GLACIAL GEOMORPHOLOGY OF THE SOUTHWEST SEGMENT OF THE CHIPPEWA LOBE MORAINE COMPLEX, WISCONSIN

DISSERTATION FOR THE DEGREE OF PH. D.

MICHIGAN STATE UNIVERSITY

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1976

This is to certify that the

thesis entitled

GLACIAL GEOMORPHOLOGY OF THE
SOUTHWEST SEGMENT OF THE CHIPPEWA
LOBE MORAINE COMPLEX, WISCONSIN

presented by

Adam Clifford Cahow

has been accepted towards fulfillment
of the requirements for

Ph.D. degree in Geography

Major professor

Date 17 May 1976

O-7639

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ABSTRACT

GLACIAL GEOMORPHOLOGY OF THE SOUTHWEST SEGMENT
OF THE CHIPPEWA LOBE MORAINE COMPLEX, WISCONSIN

By

Adam Clifford Cahow

This thesis is a detailed description and interpretation of the geomorphology of the southwestern segment of the Chippewa lobe end moraine located in Chippewa, Clark, Rusk, and Taylor Counties of Wisconsin. The study area encompasses twenty-five survey townships and covers about 900 square miles. A middle Woodfordian (Cary) age has been proposed for the Chippewa moraine located within the area.

In terms of specific objectives, the thesis represents an investigation of four major and six minor geomorphological problems. The major problems were (1) whether the middle Woodfordian end-moraine zone consists of a single moraine or is comprised of several contiguous or superposed moraines, representing more than one phase of ice-front advance and retreat; (2) to what degree, if any, the segment of the Chippewa lobe ice margin represented in this area experienced stagnation during its final wastage; (3) whether or not the high-relief morainic topography located southwest of Flambeau Ridge (a prominent bedrock high) may be of interlobate origin; and (4) the nature of the assemblage of landforms that make up the area. The minor problems that were examined include (1) the influence of the bedrock topography on ice movement, (2) regional variations in the size of the moraine, (3) the late Quaternary sequence of events in the area, (4) the extent and relationships of the loess to the geomorphic features in the area, (5) the sources

of the glacial sediments, and (6) the distribution and relative ages of periglacial features in the area.

The solution of the problems described above depended upon determining the topographic, lithologic, stratigraphic, and distributional characteristics associated with the glacial sediments. Also determined were the grain size, color, and weathering characteristics of the drift. Both field and laboratory investigative techniques were used, with emphasis upon the former. The landform mapping involved a careful examination of the available topographic maps and air photos, and an intensive investigation of morphologic and sediment conditions in the field, supplemented by laboratory grain size analysis of numerous sediment samples (using the Bouyoucos hydrometer method). Stratigraphic and drift-thickness information was obtained chiefly from examination of water-well-log data.

The significant findings of this study are: (1) the Chippewa moraine is a moraine complex, consisting of several stagnant-ice moraines, with each probably representing a separate minor oscillation phase of the Chippewa lobe; (2) marginal stagnation of the Chippewa lobe is indicated and most of the landforms in the Chippewa moraine are of stagnation origin; (3) the well-developed morainic topography southwest of Flambeau Ridge appears to be of interlobate origin, formed between two minor sublobes that were produced by a bifurcation of the ice flow at Flambeau Ridge; and (4) the Chippewa moraine complex contains a great variety of stagnant-ice landforms, collectively arranged into orderly landform assemblages.

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Adam Clifford Cahow

A DISSERTATION

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

DOCTOR OF PHILOSOPHY

Department of Geography

1976

ACKNOWLEDGMENTS

A great many persons and agencies have been helpful in the preparation of this dissertation, but I am particularly grateful to Harold A. Winters, Ph.D., my advisor, for his encouragement and untiring constructive review of my manuscript. I am also indebted to Professor Gilbert Tanner, University of Wisconsin - Eau Claire, for his technical assistance in the preparation of the illustrations in this thesis.

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CHAPTER I

INTRODUCTION

The Area

The area investigated (Figure 2) comprises most of the portic Chippewa County which lies north of Wisconsin Highway 29 and east United States Highway 53, four townships in southernmost Rusk Cour four townships in westernmost Taylor County, and the northwesternm township of Clark County. Almost square-shaped, the region is six townships or thirty-six miles across both north-south and east-wes and covers an area of about 900 square miles. The objective of th thesis is to describe and interpret the glacial deposits and landf of the southwestern segment of the Chippewa lobe moraine zone (Fig an area that has never been the subject of intensive geomorphic st or mapping. Frye, Willman, and Black (1965) indicate a probable m Woodfordian (Cary) age for the Chippewa moraine, the age accepted here. However, recent findings of a study concerned with the Supe lobe (Wright, et al., 1973) indicate that the Chippewa moraine may older (see Chapter II).

The moraine zone which lies within the thesis area ranges fro to twelve miles in width, and typically has from ten to eighty fee local relief per square mile; in a few places, however, the latter exceeds 150 feet. In general, altitudes range between 1100 to 120 feet in the ground moraine area north and east of the moraine, 105 1300 feet within the moraine, and 950 to 1100 in the outwash and c ground moraine on its distal margin.

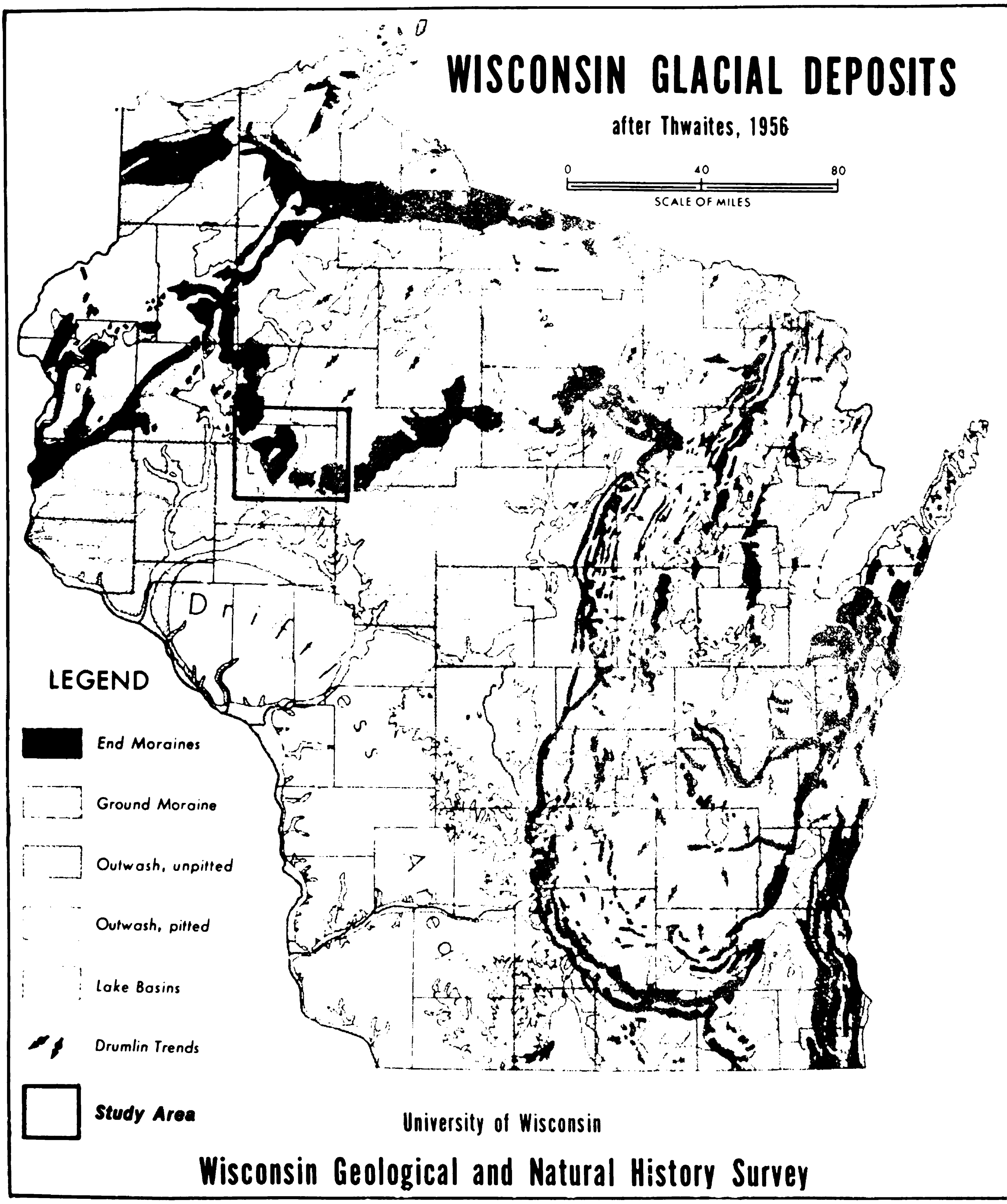
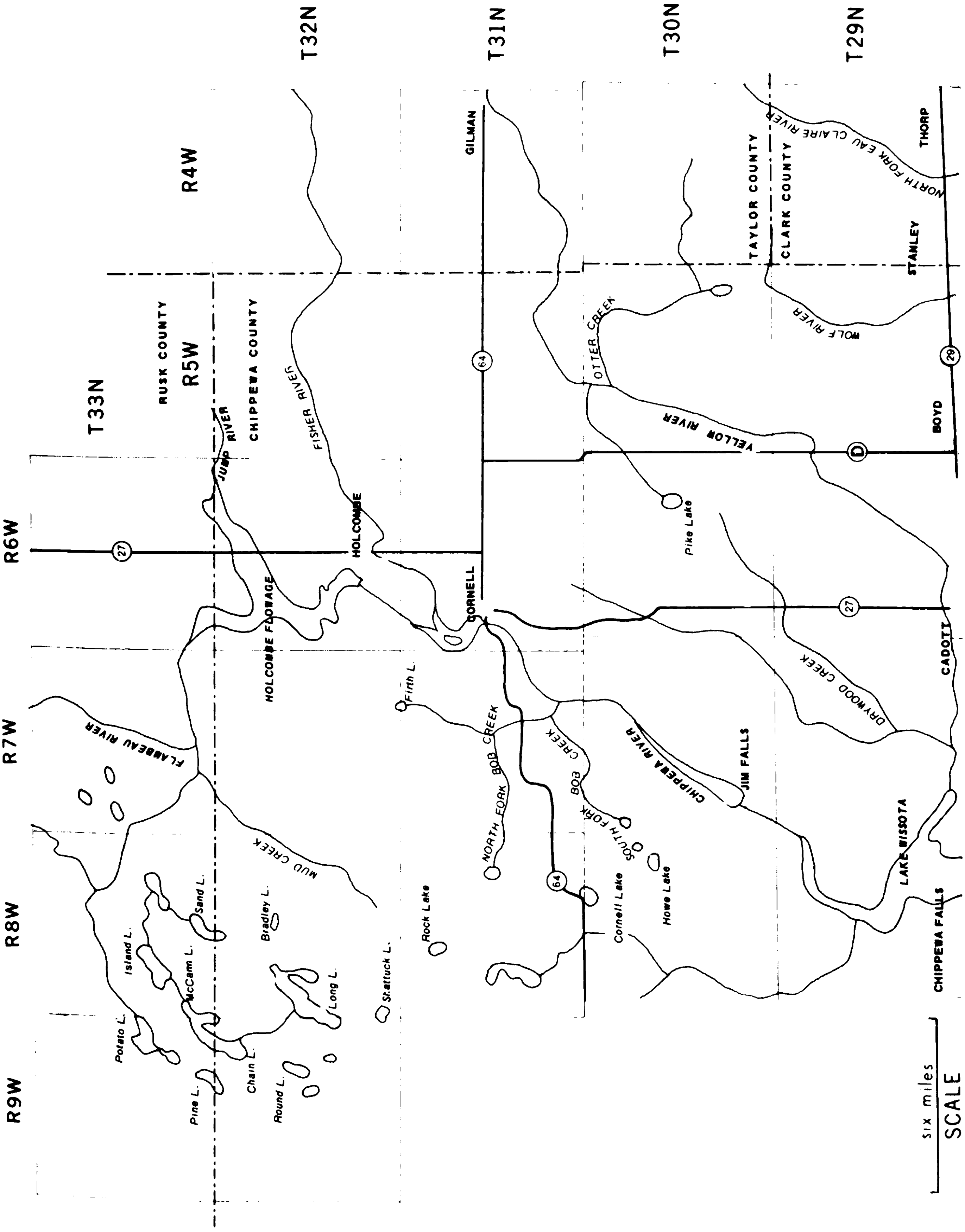


Figure 1. Wisconsin Glacial Deposits and location of the study area



T33N
T32N
T31N
T30N
T29N
R4W
R5W
R6W
R7W
R8W
R9W
RUSK COUNTY
CHIPPEWA COUNTY
TAYLOR COUNTY
CLARK COUNTY
GILMAN
THORP
STANLEY
BOYD
HOLCOMBE
CORNELL
CADOTT
JIM FALLS
CHIPPEWA FALLS
LAKE WISSOTA
JUMP RIVER
FISHER RIVER
HOLCOMBE FLOWAGE
FLAMBEAU RIVER
MUD CREEK
OTTER CREEK
YELLOW RIVER
WOLF RIVER
NORTH FORK EAU CLAIRE RIVER
DRYWOOD CREEK
CHIPPEWA RIVER
BOB CREEK
NORTH FORK BOB CREEK
SOUTH FORK
Pike Lake
Firth L.
Rock Lake
Cornell Lake
Howe Lake
Sand L.
Bradley L.
Island L.
McCann L.
Long L.
Stattuck L.
Potato L.
Chain L.
Pine L.
Round L.
27
64
29
D
six miles
SCALE

The drift associated with the Chippewa lobe consists predominan of a reddish brown sandy loam till. Igneous and metamorphic bedrock of Precambrian age underlies the drift in the northeastern part of t study area, and thin Cambrian sandstone underlies the drift in most the western and southernmost parts of the area. A map by Martin (19 p. 32), showing physiographic provinces of Wisconsin, indicates that the region covered by this study is within the southern margin of th Northern Highland and the northern fringe of the Central Plain. Lob (1950) designates the former province the Superior Upland.

The Chippewa River drains the entire study area. At numerous places along its course within Chippewa County it has cut through th drift into the bedrock. At these localities the drift exposures alo the valley sides range from 20 to 100 feet in thickness.

Problems Investigated

This study sought to resolve several major and minor geomorphic problems recognized in the area by the author.

Major Problems

1. A major objective was to determine if the Chippewa lobe end mora zone, shown on the Glacial Deposits of Wisconsin map (Figure 1) as c prised of a single moraine, actually consists of several contiguous superposed moraines, representing more than one phase of ice-front advance and retreat, or, in effect, a moraine complex.

2. A part of the study was to ascertain whether or not the end mora zone southwest of Flambeau Ridge (Figure 6) includes an interlobate tract.

3. I sought to resolve to what degree, if any, the segment of the

Chippewa lobe ice margin represented in this area experienced stagnation during its final wastage.

4. An effort was made to determine the nature of the assemblage of landforms that make up the area.

Minor Problems

1. To what degree did the bedrock topography control ice movement and formation of various landforms?

2. Why is the Woodfordian morainal zone constructed by the Chippewa lobe of apparent greater magnitude west of the Chippewa River than it is to the east?

3. What was the late Quaternary sequence of geomorphic events in this area?

4. What is the extent of loess and its relationships with the glacial sediments in the area?

5. What were the sources of the glacial sediments found within the study area?

6. Do periglacial features exist within the area?

Previous Work

Numerous writers have made incidental or general references to the glacial deposits and landforms associated with the Chippewa lobe, but only a few detailed studies have been made. Weidman (1907) and Hole (1943) discussed the district immediately to the east of the thesis area, and Mathiesen (1940) was concerned with the region adjoining the northwestern part of the study area. More recent are Master's theses by Jacobson (1969), Larson (1972), and Ranney (1964), concerned with areas located northwest, northeast, and east of the present study area respectively.

The only recent study concerned with landforms in at least part of the thesis area (Black, 1974) describes its westernmost part in a general manner and is based on limited fieldwork. To the writer's knowledge, the Chippewa lobe, among those glacial lobes recognized in north-central United States, is by far the least studied. For the thesis area, in particular, there has been no previous detailed analysis of glacial landforms.

Explanation of the drift deposits of northern United States resulting from extensive continental glaciation began in the last two decades preceding the Civil War. In an 1848 geologic report Owen refers to the "drift" or northern Wisconsin as apparently associated with some great deluge or diluvial epoch. A subsequent report by Owen (1852) includes a chapter by Whittlesey describing the drift deposits of the Wisconsin uplands near Lake Superior which states that "they are doubtless . . . connected with recent movements and upheavals on the north shore of Lake Superior, attended by intense heat, by which causes the ancient waters were powerfully agitated and charged with mud and coarse gravel" (ibid., p. 426). About this time Whittlesey began to question the diluvial theory of drift origin, attributing the drift cavities or "potash kettles" of Wisconsin to the burial and later melting out of blocks of glacier ice, and the scratches on the underlying bedrock to moving glacier ice (Whittlesey, 1860, p. 299). Four years later, Whittlesey (1864) described in much greater detail the characteristics of glacial drift and the work done by the continental glaciers. He also described numerous examples of logs and plant remains buried at various depths in the drift and exposed in wells, excavations, or along r and lake bluffs. These observations led him to suspect more than glacial advance. This article (op. cit.) includes a rough map sh the presence of glacial drift in the area from Minnesota eastward New Jersey, but the lobate character of the moraines is not indic

Bean (1937) points out that the first official statewide geo survey of Wisconsin was authorized in 1873. Prior to that time t glacial features of the Chippewa lobe were virtually unknown. Th survey results, including descriptions of the glacial deposits, w published in four volumes (Wis. Geol. and Nat. Hist. Survey, 1877 Since then there has been no additional statewide geological surv and for much of the area glaciated by the Chippewa lobe the brief accounts of that endeavor still represent the best description av

An early outgrowth of the aforementioned survey, a paper by berlain (1876-1877), contains a map showing the location and loba configuration of the Wisconsin Kettle (terminal) Moraine for an a extending from Minnesota to Ohio. Chamberlain (ibid., p. 210) st that

> within the Chippewa valley the moraine has been observed by F. H. King of the Wisconsin Survey. . . . This region is cov ered by an immense forest, mainly unsettled and untraversed, even by foot paths, so that geological exploration is diffic and expensive, and, as no industrial importance attaches to and the rock below is deeply concealed by it, I have not dee it sufficiently important to trace the belt continuously to fy the large expenditure of time and means requisite, especi as I entertain no serious doubts as to its continuity and ge position.

Chamberlain recognized more than one phase of glaciation, for he scribed a zone of glacial drift in central Wisconsin located sout the main Kettle moraine (sic) which appears older than that north of t moraine (op. cit., p. 226).

Strong (1880, p. 383) prepared a map for Volume III of the above survey, outlining the general extent of the end moraines of the Chippe and Superior lobes. The glacial formations of the entire state, including brief references to those deposited by the Chippewa lobe, are discussed in Volume I (Chamberlain, 1883a, p. 261-291). Chamberlain (1883b) summarizes what was known concerning continental glaciation at that time. He states (ibid., p. 313) that the margin of the continent ice sheet was organized into twelve major lobes, each producing a seri of great looping moraines. He also discussed the Chippewa lobe (op. cit., p. 381-382), attributing it to a flow of ice from the Lake Super basin. A later study, by Chamberlain and Salisbury (1885-1886), inclu a generalized map on which are outlined the extent of the older and younger drifts surrounding the Driftless Area and another showing the distribution of surficial deposits and generalized topography of the Chippewa valley from the Woodfordian moraine downvalley for about fift miles.

The most comprehensive investigation of the glacial geology in an area peripheral to the thesis area, that of Weidman (1907, p. 427-607) concerns the adjoining area on the east. One of the maps contained in this report (ibid., p. 434) shows four different drift sheets in weste and central Wisconsin.

Few investigations of the Chippewa lobe were published between th time of the Weidman report and World War II. One of these (Leverett, 1929) proposed a revision in the age of central Wisconsin glacial formations as given by Weidman. In this article Leverett (ibid., p. 19) identifies and delineates an area of Kansan drift southwest of Eau Claire but recognizes an "undifferentiated drift" between it and the main Wisconsinan moraine of the Chippewa valley.

Lithological analyses of the drift in eighty-seven townships, including some in Chippewa County, were undertaken by Hotchkiss, et al. (1915), as part of a comprehensive mineral lands survey. Similar investigations (Wis. Geol. and Nat. Hist. Survey, 1916-1919), by other survey teams, were conducted in the same area. More than a decade later Martin (1932, 2nd ed.) authored a book which has become the standard reference on the physical geography of Wisconsin. It contains extensive bibliographies on the glacial and other landforms of the state.

After 1940, publications on the Quaternary of Wisconsin were more abundant. Two papers concerning areas adjacent to the thesis area are of particular interest. Mathiesen (1940) delineated the early Wisconsinan drift sheets and moraines in the reentrant between the Chippewa and Superior lobes. The district mapped by him borders the northwest corner of the present thesis area. A Ph.D. dissertation by Hole (1943) examined the older drift(s?) of north-central Wisconsin in terms of depth of leaching, grain-size variations, soil-profile development, and topographic expression, among other things; but his study did not extend into Chippewa County. The glacial deposits of Wisconsin are shown on a map by Thwaites (1956).

More recently research by Robert F. Black and graduate students working under his supervision has been concerned with this area. Relevant publications include Black (1959), in which it is implied that there are no end moraines older than early Wisconsinan in the zone north of the Driftless Area, and Black, et al. (1965), reassigning the older drift of the region recognized by earlier workers as pre Wisconsinan in age as actually being similar in age to the Altonia deposits of Illinois.

The subdivision of the Wisconsinan glacial stage, and naming the substages thereof, has undergone several revisions in the last decades. One of the first studies to apply the chronological subdivisions proposed by Frye and Willman (1960) to the state of Wisc was that of Oakes (1960), who investigated the Woodfordian moraine Rock County, Wisconsin. Black (1961, p. 137) states that no glaci deposits older than Wisconsinan are recognized in the state and th "many buried ice masses in the preglacial drainageways survived fr the late Altonian advance until . . . after the Valders glaciation the north." Akers (1961) researched the clay minerals of glacial deposits in west-central Wisconsin, elaborating on a paper by Blac (1959) concerning the glacial geology of the same area. Although detailed county soils report has been published, a small-scale "So Association Map" is available for Chippewa County (U. S. Soil Cons tion Service, 1961), on which nine broad soil associations are rec

Since World War II numerous periglacial features have been re nized in western Wisconsin. Studies concerned with these features include those of Smith (1949) and Black (1969b), who described per glacial features in the Driftless Area. Black (1964; 1965) deals periglacial phenomena for the state.

Two comprehensive summaries of the glacial history and deposi Wisconsin are Black, Hole, Maher, and Freeman (1965) and Frye, Wil and Black (1965). The latter publication (ibid., Figure 1) shows tentative Valders-age glacial deposits extending across the northe

part of the Chippewa River watershed, but a later study by Black
p. 173) states that the Valderan ice did not extend into northern
Wisconsin. Andrews (1965) treats the extensive terrace system as
ciated with the valley train of the Chippewa River and its tribut

The most recent paper in which the glacial landforms of the
area are described on a local basis (Black, 1974), was prepared f
National Park Service as a background report for the Ice Age Rese
The report contains a section (ibid., p. 151-180) dealing with so
the glacial features in the western part of the study area.

Other recent studies, relevant to the problems at hand, are
Black and Rubin (1967-1968), concerning radiocarbon dates for Wis
and by Foss and Rust (1968), the latter examining soil profiles i
western Wisconsin where the stratigraphic sequence is interpreted
post-Cary (late Woodfordian) loess, underlain by thin Cary till,
in turn is underlain by late Altonian till. Another soils study
Ranney and Beatty (1969) involves the glacial deposits of Barron
located immediately northwest of Chippewa County. In recent these
Ranney (1964) and Jacobson (1969) investigated soils in the middl
western parts of the Chippewa moraine, respectively, while Larson
describes the fluted ground moraine lying immediately north-north
of the area examined in this study.

Recent maps by Dutton and Bradley (1970) show the Precambria
geology of Wisconsin, and those by Young and Hindall (1972) deal
the water resources of the Chippewa watershed. Two other sources
information, not yet published, are an open-file soils report on
County, available at the Soil Conservation Service office in Eau
and an in-progress study by the United States Geological Survey

concerning the irrigation potential of the glacial outwash districts in a region overlapping the west end of the thesis area.

Discussion of Thesis Problems

During investigation of the problems listed on preceding pages four and five the emphasis was upon determining the topographic, lithologic, stratigraphic, and distributional characteristics associated with the glacial sediments in the area. Other aspects of the sediments, such as color, grain size, and depth of weathering were also examined. Both field and laboratory investigation were used, with emphasis upon the former.* The following discussion outlines the investigative techniques that were employed during the study. Also discussed are a number of publications which were useful references or guides for this investigation.

Subdivision of the Morainal Zone

The initial step in the investigation of this problem consisted of a detailed topographic and landform mapping of the end-moraine zone, based on an intensive field examination in combination with a thorough analysis of contour maps and stereoscopic aerial photographs of the area. While photographic coverage is complete, contour maps were not available for the northern fringe of the study area at a scale larger than 1:250,000.

As far as the identification of stagnant-ice moraines is concerned, the characteristics given for end and stagnant-ice moraines by Flint (1971) and Embleton and King (1968) were used as guidelines. Numerous

*The field work for this dissertation was conducted between September, 1970 and November, 1971, excluding winter months.

topographic characteristics were examined while attempting to d
entiate drifts, locate ice-front positions, and subdivide and d
the moraine zone. Studies which aided in resolving these probl
include that by Alden (1909), and those by Ruhe (1950; 1951), a
which discuss the use of regional variations in topographic and
characteristics as a means of discriminating drift sheets of di
ages. The studies by Weidman (1907) and Hole (1943), who consi
areal topographic and erosional differences in delimiting the d
boundaries in north-central Wisconsin, were also helpful.

Thwaites (1943) used a variety of criteria in reconstructi
various ice-front positions and fluctuations in eastern Wiscons
(ibid., p. 99) points out that

> the smooth outline of the ice fronts recorded by most mora
> indicates readvance, for a very irregular border would cer
> result from melting back. Some areas mapped as terminal m
> which are not linear may be overridden outwash.

Regarding high outwash terraces, Thwaites (ibid., p. 127)
that

> following the abandonment of the easternmost . . . moraine
> ice margin melted back rather rapidly, and large amounts o
> tensely pitted outwash accumulated along it. The eastern
> of this outwash is very irregular, but devoid of a till mo

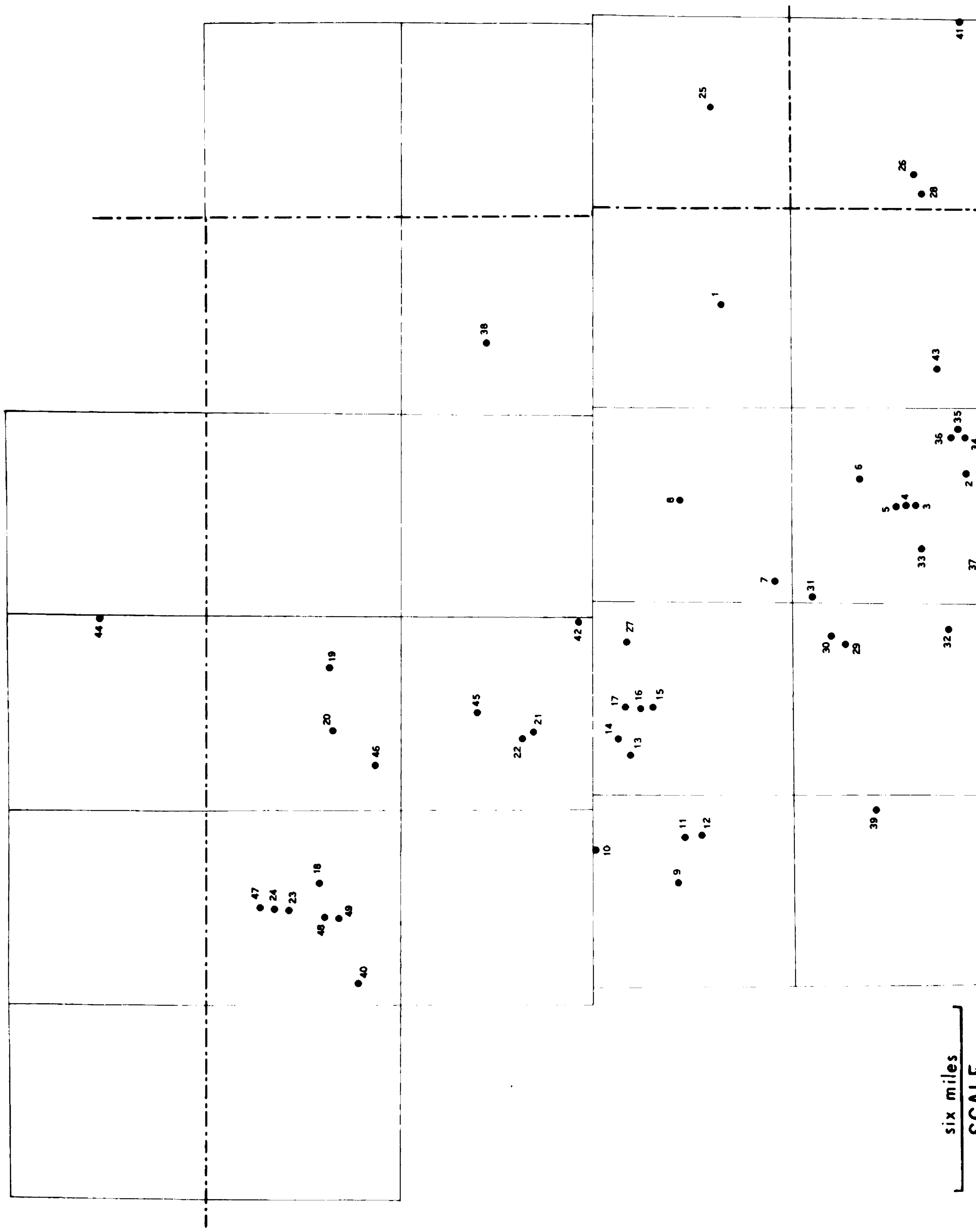
Leighton and Brophy (1966) discussed several other lines o
which may be helpful in delineating former ice margins. These
eskers and crevasse fillings, channels cut into proglacial ridg
meltwaters draining from an ice front, outwash deposits, kame d
ice-contact gravel kames, ice-margin knob-and-kettle topography
to the supposed ice front, and ridges of till with long axes pa
to the ice margin.

Thwaites (1943, p. 111) observed that "several eskers occur in rough overridden moraine in eastern Shawano County. This fact demon strates that the rough topography is inherited from an older Cary moraine." Some of the eskers in the study area were originally thou to have the same relationship, but the results of this investigatio proved otherwise, as discussed in a later section.

The discussion by Thwaites (1926, p. 316) on pitted outwash pr helpful. He states that "the presence of pitted outwash serves to distinguish recessional from terminal moraines;" i.e., the presence thick masses of pitted outwash well distal from the related moraine establishes that moraine as recessional. Mathiesen (1940) observed same relationship along the west side of the Chippewa moraine in Ba County which adjoins the northwest corner of the thesis area.

The moraine zone, hereafter termed the moraine complex, was su divided on the basis of topographic evidence and the associated gla sediments. Type localities were selected to illustrate each of the different glacial depositional environments. The findings associat with these type localities served as a basis for the general mappin the sediments in the remainder of the study area. Field mapping of sediments depended primarily on an examination of drift exposures i roadcuts, borrow pits, and in other excavations, although hand auge and shovel sampling were employed where no exposures were available where an additional check on available exposures was deemed desirab

Forty-nine samples of various sediment types--from scattered s throughout the study area (Figure 3)--were analyzed for grain size the Bouyoucos hydrometer method (Day, 1965), according to guideline


six miles

on this method prepared by the Department of Soil Science, Michigan State University. All samples came from beneath the solum.

The five sediment types and the sampling locations are as follows: (1) till from Woodfordian-age ground moraine at various sites on both the distal and proximal sides of the moraine complex; (2) till from late Altonian-age ground moraine east of the Chippewa River, at sites one to five miles beyond the maximum position of the middle Woodfordian advance; (3) till, interpreted as collapsed superglacial till, from knobs and ridges within areas of hummocky stagnation moraine; (4) glacio-lacustrine silts from several ice-walled lake plains within the moraine complex; and (5) loess from sites beyond both the distal and proximal margins of the moraine complex.

Well-log data, from several sources, were used as a means of ascertaining the drift stratigraphy in the morainal zone. These data were also used to construct a map showing thickness of the glacial sediments and to obtain information on the bedrock topography and pre-existing drainage as they relate to the surface forms.

Several topographic profiles were constructed across the moraine zone and used in the subdivision of the moraine by (1) comparing the several transverse profiles as a means of identifying intermorainal depressions and other distinctive topographic trends and (2) comparing distinctive points on each profile with well-log data and field observations of the drift for the same locality, to determine relationships between surface forms and materials. For example, many field observations were made of the larger longitudinal depressions within the moraine complex to resolve whether they are underlain by sediments of glacial, glaciofluvial, or lacustrine origin.

A limited effort was made to differentiate drifts on the basi variations in the depth of carbonate leaching; however, no conclus results were expected, for according to information by personnel o Eau Claire regional Soil Conservation Service office the unweathere drift of the Chippewa lobe is only weakly calcereous. This reflec predominantly noncalcareous Precambrian bedrock source of the drif

Is Part of the Moraine of Interlobate Origin?

