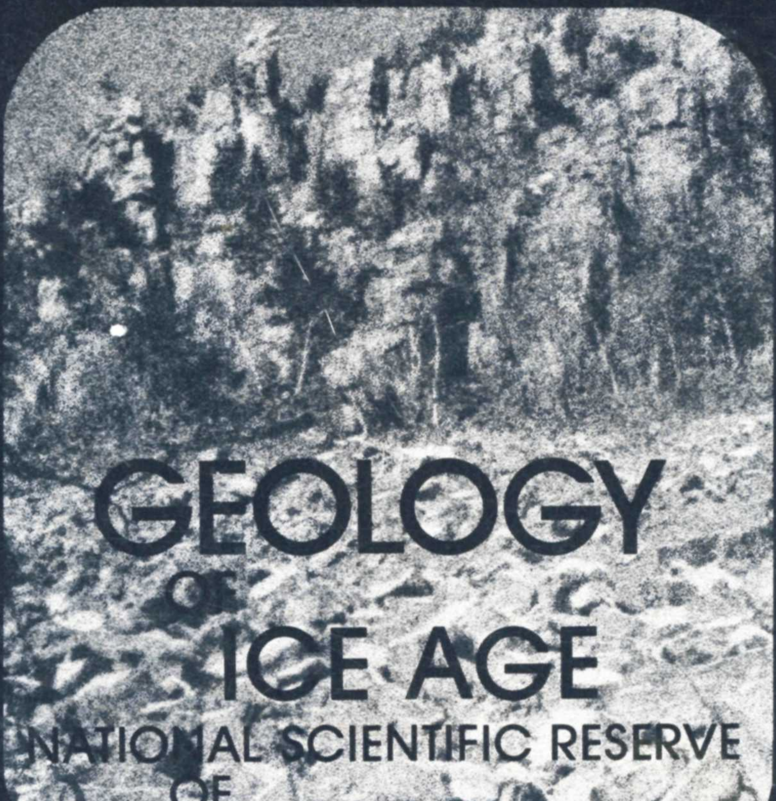

GEOLOGY OF ICE AGE NATIONAL SCIENTIFIC RESERVE OF WISCONSIN

GEOLOGY
OF
ICE AGE
NATIONAL SCIENTIFIC RESERVE
OF
WISCONSIN

Robert F. Black

University of Connecticut

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This publication is one in a series of research studies devoted to special topics which have been explored in connection with the various areas in the National Park System. It is printed at the Government Printing Office and may be purchased from the Superintendent of Documents, Government Printing Office, Washington, D.C. 20402.

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Preface

IN OCTOBER 1964, the United States Congress passed legislation charging the National Park Service, through the Secretary of the Interior and in cooperation with state and local governmental authorities, with the responsibility of formulating within 2 years a "comprehensive plan for the protection, preservation, and interpretation of outstanding examples of continental glaciation in Wisconsin." (Public Law 88-655). Through a contract[1] between the National Park Service and the University of Wisconsin, Madison, financial support was provided for a geological study of the Ice Age National Scientific Reserve. This contract provided money for some aerial photographs, field expenses, and other photographic supplies and reproductions. The study was started in February 1966, and the completed manuscript was submitted in June 1966.[2]

As expressed in the contract, the study areas include:

1. Eastern area (portions of the northern unit of the Kettle Moraine State Forest and Campbellsport drumlin area);
2. Central area (portions of Devils Lake State Park);
3. Northwestern area (portions of Chippewa County near Bloomer; and
4. Related areas (other areas in the State of Wisconsin which the Secretary of the Interior and the Governor of Wisconsin agree upon as significant examples of continental glaciation).

[1]Contract number 14-10-0529-2872

[2]Between the time of writing and printing of this book the pertinent literature on the Pleistocene as related to the Ice Age National Scientific Reserve has doubled. However, the general story has not changed, and no rewriting has been attempted.

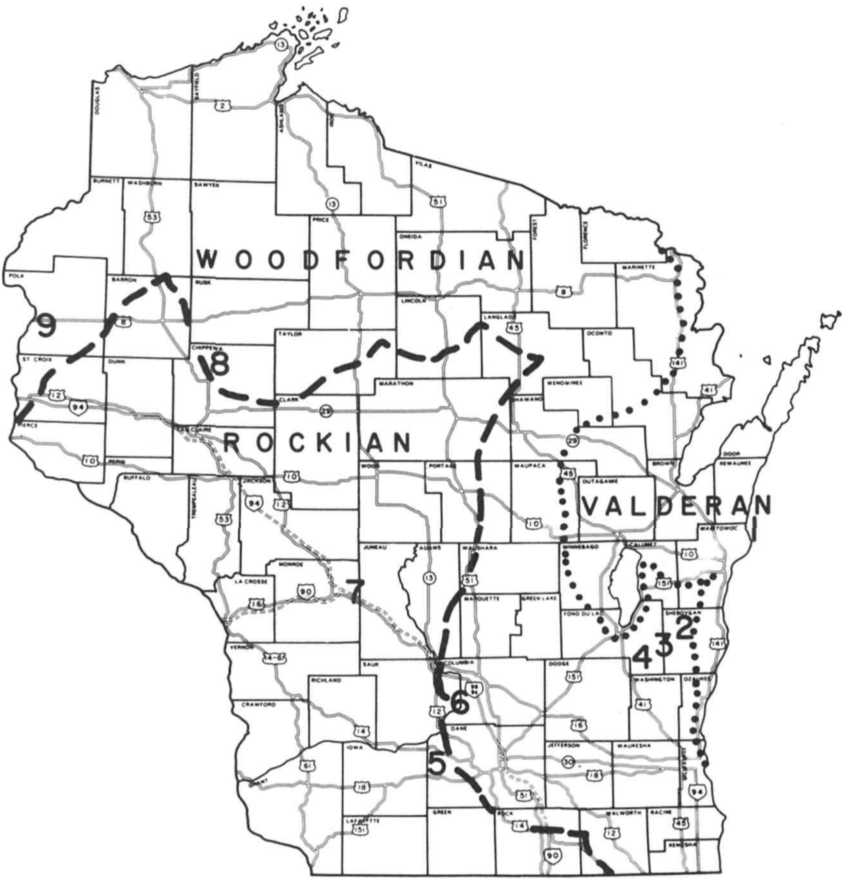

Fig. 1. County and road map of Wisconsin, showing nine areas recommended for inclusion in the Ice Age Scientific Reserve and their relation to substage glacial deposits. The nine areas are: 1) Two Creeks Forest Bed; 2) Sheboygan Marsh; 3) Northern Kettle Interlobate Moraine; 4) Campbellsport Drumlins; 5) Cross Plains Terminal Moraine; 6) Devils Lake Park; 7) Mill Bluff Pinnacles; 8) Bloomer Moraine; and 9) St. Croix Dalles Interstate Park.

The fundamental aim of the geological study of the Ice Age National Scientific Reserve was to provide basic data necessary for the formulation of a comprehensive plan for that Reserve, including:

1. The selection of representative areas and features that best illustrate continental glaciation in Wisconsin as outlined above;
2. The description and evaluation of each area selected with emphasis on the geological importance as it relates to the project purpose; and
3. The relationship of the areas and features selected in the Ice Age National Scientific Reserve to the entire scope of continental glaciation in the United States.

In consultation with representatives of the National Park Service and the Conservation Department of Wisconsin, the nine areas shown in Fig. 1 and described in Chapters 2 through 10 are recommended for inclusion in the Reserve. In certain instances the precise areas recommended for inclusion were determined not on their geology but on the bounds of practicality—some are "almost as good" geologically as other areas that are less favorably located geographically or are not already in public ownership. Some features are too large to buy or are not needed in public ownership to be preserved or appreciated. With available funds, or those likely to be forthcoming in the near future, it is not possible to include all desirable areas. Therefore, it is considered more important to spend available funds in acquiring ownership of important lands on the margins of existing state parks, such as Devils Lake, in order to preserve features there, than to start an entirely new area elsewhere which duplicates features in or adjacent to lands now in public ownership. Some of the more important features that, for a variety of reasons, cannot now be included in the Reserve are mentioned with reference to areas recommended for inclusion. It is hoped that these too will ultimately find their place in the Reserve.

This report begins with a brief review of the Pleistocene of Wisconsin as it is now understood specifically in relation to adjoining states. Detailed descriptions and evaluations of the geology of the recommended areas follow. A discussion of the Pleistocene of Wisconsin is included to show its relationship to the Pleistocene of the United States. A list of features and areas in Wisconsin that are recommended for later acquisition terminates this report.

Inasmuch as earth history is always an interpretation, and interpretors differ in their evaluations of so-called facts, no one particular interpretation may necessarily be right. Furthermore, new data are constantly forcing us to revise our story. Consequently, although coloring the report with my own interpretations and beliefs, I have attempted to point out the works of others wherein different points of view have been presented. In some instances I evaluated or rejected those views outright; others are as viable as my own. Different points of view or interpretation of past history based on what is now available to us of those ancient events is not to be deplored. Without the possibility that anyone will come up with a new fact or interpretation that provides the key to an enigma, few tourists would truly savor the Reserve, and it is through tourists and landowners, as well as scientists, that new data on the geologic past come to light every day.

In a sense this book is only a status report of the general and specific geologic setting of the recommended areas as we now know them. Much remains to be learned. A complete discussion of the earlier literature or the infinite detail of a particular area are not warranted. Appropriate basic references are cited to lead the way to others. The general story for each area is made sufficiently complete for the uninitiated reader to fit each area into the broad framework of the state and region. The description of individual areas varies in detail from one to another depending in part on the complexity and importance of an area and in part on my acquaintance with them. Devils Lake Park is emphasized because of its importance and need of expansion.

Since 1956, I have received financial support for research on the Pleistocene of Wisconsin from the Research Committee of the Graduate School of the University of Wisconsin from funds supplied by the Wisconsin Alumni Research Foundation, from National Science Foundation Grant GP-2820, from the State Highway Commission, and from the State Geological and Natural History Survey. Without the research background provided by that support, this study would not have been possible in the time allotted. I was assisted unstintingly in some field studies and in the typing, compilation, and editing of this book by my wife, Hernelda L. Black.

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1
The Pleistocene of Wisconsin

General Statement

In 1839 Charles Lyell named the Pleistocene (Gr. *pleistos* most + *kaines* new) because the life of that epoch constituted the closing stage of the transition from the geologic past to the present. Usage by many geologists has equated the Pleistocene, however, with the last Great Ice Age. Inherent with both the biologic and glacial approaches, as a common denominator, is the concept that climates over most, if not all, of the globe were distinctly cooler during the Pleistocene than the many tens of millions of years of geologic time that preceded (Emiliani 1954, 1955).

To understand fully the Pleistocene of Wisconsin we need to be cognizant of problems of nomenclature, of local physical and biological minutae, and of global and even astronomic events. This makes for a somewhat disjointed summary in order to avoid undue length, but many topics need no elaboration because the standard textbooks on the subject provide the necessary background (e.g., Flint 1957).

The generally accepted classification of the Pleistocene for North America was developed over the last 70 years in the Upper Mississippi Valley, including Wisconsin. It is now under attack by investigators elsewhere using various approaches, particularly on ocean sediments. Changes are now almost a daily occurrence in our classifications and many supposedly firm cornerstones probably will be overturned soon.

To this day authorities differ on the duration of the Pleistocene, in part because of different criteria used in defining its boundaries, but also because of differences in interpretation of available data (e.g., Ericson et al. 1963, 1964; Dreimanis 1964; Akers 1965; Krinsley and Newman 1965; Damon 1965; Flint 1965). Biologic changes and glaciation are both manifestations of climate, but they are time transgressive and are not necessarily synchronous. Clearly the biologic assemblages of that part of typical early Pleistocene that was dominated by glaciation in the north temperate latitudes had markedly earlier origins (Evernden et al. 1964). With notable exceptions, where transition beds of the Pliocene-Pleistocene boundary are recognized, no abrupt change (Flint 1965) occurs in either faunal or floral remains (Deevey 1965) or in related paleo-temperatures (Emiliani et al. 1961).

Thus, physical events and biologic changes are not yet integrated on a world-wide basis acceptable to all. Several frameworks of marine and terrestrial events are emerging. The disparity of chronologies between states or within continents, from continent to continent, or continent to ocean basin is striking (Wright and Frey 1965). It is too early to know whether we will end up with a Pleistocene Epoch based on fossils whose first part is characterized by a cooling but non-glacial climate in areas (outside Antarctica) where glaciation dominated the latter part, or whether the Pleistocene will be restricted to that time when glacial events characterized the land areas in the north temperate regions. The Pleistocene could include well over one million years (Ericson et al. 1963, 1964; Deevey 1965) or be only a few hundreds of thousands of years (Emiliani 1955, 1961).

The manifold arguments for the use of climate, biologic changes, or glaciation in defining the Pleistocene are beyond the scope of this book. For convenience the appropriate terms in the classical chronology of fourfold major stages of glaciation and threefold major interglacial stages will be used. Nonetheless, the rumblings one hears more frequently (and rightly so) that this classification must be modified (e.g., Deevey 1965) should be kept in mind. The glacial stages are named for particular glacial deposits in the respective states—Nebraskan, Kansan, Illinoian, and Wisconsinan—from oldest to youngest. The Wisconsinan Stage was originally named East Wisconsin by T. C. Chamberlin (Geikie 1894:763), shortened by him to Wisconsin in 1895, and altered by Frye and Willman (1960) to the adjectival form to make it consistent with common usage of the other stages.

The Wisconsinan Stage is aptly named for we cannot identify with certainty any non-reworked or non-buried Pleistocene deposits of pre-Wisconsinan age in the state (Black 1962).[1] The area of west-central Wisconsin, where widespread pre-Wisconsinan deposits were shown for several decades on glacial maps, has only recently been reevaluated (Black 1959a) with all surface deposits being correlated with the Wisconsinan Stage. Whether the older materials were never deposited, or, if deposited, have been removed remains debatable although both processes seem to have operated locally. Reworking of older Pleistocene deposits by Wisconsinan ice is recognized in a few places, and

[1]Very local pre-Wisconsinan deposits have since been identified outside the areas selected for the reserve; see Black, Robert F., Ned K. Bleuer, Francis D. Hole, Norman P. Lasca, and Louis J. Maher. 1970. Pleistocene geology of southern Wisconsin. Univ. Wis. Geol. and Nat. Hist. Surv. Inf. Circ. No. 15.

overrunning without removal of older deposits by younger ice also seems to have occurred. Hence, some deposits in Wisconsin now correlated with the Wisconsinan Stage may prove to be older when dating methods improve or new evidence appears.

Conversely, no unmodified landforms in Wisconsin can be said truly to be older than the Pleistocene. Unquestionably the oldest surfaces in Wisconsin are those in the classical Driftless Area of the southwest. Clearly that area was not invaded by ice of the last two major substages, the older of which covered most of the rest of the state. However, the former correlations of upland surfaces in southwest Wisconsin with early Pleistocene and pre-Pleistocene peneplain remnants (Trowbridge 1921; Bates 1939; Horberg 1946) have been discredited (Martin 1932; Thwaites 1960; Palmquist 1965).

At this time the Reserve includes part of the border of the Driftless Area. However, the early events recorded in the Driftless Area not only are of local concern but, as we shall see, affected the rest of the state as well. For example, the dissection of the uplands of southwest Wisconsin can be assigned tentatively (Palmquist 1965) to an erosion cycle comparable to one in Illinois that Frye (1963) concludes followed the Nebraskan Stage but preceded the Illinoian Stage. This cycle affected all the state, even though the evidence largely lies buried under drift outside the Driftless Area. Such problems are complicated and will be cited only with respect to the areas concerned.

Glacial Deposits of Wisconsin

The glacial deposits of Wisconsin, classified as end moraines, ground moraine, pitted and unpitted outwash, and lake deposits are shown in Fig. 2. This is a very generalized map prepared by Thwaites from data available in the files of the University of Wisconsin Geological and Natural History Survey, augmented by published geologic literature and from his interpretation of soils maps. Thwaites was the first to admit that in places the map represented only "guess work", but it still is the best available.

End moraines are deposited essentially at right angles to the direction of flow of ice at that margin and can be used to depict the lobation of the ice that formed them. Ground moraine was left behind the fronts in part as ice advanced and in part during stagnation and destruction of portions of the ice sheets. Outwash was laid down by streams outside the ice fronts, except where pitted. Pitted outwash clearly demonstrates buried ice masses of the same or an earlier advance were buried under the stream deposits (Thwaites 1926).

A very marked lobe of ice occupied the lowland of Green Bay-Lake Winnebago following the strike of

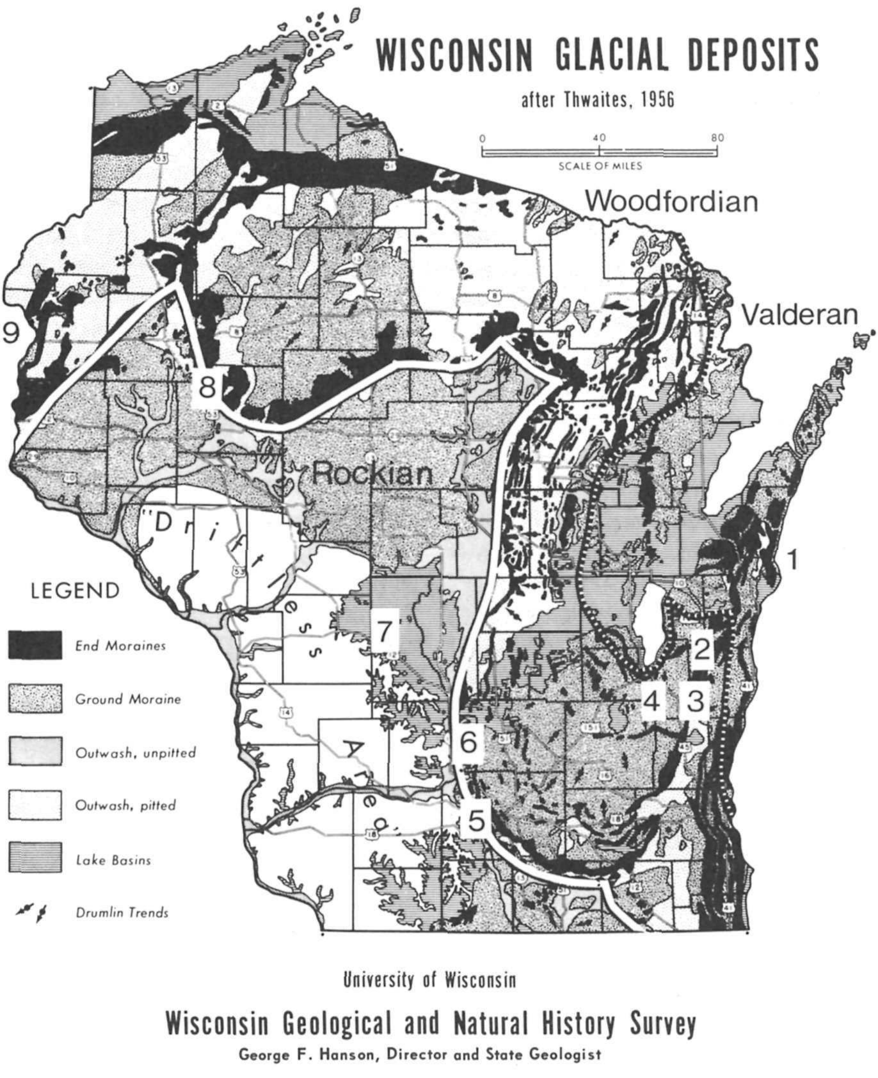

Fig. 2. Glacial deposits of Wisconsin and the nine areas recommended for inclusion in the Reserve.

the Platteville-Galena Group of dolomite (Ordovician, Fig. 3). The deployment of weak shale under the resistant dolomite of the Niagara Formation of Silurian age, both dipping gently eastward, led to the excavation of the lowland in large part by ice action. Thus, ice deployment was determined in part by rock type which controlled topography and also in part determined what the topography was to be by virtue of its ease of removal. In all instances the bedrock over which the ice traveled determined the composition of the drift deposited later.

