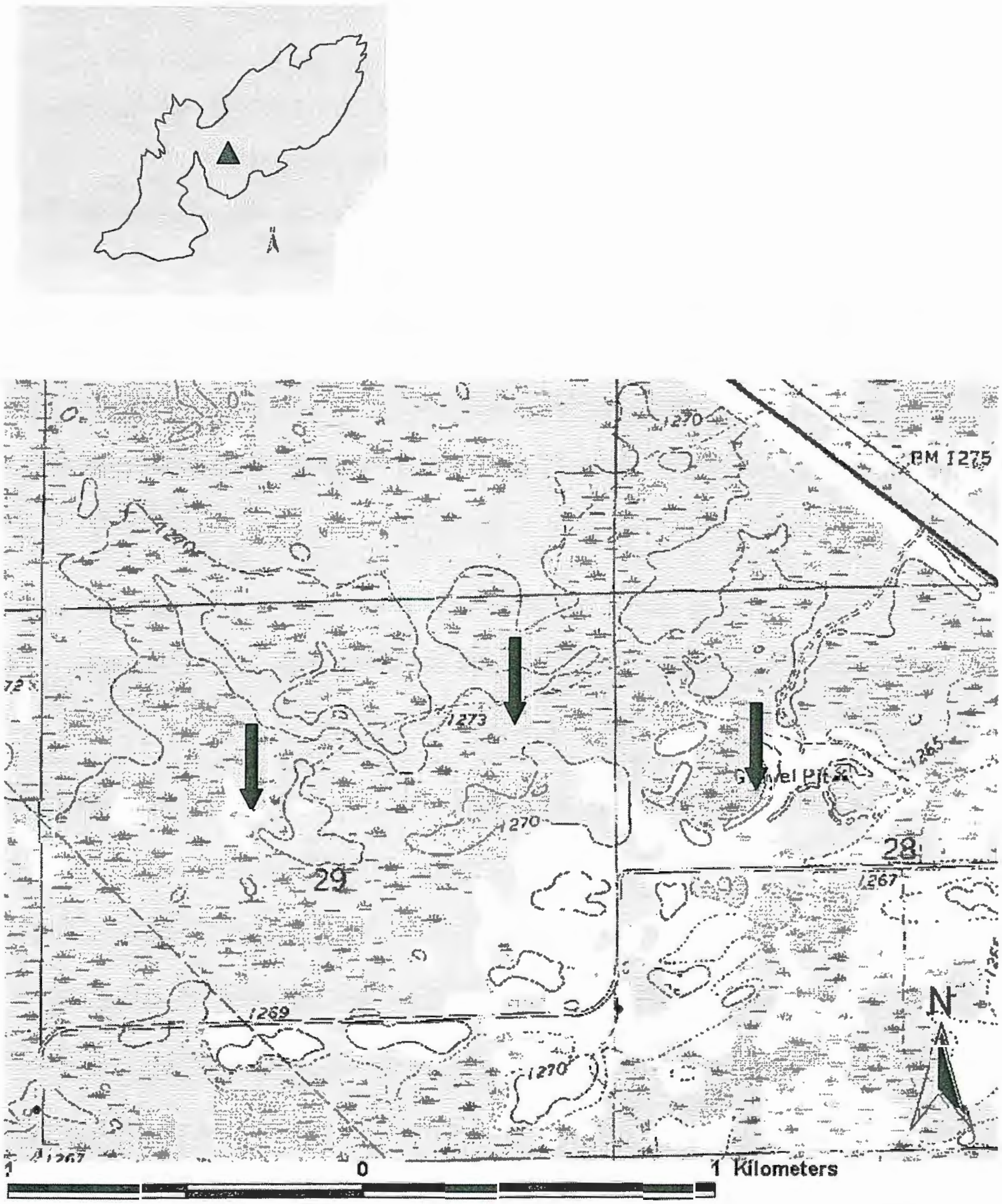


Figure 21. Crescentic dunes. Arrows indicate crescentic dunes (Swan River 7.5 minute USGS quadrangle).

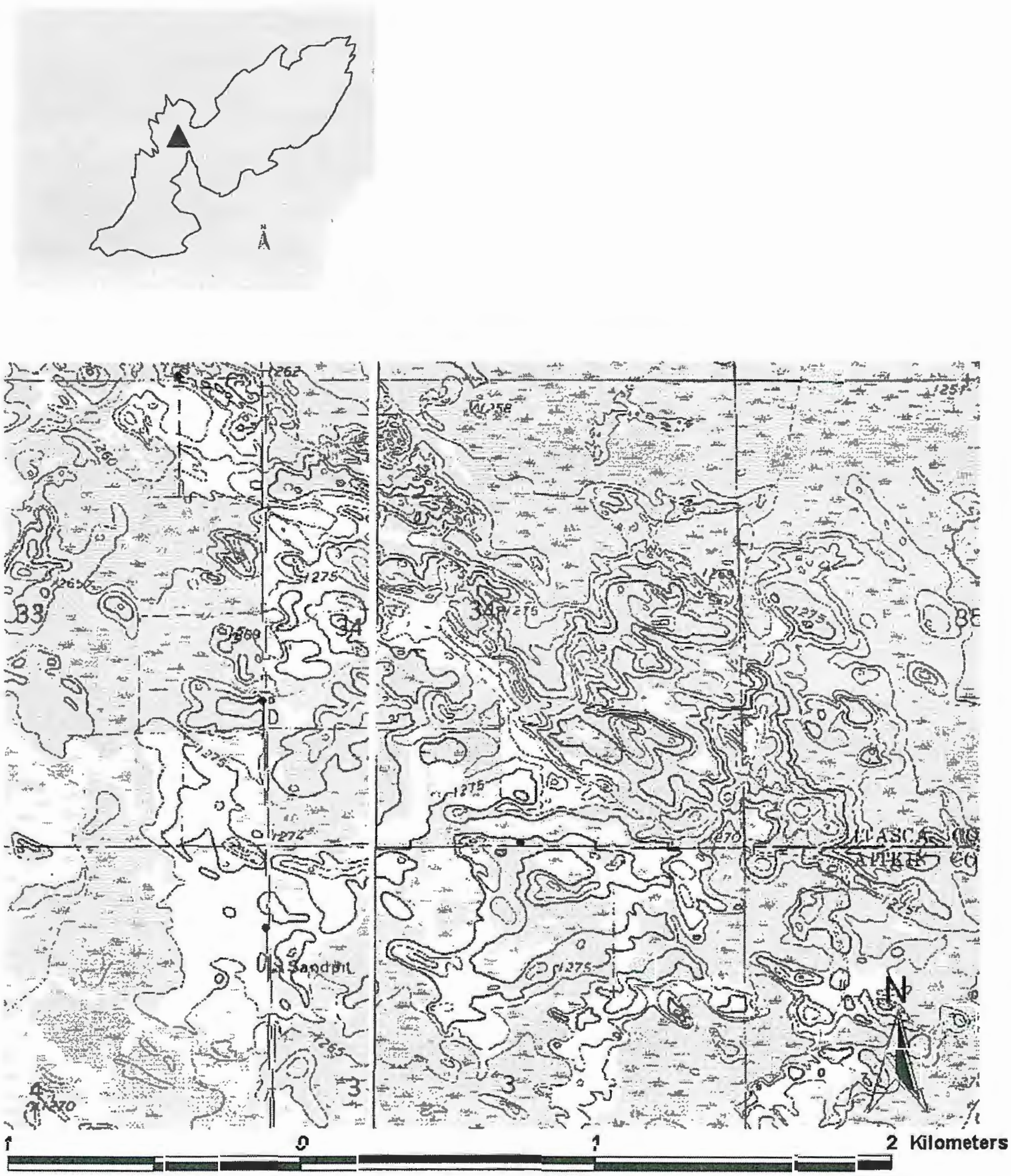


Figure 22. Distorted Dunes: combination of crescentic and longitudinal dunes. All of the elevation in this topographic map are dunes with irregular shapes (Jacobson and Split Hand 7.5 minute USGS quadrangle).

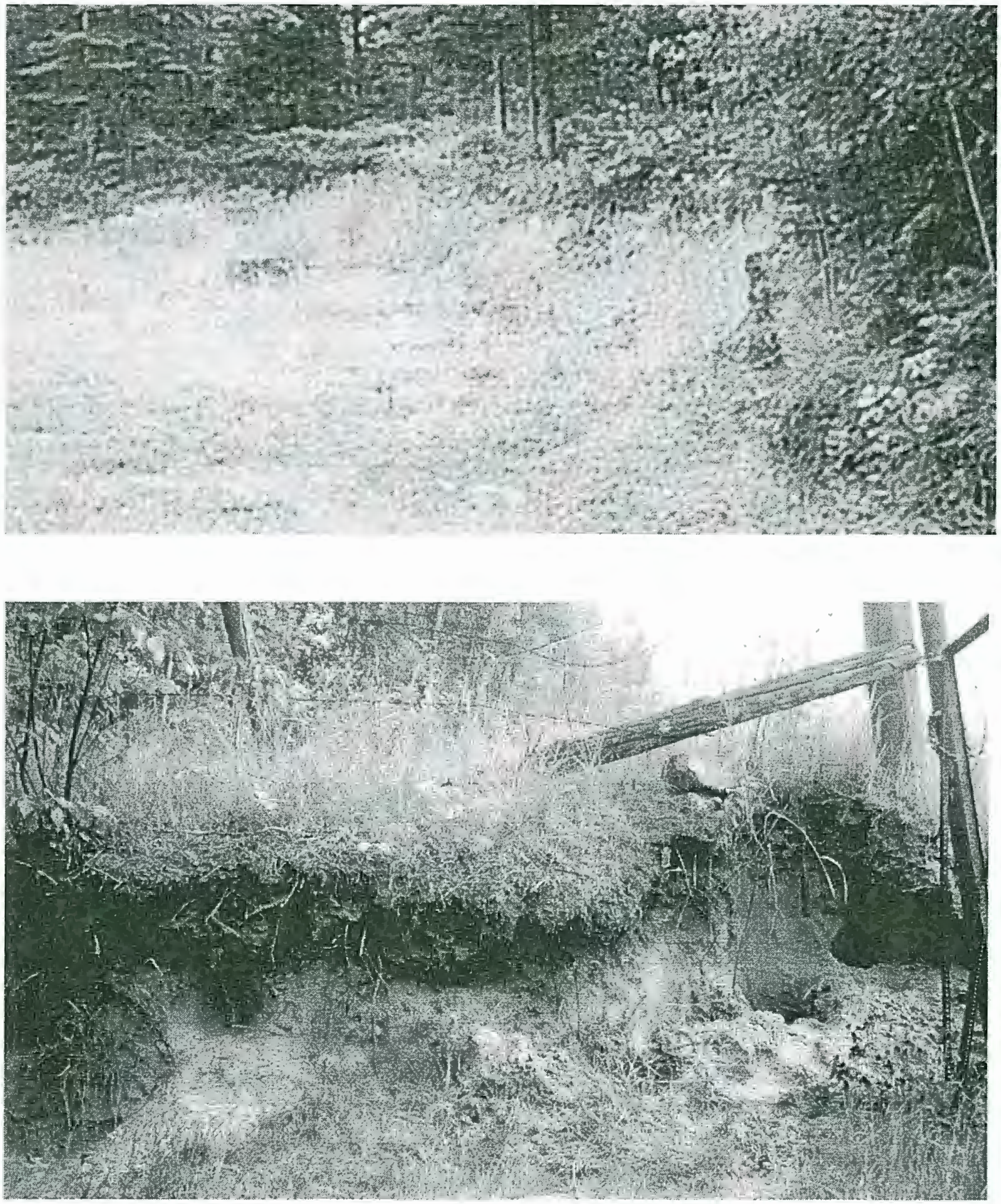


Figure 23 Dunes in cross-section

4.02 Dunes: Grain Size

The majority of the dune sand in the Aitkin and Upham basin is between .125-.065mm (3-4 Ø very fine sand) (Fig. 24). Grains move through traction, saltation, or suspension depending on their size. Grains between 0.088 and 0.354 mm are transported by saltation. Grains that are greater than 0.354 mm are transported by traction. Grains less than 0.088 mm in size can be carried by suspension. (Bagnold, 1941) The dune sand in the Aitkin and Upham basin were transported by a combination of saltation and suspension. A saltating grain can move grains that are 6 times its own diameter and 200 times its own weight and can move at speeds on the order of tens of centimeters per second (Bagnold, 1941). The grains are primarily quartz with lesser amounts of feldspar and other grains. Magnetite grains are more abundant in some locations than others and clearly visible in outcrop.