Southwest of Flambeau Ridge, in the northwestern part of Chip County, the Woodfordian moraine has a greater width and height tha elsewhere in the study area (Figure 6). This prominent and massiv section of the moraine may have formed between two minor sublobes downglacier from Flambeau Ridge.

This investigation sought to verify the aforementioned interl hypothesis by examining the sediments and morphology of this hilly and to determine if in fact it does have the characteristics assoc with interlobate moraines. The definitions by Flint (1971) and Em and King (1968) for interlobate moraines, in conjunction with Blac (1969a; 1970) descriptions of the Kettle Interlobate Moraine of Wi sin, served as a basis for recognizing the landform. Particular a tion was given to characteristics which may indicate the direction movement on either side of the suspected interlobate tract. Examp include the relative trends of linear elements, the orientation of ice-contact slopes, the direction of slope of outwash surfaces and drainage channels, and the relationships of the sediments and stru to one another on either side of the postulated interlobate tract.

Did Marginal Stagnation Occur?

The Wisconsinan moraines of the Chippewa valley are shown on th Wisconsin Glacial Deposits map (Figure 1) by Thwaites (1956) as end moraines. However, in recent years some of the areas formerly mappe as end moraines have been reinterpreted as stagnation moraines, depo ited from stagnant ice. The reports by Gravenor and Kupsch (1959), Clayton and Freers (1967), and Black (1969a) are representative of s reappraisals.

Flint (1971, p. 200) states that "the essential thing about end moraine is its close relationship to the margin of the glacier." He defines an end moraine as "a ridgelike accumulation of drift (with t commonly dominating) built along any part of the margin of an active glacier." Gravenor and Kupsch (1959, p. 51) point out that true end moraines may contain several more or less subparallel ridges "which emphasized by intervening aligned depressions. These trends define successive positions of the ice front. Generally they are gently cu in outline and regularly spaced."

By contrast glacial topography designated as "stagnation morain (Clayton, 1962; Winters, 1963) results when a wide marginal zone of glacier becomes stagnant or inactive.* Debris upon or in the stagna ice may gradually accumulate in crevasses and hollows that develop u the glacier, or drift beneath the stagnant ice mass may be squeezed into subglacial openings. In short, as the stagnant ice disintegrat into separate blocks, there is produced a jumble of knolls separated by irregular depressions, collectively called hummocky stagnation

*Gravenor and Kupsch (1959) prefer the term "disintegration moraine."

moraine (Winters, 1961, p. 39). According to Gravenor and Kupsch (1959, p. 51) moraines of this type usually lack the sharp borders of normal end moraines. Furthermore, the ridges and depressions of stagnation moraine show no preferred orientation, unless there existe a well-organized fracture pattern in the glacier prior to stagnation (ibid., p. 50).

Other features which commonly are found in areas of stagnation moraine include moraine plateaus, closed and linear ridges, and ice-walled meltwater channels (Gravenor and Kupsch, 1959). Clapperton (1968), Derbyshire (1958), and Wright (1973) have also discussed melt water channels associated with thin or disintegrating glacier margins Winters (1961) and Clayton and Cherry (1967) describe another type of plateau-like feature common in hummocky stagnation moraine, termed by them "perched lake plains" or "ice-walled lake plains."

Stalker (1960) assumes that many stagnant-ice features have an ice-press origin. He believed that many of the ridges within hummock stagnation moraine are the result of water-saturated till being squee from beneath ice blocks into openings or crevasses in the disintegrat ice. He recognized a variety of features produced by this process. Some of Stalker's ideas on the ice-pressing process appear to be infl enced by the work of Hoppe (1952) and other European studies.

Gravenor and Kupsch (1959) point out that when disintegration features are composed of ablation moraine or collapsed superglacial debris, the material may be somewhat loose, whereas if the ridges are of ice-press origin they should be composed of rather compact till similar to ground moraine. The modes of origin of superglacial debri have been discussed, among others by Embleton and King (1968), Flint (1971), Sharp (1949), and Souchez (1967).

A geologic report concerning several areas of distinctive glacial features in Wisconsin was prepared by Black (1974). It recommended a number of localities for inclusion in the Ice Age Reserve, which is being established by the State of Wisconsin jointly with the National Park Service. Black recommended that parts of the area covered by the Bloomer quadrangle, southwest of Flambeau Ridge and within the study area, be included in the Reserve, for in his view the Woodfordian moraine near Bloomer exhibits considerable evidence of deposition from stagnant ice. Black's investigations in this area were relatively limited and did not include a detailed mapping of the glacial features.

In recognition of Black's interpretation, I sought to establish the relative importance of marginal stagnation within an area affected by the Chippewa glacial lobe. In addition, an attempt was made to ascertain whether the assumed stagnation features of the thesis area resulted from the subsidence of superglacial debris and/or from the squeezing and pressing of subglacial drift by the weight of overlying ice.

Particular emphasis was given to examining the plateau-like features surrounded by hummocky moraine which are shown on topographic maps covering the Chippewa County area. Prior to this investigation the author was informed that these plateaus typically have deep silty soils, possibly of lacustrine origin.*

In summary, ascertaining the relative importance of marginal stagnation in the formation of the glacial phenomena of the thesis area was

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*Personal communication from personnel of the Eau Claire regional Soil Conservation Service office.

based on (1) determining the characteristics of the landforms and sediments that exist in the area, including their relationships, and (2) comparison of these with the characteristics given for both active- and stagnant-ice moraines. Two hypotheses concerning the formation of stagnant-ice features were given special attention: ice-pressing and disintegration of drift-mantled ice.

What Is the Assemblage of Landforms?

In general, the investigation of this problem consisted of determining whether or not the individual landform types occupy a characteristic location relative to one another. To determine this, the individual landforms were first identified, combining information obtained in the field with the study of maps and aerial photographs. The landforms were then plotted on topographic maps to determine their relative position. In this regard aerial photographs were especially useful.

Special consideration was given to determining the following: (1) the total number of distinctive landform assemblages that exist in the area; (2) the typical or relative location of each assemblage-type; (3) the location of the various landform types relative to the moraine margins; (4) whether or not the landforms are arranged in controlled or uncontrolled assemblages (as described by Gravenor and Kupsch, 1959); and (5) the conditions or factors that are responsible for the location and characteristics of the various landform assemblages. Also investigated was the question of whether or not the morphology and/or lithology of the individual landform types differ significantly, from one landform assemblage to another.

Minor Problems

Resolving the minor problems, outlined on page five, involved a variety of investigative techniques and lines of evidence, summarized by the following statements.*

1. The determination as to whether or not the bedrock topography significantly influenced the ice movement rested for the most part on an analysis of topographic evidence, such as the trends of drumlinoid ridges, ice-channel fillings, and the moraine margin.

2. The explanation of why the moraine is better developed to the west than due east of the Chippewa River was believed to involve a difference in the topography which the two sections of the ice margin had to override. Thys hypothesis was weighed along with others.

3. An attempt was made to reconstruct the late Quaternary sequence of events in this area, based on a mapping of ice-front positions, stratigraphic evidence, and landform phenomena such as morainic, periglacial, and proglacial features.

4. Establishing the extent of loess and its relationships with the glacial sediments was based mainly on the mapping of surface sediments and lateral gradations in loess grain size and thickness.

5. Pebble-count data were obtained from previous studies of this area as a means of determining the sources of the glacial sediments.

6. Periglacial features have been described in western Wisconsin by several authors, notably Black (1964; 1965 and 1969). Phenomena of this origin were investigated in instances where they appeared to be pertinent to understanding the problems outlined above.

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*Because each of the minor problems--listed on page five--is discussed in succeeding chapters under a number of headings, they are not, in general, separately identified in the Table of Contents.

Conditions Impeding Interpretations in the Moraine Complex

The following conditions seriously impeded investigations in part or all of the study area. Because of this, the thoroughness and reliability of the mapping and interpretations may vary from one district to another in the thesis area.

1. Forest cover. Somewhat more than three-fourths of the study area is in woodland, with several large tracts almost entirely forested. Since most of the field work for this study was conducted during summer months, heavy foliage severely restricted landform observations. Those districts in which the landforms have been interpreted in great detail on Figure 45 typically coincide with nonforested areas. Undoubtedly, many landforms were not identified and mapped in areas of heavy forest cover.

2. Large sections are roadless. The lack of roads in several parts of the area severely limited accessibility in these places. The interlobate tract, in particular, is crossed by only a few roads and, furthermore, is largely forested. An attempt was made to compensate for lack of field contact by an intensive study of air photos and topographic maps of the interlobate tract and in other areas posing the same problems.

3. Lack of small-contour-interval topographic maps. For the southern two-thirds of the study area the best available contour maps have a contour interval of twenty feet and a scale of 1:62,500. While small-contour-interval, large-scale topographic maps are currently being prepared for the northern part of the area, the best maps available at the time of this investigation have a scale of 1:250,000 and a fifty-foot contour interval. Since many of the stagnant-ice landforms of this area, discussed in later chapters, are relatively small or low-relief features and interpretable only when they can be examined in

detail, the available contour maps were frequently of little value when interpreting such features.

4. Obscuring effects of silt cover. At many places within the moraine zone, as well as north and south of it, the glacial sediments are covered by a loess blanket of variable thickness. Glaciolacustrine silts are also common and could not always be differentiated from wind-deposited silt. The presence of this irregular silt deposit, averaging from one to three and in places exceeding five feet in thickness, rendered determining the character of the underlying glacial sediments very difficult at times. In many places the material beneath the silt is not exposed by roadcuts or excavations and could not be ascertained without hand-augering or shoveling through the silt blanket. Mapping of the glacial sediments, therefore, was not always based on an optimum density of observations, for the amount of time and effort requisite to make the desirable number of auger borings would have been prohibitive.

The loess cover also complicates interpretation of the glacial landforms by the smoothing effect it has on the original glacial topography. In several places recently constructed roads show that the topography underlying the loess blanket has more relief and is rougher than that developed on the loess. This is particularly true along the margins of the moraine zone, where the topographic expression of the glacial features is not distinct.

Justification for this Study

The glacial landforms and deposits of the Chippewa lobe are believed to be the least studied of any glacial lobe in north-central United States, and in Wright's view (1964, p. 630) the uncertainty concerning the glacial events of this area has seriously impeded the attempts to correlate the glacial history of the Lake Michigan basin with that of the area located west of the Chippewa lobe. In view of the foregoing, the immediate objective of this study is to examine comprehensively the glacial features of this little known area. Furthermore, this information may provide a partial basis for the time-stratigraphic correlation between the glacial lobes located east and west of the Chippewa lobe.

CHAPTER II

GENERAL DESCRIPTION OF THE STUDY AREA AND NATURE OF ASSOCIATED GLACIAL DRIFT

Introduction

The extent of the Chippewa moraine complex within the study area is shown on Figure 1. As defined in this study, it includes the area designated as prominent stagnation moraine on Figure 6 and as end moraine on the map by Thwaites (1956) (see Figure 1).

The results of this investigation show that much of the moraine zone consists of several moraines, giving it a composite character. Therefore, the term "moraine complex" seems preferable. Southwest of Flambeau Ridge, however, the moraine complex is not divisible into separate moraines, but instead it consists of an impressive hilly tract that extends continuously from Flambeau Ridge to the distal edge of the moraine complex. In this area the moraine reaches its maximum height and width.

Ages of Glacial Sediments

Black and Rubin (1967-68) point out that no drift known to be older than Wisconsinan exists in the state of Wisconsin and that no radiocarbon dates have been determined for material obtained from beneath or within the drift of the Chippewa moraine. Nevertheless, it has been correlated with the Woodfordian moraines of Illinois by Frye, Willman, and Black (1965, p. 56). A quotation from Black and Rubin (1967-1968, p. 110) summarizes the situation:

> without a single radiocarbon date related to the advances of the Woodfordian ice in Wisconsin, and few to record its destruction,

> we have been dependent on morphology of forms and direction indicators to separate pulsations. These are applied with difficulty in many places but generally seem better than lithology or texture of the material involved in any one sublobe.

The term "Woodfordian" was introduced by Frye and Willman (1960) in their reclassification of the Wisconsinan glacial stage. Their classification, based on the stratigraphy of the sediments associated with the Lake Michigan lobe, combines the formerly recognized substages (Tazewell, Cary, and Mankato) into a single substage, designated as the Woodfordian, assumed as having lasted from about 22,000 until about 12,500 radiocarbon years B.P.

Wright (1964) disapproved of Frye and Willman's abandonment of established nomenclature, contending that information was inadequate to permit reliable interregional correlations of ice-margin fluctuations between Illinois and distant areas. For example, two major conditions that in Wright's view (ibid., p. 630) impeded correlations between Illinois and Minnesota are that

> the moraines around the Chippewa lobe of northern Wisconsin north of the Driftless Area have not been mapped since the days of Leverett (1929), nor dated by radiocarbon analysis.

In describing the moraines of Wisconsin, Frye, Willman, and Black (1965) indicate that the Chippewa moraine was formed contemporaneously with the St. Croix moraine of the Superior lobe and assign both to the middle Woodfordian (Cary) substage of Frye and Willman's classification (1960). However, based on the radiocarbon dating of organic matter from the ground moraine of the Superior lobe, Wright, et al. (1973) indicate a minimum radiocarbon age of about 20,000 years B.P. for the St. Croix moraine. If this interpretation is correct, the Chippewa moraine (1) can not be a correlative of the St. Croix moraine, or, (2) if it is a correlative, its age must be Tazewell (early Woodfordian), as proposed by Wright, et al. (ibid.) for the St. Croix moraine. The results of this thesis neither confirm nor deny either of these possibilities. Because the age revisions proposed by Wright were not based on field data from Wisconsin, and were not specifically proposed for the state, the middle Woodfordian-age (15,000 years) proposed for the formation of the Chippewa moraine by Frye, Willman, and Black (1965) will be accepted.

The age of the older drift(s) lying beyond the Woodfordian St. Croix and Chippewa moraines of Wisconsin has been the object of considerable debate and uncertainty. Numerous studies, but especially those of Weidman (1907), Leverett (1929), and Mathiesen (1940), have subdivided the older glacial deposits of this area into two or more drift sheets. But one of the latest reviews of the glacial deposits of Wisconsin (Frye, Willman, and Black, 1965, p. 48) asserts that the oldest drift in central Wisconsin (deposited by the Chippewa lobe) is "clearly Wisconsinan according to depth of leaching and weathering of constituents, as determined by Hole (1943)." They further state that ice of the Superior lobe advanced southeastward across west-central Wisconsin about 32,000 to 29,000 years ago, according to radiocarbon dating of spruce logs overrun or incorporated in the drift, and overrode a residual soil rich in illite and chert (evidence strongly suggestive of the lack of previous glaciation for that area). They designate the age as late Altonian but add an element of uncertainty for the age of the older drift by stating that "a pre-Wisconsinan age for some of it can neither be confirmed nor denied" (ibid., p. 54).

In a report on radiocarbon dates in Wisconsin Black and Rubin (1967-1968) give two dates of more than 45,000 years B.P. for organic matter from within or beneath the drift: one for Chippewa lobe drift in central Wisconsin, another for Superior-lobe drift in west-central Wisconsin. An additional date of 38,000 years B.P. is given for wood fragments from a deep well in northwest Wisconsin. However, Black an Rubin (op. cit.) state that the evidence is not adequate to determine whether these organic materials may have been preserved in pond depos for a considerable time prior to burial or if they were fresh at the time of burial. Recently, Bleuer (1970) recognized Illinoian and old drift in Wisconsin west of Beloit and Janesville. His interpretation may serve as at least a partial basis to question the age of the olde glacial deposits elsewhere in the state. In view of the differing in pretations, cited above, and the absence of evidence that firmly esta lishes the age of the older deposits of central and western Wisconsin the writer tends to assign the drift located beyond the middle Woodfordian moraine to late Altonian, but with the reservation that this should be considered tentative.

Bedrock Geology in the Study Area

In approximately the northeastern one-third of the area, as show on Figure 4, the bedrock on which the glacial drift rests consists of variety of Precambrian igneous and metamorphic rocks. The remainder of the area is underlain by Upper Cambrian sandstones, along with som shales, interrupted by inliers of Precambrian basement rock. Flambea Ridge, shown on Figure 4, is situated on the boundary separating the Precambrian crystallines from the Cambrian sandstones. Structurally, the entire area lies on the western flank of the southward-pitching

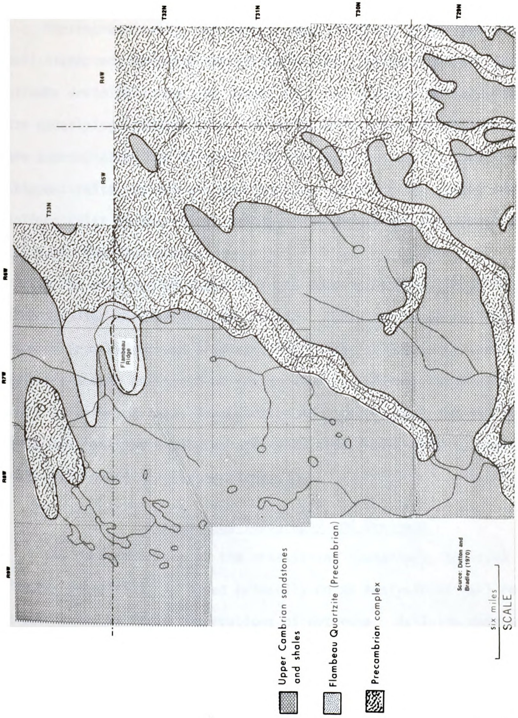


Figure 4. Bedrock geology.

Wisconsin Arch; consequently the Precambrian surface and overlying sandstones are gently inclined southwestward. The regional drainage reflects the bedrock surface and also trends southwestward.

Physiographically, the edge of the sandstone area represents a well-dissected, westward-retreating cuesta. Several of the larger streams draining across the cuesta have cut through the sandstone into the underlying basement rock. Conspicuous inliers of Precambrian rocks are exposed along the Chippewa and Yellow Rivers as far downstream as Chippewa Falls, in the southwestern part of the area. Other smaller inliers exist at several locations, including areas at Drywood and along the upper Eau Claire River.

A few outliers of Cambrian rocks subcrop beneath the drift up to several miles northeast of the cuesta. These are shown on Figure 4, as adapted from Dutton and Bradley (1970). The existence of sandstone outliers within and north of the moraine is inferred largely on the basis of well-log data, for outcrops of both the Cambrian and Precambrian are rare and exposures are restricted essentially to small areas of erosion along the larger rivers.

Preexisting Topography and Drainage

The reconstruction of the preexisting topography, believed to be largely preglacial, is based primarily on an analysis of well records, supplemented by field observations of outcrops. Well-log data were obtained from several sources but primarily from the Private Water Supply Section of the Wisconsin Department of Natural Resources.

Flambeau Ridge, an exhumed quartzitic monadnock (Martin, 1932, p. 374), was by far the most prominent preglacial landform of the area.

Its local relief presently is about 500 feet, but well records indicate drift in excess of 120 feet thick in the lowlands immediately to its east and west, suggesting a preglacial relief of at least 600 feet. A conspicuous notch is located near the east end of Flambeau Ridge, and well records from within the floor of this col indicate drift in excess of 100 feet thick.

In the remainder of the area underlain by Precambrian rocks, the preglacial topography was, in the view of Martin (1932), that of an exhumed Precambrian peneplain. Well records indicate a low-relief, relatively featureless topography on the drift-covered Precambrian surface. This appraisal is corroborated by the low relief associated with the Precambrian surface where it is exposed for several miles along the Chippewa and Yellow Rivers.

In preglacial time there were probably few hills, other than Flambeau Ridge, on the Precambrian surface with more than 100 feet of local relief, and the average was probably much less. With the exception of Flambeau Ridge, the preexisting bedrock topography on the Precambrian rocks appears to have had little influence on glacial movements, nor is it expressed in the present-day topography. The sparse well data available for the Precambrian sector are not adequate for a reconstruction of that area's preglacial drainage pattern.

In general, the bedrock in the southwestern two-thirds of the study area consists of the extensively eroded eastern margin of gently southwestward-dipping Cambrian sandstones and shales. In preglacial time the relief on this dissected cuesta, now largely obscured by the Woodfordian moraine constructed upon it, was considerably greater than on the exhumed Precambrian erosional surface to its northeast. Sandstone

outcrops are extremely rare within the moraine, except along its dis margin, but an analysis of well records reveals numerous buried sand stone uplands and large drift-filled valleys. Several of the valley were completely filled with drift and were not reoccupied with deglaciation of the area. Their presence is not always apparent fro the surface topography, and most were detected on the basis of well records. Most of the larger buried sandstone uplands, however, are expressed in the present topography of the moraine complex, even tho in some cases mantled with 100 feet or more of drift. Based on inte sive field study of this area it seems safe to state that the larger relatively smooth upland tracts within the moraine complex are typic underlain by buried sandstone uplands.

Such situations exist at about fifteen to twenty places along t sides of the Chippewa River and may be identified on Figure 8. They appear to be mainly confined to Chippewa and Rusk Counties, since we data from within the moraine indicate much less relief on the Cambri surface from Chippewa County eastward. The relief on these partiall or entirely buried sandstone uplands ranges between 150 to 250 feet in a few instances exceeds 300 feet. In overall size, relief, and morphology they resemble the sandstone uplands just beyond the dista margin of the moraine from Lake Wissota northwestward (Figure 8).

The most remarkable example of a large drift-filled valley exte from beneath the Mud Creek lowland (located just west of Flambeau Ri on Figures 5 and 6), eight miles southward to Cornell Lake, from whe it continues southeastward beneath Howe Lake and then southward bene Lake Wissota. Another major buried valley extends southwest from Cornell Lake, then south to Lake Wissota (Figure 5). The lack of we

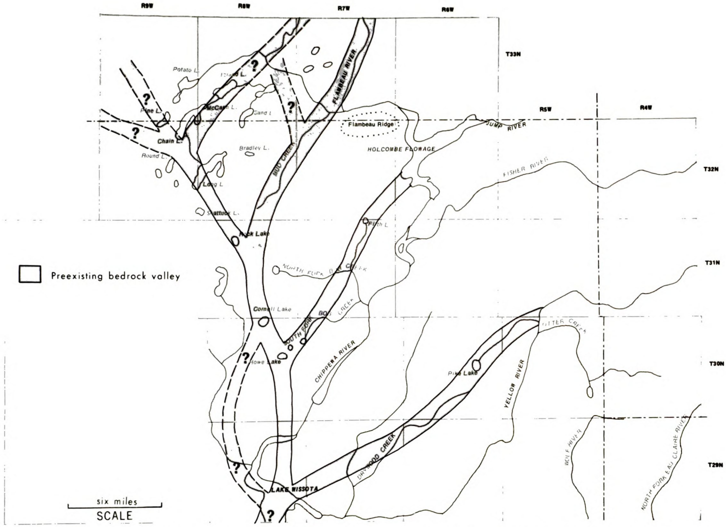


Figure 5. Preexisting bedrock valleys. The bedrock valleys shown are believed to be preglacial in age but some may be interglacial.

reaching bedrock along these routes prevents determining the exact relationships between these valleys.

Another deep drift-filled valley extends northeast-southwest beneath the moraine in the Island-Long Lake District. Within it are situated the basins of several lakes, and a number of wells near thes basins penetrate over 100 feet in depth without reaching bedrock. Ho ever, the presence of sandstone at about forty feet depth in several wells at the northeast end of Long Lake implies that the preglacial valley occupied by this lake basin does not continue directly northeastward.

It is proposed here that the above-described drift-filled valley most probably represent the preglacial courses of the Chippewa and Flambeau Rivers, though the possibility that they are in part of an interglacial age is also recognized. The evidence to substantiate th interpretation consists of the following:

1. Several deep wells located along a north-south line between Mud Creek lowland and Lake Wissota penetrate up to 185 feet in depth with encountering bedrock. The bottoms of these wells have altitudes well below that of the nearby Chippewa River, and they indicate that the floor of this buried valley is at least fifty feet lower than the bed rock floor of the modern Chippewa River between Jim Falls and Holcomb
2. The Chippewa River, south of its junction with the Flambeau River makes a large eastward bend around the eastern end of Flambeau Ridge then continues southward. This course resembles the glacial diversio of the Wisconsin River around the east end of the Baraboo Range.
3. At the south end of Long Lake, two wells, about 185 feet deep and not reaching bedrock, indicate that the bedrock surface in this local

also has an altitude below that of the bedrock floor of the Chippewa River near Holcombe.

The foregoing conditions permit the interpretation that in preglacial time the courses of both the Chippewa and Flambeau Rivers lay west of Flambeau Ridge (see Figure 5). The Chippewa River may have drained south through the drift-filled valley presently occupied by Island, McCann, Chain, and Long Lakes and then have extended southeast to join with the Flambeau River. This latter stream drained south from Mud Creek lowland through the Cornell Lake area to the locale of present Lake Wissota.

One deep well near the northwest corner of the study area suggests that another large preglacial valley may lead southeast from the present Chetek chain-of-lakes area (located near the far northwestern corner of Figure 5) to join the preglacial Chippewa near the present Long Lake. The absence of any bedrock exposures along this route lends partial support to this interpretation, but detailed subsurface data are not available. If such a buried valley does exist, it may possibly be the preglacial course of the Red Cedar River, the largest right-bank tributary of the Chippewa, and situated west of the study area. The small size of the valley presently occupied by the Red Cedar River, immediately west and southwest of Chetek, suggests that the latter was probably diverted into that position during the last glacial episode that affected the area. Before the glacial period this stream may have drained southeast to join the Chippewa in northwestern Chippewa County.

There appear to be at least two additional partially drift-filled valleys beneath the Woodfordian moraine. One is located about two miles west of and parallels the modern Chippewa River between the Holcombe

Flowage and Howe Lake (see Figures 2 and 5). This bedrock valley is marked by a continuous lowland that is not occupied by a through-flowi stream. None of the deeper wells within it, ranging between seventy t eighty feet in depth, reach bedrock. Additional evidence that the low land is associated with a preglacial valley is revealed upstream from Cornell, where a seventy-foot deep well on an island in the river does not reach bedrock. The bedrock floor of the Chippewa River, one mile downstream, has an altitude about fifty feet higher than that at the bottom of the well mentioned above. Apparently the present-day Chippe River has not reached the depth of valley-cutting attained in earlier times, nor is it following the course of the preexisting valleys in th Cornell area.

If a major preexisting valley does parallel the west bank of the Chippewa River between Holcombe and Jim Falls, perhaps this was the course of an ancestral Jump River, which drained from the point where presently enters the Holcombe Flowage to the place where it would have joined the preglacial Chippewa River a few miles northwest of Jim Fall (see Figures 2 and 5).

The remaining bedrock valley to be discussed may have been associated with the preglacial Yellow River. Well records suggest that a continuous bedrock valley extends northeast from the place where the present Yellow River enters Lake Wissota. Southwestward-flowing Drywc Creek now occupies the southern portion of this drift-filled valley, whereas the outlet of Pike Lake occupies the northern section and flov in the opposite direction. Pike Lake (Figures 2 and 5) separates thes two streams and is one of the two natural lakes (excluding small ponds east of the Chippewa River in Chippewa County.

Figure 2 indicates that the present Yellow River has a west-southwest direction of flow in western Taylor County and maintains this direction until a point about three and one-half miles inside Chippewa County, where it makes an abrupt change in direction to the south. Another unusual characteristic of the present Yellow River is its deep, narrow bedrock valley that extends from Highway D (see Figure 2) toward Cadott. Comparison with other valleys in the sandstone area reveals that this section of the Yellow River valley is almost without tributaries and that it is relatively small for this large stream.

The presence of a continuous, buried valley extending northeastward from Lake Wissota, the abrupt change in direction exhibited by the Yellow River at the point where that buried valley intercepts it, and the abnormally small size of the bedrock valley occupied by the Yellow River upstream from Cadott support the interpretation that the lower Yellow River was glacially diverted into a small preexisting valley and that the ice-block basin occupied by Pike Lake probably is situated in the large partially drift-filled valley of the preglacial Yellow River (see Figure 5).

Source and Movement of the Chippewa Lobe

Chamberlain (1883b) attributes the Chippewa lobe to ice channeled by and draining southwest from the Superior basin, between the Keweenaw and Bayfield Peninsulas. Martin (1911, p. 428) indicates that the advancing ice was profoundly affected by the preglacial highlands and lowlands, with the ice being thickest and moving fastest in the depressions. He also states that the glacial lobes from the Superior basin were less extensive than those of other ice streams. "The Michigan

lobe of the Labradoran ice extended farther south than any of the others, and the Green Bay lobe of the Labradoran ice, also having a deep axis of flow, extended nearly as far south" (Martin, 1911, p. 428). In contrast, the Keweenaw and Chippewa lobes of the Labradoran ice, obliged to advance over the highland region of upper Michigan and northern Wisconsin, did not extend as far south. Farther west the ice of the Superior lobe, draining from the west end of the Superior basin near Duluth, impinged on the northwest side of the Chippewa lobe and moved southwestward, spreading out over a much broader area than did the less active and more confined Chippewa ice.

Horberg and Anderson's (1956, p. 102a) bedrock-topography map of central North America shows that the south rim of the Superior basin, where crossed by the Chippewa lobe, has an altitude 300 to 500 feet higher than that traversed by the Superior lobe and 500 to 800 feet higher than the route traversed by the Green Bay lobe. Assuming equivalent ice-surface altitudes on all three lobes where they drained across the south rim of the Superior basin, the aforementioned values indicate that substantially thinner ice existed in the Chippewa lobe than in the Superior and Green Bay lobes. Furthermore, because the surface of the Labrador ice in the Superior basin increased in altitude to the northeast, the difference in thickness between the Green Bay and Chippewa lobes must have been greater than the figures given above.

This difference in thickness probably resulted in a slower rate of ice advance in the Chippewa lobe, and as a consequence it attained less extent than the thicker lobes to its east and west. Furthermore, the general scarcity of recessional moraines in the area glaciated by the

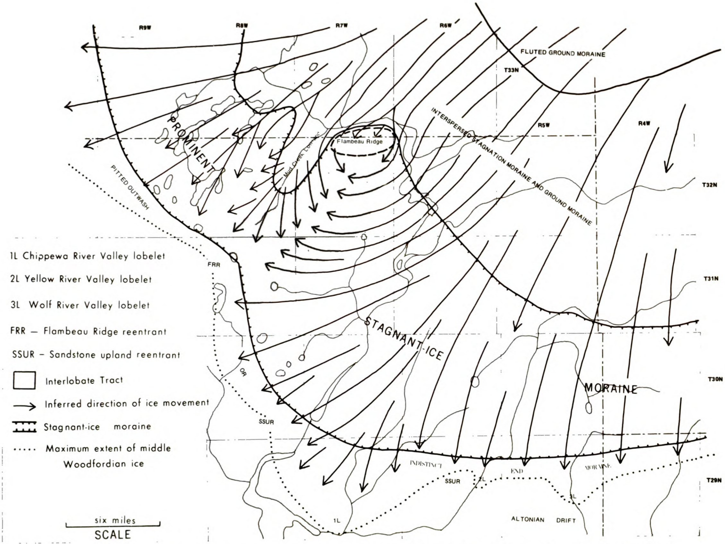
Chippewa lobe (see Figure 1) implies that during deglaciation it may not have experienced as many pulsations as did the thicker lobes of the Great Lakes area.

The pattern of ice movement in the study area, shown on Figure 6, is based on the alignments of a variety of linear features commonly oriented parallel to the direction of ice flow. These include fluted ground moraine, striae, linear disintegration ridges, and eskers. Numerous examples of the latter two classes of features are shown on Figure 45 and may be observed collectively to form a radial pattern extending outward from the central part of the Chippewa lobe. No well-preserved striae were found, but some of indistinct appearance were observed on Flambeau Ridge. Their orientations are reflected in the ice-flow lines shown on Figure 6.

Characteristics of the Drift

Although this study is primarily concerned with the morphology of the Chippewa moraine, several aspects of the drift itself will be discussed, in particular grain size and thickness. Figure 7 summarizes the results of a grain size analysis (also discussed in Chapter I) of five different sediment classes. Each class of sediment is represented by a different symbol and tends to occupy a different part of the diagram. Some of the classes, however, show grain-size overlap.

As shown on Figure 7 the till associated with the late Altonian ground moraine has the finest grain size (of the three classes of till), except where it is relatively thin and underlain by sandstone, where it often is more sandy. Slightly more coarse is the sandy clay loam/ sandy loam till from the Woodfordian ground moraine. Coarser still--



1L Chippewa River Valley lobelet

2L Yellow River Valley lobelet

3L Wolf River Valley lobelet

FRR — Flambeau Ridge reentrant

SSUR – Sandstone upland reentrant

Interlobate Tract

Inferred direction of ice movement

Stagnant-ice moraine

Maximum extent of middle Woodfordian ice

six miles

SCALE

Figure 6. Pattern of ice movement and boundaries of Chippewa moraine complex.

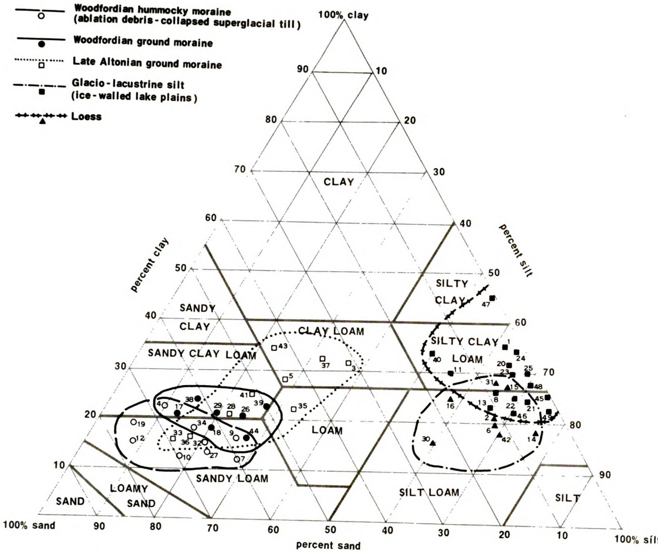


Figure 7. Grain sizes of selected sediment types.

but not markedly so--is the sandy loam till associated with hummocky stagnation moraine.