Ice followed the lowland routes, but still was thick enough to fan out normal to the concentric end moraines depicted in Fig. 2. Where ice from adjacent lobes butted against each other, interlobate moraines formed. The classic example is that between the Green Bay Lobe where it moved southeastward and the Lake Michigan Lobe where it moved westward—the world-famous Kettle Interlobate Moraine.

The end moraines mark fronts which represent distinct spans of time during which loss of ice (ablation) along the front equalled the resupply by the various processes of ice motion (Kamb 1964). Additional debris thus was being brought forward by the ice movement and dumped at the same site along the front. This requires a very delicate balance between the ice and its environment. On the scale used in Figs. 1 and 2 it is not possible to distinguish segments of end moraines as single entities where time differences are small. We know that the outer limit of the Green Bay Lobe, as depicted in Fig. 2 by relatively continuous end moraines, is actually a series of short segments of different lengths representing different times or short-term local pulsations of the ice. Such pulsations leave cross-cutting end moraines which can be shown only on maps of much larger scale. Some pulsations probably represented many centuries.

My approximation of the substage classification of the deposits, ignoring short time spans and lake deposits which are marginal and at least in part synchronous with the adjoining ice sheets, is shown in Figs. 1 and 2. All the deposits are correlated with the Wisconsinan Stage, two contrasting subdivisions of which are shown in Table 1 which also shows the generally recognized time of events in thousands of years B. P. (before the present or more exactly before 1950). A combination of terms from both classifications will be used in this paper even though they pose a problem in nomenclature for which no solution is immediately at hand. An additional term, Rockian, is used for latest Altonian time (Black 1962). A discussion of these problems, some of the reasons for the interpretation of time of events in Figs. 1 and 2, and other facets of the Pleistocene of Wisconsin follow after the description of individual

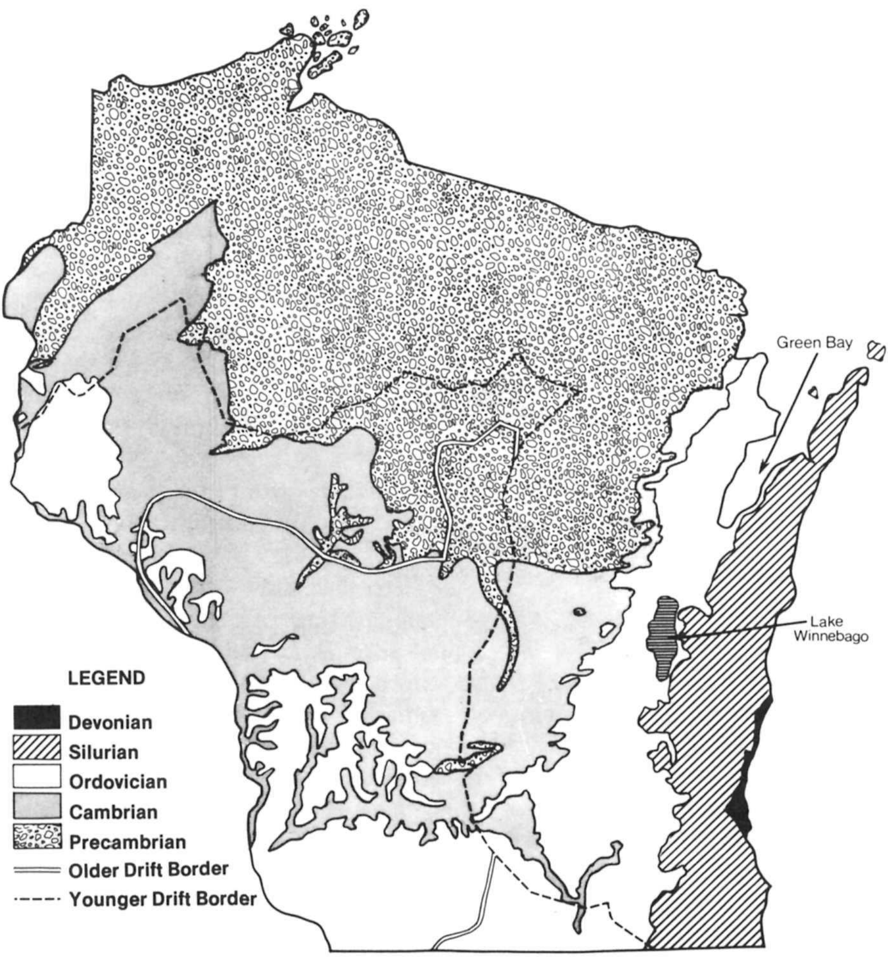

Fig. 3. Generalized geologic map of Wisconsin (after Bean 1949).

Table 1. Two contrasting classifications of the Wisconsinan Stage.

Thousands of years B. P.	Frye and Willman 1960: Glacial	Frye and Willman 1960: Nonglacial	Leighton 1960: Glacial	Leighton 1960: Nonglacial
0		Recent		Recent
5	Valderan			
10			Valders	
		Twocreekan		Two Creeks
			Mankato	
				Bowmanville
	Woodfordian		Cary	
15				St. Charles
			Tazewell	
20				Gardena
			Iowan	
25		Farmdalian		Farm Creek
30		[Rockian of	Black 1962]	
	Altonian		Farmdale	
35				
70				

areas. It is hoped that this discussion will provide further background for evaluation of the geology of individual areas and for fitting them to the regional format. More than that, however, it is hoped that the reader will be intrigued to look further into the many problems himself for possible alternative explanations.

Recommended Areas

Nine specific areas (Figs. 1, 2) are recommended for inclusion in the Ice Age National Scientific Reserve of Wisconsin. These are discussed for convenience in the order recorded by representatives of the Wisconsin Conservation Department and of the U.S. Department of the Interior in their various progress reports. They are:

1. Two Creeks Forest Bed
2. Sheboygan Marsh
3. Northern Kettle Interlobate Moraine
4. Campbellsport Drumlins
5. Cross Plains Terminal Moraine
6. Devils Lake Park
7. Mill Bluff Pinnacles
8. Bloomer Moraine
9. St. Croix Dalles Interstate Park

Detailed descriptions of boundaries of these areas, acreage, ownership, cost of acquisition, etc., have been covered in reports by others. In this report the boundaries shown in maps and sketches accompanying the description of each of these areas are predicated, within reasonable bounds of practicality, to preserve the most important features rather than all the desirable features. As such, these are minimal boundaries. Minor variations in order to follow property lines might be made locally without harm, but it is emphasized that general constriction in the boundaries can be made only to the detriment of the Reserve.

As stated in the preface, not all kinds of features attributable to the Ice Age are to be found in these nine areas. To include them all would mean an enlargement of the Reserve that would carry it to all parts of the state. It is hoped that this can be done ultimately. The recommended areas have features considered representative of many in the state, even though they are not necessarily the best features to be found there. Again, the bounds of geographic location, cost of ownership, and completeness of the geologic story to be told in a local area all played a role in the final selection of the specified areas. From the point of view of use by tourists, the Northern Kettle Interlobate Moraine, Devils Lake, Bloomer, and St. Croix Dalles areas are considered the four most important because of their size, accessibility to large metropolitan areas, and variety of features. Of these Devils Lake is by far the most important on all counts—the variety and uniqueness of its geology, scenery, fame, etc. Trebling the size of the present state park is mandatory if we are to preserve a representative sample of the variety of glacial features the area has to offer. This must be done now—not a few years from now—to avoid their loss. The recommended area is absolutely minimal; anything less will be much "too little too late."

In connection with each area, representative similar features elsewhere in the state are cited where appropriate or known. At the close of this book other features or areas of marked interest are mentioned in hope that someday they too will become a part of the Reserve or at least that their owners will treat them in the best tradition of conservation and not exploit them to the loss of mankind.

2
Two Creeks Forest Bed

Introduction

In the extreme northeast corner of Manitowoc County (Figs. 1,4,5), lying to the east of Highway 42 and along the shore of Lake Michigan, is a portion of the world-famous Two Creeks Forest Bed (the type section for the intraglacial substage of Two-creekan age, Table 1). This is an immature soil horizon or detrital organic zone with forest litter buried in lacustrine sediments which in turn are covered by glacial till and underlain by glacial till. Lake sediments locally also lie on top of the younger till. Few Pleistocene sites in the United States have gained more prominence in the recent literature than this forest bed. It provides field evidence of multiple glacial advances and retreats, of intraglacial conditions, and is a world-famous geochronological site. It is dated at 11,850 years B.P. on the average by radiocarbon (Broecker and Farrand 1963) but 19,000 years B.P. by varve analysis (Antevs 1962).

General Statement

A diagrammatic sketch of the Two Creeks horizon in the recommended area (Fig. 5) is shown in Fig. 6. A schematic section of the Lake Michigan bank is shown in Fig. 7. Representative photographs of the exposed section and of the Two Creeks horizon are shown in Figs. 8-11.

In the primary recommended area from the Manitowoc-Kewaunee County line southward, the Lake Michigan bank rises abruptly from a narrow shore to heights of 25-30 ft above lake level. The soil horizon is undulating in the bank, rising from lake level near the county line to a general level of 10-20 ft above lake level for a distance southward of about 500 ft before dropping to lake level again. Some of the undulations are attributable to initial relief on the landscape, but most result from disturbance by the overriding Valders ice. In overriding, the ice removed the apices of some undulations.

In the bank the lowermost till is clayey, locally gray or red, and compact. The overlying lacustrine deposits vary texturally from silt and clay to medium and coarse sand and locally, gravel; color is red, brown, yellow, and gray. Some bits

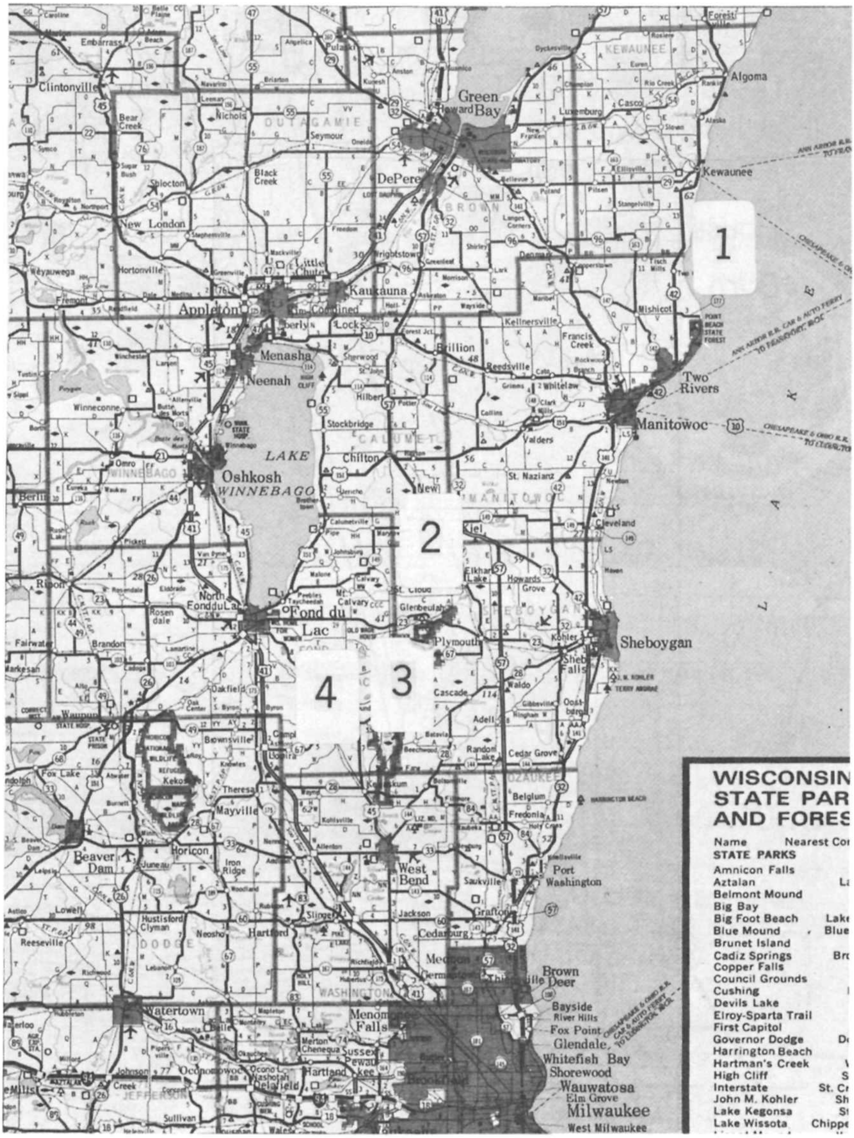

Fig. 4. Part of Wisconsin State Highway map, showing locations of areas 1-4.

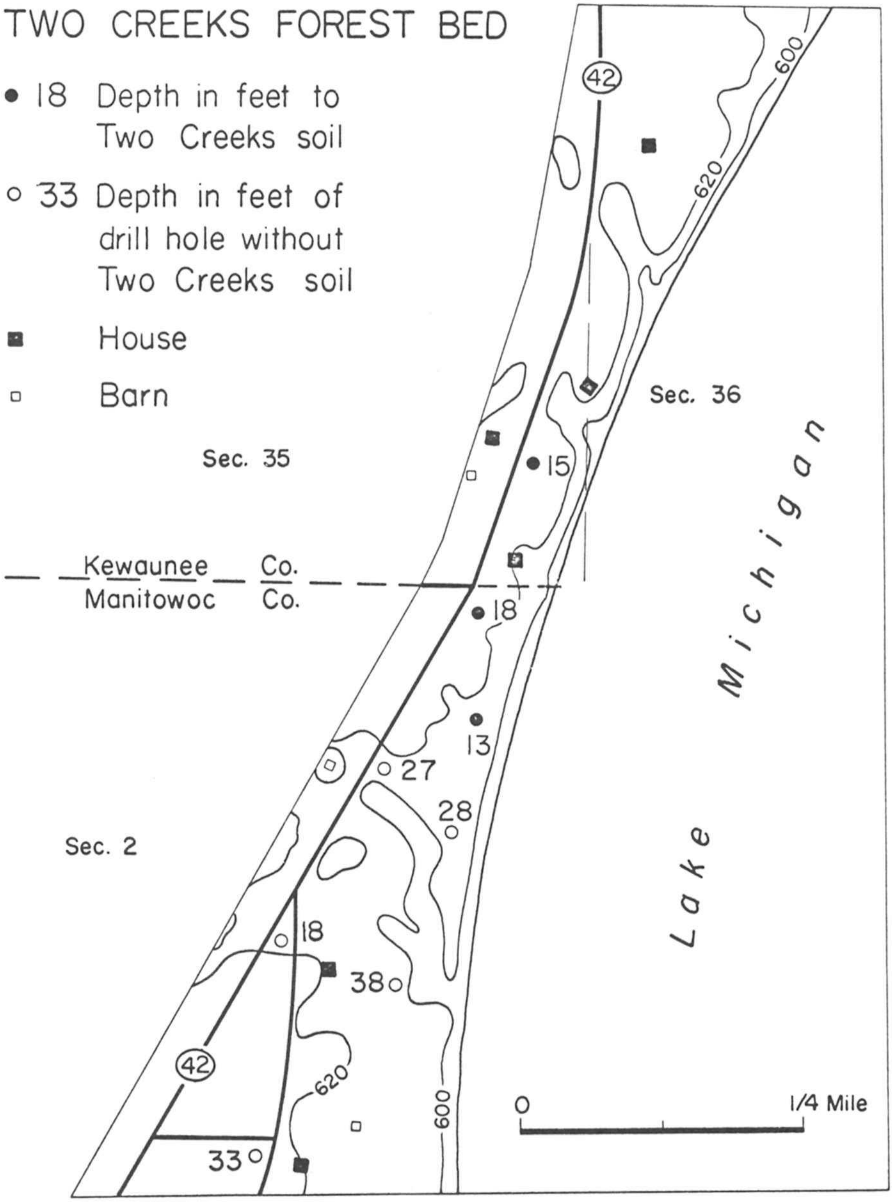

Fig. 5. Topographic map of the Two Creeks Forest Bed locality showing location of drill holes.

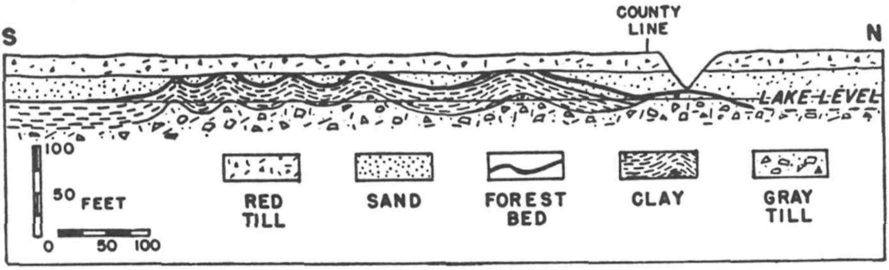

Fig. 6. The Two Creeks Forest Bed and associated deposits in the recommended area. After Thwaites and Bertrand (1957, Fig. 12).