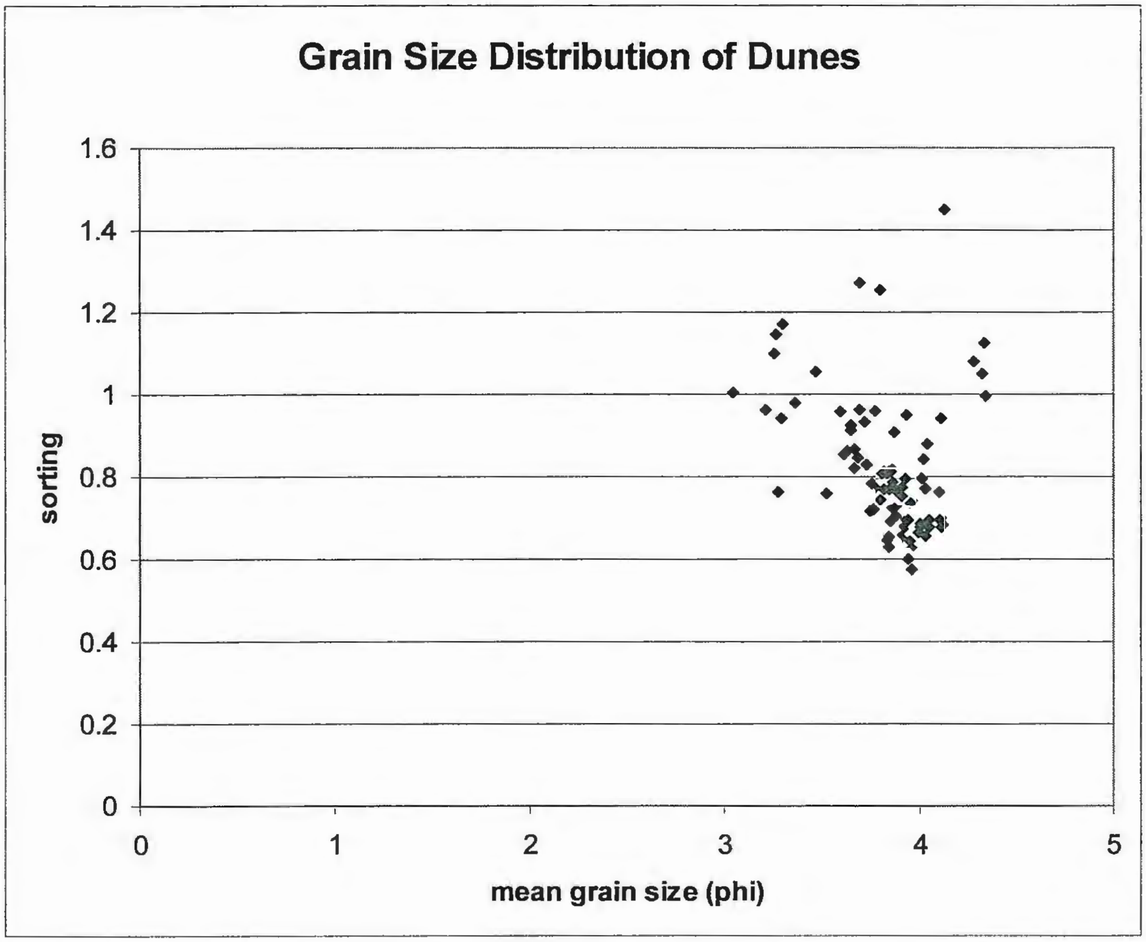


Figure 24. Grain size distribution of dune sand (n=77).

4.1 Isostatic Rebound Results

The basin was corrected for the estimated total differential rebound of 220' (67 m), and then uplifted in 20 ft.(6 m) increments. At each increment observations were made of the inlets, outlets and beaches at the various stages of uplift (Fig. 25). The paleo lake level marked in purple represents an elevation of the 1210 feet (368 m). Beaches are noted as yellow polygons.

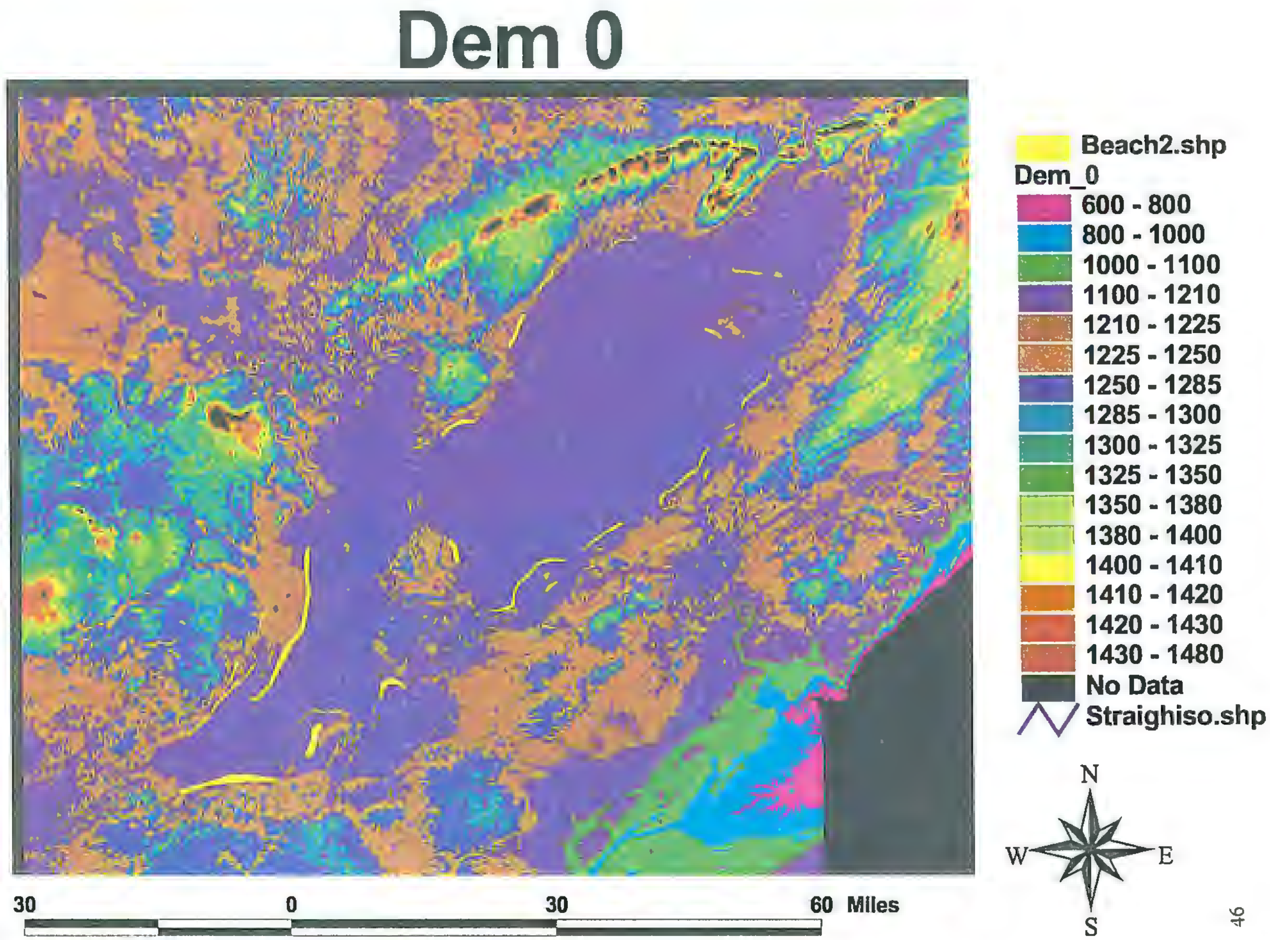


Figure 25a Digital Elevation Reconstruction of Lakes Aitkin and Upham; Maximum correction

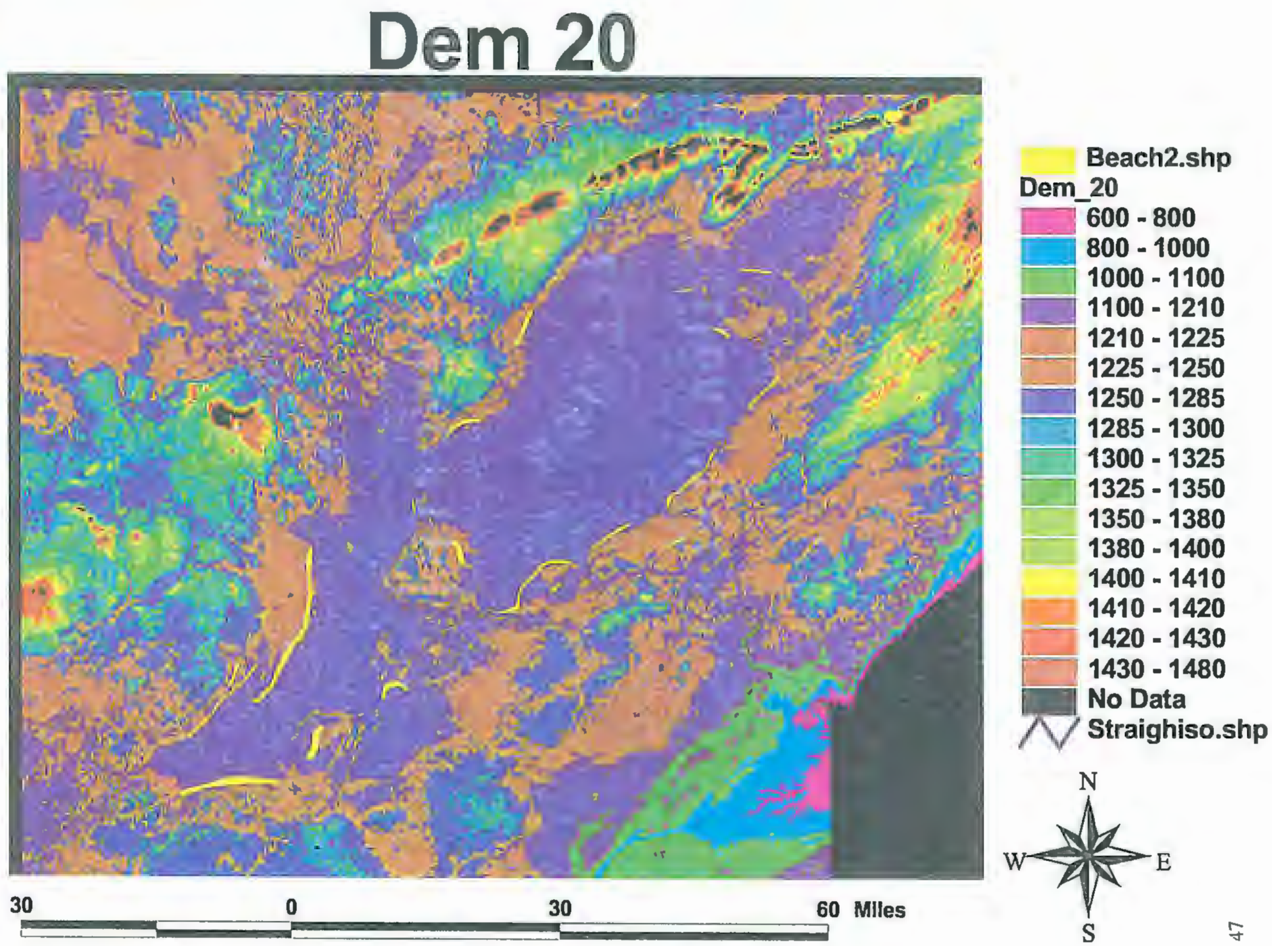


Figure 25b Digital Elevation Reconstruction of Lakes Aitkin and Upham; 22 ft of uplift.

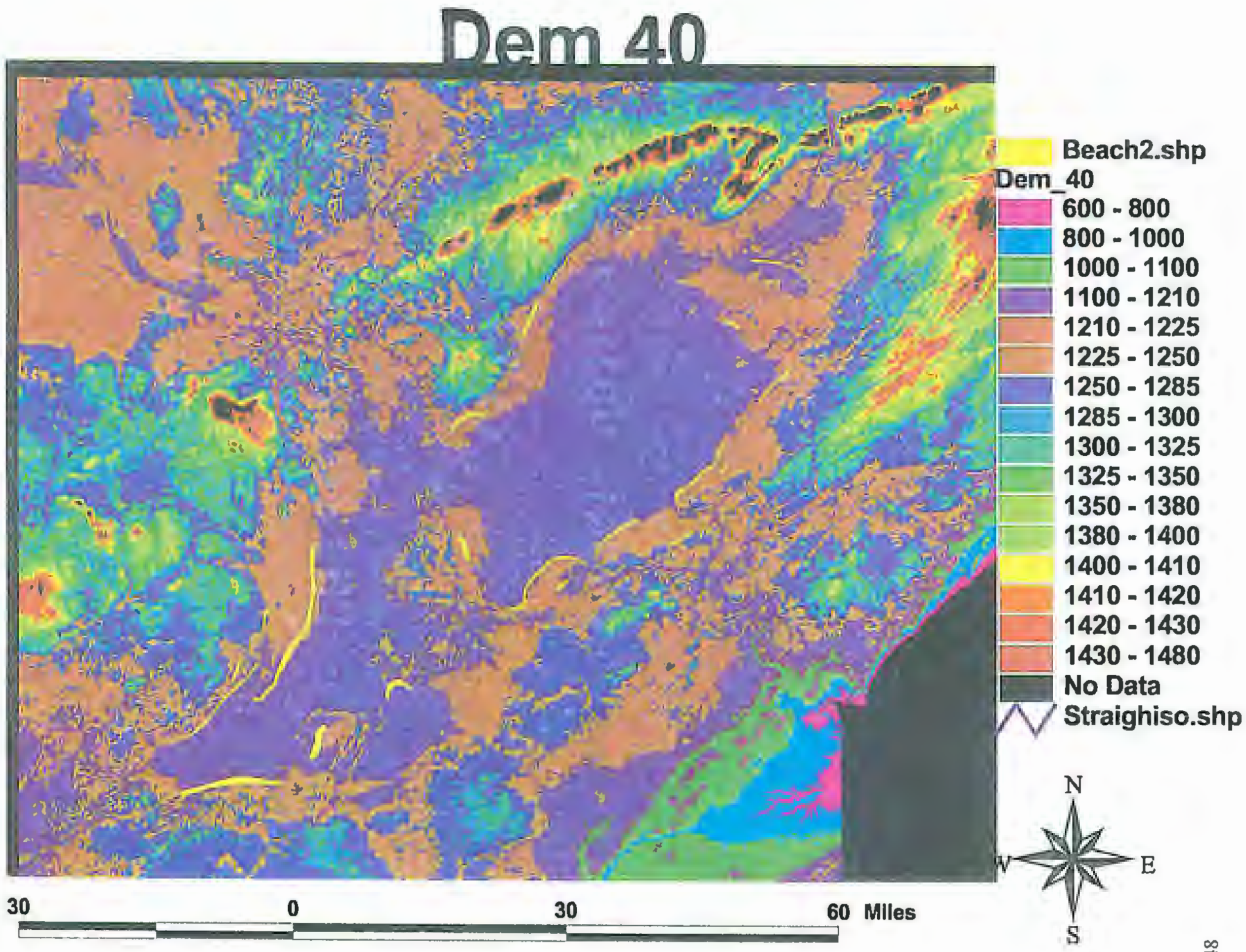


Figure 25c Digital Elevation Reconstruction of Lakes Aitkin and Upham; 44 ft of uplift.