Grain size variation is also evident within the silty materials of the area. Figure 7 shows that the loess samples collected from north and south of the moraine complex occasionally contain slightly more sand and less clay than the glaciolacustrine silts of ice-walled lake plains. Grain size within the loess tends to decrease with increasing distance from the outwash plains, which apparently served as the source areas for the loess. Similarly, the grain size of the glaciolacustrine sediments decreases with increasing distance inward from the margins of the ice-walled lake plains.

Because the various sediment types often present distinct color differences, this characteristic was sometimes useful in differentiating them in the field. The late Altonian till is typically brown-dark brown (7.5 YR 4/4).* All of the samples of middle Woodfordian till, by contrast, are commonly reddish brown (5 YR 4/4), but may vary between dark reddish brown (5 YR 3/4) and light reddish brown (5 YR 6/3); when dry it has a pale pinkish brown appearance. In places where the finer sediments are less abundant, such as in ablation deposits, the color of this till becomes brownish.

Under good drainage conditions the typical color of the loess is brown-dark brown (10 YR 4/3), although it may range toward lighter phases of yellowish brown (10 YR 5/6). The glaciolacustrine sediments are very similar in color to loess and may be in part derived from it. As a result this criterion is typically not suitable for distinguishing the

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*All of the Munsell soil colors given are for moist conditions.

two in the field. In some places the glaciolacustrine silts display a slight pinkish cast and may even have the reddish brown color characteristic of the Woodfordian-age till.

The appearance of the Woodfordian drift indicates only minor postglacial alteration by weathering. Only a few deep exposures of the Altonian till were found but these often exhibited a change in color from brownish to reddish brown with increasing depth. These color changes may indicate a greater degree of weathering than in the younger drift.

Dilute hydrochloric acid was used to test for the presence of carbonates and the depth of leaching. Effervescence was not apparent in any of the till exposures examined. This was expected for the drift of the Chippewa valley was derived largely from noncalcareous rocks. However, Hole (1943, p. 506) found several instances of locally existing carbonates in what he called the "border drift" of central Wisconsin, deposited by a pre-Woodfordian phase of the Chippewa lobe, where leaching extended only to depths of three to eleven feet. He observed considerable variation in depth to calcareous till within distances of a few tens of yards and suggested that it may be due to variations in "porosity of bedrock, texture of till, thickness of loess and colluvium, topography and drainage" (ibid., p. 510) and concludes that depth of leaching is not a reliable criterion for determining the relative age of the drift of central Wisconsin.

Even though calcium carbonate is absent or exists only in small amounts in the till of the area, marl deposits were observed in two localities in Chippewa County within the moraine complex: adjacent to Highway 64, one and one-half miles west of Cornell Lake; and about two miles southeast of Jim Falls. The lakes in which these marls were originally deposited have been filled with vegetation and both sites are presently swamps. Road excavations have exposed the marl beneath the peaty overburden.

Calcium carbonate was also found in some of the glaciolacustrine silts at depths of ten to twenty feet. Ranney (1964), in a study of several soils in north-central Wisconsin, likewise found carbonate-rich zones in glaciolacustrine sediments. He pointed out that their origin is difficult to explain, if the low carbonate content of the Woodfordian drift of the area is taken into consideration. He (ibid., p. 86) postulates that

> the waters that flowed into the glacial lakes were enriched with calcium bicarbonate through contact with calcium bearing minerals, calcium saturated clays or perhaps small amounts of carbonates in the till. Once the water was in the lakes, the carbon dioxide was presumably depleted by evaporation, warming, or probably most important, by photosynthesis of algae in the water. Depletion of carbon dioxide would cause calcium carbonate to precipitate.

Ranney suggests another possible source for the carbonate: when originally deposited, the loess blanket may have contained a calcereous fraction derived from carbonate rocks farther to the west. This soluble material was subsequently leached, some of it being precipitated in glacial as well as postglacial lakes.

A third possibility is that the drift originally contained a small amount of calcium carbonate derived from calcareous rocks from other areas, such as the southeast part of the Superior basin. The presence of occasional Ordovician to Devonian nautiloid and colony coral fossils in the drift of the Chippewa valley supports this hypothesis. Several such fossils found within the Woodfordian outwash of the Chippewa valley at Eau Claire were discovered by or reported to the author. Furthermore, a single colony coral was recovered from the middle Woodfordian moraine of western Taylor County. Similar fossils have also been reported to the author from the south shore of Lake Superior by university students at Eau Claire and Superior, Wisconsin.

Although this investigation was not basically concerned with an appraisal of drift petrology or mineralogy, field observations at many hundreds of sites indicate that the coarser-than-granular fraction is predominantly composed of Precambrian igneous and metamorphic rocks, and a small amount of Keweenawan Lake Superior sandstone. Pebble or cobble-size pieces of Cambrian sandstone are also found, especially in areas where rock of this type subcrops beneath the drift, but never account for more than a minor fraction of the drift.

In 1919 the Wisconsin Geological Survey conducted a reconnaissance geological investigation in the eastern third of Chippewa County; the results were never published but are on file with the Survey. Several pebble counts from that study, made at scattered localities, are given in Table 1. Local variations in the relative proportions of the various rock types comprising the drift--shown in Table 1--were interpreted as reflecting local variations in the bedrock traversed by the glacier.

The route traversed by the Chippewa lobe was such that masses of elemental copper were incorporated into the drift. The author found none in place but saw perhaps ten specimens of erratic copper discovered by others at scattered locations, especially west of the Chippewa River. The largest specimen, weighing approximately 2200 pounds, was found

Table 1. Relative Frequency (%) of Rock Types in Pebbles of Middle Woodfordian Till*

	Location		
	NE 1/4, sec. 21, T30N, R5W	SW 1/4, sec. 15, T29N, R6W	E 1/4, sec. 20, T29N, R5W
Granite	24	26	23
Basalt	12	5	30
Quartzite	7	18	10
Greenstone	10	14	-
Keweenawan SS	8	14	4
Porphyry	11	5	5
Gneiss	2	-	7
Iron formation	8	2	3
Rhyolite	6	-	7
Gabbro	3	5	2
Syenite/Felsite	4	6	-
Slate	-	4	-
Schist	-	1	-
Quartz	2	-	-
Chert	1	-	-
Unknown	2	-	9
Total	100	100	100

*Adapted from unpublished reports of a 1919 Wisconsin Geological Survey investigation in eastern Chippewa County, on file with the Survey in Madison. These reports are filed according to tier and range number.

in southwestern Rusk County.* Salisbury (1886) discussed copper erratics from several localities in Wisconsin and attributed their source to the Keweenawan volcanics of the Superior basin.

The drift is largely derived from hard Precambrian bedrock and contains many erratics, some of quite impressive size. The dimensions and locations of the four largest erratics observed are given below. The gneiss boulder is particularly interesting because it appears to be contained in an esker, overlain by crevasse-fill deposits.

Table 2. Largest Erratics Observed

Rock Type	Dimensions	Locations
1. Granite	30 x 24 feet	"Keystone Rock," sec. 30, T31N, R7W
2. Gabbro	25 x 12 feet	Yellow R. bridge, sec. 33, T31N, R5W
3. Granite	22 x 20 feet	N. end Fish Lake, sec. 29, T33N, R9W
4. Gneiss	18 x 10 feet	Highway 64 esker, sec. 17, T31N, R5W

Drift Thickness

Logs for approximately 1500 water wells were obtained from several sources. About 1100 were obtained from the Wisconsin Department of Natural Resources, the official repository for all well-construction reports, and approximately 200 were from unpublished Township Reports in the files of the Wisconsin Geological Survey. The remainder came from municipal agencies and homeowners. Figure 8, showing drift thickness, was compiled from the above data and supplemented by information gained by field observations on the distribution of rock outcrops.

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*The owner of this copper erratic had had it weighed at a local grain elevator.

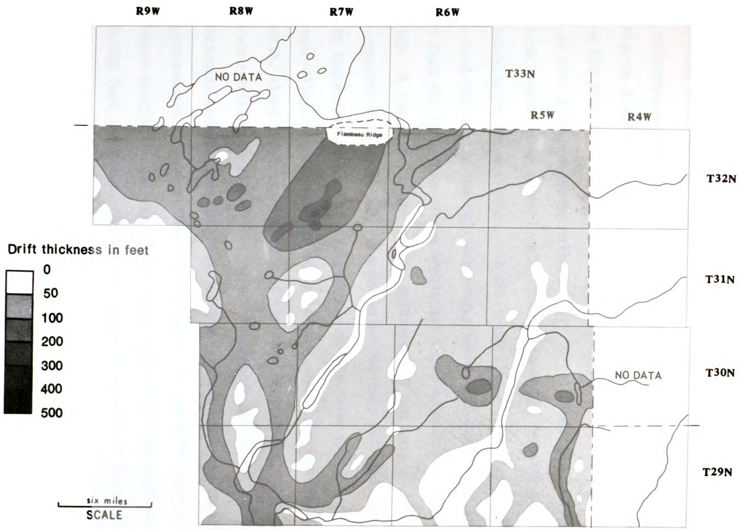

Figure 8. Drift thickness.

The greatest drift thickness is reached in the interlobate tract southwest of Flambeau Ridge (see Figures 6 and 8). In this locality the drift averages more than twice as thick as in most of the remainder of the study area. Along much of the axis of the interlobate tract drift thickness exceeds 300 feet, and beneath the highest glacial ridges (see Figures 8 and 45) it is believed to be over 400 and possibly exceeding 500 feet in thickness. The maximum thickness cannot be determined at this time because no well is known to reach bedrock. However, logs from several deep wells located farther south may provide a reasonable basis for estimates. At the south side of the city of Eau Claire test borings by the city water department indicate an altitude of 620 feet for the floor of the preglacial Chippewa valley in that locality. A few miles north, on the south edge of the city of Chippewa Falls, several deep wells were terminated at 740 feet altitude, but since none of these reached bedrock the altitude of the floor of the preglacial Chippewa valley in that locality is not known. About fifteen miles to the north, along the western margin of the moraine complex, the deepest wells were drilled to slightly below 900 feet above sea level and did not encounter bedrock. On the basis of the foregoing information it seems reasonable to suggest that the thickness of the drift overlying the deeper preglacial valleys of the interlobate tract may reach or exceed 500 feet in places, for the altitudes of several of the higher ridges of this district exceed 1370 feet and Baldy Mountain, the highest point, reaches 1390 feet.

Excluding the interlobate tract, the drift thickness within the moraine complex averages well below 200 feet. East of the Chippewa River only in the Pike Lake moraine, described in Chapter III, does the

drift thickness exceed 200 feet and then only slightly (see Figure 8). Excluding the Pike Lake moraine, in most of the area east of the Chippewa River the thickness of the drift averages between 50 to 100 feet. Based on a small number of logs, it appears that the drift thickness in the moraine complex at the far northwest part of the thesis area, north of Chain and Island Lakes, is similar to that east of the Chippewa River.

Within the moraine complex the areas with less than fifty feet of drift are (1) along the larger rivers, where at least some of the glacial deposits have been removed by erosion and (2) where bedrock hills protrude or nearly protrude through the drift. Flambeau Ridge is an especially striking, though not typical, example of the latter. The ridge has numerous quartzite outcrops and a maximum altitude several hundred feet higher than most of the surrounding drift-covered surface. Figure 8 shows the areas of less than fifty feet of drift and bedrock outcrops. It further reveals that the segment of the moraine complex with the thinnest drift is located between the Chippewa and Yellow Rivers. Due to the irregular character of the bedrock surface underlying the moraine complex, marked local variations in drift thickness are common, both east and west of the Chippewa River. Figure 8 is a generalized isopach map and because of limited data does not show all local variations in drift thickness.

In the transitional or pitted ground moraine area located on the proximal side of the moraine complex (Figure 8), the drift thickness is less, seldom exceeding 100 feet and often under fifty feet. Because there is less variation in the altitudes of both the bedrock and drift

surfaces in this area than beneath the moraine complex, the drift shows less variation in thickness, typically ranging between forty to eighty feet.

Drift Stratigraphy

The stratigraphy of the drift was studied throughout the moraine complex, in the ground moraine area on its proximal border, and to some extent in the older drift area on its distal margin. In the latter area many of the exposures reveal bedrock beneath a thin drift cover; no thick exposure of the older drift itself was found. Furthermore, none of the examined exposures contained buried loess, paleosols, weathering horizons, organic matter, or other buried features suggesting multiple glaciation. In the exposures studied in this area, no drift sequence was observed that would provide a basis for the recognition of more than one ice advance.

Field observations regarding stratigraphy in the young drift area were rather limited because of a lack of deep exposures. In only a few places along the outer margin of the moraine was the contact between the young and old drifts exposed, and these revealed no significant relationships. Elsewhere within the moraine complex the drift outcrops appeared to expose only sediments of the same age.

However, along the proximal margin of the transition zone, where it joins the fluted ground moraine (Figure 6), thin till was observed overlying outwash in three localities. This till appears to have been deposited by a minor glacial readvance which did not reach the moraine complex. The best example of till resting on outwash may be observed in several gravel pits immediately south of the village of Jump River,

which reveal from three to ten feet of till over thirty to fifty feet of coarse gravel. The other two sites are located a few miles northwest and east of Ladysmith.

The attempt to determine drift stratigraphy from well records met with numerous difficulties, some of which are herewith recounted:

1. The detail and accuracy of the well logs varied widely, and many errors were found in the over 1500 well logs examined.

2. The well drillers preparing the well reports did not all use the same terminology for similar types of material, rendering the logs impossible to compare in many cases.

3. Several well logs from localities within the moraine complex record more till layers than is typical for wells in the ground moraine area on its proximal border. Perhaps this reflects the manner of formation of the moraine. Since it is of ice-stagnation origin, the sediments comprising the Chippewa moraine complex often show relationships not typical of normal end moraines. For example, it is not uncommon to observe stratified sediments overlain by or containing several interbedded flowtills, or collapsed superglacial drift exposing gravel masses enclosed by till. Wells penetrating sediments of this type might, therefore, lead to erroneous interpretations if each mass of till recorded were interpreted to represent a separate and distinct drift sheet.

In view of these and other limitations the well logs were evaluated with considerable caution. Well records from both the study area and the older drift area tend to indicate that only one till sheet exists in the old drift area. However, many of the well logs from the area

affected by the Woodfordian glaciation are interpreted to record two tills, and on the basis of this recurring stratigraphic situation in hundreds of well records, it is suggested that two till sheets exist within that area. Although a few wells in the young drift area penetrated more than two hardpan* (till) layers, separated by sorted materials, these were not interpreted to represent additional pulsations of the glacier because of the complications in interpreting drift stratigraphy mentioned earlier.

A number of well logs record only one till within the area affected by Woodfordian glaciation. This situation might be explained if any of the following conditions prevailed: (1) degradational processes may have removed the older drift prior to the last advance; (2) the last advance may have removed the older drift before depositing a till of its own; and (3) the youngest till may have been plastered onto a preexisting till, with no distinctive horizon separating them.

Finally, several well records from the young drift area reported penetrating dark or organic material in the subsurface, but none described more than one such horizon. This, combined with the great preponderance of well logs reporting only one or two hardpan (till) layers, strongly indicates the presence of only one till sheet older than the middle Woodfordian in the Chippewa valley.

Configuration of the Outer Margin of the Moraine

Horberg and Anderson (1956, p. 108) state that "the main factors which controlled the form and extent of the glacial lobes were (1) the

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*"Hardpan" is the term commonly used by local well drillers to refer to till.

preglacial topography, (2) the configuration of the ice sheet, and (3) deflections by adjacent lobes. Among these control by the preglacial topography was undoubtedly the most important single factor, for there is a close correspondence between the glacial lobes and bedrock lowlands." They (ibid., p. 108) also point out that reentrants were formed in the distal margins of moraines where the front of an advancing glacier was slowed by bedrock uplands in its path, whereas the lobes represented the thicker and more rapidly moving ice advancing along lowlands.

Within the study area the most conspicuous reentrant in the moraine margin is that located southwest of Flambeau Ridge. Rock Lake, in sec. 3, T31N, R8W, lies near the apex of this reentrant (Figure 6). The effect of Flambeau Ridge on the glacial movement is discussed in greater detail in Chapter V.

About ten miles south of the reentrant mentioned above there is a similar, but much smaller, feature that formed when the Woodfordian ice margin was at its maximum extent. This reentrant is associated with a prominent sandstone upland which in preglacial time had a minimum of 300 feet of local relief. There is convincing evidence that at its maximum the ice covered only the east half of this upland, although it advanced a mile or two farther in the lowlands on its north and south flanks. This evidence consists of (1) an abundance of variably-shaped sandstone concretions, which have been weathered from the underlying bedrock; (2) a moderate number of sandstone outcrops; and (3) a well-dissected topography on the west half of this upland. In contrast, on the east half of the upland concretions are rare, the topography is

smoother, and glacial erratics are much more numerous. Although no distinguishable morainic topography exists to mark the limit of the ice that advanced over the east half of the upland, the evidence cited seems sufficient for the above interpretations. It should be pointed out that hummocky morainic topography does occupy the easternmost part of this upland.

East of the Chippewa River, the segment of the ice front extending three to four miles on either side of the Yellow River also appears to have been held back by sandstone uplands in its path (Figures 6 and 45). Well records indicate that these uplands stood approximately 150 to 250 feet above the preglacial terrain lying to their east and west. The inconspicuous character of the outermost morainic features of this sector may also indicate that the advance of the glacier was obstructed by these preglacial uplands. The designations 1L, 2L, and 3L, located along the boundary marking the maximum Woodfordian ice extent on Figure 6, refer to salients in the glacier margin associated with the preglacial valleys of the Chippewa River-Drywood Creek, Yellow River, and Wolf River, respectively.

The Maximum Extent of Middle Woodfordian Ice

Determining the maximum extent of the middle Woodfordian ice proved to be a major task because no single characteristic provided a basis for satisfactory results. Extensive field observations indicated that the following characteristics, taken together, appear satisfactorily to define the outer limit of the middle Woodfordian advance, shown on Figure 6.

1. Topography. The topography of the Altonian or old-drift area is typically smoother; is characterized by longer, gentler slopes; and exhibits fewer small hills than the Woodfordian drift area. Where the boundary of the hummocky topography was distinct, the maximum position of the Woodfordian ice front was drawn along the distal border of the hummocky zone. In general, topographic differences of the types described above were useful in establishing the maximum position of the Woodfordian ice front, except within that segment of the moraine complex located between the Chippewa and Yellow Rivers, where there is usually not a sharp topographic discontinuity separating the two drift areas.

2. Drainage integration. The older drift area exhibits a more orderly and advanced drainage network than does the young drift. This characteristic was of greatest assistance in establishing the Woodfordian/Altonian drift boundary in the area east of the Yellow River (Figure 6).

3. Swamps and poorly drained depressions. Besides there being a greater number of poorly drained depressions north and east of what is interpreted as the middle Woodfordian front, the swamps of that area are more confined to distinct topographic depressions than in the older drift area. For example, most of the poorly drained lands within the older drift region are situated on broad floors of low-gradient, outwash-dammed tributary valleys. In contrast, those of the young drift area are typically situated within hummocky topography.

4. Lakes and ponds. West of the Chippewa River the area affected by middle Woodfordian glaciation contains a great many lakes and an unusually high density of small ponds which markedly contrasts with the almost total lack of such features in the older drift region. On the other hand, both the young and old drift east of the Chippewa River

show a virtual absence of lakes and ponds, with the result that this characteristic proved of little value in establishing the limit of the Woodfordian ice in that area.

5. Color of the drift. There appears to be a definite color difference, in the upper few feet at least, between the Altonian and Woodfordian tills. This characteristic proved to be one of the more useful in delimiting the middle Woodfordian border in the uplands east of the Chippewa River, for in these localities the young and old till areas are not separated by middle Woodfordian outwash. On the basis of numerous Munsell color observations, made at depths of three to six feet on both sides of the suspected boundary, reddish-brown proved to be the typical color of the till within the young drift area, whereas yellowish-brown or light brown was the usual color of the older till. However, till color itself does not always serve as an adequate basis to define the middle Woodfordian advance, for the till deposited near the margin of this ice advance often consists of reworked, somewhat altered, yellowish-brown Altonian till, mixed with varying amounts of younger reddish-brown till (Black, 1959). Due to the lack of a continuous, well-defined color boundary between the two drift areas, till color had to be used in conjunction with other criteria to differentiate them.

6. Rock outcrops. Natural outcrops and roadcuts exposing bedrock are common in most of the older drift area. However, they are rare in the young drift area, except along the southern fringe of the moraine complex between the Chippewa and Yellow Rivers where sandstone occasionally outcrops within an area of generally thin drift. The boundary between the Altonian and Woodfordian drift areas was interpreted as approximately

following a line which separates the virtually outcrop-free area from the area of frequent bedrock outcrops.

7. Periglacial features. Several types of periglacial features are common in the Altonian drift but exceedingly rare in the young drift area. These include ice-wedge casts, exposed in most roadcuts through Altonian kamic features; numerous and well-developed ventifacts; and occasional eolian sand deposits overlying a layer of wind-faceted stones. Collectively, these features were very helpful in delimiting the young drift. That is, the boundary separating Altonian from Woodfordian drift was drawn between the area containing numerous periglacial features and that where these features are scarce or absent.

8. Proglacial meltwater channels cut into or across ridges. The features of this nature are believed to have been formed by meltwaters draining southward from the Woodfordian ice front when it stood at its maximum position. Six instances of such features, located between a point two miles north-northwest of Cadott and the Yellow River, four miles to the east (Figure 45), served as a partial basis for delimiting the middle Woodfordian advance in that area.

Another interesting example of a valley apparently cut by middle Woodfordian meltwaters draining across a ridge peripheral to the glacier exists along the headwaters of Goggle-eye Creek, about two miles north-northeast of Thorp, in the southeastern corner of the study area. Although this forty-foot deep valley lies about one-half mile outside the east boundary of the study area, it is discussed here because it is an important feature in establishing the extent of the middle Woodfordian advance in that locality. This valley appears to have been

cut by water draining southward across a bedrock ridge from a small proglacial lake that formed between the ice front and the ridge. The segment of Goggle-eye Creek located north of the low ridge possesses several small barbed tributaries, implying that the preglacial drainage north of the ridge was locally toward the north. The foregoing conditions, in addition to the small size of Goggle-eye Creek relative to its valley, suggests that it is an underfit stream, occupying a valley cut by meltwater.

9. Pitted outwash. Thwaites (1926) discussed the diagnostic importance of pitted outwash in ascertaining the extent of individual ice advances. His comments suggest that where a moraine is fronted by equivalent-age pitted outwash, the minimum extent of the glacier producing the outwash is marked by the outermost extent of the pitted outwash, assuming both the buried ice and its outwash cover were from the same glacier.

West of the Chippewa River the moraine is fronted almost continuously by pitted outwash, which itself is commonly fronted by unpitted outwash; in this area the proposed boundary of the middle Woodfordian advance essentially follows the outer limit of the pitted outwash.

East of the Chippewa River, however, outwash is more localized, being mainly confined to the larger valleys and lowlands. Pitted outwash was used to establish the middle Woodfordian boundary in the outwash area immediately east of Lake Wissota, in Hay Creek valley two miles northeast of Boyd, in the Wolf River lowland immediately northwest of Stanley, and in the Eau Claire River valley, two miles northwest of Thorp.

10. Eskers. In two localities eskers, assumed to be of Woodfordian age, were of assistance in defining the minimum extent of ice associated

with this advance. At the junction of Highways 178 and 64, twelve miles due north of Chippewa Falls, a well-defined esker, partially buried by low-relief middle Woodfordian pitted outwash, lies just east of the proposed front but to the distal side of the main moraine complex.

At the second location, about three miles north-northwest of Cadott (Figure 45), glacial features are poorly developed, except for two eskers that are interpreted to be of Woodfordian age. In this locality these eskers provide a basis for establishing a minimum position of the maximum Woodfordian ice extent.

CHAPTER III

EVIDENCE FOR SUBDIVISION OF THE MORAINE

Introduction

In this chapter evidence is presented which pertains to two of the major problems (stated in Chapter I) that were investigated in this study. One of these was to determine if the Chippewa lobe end-moraine zone, shown on Figure 1 as comprised of a single moraine, may be subdivided into a complex of smaller moraines. A second major problem was to ascertain whether or not the end-moraine zone southwest of Flambeau Ridge (Figure 6) includes an interlobate tract. The two problems will be discussed separately, in the order listed above.

Evidence for a Moraine Complex

On the Wisconsin Glacial Deposits map (Thwaites, 1956) no subdivisions are shown for the Chippewa moraine (Figure 1). The results of this investigation, however, suggest that it is a moraine complex, representing more than one phase of ice-front activity. That is, the Chippewa moraine typically consists of a number of semiparallel belts of hummocky terrain, separated by discontinuous lowlands, the latter underlain by stratified sediments. Furthermore, within each of the hummocky belts there exists a great variety of landform types, whose morphology and relationships collectively indicate that the various hummocky belts evolved--in general--independent of each other.

The composite character of the moraine complex, although well expressed east of the Chippewa River, could not be established for the segment lying west of the Chippewa. There are at least two reasons for

this. First, the proposed interlobate tract (Figure 9), extending nearly the full width of the moraine complex and accounting for about one-half of the area west of the Chippewa River, consists of a single hilly tract. Second, the remaining area, or that lying north of the interlobate tract, falls largely in the area not covered by large-scale, small-contour interval maps. Although field observations suggest that this part of the moraine zone is composed of several moraines, confirmation of this is difficult because of the lack of adequate topographic maps.

East of the Chippewa River the number of moraines comprising the moraine complex varies, with the greatest number located between the Chippewa and Yellow Rivers. The total number is at least six and may be as high as eight or nine. These moraines are shown on Figure 9. The uncertainty over their number arises because the proximal margin of the moraine complex is not composed of distinct morainic belts (Figure 6). Rather, it consists of disjunct areas of hummocky topography. In places these seem to line up into belts; it is on this basis that the moraines lying north of Highway 64 (Figures 9 and 45) were delimited.

Several transverse profiles of the moraine complex were constructed, three of which are reproduced in Figure 10. On these profiles, as well as on Figure 9, the moraines are named. The communities or physical features after which they were named are shown on Figure 45. Due to their relatively small sizes and variations in longitudinal development, all of the moraines may not be recognized on a given profile, but collectively the profiles show most of the moraines recognized on Figure 9.

Only three of the moraines have continuity and are large enough to be recognized easily on maps or in the field. These three from south

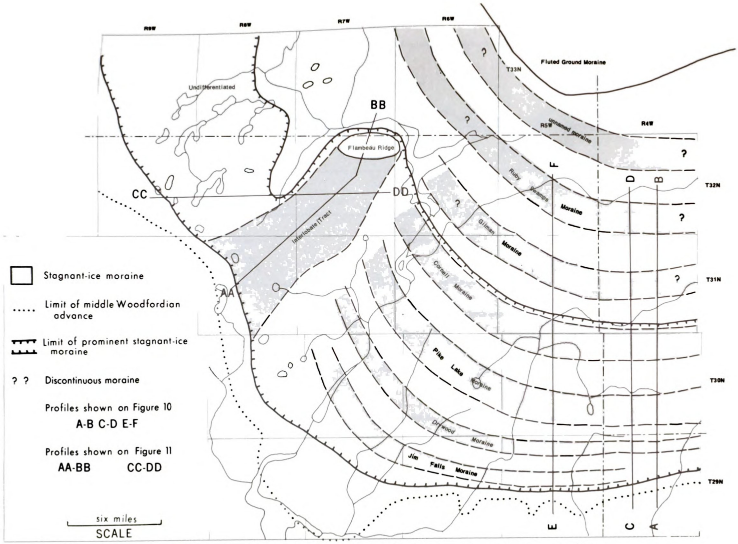


Figure 9. Subunits of the Chippewa moraine complex.

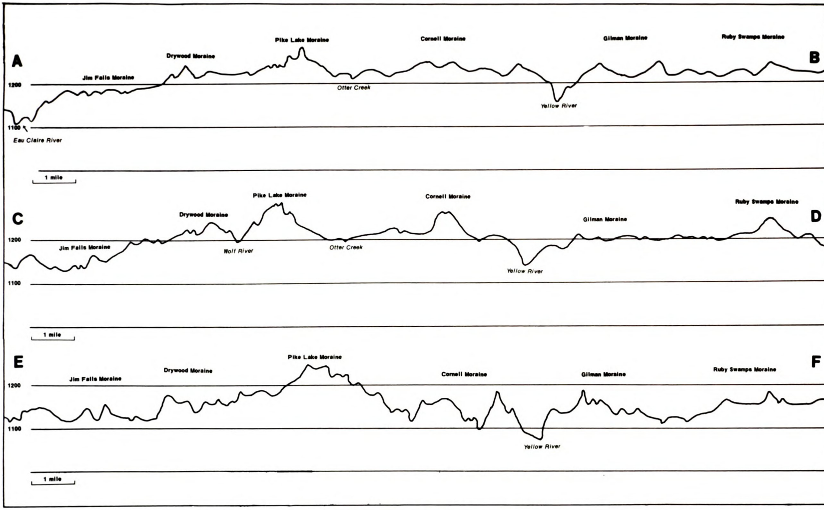


Figure 10. Transverse profiles of the Chippewa moraine complex. The locations of the above profiles are shown on Figure 9. The Jim Falls Moraine, the outermost discernible moraine of the complex, is not well expressed in the eastern part of the study area.

to north are Drywood, Pike Lake, and Cornell, of which the Pike Lake moraine is the largest.

The individual moraines average about one to two miles in width and show equivalent distances in their spacings. Their sizes are typically not impressive, with their maximum overall relief commonly between 50 to 150 feet and their local relief usually 20 to 60 feet.

In some parts of the moraine complex its composite (belted) character is not apparent. There are several reasons for this circumstance: (1) In places successive moraines merge. (2) The topography in thin drift areas often reflects the preglacial bedrock topography more than it does the pattern of the moraines. This is especially true in the southern part of the moraine complex east of the Chippewa River and to some extent in southwestern Rusk County, north of Island Lake (see Figure 9). Several of the buried sandstone uplands of these sectors are shown on Figure 8. (3) The inner transitional zone, bordering the proximal side of the moraine, shows only a weak tendency toward separate moraines. And (4) the individual moraines are occasionally interrupted by gaps which, in the case of the poorly developed moraines, makes them difficult to map. These gaps may reflect drainage routes across the moraine(s) during deglaciation, places where originally low sections of the moraine were buried by outwash, or possibly, the gaps may indicate that a moraine was never developed at these places.

Evidence for an Interlobate Tract

The term "interlobate moraine" was introduced by Chamberlain (1883a, p. 276) to describe a section of an end moraine formed "between two adjacent glacial lobes." Chamberlain (1883b, p. 302) also referred

to an interlobate moraine as originating by "the joint action of two glacial lobes pushing their marginal moraines together and producing a common one along the line of their contact. They are terminal moraines in character but interlobate in position." In the same publication he described the morainic loops which extend across the northern United States from the Great Plains to the Atlantic Coast, characterizing them as reaching their greatest height and roughness in interlobate locations, where the morainic loops join.

Southwest of Flambeau Ridge, in the northwestern part of Chippewa County, there is evidence that the Woodfordian end moraine is of interlobate origin. Topographically, this part of the moraine complex consists of a high range of hills oriented perpendicularly to and extending northeastward from the proximal side of the moraine complex (Figure 6). Terminating at Flambeau Ridge, this four- to six-mile-wide range of hills is the most prominent part of the moraine. Figures 11 and 12 illustrate the foregoing generalization. Figure 11 shows that the altitudes and local relief of the hypothesized interlobate tract (right-hand portion of the lower profile) are conspicuously greater than in other parts of the moraine (left-hand part of the same profile).

Several phenomena suggest that Flambeau Ridge did influence the movement of the glacier margin and, thereby, the formation of the moraine to its southwest. Perhaps the most convincing evidence is the existence of a conspicuous reentrant in the distal margin of the moraine southwest of Flambeau Ridge (Figure 6). The presence of this reentrant suggests that the southwestward movement of the glacier in this area was impeded by Flambeau Ridge. This, in effect, resulted in the formation of two minor sublobes of ice downglacier from Flambeau Ridge.

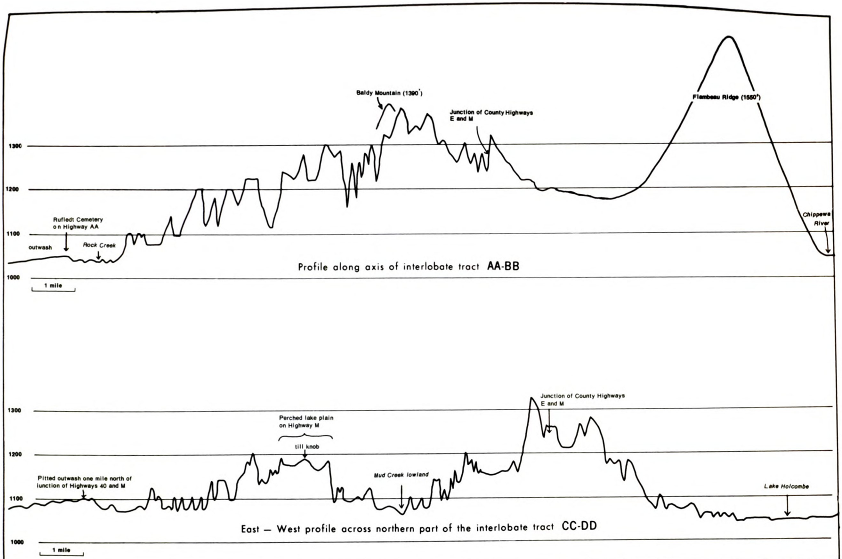


Figure 11. Topographic profiles of the interlobate tract. See Figure 9 for the location of the above profiles.