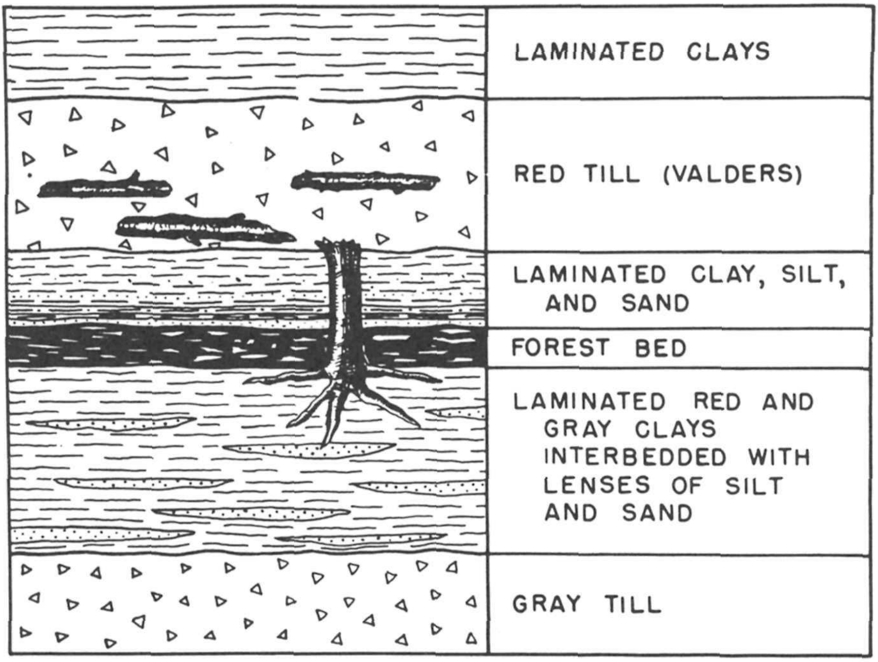

Fig. 7. Idealized Two Creeks section. After Prouty (1960, Fig. 12).

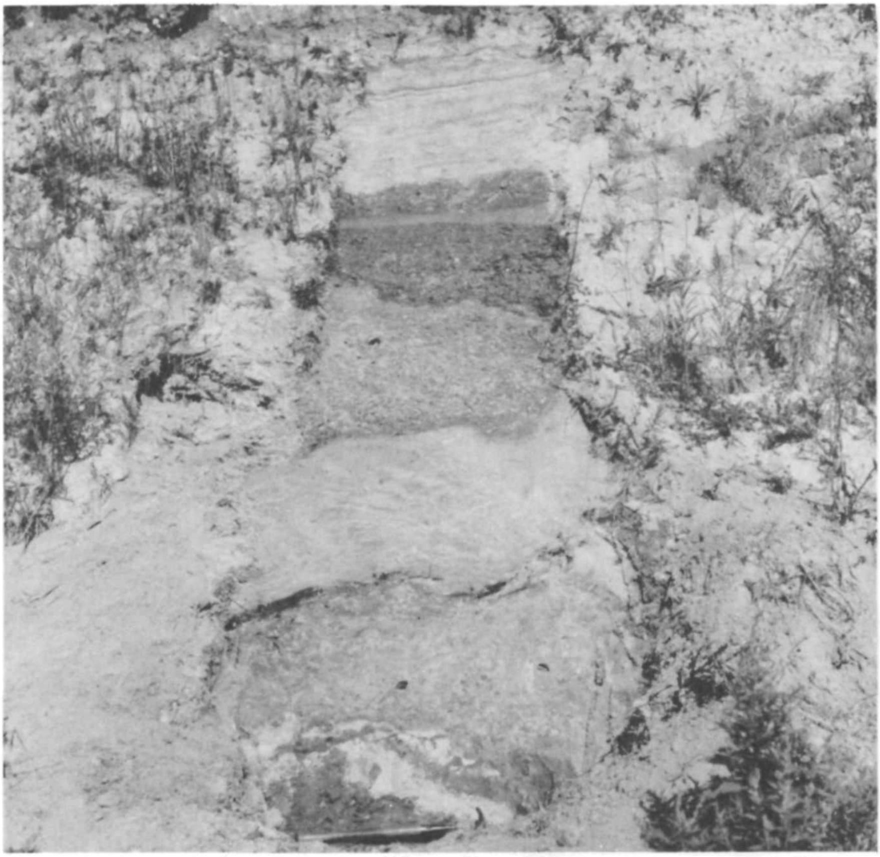

Fig. 8. Two Creeks section from bottom up of red till, lacustrine sand, organic Two Creeks horizon, lacustrine sand, red till, and lacustrine sand at top. Army entrenching shovel at base is 18 inches long.

of wood are present in the lacustrine sand below the forest bed proper. The forest bed is locally an *in situ* soil profile a few inches thick or a detrital zone of organic matter, dark brown to black, with easily recognized needles, cones, twigs, and logs mostly of spruce. Most early investigators who worked there, when Lake Michigan water levels were higher than in 1966 (Thwaites and Bertrand 1957)(Fig. 11), saw stumps rooted in the horizon. Some rose to the top of the overlying lacustrine

Fig. 9. Detail of Fig. 8, showing Two Creeks horizon marked by 35 mm film cassette.

deposits, with much detrital organic matter, where they were cut off by the overriding Valders ice. It left red clayey till on top of the lacustrine beds that was similar to the red till below, but with numerous logs and other organic matter. In turn it is covered locally by a few inches to a few feet of yellow-brown and red-brown lake sand and local dunes. Details of the sections vary from point to point along the bank. The lacustrine beds particularly may be absent or are locally as much as 10 ft thick. Depth to the forest bed in drill holes is shown in Fig. 5 by solid dots with numbers; open circles show depth of holes in which I did not encounter or recognize the forest bed.

If later expansion of the primary area is contemplated, exploration northward in Kewaunee County is urged. I have not seen the forest bed exposed along the shore in Kewaunee County, but I found it in a drill hole along Highway 42. It was not found south of the recommended area either along the shore or inland.

Detailed Description

Whittlesey (Owen 1852:463) first mentioned the presence of wood in a well at Appleton on the Fox River between Fond du Lac and Green Bay in east-central Wisconsin. Goldthwait (1907:61), while studying abandoned shorelines in eastern Wisconsin, was the first to find and record in print exposures of the organic soil horizon with logs near Two Creeks. The initial find was south of the recommended area but it has been largely destroyed by lake shore erosion. What remains is too close to water level (0-3 ft) to develop.

Goldthwait (1907:61) wrote:

> Two miles south of the village of Two Creeks (in section 24) the freshly cut lake cliff showed in July, 1905, a remarkable cross-section of an interglacial forest bed. Laminated red clays formed the base of the section, up to two or three feet above the water. Above this, and separating it from a twelve-foot sheet of stony red till was a conspicuous bed of peat, sticks, logs and large tree-trunks, which unmistakably represent a glaciated forest. . . . The till immediately above the forest bed, besides containing characteristic subangular striated stones and red clay similar to the clay in the stratified beds below, all absolutely unassorted, was plentifully mixed with broken branches and twigs. In the underlying forest bed the stumps were well preserved, the wood being soft and spongy like rotten rubber, but retaining all the appearance of its original structure. Several logs and stumps lay pointing significantly towards the southwest, the direction in which the ice sheet probably moved at this place. One little stump, however, . . . with its ramifying roots firmly fixed in the laminated red clays, stood erect as when it grew there, but it had been broken short off at the top, where the ice sheet, dragging its ground-moraine along had snapped off the top without uprooting the tree. Around each root the red clay was discolored to a light drab, showing the effect of acids derived by decay, in contact with the iron-bearing clays. There was no mistaking the only half-excavated condition of the deposit. Clearly this surficial sheet of red till records a final advance of the ice sheet over a surface of laminated red clays, which here, at least had been clothed with a forest. The trees were broken and generally overturned by the ice, and buried beneath the twelve-foot sheet of till. The wonder is that so much of the over-ridden forest should be preserved, and at least one stump in it remain erect.

Goldthwait (1907:59) also mentioned that at one point in the vicinity of Manitowoc a peat bed 3 ft thick formed the upper part of the 10 ft cliff, with laminated clays containing sticks and branches below. In another place a bed of old logs and sticks lay buried beneath 15 ft of clay, near the base of the cliff. Thus were recorded in part, the location and description of two segments of the Two Creeks Forest Bed.

Fig. 10. Two Creeks organic horizon of detrital twigs, needles, cones, etc., with pencil for scale.

Fig. 11. Valders till with logs and holes where wood has just been removed.

No immediate study was made of the forest bed although Thwaites visited the area several times between 1922 and 1930. Wilson (1932) undertook a preliminary investigation and later amplified his work (Wilson 1936). He first studied the forest bed where it was exposed for 0.5 mile along the lake shore in secs. 11 and 13, T. 21 N., R. 25 E. He also mentioned that the same forest bed was exposed 3 miles to the north on the lake shore and in a ravine about 0.25 mile to the west in sec. 35, T. 22 N., R. 24 E., Kewaunee County.

Wilson (1932) studied closely the forest bed for only about 100 ft along the lake shore and through a vertical range of only several inches. He interpreted the forest bed to lie on top of varved clays and silts and under additional lacustrine silts and sands deposited between the retreat of ice and its readvance which laid down till on top. Locally the lake beds were 12 ft thick. The till on top of the lacustrine sediments was about 8 ft thick. At the site most of the wood has been spruce (*Picea mariana* and *P. canadensis*) and hemlock (*Tsuga*). The wood is soft and easily broken and checks and breaks into short sections on drying. Tissues, however, are not destroyed and microscopic sections can be made of them. Where wood and peat are in contact with the red till, there is a zone in the clay a few inches wide of greenish gray color due to deoxidation of the iron. The logs occur most frequently in the lacustrine sediments directly above the forest bed, but are also in the overlying till. Wilson found one stump *in situ* with the butt of the broken log almost attached. The roots of this stump extended along the forest bed peat. A bracket fungus was found on a portion of the root that had been exposed above the ground during the interstadial period. The bracket fungus is *Polyporus* but the species was not determined. All the logs that had not been broken by subsequent handling showed ragged splintered ends as a consequence of glacial action.

Wilson (1932) studied the growth rings in sections of six logs. The greatest number of rings in one section was 82; the average was about 60. Five of the logs showed by the width of successive rings a marked decrease in the rate of growth in the last 12 years of the Two Creeks site. One log taken from the red till directly above the forest bed showed little decrease until the last year of its growth. That particular log was white spruce (?), *P. canadensis*. The first five logs were considered by Wilson to represent the growing conditions of the forest bed at the site whereas the log taken from the till above was considered to have been transported by ice from a different environment farther north. When the log taken from the red till was compared with the others, an extreme difference in size and growth rate was noticeable.

That log was twice the diameter of any of the others although it had only the average number of growth rings. The width of the rings did not agree with that of the forest bed trees.

The growth rings could not be compared exactly with reference to particular years, because it was not known whether all the trees were destroyed in the same year or whether they were all alive at the time the ice advanced. However, Wilson considered it probable that the largest log, having been brought in by the ice, was felled several years before the trees in the site studied.

Detailed study of wood sections by Wilson showed that certain small rings of the forest bed trees occurred at years approximately corresponding to those in which wide growth rings occurred in the transported log from the overlying till, and vice versa. If excessive moisture was one of the primary factors for small growth rings in the trees, as is suggested by the character of the flora and fauna, then trees growing in higher ground would not have been similarly affected and probably would do better in wet years.

The moss floor of the forest bed comprised the most extensive group of plants found in the remains. The moss material, identified by L. S. Cheney (Wilson 1932:38), was divided into 19 species. All the mosses are of existing species but are in general more northerly in their modern distribution than the Two Creeks Forest Bed location. Nearly all are found in northern Wisconsin, but the present southern limits of a few are in Canada.

Peat in the forest bed was poorly formed and in some parts of the exposure was wanting entirely. Wilson (1932) concluded from this, as well as from some other organic remains, that the Two Creeks Forest Bed was not exactly a lowland forest but rather a dry forest at one stage of its existence. In places the mosses and other plant remains accumulated as a silty peat such as can be found in any spruce forest today. It is from this peat that the microfossils were secured.

Seven species of mollusks were identified from the forest bed by F. C. Baker to whom Wilson sent specimens. These were from three levels in the forest bed. They agree ecologically with other organic remains from their respective horizons. One Pleistocene form was reported—from the clay immediately beneath the forest bed. The individuals in higher levels represented existing species.

The mollusk *Fossaria dalli* (Baker) was considered Pleistocene on the basis of its large size. Its habitat was wet mud above water. Two other species of mollusks, *Pupilla muscorum* (Linn.) and *Succinea avara* (Say.), represent forest forms and suggest arrival of trees at the same time as a few grasses and mosses. Directly on the surface of the clay occur spruce cones, needles, and forest mosses mixed. Mixed with the mosses are shells of land mollusks *Succinea avara* and *Vertigo ventricosa* (Morse). One moss was peculiarly restricted to the lowest level of the forest bed. This is *Bryum cyclophyllum* (Schwaegr.) —a forest form that seems to have been first to establish itself on the Two Creeks Forest floor. Other plants that appear in this horizon are grasses, heaths, birch, jack pine (*Pinus banksiana* Lamb.), and a species of spleenwort (*Asplenium*). These are represented only by a few pollen grains and spores. Fungi were abundant; some were lichens and others were representative of *Dematicae*. Bark beetle excavations were found on the logs and may represent two genera.

Culberson (1955) with the aid of W. C. Steere found eight species of mosses which are associated with floras of more northern affinities.

Wilson (1932, 1936) thus recorded an early phase with aquatic and semiaquatic mollusks, an intermediate phase with moist to dry woodland mosses, and a final phase of flooding with aquatic mollusks and mosses. The pollen spectra in Fig. 12 by West (1961) give additional details of the vegetational changes associated with these changes in water level. The abundance of soapberry (*Shepherdia canadensis*) pollen at the base of the sediments indicates early colonization by this shrub of the land surface exposed by the lowering of the lake level (West 1961). *Shepherdia canadensis* is a northern and mountain plant found in forest clearings and on sandy shores particularly in the boreal spruce forests. Thus the plant's behavior at the beginning of the Two Creeks interval exactly parallels its present behavior. The phase with *Shepherdia* was short lived, and *Picea* forest succeeded the pioneer community as indicated by the high frequencies of *Picea* pollen. White spruce dominated over black spruce. The flooding of the forest litter by the upper silts was accompanied by a large decrease in pollen frequency. At the same time the non arboreal pollen total, including *Ambrosia, Artemesia,* and composite, rises slightly. This may reflect the opening out of the regional forests associated with the Valders readvance, but there is also the possibility that some of the pollen in the silts are secondarily derived.

The vegetation of the Two Creeks interval is clearly boreal in character. Originally Wilson (1932) compared its climate with that of northern Minnesota today. In his later paper (1936) he suggested that the climate was not necessarily as severe as this, for the plants also represent pioneer organisms of denuded areas under certain conditions and are not reliable indicators of a severe climate. West (1961) finds the interpretation of the pollen profiles from this and other nearby sites to be complicated and open to more than one interpretation. He concludes,

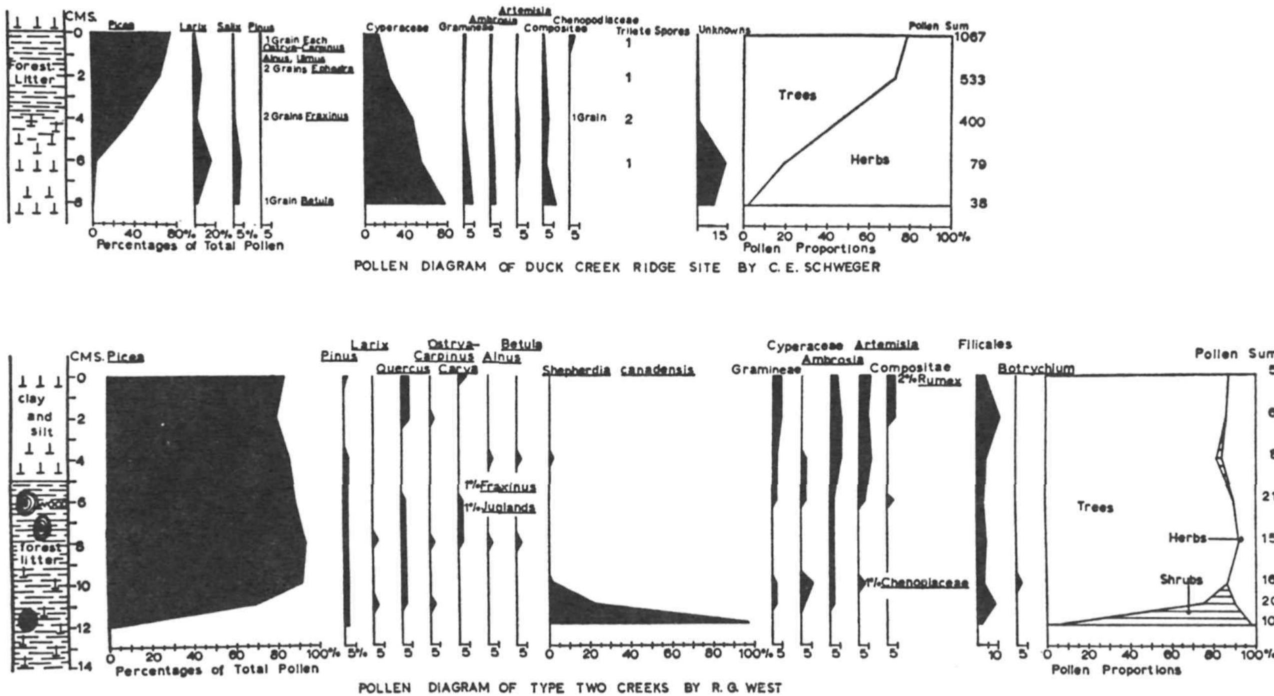

Fig. 12. Pollen profiles of Twocreeken material. After Black et al. (1965, Fig. 8-2).

however, that the spruce forest was able to survive along the margin of the Valders ice, although with openings. At least the climate of Two Creeks time was not necessarily much more severe than that of today in the area (Schweger 1966)—only simple pioneering boreal wetland species are present. Roy (1964) in a study of the Pleistocene non-marine mollusca of northeast Wisconsin concluded also that these species represented climates very similar to that of northern Minnesota today.