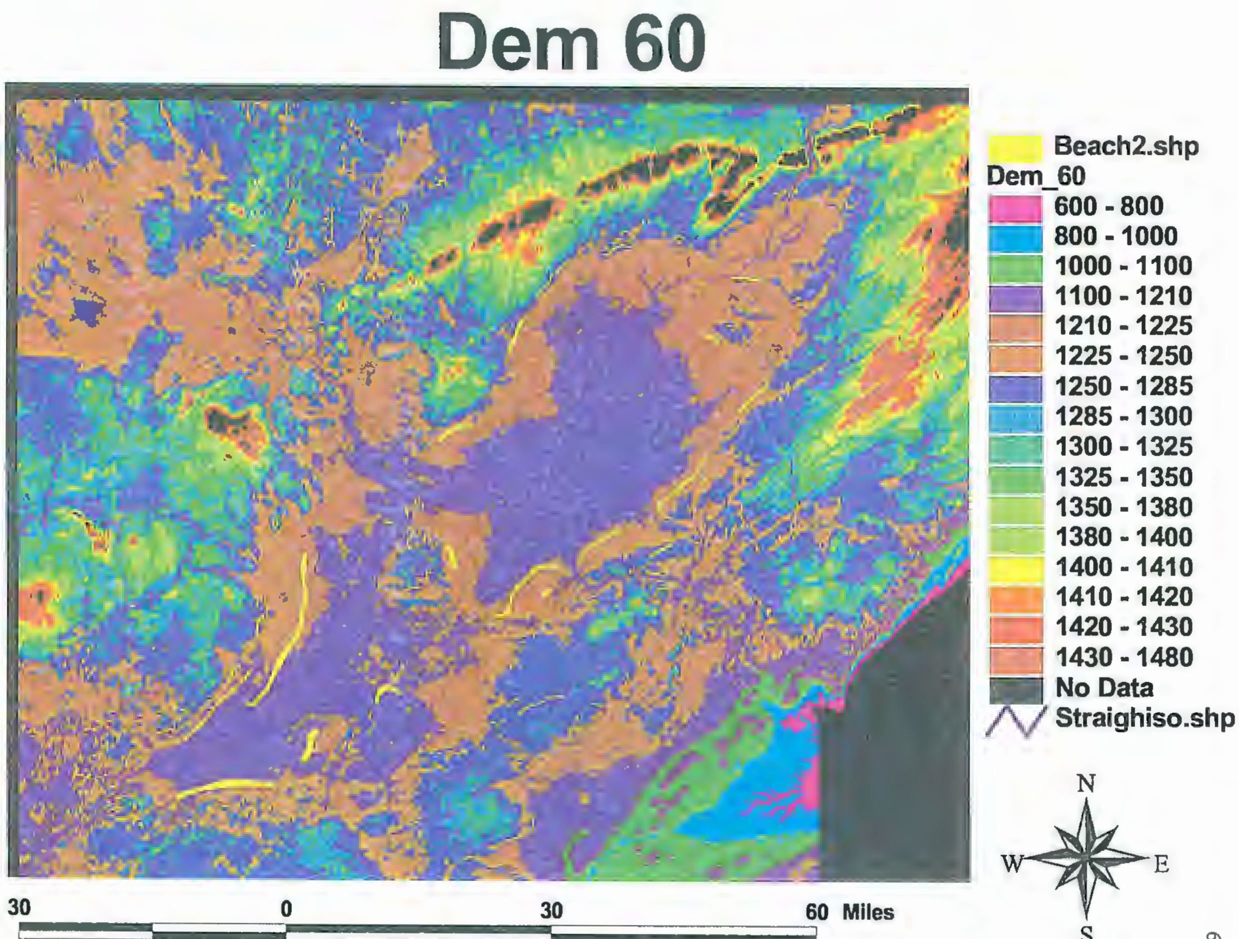


Figure 25d Digital Elevation Reconstruction of Lakes Aitkin and Upham; 66 ft of uplift

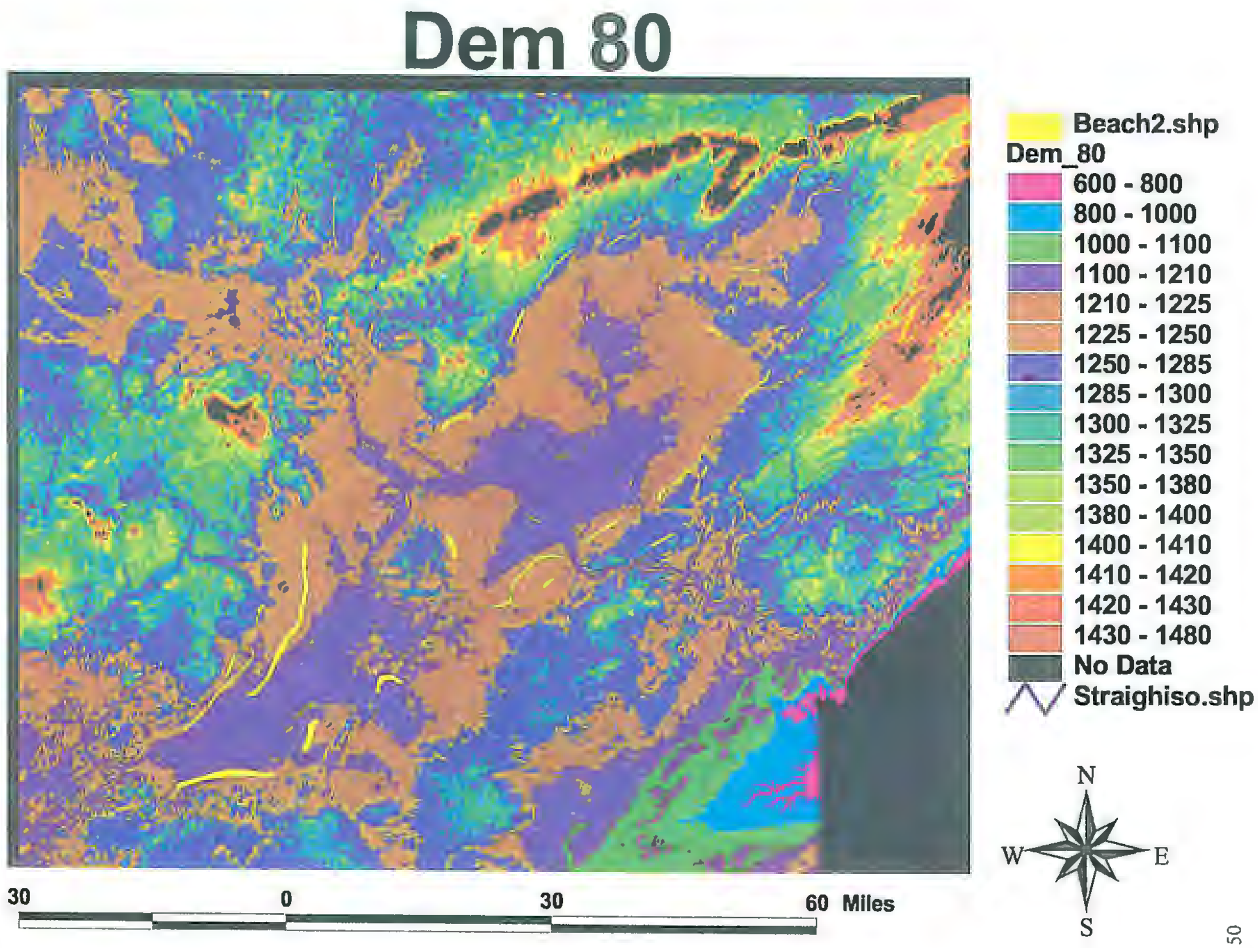


Figure 25e Digital Elevation Reconstruction of Lakes Aitkin and Upham; 88 ft of uplift

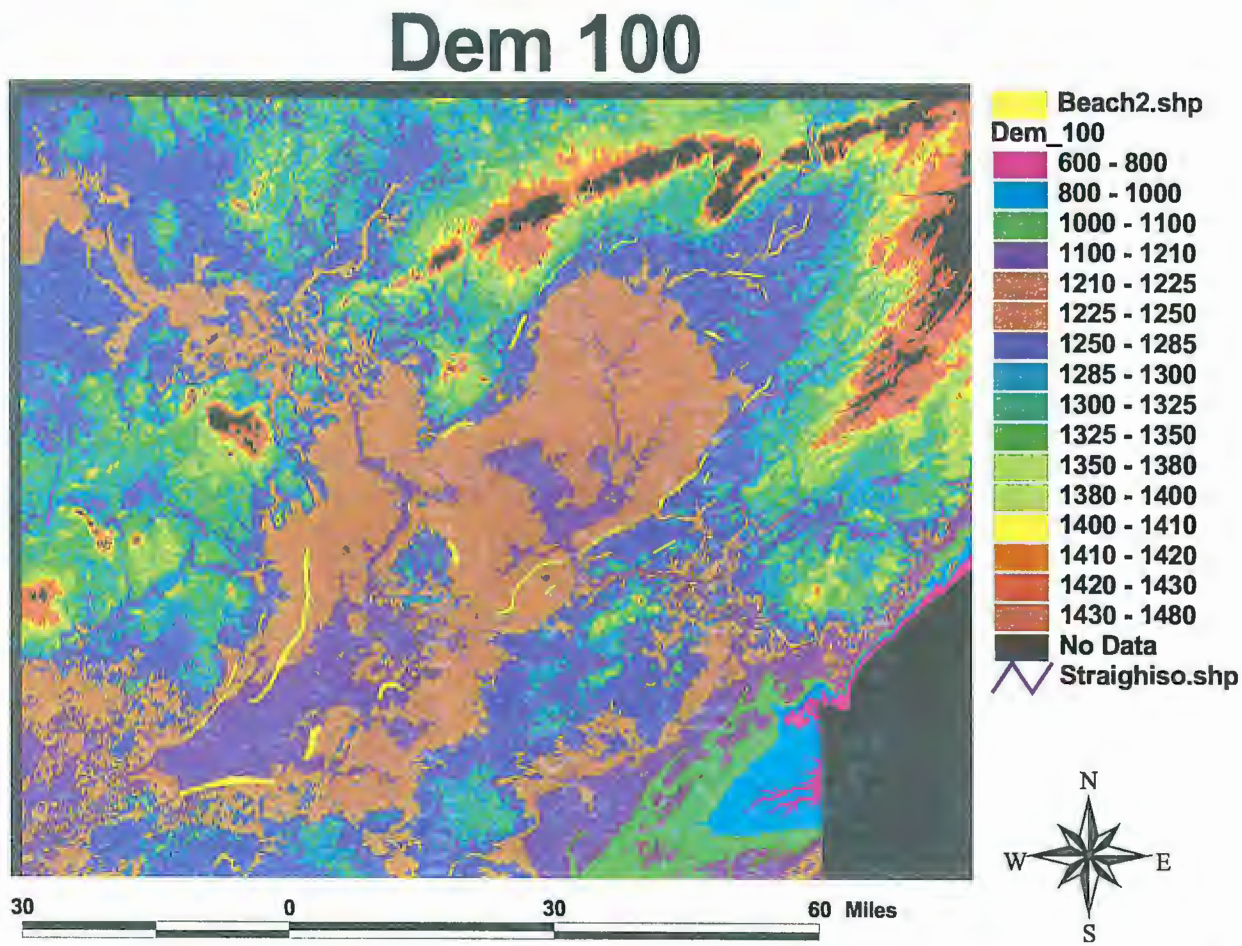


Figure 25f Digital Elevation Reconstruction of Lakes Aitkin and Upham; 110 ft of uplift.