Figure 12. Westward-looking view of Flambeau Ridge and the interlobate morainic tract on its south, from a fire-lookout tower about twelve miles east-southeast of Flambeau Ridge. The interlobate tract forms the left-hand two-thirds of the skyline.

The overall morphology of the impressive hilly tract southwest of Flambeau Ridge also suggests the foregoing interpretation. As shown on Figure 11 the highest altitudes of this tract are located along its axis. While the greater part of the tract consists of a jumble of irregularly-shaped ridges composed of till, intensive field study indicates that the series of prominent hills along its axis consist predominantly of stratified drift. These relationships, combined with the data on drift thickness (Figure 8), suggest that the hilly tract southwest of Flambeau Ridge formed between two minor glacial sublobes and is, therefore, of interlobate origin.

CHAPTER IV

EVIDENCE FOR MARGINAL STAGNATION

Introduction

One of the major problems of this study was to determine whether or not the Chippewa lobe did experience marginal stagnation. The results of this investigation indicate strongly that this condition did occur and, of greater significance, that the landforms of the Chippewa moraine complex are largely the result of stagnant-ice disintegration, rather than an orderly retreat of an active margin of the ice sheet.

Evidence for Marginal Stagnation

Numerous types of evidence exist which demonstrate that extensive stagnation took place during the deglaciation of the area. For convenience of discussion, the various types may be arranged into two basic lines. The first consists of evidence--both topographical and sedimentological in nature--indicating marginal stagnation, but not necessarily the existence of large amounts of buried ice. The second line of evidence, again concerning both topographic and lithologic conditions, demonstrates the presence of extensive areas of drift-mantled ice. The discussion of both lines of evidence includes some interpretation, but only to the extent needed to establish its relevance or significance.

Evidence for stagnant ice

Eight different types of geomorphic evidence indicate that marginal stagnation was a significant mode of deglaciation of the Chippewa lobe at the time of formation of the Chippewa moraine complex. The evidence in this category includes phenomena from a variety of relative locations.

Character of outer margin of moraine.--As described in Chapter II the distal margin of the zone of conspicuous morainic topography seldom coincides with the outer limit of the middle Woodfordian advance. In particular, fronting the moraine complex is a one- to five-mile-wide belt consisting of pitted outwash and/or indistinct morainic topography, the latter containing eskers and kames indicative of stagnant ice. Within the present study area it appears that very soon after the middle Woodfordian ice had reached its maximum position, the marginal part of the glacier stagnated, before it had had sufficient time to construct more than a small discontinuous moraine.

Hummocky nature of moraine complex.--Most of the Chippewa moraine complex consists, topographically, of small, closely-spaced, irregularly-shaped hills and depressions which, collectively, display considerable uniformity in size (Figure 19). The knolls do not seem to be aligned into ridges and no dominant trends are discernible. Moreover, the hills are predominantly composed of till but, occasionally, may contain small amounts of gravel or washed till. In describing the differences between moraines of active-ice and stagnant-ice origin, Gravenor and Kupsch (1959, p. 50) state that hummocky topography of the type described above is diagnostic of moraines of stagnant-ice origin.

Northward-sloping outwash plains.--The presence of northward-sloping "outwash plains" at several places along the proximal side of the moraine complex--notably in the Mud Creek lowland (Figure 6)--

represents an anomaly in terms of the typical relationships of outwash plains. Normally an outwash plain borders the distal side of a moraine and slopes away from the area occupied by the active glacier. The northward-slope of some outwash plains on the proximal margin of the Chippewa moraine complex suggests that the outwash was deposited by northward-draining meltwater, derived from the extensive stagnant ice within the moraine--after withdrawal of the glacier.

Ice-contact features.--The Chippewa moraine complex contains numerous hills, ridges and mesa-like features consisting entirely, or in part, of stratified drift. In his classification of glacial sediments Flint (1972, p. 183) states that

> stratified drift (or washed drift) falls into two classes, based on the characteristics that indicate the environment of deposition. One class, built upon or immediately adjacent to glacier ice, is ice-contact stratified drift, or more simply, ice-contact features, a term that emphasizes surface form as well as internal characteristics. The other class consists of proglacial sediments which, although of glacial origin, are built beyond the glacier itself, in streams or lakes. The two classes grade into each other at the margin of the glacier.

According to the above classification, landforms built in contact with stagnant ice and composed of stratified drift are termed "ice-contact features." Internally, three general characteristics distinguish the sediments of such features from non-ice-contact stratified drift: (1) extreme range and abrupt changes in grain size, (2) inclusion of bodies of till such as "flowtill" (Hartshorn, 1958; Boulton, 1968) resulting from superglacial mudflow, and (3) marked deformation of the stratification.

Ice-contact features with the foregoing characteristics are distributed widely within the Chippewa moraine complex and indicate the widespread existence of stagnant ice during deglaciation. In the following paragraphs several specific types of ice-contact features are discussed.

<u>Glaciolacustrine features</u>.--Glaciolacustrine sediments (stratified clays, silts and associated sands, deposited in glacial lakes) are common within the moraine complex. In relative location they typically underlie mesa-form hills or perched plains whose surfaces are normally concave upward. They appear identical to the perched lacustrine plains described by Winters (1961, p. 20) and the ice-walled-lake plains described by Clayton and Cherry (1967), who interpret them as consisting of sediments deposited in lakes enclosed by stagnant glacier ice. Within the study area the features of this type are commonly bounded on all sides by ice-contact slopes, but their surfaces are usually unpitted, indicating that the ice-walled lakes were, for the greater part, not underlain by stagnant ice.

<u>Perched abandoned drainage channels</u>.--One of the more significant types of topographic evidence for the former existence of stagnant ice within the moraine complex consists of the numerous remnants of perched abandoned stream channels. The typical feature of this type consists of a small valley incised into the surface of a perched lake plain which, itself, rises above the surrounding hummocky topography.

These abandoned perched valleys appear to have been formed in the following manner. Some of the ice-walled lakes, before their drainage, had had streams flowing into and out of them. When the stagnant ice enclosing these lakes had melted down sufficiently to cause drainage of the lakes, but not the destruction of the streams draining into them, the streams then flowed across the beds of the drained lakes. Disintegration of the buried ice resulted ultimately in complete destruction of the superglacial streams, but not before at least some of them had eroded valleys into their respective lake beds. In effect, then, the perched valleys represent short remnants of superglacial stream channels, preserved only where the channels were not underlain by stagnant ice.

Perched linear gravel plains.--At several places within the study area flat and linear tracts exist that are underlain by deposits of stratified sands and gravels which may extend from the margin of the moraine into its interior. In all cases they slope toward the moraine margin, are of a higher altitude than the hummocky topography that borders them, and are bounded by ice-contact slopes. They appear to represent sands and gravels deposited by streams that flowed in ice-walled trenches. Melting of the adjacent ice left these linear "outwash" features perched above the adjoining hummocky topography.

Eskers and kames.--Numerous examples of both eskers and kames are present in the study area. At least nineteen eskers were identified (Figure 45), most of which are associated with the pitted ground moraine area on the proximal margin of the moraine complex. As pointed out earlier, eskers, as well as kames, are also found at scattered locations within the zone of poorly-expressed morainic topography bordering the distal margin of the moraine. The greatest abundance of kames, however, is within the moraine. In summary, eskers and kames are common within the area and, according to Flint (1971, p. 207), are usually indicative of marginal stagnation of a glacier.

Evidence for buried ice and superglacial drift

Although the foregoing evidence does strongly indicate marginal stagnation, it is not offered as proof that the stagnant ice was thickly mantled with superglacial drift. However, evidence does exist which indicates that condition.

High-relief perched lake plains.--At least several dozen perched lake plains whose local relief exceed forty feet exist within the study area at widely scattered locations. Several of these are more than one mile in diameter. One of the most fundamental questions concerning these features is the origin of the sediments comprising them. As interpreted by Gravenor and Kupsch (1959), Clayton (1962), Winters (1963), and Parizek (1969), perched lake plains consist of sediments washed into ice-walled lakes from drift-mantled stagnant ice enclosing them. Clayton (1962, p. 37) uses the term "superglacial drift" for the ablation drift mantling the stagnant margin of a glacier.

The principal lines of evidence indicating that the stagnant ice of the Chippewa moraine complex was mantled by superglacial drift during formation of the perched lake plains concern the grain-size characteristics and thickness of the glaciolacustrine sediments. Although many of the perched lake plains are characterized by sands and gravels in their margins, their central parts usually consist of clay and silt, indicating that the rate of meltwater production on the enclosing ice--and, thus, runoff into the ice-walled lakes--was relatively low. Additionally, a large number of the lake plains are underlain by glaciolacustrine sediments exceeding forty feet in thickness and, in a few cases, more than seventy-five feet. This indicates that at least some of the ice-walled

lakes lasted for an extended period of time. Such conditions would have been possible only if the stagnant ice enclosing them melted slowly. The slow rate of melting indicates that the ice was largely covered with a blanket of thick superglacial drift.

Interbedded flowtills.--Many of the roadcuts through the margins of perched lake plains exhibit one or more till layers of local extent--from a few inches to several feet thick--interbedded with ice-contact stratified sediments (see Figures 13, 14, and 15). Both Hartshorn (1958) and Boulton (1968) maintain that till layers of this nature could only be on "flowtill" origin, derived from water-saturated superglacial till. In conclusion, the numerous instances of this phenomenon within the Chippewa moraine complex are interpreted as indicating that during deglaciation the stagnant ice was extensively mantled by superglacial till.

Relationships of the intermorainal lowlands.--East of the Chippewa River the moraine complex generally consists of several semi-parallel moraines, separated by discontinuous linear depressions. In many places these depressions are underlain by unpitted stratified drift. But of particular importance is the fact that the surfaces of the latter areas are typically concave upward, with coarser sediments near their margins, and with their margins bounded by ice-contact slopes. In addition, ice-block basins exist within the moraines, located both to south and north of the lowlands, that often have bottoms extending well below the altitudes of the lowland plains separating the moraines. The foregoing conditions indicate that (1) sediments were washed into the intermorainal depressions from both sides, concurrently, accounting for their concave surface form as well as the decrease in sediment grain size from

Figure 13. Flowtills in gravel pit wall. Location: about one-fourth mile north of the west end of Jim Falls dam on the Chippewa River. The three- to four-foot-thick layers of unsorted stony material in the upper and middle parts of the pit wall consist of flowtill. Stratified sediments (glaciolacustrine ?) separate the two flowtills as well as underlie (behind the shovel) the lower one.

Figure 14. Multiple flowtills exposed in a roadcut through the rim of an ice-walled lake plain in sec. 10, T32N, R7W. The stony flowtill layers are separated by darker layers, containing stratified silts, sands, and gravels. The ice-contact slope of this rim is just out of view to the right side of the photo. Fault structures, typical of ice-contact features, are visible in the right half of the photo. See Figure 15 for greater detail on this same exposure.

Figure 15. Closeup view of multiple flowtills exposed in a roadcut through the rim of an ice-walled lake plain in sec. 10, T32N, R7W (also shown on Figure 14). The dark layers consist of well-stratified glaciolacustrine sediments (see left side of photo from the shovel upward to the top of the roadcut), but the stony, unsorted materials separating the dark layers (stratified) are of flowtill origin. See Figure 14 for additional descriptive information.

their margins inward; (2) that the moraines were at least in part ice-cored, accounting for the relatively low altitudes of the ice-block basins within them; and (3) that the stratified sediments of these inter-morainal depressions could only have been derived from the superglacial drift which blanketed the ice-cored moraines, and not from an active ice front.

Buried gravel masses.--Numerous roadcuts within the areas of hummocky topography expose variably-shaped deposits of gravel within till. Moreover, some of these buried gravel deposits contain well-sorted materials but show absolutely no stratification. These relationships are interpreted here to indicate that gravels deposited by surface runoff in the depressions of the superglacial topography were later buried by superglacial slump or flow till derived from slopes adjoining these depressions. Or this phenomenon may have resulted when superglacial alluvium was mixed with superglacial till during the disintegration of the drift-mantled stagnant ice.

Buried silt tongues.--Numerous roadcuts within areas of hummocky topography reveal tongues of silt extending downward from the surface into the till of the morainic ridges. The number, size, shape, and relative orientation of these tongues varies greatly. They number between two to five tongues per morainic ridge, range between one and ten feet in length, and vary from a few inches to five feet in diameter. They are occasionally branched, may pinch-out to reappear a few feet away, and vary between vertical to nearly horizontal in orientation. These silts may in some cases have been of loessal origin, but more

likely they were originally deposited on top of superglacial till in ponds in ice fractures or openings in the stagnant ice. When the buried ice disintegrated, rearrangement of the superglacial materials resulted in a mixing of the superglacial silt and till.

Till-capped eskers.--About one-half of the eskers identified within the thesis area reveal either a continuous or discontinuous till mantle up to fifteen feet thick, underlain by esker gravels. This till is interpreted as representing superglacial till, let down onto the eskers when the overlying drift-mantled ice melted. The above interpretation is proposed because the eskers show no structures or features, such as completely disrupted bedding or gaps, that would indicate a glacial readvance. If these interpretations are correct, then the widespread distribution of till-mantled eskers within the area represents additional evidence of the presence of extensive drift-covered stagnant ice during deglaciation.

In summary, the purpose of this discussion has been to show that a large variety of phenomena exists which indicates that marginal stagnation was the predominant mode of deglaciation during the formation of the Chippewa moraine complex. The characteristics of the landforms and sediments associated with that feature also indicate the existence of widespread superglacial drift during the disintegration of the stagnant ice.

CHAPTER V

PROPOSED THEORY OF FORMATION OF THE CHIPPEWA MORAINE

Introduction

In Chapter III it was stated that the distal margin of the zone of conspicuous morainic topography seldom coincides with the outer limit of the glacial advance of middle Woodfordian age (see Figure 6). The primary reason for this interpretation is the fact that an end moraine--defined as a ridge composed of debris that accumulated at the margin of an active glacier (Flint, 1971, p. 200)--is virtually absent. It appears that soon after the Chippewa lobe had reached its maximum extent, but before it had had time to construct more than a small discontinuous end moraine, the marginal part of the glacier, from one-half to several miles in width, thinned and stagnated. Thus, the conspicuous part of the moraine complex is not a true end moraine, but consists of hummocky stagnation moraine.

Formation of the Noninterlobate Section of the Moraine Complex

On the basis of the evidence presented in Chapters III and IV the following manner of formation is proposed for the Chippewa moraine complex. During its initial advance the Chippewa lobe probably had a positive economy, and advance continued until marginal thinning led to a stabilization of the ice front. In Flint's (1971, p. 47) words,

> the profile of the terminus of a glacier with a strongly positive economy is a smooth, steep curve, so that the ice is comparatively thick even a short distance upglacier from its margin. However, termini of glaciers with a negative economy or with an inactive regimen can present a different appearance.

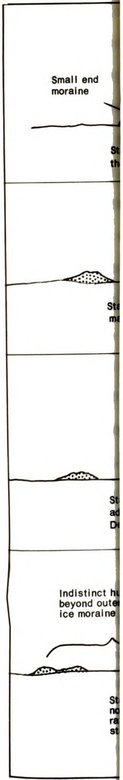
In a paper concerning the Greenland icecap Bishop (1957) points out that when the margin of the ice sheet is in a steady-state condition, a narrow zone at the margin will be so thin (200 to 250 feet thick) that the rate of creep within the zone will be negligible. The zone acts as a stagnant barrier that impedes the flow of the thicker ice upstream; that ice, flowing compressively, rides up over the barrier, forming steeply dipping shears or thrust faults. Should a negative economy ensue (ablation greater than accumulation), the stagnant barrier widens and the zone of overthrusting migrates upglacier. Rock debris, brought up from the subglacial floor by shears, reaches the ice surface where it is gradually released by ablation and may cover the ice; thereafter, it is slowly let down onto the ground by wastage of the buried ice to form stagnation moraine.

Souchez (1967) mentions another mechanism bringing debris up from the base of the glacier which may lead to the formation of an ablation blanket of fine-grained superglacial debris near the glacier margin. This process occurs before marginal stagnation sets in and consists of the formation of debris layers (termed "Wertman debris layers") which form along subhorizontal micro-shear or flow planes within the glacier. They are probably the result of the "freezing-in" process at the glacier bed and the upwarping of flow lines at the margin of the glacier. The dip of these shear planes, developed along upwarped flow lines, is much less than that of the fault planes found in the zone of active overthrusting, formed after the margin becomes too thin to flow. The fault planes were observed by Souchez to cut the low-angle Wertman debris layers, indicating that they formed after the latter.

In the case of the part of the Chippewa lobe under consideration, several successive zones of overthrusting, parallel to the ice margin as well as to one another, seemed to have formed before the glacier ceased to be active. The stagnant-ice zone that was located immediately beyond the outermost zone of overthrusting appears to have ranged from one to five miles in width. The widest part of it is represented by a broad belt of pitted outwash, discussed by Mathiesen (1940), which adjoins the distal margin of the moraine from northwestern Chippewa County northward (Figure 6). The Chetek chain of lakes and the lakes near Rice Lake--located from five to fifteen miles northwest of the study area--occupy ice-block basins in this pitted outwash.

Apparently, the outer stagnant zone was of greater width along this section of the glacier margin because the latter had overrun higher bed-rock uplands (the Blue Hills and sandstone uplands bordering them on the south) than had the ice margin farther south and east. Therefore, when marginal thinning occurred, the contact between the stagnant and active parts of the glacier was displaced farther upglacier for this section of the ice margin than was the case farther southeast, where lower bedrock surfaces resulted in slightly thicker ice in the marginal zone and, consequently, a narrower stagnant margin.

The zones of overthrusting, and their resulting ice-cored moraines, probably originated in the following manner (see Figure 16). Soon after the ice had reached its maximum extent, the margin of the ice became so thin that there was a virtual cessation of ice movement in marginal areas. This was followed by a readvance of the active ice that overrode the edge of the stagnant ice, but never advanced far enough to cover it to a significant degree.



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The zone of overthrusting migrated slowly upglacier, perhaps because of continuous thinning and progressive stagnation of the ice immediately upglacier of the overthrusting zone, due to a slight negative economy for the glacier. Active overthrusting was eventually terminated by a marked thinning of the ice upglacier from the overthrusting zone. During this relatively inactive phase, thinning in the ablation zone produced a belt of stagnant ice one to two miles wide bordering the upglacier side of the previous zone of overthrusting. This was followed by another phase of rethickening, reactivation of glacial advance (see Figure 16), and the formation of a second zone of overthrusting, separated from the original overthrusting zone by the belt of relatively clean stagnant ice. This general sequence of events was apparently repeated several times, for the moraine complex typically consists of several semiparallel belts of hummocky terrain. As envisioned here, each major zone of overthrusting eventually became manifested as a separate ice-cored moraine, to result in a belt of hummocky moraine after disintegration.

The small size of these moraines may reflect the relative lack of vigor of the Chippewa lobe, as well as the hard Precambrian rocks it traversed, which yielded less material for the glacier's deposition at its margin than might have been deposited had it crossed softer rocks. Also, the duration of construction of the Chippewa moraine may have been less than was the case for the construction of the end moraines of nearby ice lobes (the massive St. Croix moraine, e.g.).

Although the composite character of the moraine complex is discernible east of the Chippewa River, it could not be established for the segment lying west of the Chippewa. The reasons for this were discussed

in Chapter III. East of the Chippewa River the number of stagnation moraines comprising the moraine complex varies, with the greatest number located between the Chippewa and Yellow Rivers (Figure 9). This segment of the moraine complex will be referred to as the axial zone, for it represents the general zone occupied by the northeast-southwest-trending longitudinal axis of the Chippewa lobe. Within the axial zone the ice reached its greatest thickness and mobility. The outermost stagnation moraine of the axial zone (Figure 9) cannot be traced eastward much beyond the Yellow River, suggesting that the outermost zone of over-thrusting in the axial zone was not expressed in the ice margin farther east.

Two additional characteristics of the stagnation moraines in the axial zone merit attention. Not only are they of smaller size in this segment of the moraine complex, but they display slightly greater spacing between successive moraines than elsewhere. Their relatively small size may reflect less marginal overthrusting in the axial zone, due to thicker, more mobile ice, combined with a more continuous upglacier migration of the locus of overthrusting for this part of the ice margin. The latter condition would preclude the eventual formation of distinct stagnation moraines, since the superglacial debris would have been less confined to distinct belts than elsewhere along the ice margin. The wider spacing between moraines in the axial zone suggests a faster rate of retreat of successive zones of overthrusting than elsewhere along the glacier margin.

In Chapter III it was established that at least six and possibly as many as nine stagnation moraines comprise the moraine complex. The uncertainty regarding the exact number arises because the disjunct mo-

rainic areas on the proximal margin of the moraine complex (north of Highway 64) are not distributed in distinct belts.

A recent study by Larson (1972) analyzes an area of parallel drainage contiguous with the northeast corner of the study area. Larson proposed that the parallel drainage pattern of that area is associated with fluted ground moraine (Figure 6). This physiographic district, whose topography was formed beneath active ice, is separated from the moraine complex by a transitional zone, ten to fifteen miles in width. This zone contains numerous areas of poorly developed drumlinoid topography, separated by tracts or belts of stagnation moraine. Both types of topography are shown on Figure 45. The areas of hummocky stagnation moraine are distinctly different from the smoother ground moraine which separates them but are quite similar in both array of landforms and sediments to the larger stagnation moraines located to the south and west.

Distinct belts of stagnation moraine are absent north of Highway 64, possibly because after the zone of overthrusting had retreated north of this line, it slowly but continuously shifted upglacier, not making prolonged intermittent halts as was the case south of Highway 64. Apparently, the transition zone described above separated the active central part of the Chippewa lobe from the marginal area of stagnant ice. The drumlinoid parts of the transition zone suggest that the action of the ice in these areas consisted primarily of basal shearing and slippage, with negligible overthrusting within the glacier. However, other parts of the same zone apparently experienced moderate overthrusting, as evidenced by the scattered areas of hummocky stagnation moraine. Wastage of the resulting debris-laden ice in these localities resulted in

topography distinctly different from that produced by melting of the intervening sections of relatively clean ice. Figure 6 shows the extent of the transitional zone of ground moraine/stagnation moraine, relative to the area of fluted ground moraine on its proximal side and the prominent stagnation moraine on its distal side.

Formation of the Interlobate Tract

Evidence was presented in Chapter III indicating that Flambeau Ridge, a conspicuous bedrock feature, was a major influence on the formation of the moraine complex located to its immediate southwest. In a paraphrase of the words of Horberg and Anderson (1956, p. 109), where a highland retarded an advancing ice margin, the ice immediately downglacier from that highland experienced a relative increase in ablation, inducing flow convergence toward the locus of thinning.* If one applies this principle to the Chippewa lobe, it appears that Flambeau Ridge caused a bifurcation of the ice flow, i. e., two minor sublobes, which reconverged to its leeward (Figure 6). The impressive morainic topography lying southwest of Flambeau Ridge is, therefore, interpreted to be of interlobate origin.

Black (1970), in reevaluating the origin of the glacial features of the Kettle Interlobate Moraine of southeastern Wisconsin, incorporated some of the recent findings on the formation of stagnant-ice landforms and glacial deposition in interlobate environments. He states that this

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*Given equal ablation rates, the ice surface downglacier from an overridden highland will melt down more rapidly than will the ice at either side of this location. This results because the ice lost by ablation on the part of a glacier distal of a highland is not replenished as rapidly as where the ice movement is unobstructed and the flow velocity, therefore, more rapid.

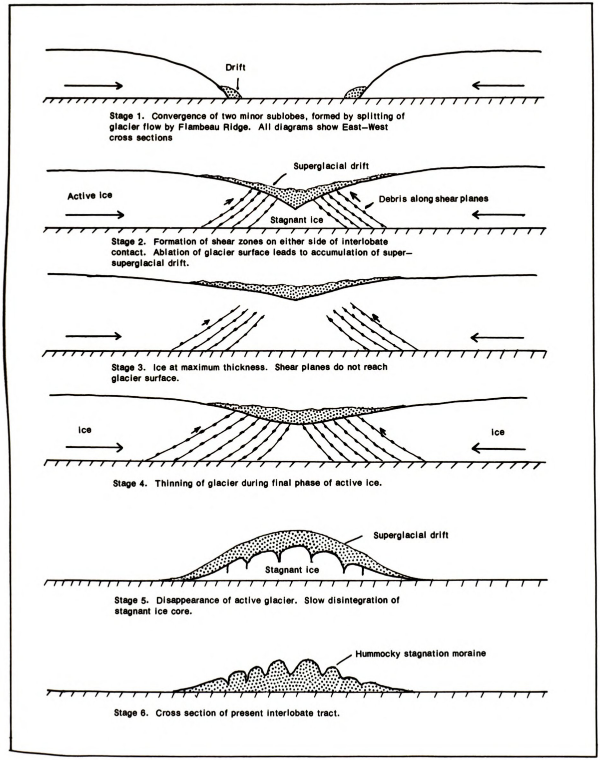

Figure 17. Manner of formation of the interlobate tract. All diagrams represent east-west cross sections of the interlobate tract whose location is shown on Figure 9. The stages shown are hypothetical and adapted from Black (1970, p. 38).

moraine is thought to be of stagnation origin and consists largely of collapsed superglacial debris that had accumulated on the ice in the contact zone between the Green Bay and Lake Michigan lobes. To explain the deposit of superglacial debris within the interlobate zone, Black hypothesizes that shortly after the two lobes had converged, the thicker ice of both lobes, upglacier from the junction, continued to move forward. In Black's view (1970, p. 37)

> ablation . . . at the junction of the two lobes would be countered by ice movement from the base of the ice sheet diagonally upward to the junction at the surface. Upward flow at the terminal zone of glaciers is commonly at angles of 10 to 45 degrees . . . toward the surface to replace ice lost in the ablation zone and to maintain the surface profile of the glacier. (This upward moving ice carried with it debris from the base of the glacier.) When the ice was at its maximum thickness at the junction, the basal debris may not have reached the surface. As the ice thinned during the waning . . . of the glacier, that debris would intersect the surface. . . .
>
> However, at that time the thicker ice back from the junction was continuing to move forward even though the terminal zone was stagnant. The shear planes and flow layers that brought debris up from the base presumably angled downward and away from the actual surface junction of the two lobes. . . . As thinning continued to perhaps 200 to 300 feet of ice, fractures penetrated in favorable places, aided by meltwaters, to the bottom of the ice. In them moulin kames, eskers, and crevasse fills begin to form. . . . Many large fractures were fed not just with running water, but also with mud flows, debris slides and the like. Ponded water in some also trapped deltas and lacustrine sediments. . . . In the interlobate area all the landforms are believed to have resulted from ice stagnation and melting of blocks of appropriate geometry to fit the surface depressions.

The above interpretations of the Kettle Interlobate Moraine seem, at least in part, to explain the array of features and sediments present in the Flambeau Ridge interlobate tract and are herewith proposed as the manner of formation for this section of the Chippewa moraine. Figure 17 illustrates the proposed manner of formation of the Flambeau Ridge interlobate tract.

CHAPTER VI

PROPOSED THEORY OF ICE WASTAGE

Introduction

There can be little doubt that the landforms and sediments of the Chippewa moraine complex, described in Chapters III, IV and VII, are predominantly of ice-stagnation origin, since they are similar or identical to phenomena assigned this mode of formation in several areas of Europe and North America. Some of the more frequently cited studies dealing with ice-stagnation features include Mannerfelt (1945) and Hoppe (1952) in Scandinavia; Niewiarowski (1963) in Poland; Flint (1930) in Ireland; Flint (1928 and 1929), Hartshorn (1958), and Kaye (1960) in New England; Gravenor and Kupsch (1959) and Stalker (1960) in the Canadian Great Plains; and numerous studies in North Dakota, largely summarized by Clayton and Freers (1967). Among the above, Flint's studies are not primarily concerned with hummocky stagnation moraine and are, therefore, not especially helpful in understanding the Chippewa moraine complex.

The remaining studies propose two principal modes of formation of stagnation moraines. One of these is known as the ice-pressing theory, for which Hoppe (1952) and Stalker (1960) have been strong advocates. They and others have suggested that continental glaciers were relatively clean and that stagnation features are largely the result of the squeezing of water-saturated subglacial till up into holes and crevasses in the base of disintegrating stagnant-ice masses. Clayton (1967, p. 36) questioned Stalker's interpretation by noting that

> Wertman (1966) has shown that continental glaciers should theoretically have englacial drift in a basal zone several hundred feet

thick, and Behrendt (1963) has shown that this basal zone in glaciers in Antarctica contains several percent of englacial debris. Flint (1971, p. 171) discounts the appropriateness of ice-pressing for many areas by observing that "whatever its other limitations, ice-pressing seems to be confined to drift rich in clay and to areas of low relief."

The second and most accepted explanation of stagnation moraines has been elaborated by Gravenor and Kupsch (1959), Kaye (1960), Winters (1961), Clayton and Freers (1967), and Parizek (1969). In this mode of formation, ablation drift, which earlier constituted superglacial drift on stagnant ice, was let down as the ice disintegrated--accompanied by repeated slump, debris flow, solifluction, and other processes--and reached the ground after repeated redeposition.

Clayton (1967) describes several lines of evidence which indicate that the stagnant ice of the Missouri Coteau was mantled with thick superglacial drift and that it persisted for several thousand years after the main ice sheet had disappeared from that area; the latter condition is also inferred for Wisconsin by Black and Rubin (1967-68) and for Minnesota by Florin and Wright (1969). Discussion of superglacial till is also contained in Hartshorn (1958) and Boulton (1968), who describe till layers overlying and interbedded with ice-contact-stratified drift which they maintain could only be of "flowtill" origin, derived from water-saturated superglacial till.

The Chippewa moraine complex also contains many instances of flow-till, high-relief ice-walled lake plains, and various stagnation features which can best be explained as having originated from the disintegration of stagnant ice bearing a thick cover of ablation debris.

As stated in the previous chapter the zone of marginal stagnation within the Chippewa lobe intermittently widened. Consequently, the zone of overthrusting underwent an intermittent upglacier migration, thereby producing several semiparallel debris-rich zones in the stagnant ice. Ablation of the stagnant part of the glacier gradually produced a thick blanket of superglacial debris over each of the debris-rich former shearing belts. But the zones of relatively clean ice separating the shear belts did not develop thick superglacial debris blankets and therefore ablated more rapidly. As a result, it is probable that several broad semi-parallel ice-cored moraines formed that were separated by much thinner debris-mantled stagnant ice or by no ice at all (Figure 16).

The interlobate tract, however, constituted an exception to the above pattern during deglaciation, for it consisted of a single three- to five-mile-wide prominent ice-cored morainic ridge, extending about ten miles southwestward from Flambeau Ridge. That is, the greater thickness of the superglacial drift in this area resulted in a significantly greater thickness of stagnant ice than elsewhere within the moraine complex, and probably a different manner of ice disintegration as well.

Thickness of the Stagnant Ice

The thickness of the stagnant ice of the Chippewa moraine complex during deglaciation is not known with any certainty, but the following discussion may provide at least a partial basis for an understanding of this matter. Florin and Wright (1969, p. 700), in a paper on stagnant ice in Minnesota, state that

> buried ice in moraines may be so thick that it is difficult to reconstruct the morphology of ice-cored moraines. For example, the St. Croix moraine, when ice cored, may have stood 600 to 900

> feet above the surrounding terrain after the retreat of the active ice, rather than less than 300 feet, as today. It was at least high enough to confine the southern edge of the Grantsburg sublobe, which advanced over the terrain that was bared by the retreating Superior lobe.

Relative to the ice thickness in existing glaciers, Clayton (1964) states that the stagnant margin of the Martin River Glacier in Alaska contains many funnel-shaped sinkholes that range from 50 to 300 feet in depth. In a later paper Clayton (1967, p. 36) points out that "topographic considerations suggest that the stagnant ice on the Missouri Coteau averaged 300 feet thick. . . ."

The greatest thickness of stagnant ice within the Chippewa moraine was probably in the interlobate tract, for the higher "kames" and ice-walled lake plains in this tract range between 100 to 200 feet of local relief. Except in their margins, the lack of collapse structures in the upper part of these ice-contact features indicates that they were deposited on the ground, no doubt after the stagnant ice was perforated with numerous melt depressions during the final stages of its disintegration. Assuming a considerable amount of thinning before the stagnant ice had reached an advanced stage of disintegration, it seems reasonable to postulate a thickness of 300 to 400 feet during the initial phase of stagnation in the interlobate tract, and it may have exceeded this amount.

Judging from present local relief alone, it would appear that during deglaciation the ice thickness in the ice-cored moraines east of the Chippewa River was much less. In the early phases of stagnation it may have ranged between 100 to 200 feet for the present smaller stagnation moraines and 300 feet or more in thickness in parts of the Pike Lake moraine (see Figure 9). As in the interlobate tract, the thickness of the stagnant ice in the smaller ice-cored moraines east of the Chippewa River was probably much less than the above figures during the later stages of their wastage, when most of the ice-walled lake plains and other stagnation features were formed.