Pollen studies of other sites of Two Creeks age in Wisconsin have also been done by West (1961), by Schweger (1966) *in* Black et al. (1965:56-81) (Fig. 12), and by others whose work has not been published. Local variations in the forest litter and macrofossils at these various Two Creeks locations also are appearing (Black et al. 1965: 56-81).

Wood fragments of Twocreekan age are especially common in the Valders till, but locations in eastern Wisconsin where Twocreekan soil profiles are *in situ* are less common. Particularly good exposures have been seen in borrow pits in the SW¼ and NE¼SE¼ sec. 19, T. 23 N., R. 19 E., Outagamie County (Piette 1963). Another is in the SE¼NW¼ sec. 15, T. 22 N., R. 15 E. Detrital organic litter of Twocreekan age in lacustrine sediments is found at several places, such as the SE¼SE¼ sec. 22, T. 24 N., R. 21 E., Brown County, and the SW¼SW¼ sec. 6, T. 21 N., R. 23 E., in Manitowoc County. All these sites are in borrow pits, are of very limited extent, and do not lend themselves to use by the public. The type section remains unique.

The controversy of the varve-dated chronology calling for Two Creeks to be 19,000 years old (Antevs 1962) versus the radiocarbon dates of 11,850 years (Broecker and Farrand 1963) requires that we examine information available for much of northeastern United States and Canada as well as the European transatlantic correlations. This goes far beyond the scope of this book. Suffice it to say that the radiocarbon-controlled chronology has been accepted by a majority of workers.

When it is recognized that the forest bed is established on lacustrine sediments and yet is covered by lacustrine sediments, all of which in turn lie between two tills, something of the magnitude of the glacial history inferred becomes apparent. To this we must add still more lacustrine sediments and wind blown materials on top of the younger till at the Two Creeks Forest Bed locality. This means we must take into account at least three lakes whose levels have been up to a point more than 30 ft higher than that of present Lake Michigan. One lake followed the basal till, one swamped the forest bed horizon, and one came in on top of the younger till. These fluctuations are of an order of magnitude beyond that which can be achieved merely by increasing precipitation. Changes in the outlet or outlets of Lake Michigan were involved. We cannot confine our analysis of this problem only to a study of Lake Michigan. All of the Great Lakes (Hough 1958) must be taken into account and their story integrated with the Pleistocene history of the St. Lawrence, Hudson, and Mississippi river valleys.

Precise correlation of the age of the till at the base of the cliff has not been made. It is certainly late Woodfordian, possibly a younger unit of the Cary or subsequent slight readvance such as the Mankato or Port Huron of other states. The till locally is gray but I found mostly red. Gray drift is supposedly characteristic of the Port Huron of Michigan (Wayne and Zumberge 1965:63-84), whereas red drift that is post-Cary and pre-Two Creeks is generally considered representative of the Mankato of Minnesota (Wright and Ruhe 1965:29-41). The Port Huron Moraine (Wayne and Zumberge 1965:63-84) was described by Taylor (Leverett and Taylor 1915:293) as "one of the best developed and most clearly defined moraines in the Great Lakes region" and this status has been accorded this moraine by every glacial geologist who has worked in Michigan since that time. The Port Huron Moraine was dated by Hough (1958:278) at 13,000 years B.P. It was correlated across Lake Michigan by Thwaites and Bertrand (1957) (Fig. 1) with an unnamed moraine near Sheboygan, Wis. Presumably the Port Huron then would extend to the north and encompass the Two Creeks site.

If the above correlation is correct that the basal till at the Two Creeks Forest Bed is of Port Huron equivalent, then the history of the lake sequence would begin at about 12,500-13,000 years B.P. A diagrammatic depiction of two contrasting interpretations of the fluctuations of water level of the post-Cary lakes in the Lake Michigan basin is shown in Fig. 13. The differences of opinion of interpretation of field data between Bretz (1959, 1964, 1966) and of Hough (1958, 1963, 1966) are by no means resolved. I would agree with Bretz (1966) that a lake level at 620 ft (equal to the level along Highway 42 in the recommended area) was post-Valders further south, near Port Washington. It seems likely that the lake sands on the Valders till at Two Creeks are local in occurrence but a Calumet level of Glacial Lake Chicago has not yet been ruled out entirely. Discussion of this problem goes far beyond the scope of this book and includes glacial lakes and drainage in northern United States and in Canada from the Rockies to the Atlantic and from Hudson Bay to the Gulf of Mexico. It even involves indirectly the arguments of Antevs (1962) on transatlantic correlations and dating. Although complex, the history of the Great Lakes is truly a fascinating subject (Hough 1958).

When one considers the magnitude of water fluctuations through

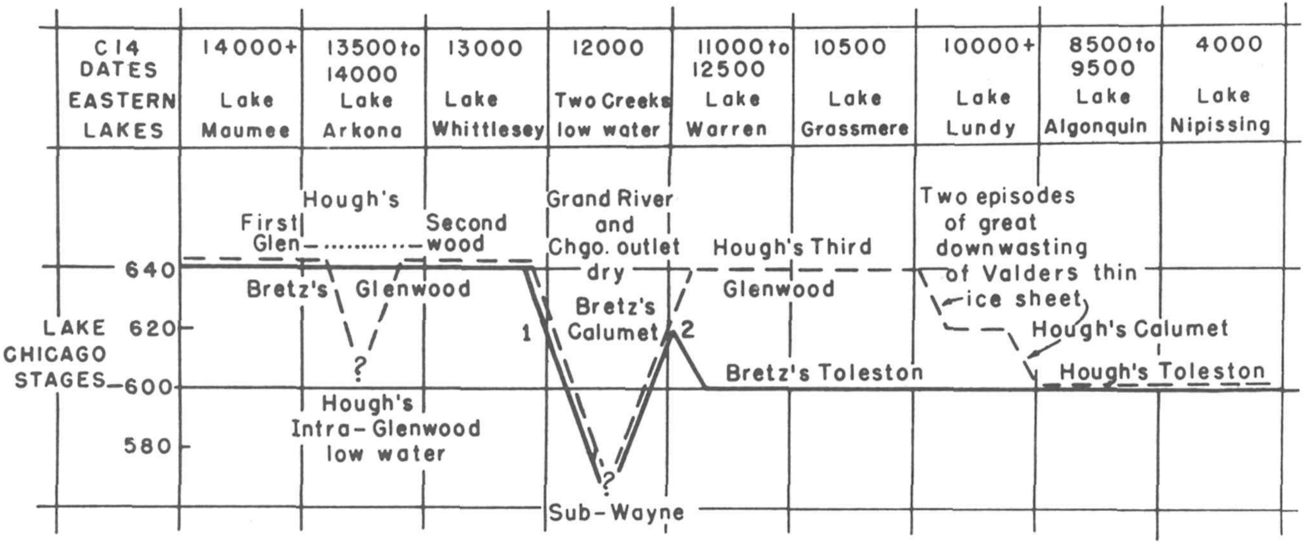

Fig. 13. Diagrammatic depiction of two contrasting interpretations of fluctuations of Glacial Lake Chicago. After Bretz (1959, Fig. 3).

hundreds of feet during only some thousands of years, the present day fluctuations of a few feet are relatively insignificant. Nonetheless, exceedingly rapid shoreline erosion [up to 40 ft per year at Manitowoc in 1905 (Goldthwait 1907)] forces one to appreciate some of the consequences of minor lake level fluctuations. In 1966 water levels in Lake Michigan had been low for several years. As a result, shoreline erosion at the Two Creeks Forest Bed has been minimal. When the site was first found by Goldthwait and subsequently when Wilson had an opportunity to examine the location, water levels were relatively high. This permitted shore erosion to expose the forest bed which in 1966 was covered by slump and vegetation. Should water levels again rise by 2 or 3 ft, additional shore erosion can be expected. It is for this reason that those portions of the Two Creeks Forest Bed now close to water level cannot be recommended for inclusion in the Reserve. Drainage, shore erosion, slumping, and other problems would make it difficult to have displays for public use.[1]

[1]Since 1966 water level has risen and points north and south of the recommended area, that supplied sand for the beach, have been protected by large blocks of dolomite at new nuclear power plants. Hence, rapid shore erosion is occurring at a dug exposure of the Two Creeks soil at the site, which is protected by a temporary shelter built during preparation of a documentary film (Black, Robert F., David L. Clark, and Thomas E. Hendrix. 1968. Two Creeks Buried Forest Project —CIC Instructional Improvement Program. J. Geol. Ed. 16:139-140. See also Black, Robert F. 1970. Glacial geology of Two Creeks Forest Bed, Valderan type locality, and Northern Kettle Moraine State Forest. Univ. Wis. Geol. and Nat. Hist. Surv. Inf. Circ. No. 13).

3
Sheboygan Marsh

Sheboygan Marsh, almost 20 mile^2 in area, lies directly behind the front of the Green Bay Lobe where it abutted ice of the Lake Michigan Lobe to form the Kettle Interlobate Moraine in Sheboygan County (Figs. 1, 2, 4, 14, 15). Sheboygan River runs easterly through the center of the marsh which is partly occupied by Sheboygan Lake. Water level is controlled at 907 ft above sea level by a dam at the northeast corner of the marsh. Much of the marsh is a county park and wildlife refuge. One of the best views of the marsh (Fig. 15) is from the terminus of the town road which serves the house on the bluff on the north side of Sheboygan Lake, in the west center of sec. 14, T. 16 N., R. 20 E. A public overlook at this location is highly desirable.

Details of the history of the marsh, a former glacial lake, and of its deposits are lacking. Bedrock supported hills rise more than 100 ft above the marsh on much of its north, south, and west margins (Fig. 14). The Kettle Interlobate Moraine forms the east margin. Alden (1918:46, PI.II) shows that a deep pre-glacial valley trended southeastward through Sheboygan Marsh and Elkhart Lake. On the west side of that lake a well 240 ft deep penetrated only unconsolidated materials. Seemingly, more than 100 ft of sediment fills Sheboygan Marsh, of which some at least is marl (Alden 1918:47). The Sheboygan River undoubtedly has contributed much material to the marsh, and an unknown but appreciable proportion probably came from the local vegetation added year after year since the late Woodfordian or Cary ice disappeared from the area. But why is the marsh almost filled while Elkhart Lake is still 113 ft deep? Was its buried ice that much thicker or has more detritus been brought to Sheboygan Marsh? Both surmises seem plausible, but the story remains to be worked out.

What role either the Cary or Valders melt waters played in filling the marsh is not known. However, the marsh lay between the Valders ice that filled the area of Lake Winnebago to the northwest and of Lake Michigan to the east. Sheboygan River would have been dammed

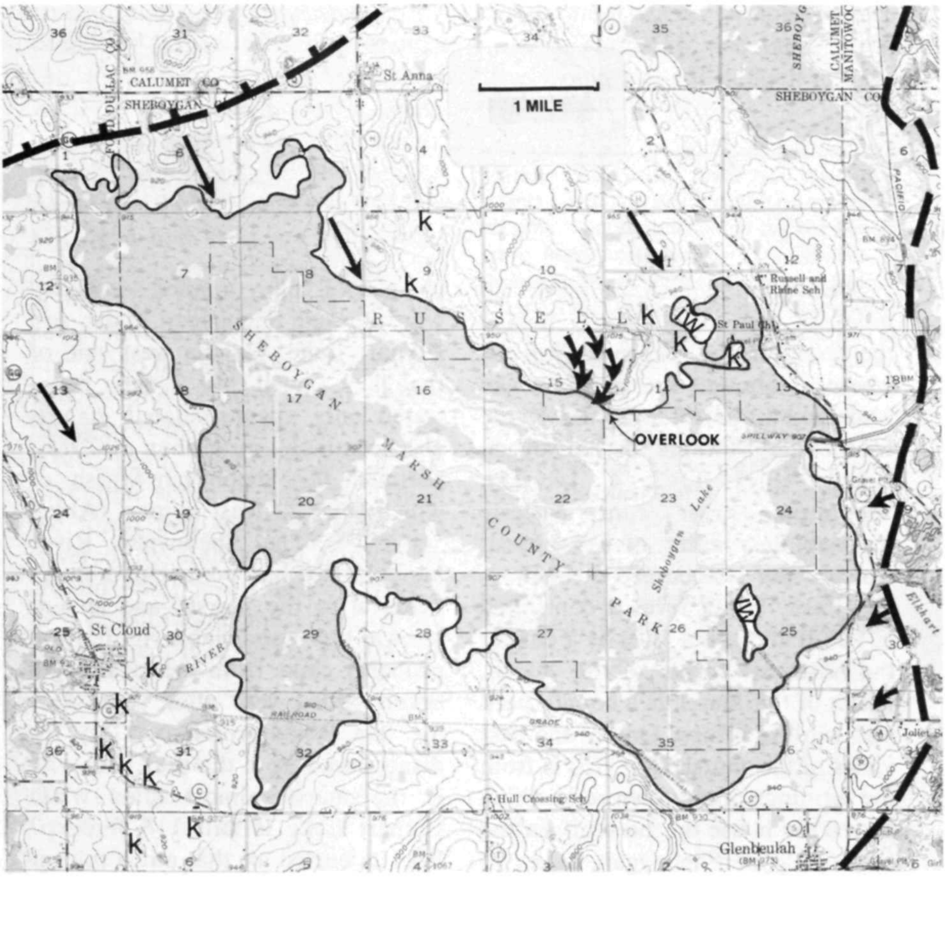

Fig. 14. Topographic map of Sheboygan Marsh, a former glacial lake, and its relationship to the Kettle Interlobate moraine whose west border is shown by a heavy broken line. The position of a retreatal moraine of Woodfordian age is shown northwest of the marsh by ticked line. Directions of ice movement are indicated by long straight arrows and of water flow by short arrows; some local kames are indicated by the letter "k"; and two ice-walled lake deposits by the letters "iw". Part of U.S. Geological Survey topographic map-Sheboygan.

Fig. 15. View southward of Sheboygan Marsh, from recommended overlook on the north side (Fig. 14).

on the east, undoubtedly raising water levels in the area of the Marsh. Furthermore, the South Branch of the Manitowoc River would have had its outlet cut off by the Valders ice on the east, but at the same time it would have been receiving water from the Valders ice in the Lake Winnebago area. The divide between Sheboygan Marsh and the South Fork of the Manitowoc River to the west is only about 950 ft above sea level. It seems likely that both areas were once part of an extended glacial lake, but no research has been done on this problem.

The bedrock-supported hills that rise above the Marsh on all sides but the east show clearly by drumlinoid and fluted forms the southeasterly direction of flow of the ice of the Green Bay Lobe. During stagnation of the Cary ice, water flowed across and around a number of those hills. The most striking channels are north of Sheboygan Lake in secs. 14 and 15. Water from off the ice in what is now Elkhart Lake also flowed southwesterly. The area of Glenbeulah, northward and also southwestward to Greenbush in the Mullet River Valley, has the surficial appearance of having been "washed over." Sand of possible lacustrine origin is common in that area. A small morainal area of knob and swale topography lies west of Glenbeulah.

Around the Marsh in various places are small irregular ice-contact features. Most are kames; Alden (1918) mapped a small esker north of the Marsh but most of it has been removed. A few of these features are shown in Fig. 14. The small area enclosed by black line in Fig. 14 in the southeastern part of the Marsh appears from inspection of aerial photographs to be an ice-walled lake area (a former lake that received sediment from the enclosing glacial ice) produced during stagnation of the Cary ice. Another appears in the northeastern corner of the Marsh. No borings or excavations of these features or of the adjacent marsh sediments have been attempted.

Sheboygan Marsh thus is the final vestige of a large glacial lake into which many tens of feet of fill have been laid down by wash from the surrounding hillsides, from the Sheboygan River, and from the growth of vegetation within the lake. The details of its history have not been reconstructed. It seems typical of many large lakes that originated during the Woodfordian and Valderan glaciations. Others, like Horicon Marsh in Dodge County, are as far advanced or even farther in their maturation. Others, such as Lake Winnebago, are doomed ultimately to the same fate. Man usually increases the rate of maturation; he has difficulty in retarding it.

4
Northern Kettle Interlobate Moraine

Introduction

The Northern Kettle Interlobate Moraine is, as the name implies, a moraine with numerous kettles, formed between two lobes of ice—the Green Bay and Lake Michigan Lobes (Alden 1918:308-309) (Figs. 1, 2, 4, 16, 17). The moraine contains a variety of glacial features some of which were among the first in the country to be well described (Chamberlin 1877:199-246, 1878) and which remain today among the best in the world. Glacial features are well represented in the Northern Kettle Moraine State Forest which extends from the vicinity of Glenbeulah in Sheboygan County southwesterly and then southerly 20 miles to the vicinity of County Highway H about 3 miles south of Kewaskum, in Washington County (Figs. 1, 4, 17). The forest encompasses much of the area along the common boundary of Fond du Lac and Sheboygan counties. The area recommended for inclusion in the Reserve includes the entire Northern Kettle Moraine State Forest plus principally an area within the north-central part that is not now in the forest . The recommended addition contains one of the most striking groups of moulin kames (conical hills of drift deposited under the ice) to be found anywhere in the world (Fig. 17). [Moulin (möulań) Fr., is defined by Webster's dictionary as a nearly vertical shaft in a glacier into which a stream of water pours. The debris carried in by the water is piled up at the base of the moulin, building the moulin kame.]

The Interlobate Moraine was mapped by Alden (1918) as part of a reconnaissance in southeastern Wisconsin and has hardly been touched since. Much study is needed to understand fully the history of individual forms or even of many large units. Different interpretations are possible within the framework of existing data. However, it seems clear that several at least local fluctuations of the two lobes were involved during Woodfordian time. The junction thus is a zone of partial mixing or interstratifying of material from each lobe. Outwash gravel and other glacial deposits were reworked and redeposited, commonly on preexisting ice, as the junction shifted back and forth.