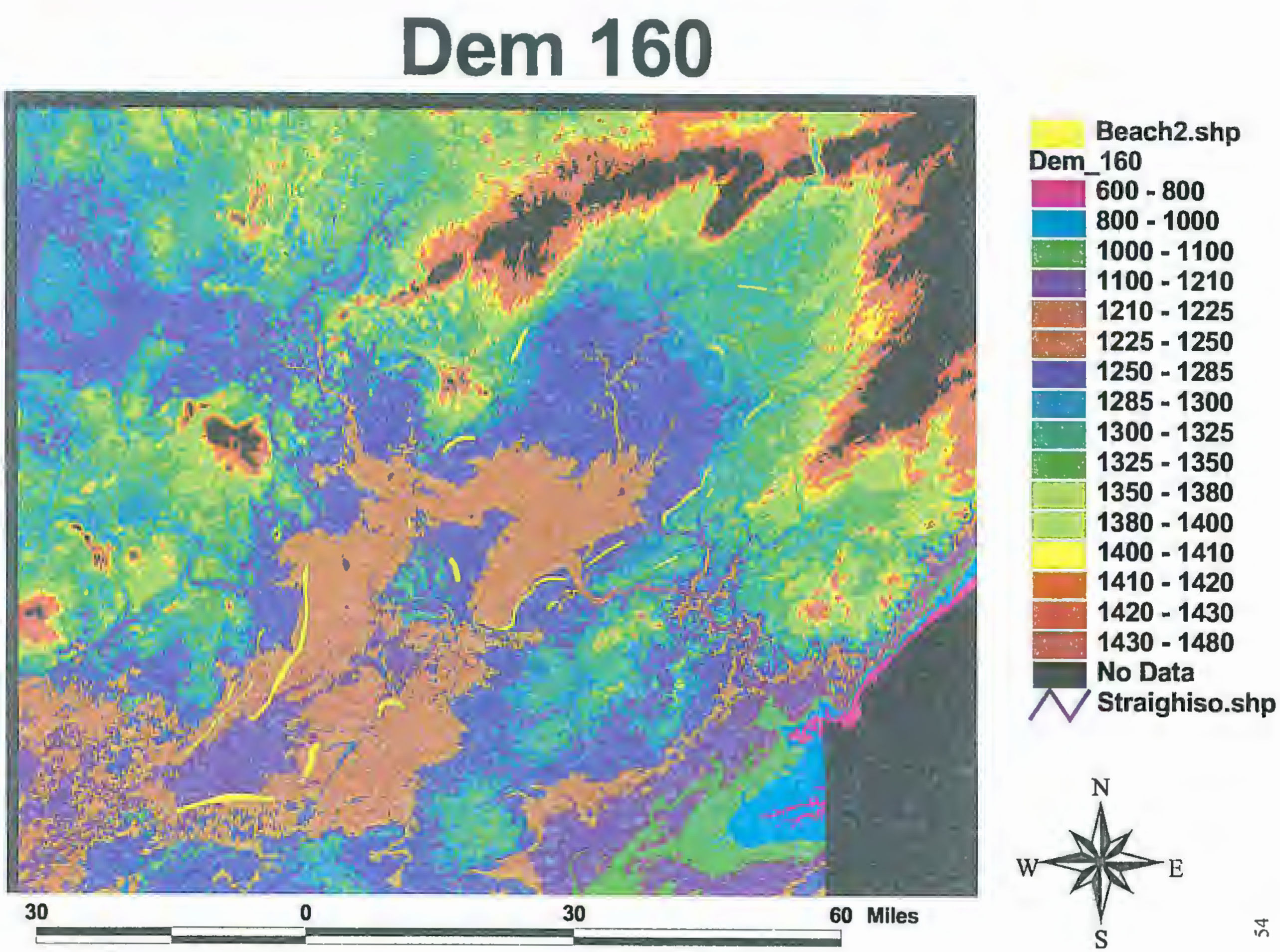


Figure 25i Digital Elevation Reconstruction of Lakes Aitkin and Upham; 176 ft of uplift.

Dem 180

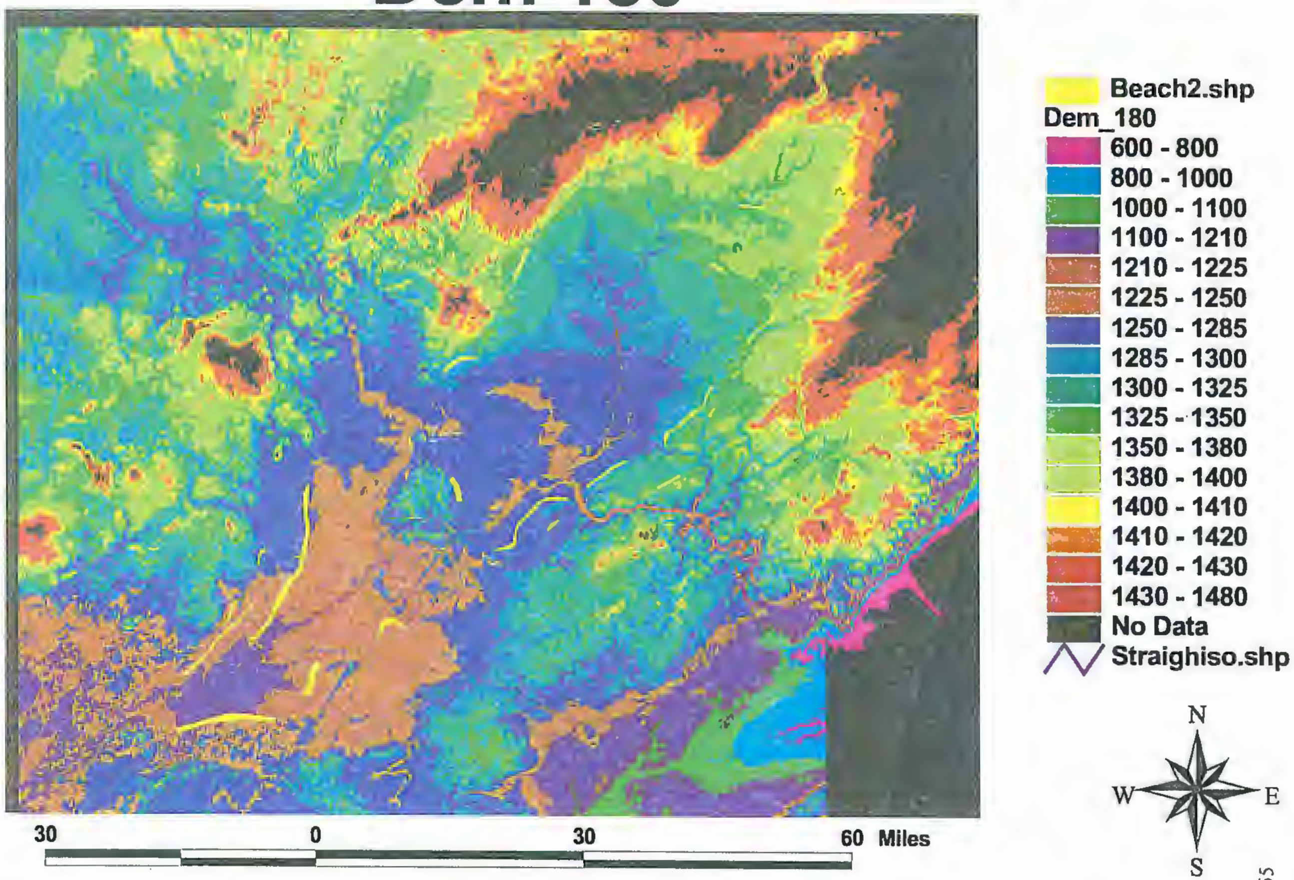


Figure 25j Digital Elevation Reconstruction of Lakes Aitkin and Upham; 198 ft of uplift.

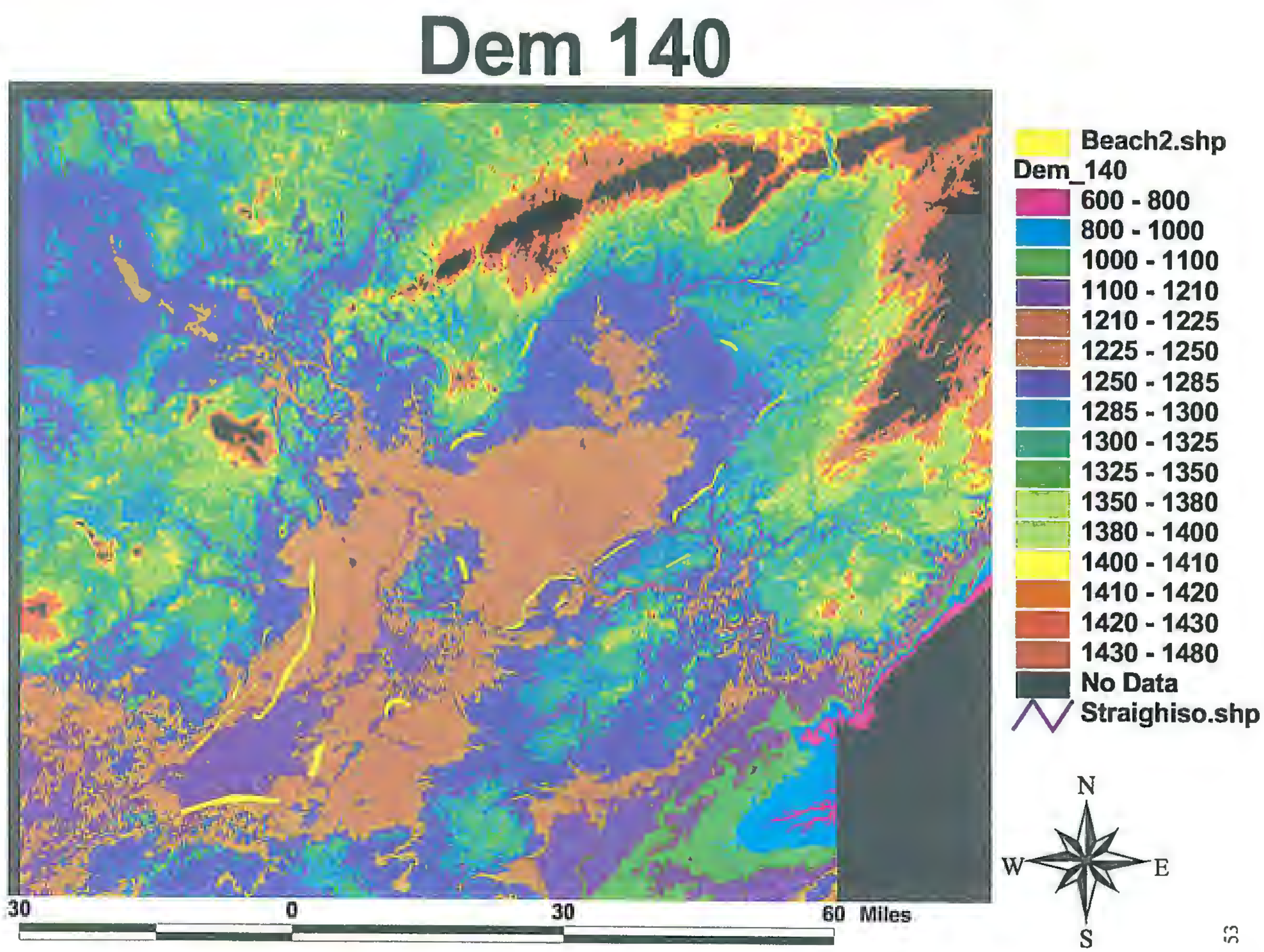


Figure 25h Digital Elevation Reconstruction of Lakes Aitkin and Upham; 154 ft of uplift.