Superglacial Debris Thickness

The thickness of the superglacial drift insulating the buried ice is not precisely known; nevertheless, several lines of evidence permit some estimations. First, the volume of drift comprising the individual stagnation moraines, estimated by considering both the local relief and spacing of the morainic hills, suggests that the superglacial drift blanket, before disintegration, averaged about five to twenty feet thick over the smaller ice-cored moraines to at least a few tens of feet thickness in the interlobate tract; in the latter area it may even have exceeded 100 feet in places. Second, the thick accumulations of lacustrine silts in some of the ice-walled lake plains (75-foot thickness in one power-auger test hole) indicates an extended lifespan for some of these lakes, only possible if the ice enclosing them was well insulated from surface melting by a thick cover of superglacial drift.

The actual thickness of the superglacial drift of the ice-cored moraines may well have exceeded the values given above, if the observations by Clayton (1967) in the Missouri Coteau are applicable here. He assumes that the thickness of the superglacial drift in that case was roughly equal to the local relief of the present dead-ice moraine (hummocky stagnation moraine) of that area. As support for his contention that the dead-ice moraine was derived from superglacial drift, Clayton (ibid., p. 40) describes the debris-mantled, stagnant margins of the Martin River, Bering, and Malaspina Glaciers of Alaska. Along a marginal zone of the Martin River Glacier, the superglacial drift averages between ten and twenty feet in thickness.

Melting Rates and Duration of the Buried Stagnant Ice

Wastage of buried ice proceeds both from the base and from the surface. Basal melting results from the escape of heat from the earth's interior. A considerable amount of sediment may have been released from the base of the stagnant ice by basal melting; perhaps deposition by this process in stagnant-ice areas has not received proper attention. For example, Florin and Wright (1969, p. 701) state that in the Minnesota area about 30 cal/cm^2 of geothermal heat were probably delivered annually to the base of the stagnant ice, enough to melt 0.4 cm of buried ice each year or about 30 feet in 2000 years, provided that the temperature of the ice was close to the melting point. From basal melting alone, therefore, it would take more than 6000 years to melt a 100-foot thick mass of buried ice.

The rate of surface melting of the stagnant ice was controlled by several factors, including the following: (1) thickness of the superglacial debris mantle (discussed at length after this enumeration); (2) degree of crevassing and topography of the ice surface, which influenced the amount of ice-surface area exposed to melting, as well as the ease with which meltwater drained through and out of the stagnant ice; (3) permeability of the superglacial drift, which influenced both the amount of "warm" surface water percolating downward to the ice, assisting in its melting, and the rate of lateral drainage of groundwater (meltwater) through the superglacial drift; (4) topography of the ice-cored moraines, which affected vertical and lateral drainage of the groundwater, of both surface and meltwater origins;* (5) air temperature, which influenced the amount of heat conducted downward to the buried ice for melting purposes; and (6) form, amount, and temperature of the precipitation, which conditioned the amount of water and heat transferred downward to the buried ice from the surface.

Among the foregoing variables, certainly the most important was the superglacial drift thickness. Surface melting in summer is greatly inhibited by a mantle of drift only a few feet thick if the mean annual temperature is close to 0^{o} C (Florin and Wright, 1969). Embleton and King (1968), in describing modern ice-cored moraines in Scandinavia, note that

> experiments were carried out to determine the effect of different thicknesses of dirt on the ablation of the ice. The results showed that when the ablation of clean ice was 4.5 cm/day, the ablation rate decreased when the dirt cover exceeded 0.5 cm. It fell to 3 cm/day with a dirt cover of 6 cm and to less than 1 cm/day when the dirt cover exceeded 20 cm.

The melting rates of thickly-mantled stagnant ice of Pleistocene continental glaciers were apparently much lower than the latter figure, for Clayton (1967, p. 36) points out that on the Coteau of North Dakota "the stagnant ice lasted for at least 3000 years." Radiocarbon dates indicate that most of North Dakota northeast of the Coteau was free of

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*Lateral movement of groundwater in the ice-cored moraines was no doubt relatively rapid during their early phases of wastage, when they had their highest relief. The steep and unstable slopes of such a moraine would provide not only rapid drainage, but would permit slump and flowage of the surface material (when saturated), which enhanced melting by exposing the buried ice to the sun. As melting continued, the thickness of the superglacial debris increased, while the local relief of the superglacial topography decreased; both favored a gradual reduction in the wastage rate.

ice after 12,000 years B.P., yet buried stagnant ice persisted on the Coteau until as recently as 9000 B.P. Clayton concludes by noting that

> using an average of about 300 feet of buried ice, which required about 3000 years to melt, the ice-melt rate was roughly an inch per year. At first the rate was probably much greater when the superglacial drift was thin. Later, as more drift became concentrated on the ice surface, the rate was probably less than an inch per year.

Black and Rubin (1967-68, p. 111) note that radiocarbon dates from basal peat and pond deposits in eastern Wisconsin, south of the terminal position of Valders glaciation, "confirm" that deglaciation of the Woodfordian ice, and hence the beginning of Two Creekan time, must have taken place about 12,000 to 13,000 years ago; they further indicate that these dates do not represent the time of disappearance of active ice from the area but only the dates when organic matter began to accumulate or be preserved in ice-block-basin lakes. Buried ice blocks in that area were believed to have frequently lasted several centuries or more. In the same paper Black and Rubin interpret a radiocarbon date of 7650 years B.P. from the bottom of a kettle lake in northern Wisconsin as follows:

> it is a minimal date for organic accumulation, but time must be allowed for the thaw of the buried ice to produce the kettle. Drumlins made by Valderan ice have been dropped into some lakes by post-Valderan thaw of buried ice of Woodfordian and late Altonian age.

Although no radiocarbon dates are available for the Chippewa moraine Frye, Willman, and Black (1965) indicate that it is at least middle Woodfordian in age. However, this age is now in question for Wright, et al. (1973) have reinterpreted the age of the St. Croix moraine, believed to be a correlative of the Chippewa moraine (Frye, Willman, and Black, op. cit.), as being at least 20,000 years. Florin and Wright (1969), in a paper dealing with the duration of buried ice, state that stagnant ice persisted in the St. Croix moraine for several thousand years after the disappearance of the active ice. For example, they cite (ibid., p. 36) two radiocarbon dates of about 11,800 years B.P. for wood from the base of a dead-ice hollow on the St. Croix moraine of eastern Minnesota, almost directly west of Eau Claire, Wisconsin. These figures imply that melting of the stagnant ice coring the St. Croix moraine may have taken as long as 8,000 years. Because of the uncertainty concerning the age of the Chippewa moraine, no reliable estimate can be given for the maximum possible duration of stagnant ice in the thesis area, except to suggest that at least some of it may have lasted for several thousand years.

The climatic conditions of the Chippewa County area during wastage of the stagnant ice may have been similar to those inferred for eastern Minnesota by Cushing (1967), Wright (1969), and Florin and Wright (1969) on the basis of radiocarbon and paleoecological evidence. Wright (1969, p. 104), in summarizing the late glacial history of Kirchner Marsh on the St. Croix moraine--about 100 miles west-southwest of the area here under consideration--states that tundra conditions lasted from the time of disappearance of the active glacier until about 14,000 years ago. At that time spruce forest was established, which lasted until about 11,000 years ago, when it began, in turn, to be replaced by a broadleaf deciduous forest.

According to Florin and Wright (1969, p. 702), when tundra vegetation prevailed, "the mean annual air temperature must have been at least as low as 23° F, as it is today at the tundra border in northern Manitoba." However, they add that discontinuous permafrost, thermally

analogous to buried glacier ice, is today found where the mean annual temperature is less than 30^{0} to 31^{0}F. This raises the question of how the buried ice was melted to form the ice-block lakes, whose sediments contain the fossil tundra pollen. In Florin and Wright's view (ibid., p. 702),

> one can only conclude that in some circumstances buried ice might not persist even under a mantle of drift and even with below freezing mean annual temperature. Appeal must be made to local surface melting despite the mantle, analogous to the manner of formation of thaw lakes in continuous permafrost in northern Alaska and Yukon. Such local melting can be initiated by accidental breaching of the protective mantle, exposing the ice to the sun.

They further note that during the spruce-dominant period at Kirchner Marsh, roughly 14,000 to 11,000 years ago, the climate appears to have been comparable to that found today in southern Manitoba, where the mean annual temperature is about 36^{0}F.

In view of the modern climatic similarity between eastern Minnesota and the Chippewa County area, it seems reasonable to conclude that during the interval given above climatic conditions similar to those at Kirchner Marsh may also have existed in the Chippewa moraine area.* In regard to this interpretation of the wastage of stagnant ice, a statement by Florin and Wright (1969, p. 699) seems appropriate:

> climate cool enough for buried ice to persist in these regions apparently lasted until the accelerated climatic change that brought about not only the final retreat of the Wisconsin ice from the Great Lakes region at the end of the Valders phase 11,000 years ago, but also the transformation of the boreal spruce forest (which had occupied the ice-cored moraines) to a deciduous or mixed coniferous-deciduous forest.

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*For purposes of comparison with the above temperatures, the present mean annual temperature at Weyerhauser, Wisconsin, located near the northwest corner of the study area, is 42^{0}F (Wis. Crop Reporting Service, 1961).

Theory of Wastage of the Buried Ice

Except for the final stages of ice wastage the topography of the Chippewa ice-cored moraine complex cannot be reconstructed with any high degree of confidence. Very likely the surface configuration underwent repeated changes and frequently topographic reversals as the last vestiges of the glacial ice melted. Important factors affecting topographic development at that time include the amount of superglacial drift thickness and the associated variations in the melting rate of the buried ice. Variations in the superglacial drift thickness probably reflected either or both localized differences in the amount of debris entrained in the stagnant ice or changes in the original thickness of the mantle due to washing, sliding, or flowage of superglacial drift.

These conditions are illustrated by the following hypothetical example. Ice-cored hills in areas of drift-mantled stagnant ice formed where the original drift cover was thickest. At these sites the buried ice melted more slowly than on the surrounding less well-insulated stagnant ice. Eventually, because of less ablation, these places became ice-cored hills, higher in altitude than the surrounding stagnant-ice topography. Once an ice-cored hill formed, its drift cover slid, flowed, or was washed off its slopes. But eventually the thickness of its drift cover became less than that of the surrounding ice. As a result ablation rates on the ice-cored hill became greater than on the adjacent lower ice surfaces, culminating in the destruction of the hill. The ablation rate may have remained greater than normal at this site until the ice surface had melted down well below that of the surrounding ice. This depression may have continued deepening until it evolved into a sinkhole in the stagnant ice, or, once the depression had formed, superglacial drift may have been transferred into it from the enclosing, higher ice surface, which increased the thickness of the insulating layer in the depression and reduced that of the surrounding ice. The locus of maximum downwasting might then have shifted back to the ice enclosing the depression. With this reversal in the relative downwastage rates, the depression might eventually again become an ice-cored hill, starting a second cycle of topographic reversal. In any given part of the stagnant ice such topographic reversals might conceivably have occurred several times during the overall wastage of the ice.

In the initial phase of wastage rapid melting of the stagnant ice and resulting rearrangement of the superglacial debris may have proceeded as a continuum of topographic reversals as described in preceding paragraphs.

During the middle, and longest, stage of wastage the buried ice melted more slowly and the superglacial drift topography was perhaps relatively stable. But in the final stage, after the buried ice had become highly perforated with water-filled ablation hollows and had begun to disintegrate into separate blocks, due in part to more rapid ablation along the original fractures in the glacier, the topography of the moraine complex changed more rapidly. These rapid topographic changes were due to a more active rearrangement of the superglacial drift covering the now highly-roughened, stagnant ice.

The processes by which some stagnant-ice features may have formed have been compared to those that operate in karst landscapes (Clayton, 1964). In describing the drift-mantled, stagnant part of the Martin River Glacier of Alaska, Clayton (1967, p. 42) indicates that the topography is very rugged, with "local relief averaging about 150 to 200 feet, although depressions with as much as 350 feet were observed. The surface is pock-marked by roughly circular sinkholes similar to those found in limestone areas."

Although the removal of ice in the formation of "glacial karst" (Clayton, 1964) is by melting, rather than by chemical action, the resulting features are somewhat similar to those in an area of karst topography. The prerequisites for the formation of glacial karst, according to Clayton, are stagnant ice with a thick insulating cover of ablation drift. These conditions existed in the Chippewa moraine during deglaciation; therefore, glacial karst may also have occurred.

In describing the glacial karst features of the Martin River Glacier, Clayton (1964) points out that the most striking features are funnel-shaped sinkholes. Most, in Clayton's words (ibid., p. 109) "are 200 to 1200 feet across, and there are as many as thirty in a square kilometer. They are 50 to 300 feet deep and many hold small lakes." In Clayton's view, sinkholes in stagnant ice, as in limestone areas, may originate from either "solution" or collapse or a combination of the two. Gravenor (1955) also believed that a stagnant-ice sheet may develop "ice-sinks" when the roofs of ice caves collapse.

Clayton (1964) points out that all stages of the karst cycle are represented on the Martin River Glacier. On the near-inactive ice, immediately upglacier from the stagnant terminal zone, there is much surface drainage, but the streams disappear or "sink" into shaft-like sinkholes (moulins); there are no superglacial lakes in this zone and superglacial till is sparse or absent.

In contrast, the features and conditions in the inner part of the stagnant zone resemble those of the mature karst stage on limestone. The ice is mantled with superglacial drift but is not as thick as in the outer part of the stagnant zone. Drainage is largely englacial or subglacial, ice-caves exist, and funnel-shaped, often water-filled sinkholes are common. The ablation till is very unstable in this area; it is frequently sliding down the steep sides of the numerous sinkholes.

The most advanced stage of glacial karst is found in the outer part of the stagnant zone. Here features that might be considered the equivalent of karst windows, uvalas, poljes, and hums are found. Furthermore, at least some of the streams again have surface routes, due to a more advanced stage of ice disintegration. The ice-walled lakes of this zone are underlain by solid ground rather than ice and are insulated from adjacent ice by ablation till; they may therefore be relatively clear, not frigid, and may even contain freshwater plants and animals. In contrast, the lakes in the inner part of the stagnant zone have unstable shorelines and therefore are turbid, and they are relatively cold because of contact with the enclosing ice or their small size; due to these conditions they are barren of life. Clayton (op. cit., p. 110) further states that where the superglacial till on the Martin River Glacier is at least five feet thick and stable, it supports a mature spruce-hemlock forest.

Gravenor and Kupsch (1959) and Clayton (1967) invoke the concepts of glacial karst in interpreting certain of the stagnant-ice phenomena of the northern Great Plains. For example, Gravenor and Kupsch (op. cit., p. 52-53), in describing the stagnation features of the Canadian

Great Plains, believe that where the superglacial till was thick, had a high silt and clay content, and was saturated, it flowed into the sinkholes in the stagnant ice and, when the enclosing ice had melted, formed the "prairie mounds" and "circular disintegration ridges" (see Chapter VII).

A form of ice wastage somewhat related to Clayton's glacial karst model was proposed by Hoppe (1952). He described examples of a type of stagnant-ice landform composed mainly of stratified drift and termed "sediment plateau," which he interprets as having been deposited subglacially, by meltwater, in large cavities in the base of stagnant ice. The plateaus, though composed primarily of sand and gravel, are veneered with a thin cover of till. While Hoppe does not rule out the possibility that this till veneer represents ablation debris derived from overlying or adjacent "dirty" ice, he asserts that the stratified drift "must have been deposited subglacially" (op. cit., p. 47).

According to Flint (1971, p. 211), the widespread existence of eskers and moulin kames in areas of former stagnant-ice sheets appears to establish that subglacial deposition of stratified drift by meltwater was important, because most are believed to be basal features. But that relatively large hollows were ablated in the base of stagnant-ice masses and filled with meltwater deposits is less well established. Hoppe's (1952) interpretation that some stratified-drift features in stagnation moraines (excluding eskers and moulin kames) are the result of subglacial sedimentation may be correct. In the Chippewa moraine complex, however, the morphological and sedimentological characteristics of at least the larger ice-contact-stratified drift features do not indicate that they were formed subglacially. Yet this process of ice wastage cannot be ruled out as having operated, as shown by Florin and Wright's (1969) data on the probable basal melting rate of stagnant ice by geothermal heat.

The Chippewa moraine complex contains a great variety of features composed of ice-contact-stratified drift, but the greater part of the complex consists of hummocky stagnation moraine. The dominant depositional process in the formation of this landform type was apparently flowage and sliding. Within the Missouri Coteau of North Dakota the "evidence for this comes in part from an analogy with the collapsed superglacial lake sediments that are associated with dead-ice moraine; both have similar topography and are thought to be of similar origin" (Clayton, 1967, p. 29). In that area the contortion and disruption of the bedding of collapsed lake sediment indicate that sliding and flowage were the dominant processes during collapse of superglacial lake sediments. Within the Chippewa moraine complex numerous exposures reveal one or more layers of flow till, interbedded with stratified deposits, indicating that during its disintegration the stagnant ice was mantled, at least in part, by superglacial till.

It has been established that the thickness of the superglacial drift influenced the rate of melting of the buried ice, which in turn controlled the amount of water in the superglacial till. This in turn affected the flowage tendency of the superglacial till, which controlled the final slope angles after disintegration. Clayton (1967) cites work done in Saskatchewan which suggests that in any given area of dead-ice moraine there exists, within narrow limits, a characteristic maximum slope angle. This slope angle approaches the steepest equilibrium slope that existed during the period of collapse.

In a discussion of the glacial history and features of the Missouri Coteau, Clayton (1967, p. 30) states that in those "parts of the Coteau where the maximum slope angles in dead-ice moraine are low (four degrees or less) the superglacial drift was relatively thin." The local relief of such areas is also low and the ice-walled lake sediments are largely sands and gravels, indicating rapid deposition; the superglacial environment in these areas was <u>unstable</u>. Where maximum slope angles in dead-ice moraine (hummocky stagnation moraine) are steep, the superglacial drift was thick. In these areas of the Coteau the local relief is high, the ice-walled lake sediments indicate slow deposition, and the superglacial environment was <u>stable</u>.

The interlobate tract of the thesis area, characterized by high relief, steep slopes, and thick ice-walled lake silts, represents the best example of a stable depositional environment. Unstable depositional conditions during wastage of the buried ice are probably best exemplified by the axial zone, immediately east of the Chippewa River; here stagnation moraine commonly exhibits low relief, gentle to moderate slope angles, and few good examples of ice-walled lake plains, especially those containing thick silt accumulations, indicating slow deposition.

The influence of crevasse and fracture patterns in glaciers on the landform shapes that develop during wastage of stagnant ice have been discussed by Hoppe (1952), Gravenor and Kupsch (1959), and Stalker (1960). In a discussion of the stagnation features of western Canada, Gravenor and Kupsch (op. cit.) interpret the hummocky moraine of that area as resulting from the slow disintegration of drift-mantled glacial ice. The term "disintegration" is used to describe the manner of wastage of the ice. It appears that, as the buried ice melted, it must have separated into a number of smaller blocks.

The melting of a stagnant glacier and its gradual separation into detached ice masses, according to Gravenor and Kupsch, may take place under uncontrolled conditions, in which case there is no evidence of alignment of the resulting landforms; on the other hand, the stagnant ice may break up, or disintegrate, along aligned weaknesses within the ice, where melting is more rapid, and give a pattern to the resulting deposits. This gives rise to the controlled type of disintegration. The pattern that results from this manner of disintegration is inherited from a more active phase of glacier motion, when the ice developed a pattern of shear planes, fractures, and crevasses associated with its movement. In controlled disintegration the linear zones along which the decaying ice separated are normally aligned parallel or perpendicular to the direction of ice movement; in some cases, however, they may be aligned at near 45-degree angles to the ice movement (Gravenor and Kupsch, 1959, p. 53).

As described by Embleton and King (1968, p. 396), "where the disintegration was uncontrolled, there is no pattern apparent in the topography. Ridges, hollows, and mounds are distributed indiscriminantly." Most of the Chippewa moraine exhibits the foregoing conditions; thus, disintegration appears to have been largely uncontrolled.

Controlled-disintegration landforms are widespread in the northern Great Plains, in the view of Gravenor and Kupsch (1959). Landforms of this type are composed chiefly of till with a thin covering of gravel

on the till ridges, but some are of stratified material. Where the major fractures in the stagnant ice (inherited from live ice) were parallel, the resulting disintegration ridges and hollows show alignment in only one direction; but where both radial (or longitudinal) and transverse crevasses were present, the topography after disintegration consists of a rectangular pattern of ridges and hollows.

In the Chippewa moraine complex the controlled variety of disintegration appears to have been relatively uncommon; it is best developed in the interlobate tract, where there is sometimes a semirectangular or parallel system of disintegration ridges and dead-ice hollows.

The relative size and spacing of the dead-ice knobs and hollows provides a clue to the manner of disintegration of the stagnant ice. In Clayton's (1967, p. 31) words, "a depression in dead-ice moraine was a hill of stagnant ice and a hill in dead-ice moraine was an ice depression during the last phase of melting of the stagnant ice." He adds that "the topographic density of the stagnating ice must, therefore, have been the same as that of the present dead-ice moraine."

Within the Chippewa moraine complex the small size and close spacing of the knobs and kettles cause areas of hummocky stagnation moraine to have relatively high topographic densities. During disintegration, the ablation depressions in the stagnant ice probably did not extend much below a depth of 150 to 200 feet, because ice of temperate glaciers tends to become plastic at greater thicknesses (Flint, 1971, p. 36). As envisioned by Clayton (1967, p. 31), "if maximum probable ice slopes were about 40 degrees, and maximum depression depths were 200 feet, the average ice depression would have been about 500 feet in diameter." Assuming that the stagnant ice was thoroughly pitted and

furrowed with depressions of this size and spacing, the topographic features in the resultant stagnation moraine should have similar dimensions.

The dimensions of the till knobs and intervening dead-ice hollows most nearly approach the above figures in the interlobate tract and adjoining areas, and again in the better developed stagnation moraines east of the Chippewa River. For example, at the southwest end of the interlobate tract and in the area immediately northwest of it until Long Lake--or T31N, R8W and T32N, R8W (see Figure 45)--there are several dozen small, typically elongate, kettle or dead-ice hollow lakes. With few exceptions their widths range between 300 to 1000 feet and they are separated by knobs or ridges of similar width. East of the Chippewa River, where the thickness of the stagnant ice during disintegration appears to have been thinner, the size and spacing of the knobs and hollows is often less than that in the interlobate tract, which suggests that the topographic density of pits and furrows on the stagnant ice was influenced by its thickness.

In summary, where knob-and-kettle topography is well expressed, the relative uniformity in the size of the till knobs and intervening dead-ice hollows suggests that the stagnant ice disintegrated into blocks of roughly uniform size, the latter somewhat dependent on the thickness of the ice during its final disintegration.

Bottom-altitudes of the dead-ice hollows within the moraines, relative to the altitudes of the "lowlands" adjoining the moraines, provide some insight into the manner of deposition of the moraines of the Chippewa moraine complex. The bottoms of the deeper or larger dead-ice depressions are often similar to, or only slightly higher than, the altitudes of the bordering lowlands, and in some cases even lower. This condition holds true whether the bordering lowlands are underlain by outwash or ground moraine. Certainly such conditions could not exist if these were true moraines.

There are at least two possible explanations that may account for such circumstances. If the material comprising the knobs and ridges consists largely of collapsed superglacial till deposited between blocks of a disintegrating stagnant-ice mass, as appears to be the case, then the depressions are simply sites of non-deposition and their bottoms should have altitudes similar to those of adjoining lowlands, where little or no superglacial drift was deposited. An alternate explanation, preferred by Hoppe (1952) and Stalker (1960), is that after the glacier had stagnated and had begun to disintegrate into blocks, water-saturated, highly plastic till underlying individual ice blocks was squeezed or pressed from beneath them into the intervening ice-free spaces. After the ice had melted, till "squeezed" from beneath the ice blocks became the present till ridges.

Although the ice-pressing mechanism may have operated during the formation of the Chippewa moraine, major arguments against its being of more than minor importance are the high local relief of the till ridges in the interlobate tract, the belted pattern of the stagnation moraines east of the Chippewa River, and the fact that these moraines rise well above the lowlands separating them, as well as those adjoining them.

The significance of these conditions is as follows: Stalker (1960, p. 21) notes that "under favorable conditions with ice 200 feet thick,

assuming an ice density of 0.8 and a till density of 2.0, a ridge of about 80 feet high could be constructed." Since the local relief of some of the till ridges in the interlobate tract exceeds 150 feet, it hardly seems possible that they are entirely of ice-press origin. Furthermore, it is improbable that the ice-pressing mechanism could have been restricted to distinct belts, with no ice-pressing beneath the stagnant ice separating these belts. Finally, the drift is significantly thicker and the average altitude of the land surface higher within the zones of hummocky topography than in the intervening flatter areas. Such conditions are not explainable by the ice-pressing process, but understandable if (1) the hills in these areas of hummocky stagnation moraine are interpreted as composed of collapsed superglacial till and (2) that before disintegration the superglacial drift had been mainly restricted to those areas where hummocky stagnation moraine exists today.

Drainage Conditions Prior to and After Disintegration of the Buried Ice

Surface drainage conditions in the moraine area before the melting of the buried ice, must have been different than those of the present. The modern surface drainage is essentially nonintegrated, with the topography consisting of many small hills and intervening depressions. But before wastage of the buried ice, the superglacial drainage was partially integrated, with more surface streams, fewer undrained depressions, and very likely a lower topographic density than at present.

These interpretations are based on several lines of evidence, the most convincing of which consists of numerous short remnants of superglacial stream channels, preserved only in those places where the

channels were not underlain by stagnant ice. Where the superglacial streams drained across drift-mantled ice, their channels were destroyed with disintegration of the buried ice. Preserved remnants of "valleys" cut by superglacial streams are most commonly situated on ice-walled lake plains and are found at many locations throughout the Chippewa moraine complex. These channel remnants probably originated in the following manner. The streams which drained into and out of the ice-walled lakes were superposed on their drained floors, and had begun to cut valleys into them, when the surface of the stagnant ice had down-wasted sufficiently to permit drainage or destruction of the lakes. But soon thereafter the superglacial streams became extinct, apparently because they could not maintain themselves across the superglacial topography, which became increasingly rougher as disintegration progressed.

The largest preserved valley remnant of this type is about twenty-five feet deep, is nearly one mile in length, and has a meandering course across the top of a perched lake plain in sec. 27, T32N, R8W. The ubiquitous distribution of this type of feature indicates that most of the larger ice-walled lakes had streams draining into and out of them, unlike many of the modern lakes of this area.

The presence of fish in landlocked lakes is another circumstance indicating better drainage integration during the stagnant-ice period than at present. Most of the lakes without outlets contained fish when this area was first settled by Europeans. Assuming that these land-locked lakes were not stocked with fish by Indians prior to European settlement, their presence may indicate that at one time these lakes, or their ancestors, had outlets up which fish must have migrated from the

Chippewa River and points south. With few exceptions these small lakes show no topographic signs of having been connected in the past.

The above condition may indicate that during the closing stages of ice wastage, and after the climate had warmed considerably, the superglacial topography held many relatively warm lakes, connected by superglacial streams. Fish swam up these superglacial streams to occupy the lakes. The gradual disintegration of the buried ice led to the destruction of the ice-walled lakes, formation of many new lakes occupying the depressions left by the last blocks of ice to melt, and disintegration of the drainage pattern; the latter event led to the landlocked character of many of the present lakes.

On the basis of the above suppositions, the ice-walled lake sediments of the Chippewa moraine are presumed to contain fossils of aquatic life forms; the close similarity between the ice-walled lake plains and other stagnation features of this area with those on the Missouri Coteau, where ice-walled lake sediments commonly contain fish and mollusk fossils (Clayton and Freers, 1967), supports that hypothesis. However, the hypothesis remains as yet unverified by field work.

Collapsed superglacial alluvium, common on the Missouri Coteau (Clayton and Freers, 1967), may also be used to reconstruct superglacial drainage conditions. This type of sediment appears to be scarce in the Chippewa moraine, with the explanation for that postulated condition unknown. Whether the apparent scarcity of this material means that the Chippewa moraine carried fewer superglacial streams than indicated for the Coteau also remains unresolved. Possibly, the narrow width of the individual ice-cored moraines comprising the Chippewa moraine complex precluded the formation of superglacial streams large enough to carry the quantities of alluvium deposited by those draining across the much broader stagnant-ice zone of the Coteau.

Despite the apparent scarcity of superglacial stream sediments the presence of several hundred well-developed examples of moderate- to high-relief ice-walled lake plains and ice-walled gravel trains (see Figure 45) indicates that a substantial sediment load was carried by superglacial streams. Perhaps further investigations will reveal that collapsed alluvium is more common than present knowledge suggests.

In summary, several types of phenomena infer better drainage integration and more surface runoff in the superglacial environment than currently exists. The disintegration of the stagnant ice appears to have destroyed a well-developed superglacial drainage pattern, replacing it with a maze of small, undrained depressions and nonintegrated drainage, almost unchanged since deglaciation.

CHAPTER VII

STAGNANT-ICE LANDFORMS OF THE CHIPPEWA MORAINE COMPLEX

Introduction

The Chippewa moraine complex, largely of ice-stagnation origin, contains a great variety of stagnant-ice landforms and features. Collectively, these features comprise orderly landform assemblages, with each landform type occupying a characteristic location relative to the other types.

Stagnant-Ice Landform Types

Individual landforms were classified mainly on the basis of three criteria: (1) morphology, (2) lithology and internal structure, and (3) apparent mode of formation (genesis or origin). On the basis of these criteria more than twenty different types of stagnant-ice landforms were identified. They are not, however, all distinctly different from each other. Rather, the various types comprise several gradational series, and the series, in turn, may comprise a continuum. Some landforms consist of only one type of sediment, deposited by a single depositional process, while others are more complex.

In the following discussion frequent reference will be made to Figure 45 (end pocket), which shows the location of several hundred stagnant-ice landforms. It is recommended that the reader also refer to Figures 9 and 41 when considering landform distributions.

Hummocky stagnation moraine

Hummocky stagnation moraine (Winters, 1963, p. 39) is the most representative and extensive landform type in the Chippewa moraine, and may account for 75 per cent or more of the morainic topography in the thesis area. Gravenor and Kupsch (1959, p. 50) have characterized this type of topography as consisting of "a non-descript jumble of knolls and mounds of glacial debris, separated by irregular depressions. the knolls do not align into ridges and no dominant trends are discernible." They further note that "knob and kettle" topography of this type (Figure 18) is diagnostic of moraines of stagnant-ice origin.

Previously (Chapter VI) it has been established that the ridges and mounds in hummocky stagnation moraine are predominantly composed of collapsed superglacial till deposited during disintegration of stagnant ice and that the local relief and slope angles of the knobs reflect the relative instability of the superglacial environment. According to Clayton (1967, p. 30), the superglacial drift reached the ground primarily by gravitative transfer of various types during disintegration of the stagnant ice.

The ridges and depressions of hummocky stagnation moraine, although varying greatly in size, are commonly closely spaced, averaging a few hundred feet in diameter (Figure 19). In overall arrangement the ridges function as partitions that divide the topography into numerous distinct, frequently closed depressions. As a result, the surface drainage for a large part of the moraine complex is nonintegrated. East of the Chippewa River remarkably few of these depressions contain ponds, yet most have swampy floors. West of the river, in striking contrast, the depressions average greater depth and many contain ponds (Figure 20).

Figure 18. Representative hummocky stagnation moraine in the W1/2, sec. 32, T32N, R8W (southwestern part of the interlobate tract).

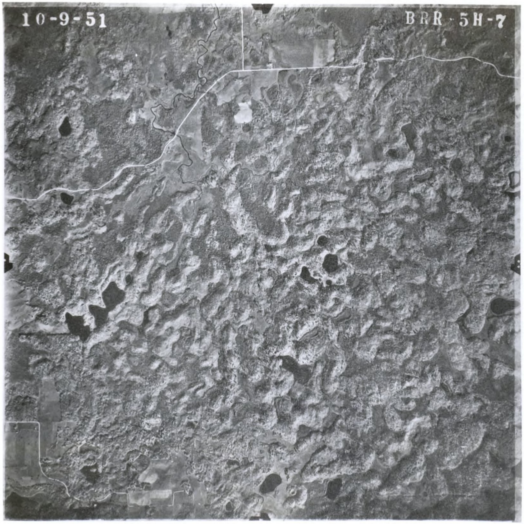

Figure 19. Air photo showing well-developed hummocky stagnation moraine. Location: western margin of the interlobate tract at the south end of Mud Creek Lowland (see Figure 6). Photo: United States Department of Agriculture - Agricultural Stabilization and Conservation Service.

Figure 20. Pond in a dead-ice hollow (depression in hummocky stagnation moraine) in sec. 32, T32N, R8W.

The ridges, composed mainly of till, are sometimes capped by variably sorted sand and gravel, and occasionally contain pockets or lenses of sand and gravel.

Dead-ice hollow

The often deep, closed depressions, so characteristic of hummocky stagnation moraine, have been given various names. Hoppe (1952, p. 5) preferred the designation "dead-ice hollow" and Black (1974, p. 152) referred to them as "swales." The author prefers the term proposed by Hoppe (op. cit.). In his words, "the depressions in dead-ice moraine [termed "hummocky stagnation moraine" in this study] may be considered as places where the last remnants of the ice melted. In other words, they are dead-ice hollows" (Figure 20).

Depressions of this type vary from a few to a hundred feet or more in depth. They are best developed in the interlobate tract, where they typically exhibit local reliefs of thirty to fifty feet, but with some exceeding 100 feet. The diameter of dead-ice hollows may vary from as low as a few tens of feet to more than a mile. The great majority, however, have diameters of less than 1000 feet (see Figure 19). Their shapes also vary considerably. Most are slightly elongated but display irregular outlines. Except in the interlobate tract there is no discernible preferred orientation.