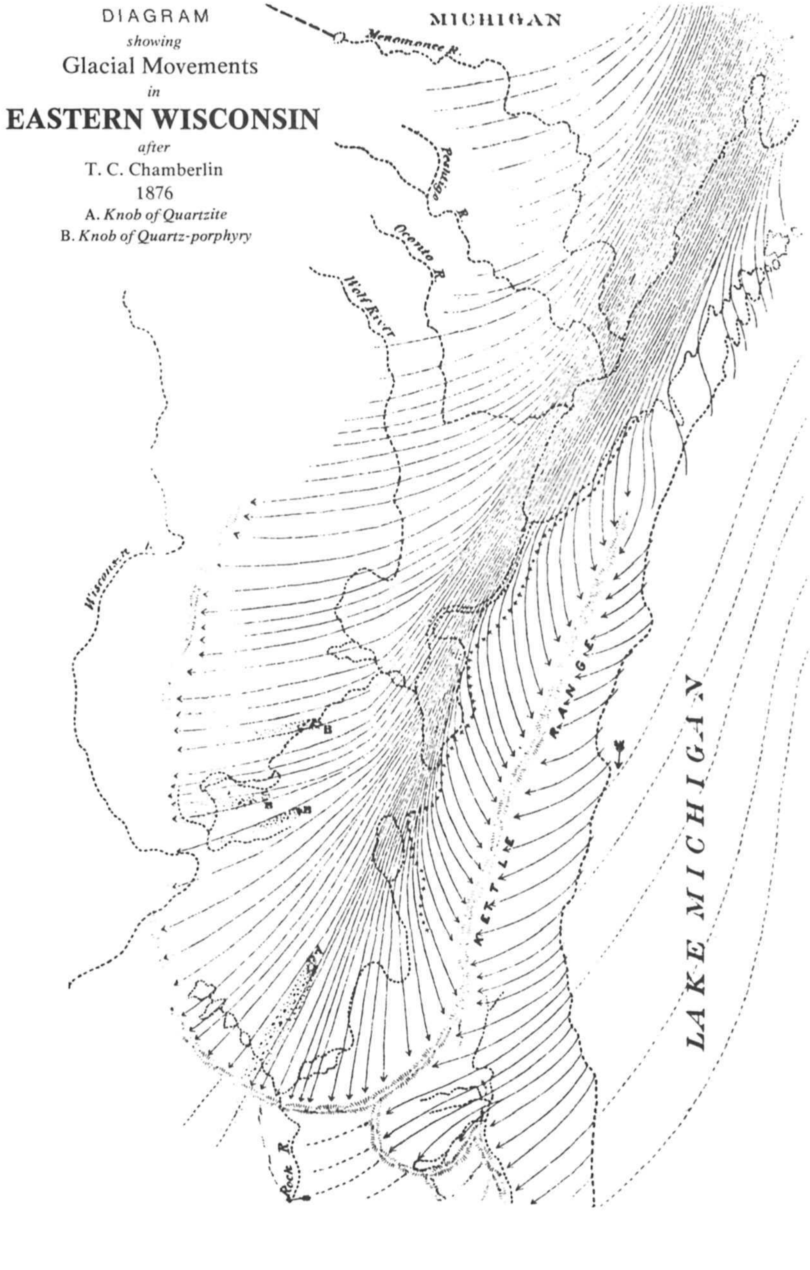

Fig. 16. Diagram showing glacial movements in eastern Wisconsin. After Chamberlin (1878). The Kettle Range is now called the Kettle Interlobate Moraine.

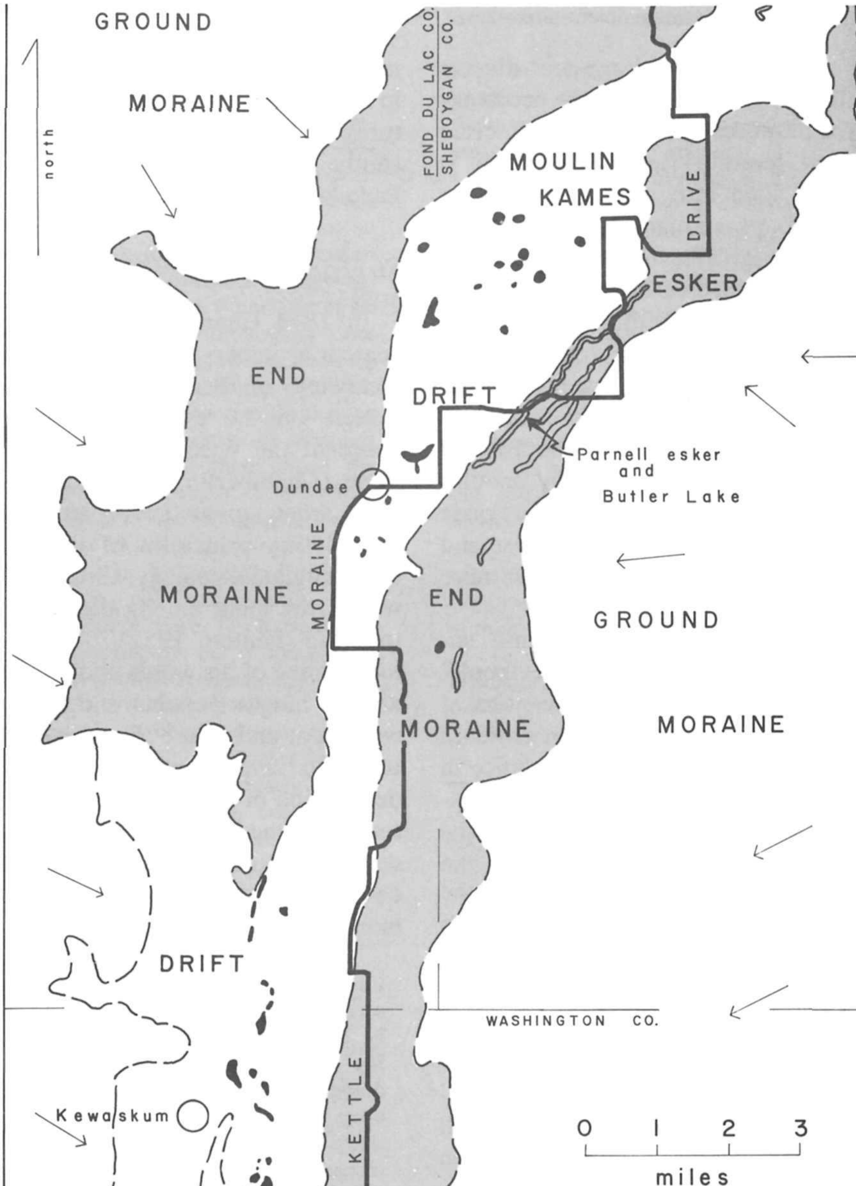

Fig. 17. Generalized glacial map of part of the Northern Kettle Interlobate Moraine, showing ground moraine with drumlins, end moraine or stagnate-ice and dead-ice moraine, stratified drift, some moulin kames (closed circles), some eskers and crevasse fills (sinuous forms), and ice movement (arrows).

The area is so large and diverse that it is not feasible nor necessary for purposes of this book to describe each feature. Rather, the area is subdivided into mappable units or groups of similar geomorphic features (Fig. 17). These are not pure units because of the almost infinite detail available within any relatively small segment of the forest. Nonetheless they serve to emphasize such features as end moraines and stagnate-ice or "dead-ice" moraine of knob and swale topography, moulin kames, outwash, eskers, crevasse fills, kettles, and the like. These and other features are described more fully later.

Because of its variety and superb development of "textbook" features, its proximity to centers of population and heavy recreational use, and its historical importance in the development of concepts in glacial geology, this area is one of the most important in the state that can be recommended for inclusion in the Reserve. It deserves every attention it can be given; the recommended expansion of the state-owned forest is absolutely minimal. Further expansion in spite of high land values is exceedingly desirable.

Numerous glacial features on all sides of the recommended area should be included in the Reserve even though they are similar to features within the recommended area. Because of their value for construction aggregates, many depositional glacial features have been destroyed, and others will be lost or modified unless they can be included in the Reserve. Other glacial features not likely to be altered significantly or lost to mankind if not included are also mentioned.

Surface Features

In 1876 Chamberlin orally presented a paper to the Wisconsin Academy of Sciences, Arts, and Letters on the extent and significance of the Wisconsin Kettle Moraine (Chamberlin 1878). In those days when great geologists were formulating principles of the concepts of glacial geology, Chamberlin was a true giant among them (Fenton and Fenton 1952). Although today some of his words and phrases are no longer popular and editors would cut and prune his remarks in order to save space, Chamberlin's description of the moraine bears the test of time so well that I feel compelled to quote him directly. In describing the surface form of the moraine he wrote:

> The superficial aspect of the formation is that of an irregular, intricate series of drift ridges and hills of rapidly, but often very gracefully, undulating contour, consisting of rounded domes, conical peaks, winding and, occasionally, geniculated ridges, short, sharp spurs, mounds, knolls and hummocks, promiscuously arranged, accompanied by corresponding depressions, that are even more striking in character. These depressions, which, to casual observation, constitute the most peculiar and obtrusive feature of the range, and give rise to its descriptive name in Wisconsin, are variously known as 'Potash kettles', 'Pot holes',

'Pots and kettles', 'Sinks', etc. Those that have most arrested popular attention are circular in outline and symmetrical in form, not unlike the homely utensils that have given them names. But it is important to observe that the most of these depressions are not so symmetrical as to merit the application of these terms. Occasionally, they approach the form of a funnel, or of an inverted bell, while the shallow ones are mere saucer-like hollows, and others are rudely oval, oblong, elliptical, or are extended into trough-like, or even winding hollows, while irregular departures from all these forms are most common. In depth, these cavities vary from the merest indentation of the surface to bowls sixty feet or more deep, while in the irregular forms the descent is not unfrequently one hundred feet or more. The slope of the sides varies greatly, but in the deeper ones it very often reaches an angle of 30° or 35° with the horizon, or, in other words, is about as steep as the material will lie. In horizontal dimensions, those that are popularly recognized as 'kettles' seldom exceed 500 feet in diameter, but, structurally considered, they cannot be limited to this dimension, and it may be difficult to assign definite limits to them. One of the peculiarities of the range is the large number of small lakes, without inlet or outlet, that dot its course. Some of these are mere ponds of water at the bottom of typical kettles, and from this, they graduate by imperceptible degrees into lakes of two or three miles in diameter. These are simply kettles on a large scale.

Next to the depressions themselves, the most striking feature of this singular formation is their counterpart in the form of rounded hills and hillocks, that may, not inaptly, be styled inverted kettles. These give to the surface an irregularity sometimes fittingly designated 'knobby drift'. The trough-like, winding hollows have their correlatives in sharp serpentine ridges. The combined effect of these elevations and depressions is to give to the surface an entirely distinctive character.

These features may be regarded, however, as subordinate elements of the main range, since these hillocks and hollows are variously distributed over its surface. They are usually most abundant upon the more abrupt face of the range, but occur, in greater or less degree, on all sides of it, and in various situations. Not unfrequently, they occur distributed over comparatively level areas, adjacent to the range. Sometimes the kettles prevail in the valleys, the adjacent ridges being free from them; and, again, the reverse is the case, or they are promiscuously distributed over both. These facts are important in considering the question of their origin.

The range itself is of composite character, being made up of a series of rudely parallel ridges, that unite, interlock, separate, appear and disappear in an eccentric and intricate manner. Several of these subordinate ridges are often clearly discernible. It is usually between the component ridges, and occupying depressions, evidently caused by their divergence, that most of the larger lakes associated with the range are found. Ridges, running across the trend of the range, as well as traverse spurs extending out from it, are not uncommon features. The component ridges are themselves exceedingly irregular in height and breadth, being often much broken and interrupted. The united effect of all the foregoing features is to give to the formation a strikingly irregular and complicated aspect. (Chamberlin 1878:202-204).

Chamberlin was referring to the surficial features of the end moraine of what is now called the late Woodfordian or Cary ice as it was deployed through the entire state of Wisconsin and not just the interlobate moraine in what is now the Northern Kettle Moraine State Forest. Nonetheless, his description can scarcely be improved upon for the recommended area. In speaking of the nature of the material, Chamberlin (1878:205) emphasized that "... *all* the four forms of material common to drift, vis: clay, sand, gravel, and boulders, enter largely into the constitution of the Kettle range, in its typical development. Of these, gravel is the most conspicuous element, *exposed to observation.*" Chamberlin (1878:210) further recognized that most bedrock units in Wisconsin and Upper Michigan were represented in any one section of the drift, including native copper from Keweenaw Peninsula, but that the bulk of the drift was derived locally. Thus, most gravel is composed of the local white to very light gray Silurian dolomite (Fig. 3), well rounded by water work (Fig. 18).

However, we now know that

Fig. 18. Typical road cut, showing well-rounded gravel dominantly of white to very light gray Niagaran dolomite, 1.5 miles south of Glenbeulah. (NW¼ NE¼ sec. 13, T. 15 N., R. 20 E.). This is reworked outwash.

more than one local advance of ice was involved, spanning probably several thousands of years, and that reworking of outwash gravel by later advances was commonplace. Hence, some constructional forms contain nonstratified gravel instead of till. Deposition of the gravel directly from ice without water working took place.

Other details of the moraine in Wisconsin were presented, and it was compared with its counterpart in other states (Chamberlin 1877, 1883a). In the latter paper, the term "interlobate moraine" was first introduced (Chamberlin 1883a:276) and properly diagnosed as to origin in contrast to normal medial moraines. A reconstruction of the ice flow directions (Fig. 16) demonstrates conclusively the lobate character of the ice and the opposing movements at the junction of the two lobes. This gross story has changed little in the intervening century.

Chamberlin's important role in the development of the concepts of glacial geology would not have been

Fig. 19. Small symmetrical kettle with shallow pond in typical knob and swale topography. West side of County Highway A, NW¼SW¼ sec. 14, T. 15 N., R. 20 E.

possible were it not for the clear observations and lucid writings of his predecessors of whom, in connection with the Kettle Interlobate Moraine, only Charles Whittlesey will be singled out. It was Whittlesey who, in the mid-1800s, first recognized the "kettle moraine" and correctly interpreted the origin of the kettle holes (Fig. 19) to buried glacial ice rather than drifting icebergs as was in vogue at the time (White 1964). This was truly astonishing insight, and is but one of the major accomplishments of that amazing man.

The Greenbush Kettle, 2 miles south of Greenbush on the Kettle Moraine drive, has been favored with a geological marker sign for years. It is one of the most symmetrical deep circular depressions visible from the road. Many others are more irregular (Fig. 20) but just as typical, with or without water.

In brief the Northern Kettle Interlobate Moraine is conspicuous because of its more abrupt irregularity and sharpness of feature (Fig. 21) compared to the undulating ground moraine with smoothly contoured drumlins and till-covered bedrock rises on both sides (Fig. 22). The light-grey gravel of the Interlobate Moraine also contrasts markedly with the reddish brown and light yellowish brown sandy till of the ground moraine. Neither its maximum elevation (1311 ft at Parnell tower, sec. 10, T. 14 N., R. 20 E.) (Fig. 23) nor its general relief of 100-200 ft are significantly different from the till plains and drumlins adjoining. However, it is characterized by major lowlands at 950-1000 ft, such as that occupied by Long Lake (Fig. 24) and the East Branch of the Milwaukee River. The flatness of such lowlands and the abrupt rise of drift deposits flanking them also emphasize the glacial features (Fig. 25). Farming of the lowlands contrasts with the wooded drift hills to spice the view.

Drainage

The Kettle Interlobate Moraine lacks an integrated drainage network. Many closed depressions drain through the coarse gravel below and do not need surface streams. Others intersect the ground water table and have perennial ponds or lakes. Elkhart Lake (Fig. 26), a large kettle north of the forest , with high land around it, drains westward to Sheboygan Marsh and the Sheboygan River. Crystal Lake, next south of Elkhart Lake, has no outlet. However, Mullet River flows by only 0.25 mile to the southwest in its arc around the north end of the Kettle Moraine Forest, and then southeasterly and eastward in a tortuous route to join the Sheboygan River at Sheboygan Falls. Interestingly those two rivers have adjoining headwaters, and their uppermost courses are parallel yet flowing in opposite directions about 1 mile apart northwest of Long Lake. Both rivers have very intricate courses to Lake Michigan,

Fig. 20. Part of the large irregular kettle without water in stagnate-ice moraine, 2 miles south of Greenbush. View northeastward (SW¼ sec. 23, T. 15, N., R. 20 E.).

Fig. 21. Typical topography in stagnate-ice moraine, showing stony knobs, 1.5 miles south of Glenbeulah. View northeastward (NW¼ NE¼ sec. 13, T. 15 N., R. 20 E.).

Fig. 22. Ground moraine of the Green Bay Lobe 1.5 miles west of Glenbeulah. View westward (NW¼ NW¼ sec. 2, T. 15 N., R. 20 E.).

Fig. 23. Main portion of the Kettle Interlobate Moraine culminating at the Parnell observation tower in center. View northeastward across swampy bottom land with numerous glacial erratics.

Fig. 24. Long Lake, looking northward, from the Wisconsin State camp ground.

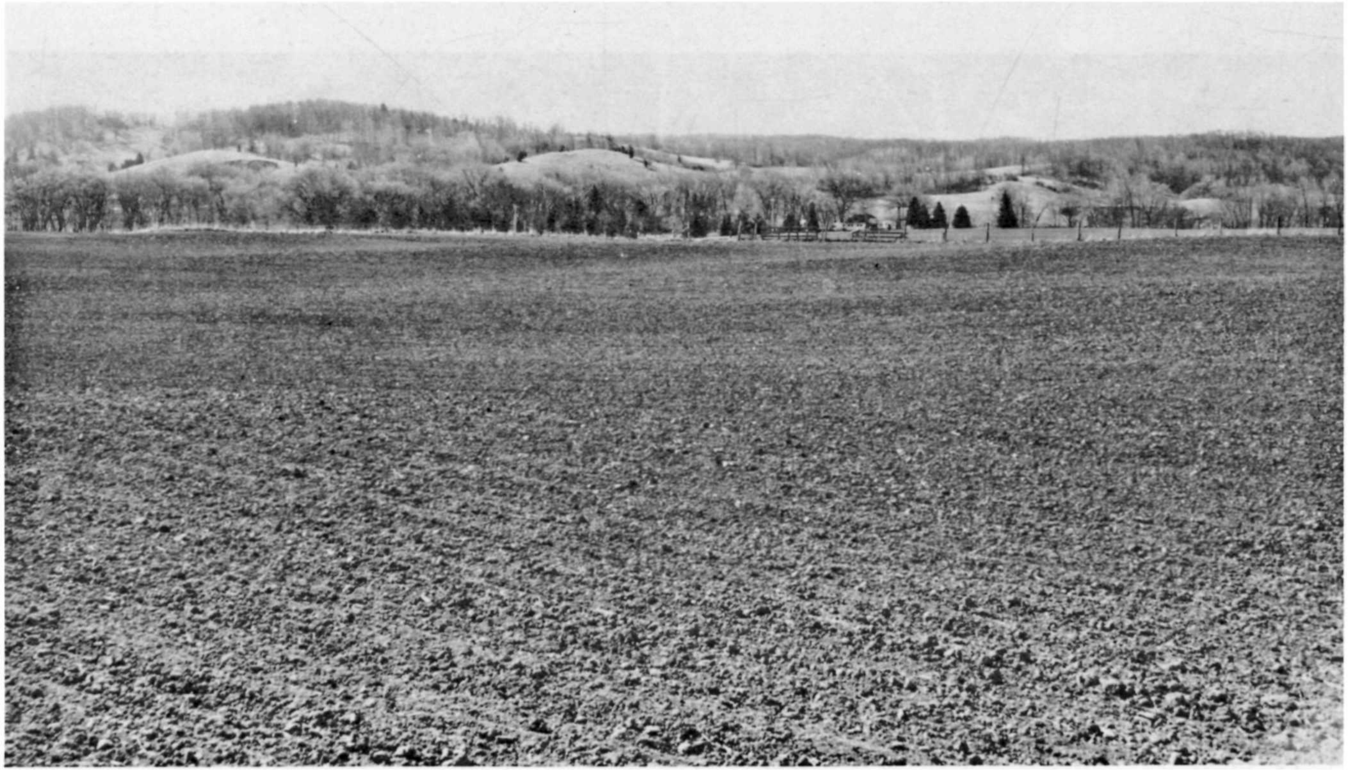

Fig. 25. Main Kettle Interlobate Moraine immediately south of Glenbeulah, looking southeastward.