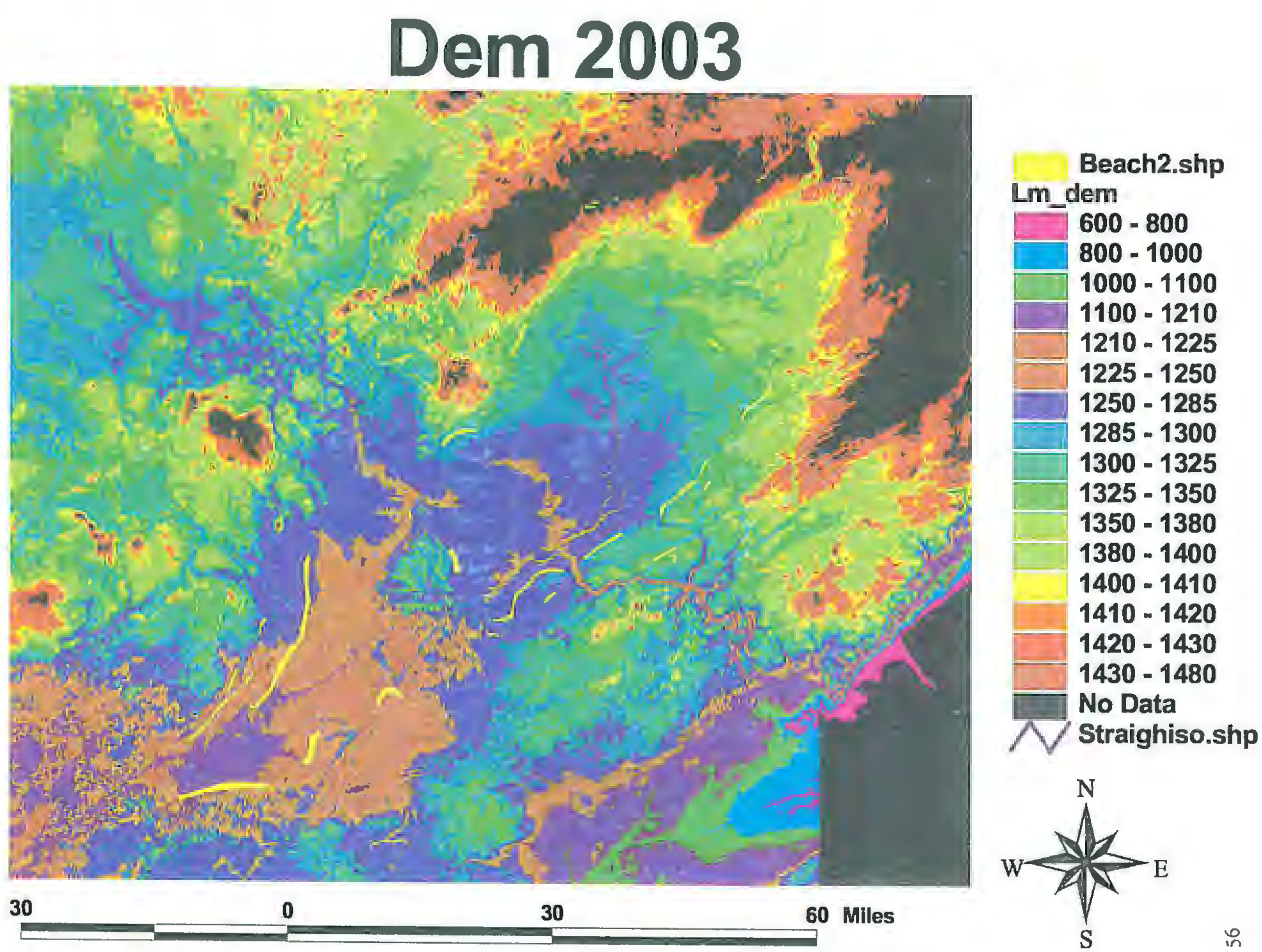


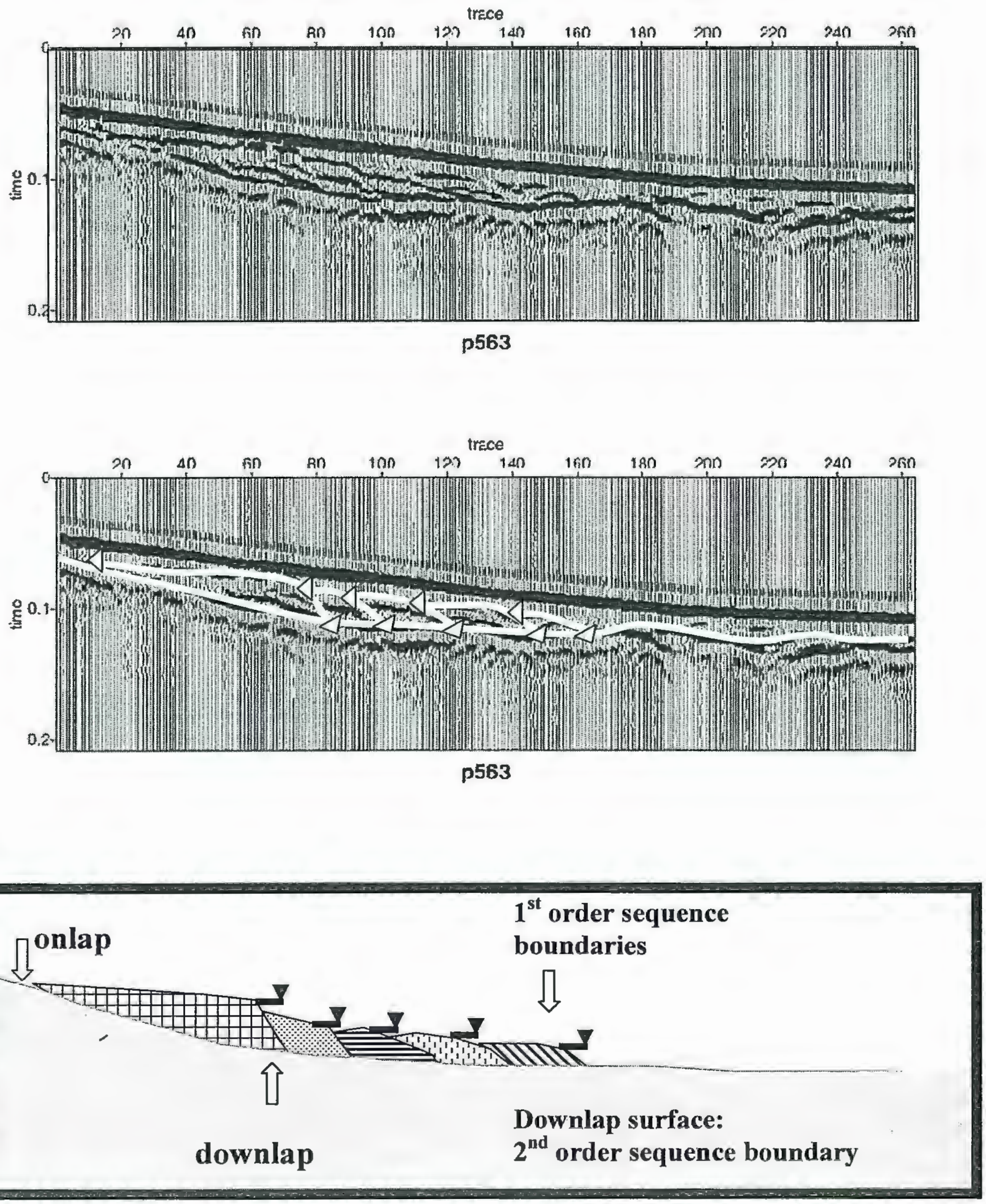
Figure 25k Digital Elevation Reconstruction of Lakes Aitkin and Upham; 220 ft of uplift.

4.11 Ground Penetrating Radar

The GPR penetrated to a depth of about 5 meters. below the surface before becoming attenuated by a clayey till. The deepest reflector is strong but discontinuous and displays an irregular surface that marks the base of the sands. Using terminology outlined by Bristow (1995), the Upham shoreline transect exhibits first and second order sequence boundaries (Fig. 26). First order sequence boundaries are characterized by horizontal, curved, and down-dip reflectors. They represent small unconformities that truncate against the second order sequence boundary. The second order sequence boundary is characterized by a strong continuous reflector at the base of the profile. It represents a larger and broader unconformity than the first order. The second order sequence boundary is interpreted as the till basement that slopes gently basin ward. It forms the downlap surface for later first order deposits.

The first order sequence boundaries are the smaller deposits that truncate onto the second order sequence boundary. There is a series of 5 first order surfaces that exhibit horizontal, curved, and down-dip basin-ward reflectors. These reflectors indicate bedforms that exhibit onlap (up-dip) and downlap (down-dip) that resulted from lake deposition. The 5 bedforms range in thickness from 1.5 m. to 1 m., as estimated from the time traveled.

Figure 26. GPR results of the Lake Upham shoreline tract.



4.2 Lake Core: Whole-Core Magnetic Susceptibility Record and Dates

Magnetic susceptibility ranged between 2 and 30 SI units. The largest values were at the base of the core, where three major peaks in magnetic susceptibility are observed in the bottom 1.3 meters. Magnetic susceptibility then dropped off to minimal values at about 6.5 meters and then rose steadily toward the core top with minor peaks and valleys. The three major peaks are within the massive carbonate-rich unit. They do not coincide with laminar beds of silt or sand that may have been deposited by slumping, therefore are considered to be deposited by eolian processes.

The peaks in magnetic susceptibility are all bracketed by the AMS dates. The AMS dates, both reported in C^{14} years B.P., were 10,108 +/- 74 at the lower extent and 6,635 +/- 56 at the upper extent.

Basal Date: At 8.2 meters down core
Date Measured May 23, 2002
Target # 1162 — sample wt (mg.) 0.96
mg C in target 0.56
AMS Lab Arizona — AMS lab # AA48637
$\Delta^{13}C$ value used: -25 0/00 PDB (estimated value from AMS lab)
C14 Fraction of Modern 0.2841 — Fm Error (+/-) 0.0026
REPORTED C^{14} AGE YRS BP *10,108*   AGE ERROR (+/-) *74*

Upper Date: At 6.9 meters down core
Date Measured May 23, 2002
Target # 1165 — sample wt (mg.) N/A
mg C in target 0.19
AMS Lab Arizona — AMS lab # AA48636
$\Delta^{13}C$ value used: -25 0/00 PDB (estimated value from AMS lab)
C14 Fraction of Modern 0.4378 — Fm Error (+/-) 0.0030

REPORTED ^{14}C AGE YRS BP *6,635*   AGE ERROR (+/-) *56*

According to calculations based on age-depth record and the magnetic susceptibility peaks, there was increased influx of clastic sediments circa 9,800, 9,300, and 7,400 yr B.P. (Fig. 27). There are no peaks after 6,635 yr B.P.

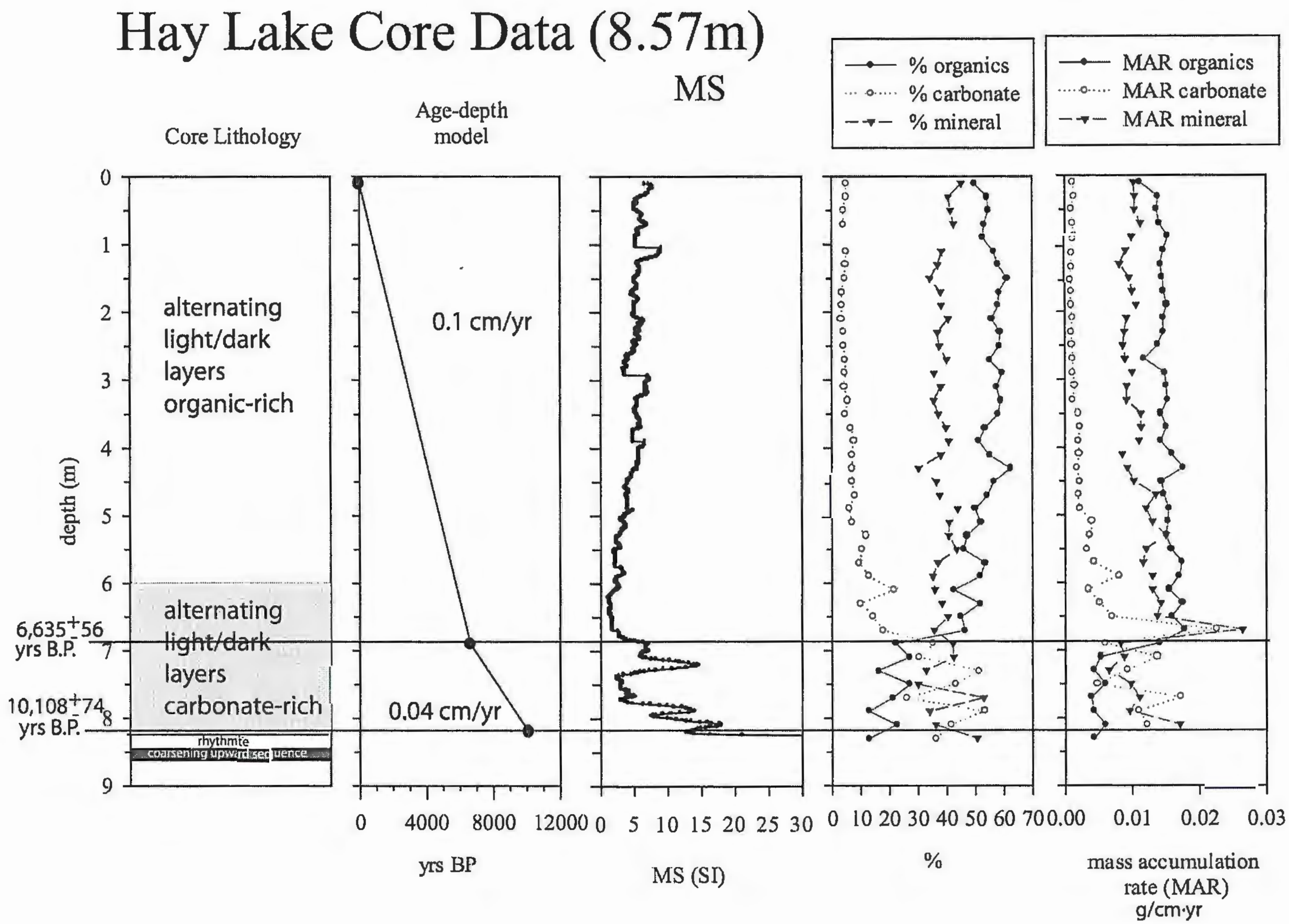


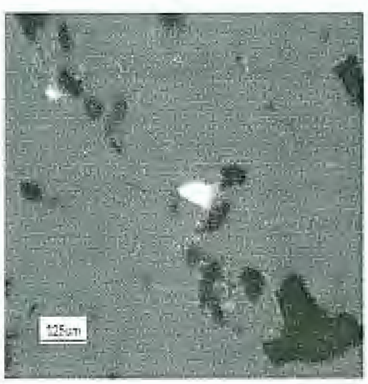
Figure 27. Hay Lake sediment core (8.57m): 1. Lithology 2. Age-depth model 3. Magnetic susceptibility 4. Composition 5. Mass accumulation rate.