Ice-contact slope

The typical ice-contact slope consists of a steep slope that represents the surface of a mass of sediment deposited on a firm ground surface, but against a steep supporting wall of glacier ice. When the ice melted, the drift banked against it slumped. Consequently, the

sediments making up such slopes typically stand near the angle of repose. Collapse structures commonly observed in the associated stratified sediments provide evidence that slumping took place during the formation of ice-contact slopes.

Ice-contact slopes are often conspicuous and useful in the interpretation of other stagnant-ice landforms. For example, many of the landforms shown on Figure 45 (end pocket) were first identified by their ice-contact slopes (see Figure 21).

Ice-walled lake plain (perched-type lake plain)

An ice-walled lake plain (Clayton, 1962, p. 39) is an elevated area of smooth and nearly flat topography underlain (in the ideal case) by horizontally bedded clays, silts, sands, and gravels (and, in some cases, till near the margin) that were deposited in a glacial lake that was restricted at least in part by stagnant ice. Ice-walled lake plains (perched lake plains) are some of the most characteristic features of the Chippewa moraine, as well as of glacial stagnation in general.

Marked variations exist among the perched-type ice-walled lake plains of the study area. They do, however, share several common characteristics. They usually are (1) subcircular, (2) concave upward in profile, (3) mesa- or plateau-like in form, (4) bounded by ice-contact slopes, (5) rimmed by low ridges, and (6) topographically higher than the intervening areas of hummocky stagnation moraine. Figures 21, 22, 23, and 45 demonstrate these aspects.

A small number of lake plains display forms suggesting two or more stages in their development. Figure 26 illustrates this situation. Additional examples of multistage lake plains include several of the high-relief lake plains of the interlobate tract, which display

Figure 21. View toward the ice-contact slope on the southwest margin of the ice-walled lake plain in secs. 21, 22, 27 and 28, T30N, R5W. The surface of the lake plain extends into the distance (toward the upper right) behind the top of the ice-contact slope.

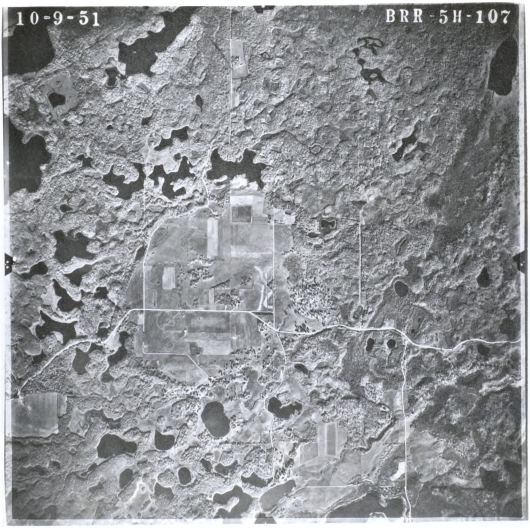

Figure 22. Air photo showing an ice-walled lake plain surrounded by hummocky stagnation moraine. County Highway M crosses the large ice-walled lake plain in the center of the photo in secs. 21 and 28, T32N, R8W. Another perched lake plain is situated just to the right of the north-south centerline at the bottom of the photo. The meandering linear feature on the surface of this lake plain is an abandoned valley, formed before disintegration of the buried stagnant ice (Figures 29 and 30 concern this same feature). Photo: United States Department of Agriculture - Agricultural Stabilization and Conservation Service.

Figure 23. Rim and ice-contact slope of ice-walled lake plain in sec. 15, T30N, R8W, two and one-half miles northeast of Eagleton (see Figure 45). The trees in the background are located on hummocky stagnation moraine. The surface of the perched plain occupies the foreground. The cornfield in the middle distance is located at the head of an indentation in the lake plain margin.

Figure 24. Eastward view along the southeast margin of the Boot Lake ice-walled lake plain in secs. 10 and 11, T30N, R8W. Buildings in background are situated on the rim of the lake plain, whose surface slopes toward the left. An ice-contact slope bounds the right side of the plain. The mound on the right margin is part of the hummocky stagnation moraine surrounding the lake plain (Figure 25 illustrates this relationship).

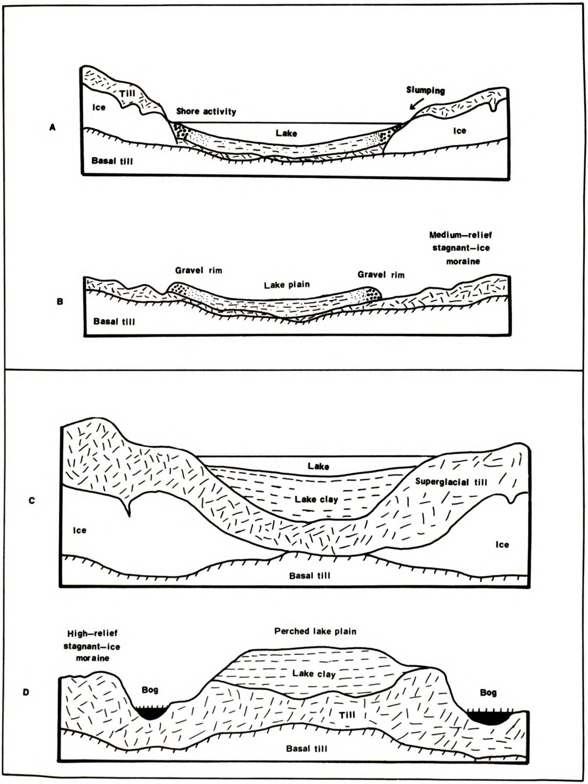

Figure 25. Formation and characteristics of ice-walled lake plains. A and B show lake plains formed in unstable depositional environments. C and D show lake plains formed in stable depositional environments. Source: Clayton (1967, Figure B-9).

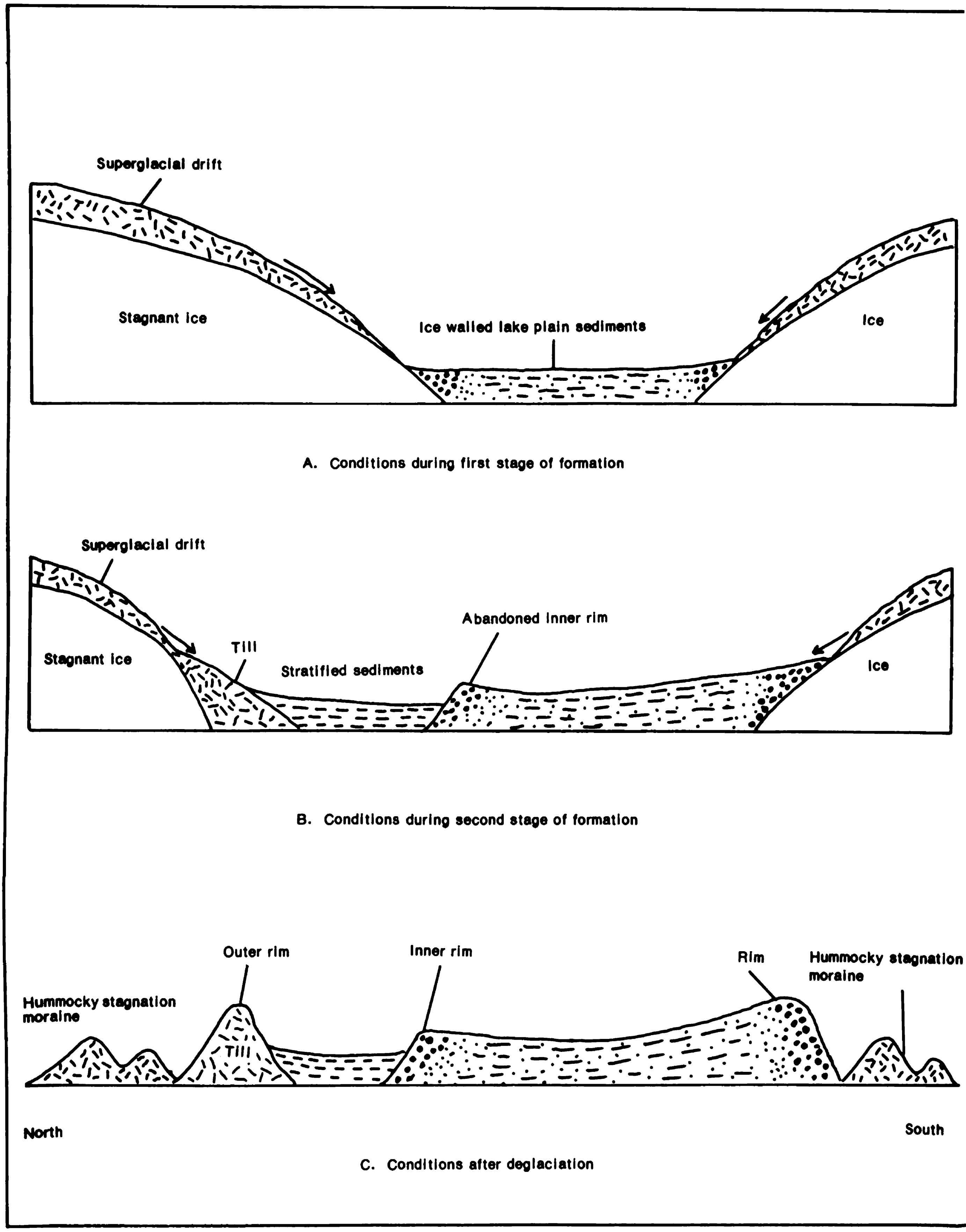

Figure 26. Manner of formation of the multistage western end of the ice-walled lake plain in sec. 35, T31N, R8W. All diagrams show north-south cross sections.

Figure 27. Deltaic bedding exposed in a roadcut through the western margin of the ice-walled lake plain in secs. 7 and 18, T32N, R7W. The ice-contact slope of this lake plain faces toward the background of the photo. Figure 28 represents a closeup view of this same exposure.

Figure 28. Closeup view of roadcut exposed in Figure 27. The bedding of the sediments was exposed by shoveling, the effects of which are visible just above the car. Dip of the bedding is toward the right.

Figure 29. Abandoned valley entrenched into the surface of the ice walled lake plain in sec. 27, T32N, R8W. This valley was cut by an east ward-draining stream (toward the background of the photo) after the ice-walled lake had drained, but before the stagnant ice surrounding the lak sediments had fully disintegrated (see Figures 22 and 45 for further information concerning the relative location of this feature).

Figure 30. Hanging-valley relationship of the abandoned valley described on Figure 29. The car is situated at the upstream end of the valley-remnant crossing this perched plain. The lake occupies a depression in the hummocky stagnation moraine bordering the lake plain.

conspicuous marginal terraces, situated as much as several tens of feet below the summit plain. These terraces are a variety of kame terrace, and probably consist of sediments deposited in a moat separating the receding ice wall from the highest part of the perched plain. As described on preceding pages, many of the ice-walled lake plains are incised by small valleys, five to twenty-five feet in depth, that were formed by streams flowing across them immediately after they had been drained but before the enclosing ice had melted below the surface of the plains (Figure 29).

<u>Intermorainal ice-walled lake plain (lowland-type lake plain)</u>

Lake plains of this type formed in the depressions separating the individual stagnant-ice moraines (Figure 45). During their formation they were ice-walled, at least in part. They are rimmed by low ridges and therefore tend to have a dished appearance. Small tracts of hummocky stagnation moraine are enclosed by and rise above the surface of some intermorainal ice-walled lake plains, suggesting that in the subsurface features of this type are underlain by hummocky stagnation moraine.

<u>Incipient ice-walled lake plain</u>

Features of this type (see Figure 41) resemble ice-walled or perched lake plains in many ways, with a number of significant exceptions. Their overall shapes, rather than being flat or dished, might be more aptly described as resembling rimmed, semicircular ramps, with their rims breached on one side. These breaches are not the result of stream erosion, but gaps of nondeposition. The floor of the central depression,

Figure 31. Southward-looking view of Baldy Mountain, an example of a high-relief, stable-environment ice-walled lake plain. The altitude of the summit of this feature, located in sec. 5, T31N, R7W and sec. 32, T32N, R7W, is about 185 feet higher than the lakes in its immediate vicinity.

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The term "incipient" lake plain is considered appropriate for features of this type, because they represent lake-plain features whose development was terminated before they had attained the characteristics of normal lake plains, or they may represent lake plains which formed under less than ideal conditions. For example, where the ice on one side of an ice-walled lake was very thin, the ice surface at that point may not have been high enough to supply sediments to the lake; therefore, no rim developed at that point.

Transmorainal meltwater depressions

West of the Chippewa River the Chippewa moraine complex is traversed by two large linear depressions (Figure 42), neither of which serves as a routeway for a modern stream crossing the moraine nor displays the typical morphology of major glacial spillways.* Nevertheless, meltwater appears to have played a major role in their formation.

Excluding the ice-block basins which pit their surfaces, these gaps in the moraine are believed to represent places where meltwater, draining across the ice-cored moraine from the active glacier situated on its

*The individual transmorainal lowlands are discussed in Chapter VIII, under the heading of their respective physiographic districts (Figure 42).

proximal side, removed much or all of the superglacial debris. At the time they were active, the floors of these spillways presumably stood several tens-of-feet higher than the present altitudes of the same localities. Eventual disintegration of the underlying ice lead to the destruction of the original form of the superglacial spillways, as discrete morphologic features, but their former existence is still vaguely apparent.

Branching ice-walled alluvial-lacustrine train (lowland-type train)

Branching ice-walled alluvial-lacustrine trains are found principally in the area between the interlobate tract and the Chippewa River and again in the lake district located northwest of the interlobate tract (Figure 45). Linear branching plains of this type are underlain by alluvial (including both glacial and nonglacial stream sediments) and/or lacustrine sediments and possess asymmetrical rims composed of till or stratified sediments, bearing ice-contact slopes on their outer sides. Ideally, they are bordered by hummocky stagnation moraine on all sides, except for those places where streams drained from them during their formation. Some of these lowlands are occupied by modern streams.

The ice underlying these lowlands melted earlier than in adjoining areas, probably because the superglacial streams following these routes had removed much of the insulative superglacial drift. When the lowlands became ice-free, they were occupied by streams, or by lakes connected by streams. Sediments were then deposited in them, derived from the adjoining drift-mantled stagnant ice. Lacustrine sediments were deposited where the lowlands held ponded water; elsewhere, "alluvium" was deposited,

except at some places in the rims where till was deposited by mass wastage.

The streams occupying the trunk lowlands apparently had tributaries, and ice-walled tributary lowlands were melted out along them; within the latter waterlaid sediments were also deposited. In places, the lowlands were underlain by thin stagnant ice, until after active sedimentation had ceased. As a result, the floors of the ice-walled alluvial-lacustrine lowlands are sometimes pitted. Due to the complex character of these lowlands and their surroundings, it is in most cases impossible to reconstruct the drainage conditions associated with them at the time of their formation.

Ice-walled outwash plain

Clayton (1962, p. 39) states that "ice-walled outwash plains are similar to ordinary outwash plains but are more than half surrounded by an outward-facing ice-contact face." Features of this type consist of outwash that was deposited in large indentations in the ice margin and in some cases display rimmed margins. They are not common in the study area.

Linear disintegration ridge

In this study a linear disintegration ridge is defined as a narrow, commonly steep-sided, distinctly linear ridge, usually less than one mile in length and consisting of till, sorted drift, or a combination of both types; eskers are not included. As proposed by Gravenor and Kupsch (1959, p. 54), ridges of this type are composed of material that was deposited during the disintegration of stagnant or near-stagnant glacier

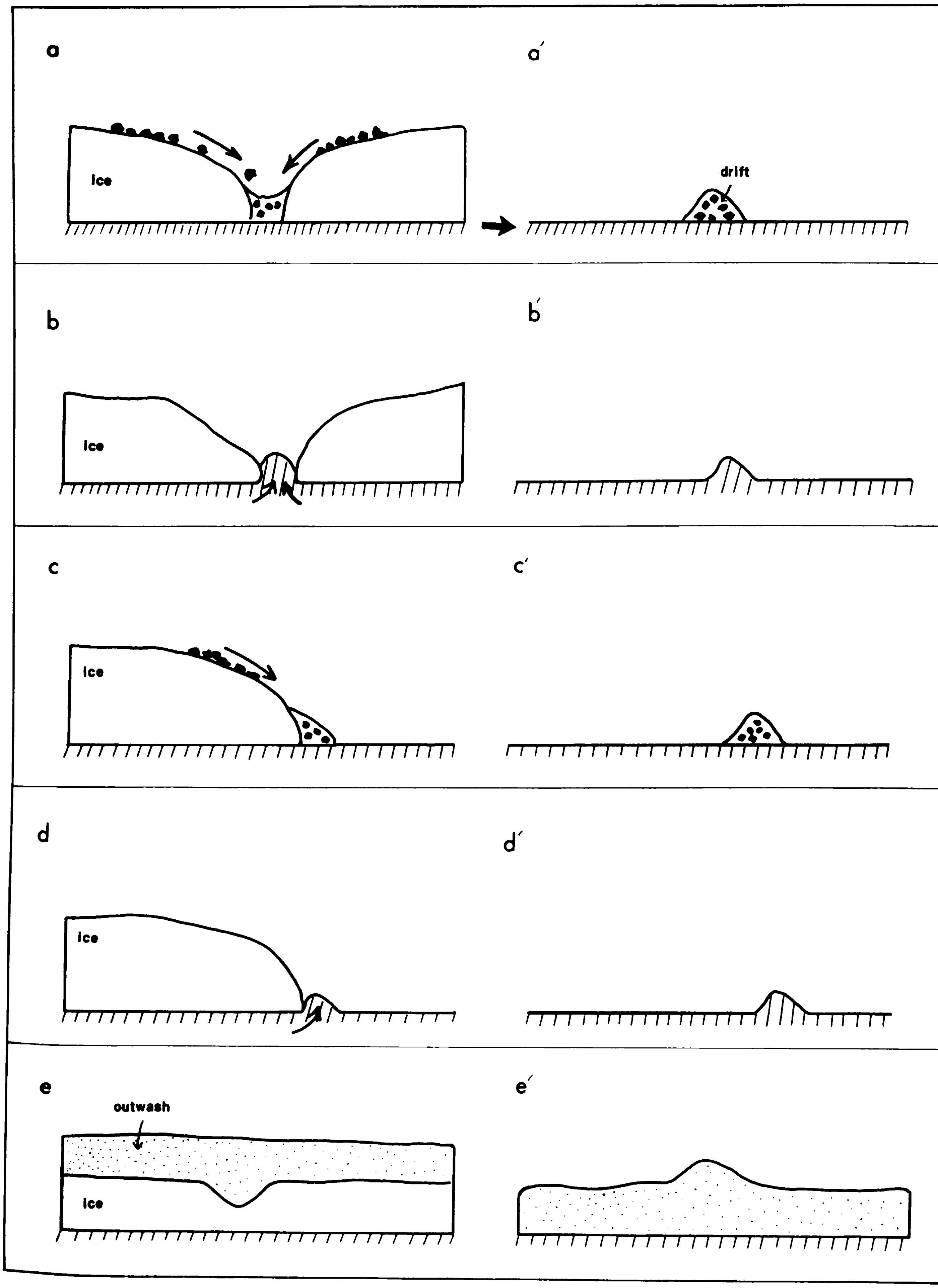

Figure 32. Five modes of forming linear disintegration ridges. All drawings are diagrammatic cross sections. Source: Clayton (1967, Figure A-1).

Figure 33. Linear disintegration ridge in sec. 17, T30N, R7W. This feature has about fifteen feet of local relief and appears to consist of till. The even skyline represents a perched lake plain with a surface about fifteen feet higher than the ridge in the foreground.

ice. Clayton (1967, Figure A-1) described six ways by which linear disintegration ridges may be formed. These are shown on Figure 32.

The linear disintegration ridges of the study area are clearly members of a gradational series. They show similarities to, and may grade into, almost any of the other disintegration features, except perhaps some of the larger non-linear features, such as ice-walled lake plains. In total, about forty-five to fifty disintegration ridges were identified within the thesis area.

Circular disintegration ridge

The circular disintegration ridges of the Chippewa moraine complex form nearly circular, mesa-like hills, with a central depression that may vary in size. The outward-facing surfaces of the circular ridges are steep and appear to be ice-contact slopes, but they possess gently- to moderately-inclined inner slopes that grade into the floors of the central depressions. The altitude of these central depressions is distinctly higher than the ground surface in the dead-ice depressions adjacent to the circular disintegration ridges. These characteristics are illustrated by Figures 41, 34, and 35.

Gravenor (1955) was one of the first to describe the features herein designated circular disintegration ridges, although he termed them "closed disintegration ridges". Clayton (1967, p. 32), employing many of Gravenor's ideas, described the steps in the formation of these features (see Figure 35) as follows:

> circular disintegration ridges formed by (a) sliding or flowing of superglacial till . . . into a sinkhole in the stagnant ice, (b) inversion of topography as a result of the insulating effect of the drift in the bottom of the sinkhole, followed by mass movement of this drift away from the center and down the sides of the buried ice core, and (c) melting of the ice core.

Figure 34. View of a roadcut through a circular disintegration ridge. This feature is located along Highway G (six miles north-northwest of Stanley), on the south margin of the moraine. Low-relief hummocky stagnation moraine is visible in the background.

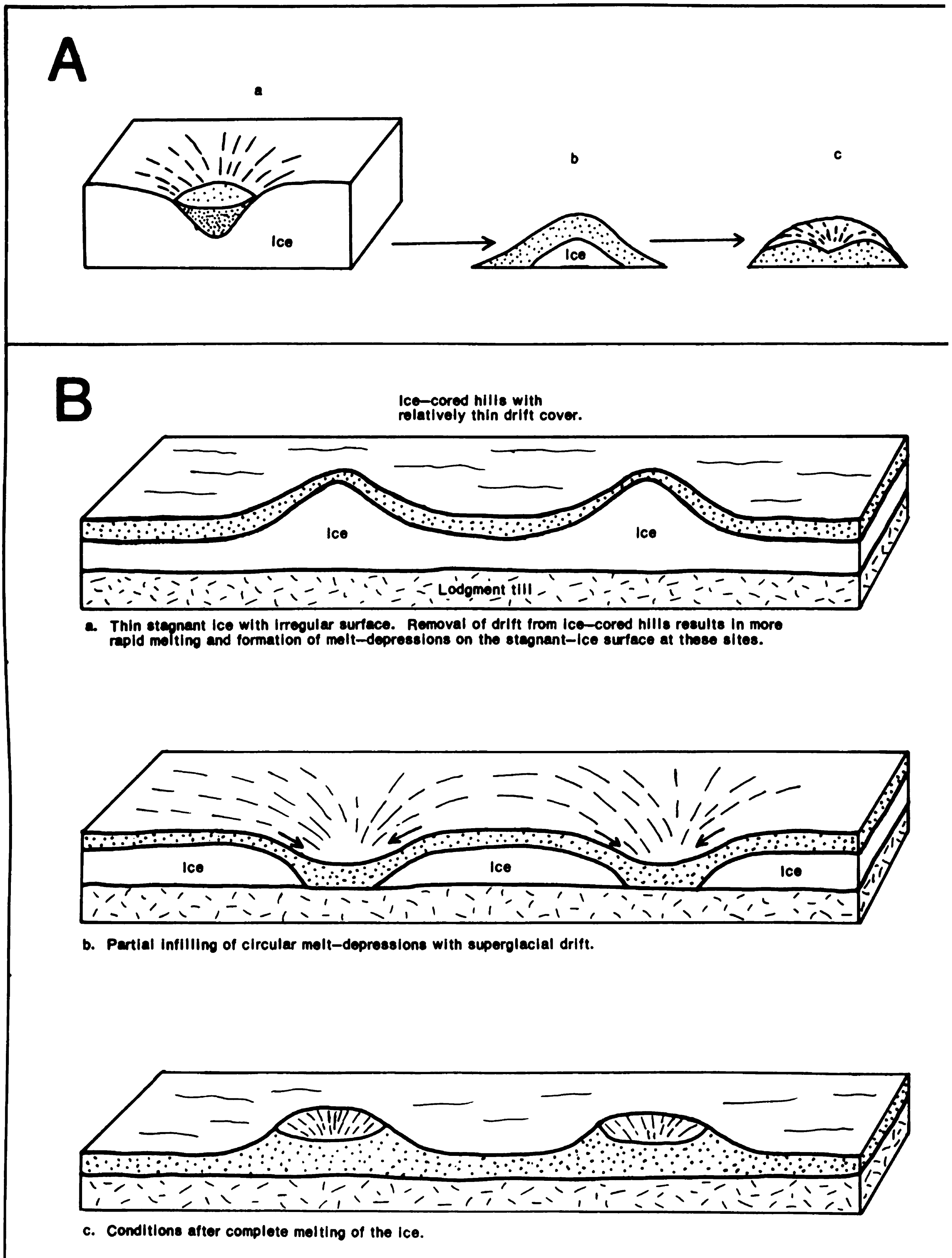

Figure 35. Manner of formation of circular disintegration ridges. Sequence A: formation according to Clayton (1967, Figure A-2). Sequence B: alternate mode of formation proposed by author.

Because their central depressions occasionally contain more water-laid sediments than might reasonably be expected to have been derived from the inner slopes of the circular disintegration ridges in postglacial time and because the steep outer slopes of the circular disintegration ridges of the Chippewa moraine complex seem to be ice-contact slopes, the author proposes an alternate hypothesis (see Figure 35) for the formation of these features to account for the above-cited "anomalies" to Gravenor's model.

Semicircular disintegration ridge

Features of this type resemble coastal blow-out dunes in several respects: they are parabolic in plan view and asymmetrical in side view (Figure 36), as well as highest on their convex ends and lowest on their concave ends (see Figure 41). The central depression of a semicircular disintegration ridge is always higher than the altitude of the dead-ice depression enclosing the convex side of the feature, and it slopes toward the open end of the ridge, as do the terminal parts of the ridge. The ridge has a steep ice-contact slope on its convex side but slopes with a moderate inclination toward the central depression.

Semicircular disintegration ridges appear to have formed in melt-embayments in the sloping margins of debris-mantled stagnant-ice masses. They are believed to usually consist of till transported into these melt-embayments by mass wastage; the high, convex end of a given ridge originally faced toward the central, thickest part of the ice mass, whereas its open end faced toward the adjoining ice-free area. Figures 36 and 41 illustrate the general morphological characteristics of semicircular disintegration ridges.

Figure 36. Side-profile view of a semicircular disintegration ridge. The high convex-end of the feature is on the left side of the photo; the open-end of the feature faces toward the right. See Figure 41 for a perspective view of features of this type.

Stagnant-ice terrace

A stagnant-ice terrace is simply a terrace in an area of stagnant-ice moraine, which is bounded by an ice-contact slope on one side and higher ground on the other. In effect, it is a shelf-like feature that interrupts an otherwise steeper slope. Stagnant-ice terraces may consist of till, glaciofluvial sediments, colluvium, or any combination of these. They contain material which collected in a depression separating a buried stagnant-ice mass from a nearby hill or valley wall.

The sediments comprising a given stagnant-ice terrace may have flowed, slid, or been washed into a depression either from the adjoining ice on one side or the ice-free area on its other. When the ice melted, the materials were left standing above the depression formerly occupied by the ice mass against which they had rested.

Stagnant-ice-border linear ramp

Features of this type, though constituting some of the larger, more prominent landforms of the study area, are not as numerous as most of the other types. A stagnant-ice-border linear ramp consists of material deposited at the margin of a stagnant-ice mass whose border was relatively straight. Ramps of this type are asymmetrical in cross-section and in overall form resemble small homoclinal ridges or cuestas (Figure 41). Their steep sides are ice-contact slopes and may be somewhat irregular, but their distal slopes are usually smooth and of gentle to moderate inclination.

Stagnant-ice-border linear ramps, although consisting of till, may also contain, or consist entirely of, stratified sediments. Genetically, they are a variety of the linear disintegration ridge. The dominant depositional process in their formation was apparently flowage of

Figure 37. Oblique view of the distal side of the stagnant-ice-border linear ramp in sec. 36, T32N, R33W. The surface of the ramp slop toward the lower right. Trees occupy the ice-contact slope and the hummocky stagnation moraine adjoining the proximal side of this feature. See Figure 41 for a perspective view of features of this type.

saturated superglacial till off a stagnant-ice mass into the adjacent ice-free area. In some cases superglacial material was washed off the ice and deposited as stratified drift; lacustrine sediments were also deposited in the ramps where the ice was bordered by a lake.

Where largely composed of till, their smooth distal slopes were presumably graded by mass wasting and running water before the adjacent stagnant ice had melted; when the latter event occurred, the watersheds for their distal slopes were largely destroyed, thus showing only minor postglacial modification by running water.

Stagnant-ice-border linear ramps did not form continuously along the margins of the ice-cored moraines (see Figure 41). Their discontinuous nature may reflect to some extent local variations in the amount of water that drained from the ice margins: where excessive, linear ramp development may have been inhibited. For example, in some cases linear ramps have conspicuous notches cut by superglacial streams that drained across the ramps into the adjoining ice-free area. In other places, ice-border linear ramps may not have formed (or be present) because (1) the slope of the stagnant-ice surface was not favorable for flowage of the superglacial till; (2) the ice margin was highly embayed, irregular, or rapidly receding; (3) the amount of superglacial till may not have been adequate to construct an ice-border ramp of any consequence; or (4) they may have formed in places, but were later buried by outwash.

Stagnant-ice-border semicircular ramp

Features of this type resemble stagnant-ice-border linear ramps in several aspects of their morphology and genesis; however, instead of forming at the edge of stagnant-ice masses whose margins were straight,

Figure 38. Side-profile view of the stagnant-ice-border semicircular ramp in sec. 29, T32N, R4W. Its steep ice-contact slope is inclined toward the left and its surface slopes gently toward the right. This feature contains both till and gravel. A gravel pit and gravel pile are situated on its near-side. See Figure 39 for a closeup view of the gravel pit wall.

Figure 39. View of the complex sediments associated with the stagnant-ice-border semicircular ramp described on Figure 38. The boulder, about four feet in diameter, came from the till exposed in the right wall of the pit. Well-stratified gravel is exposed in the pit wall just to the left of the boulder. This gravel is overlain by till in the upper left.

they formed in arcuate-shaped melt bays in the margins of the stagnant ice.

Stagnant-ice-border semicircular ramps are asymmetrical when viewed from the side (Figure 38), show a convex-outward form on their ice-contact slopes and gently to moderately sloping, flat surfaces on their distal sides; the latter merge with no apparent break into the typically low-relief topography that adjoins their distal sides (Figure 41). They may consist of till, stratified sediments, or a combination of both types. To the writer's knowledge stagnant-ice-border semicircular ramps have not been previously described in the literature.

Eskers

Eskers are distinguished from crevasse fillings (linear disintegration ridges) by their greater sinuosity and length and their tendency to traverse hills. The latter condition is possible because subglacial streams may have flowed under hydrostatic pressure, whereas the streams flowing in crevasses did not.

At least nineteen eskers were identified in the study area, and five are located short distances outside the area. Those within the thesis area are shown on Figure 45. Since not all of the eskers have been excavated for gravel, the lithologies of some of them could not be examined. In at least nine of the ninetten recognized in the thesis area a continuous or discontinuous layer of till up to fifteen feet in thickness overlies the gravels. This till is interpreted as of principally superglacial origin, let down onto the eskers from the overlying drift-ridden ice. The above interpretation is proposed because the eskers

show no structures of features, such as completely disrupted bedding or gaps, that would indicate a glacial readvance.

The author has also identified more than fifty eskerine ridges in and near Rusk County, to the north of the present study area. These eskers, in addition to those described for the thesis area, offer further evidence of the widespread stagnantion associated with the deglaciation of the Chippewa lobe.

Ice-walled gravel train (perched-type train)

An ice-walled gravel train (Winters, 1963, p. 44) consists of stream sediments (sand and gravel) deposited in an open trench in the surface of a debris-mantled glacier. Deposition of sediments of this type may have involved streams produced by normal runoff, as well as by meltwater (Clayton, 1967, p. 44).

The ice-walled gravel trains of the thesis area are generally small, compared to those described by Winters (1963) and Clayton (1967) for the Missouri Coteau. Possibly, long, well-developed ice-walled gravel trains are not present because the narrow width of the individual stagnant-ice moraines comprising the moraine complex precluded superglacial streams of the size necessary to construct large features of this type.

Kames

The Chippewa moraine complex contains few kames that fit Holmes' (1947, p. 248) ideal type, that is, "an isolated symmetrical hill, consisting chiefly of gravel or sand." Instead, most of the kame-like features of the area are variants of such stagnant-ice landforms as ice-margin gravel ramps, crevasse fillings, and stagnant-ice "sinkhole

fillings." The distinct scarcity of gravel pits in kames is further evidence of the area's lack of well-developed kames.

As a group, the kame-like features recognized do not display morphological uniformity. Most were, therefore, not classified as kames but, instead, assigned to the landform type they individually bear closest resemblance to. The remainder, those roughly similar to the above definition of a kame, are identified on Figure 45.

Kame terrace

Flint (1971, p. 209) indicates that a kame terrace consists of an accumulation of stratified drift, deposited chiefly by stream action, between a glacier and the side of a valley, and left as a constructional terrace after disappearance of the glacier. Furthermore, the edge of the terrace which bordered the ice typically shows a prominent pitted or irregular ice-contact slope. Only one example of a feature that closely matches this definition was identified. It is located in a relatively deep, narrow part of the Yellow River valley. Also present, particularly in the Jim Falls area, are several large, terrace-like features resembling the "outwash terraces" described by Flint (1971, p. 189). Both the kame and outwash terraces formed early and predate the disintegration of the ice-cored moraines.