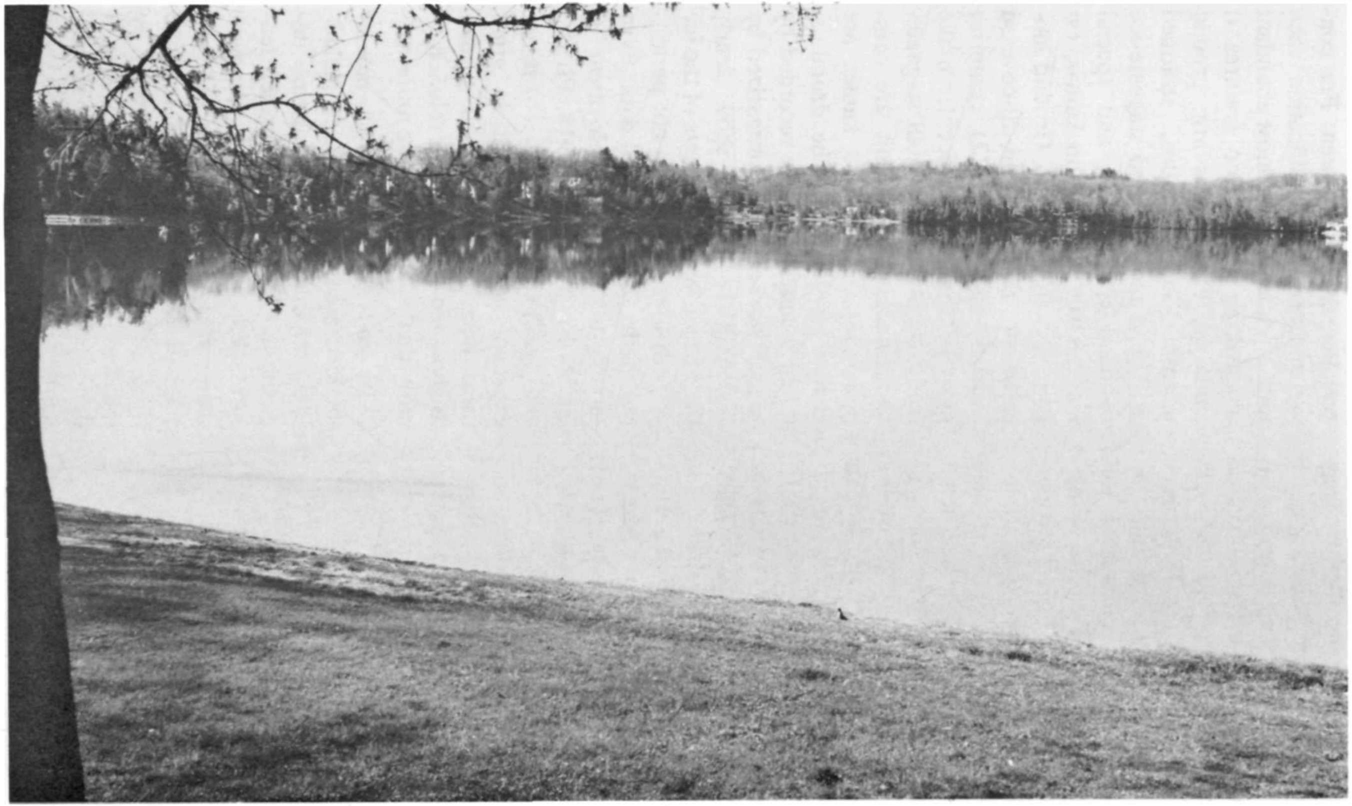

Fig. 26. Elkhart Lake, a large kettle, looking southwestward.

probably in part controlled by fracture patterns in the stagnating ice which permitted the supraglacial streams to superpose themselves on the underlying drift and bedrock. The East Branch of the Milwaukee River, flowing southward into the Milwaukee River southeast of Kewaskum, drains most of the Northern Kettle Moraine Forest proper. Its course follows the trend of the moraine and generally lies almost precisely on the reconstructed boundary between the two lobes of ice. (This is somewhat west of the boundary indicated by Alden 1918, Pl. III). Probably its origin dates back to the initial abutment of the ice of the two lobes where it developed in the axial depression along that junction. Apparently it has remained in that position since.

In the wastage of the Lake Michigan Lobe, however, additional channels were formed on the stagnating ice. Mink Creek lies in a channel that starts about 2 miles northeast of Parnell and flows generally southerly past Beechwood in a course with abrupt right-angle bends. These seem also to reflect the fracture pattern of the ice as the initial stream was let down on the surface below. Many other examples exist in the area, but no field study of any of them has been attempted. They need to be integrated into the history of the moraine.

Glacial Units

Figure 17 shows the distribution of drift features that characterize certain parts of the area. For convenience in the classification each unit is named for the most abundant or striking feature or features it contains. These units are: ground moraine and drumlins, stratified drift, end moraine and stagnate-ice or dead-ice moraine, and special features such as moulin kames, eskers, and crevasse fills. Ground moraine with drumlins and till-covered bedrock rises (Fig. 22) comprise most of the area up-ice from the front of both lobes. Small stagnate-ice features in that unit are common, but only a few kames are shown on the map. The drumlins, fluted forms, and striae recorded by earlier workers and summarized by Alden (1918, Pl. IV) show clearly the former last movements of the ice of both lobes. Time has not permitted me to add more field data. Even though the general deployment of ice shown by Alden (1918, Pl. IV) and by Chamberlin (Fig. 16) is not expected to be changed in gross form, detailed field work is needed to show ice movement in relation to individual segments of the moraine.

In and adjacent to the recommended area stratified drift, including outwash, glacial-lacustrine deposits, and other water-formed features (Fig. 27) are more prevalent than end moraine (Fig. 25) and stagnate-ice or dead-ice moraine (Fig. 28) formed more directly by the ice. The washed surfaces and deposits reflect in part the cleaner ice of the two lobes juxtaposed and in part the concentration of runoff

Fig. 27. Washed drift plain in foreground with crevasse fill on left of higher moulin kame or ice-walled lake area. View eastward from County Highway A, 2 miles south of Greenbush.

Fig. 28. Crevasse fill or elongate ridge of drift in stagnate-ice moraine, immediately southeast of Glenbeulah. View southwestward.

along the junction of the two lobes. The normal surface gradient up-ice in each lobe would have led water to the junction of the lobes, from which its escape could only have been to the south along that junction. Such water-worked stratified drift varies in size from the coarse bouldery material of glacial streams to the sand, silt, and clay in ponded water. Drift obviously has formed in places on buried ice blocks to leave pitted outwash; elsewhere it seems that entire portions of stream beds or lake sediments have been dropped down as continuous ice below melted out. Most parts of the well-washed drift, however, were formed adjacent to ice, but not on it. Original stratification is preserved.

Even during deglaciation the widening and northward migrating gap between the two lobes effectively concentrated glacial-fluvial activity between the lobes. Thus it was the locus for many striking forms. Eskers (Figs. 29-32) and moulin kames (Figs. 33-37) formed under the stagnate ice by subglacial streams fed through moulins or openings through the ice sheet. Their subglacial waters also flowed toward that same gap. Crevasse fills (Fig. 28), topographically commonly like eskers, were formed in crevasses open to the sky in part by supraglacial streams and in part by mass movement of surface debris into the crevasses.

Moulin kames are scattered throughout and adjacent to the recommended area, but none is better developed or displayed than those in the group northeast of Long Lake. There, in the widest part of the washed drift area, are some of the best moulin kames to be found anywhere in the world. Beautifully conical hills, such as McMullen (Fig. 33), Garriety (Figs. 34 and 35), Connor and Johnson (Fig. 36), rise at the angle of repose of the material more than 100 ft above the flat, washed, drift plain surrounding them. Numerous smaller kames, only a few tens of feet high, are commonly less conspicuous among the drift ridges. Many are just as symmetrical as the larger ones in the lowlands (Fig. 37). Other more irregular moulin kames, such as Dundee Mountain, are also present and grade into crevasse fills or into ice-walled lake areas (openings so enlarged that lakes formed within the glacial ice walls). Such forms originated where melt waters on the ice dropped through moulins or crevasses, dumping their detritus at the base. Openings ranged from nearly vertical circular pipes (moulins) to very elongate fractures and rounded to irregular openings; commonly water and debris was fed into the fractures at more than one place along the sides and ends of crevasses, building irregular forms below the ice. 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 delta and lacustrine sediments. Thus, the material in such features as moulin kames and

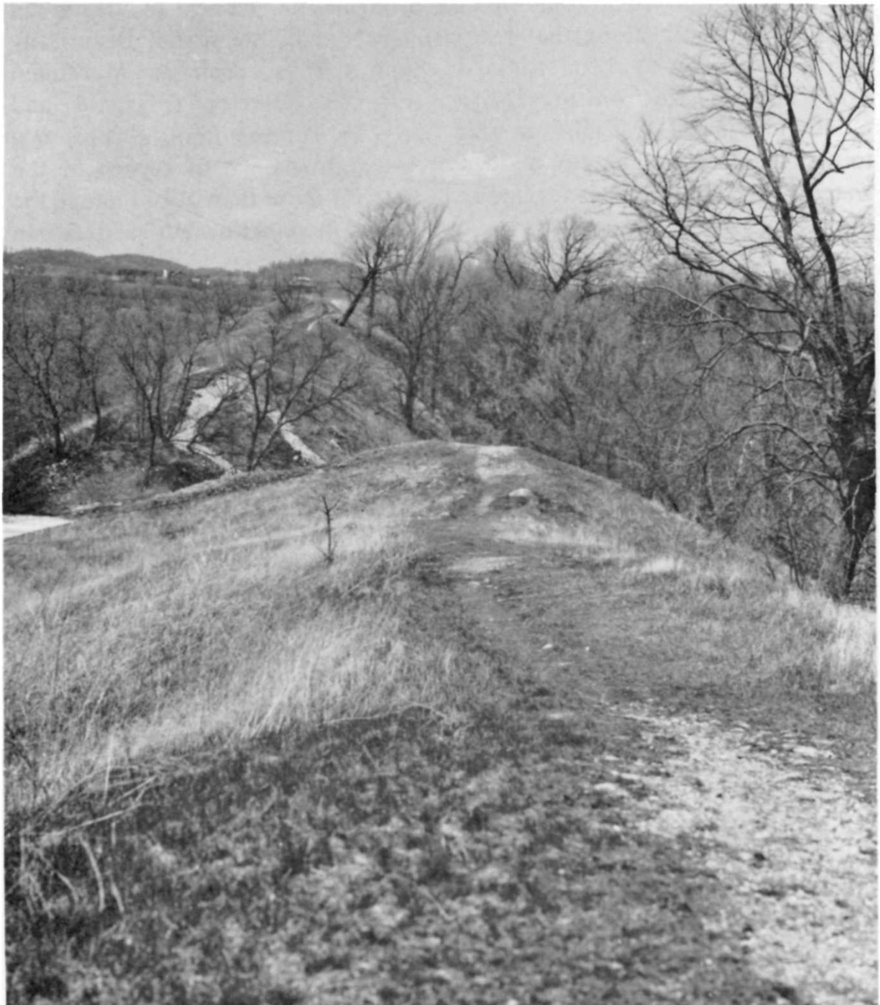

Fig. 29. Parnell esker at Butler Lake. The breach in the esker is manmade. View northeastward.

Fig. 30. Parnell esker, on skyline, in the NE¼ sec. 20, T. 14 N., R. 20 E. View eastward.

Fig. 31. Cross section of Parnell esker by County Highway V, north center of sec. 20, T. 14 N., R. 20 E.

Fig. 32. Butler Lake, looking northeastward, from top of Parnell esker. Cross section of esker in Fig. 31 is visible in distance, left of center.

Fig. 33. McMullen Hill, a moulin kame, looking northward, with Connor Hill, another moulin kame, on the right.

Fig. 34. Garriety Hill, a moulin kame opened by road work. View eastward.

Fig. 35. Close-up of material comprising Garriety Hill.

Fig. 36. Johnson Hill, a moulin kame, looking westward.

Fig. 37. Unnamed small moulin kame, 0.3 mile southeast of Dundee. View southward.

crevasse fills ranges from normal till, through the available sizes of water-transported material, to ponded sediments. The cross section of Garriety Hill is typical (Figs. 34, 35). It shows rounded to angular gravel, sand, silt, and clay deposited as unsorted till in irregular masses, and as sorted sediment in alluvial fans, pond sediments, and the like. Water that formed the northern group of moulin kames drained westward under the ice to join the drainageway through Long Lake Valley. Their channels are readily discernible on aerial photographs.

The end moraine and stagnate-ice or dead-ice moraine are not differentiated in Fig. 17 because of their general similarity of origin. The terms are used loosely here for lack of detailed understanding of their genesis. They might have been subdivided for descriptive purposes into those areas characterized by elongate ridges and valleys and those with circular knobs and swales. In the interlobate area all are believed to result from ice stagnation and the melting out of blocks of ice of the appropriate geometry to fit the surface depressions. Such geometry is predicated on the movement of the ice at the time the ice and debris were mixed, on its fractures, or on the manner of burial by overriding ice, outwash, debris slides, etc.

The detailed deployment of the moraines in the Northern Kettle Moraine Forest is of considerable interest in the reconstruction of events as related to the flow of ice. From the vicinity of Kewaskum north to Long Lake, the trend of the Interlobate Moraine is almost north.

From Long Lake the Interlobate Moraine turns fairly abruptly to the northeast to Elkhart Lake where it again swings to the north. At least part of the explanation of the bend may lie in the topography of the bedrock which unquestionably has exercised some control on the deployment of the ice. The deep preglacial valley at Sheboygan Marsh must have provided relatively easy access for the ice of the Green Bay Lobe, leading it more rapidly and further to the southeast than was possible over the bedrock hills south of that marsh. They restrained the ice of the Green Bay Lobe, allowing the ice of the Lake Michigan Lobe to push farther westward. Such kinks and bends in the terminal area are commonplace along the entire late Woodfordian front. They are of considerable importance in understanding the development of the features found in the Northern Kettle Interlobate Moraine, but time does not permit their reconstruction here. Much field work is needed to unravel their history.

Small moulin kames in the stagnate-ice moraines are probably contemporaneous with the related features. However, the precise timing of the formation of the main group of moulin kames versus the main moraines to west and east is conjectural.

I hypothesize (Black 1970) that shortly after the two lobes butted together, the thickness of ice gradually increased from 100 to 300 ft at the start to a thickness perhaps of several thousand feet when the ice extended southward into the center of Illinois. Ablation (loss of ice) particularly by melting aided by a surface stream at the junction of the two lobes would be countered by ice movement from the base of the ice sheet diagonally upward to that junction at the surface. Upward flow at the terminal zones of glaciers is commonly at angles of 10-45°, bringing debris from the base toward the surface to replace ice lost in the ablation zone and to maintain the surface profile of the glacier. When 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 Cary glaciation (late Woodfordian), it would intersect the surface. As thinning continued to perhaps 200 or 300 ft of ice, fractures aided by melt waters penetrated in favorable places to the bottom of the glacier. In them the moulin kames, eskers, and crevasse fills began to grow. However, at that time the thicker ice back from the junction was continuing to move forward even though the terminal zone was stagnate. The shear planes and flow layers that brought debris up from the base presumably angled obliquely downward and away from the actual surface junction of the two lobes to the general location of the main moraines on both sides of the inner washed area. At the locus of the moraines, basal ice and debris were interstratified by flow of ice while the basal ice closer to the junction was stagnated and remained relatively free of debris. Thus the two main moraines, one for each lobe, are in a sense end moraines even though they do not mark the terminal position of the ice nor were they deposited at the outer edge of the ice. They represent the outer edge of the active ice for each lobe and were separated by a zone of stagnate ice shaped like a very broad, low wedge with its apex upward, at least during the waning of the glaciation. It seems relatively clear that stagnation took place over much of the area because so many small ice-contact, washed-drift features are superposed on all other forms.

Conclusion

Many details of the reconstruction of the events that led to the surface features in the Northern Kettle Interlobate Moraine are imperfectly known. New topographic maps and aerial photographs unavailable to Alden (1918) now permit an analysis of surface forms to be made in far more detail than was possible for him in his reconnaissance study. Surface analysis, however, is only part of the story. Serious mistakes have been made in the past in the interpretation of glacial forms by morphology alone. Subsurface exploration must be carried on concurrently before a firm foundation can be laid that would permit us to change significantly the gross picture of the Kettle Interlobate Moraine as commonly accepted. Such detailed study has had little economic incentive, but should be undertaken before gravel pit operations remove or modify evidence that might be the key to part of the story. A beautiful story can be constructed on evidence available, but an even larger part of the story is still unsupported in fact. The prospects in future study are especially intriguing.

Thus, in brief, the heavy use of the area for recreation and consequent loss of land for cottages and commercial development require our immediate action to preserve many glacial forms, like kames, eskers, and stagnate-ice features. Demands for gravel are increasing and many glacial forms are being removed in toto. We must protect not only the many striking forms but also the "normal" forms now before they are exploited. Many shown in the mapped area (Fig. 17) are outside of the recommended area. It is hoped that some of the better ones ultimately will find their place in the Reserve. If not, the gravelly deposits will disappear as have the moulin kames and crevasse fills immediately east of Kewaskum, on the north side of Highway 28.