4.21 Lake Core: Sediments

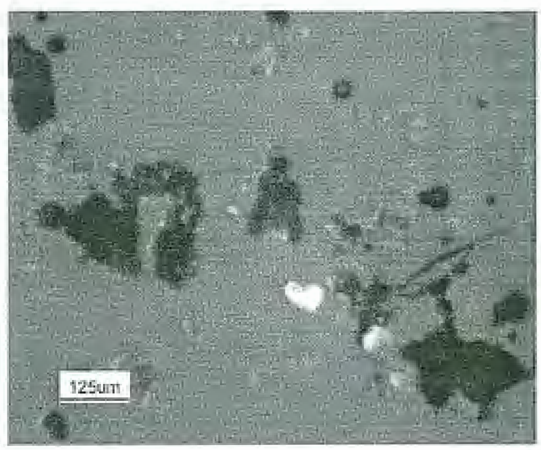
The basal sequence of the core, from 8.57 to 8.46 (11 cm), is a coarsening upward sequence of fine to coarse grain sands. Directly above this sequence from 8.46 to 8.3 (16 cm) is a series of 4 rhythmites consisting of alternating layers of fine sand and silt (Fig. 27). The layers have varying thicknesses with the sand layers slightly thicker than the silt layers. Above the rhythmites from 8.3 there are alternate light and dark layered sediments. The layers are of varied thicknesses and display no recognizable thickness pattern. The rhythmites abruptly transition to marl at 8.3 m. The ratio of inorganic carbon to organic carbon varies inversely and there is a gradational change to predominantly organic carbon at 6.3m.

Particular attention was paid to smear slides in areas where magnetic susceptibility peaked. The clastic grains were typically silt to very fine sand quartz with some heavy minerals (Fig. 28).

Sedimentation rates are greater (0.1 cm/ yr) in the upper part of the core between 0 and 6.9m and less in the lower part (0.04cm / yr) between 8.2 and 6.9m.



Quartz grain at 7.19m down core (at peak magnetic susceptibility)



Quartz grains at 7.88m downcore

Figure 28. Smear slide of Hay Lake sediments; quarts grains at 7.88 m.

Chapter 5 ANALYSES

5.0 Discussion

The Aitkin and Upham basin, which was scoured by the over-riding Rainy and Superior Lobes from the northeast, was occupied by two phases of glacial lakes during late Wisconsinan glaciation. The retreat of the Rainy and Superior Lobes from their maximum positions resulted in a series of ice-cored recessional moraines that bordered the Aitkin and Upham basin and led to the ponding of water and subsequent formation of Glacial Lakes Aitkin and Upham I (Wright, 1972; Hobbs, 1983; Lehr and Hobbs, 1992). The retreat of the ice from the Mille-Lac and Outing moraines (Mooers, 1988) led to the formation of Glacial Lake Aitkin I (Fig. 3). Continued retreat of the Rainy lobe from the Sandy Lake moraine led to the formation of Glacial Lake Upham I. Although the exact age of Glacial Lake Aitkin and Upham I is not known it was after the Rainy Lobe reached its maximum limit at the St. Croix moraine,~15,500 yr B.P. (Clayton and Moran, 1982; Mooers and Lehr, 1997), and before the St. Louis sublobe advanced into the Aitkin and Upham basin from the northwest.

The timing of the Rainy Lobe retreat with respect to the St. Louis sublobe advance is recorded in an outcrop at Cherry, MN (Fig. 4). Sediments here are interpreted as subaqueously deposited Rainy Lobe outwash overlain by St. Louis sublobe till that contains Lake Upham I sediments that include St. Louis sublobe sediments and fine-grained Rainy Lobe lithologies (Fig. 5). This implies that the Rainy Lobe was still south of the Giants Range when the St. Louis sublobe advanced into the Aitkin and Upham basin.

The St. Louis sublobe ice in the basin stagnated and the earliest phases of Glacial Lake Aitkin and Upham II came into existence. During this earliest phase water drained around the ice margin, possibly through a chain of small lakes, and drained through the highest outlets, the Uskabwanka and Chicken channels in the Upham basin and possibly the Snake (1250 ft./ 381m.) channel in the Aitkin basin (Hobbs, 1983)(Fig. 4 and 29). The elevations of the drainage channels are given in parenthesis, some of which are taken from Hobbs (1983), refer to the channel bottom. These drainage patterns are shown on DEM 0 (Fig. 25). The aforementioned outlets lie above the highest main strandlines of Lake Aitkin and Upham, therefore, they were likely associated with small ice-dammed lakes along the margin rather than the main stages of the lakes.

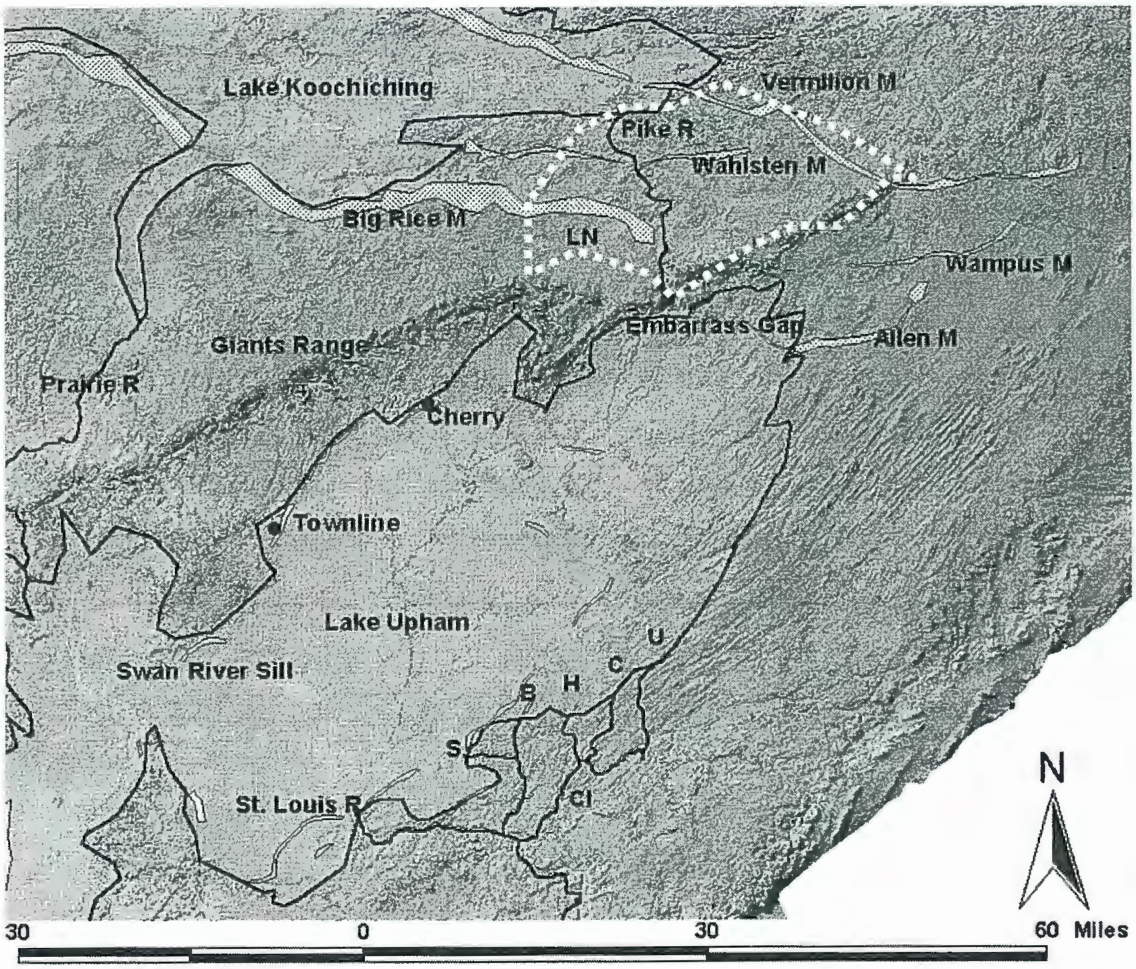


Figure 29 Significant features of the region. LN: Lake Norwood extent is indicated by a dashed line.

Continued retreat of the Rainy Lobe resulted in the ponding of water between the retreating ice an the Giants Range forming Glacial Lake Norwood (Fig. 29) (Winchell, 1901; Hobbs, 1983; Lehr and Hobbs, 1992; Bjork, 1988). Further retreat of the ice margin to the north led to the expansion of Lake Norwood and the deposition of several recessional moraines: Big Rice/Allen, Wahlsten/Wampus and Vermilion from south to north (Fig. 29) (Lehr and Hobbs, 1992). Lake Norwood drained south through the Embarrass Gap where it entered Lake Upham II at an elevation of 1450 ft. / 443 m. (Hobbs, 1983). This was the first major inlet to Glacial Lakes Aitkin and Upham. During this time the lake had a significant amount of stagnant ice in the basin and it is probable that Lakes Aitkin and Upham were separated by stagnant ice along the Swan River Sill.

Lake Norwood increased in extent and the level dropped to 1430 ft./ 436 m. and eventually to 1400 ft./ 427 m.. Hobbs (1983) identifies this new stage as Lake Koochiching (Fig. 29). Lake Koochiching drained south along the Pike River and through the Embarrass Gap into Lake Upham (Hobbs, 1983; Lehr and Hobbs, 1992; Leverett, 1932). During this period the upper beaches (1380-1350 ft. / 420-411 m.) in the NE most part of Lake Upham formed. At this level the outlets for Lake Upham were likely through the Hellwig (1330 ft. / 406 m.), Birch (1320 ft. / 401 m.), and Spider (1300 ft. / 397 m.) channels (Fig. 24 DEM 0 and Fig. 29) The elevation of the outlet channels and the elevations of the upper beaches in Lake Upham are not directly comparable here because of the effect of isostatic rebound.

After the time of upper beach formation, Lake Upham experienced a drop in water level documented at Townline beach (Fig. 26 and 29). The GPR results are interpreted as a down stepping of shoreline deposits. The progradation of the bedform resulted in a constructional shoreline and indicates regression of the lake. The regression is an example of forced regression as discussed by Posamentier (1999). Forced regression takes place when there is a relative sea-level fall that progressively exposes the sea (or lake) floor, thereby causing the shoreline to migrate seaward (Posamentier and Allen, 1999, p. 23).

The lake level drop eventually led to the abandonment of the Hellwig, Birch, and Spider channels. Baker (1965, p. 602 and 608) documents a sequence of marl in the Spider Creek outlet on the eastern-most edge of the Upham basin with a date near the base of ~13,000±400 (W-1234) yr B.P (Fig. 30) (Hobbs, 1983). Since marl formation requires shallow still water, the marl must post-date the drainage through the channel. This would place the minimum age of Lake Aitkin and Upham, and therefore the maximum limit of the St. Louis sublobe, prior to ~13,000 yr B.P. This date is consistent with the other findings in this report. However, Baker (1964) expressed concern that the date was too old and could possibly have been contaminated by lignite.

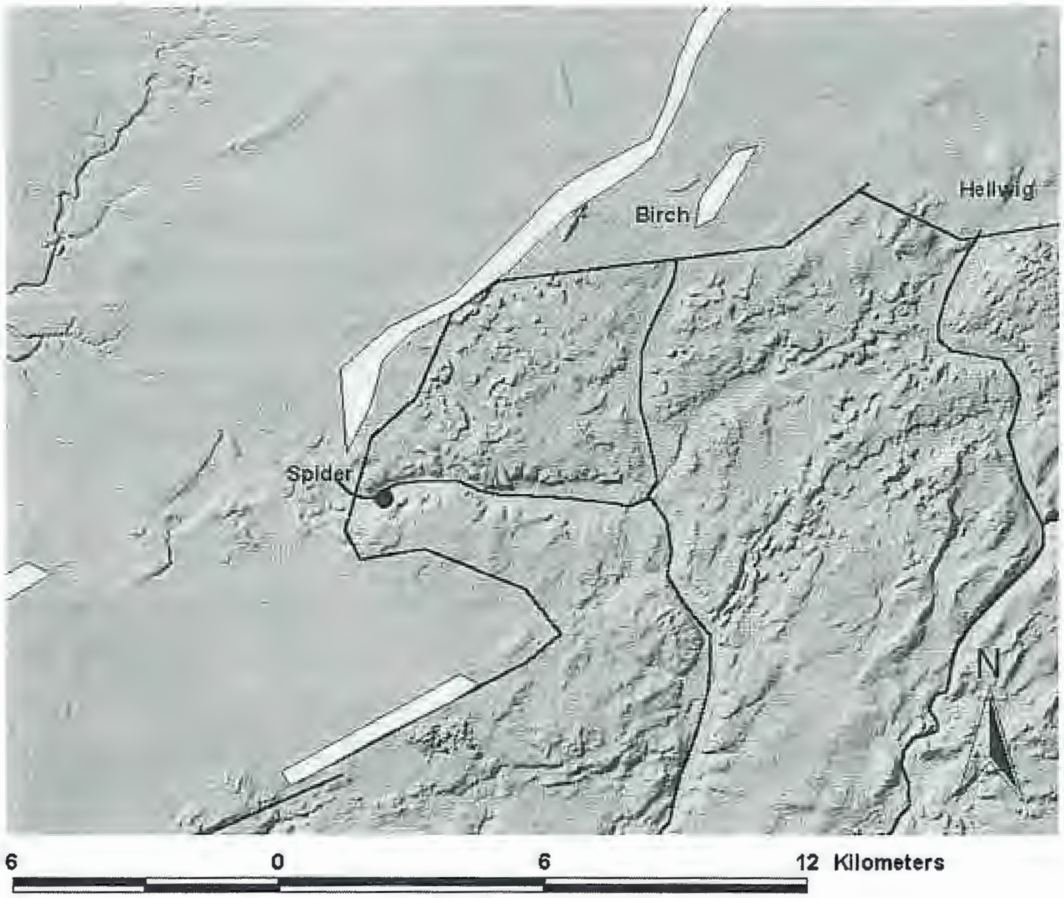


Figure 30. Three early outlets for Lake Upham; Hellwig, Birch, and Spider Creeks. Spider Creek marl location is marked with a circle. Lake Upham border delineated and polygons represent beaches.