Collapsed outwash topography

The characteristics for this kind of topography were described by Clayton (1967, p. 38); he states that

> collapsed outwash topography is similar to hummocky stagnation moraine, but it is composed of outwash . . . rather than till. It was formed during the melting of stagnant ice on which a superglacial outwash plain . . . had been deposited. Collapsed outwash

Figure 40. Stagnant-ice boulder concentration zone in sec. 30, T30N, R5W. This boulder field has an extent of several acres and is more or less linear in spatial layout.

> topography is defined as having less than one-half of its area flat and uncollapsed to distinguish it from pitted outwash, which has greater than half its area flat and uncollapsed.

Several localities in the thesis area are very similar to the above description of collapsed outwash topography. These are discussed in the landforms assemblage section, later in this chapter.

Because of the presumed extended duration and slow rate of melting of the buried ice, some of the material comprising the collapsed outwash in the study area is believed to have been deposited by nonglacial streams, as well as glacial meltwater. If so, this type of outwash may be related to what Clayton (1967, p. 32) termed "collapsed stream sediment," a designation he uses for the alluvium deposited by the superglacial streams of the Missouri Coteau stagnation moraine; in this case he defines "alluvium" as any stream sediment, whether deposited by streams of glacial or nonglacial origin.

Stagnant-ice boulder-concentration zone

Boulder concentrations of this type are not landforms, per se, yet do constitute distinctive landscape features and are, therefore, discussed in this section. The topographic setting of boulder-concentration zones suggests that they may represent a lag concentrate, formed on the surface of the stagnant ice, near hollows in the ice. When the stagnant ice separating the various ice-walled features finally melted, the boulder concentrations on the ice surfaces were let down in the resulting stagnant-ice depressions. These boulder concentrations may also represent collapsed boulder accumulations from the channels of superglacial streams. Or, they may represent superglacial stones that

were concentrated by mass wastage in crevasses or at the base of steep ice-cored ridges during the disintegration of the stagnant ice.

Landform Assemblages

One of the four major objectives of this study, stated in Chapter I, was "to determine the nature of the assemblage of landforms that make up the area." The results of this study indicate that the various topographic features are usually arranged into orderly assemblages of landform types.

There are two basic landform assemblages associated with the Chippewa moraine complex: (1) that associated with the interior of the moraine and (2) that associated with the moraine margins. Some landform types, however, exist in a variety of relative settings, both within and outside of the moraine. They will be examined separately.

Assemblage of Landforms Within the Moraine

As shown on Figures 41 and 45 the landform assemblage associated with the interior of a moraine of stagnant-ice origin consists of a variety of ice-contact features, interspersed with hummocky stagnation moraine--the latter by far the most widespread landform type. In addition to hummocky stagnation moraine the more common landform types, in approximate descending order of topographic extent, are: (1) ice-walled lake plains, (2) incipient ice-walled lake plains, (3) circular disintegration ridges, and (4) stagnant-ice terraces. The remaining landform types of this assemblage--all of which are described below--are relatively uncommon.

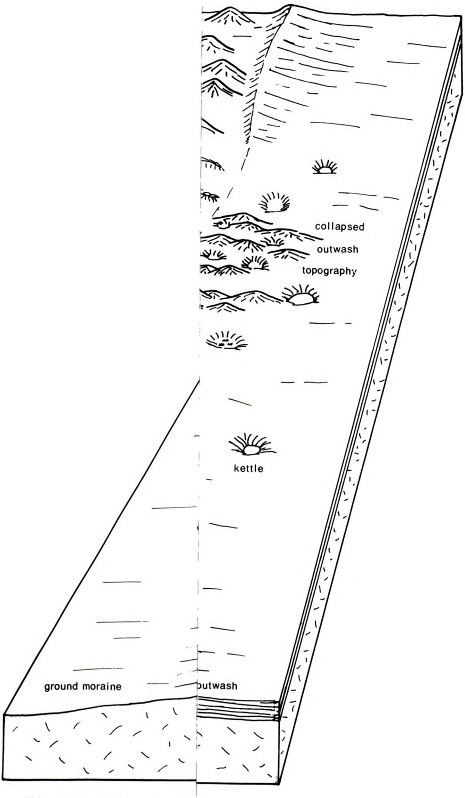

Figure 41. Asseation
moraine, shown other
landform types

Within the moraine complex east of the Chippewa River, the hummocky topography is arranged in a number of linear zones, reflecting the distinctive mode of formation of this part of the moraine. West of the Chippewa, however, hummocky topography extends more or less continuously across the full width of the moraine complex (Figure 9). In the transitional belt (pitted ground moraine zone) bordering the entire proximal margin of the moraine complex hummocky stagnation moraine is discontinuous.

Considerable regional variation exists within the areas of hummocky stagnation moraine with respect to local relief and slope angles. For example, low-relief hummocky stagnation moraine, with slopes of gentle to moderate inclination, is characteristic of the distal margin of the moraine complex east of the Chippewa River, indicating that in this area the superglacial till was relatively thin. In marked contrast are the steep slopes and high relief of the interlobate tract. These topographic characteristics suggest that the stagnant ice of that area was covered with thick superglacial drift, which fostered a stable superglacial environment--as defined by Clayton (1967, p. 30)--during disintegration.

In the interlobate tract, where the hummocky topography is best developed, the local relief usually ranges between thirty to eighty feet, but exceeds 100 feet in several places. Outside the interlobate tract hummocky topography of moderate local relief--that is, commonly ranging between twenty to fifty feet--is found (1) in the part of the moraine extending northwest from the interlobate tract, (2) in the Pike Lake moraine east of the Chippewa River (see Figure 9), and (3) in all of the stagnant-ice moraines east of the Yellow River. Finally, subdued hummocky moraine topography, typically ranging between five to thirty feet

of local relief, is found (1) between the Chippewa and Yellow Rivers (except in the prominent Pike Lake moraine) and (2) along the proximal and distal margins of the Chippewa moraine complex (Figure 45).

Ice-walled lake plains (perched lake plains) are second only to hummocky stagnation moraine* in abundance and account for some of the highest features within the Chippewa moraine complex. They are usually located well back from the margins of their respective moraines. In the final stage of ice wastage disintegration of the debris-mantled ice that enclosed the ice-walled lakes produced the hummocky stagnation moraine that characteristically surrounds perched lake plains (Figure 22). In terms of relative age, therefore, the perched lake plains were formed prior to the intervening areas of hummocky stagnation moraine.

Perched lake plains, although scattered widely throughout the moraine, do not exhibit a homogeneous distribution. They are very common and well developed in areas of conspicuous hummocky stagnation moraine but, conversely, are scarce and poorly defined in those areas of hummocky topography that are indistinctly expressed or of limited extent.

Lake plains that formed where the superglacial drift (and associated buried ice) was thick are perched well above the surrounding hummocky stagnation moraine, whereas the lake plains that formed in areas of thin superglacial drift (and correspondingly thin buried ice) are commonly lower.

Not infrequently the individual lake plains are separated from the surrounding hummocky stagnation moraine by distinct moat-like depressions

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*Dead-ice hollows, though common, are an integral part of hummocky stagnation moraine and show the same geographic patterns as that of hummocky stagnation moraine, described previously. They will not, therefore, be discussed separately.

which formed at places where most of the drift on the ice surface immediately adjoining the lake had creeped, flowed, or been washed into the lake. Thus, when the ice enclosing the lake melted, the thinner-than-average drift cover on the ice nearest the lake plain led to the formation of a depression or moat at the foot of the lake-plain rim. Depressions of this type are common in parts of the study area, and one of the most distinctive characteristics of lake plains is the ring of lakes, ponds, or swamps which encircle them. The small lakes encircling the lake plain in Figure 22 illustrates this relationship.

Incipient ice-walled lake plains differ markedly from perched lake plains, both in terms of morphology and relative location. Lake plains of the incipient type tend to have asymmetrical longitudinal profiles and are located relatively near the moraine border. As pointed out earlier, incipient ice-walled lake plains appear to have formed near the margins of stagnant-ice masses or where the surface of the stagnant ice sloped markedly in one direction. Under these conditions high rims developed along that part of the lake margin bounded by thicker ice, but little or no rim developed on the side bordered by thinner ice. Disintegration of the enclosing ice produced a lake plain that, though surrounded by hummocky stagnation moraine, is characterized by a surface that slopes conspicuously toward the moraine margin nearby.

Ice-walled lake plains of the incipient type are very common in the stagnant-ice moraines east of the Chippewa River. If the above interpretations are correct, then it is only logical that lake plains with incomplete rims should be common in that area, for the ice surface in the relatively narrow ice-cored moraines east of the Chippewa River

should have sloped steeply away from the axes of these moraines toward their margins.

The circular disintegration ridges of the thesis area are, in the majority, located within areas of hummocky stagnation moraine and appear to have formed after the ice-walled lake plains that have a higher altitude. Furthermore, they seem to be most common near the lake plains and may, collectively, completely encircle them.

Circular disintegration ridges are best developed in low- and moderate-relief hummocky stagnation moraine but uncommon in high-relief hummocky stagnation moraine. They are, therefore, quite common east of the Chippewa River but extremely scarce in the high-relief moraine west of the Chippewa.

Circular disintegration ridges sometimes exist in an aligned series, in which the rims of the individual circular disintegration ridges are interrupted at those places where the centerline of the series passes through them. Clayton (1967, p. 32) attributes these breaches to the melting of an ice-cored crevasse filling that underlay the series. If this interpretation is correct, the circular disintegration ridges of such an aligned series apparently all developed along the same or at least closely related crevasses.

Stagnant-ice terraces are also common features within areas of hummocky stagnation moraine. Features of this type are highly variable in their size, morphology, and relative location and, individually, represent material which collected in a depression separating a buried stagnant-ice mass from a nearby hill or valley wall. They are not necessarily linear features and may be found at any height, from near the bottoms of valleys to near the tops of hills or upland features.

Stagnant-ice terraces tend to be more conspicuous, and may actually be more abundant, in areas where hummocky stagnation moraine is poorly developed, such as along the distal part of the moraine complex east of the Chippewa River and around the flanks and bases of till-covered preglacial uplands, both east and west of the Chippewa River.

The less common features of the landform assemblage under discussion include (1) intermorainal ice-walled lake plains, (2) ice-walled gravel trains, (3) stagnant-ice boulder-concentration zones, (4) branching ice-walled alluvial-lacustrine trains, and (5) transmorainal lowlands. Intermorainal ice-walled lake plains (lowland-type lake plains) formed in the depressions separating the individual stagnant-ice moraines (Figures 9 and 45) and are, therefore, found principally east of the Chippewa River. Though typically encircled by hummocky stagnation moraine they are, in a few localities, closely-spaced and form nearly unbroken chains. Because of their lowland setting intermorainal ice-walled lake plains are not nearly as conspicuous as lake plains of the perched type, situated within the moraines. In addition, their surfaces are commonly poorly drained and uncultivated, in contrast to the well-drained, intensively cultivated perched lake plains.

The ice-walled gravel trains of the thesis area exhibit marked variations in size, morphology and composition. Some resemble crevasse fillings, while others have more the form of narrow tongues of ice-walled outwash. The ideal ice-walled gravel train, however, consists of a flat-surfaced, linear gravel plain, perched above flanking areas of hummocky stagnation moraine. If originally deposited on solid ground, its surface slopes toward the moraine margin, where it may merge with an outwash plain.

Perhaps the largest and best developed example of a gravel train within the study area is that found bordering the east side of Round Lake. It extends from sec. 9 to sec. 23, T32N, R9W. This feature slopes gently south, is at least two miles in length, averages about one-fourth mile in width, is about ten to twenty feet in height, and is partially encircled by hummocky stagnation moraine. Several other ice-walled gravel trains--or closely related features--extend down and out of the moraine west of the Chippewa River (Figure 45), but no features of this type were identified east of the Chippewa.

Stagnant-ice boulder-concentration zones, though not common, are conspicuous features of the moraine interior. They may be situated within the bottoms of stagnant-ice depressions, as well as upon the slopes of the associated ridges. They appear to be most common in the areas of hummocky stagnation moraine immediately surrounding ice-walled lake plains, or other features that formed in ice-walled environments.

Branching ice-walled alluvial-lacustrine trains are uncommon and found principally in the area between the interlobate tract and the Chippewa River and again in the lake district located northwest of the interlobate tract (Figure 45). Ideally, they are bordered by hummocky stagnation moraine on all sides, except for those places where streams drained from them during their formation. Some of these features are occupied by modern streams; others are not.

Transmorainal meltwater depressions are rare and are not a part of the basic assemblage of landforms. However, where found they are important because they represent gaps in the moraine. In other words, they interrupt the moraine and divide it into segments. They are, of course, bordered by hummocky stagnation moraine.

Landform Assemblage of the Moraine Margin

The types of landforms associated with a given part of the moraine margin depends largely upon the relative altitude of that locality. In those places where the moraine margin occupies an upland location the typical features are (1) stagnant-ice-border linear ramps, (2) stagnant-ice-border semicircular ramps, and (3) semicircular disintegration ridges. But where the moraine is bordered by lowland, outwash-type features (such as kame terraces) are more typical of its margin.

Stagnant-ice-border linear ramps usually trend parallel to and typically mark the edges of tracts or belts of stagnant-ice moraine; in this respect they may be found on either the proximal or distal sides of the individual stagnant-ice moraines. However, as brought out earlier, they did not form continuously along the margin of the ice-cored moraines. Furthermore, in some instances they were apparently formed at the margins of isolated ice masses and show no distinctive trends.

Stagnant-ice-border linear ramps tend to be bordered on their ice-contact sides by hummocky stagnation moraine and on their distal sides by ground moraine, outwash, or ice-walled lake sediments (Figure 41). One of the best defined linear ramps is that shown on Figure 37. That particular ramp is bordered by a flat outwash plain, in cropland, on its west and hummocky stagnation moraine, in forest, on its east.

Stagnant-ice-border semicircular ramps exhibit the same relative location as do linear ramps. They differ from them in form only. The most favorable environment for the formation of semicircular ramps seems to have been where the margin of the stagnant ice was relatively thin, for they are far more common east of the Chippewa River than west

of it. As is true of linear ramps, their ice-contact slopes commonly face toward hummocky stagnation moraine.

Semicircular disintegration ridges may exist anywhere within the moraine complex, as shown on Figure 45, but seem to be more common along the margins of the moraines. Furthermore, they are far more abundant east of the Chippewa River than west of it. This probably indicates that thinner ice existed in the former area, which was more favorable for the formation of features of this type. Often two or more of them may be found in the same locality and at several places within the moraine there exist an aligned series of semicircular disintegration ridges. The ridges of a given series collectively appear to mark progressive elongation of their respective melt-embayments.

Ice-walled outwash plains consist of outwash that was deposited in large indentations in the margin of the stagnant ice that formerly cored the moraine. Their surfaces slope toward and merge into the larger proglacial outwash plains bordering the moraine. Because of these relationships ice-walled outwash plains may be bounded on three sides by hummocky stagnation moraine. In effect, then, ice-walled outwash plains represent lobate- or irregularly-shaped extensions of proglacial outwash plains into the moraine itself.

Areas of collapsed outwash topography, of varying size, exist along both the distal and proximal margins of the moraine complex, as well as along the larger meltwater routeways throughout the study area. Probably the most pitted and certainly the largest area of collapsed outwash topography extends along the distal margin of the moraine complex from the northwest corner of the thesis area to a point about four miles northwest of the Chippewa River, a distance of about twenty miles. This belt of collapsed outwash topography, about one-half to one and one-half miles in width, separates the hummocky stagnation moraine on its proximal gradational border from the area of pitted outwash along its distal border. It is virtually impossible to determine where hummocky stagnation moraine ends and collapsed outwash topography begins on the basis of topography alone.

The kame and outwash terraces of the moraine margin are uncommon. The best examples are located along the distal margin of the moraine, near the Chippewa and Yellow Rivers.

Landforms Not Restricted to the Moraine

Most of the eskers of the study area are associated with the ground-moraine area, located on the proximal side of the moraine complex, and are commonly situated in the floors of valleys which lay parallel to the direction of ice flow. They tend, therefore, to be oriented perpendicularly to the trend of the moraine complex. Those eskers situated within the moraine complex are in some cases almost impossible to identify because of the hilly topography of these areas, or because portions of them are completely obscured by a thick covering of collapsed superglacial till.

Linear disintegration ridges, though not abundant, exist at a number of places. They are best developed in the transitional area north of Highway 64 and east of the Chippewa River, where they commonly range from one-half to one mile in length and are straight and aligned parallel to the direction of ice movement in their respective localities. In this area they might be considered as "controlled disintegration" features, as defined by Gravenor and Kupsch (1959, p. 54).

In contrast to the above, the linear disintegration ridges located within the areas of hummocky stagnation moraine east of the Chippewa River and south of Highway 64 show no preferred orientation. They appear to be more in the nature of the random ice-fracture fillings described by Kaye (1960, p. 360) and probably represent "uncontrolled disintegration" of the ice, as defined by Gravenor and Kupsch (op. cit.).

The linear disintegration ridges west of the Chippewa River within the moraine are also highly variable, but are larger and show more preferred orientation than those just described. Most of the larger linear disintegration ridges west of the Chippewa are crudely aligned, either parallel to or at right angles to the direction of last ice flow in their respective areas, implying that disintegration was partially controlled.

The study area contains few well-defined kames, that is, distinct rounded hills composed of sand and gravel. The hills referred to herein as kamic or kame-like features all contain at least some sorted material, but morphologically, as well as lithologically, they vary widely among themselves. As a group, they probably formed in several ways, are related to a wide variety of stagnant-ice landform types, and display no distinctive locational pattern.

The morphology and frequency of kame-like features varies from one part of the study area to another. They are most conspicuous, although not necessarily more abundant in the outermost, poorly developed part of the moraine complex, east of the Chippewa River. Mound-like kames, perhaps as many as a few dozen in number, are scattered throughout this sector. Just northwest of Stanley, in sec. 23, T29N, R5W is an east-west trending string of conspicuous, often linear kames. These appear

to have formed along the glacier margin or in a crevasse, parallel to and slightly back of the margin.

West of the Chippewa River, in the interlobate tract, and immediately west of it, are several excellent examples of hills or ridges that may be interpreted as kames, or features gradational between kames and perched lake plains. Because they contain glaciolacustrine sediments, in addition to glaciofluvial deposits, the author tentatively interprets these features as representing sinkhole fillings; that is, they consist of sediments deposited in deep, small-diameter ice-walled lakes.

Postglacial Modification of the Stagnant-Ice Landforms

The great majority of the stagnant-ice landforms are remarkably well preserved. Numerous rather fragile forms exist, which certainly could not be preserved had there been significant postglacial morphological modification of the stagnant-ice landforms.

The largest modifications have taken place along drainageways. Despite the incisement of the surfaces of some of the larger perched-type ice-walled lake plains by postglacial ravines, their rims are strikingly well preserved. By comparison, the lowland-type ice-walled lake plains have experienced greater postglacial modification because they are traversed by more numerous and larger drainageways. Several of them contain entrenched streams, which are descendants of streams whose courses were in part superglacial at the time entrenchment was initiated.

Perhaps the most obvious evidence of postglacial modification of the landscape may be found along streams that drain through rough

morainic topography, but these are not abundant. The valleys of these streams are segmented, alternating between narrow erosional segments and wide depositional segments. Where a stream initially crossed morainic ridges, it has cut notches into them; the floors of the intervening dead-ice hollows or basins are commonly poorly drained and underlain by fluvial or lacustrine sediments, or both. This kind of modification of the glacial topography is especially conspicuous in the rough interlobate tract, where the drainage is still largely nonintegrated.

Although postglacial modification of the glacial landforms is very apparent along streams, postglacial filling of depressions in the glacial topography has not been restricted to basins of this type. Probably the most frequent and widespread type of postglacial modification is represented by the many small basins, scattered throughout the moraine, that have been partially or completely filled with organic and inorganic sediments. Very likely, ponds were more numerous east of the Chippewa River at the close of the glacial epoch, but most have since been filled.

CHAPTER VIII

PHYSIOGRAPHIC DISTRICTS

From the spatial standpoint, significant geomorphic differences do exist within the thesis area. To aid in the description of the basic topographic patterns of the area, it has been subdivided into nine physiographic districts. Throughout this chapter, in all matters pertaining to the location of these physiographic districts, the reader is advised to consult Figure 42. In some instances the districts may be referred to as tracts, zones, or segments.

District 1. Outer Transitional Zone

The Outer Transitional Zone fringes the distal margin of the moraine complex (Figure 42). This district, because it was the first to be abandoned by active ice of middle Woodfordian age and was occupied but a short time by it, consists of a belt of poorly expressed morainic topography and of outwash deposited over stagnant ice. With regard to the latter condition, the outer margin of the moraine west of the Chippewa River is commonly well expressed and for most of its extent merges directly into collapsed or pitted outwash; weak morainic topography is, therefore, not extensive along this part of the moraine front.

East of the Chippewa River, however, much of the middle Woodfordian ice border, when at its maximum, occupied rolling bedrock-supported uplands. Consequently, this segment of the moraine complex is more

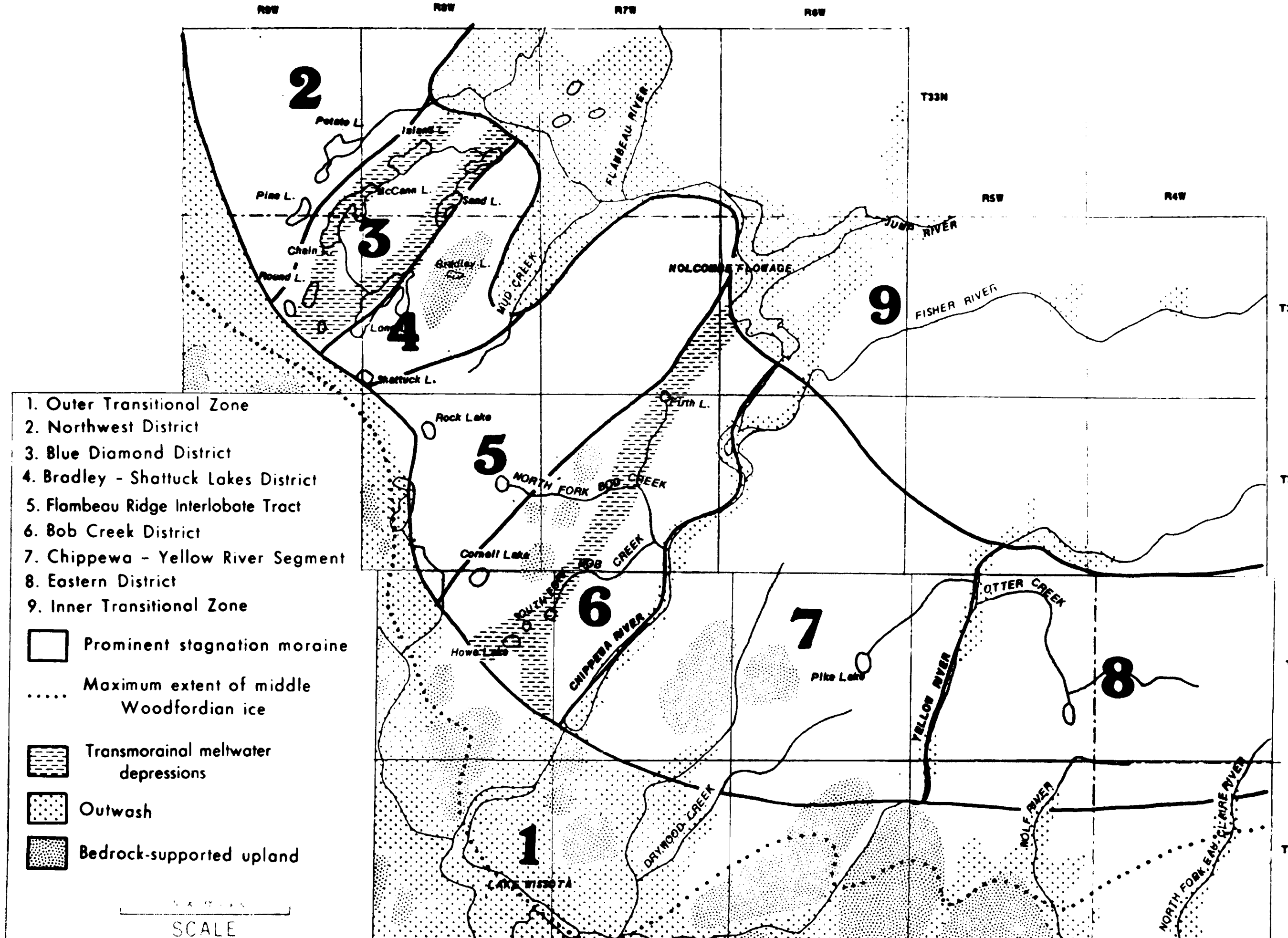

Figure 42. Physiographic Districts.

extensively fronted by poorly developed morainic topography than that west of the Chippewa.

District 2. Northwest District

District 2 consists of the northwestern part of the moraine complex. It is bounded on its southeast, and separated from the remainder of the moraine complex, by the large transmorainal lowlands that comprise District 3. In its overall form the northwest segment of the moraine complex may show more similarity to its counterparts east of the Chippewa River than it does to the remainder of the moraine complex west of the Chippewa.

Nevertheless, the Northwest District differs from the physiographic districts east of the Chippewa River in several important respects. Though it has about the same width (eight to ten miles) as the moraine complex east of the Chippewa, the northwest segment consists of more hilly morainic topography. In addition, the existence of several moraines in the moraine complex, so obvious east of the Chippewa River, cannot be easily discerned in the northwest segment. Finally, well-drilling reports indicate that the prominent morainic topography of the Northwest District is in part bedrock-controlled, a circumstance not found in the easternmost part of the morainic complex.

District 3. Blue Diamond District

Three aspects of this district--known locally as the Blue Diamond lake area--give it a distinctive character. First, it consists of two large transmorainal lowlands that constitute pronounced gaps in the moraine, separated by an elongated, southwest-northeast trending belt

of stagnation moraine. Second, the majority of the larger lakes of the study area are situated in these lowlands. Third, the belt of stagnation moraine separating the lowlands has both less altitude and local relief than exist in the moraine on either side of this district.

The floors of these lowlands show a complex pattern of sediments. In scattered localities they are underlain by till, but the surface material consists primarily of glaciofluvial and glaciolacustrine sediments.

The northwest boundary of this district follows the northwestern shores of, from south to north, Round, Chain, McCann, and Island Lakes. Possibly Pine and Potato Lakes, lying just northwest of this district, should also be part of it. Both lakes are elongated and lie parallel to the lakes of the Blue Diamond District. But because they do not occupy a distinct transmorainal lowland, they have been excluded. Perhaps further study will show that the history of the Pine-Island Lakes depression was similar to that of the transmorainal lowlands under discussion, and if so, the former depression should be combined with this district.

District 4. Bradley-Shattuck Lakes District

This district is named for Bradley and Shattuck Lakes, located in its central and southernmost parts, respectively. Except for its southeast side, this segment of the moraine is topographically well defined. On its northwest side it overlooks the Salisbury-Long-Sand-Mud-Rice Lakes transmorainal lowland. On the north half of its southeast side its higher altitude makes it clearly distinguishable from the Mud Creek

lowland, underlain by outwash, but on the south half of its southeast side it merges, with no break, into the interlobate tract.

The morainic topography of this district is very well expressed, exceeded in local relief only by that along the axis of the interlobate tract. On the basis of one well log, indicating sandstone at a depth of seventy-five feet near the center of this district, in addition to several others, at the north end of Long Lake, immediately west of the foregoing site, reporting sandstone at less than fifty feet depth, it appears that at least part of the greater relief of the northern half of this high morainic tract is the result of higher bedrock altitudes. In the southern half of this district, however, no well was recorded with less than 100 feet of drift and any direct effect of bedrock on topography is not apparent. The southern half of the district also is characterized by a greater number of deep, closed, dead-ice hollows than any other part of the moraine complex.

District 5. Flambeau Ridge Interlobate Tract

Containing by far the most impressive morainic topography of the thesis area, the local relief of this tract averages roughly twice that of the remainder of the moraine complex. The highest altitudes of the tract are found along its axis and consist of the string of ice-walled lake plains and kame-like features that extend southwest from Flambeau Ridge, shown on Figure 45. These stagnant-ice features, the most prominent in the thesis area, were formed along the contact zone between the two sublobes produced by Flambeau Ridge.

When observed in longitudinal profile, the highest part of the interlobate axis is seen to be located several miles southwest of

Flambeau Ridge (Figures 11 and 12). Apparently, less debris-laden ice converged along the interlobate contact immediately downglacier of Flambeau Ridge than was the case a few miles to the southwest (see Figure 6); ultimately, the middle segment of the interlobate tract was left with a higher altitude.

District 6. Bob Creek District

Bounded by the interlobate tract on the west and the Chippewa River on its east, this district is named for Bob Creek, which drains roughly half its area. Possessing considerable topographic diversity, this segment of the moraine complex was one of the more difficult to interpret.

The district may be subdivided into three relatively distinct subdistricts. The most obvious subdistrict consists of the large transmorainal lowland that parallels the Chippewa River and extends south-southwest from approximately the west side of the Holcombe Flowage across the moraine complex. This lowland is underlain by a complex pattern of glaciolacustrine and glaciofluvial sediments, interrupted in a few places by till associated with low-relief hummocky stagnation moraine.

While the northern one-third of this transmorainal lowland adjoins the eastern base of the interlobate tract, the southern two-thirds diverge from it. A second subdistrict, therefore, consists of the area situated between the interlobate tract and the southern two-thirds of the transmorainal lowland. The western side of this subdistrict exhibits moderately well-developed stagnation moraine, in contrast with its east

side, consisting principally of several relatively smooth bedrock-supported uplands (Figure 45), enclosed by hummocky stagnation moraine.

The third subdistrict consists of the two- to three-miles-wide strip of morainic topography located between the transmorainal lowland and the Chippewa River to its east. Topographic patterns in this subdistrict are extremely difficult to interpret. Several of the stagnation moraines identifiable east of the Chippewa River may be traced into this subdistrict, but their extent and nature are not apparent (see Figures 9 and 45).

District 7. Chippewa-Yellow River Segment

The segment of the moraine complex between the Chippewa and Yellow Rivers possesses several characteristics that distinguish it from adjoining districts. The morainic topography of this district is of less magnitude than in any other segment of the moraine complex and, with the exception of the Pike Lake moraine (Figure 9) is, on the whole, not well expressed. While several stagnation moraines may be positively identified, the individual moraines are not as easily differentiated as in the next district east. Moreover, the topography of the southern half of this district reflects more bedrock control than any other district (Figure 8).

District 8. The Eastern District

The Eastern District includes that part of the moraine complex bounded on the west by the Yellow River and on the east by the eastern boundary of the area studied. Two aspects of this district set it apart from the others: (1) its compound character, consisting of several

well-defined stagnation moraines, and perhaps of even greater distinctiveness, (2) its unusual array of stagnant-ice landforms.*

Though well developed, the morainic topography of this district is not as rugged as in those districts from the interlobate tract on northwestward. Also, this district contains no major bedrock-supported uplands.

District 9. Inner Transitional Zone

The Inner Transitional Zone, the last to be vacated by the active-ice margin of the Chippewa lobe, consists of an eight- to twelve-mile-wide strip bordering the proximal side of the moraine complex. Its distal and proximal margins are roughly parallel, and while the latter is not shown on Figure 42, it lies just outside the northeast corner of the study area; for reference purposes, Figure 5 shows both boundaries of this zone.

The topography of the district is a mixture of forms resulting from active- and stagnant-ice environments. That is, tracts of subdued to moderately well-developed hummocky stagnation moraine, as well as many large dead-ice depressions, are found throughout the district. These are interspersed with slightly rolling ground moraine tracts, the latter with drumlinoid characteristics.

Glacial erosional features, virtually non-existent in the moraine complex, are found at several scattered localities in or near the Inner Transitional Zone. The best example within this district consists of

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*In the author's view, the landform assemblages of this sector are so outstanding that the area is recommended to those interested in the subject as a type-locality for glacial landforms resulting from extensive marginal stagnation.

Figure 43. Ice-scoured bedrock knobs (roches moutonnees) located in the floor of Christmas Creek valley in sec. 14, T31N, R5W. This is one of the few localities in the study area where glacial erosional features are evident. The direction of ice movement was from right to left.

several small roches moutonnees, developed on low-relief Precambrian bedrock knobs. The latter are situated in the broad floor of Christmas Creek valley, about nine miles east of Cornell, in the NE 1/4, NW 1/4, sec. 14, T31N, R5W. Figure 43 is representative of these features. About five miles east-southeast, a 100-yard-long dike-like bedrock ridge, bearing a form suggestive of ice molding, parallels the east side of the Yellow River, in the NE 1/4, sec. 27, T31N, R4W. The largest roche moutonnee, situated along the boundary between the interlobate tract and the Inner Transitional Zone, consists of the easternmost knob of Flambeau Ridge, in the SW 1/4, sec. 6, T32N, R6W. The smooth stoss side of this knob faces northeast and is almost 100 yards wide; the entire knob has about fifty feet of local relief.

CHAPTER IX

DISTRIBUTION OF OUTWASH AND LOESS

This chapter is primarily concerned with the distributional characteristics of outwash and loess. The discussion of this aspect of these sediments has been delayed until this point because certain of their spatial relationships could not be well demonstrated until after the characteristics of stagnant-ice landforms, as well as topographic patterns within the study area, had been established. The distributions of other types of Quaternary sediments are discussed in Chapters II and VII.