The Northern Kettle Interlobate Moraine differs considerably in detail from the Southern Kettle Interlobate Moraine, but their gross chronology and origin have been similar (Alden 1918). The latter would also make a very desirable addition to the Reserve.

5 Campbellsport Drumlins

Drumlins have been by far the longest known and best known streamline molded forms (Flint 1957:66). Charlesworth (1957: 389-403) lists 302 citations to the literature extant at the time, and many other papers on drumlins have appeared since. To this day the origin of drumlins remains conjectural, only in part because different streamline molded forms are called drumlins. Alden (1918:253-256, Pl. III) last mapped and described the drumlins of southeastern Wisconsin where about 5000 drumlins are recognized (Fig. 38). Drumlins and drumlinoidal ridges and flutings in the Campbellsport area (Figs. 1, 2, 4, 39) were included on Alden's geologic map, but not specifically mentioned in the text. Neither were they among those mapped and described in an earlier report (Alden 1905) which still provides us with the bulk of our information on the drumlins of Wisconsin.

Alden (1918:253-255) wrote:

> The writer prefers to restrict the name drumlin to those drift hills which show clearly the molding effect of the advancing ice. The typical drumlin of southeastern Wisconsin may be said to be a hill of glacial drift which approximates the form of a segment of an elongated ovid, of which the widest part of the basal outline and the highest point of the crest are generally not more distant from the stross end than one-third the length of the major axis, and whose major axis is oriented parallel to the direction of the movement of the glacier which formed it. From this type the forms vary on the one hand to elongated narrow ridges, some of which attain lengths of 3 to 4 miles, and on the other hand to nearly equiaxial dome-shaped or mammillary hills. Exceptional variations are double-tailed and double and triple crested forms, and ridges with subordinate crests overlapping in echelon. The longer forms were developed principally in the region of axial movement of the glacier and in the tract west of the Niagara escarpment, where there was good opportunity for the incorporation of sticky aluminous clay derived from the Cincinnati shale [part of Maquoketa, Fig. 3]. The presence of the adhesive clay doubtless facilitated the building up of drumlins by the plastering-on process, and the elongation of the forms may be some function of the more rapid movement along the Winnebago-Rock River trough. There is a general shortening of the forms progressively toward the limit of the advance, where the rate of

flow was retarded owing to the thinning and wide radial spreading of the moving ice. The shortening of the forms on the upland east of the Niagara escarpment may be due to the retarding effect of this escarpment on the ice overriding it in addition to that resulting from thinning and radial spreading.

. . .

With but few exceptions the partial sections of drumlins seen by the writer exposed compact structureless clayey till like that composing most of the rest of the ground-moraine deposit over the limestone areas. In some places the till is semistratified, with somewhat indefinite bands that curve conformably with the surface contours and suggest that the hill has been built up by the addition of more or less definite layers.

Layers of stratified sand and gravel or stoneless silt are exposed in sections of few drumlins. These are in some cases folded, and it is not clear that they have any definite relation to the drumlin structure as such.

A fairly definite cleavage in the clayey matrix of the till developed parallel to the curved surface suggests the effects of pressure of the overriding ice but may in reality be the result of successive additions of thin layers of adhesive clayey material. . . . Evidence indicating the absence of rock cores from the drumlins has been collected throughout the whole drumlin area.

. . .

Estimates based on pebbles collected from drift composing drumlins show . . . that about 91 per cent of the coarser material is of local derivation. If the analyses included also the finer material comprising the matrix of the till the percentage of local material present might be found to be even larger. This high percentage of local material indicates that the drumlins are composed of drift accumulated at or near the base of the ice and transported for comparatively short distances.

. . .

The drumlins were formed . . . in those parts of the area of the Green Bay Glacier where the lines of movement were radiating very notably as the ice spread to the curved margin of the lobe. The lines of movement bounding any drumlin-forming segment of this glacier from the north limit of the drumlins in the area of that segment to the peripheral margin of the lobe show that the amount of this spreading is very considerable—much greater than that which took place in similar segments of the glacier of equal initial width but within whose area no, or few, drumlins were formed. This relation gives rise to the suggestion that radiation was an important factor in drumlin formation. . . .

On Alden's map (1918, Pl. III) distinction was made between drumlins according to the description above and drumlinoidal ridges and fluted forms of similar shape and dimensions. No list of criteria distinguishing between the two were set up, although such forms with bedrock cores seem automatically to be called drumlinoidal. Such practice is generally, but not universally, followed today. Obviously by surface inspection alone, it is not always possible to distinguish true drumlins in Alden's sense from drumlinoidal ridges and fluted forms, and I make no pretext of doing so. Stratification of till (non water-worked) and drift (waterworked) in drumlins in Wisconsin seen by me is much more complicated than Alden described, but details cannot be entered here.

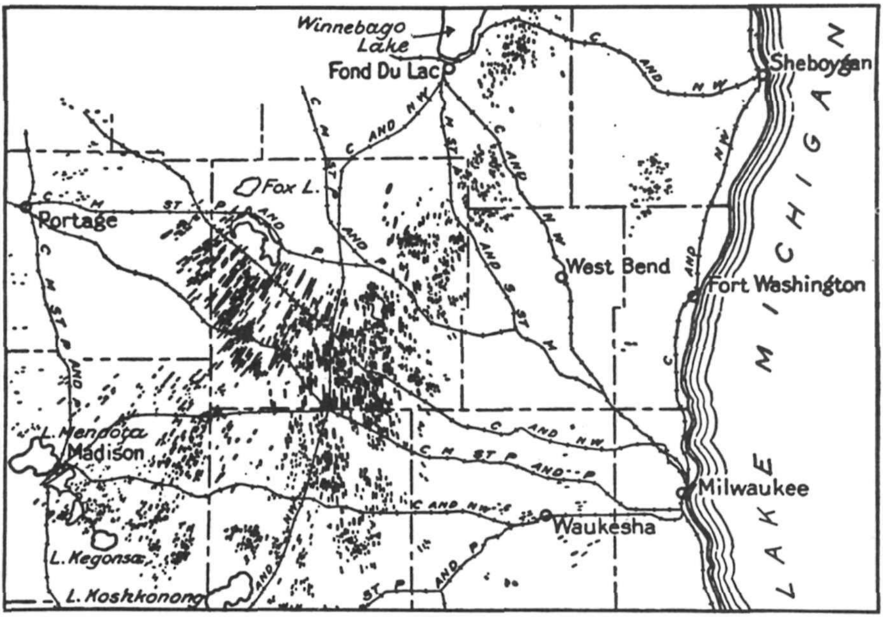

Fig. 38. Drumlins of southeastern Wisconsin. After Alden (1918) via Prouty (1960, Fig. 7).

The specific drumlins and drumlinoidal forms of the Campbellsport area doubtless were specified in the Act because of their proximity to the Northern Kettle Interlobate Moraine rather than because of their uniqueness or unusual degree of development. The best drumlins (including drumlinoidal forms) in the vicinity of Campbellsport are centered 4 miles northwest of that town (Fig. 39). There the drumlins are generally 60-120 ft high, rounded, irregular, to elliptical. Elongation ratios commonly are less than 2:1. Some of the drumlinoidal forms are believed to have bedrock cores. The more symmetrical and elliptical drumlins clearly show deployment of the ice, but only minor flutes on the irregular forms show it.

The Campbellsport drumlins, although typical of many in Wisconsin, are neither the highest, longest, largest, nor most symmetrical. They are closest to the Northern Kettle Interlobate Moraine and provide good views of such features from county roads not heavily traveled. In the area of better drumlins and drumlinoidal forms shown in Fig. 39 are some locations of possible roadside overlooks, not all of which are needed to give the visitor a good

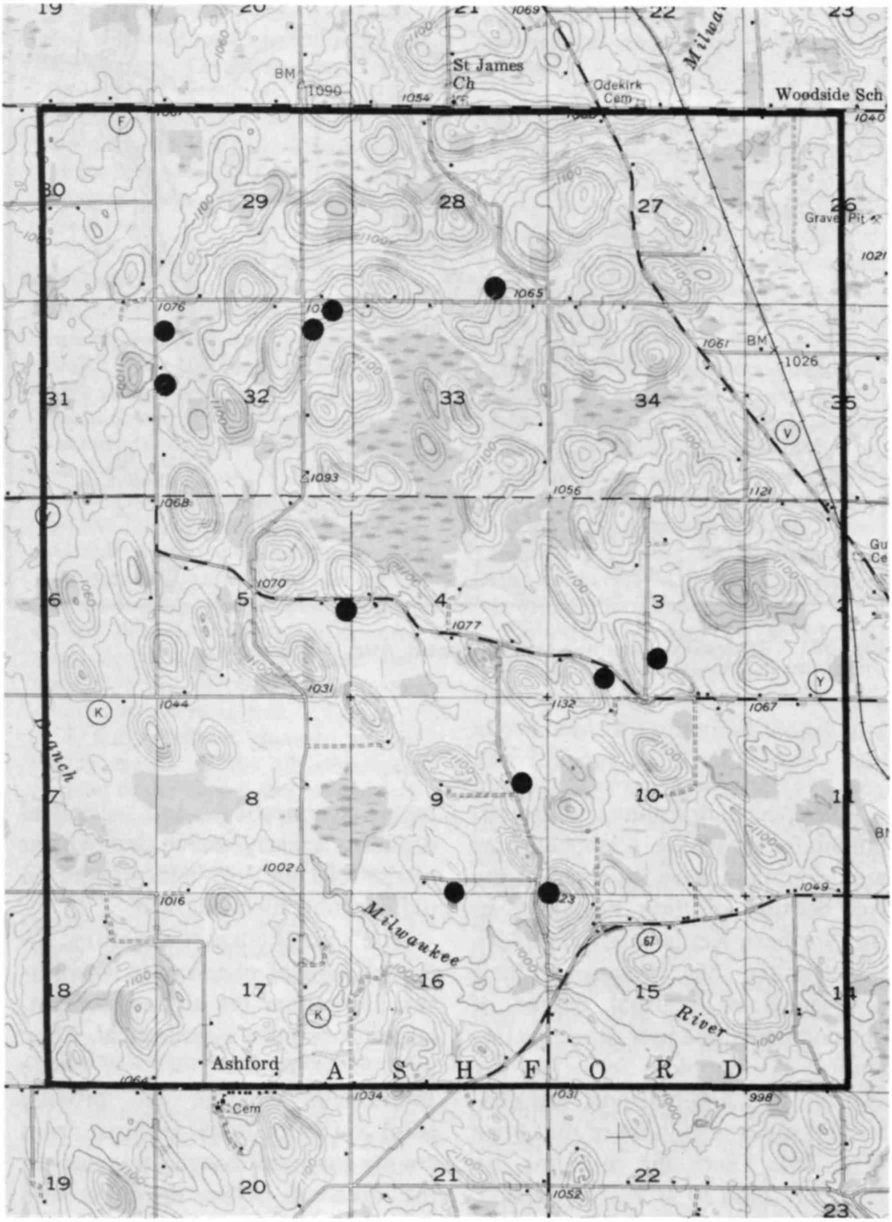

Fig. 39. Part of U.S. Geological Survey Topographic Quadrangle—Campbellsport, showing general area of drumlins recommended for the Reserve. Possible waysides are shown with black dots. Scale 1 mile per inch.

view of the drumlins. Typical views of the Campbellsport drumlins, from some of the recommended overlooks, are recorded in Figs. 40-42.

Inasmuch as most of these drumlins seem to be composed of the pale yellow-brown sandy calcareous till typical of the drumlins and ground moraine over a broad area of the northern part of Green Bay Lobe, it is not likely that many would be destroyed if they were not in public ownership. The till makes good farm land, a relatively firm base for construction, but is only poorly suited for construction aggregate. Consequently, available funds should go mainly toward acquisition of land associated with the Northern Kettle Interlobate Moraine, Devils Lake Park, and the Bloomer Moraine. Two or three of the proposed roadsides shown in Fig. 39, for views of the drumlins, seem sufficient to me to integrate these features into the Reserve.

Larger, well-developed, rounded drumlins may be seen readily from State Highway 23, County Highway AA, and connecting town roads about 5 miles east of Fond du Lac (Figs. 4, 43). Elongated drumlins, with ratios of 5-10:1, are well displayed in the area south of Mayville and south and east of Horicon (Figs. 4, 44). Many drumlins south of Juneau and southwest of Beaver Dam are even more elongated and larger (Figs. 4, 45). Drumlins are less well developed in northern Wisconsin.

Fabric studies of the stones and finer particles in the till in the drumlins in Wisconsin have not been attempted. Few mineralogical and textural data are available. Obviously much subsurface geological study is needed to describe the drumlins adequately and to reconstruct their possible origin.

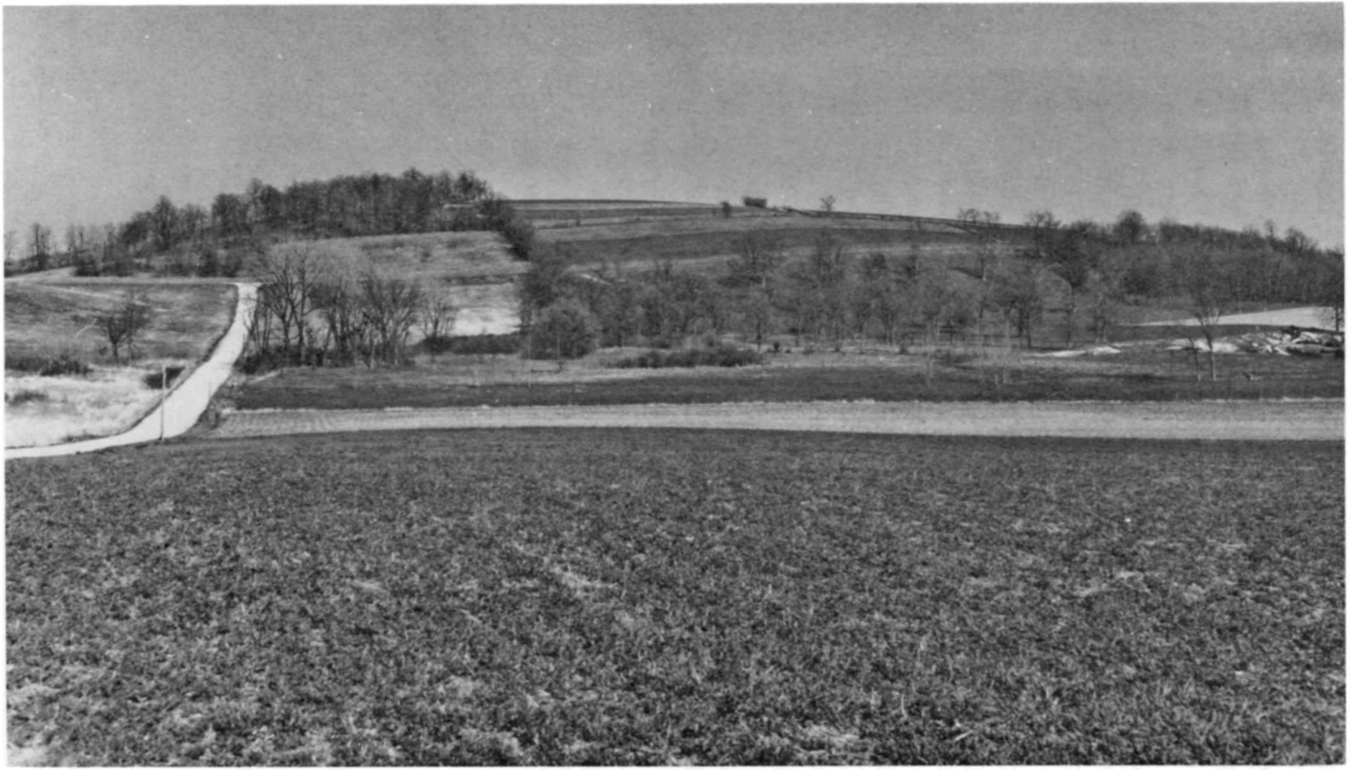

Fig. 40. View eastward of the long-profile of the drumlin with an elevation of 1123 ft, centered at the corners of secs. 9, 10, 15, and 16 (Fig. 39). The steeper stoss (up ice) side shows clearly on the left.

Fig. 41. View southeastward of the cross-profile of the drumlin in the north-central part of sec. 10. Camera position is on County Highway Y, in the saddle in the SE¼ SW¼ sec. 3 (Fig. 39).

Fig. 42. View north-northwest of cross-profile of the drumlin in the center of sec. 28 as seen from the possible overlook on the town road, 0.25 mile west of the southeast corner of that section (Fig. 39).

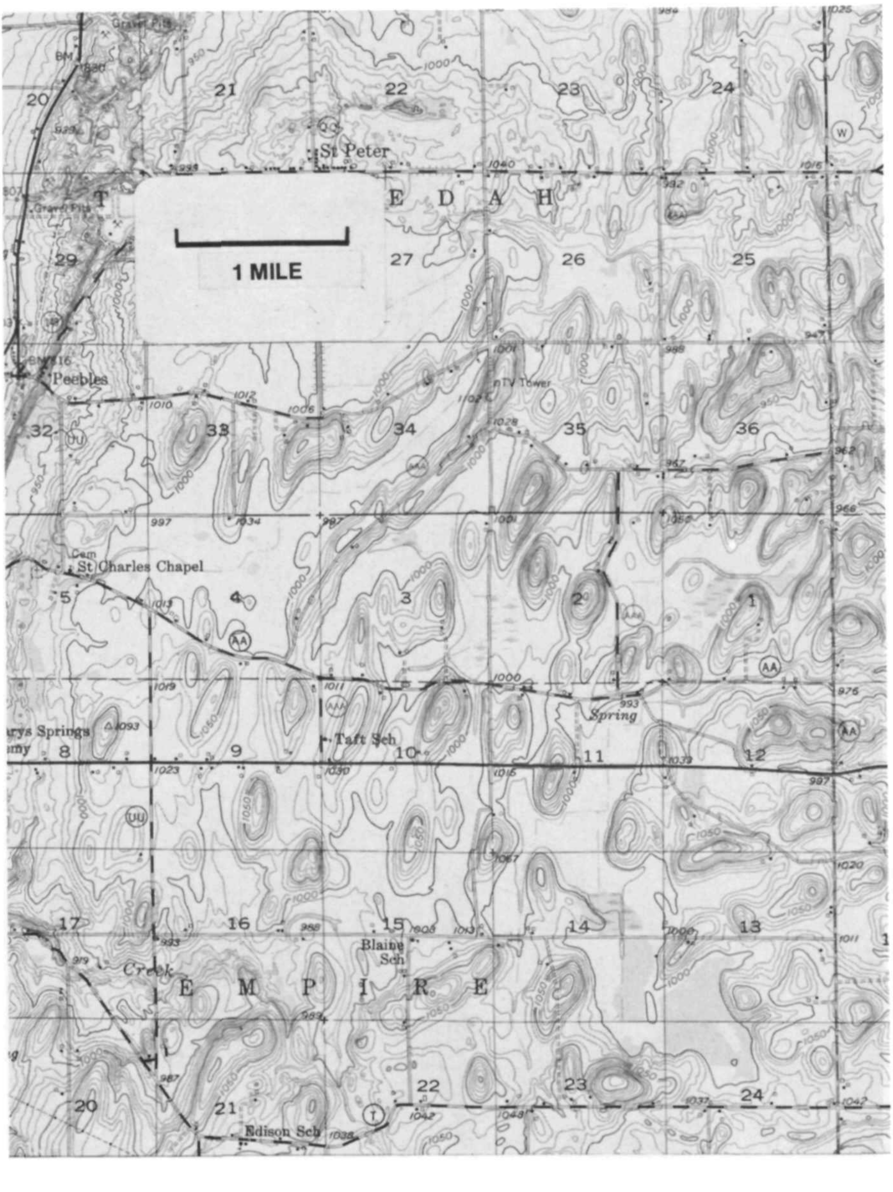

Fig. 43. Part of U.S. Geological Survey Topographic Quadrangle—Fond du Lac, showing drumlins.