At some time the inflow halted through the Embarrass Gap because lower outlets for Lake Koochiching opened a short distance to the SW along the Prairie River and flow was diverted into Lake Aitkin south of Grand Rapids, MN (Fig. 4 and 29). Although it is not known exactly when the Embarrass Gap was abandoned, it must have occurred by 10,200 yr B.P. on the basis of the transition from clastic to organic lake sedimentation in Sabin Lake, which is in the Embarrass Gap (Bjork, 1988, p. 34; Lehr and Hobbs, 1992).

Fenton suggests (1983, p. 52) that the Prairie River was initiated after 12,300 and before 10,800 yr B.P. based on the chronology of the Lake Agassiz basin to the west. Clayton (1983) suggests 11,500 yr B.P. for the Prairie River inception.

The Prairie River enters Lake Aitkin at an elevation of 1300 ft./ 397 m., which is when Lake Koochiching was at an elevation of 1350 ft./411 m. (Hobbs, 1983). The implications of the diversion from the Embarrass Gap to the Prairie River is a 50 ft. / 15 m. drop in the level of Lake Koochiching from its previous level at 1400 ft. / 427 m.. The large inflow associated with the Prairie River from Lake Koochiching resulted in the deposition of a large underflow fan that extends 50km outward from its apex (Hobbs, 1983). There are ~30 ft./ 9 m. of the underflow fan exposed down to a basal clay layer at the Mississippi River a few miles northwest of Jacobson, MN (Fig. 19). The age of the Prairie River inlet suggested by Clayton (1983), 11,500 yr B.P., is consistent with another record in Lake Aitkin. A paleosol near Aitkin, MN, with a date of 11,635 yr B.P., (Farnum et al., 1964) (Fig. 4) is overlain by a thin marl and 3 feet of clay. This deposit is interpreted to be the result of the beginning of the Prairie River inflow sometime after 11,635 yr B.P. Conditions in the Aitkin basin prior to the initiation of the Prairie River potentially reached a steady state at a low level, subaerially exposing some sediments resulting in soil formation as is documented by Farnum et al. (1964). However, the initiation of the Prairie River would have led to a rise in lake level even if only temporarily because of the 50 feet adjustment in Lake Koochiching. A marl with dates that cluster around 10,000 year B.P. (Hobbs, 1983) sampled just southwest of the Swan River Sill, ~30 miles northeast of Aitkin, suggests continuity throughout the Aitkin basin at this time period.

An alternative scenario that may explain the stratigraphy at the Farnum et al. (1964) site would be the isostatic uplift of the northeast portion of the lake causing the western shore of Lake Aitkin to transgress.

Although it is possible that water exchange between Lake Aitkin and Upham across the Swan River Sill (Fig. 4) existed prior to the inception of the Prairie River inlet, the initiation of the Prairie River after 11,635 yr BP began the bulk of the water exchange between the lakes. The majority of the flow was from Lake Aitkin to Lake Upham in the direction of the St. Louis River, which was the likely outlet at this time. The flow from Lake Aitkin to Lake Upham is supported by the existence of a spit just southeast of the Swan River Sill, herein named the Ball Bluff Spit (Fig. 31). The spit extends west to east, from the western border of Lake Upham where it intersects the Sill toward the St. Louis River outlet. On the spit there is clay overlain by 1 meter of fine-grained underflow sands. This deposit is interpreted as the inception of the Prairie River and subsequent inundation of sediment over a lake floor previously in a lake-clay depositional environment. During this time extensive beach deposits were forming in Lake Aitkin and Upham, DEM 20-40 (Fig. 25).

Once Lake Koochiching began to flow into Lake Climax, the Prairie River was abandoned. Hobbs (1983) indicates that water could have still been flowing into Lake Aitkin from the Mississippi River, which has a prominent terrace at 1275 ft. (389 m.) Eventually, however, all flow into Lake Aitkin and Upham was cut off. Isostatic rebound continued to drain the lakes, which led to the separation of Lake Aitkin and Upham. Ultimately, the two lakes drained completely through their respective outlets; Upham through the St. Louis River and Aitkin through the Mississippi River, DEM 60-80 (Fig. 25).

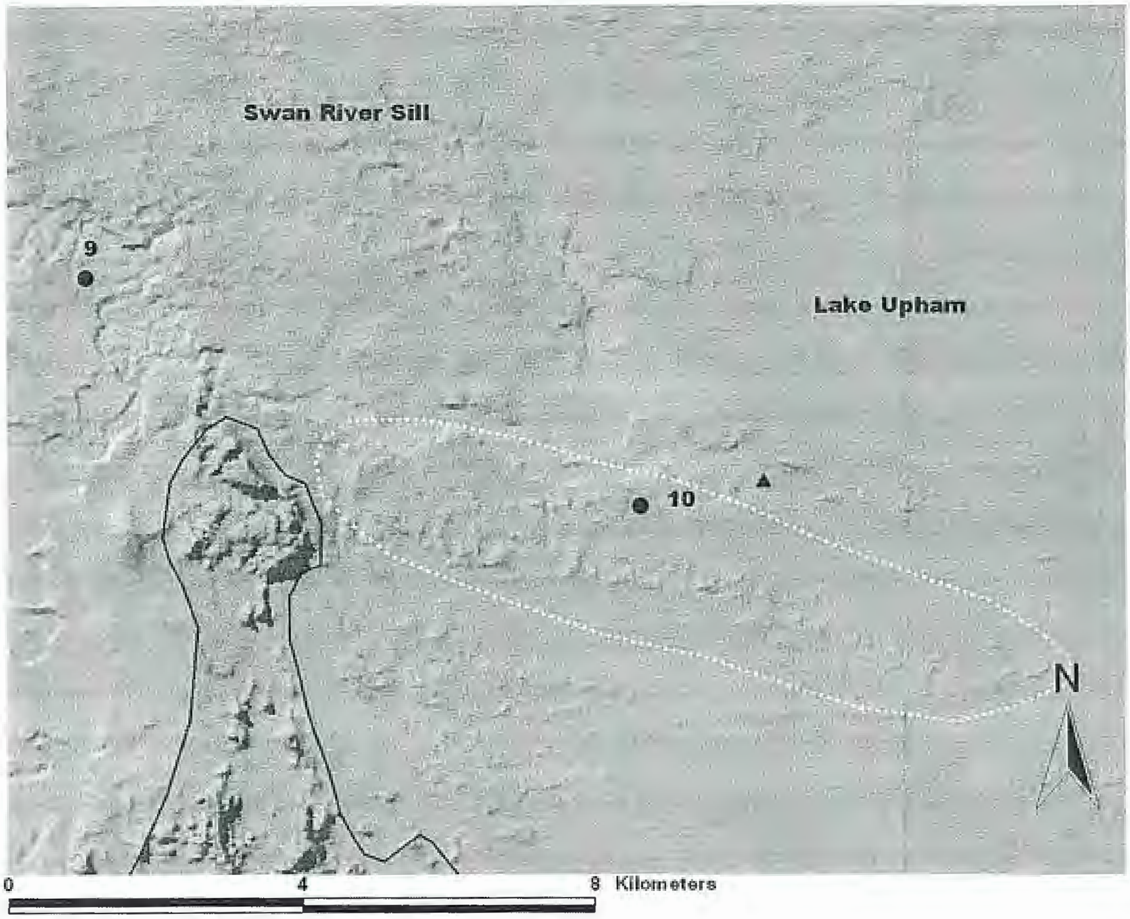


Fig 31. Ball Bluff Spit is within dashed line. Circle labeled 10 indicates where clay is overlain by 1 meter of underflow sands. Triangle indicates field site that was lake clay at the surface.

As the two lakes drained littoral areas of lake sediment became exposed and were subject to eolian activity. Dunes formed sporadically throughout the basin with respect to regions of appropriate sized sediment source. Dunes, as opposed to sand sheets or drifts, are defined by Bagnold (1941) as mound or hill of sand that rises to a single summit and occurs alone, attached in colonies, or in the form of chains. They are best developed on flat, featureless topography and are able to migrate while retaining their shape. The most well developed dunes have a 'slip face', the limit of steepness imposed by an angle of shear, a feature that is absent in most of the dunes in the Aitkin and Upham basin. Sand accumulation in the form of dunes is related to the relative strength, duration and direction of alternating periods of strong and gentle winds, which can generally be thought of as its ability to cause sand movement (Bagnold, 1941, p.178). Sand accumulation is also affected by several other factors: sediment source, wind (strength, duration, and direction), topography, climate, vegetation and hydrology.

Dunes in the Aitkin and Upham basin are most densely distributed near the Prairie River inlet in the Aitkin basin and across the Swan River Sill into the lower Upham basin. The 4 phi mean grain size of the dunes sampled in these areas in the basin match that of the underflow sediments, which have been mapped in the Aitkin and Itasca Soil Survey (Nyberg, 1999 and 1985) as Zimmerman, Cowhorn, and Wawina soils composed of fine to very fine sands of lacustrine origin. The dune clusters that extend beyond the Swan River Sill into Upham beyond the extent of the underflow fan can be explained by the transport of underflow sediments from Lake Aitkin to Lake Upham during the time of active water exchange between lakes. In fact, there is a prominent dune cluster directly on the Ball Bluff Spit (Fig. 18). The location of the dune clusters corresponds to the regions of sand source, which in the Aitkin and Upham basin correlates well with the location of Prairie River underflow sands.

The large number of longitudinal dunes and elongate clusters in a NW-SE direction suggest a NW-SE wind direction (Fig. 32). As Bagnold (1941) described "dunes tend to occur in belts or chains, whose direction coincides with that of the resultant long-period sand vector Q" (p. 281, 1941) with Q being sand flow because of the sum of the strong and gentle wind directions, and the width of the belt at right angles to Q. The absence of dune formation near the Embarrass Gap inlet could indicate a lack of appropriate sized sediment to form dunes, or possibly the topographic high of the Iron Range upwind of the basin blocked the wind from the Upham basin. The Iron Range rises 410 ft./ 125 m. in relief above the basin below, compared with 100 ft./ 30 m. of relief upwind of the Aitkin basin near the Prairie River inlet, which would affect the fetch of the NW-SE winds.