Distribution of Outwash

Outwash is normally defined as sand and gravel "washed out" from and deposited by meltwater streams issuing from a glacier (Flint, 1971, p. 185). However, to accommodate all of the outwash-like deposits of the thesis area, outwash will be defined here as including sand and gravel deposited upon, along, or beyond the margin of a mass of glacial ice by streams, at least in part of meltwater origin. The latter stipulation is included because the upper layers of some of the sand and gravel deposits bordering the stagnant-ice moraines are thought to consist in part of nonglacial stream sediments.

As is true of outwash deposits in general, those of this area show a distinct grain-size decrease with increasing distance from their respective sources. Three types of outwash topography are recognized: collapsed, pitted, and nonpitted. The first two were deposited in an

ice-contact environment. Collapsed outwash topography represents outwash that was deposited over continuous or nearly continuous stagnant ice, whereas a pitted outwash plain represents outwash deposited over isolated blocks of stagnant ice. The latter type of outwash topography has more than fifty per cent of its surface unpitted, whereas the former type has less than fifty per cent unpitted. Nonpitted outwash plains consist of sand and gravel not deposited over stagnant ice and are, therefore, termed "proglacial" (Winters, 1963, p. 23).

Nonpitted outwash plains typically slope gently away from the former position of the ice margin. Pitted outwash plains may show a similar slope, but collapsed outwash topography is often hilly, broken topography, showing no noticeable horizontal trend with respect to altitude and perhaps even decreasing in altitude in the upglacier direction. All three of these types of outwash topography are present in the thesis area.

The distribution of outwash deposits was discussed in Chapter VII and is shown on Figure 45. Examination of this map will reveal that outwash is found in two general environments: (1) bordering the distal margin of the moraine and (2) in association with the larger glacial and postglacial drainageways.

The most extensive outwash area borders the outer margin of the moraine from the Chippewa River northwestward. In this area, the glacier margin bordered an almost continuous lowland; therefore, outwash adjoins most of the distal margin of the moraine. However, the presence of sandstone uplands one to three miles west of the glacier caused the outwash to be confined to a long narrow belt. Well records for this area indicate that the outwash averages over 100 feet in thickness and may exceed 150 feet in some parts of it. Beginning with nonpitted or slightly pitted outwash on the west, this outwash area grades eastward, through pitted outwash, into rolling to hilly collapsed outwash topography, which in turn grades into hummocky stagnation moraine.

East of the Chippewa River the outwash located beyond the moraine is less extensive and confined to valleys or lowlands separated by bedrock-supported uplands. East of the Yellow River, in particular, outwash is less extensive than one might expect. The author suspects this may reflect the relatively early stagnation of the glacier margin in this area, resulting in increased insulation by superglacial drift; this greatly reduced the rate of meltwater production and thereby the amount of outwash that was deposited. Furthermore, the greater size of the moraine west of the Chippewa River, as well as the greater amount of moraine-border outwash, may indicate that a greater amount of debris-laden ice reached the margin of the southwestward-advancing glacier in that area than east of the Chippewa, where the moraine, in relative terms, was formed more in a lateral position.

A rather anomalous feature of the outwash distribution is the discontinuous nature and limited extent of the valley train in the Yellow River valley, upstream from Cadott. The approximately north-south-aligned part of this valley, in particular, contains very little outwash; its narrow width and steep gradient were apparently not conducive to outwash deposition.

The part of the Chippewa River valley lying within the moraine complex also lacks a well-developed valley train, probably for the same reasons as given for the Yellow River valley. The presence of hummocky stagnation moraine bordering these two streams in many places suggests that during deglaciation both cut into and were essentially confined by debris-mantled stagnant ice. In addition, their steep gradients across the moraine complex were not conducive to meandering; thus, the glacial-Chippewa and Yellow Rivers did not erode broad valleys within the moraine complex. In summary, they formed only narrow, discontinuous valley trains within the moraine.

Bordering the inner margin of the moraine complex are three sizeable areas of outwash, all of which are pitted or adjoined by large dead-ice depressions, indicating that they were deposited over or partially in contact with stagnant ice. One of these areas borders Lake Holcombe on the Chippewa River and extends up the lower Jump River valley for several miles (Figure 45). Another occupies the Mud Creek lowland, immediately west of Flambeau Ridge, and merges into the third, and largest of the three, which borders the Chippewa River upstream from its junction with the Flambeau.

Immediately downvalley from each of these outwash areas the Chippewa River is underlain by till. This implies that (1) the outwash of the forementioned localities was deposited in low-gradient sections of the glacial-Chippewa (and its tributaries) that existed upstream from till "uplands" located athwart the course of the river and (2) that the outwash was deposited before the stream had entrenched itself into these till "uplands."

As a final note on the outwash features of the Chippewa valley, reference will be made to the Wissota terrace, a term proposed by Andrews (1965) for the terrace representing the valley train of the Chippewa valley and valleys tributary to it. The Wissota terrace was named for the prominent, high outwash terrace overlooking the Chippewa River in the vicinity of Lake Wissota. The term also applies to the extensive outwash surface fronting the Chippewa moraine from the Chippewa River northwestward to the northwest corner of the thesis area, a distance of about twenty-seven miles.

Distribution of Loess

In accordance with the nomenclature proposed by Frye, Willman, and Black (1965, p. 56) the Woodfordian eolian silts lying beyond the Chippewa moraine are assigned to the Peoria loess rock-stratigraphic unit, whereas those overlying the Chippewa moraine (of Woodfordian age) are assigned to the Richland loess, a subdivision of the Peoria loess.

Loess of over one foot in thickness covers a large part of the study area, yet there are several sectors where it is thin or absent. Despite its discontinuous distribution and marked local variation in thickness, it reveals a meaningful pattern.

The distribution of loess seems to have been controlled primarily by (1) the distribution of outwash plains which served as the principal source for the loess;* (2) prevailing winds, with loess typically being

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*At least some of the loess may have been derived from silts of cryogenic origin, in particular, from upland sites in the Altonian drift-area, fronting the Woodfordian moraine. The basis for this reasoning is as follows: Woodfordian-age ice-wedge casts exist at numerous locations within the Altonian drift (Black, 1965), indicating that cryogenic activity was important in that area at that time. It is believed (Berg, 1964; Taber, 1953; Troll, 1958; Washburn, 1973, are representative) that in such an environment the formation of silt-size soil particles by frost action (frost weathering) may be significant. Conditions during Woodfordian time were ideal for the removal of such silts by wind erosion: vegetation was scarce (Black, 1965) and wind erosion was active (Smith, 1949). The relative abundance of relict ventifacts at upland sites in the Altonian drift-area (observed during this investigation) indicates that intense wind erosion--and, presumably, the deflation of silts formed as described above--did occur.

most extensive and thickest to the east of the outwash areas; and (3) topographic conditions, with loess deposits often restricted to or reaching their greatest thickness on the summits or leeward slopes (relative to the loess source) of the higher ridges.

With regard to the latter variable, three anomalous sites, in particular, are worthy of attention. At one of these, a ridgetop about eight miles west-northwest of Lake Wissota, loess about ten and one-half feet thick was observed, although a roadcut across the top of a high ridge about five miles southeast of Lake Wissota shows a loess cover of twelve feet, the thickest observed in this part of Wisconsin. More than eight feet of loess was noted on the southern slopes of the high bedrock upland located two miles north-northwest of Lake Wissota. The thick loess deposits on these higher ridges within the older drift area, west and south of the moraine, where loess is often totally lacking, suggests that silt-laden winds moving west and south from the outwash plains fronting the moraine did not, in general, deposit much loess, except where topographic conditions were favorable. Examples of possibly favorable environments were (1) wind-shadow locations associated with high ridges and (2) locations where silt-laden winds experienced a decrease in velocity, such as where they were forced to rise over a topographic barrier.

Loess also is generally thin or absent on the extensive outwash plains at the distal margin of the moraine west of the Chippewa River. Where found, it probably indicates that although outwash deposition had ceased at that site, meltwater activity was still active in some nearby section. For instance, over six feet of loess was encountered in auger borings on an outwash terrace two miles north of Lake Wissota on the west side of the Chippewa River, in sec. 2, T29N, R8W. This terrace has an altitude about forty feet higher than the main outwash surface of the Wissota area.

Within the moraine complex, loess is scarce northwest of the interlobate tract. It is found, however, in scattered localities between the interlobate tract and the Chippewa River, at thicknesses of from one to four feet, and again on some of the uplands bordering the north and west sides of the extensive outwash area along the Chippewa River above its junction with the Flambeau.

East of the Chippewa River loess is more widespread, but as is true west of the Chippewa, it is of restricted extent within the moraine complex, particularly in the rougher hummocky areas (also true west of the Chippewa), where disintegration of the buried stagnant ice probably took place after loess deposition had ended.

The localities of notable loess accumulation east of the Chippewa River seem for the most part, to be located east of, or in an overlapping relationship with, the eastern sides of local outwash plains, shown on Figures 42 and 45. Within the moraine complex, loess from one to three feet in thickness is found in several localities immediately east of the Chippewa River (see Figure 44), from Jim Falls to Holcombe, particularly where the topography is least hummocky and a valley train is present along the Chippewa (see Figure 45).

In the area south of the well-developed part of the moraine complex, loess is almost totally lacking in the western half of the sector located between the Chippewa and Yellow Rivers but ranges from a few inches to three feet or more in thickness in the eastern half of the

Figure 44. Loess capping stony till in a roadcut. This site is located about two and one-half miles south-southwest of Cornell in sec. 36, T31N, R7W. The source of the loess was the valley train along the Chippewa River a few miles west of this locality.

same sector. The loess source for the latter area is believed to have been the large outwash plain around Lake Wissota. The absence of loess immediately to the east of the Wissota outwash plain probably indicates that the winds coming across this extensive plain were relatively strong, as evidenced by the eolian sands and ventifacts of that area, and were, therefore, able to suspend their silt loads for greater distances on the east than was true for the smaller outwash plains of the thesis area. As another possibility, the outwash deposited by the turbulent glacial-Chippewa River in the Wissota area may not have contained as much silt as the outwash deposited in other areas by smaller meltwater streams.

Relatively thick loess accumulations, two to four feet in depth, are common in a three- to four-mile-wide strip adjoining the south boundary of the thesis area, between the Yellow River and the Chippewa County line, about eight miles eastward. Within that area, the deep loess deposits on the uplands directly east of the north-south part of the Yellow River are somewhat of an anomaly, for there is no valley train in this part of the Yellow River valley. It appears that the eolian silts of this locality were derived from the outwash plain in the vicinity of present-day Lake Wissota. If so, they were blown up the east-west-trending part of the Yellow River valley (see Figure 45), to where the valley angles abruptly northward, at which point the eastward-moving winds left the valley and deposited their silt loads on the uplands to the east. More than eight feet of loess was observed on one eastward-facing slope in this locality.

Farther to the east, loess several feet in thickness overlies some of the higher parts of the pitted outwash district that extends between

Boyd and Stanley (Figure 45). Though highly atypical, more than ten feet of loess was observed on this outwash plain, at a site located about two miles directly northeast of Boyd. Just north of Stanley deep loess has been stripped from the outwash and placed into large piles, preparatory to commercial gravel operations. North of the moraine complex thick loess, over five feet in places, is quite extensive to the northeast, east, southeast, and overlapping the east side of the large outwash area that extends up the Jump River valley from the Chippewa River (see Figure 45).

CHAPTER X

HYDROGRAPHIC FEATURES

This chapter is principally concerned with the distribution and morphology of selected hydrographic features, as they relate to glaciation, particularly ice stagnation.

Drainage Patterns of the Moraine

One of the more significant aspects of the Chippewa moraine complex is its influence on the stream pattern of Wisconsin. Only two rivers traverse its full width in the entire 125-mile distance between where it junctions with the St. Croix moraine on the west and where it is crossed by the Wisconsin River on the east. Both of these rivers, the Chippewa and the Yellow, cross the moraine complex within the study area. In effect, then, it is one of the more important, and better defined, drainage divides in Wisconsin.

Because of the limited number of water gaps in the Chippewa moraine complex, it seems obvious that the Chippewa and Yellow Rivers must have carried a great deal of glacial meltwater during the deglaciation phase of the Chippewa lobe. As noted previously, both these rivers, where they cross the moraine complex, have in many places cut through the drift to the resistant Precambrian surface. It has also been established that they had cut their valleys even before the end of Woodfordian glaciation, that is, before all of the remaining ice had completely melted. The apparently rapid cutting of the valleys of these two rivers was probably due to erosion achieved by the large quantities of meltwater.

With the exception of the Chippewa, Flambeau, and Jump Rivers, the streams of the transitional area bordering the proximal margin of the moraine complex (see Figure 45) are small and do not occupy well-entrenched valleys. Moreover, there are numerous sectors where streams are totally lacking, in particular the sector bounded by Highway 64 and the Jump and Chippewa Rivers. These parts of the transition area are characterized by numerous large, poorly-drained depressions of stagnant-ice origin. In some localities drumlinoid topography, as well as controlled-type disintegration, has given the poorly-drained depressions a distinctly linear character.

Within the Chippewa moraine complex proper, there also are a great many poorly-drained or undrained depressions, but in this area they are usually smaller than those in the transitional area. The drainage pattern in areas of hummocky stagnation moraine is typically highly disorganized, and probably over one-half of the total area of hummocky stagnation moraine is not part of an integrated surficial drainage system. Streams are also relatively small in the areas of hummocky stagnation moraine, with their dimensions probably averaging less than those in the ground moraine areas. This is because the highly irregular surface of the hummocky areas has not been conducive to the formation of extensive drainage systems.

As mentioned earlier, the stream valleys in areas of hummocky stagnation moraine are often variable in depth and width. The valleys are shallow and wide where the streams flow through dead-ice hollows but deep and narrow where the same streams are incised into the ridges separating the hollows. This sharply contrasts with the stream valleys of more or less uniform depth and width in the ground-moraine areas. Furthermore, the stream valleys within hummocky stagnation moraine are often incised into lacustrine or fluvial sediments, a condition much less common in the ground moraine areas.

East of the Chippewa River the various stagnation moraines, when viewed collectively, have resulted in the formation of a crude trellisform drainage pattern. The streams trending along the intermorainal depressions are frequently incised into glaciolacustrine sediments and, therefore, are bordered by flat lake-plain topography. In addition, a large proportion of the streams comprising the Bob Creek drainage system, situated between the interlobate tract and the Chippewa River, are incised into relatively flat lacustrine plains.

The typically shallow valleys of the smaller streams of the moraine complex are in sharp contrast with the relatively deep bedrock-floored valleys of the Chippewa and Yellow Rivers. The latter two valleys were cut by meltwater streams, and came into existence before the stagnant ice had melted. By comparison, the eroded valleys of the smaller intermorainal streams are mainly postglacial in age.

Lakes

In the study region there are more than 500 lakes of glacial origin that exceed an area of one-half acre (Sather and Threinen, 1963), and all appear to occupy basins of ice-block origin. Despite the large total the vast majority are found west of the Flambeau and Chippewa Rivers, and there are extensive areas with virtually no lakes.

At the time of European settlement there were only two natural lakes with an area of more than fifteen acres east of the Flambeau-Chippewa dividing line (see Figure 45). One of these, Otter Lake, has since been expanded by a dam. With two or three exceptions the remaining few dozen lakes and ponds east of the Flambeau-Chippewa are well below five acres in area.

The striking scarcity of lakes in the stagnation moraines east of the Flambeau-Chippewa (also discussed in Chapter VII) probably indicates that the relatively thin superglacial drift of that area was not favorable to the eventual formation of deep, enclosed dead-ice hollows. The two largest lakes of the area, Pike and Otter, both of which exceed forty feet in depth, appear to owe their existence to special circumstances: both are located in the largest of the stagnation moraines east of the Chippewa River, and both appear to occupy ice-block basins in north-south-trending preglacial valleys. But with the exception of the basins occupied by these lakes, the preglacial valleys are essentially drift-filled.

One of the most striking characteristics is the large number of lakes of glacial origin west of the Flambeau and Chippewa Rivers, with approximately 450 exceeding one-half acre in size. The author believes that the larger number of lakes west than east of the Chippewa River resulted from (1) thicker stagnant ice during disintegration due to (2) a superglacial drift cover of greater vertical thickness. When the stagnant ice west of the Chippewa disintegrated into blocks, it may still have been 100 or more feet in thickness. This, combined with the thick superglacial drift of the area, led to the formation of high-relief inter-block (ice-fracture) fillings or, conversely, deep dead-ice hollows upon complete melting of the ice. In essence, the greater depth and more complete closure of the dead-ice depressions west of the Chippewa River favor more lakes here than east of the Chippewa.

The distribution of lakes west of the Flambeau-Chippewa line is by no means uniform. Three distinct lake concentrations may be recognized. Two of these, both located near the southwest end of the interlobate tract, consist for the most part of small lakes situated in areas of hummocky stagnation moraine. The third, represented by Long and Island Lakes, located in the north and south parts of the concentration, respectively, contains the majority of large lakes of the study area.

The two concentrations of small lakes are located at the apex and on either side of the reentrant separating the two glacial sublobes produced by Flambeau Ridge (see Figure 45). One concentration, formed along the margin of the sublobe that came around the east end of Flambeau Ridge, is for the greater part situated in the E 1/2, T31N, R8W. The other concentration, originating along the margin of the sublobe that extended around the west end of Flambeau Ridge, is largely confined to the S 1/3, T32N, R8W. The contact between the two sublobes approximately follows the line of high kame-like features extending northeastward from the north end of Marshmiller Lake, in sec. 17, T31N, R8W.

Both lake concentrations appear to have formed where the glacier margin had a small radius of curvature, which apparently induced severe crevassing. The intense fracturing of the glacier margin in these localities fostered disintegration of the stagnant ice into numerous small blocks and was ideal for the formation of many small dead-ice hollow lakes. Ice disintegration in these two localities was, to a degree, of the controlled variety, for in both areas the long axes of the lakes tend to be aligned either parallel or perpendicular to the respective sublobe margin.

The third lake concentration is comprised mainly of several large, elongated lakes (exemplified by Long and Island Lakes), oriented southwest-northeast and situated within two large transmorainal lowlands located about five miles northwest of the interlobate tract. As interpreted in Chapter VII, these pronounced transverse lowlands extending the full width of the moraine represent routes where meltwater, spilling across the ice-cored moraine, had removed much of the superglacial drift. In consequence, when the stagnant ice underlying the spillway floors melted, there remained a series of large linear basins (now occupied by lakes) due to the absence, or scarcity, of superglacial drift in these localities. Finally, it should be noted that the deep basins occupied by these lake chains may also coincide with southwest-northeast oriented ancestral river valleys. If so, the lakes may owe part of their existence to this circumstance.

Abandoned Drainage Channels

Some of the drainage channels (and valleys) within the area are believed to have formed during the glacial epoch but have remained unused, or occupied only by underfit streams, since deglaciation. Although glacial meltwater was presumably involved in the formation of all of these features, some are believed to have been affected by streams, in part at least, of nonglacial origin.

Several types of abandoned drainage features are present, varying greatly in size. For instance, in the section on ice-walled lake plains in Chapter VII, it was established that the surfaces of more than twenty of these perched plains are incised by shallow, abandoned stream valleys that formed after the ice-walled lakes had drained but before the superglacial drainage systems were destroyed by disintegration of the stagnant ice. These features are shown on Figure 45.

The dozen or more abandoned drainage channels that exist at or just beyond the moraine margins represent another type. These were formed by streams that drained away from the glacier when near its maximum extent or by streams that originated within the moraines before disintegration of the buried ice. The two largest "dry valleys" in this group were made by streams whose courses were in part superglacial. One parallels part of the west margin of the moraine and extends from sec. 23, T32N, R9W, and the other leads westward toward Lake Wissota from sec. 28, T29N, R7W. Both valleys are ten to twenty feet in depth and up to one-quarter mile in width.

A third class of abandoned drainageways includes the numerous, often interconnected meltwater channels formed by the glacial-Jump and glacial-Chippewa Rivers (see Figure 45). Those made by the glacial-Jump are located in an area of pitted outwash east and southeast of Lake Holcombe in the NE 1/4, T32N, R6W. The floors of the meltwater channels in this area are not all graded to each other, indicating they were not all occupied simultaneously. Again, in the large outwash area flanking the Chippewa River, just above its junction with the Flambeau, there are several interconnected meltwater channels formed by the glacial-Chippewa River.

Probably the single most impressive spillway is that which extends around the northeast, east, and southeast sides of Lake Holcombe (see

Figure 45). The glacial-Chippewa River followed this route until the stagnant ice occupying the western part of the present Lake Holcombe had melted sufficiently to open a lower route two and one-half miles west, that is, the modern route. The southernmost part of this glacial drainageway, or that lying south of the latitude of Holcombe (see Figure 45), will be given special attention. It appears that for a considerable time after the Chippewa River had assumed its present course above Holcombe it continued to occupy the large valley now followed by the lower Fisher River, from Holcombe southward. At that time the area immediately southwest of Holcombe was underlain by thick stagnant ice. When this ice melted, the Chippewa River was diverted from its Fisher River route to the present route leading southwestward from Holcombe.

The presence of numerous exposures of Precambrian rocks, some bearing potholes, in the broad floor of the lower Fisher River valley indicates that the glacial-Chippewa River had cut down to bedrock in this locality before the lower, modern route to the west was opened by the melting of buried stagnant ice. If these interpretations are correct, it would appear that the Chippewa River, which is not significantly entrenched into the Precambrian surface, had essentially reached its present depth before all of the buried ice had melted. Additional evidence to support this interpretation is presented in Chapter XI, which summarizes the Quaternary history of the area.

At the east side of the village of Jim Falls, on the Chippewa River upstream from Lake Wissota, is a distinct abandoned stream valley (Figure 45), whose upper and lower ends connect with the modern Chippewa valley, but at slightly higher altitudes than that of the modern valley.

The tentative interpretation is that during late glacial or early post-glacial time the present townsite of Jim Falls was an island in the Chippewa River. More rapid downcutting by the channel on the west side of the island ultimately resulted in abandonment of that part of the river channel lying east of the island. Subsequent downcutting by the Chippewa in the Jim Falls locality has lowered the floor of the river several tens of feet below its level at the time the channel located on the east side of Jim Falls was abandoned.

CHAPTER XI

SUMMARY OF THE LATE PLEISTOCENE HISTORY

Glacial History

Work by Hole (1943) shows that, based on the depth of leaching and weathering, the oldest drift of central Wisconsin is clearly Wisconsinan in age. Because radiocarbon dating indicates that its correlative drift in western Wisconsin is only about 30,000 years old, Black, et al. (1965) conclude that the oldest surface drift deposited beyond both the St. Croix and Chippewa moraines is late Altonian in age.

The studies by Hole (1943) in central Wisconsin and Black (1959) in western Wisconsin further indicate that there is no conclusive evidence for pre-Wisconsin drift in the subsurface of these areas, despite interpretations to the contrary by Weidman (1907; 1913), in central Wisconsin, and Chamberlain (1910), Leverett (1929), and Flint, et al. (1959), in western Wisconsin. Nonetheless, statements contained in Frye, Willman, and Black (1965) and Black and Rubin (1967-68) imply that the pre-Wisconsin glacial history of west-central Wisconsin has not been satisfactorily resolved.

Early Woodfordian drift was not recognized in the thesis area. However, Frye, Willman, and Black (1965, p. 43) show surface drift of this age in the northern part of the reentrant between the Chippewa and St. Croix moraines. If this is correct, then drift of this age presumably underlies at least part of the thesis area in the subsurface.

Although there are no radiocarbon dates for the Chippewa moraine (Black and Rubin, 1967-68), it has been, since the studies of Chamberlain

(1876-77; 1883a; and 1883b), correlated with the St. Croix moraine. But the validity of this correlation is now uncertain, for, while Frye, Willman, and Black (1965) indicate a probable middle Woodfordian age (15,000 years B.P.) for both the Chippewa and St. Croix moraines, Wright, et al. (1973) interpret the St. Croix moraine to be of early Woodfordian-age (about 20,000 years B.P.).

On the basis of the interpretations by Frye, Willman, and Black (op. cit.), it would appear that the active margin of the Chippewa lobe receded from the position of the Chippewa moraine about 15,000 years ago. However, the stagnant-ice core of the Chippewa moraine may have persisted for several thousand years thereafter. This means that some of the stagnant-ice landforms of the Chippewa moraine are probably of late Woodfordian age (12,500 to 14,000 years B.P.) and may even be younger. This may be the case because paleoecological studies and radiocarbon dating of the basal organic sediments in ice-block lakes of the nearby St. Croix moraine of eastern Minnesota (Florin and Wright, 1969) show that stagnant ice persisted there until after the onset of the pronounced warming trend of the Two Creekan interval, which began about 12,500 B.P. (Frye, Willman, and Black, 1965).

Till overlies thick outwash deposits in several places north, east, and southeast of Ladysmith, along the contact between the transitional zone of this study and the ground moraine area to its northeast. This indicates a minor late-middle Woodfordian readvance of the Chippewa lobe before it began its final disintegration. The presence of numerous eskers in the ground moraine area immediately to the northeast of these sites, however, indicates that the margin of the Chippewa lobe did not retreat in an orderly manner; instead an extensive marginal area underwent large-scale stagnation soon after the above-cited minor readvance.

In late Woodfordian time the Chippewa lobe constructed a large end moraine, extending across the northern part of Wisconsin in the headwaters area of the Chippewa-Flambeau watershed (see Figure 1). The abundant quantity of meltwater coming from this ice front very likely played an important role in the formation of the Chippewa valley. In particular, it may have assisted in the rapid entrenchment of the Chippewa River down to the level of the Precambrian surface, which appears to have taken place even before complete melting of the buried ice of the Chippewa moraine.

Black's (1969c) work on the extent of Valderan ice in the Upper Peninsula of Michigan seems to indicate that very little, if any, meltwater from the Valderan ice front drained by way of the headwaters of the Flambeau and, from it, to the Chippewa River. Assuming this to have been the case, the Chippewa valley did not receive large quantities of meltwater after the close of Woodfordian time. However, after Woodfordian time it may have carried some meltwater from the melting of stagnant-ice masses in northern Wisconsin.

Periglacial Activity

Periglacial features, especially ice-wedge casts, were observed at many localities in the Altonian drift area, but only two localities were identified that reveal features of probable permafrost origin within the Chippewa moraine complex. This accords well with the findings of Black (1964; 1965), who, in describing the periglacial features of Wisconsin, points out that ice-wedge casts, although found in Woodfordian

drift in a few localities, are found principally in Altonian drift. Most of the ice-wedge casts, therefore, must have formed during the period between Altonian and middle Woodfordian time. However, Black states (1964, p. 26) that permafrost may have existed in Wisconsin from the end of the Altonian until the end of the Woodfordian, or from 30,000 to 12,500 years ago. This interpretation is substantiated in part by Florin and Wright (1969). They indicate that the persistence of buried ice in the St. Croix moraine (in some cases until late Woodfordian time) was associated with the existence of tundra conditions in that area. Because the stagnant-ice core of the Chippewa moraine is believed to have originated about 15,000 years ago and because the stagnant ice appears to have had an extended life-span, it seems reasonable to assume that a periglacial environment also persisted in this area until late Woodfordian time. If this was the case, then cryogenic and periglacial eolian activity may have continued to operate in the Chippewa moraine-area until late Woodfordian time.

Within the moraine complex two instances of features of apparent periglacial origin, both of presumed late Woodfordian age, are (1) intensively frost-rived quartzite exposed on the eastern end of Flambeau Ridge and (2) a series of poorly developed ice-wedge casts and involutions, exposed in an approximately forty-foot-long roadcut excavated in stratified sands and gravels in the SE 1/4, sec. 12, T32N, R8W, on the inner margin of the moraine complex. It is possible that additional instances of permafrost phenomena exist in the thesis area, but were unrecognized during this investigation.

Other phenomena of the thesis area, presumed to be at least in part of periglacial origin, include the eolian sands and abundant ventifacts of the Altonian drift-area. Also, some of the loess of the area may be of cryogenic origin.

Finally, the possibility exists that at least some of the hill-slopes of the area, in existence before late Woodfordian time, owe their convexo-concave profiles in part to solifluctional and related activities. If such hillslopes do exist, as is the case in the Driftless Area of Wisconsin (Black, 1964), they provide further evidence of the role of periglacial activity in the development of the geomorphic phenomena of this area.

Comparison of the Chippewa Lobe with the Other Glacial Lobes of Wisconsin

In comparison to the other major Woodfordian-age glacial lobes that affected Wisconsin (the Superior, Green Bay, and Lake Michigan lobes), the Chippewa lobe did not advance as far south, generally did not construct moraines as large, nor did it construct as many moraines as the other lobes. Furthermore, its ground moraine, while drumlinoid or coarsely-fluted over large areas, does not contain numerous, well-defined drumlins, as do the ground moraines of the Superior and Green Bay lobes.

Arranged on a scale of relative activity, the Chippewa lobe, among the four cited above, would probably be classified as the least active, and the Michigan lobe as the most active. Not only did the Michigan lobe construct a far greater number of moraines than did the Chippewa, but most of them (Frye, Willman, and Black, 1965, p. 55) are comprised of material deposited from an active ice front. By contrast, the Chippewa moraine complex, which represents the bulk of the morainic

topography associated with middle Woodfordian Chippewa ice, is essentially of stagnant-ice origin.

The relative smaller size and lesser activity of the Chippewa lobe, as manifested by the small number of moraines it constructed, its limited southward extent, and the widespread evidence of marginal stagnation, was probably due in large part to its comparative thinness. Because the subglacial surface traversed by the Chippewa lobe--across the highland south of Lake Superior--was several hundred feet higher than the routes traveled by the other lobes, the thickness of the Chippewa ice was correspondingly several hundred feet less, its hydrologic budget probably marginally positive at best, and its advance rate, as a consequence, relatively slow.

CHAPTER XII

SUMMARY AND RECOMMENDATIONS FOR FUTURE STUDY

Summary

This study has been primarily concerned with the mode of formation of the Chippewa moraine complex. It has been determined that marginal stagnation of the Chippewa lobe occurred on an extensive scale. As a result most of the glacial landforms in the Chippewa moraine complex are, at least to some degree, of stagnation origin.

A large variety of stagnant-ice landform types are present and have been described in detail. The various landform types comprise a gradational series, with many of the individual geomorphic features of complex origin and showing characteristics of two or more of the landform types. The overall arrangement of the geomorphic features reveals distinct landform assemblages in which the various landforms have an orderly arrangement relative to one another. These assemblages have been described and interpreted. Several previously unrecognized stagnant-ice landform types were identified and hypotheses on their formation were proposed, notably (1) incipient ice-walled lake plains, (2) semicircular disintegration ridges, and (3) stagnant-ice-border semicircular ramps.

The stagnant-ice landforms were largely derived from superglacial or ablation drift. Marked topographic variations from one district to another are apparent, in part reflecting variations in the manner of deposition of the superglacial drift and the characteristics of the

underlying ice. Areas of rough, high-relief topography were formed where stable depositional environments existed during disintegration of the stagnant ice, whereas less rugged morainic topography developed in the areas of thinner superglacial drift, where unstable depositional environments prevailed.

The Chippewa moraine, particularly east of the Chippewa River, consists of several stagnant-ice moraines. The moraine is, therefore, not the result of a single stillstand, as implied by Figure 1, and is, accordingly, referred to as a moraine complex in this study. Several hypotheses were proposed on the overall manner of formation of this moraine complex.

The morainic topography, in places, reflects the bedrock topography. Of special significance in this respect is the relationship of Flambeau Ridge, a prominent bedrock high situated on the inner margin of the moraine, to the overall morphology of the moraine complex. Flambeau Ridge played an important role in the formation of the morainic topography to its southwest, by far the most rugged and impressive part of the entire moraine complex. It was established that this section of the moraine complex is of interlobate origin. Hopefully, some of the points elucidated on the formation of this interlobate tract may provide a better understanding of interlobate moraines elsewhere.

In addition, this study yielded a substantial amount of additional information on the geomorphology of the area, including what appears to be the buried ancestral and very possibly the preglacial valley of the Chippewa River which extends southward from the junction of the modern Chippewa and Flambeau Rivers. The late-Pleistocene glacial history has

been interpreted and major conclusions are that: (1) at least some of the buried stagnant ice seems to have had a prolonged duration; (2) loess deposition appears to have ceased before disintegration of the buried ice had been completed; (3) the geomorphology of the Chippewa Valley indicates that the entrenchment of the Chippewa River to the Pre-cambrian surface had been accomplished even before all of the buried ice had disintegrated (wasted away by progressive fragmentation); and (4) the various stagnant-ice landforms differ in their relative ages--some formed during the early phases of glaciation while others may have formed as much as several thousand years later during the final disintegration of the better insulated stagnant-ice masses.

The Chippewa moraine complex is, in general, larger and more prominent to the west, than east, of the Chippewa River. Conditions that account for this situation are explained, with special reference to the presence of the rugged interlobate tract southwest of Flambeau Ridge.

Other findings include (1) a determination of the factors that controlled the distribution of loess deposits in the thesis area, and (2) the identification of several erratic Ordovician- to Devonian-age fossils, as well as of other erratic materials in the drift that may be useful in establishing the provenance of the material comprising the Chippewa moraine.

Recommendations for Future Study

Important problems meriting attention in the future are (1) the determination of the specific age of the glacial deposits of the area, (2) establishing the morphological and stratigraphic relationships of the Chippewa moraine with the moraines formed by neighboring glacial

lobes, and (3) determining the stratigraphy of the Pleistocene sediments of the area by means of deep drilling.

Other problems needing attention include (1) establishing the amount of time that stagnant ice existed within the Chippewa moraine, (2) ascertaining whether or not the ice-walled lake plains of the area contain fossiliferous sediments, and (3) conducting till-fabric studies of the stagnant-ice landforms as a means of determining the relative importance of ice-pressing in the formation of these features.

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