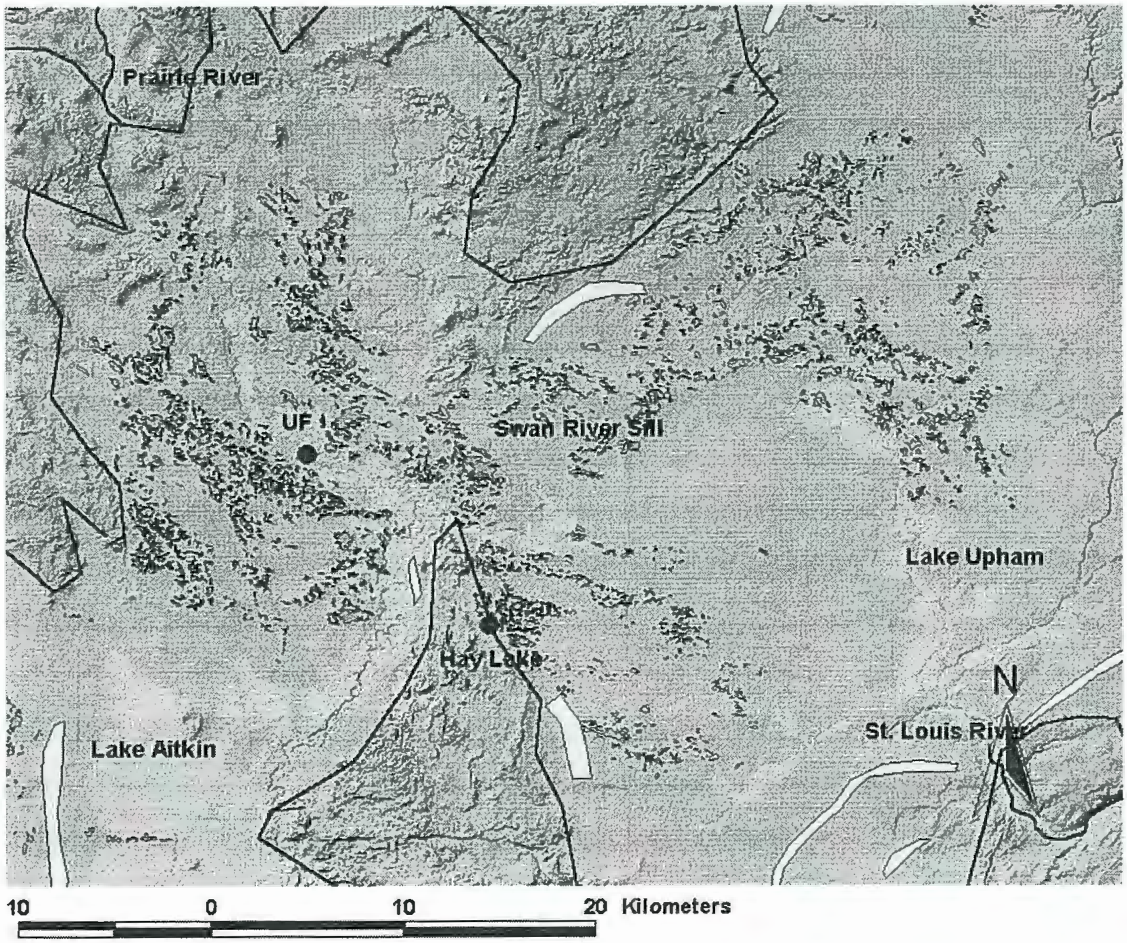


Figure 32. Trend in dune clusters indicate a NW-SE wind direction.

Dunes in the Aitkin and Upham basin are an example of the strong relationship between source sands and dune formation. Though the Aitkin and Upham dunes were likely forming continually as the littoral regions of the lake were exposed, the primary periods of dune formation were coincident with the eolian activity recorded in the Hay Lake sediment core at 9,800, 9,300 and 7, 300 yr B.P. Near Hay Lake, Lake Upham sediments are sufficiently exposed by DEM 40-60 (Fig. 25) and could account for the two earlier eolian peaks, 9,800 and 9,300 yr B.P. However, Lake Aitkin sediments are not sufficiently exposed until DEM 80 (Fig. 25) and could account for the latest eolian peak of 7,300 yr B.P.

Dunes throughout the Midwest and elsewhere have been interpreted to have formed upon exposure of source either of a valley train of a river with declining water levels or in a post glacial pro-glacial landscape (Holliday, 1995; Winspear and Pie, 1996; Mush and Holliday, 2001; Warren, 1976; Mush et al., 1999; Seppala 1993). However, in many cases those dunes are vegetated then subsequently reactivated as a result of changing climatic conditions, as was common throughout the Mid-Holocene (Lemmon et al., 1998; Wolfe et al., 2000; Win spear and Pie 1996; Stokes and Gaylord, 1993). This was particularly commonplace and still occurs today (Wolfe, 2001) in existing arid environments and Prairie ecotone regions. However, in Minnesota where we have a diverse climatic regime that contains Prairie, Deciduous, and Conifer/Hardwood ecotones it is not obvious that a climatic shift to warmer drier conditions would be enough to destabilize a landscape and cause dune formation. It is conceivable that warmer drier conditions could lower rivers and lakes exposing their littoral areas to dune formation as happened at Lake Winnibigoshish (Grigal et al., 1976). This, however, would limit the locations in which dunes would form.

5.1 Conclusions

The two phases of Glacial Lakes Aitkin and Upham formed as a function of glacial ice retreat and isostatic rebound. The evolution of the lakes is summarized in figure 33.

The dunes in the Aitkin and Upham basin formed as the lakes drained incrementally exposing littoral areas to eolian processes. Since the discovery that the

Winnibigoshish dunes are middle Holocene in age (Grigal, et al., 1976), other studies have used this revelation in combination with their research to imply that many of the dunes in Minnesota were also formed during this period. Dean (1997) and Keen and Shane (1990) both have strong middle Holocene eolian signals in lake sediment cores using different proxies. Keen and Shane (1990) use their magnetic susceptibility record from a lake core to suggest Lake Ann dunes formed during the middle Holocene. However, given the circumstances that led to the Winnibigoshish dunes (Grigal et al., 1976) and what would be involved in forming the Lake Ann dunes, it is a viable comparison. The Winnibigoshish dunes formed when drawdown of the lake exposed littoral areas to eolian activity, which led to the formation of dunes. In the Lake Ann region, however, there would have needed to be devegetation before the formation of any dunes would occur. Even with the shifting of the Prairie/Forest border during the middle Holocene, it is unclear how shifting to a Prairie ecotone equates with dune formation. It is more likely, given the proximity of the lake cores studied by Dean (1997) and Keen and Shane (1990), ie. Elk Lake and Ann Lake (Fig. 2), to the Prairie/Forest border that their records actually record dusty conditions on the Prairie and not necessarily dune formation during the middle Holocene. The Aitkin and Upham basin, however, which is more distal from the shifting Prairie/Forest border, can accurately record the local conditions. The strong correlation between dune location and source sands combined with the record of eolian activity in the Hay Lake sediment core show that the Aitkin and Upham dunes formed immediately upon exposure of source sands. The peaks of magnetic susceptibility at 9,800 and 9,300 yr B.P. coincide with the estimated time of drainage of Lake Upham when sands would have been susceptible to eolian activity. The magnetic susceptibility peak at 7,300 yr B.P. coincides with the final drainage of Lake Aitkin.

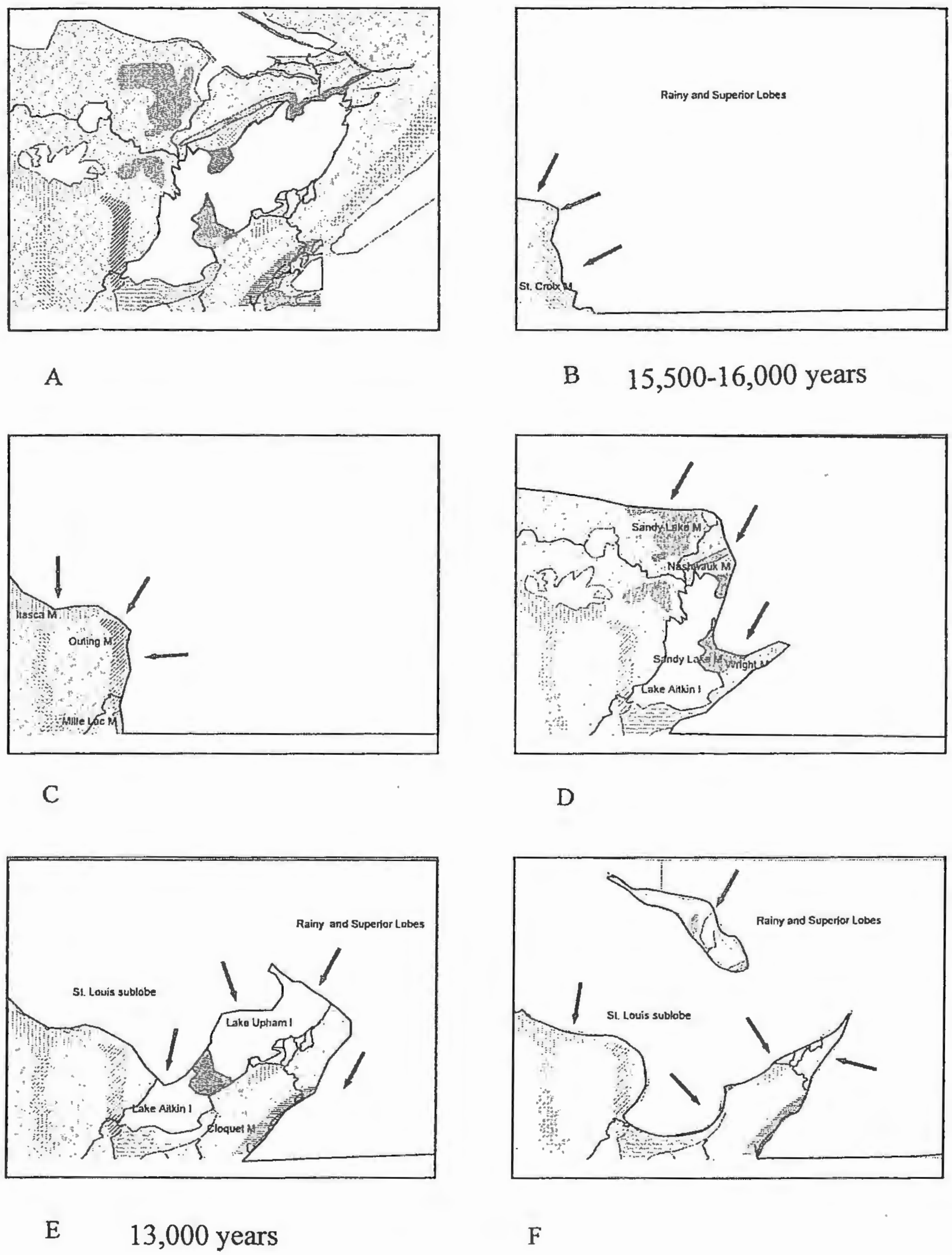
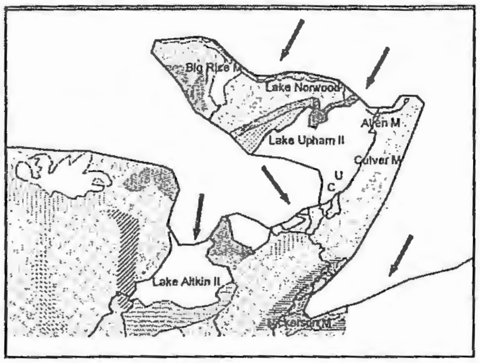


Figure 33. Glacial History of the Aitkin and Upham basin.



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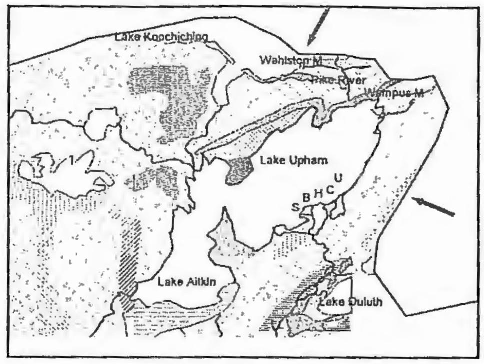
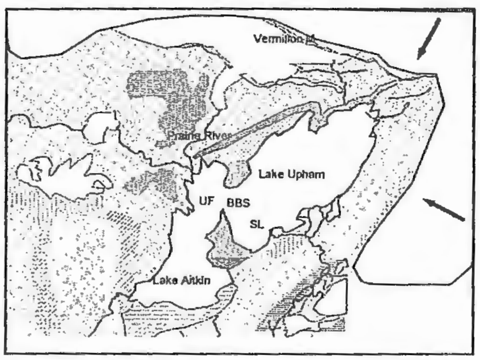


H 11,700 years



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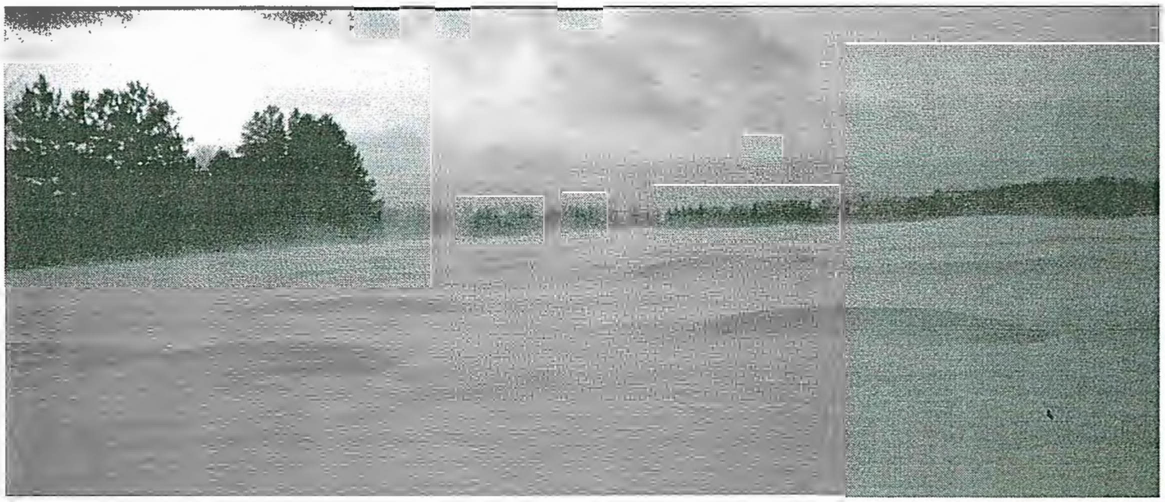

Figure 34 Dunes forming in the Upham basin today- snow dunes.

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