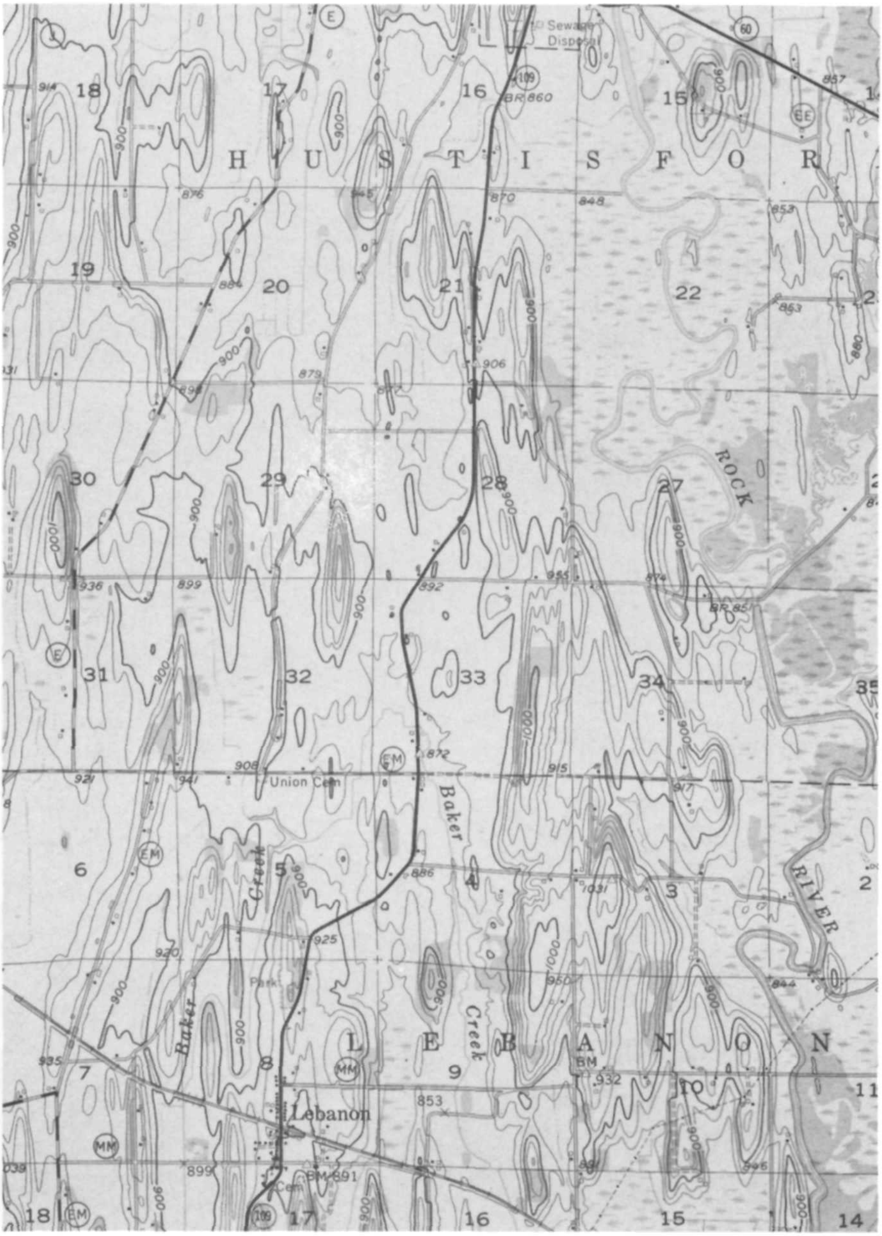

Fig. 44. Part of U.S. Geological Survey Topographic Quadrangle—Horicon, showing drumlins. Scale 1 mile per inch.

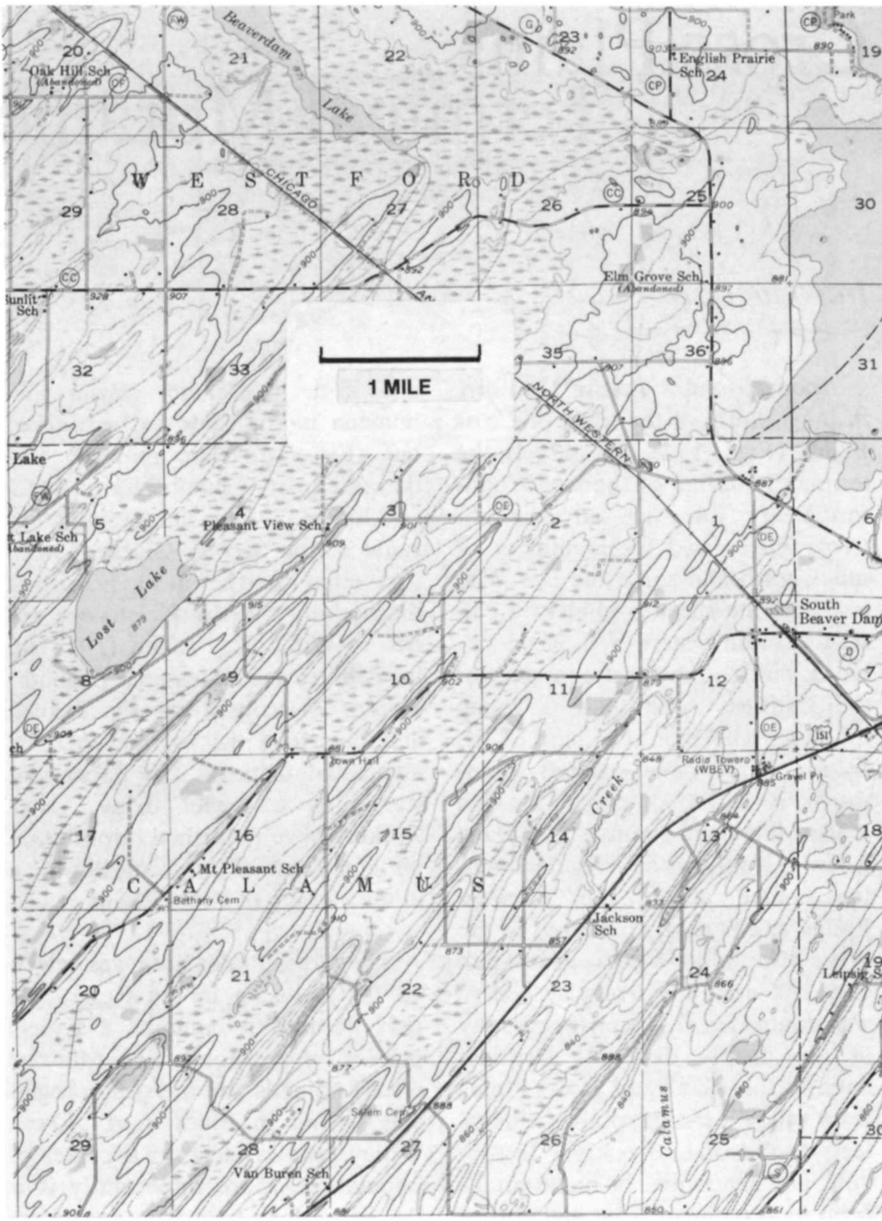

Fig. 45. Part of U.S. Geological Survey Topographic Quadrangle—Beaver Dam, showing drumlins.

6 Cross Plains Terminal Moraine

Introduction

About 10 miles west of Madison, in the vicinity of the town of Cross Plains (Figs. 1, 2, 46, 47) is the terminal moraine of the late Woodfordian or Cary ice advance. At Cross Plains and for a number of miles north and south the late Wisconsinan ice sheet ground to a halt on the southwestern Wisconsin uplands, marking the east boundary of the Driftless Area with a young moraine. The moraine, part of the Johnstown Moraine, extends southward in a broad curve through Adams and Sauk counties into Dane County. It follows an irregular looping course across the Baraboo Range and is partly obscured in the Wisconsin River Valley near Sauk City. Its minutely irregular course was controlled by local topography in the deeply dissected Driftless Area in the vicinity of Cross Plains from where it trends south-southeasterly to Verona, Brooklyn, Evansville, and Janesville. Its name comes from its prominent front and abrupt reentrant angle near the town of Johnstown, east of Janesville.

The deployment of the Johnstown Moraine of the Green Bay Lobe was one of the first major glacial phenomena in the state to be worked out (Fig. 16). The moraine was described by Chamberlin (1883a, 1883b:261-298) as the terminal moraine of the Second Glacial Epoch and cited as the most important discontinuity in the Pleistocene epoch in Wisconsin. Alden (1918) in his detailed reconnaissance of southeastern Wisconsin clearly defined and described the moraine and its associated features. His paper still stands as a model today. Later workers have published information on local areas of the front, but none in the vicinity of Cross Plains.

The Cross Plains area was selected for inclusion in the Reserve in part because it contains a typical portion of the Johnstown Moraine on the uplands and a typical proglacial stream in Black Earth Creek Valley, and is close to a center of population. More importantly it is the only place known to me where the terminal moraine rests directly on well exposed, weathered dolomitic bedrock and where small marginal proglacial lakes, a marginal

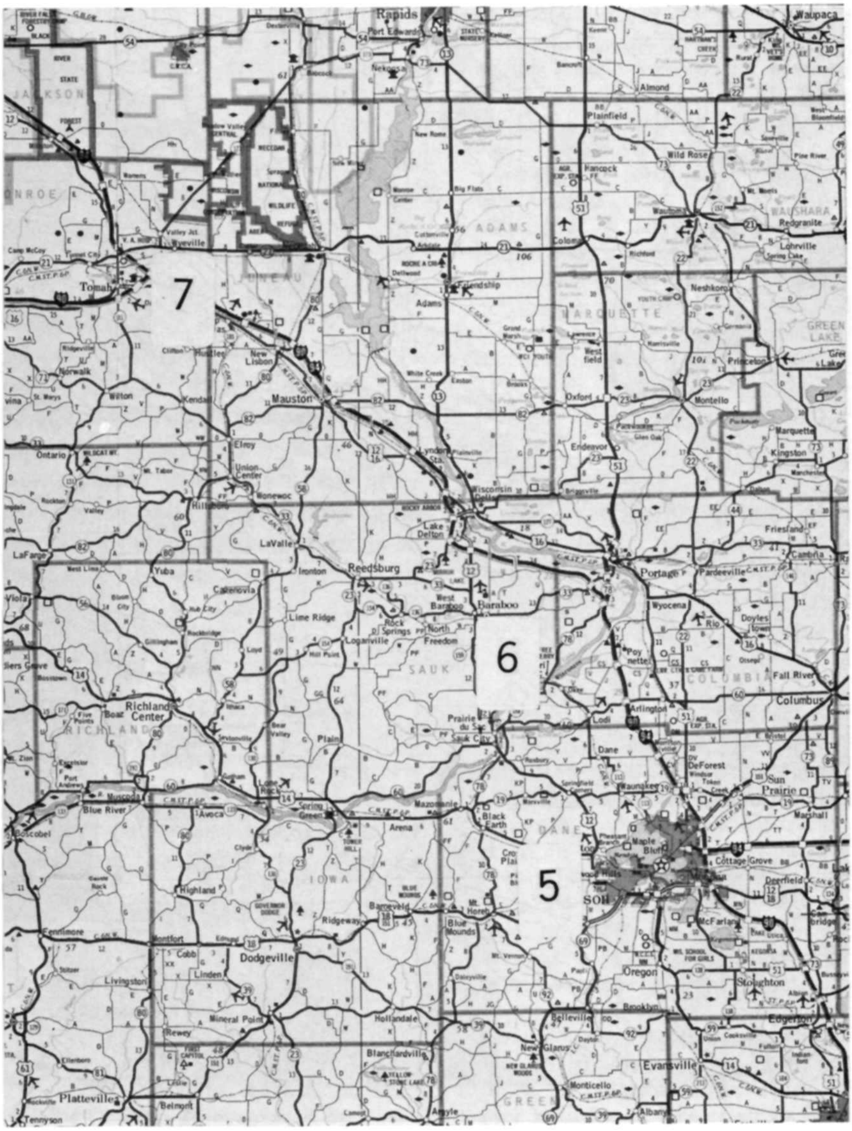

Fig. 46. Part of Wisconsin State Highway map, showing locations of areas 5-7.

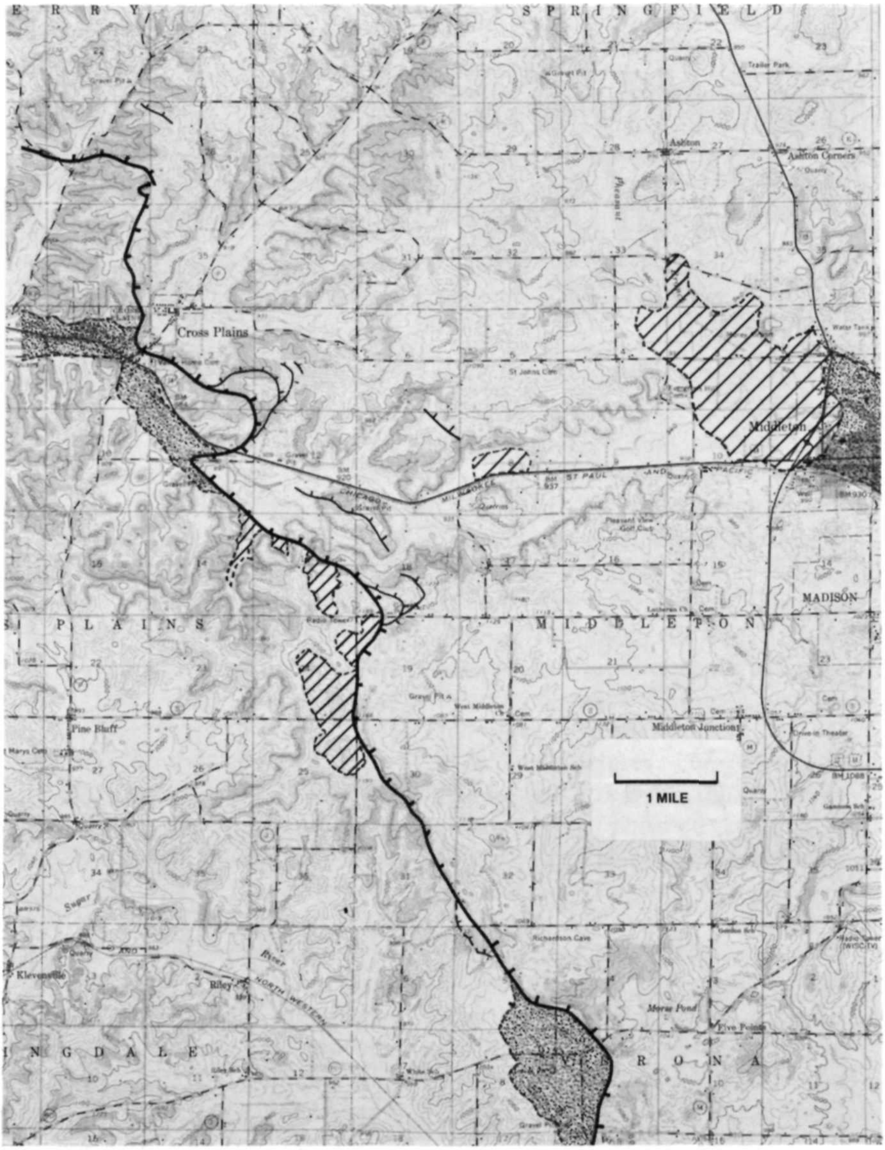

Fig. 47. Part of U.S. Geological Survey Topographic Quadrangle—Cross Plains showing the main Johnstown Moraine (heavy ticked line), some retreatal moraines (lighter ticked lines), a brief advance beyond the main front (light ticked broken line), some glacial lakes (diagonal lines), and outwash (dots).

drainageway, and a subglacial drainageway may all be seen in a small area. Most of the length of the terminal moraine in southern and central Wisconsin fronts on broad outwash plains, in large proglacial lakes, or against older drift. There the relation of the moraine to its adjacent features is clear, but the observer must visit a large area with map in hand to appreciate it. In contrast, the various glacial features associated with the moraine in the vicinity of Cross Plains are more varied and yet as definitive as one could hope to see, all preserved in a neat little package. The area is one of increasing urbanization, and preservation of parts of the front and its associated phenomena can only be assured in the Reserve.

General Description of the Moraine

Alden (1918:212-213) described the distribution and topographic relations of the Johnstown Moraine between Verona and the Wisconsin River as follows:

> Crossing an ancient valley at Verona, where it is cut through by Badger Creek, the moraine continues northwest up a second old tributary to Sugar River. For 1½ miles north of the line of the Chicago, Milwaukee & St. Paul Railway the glacier occupied the valley and left its moraine crowded against the west slope, being separated therefrom by only a sharp narrow ravine 35 to 40 feet in depth. One side of this ravine, which was probably kept open by the southward flow of the glacial waters, is of nearly bare Lower Magnesian limestone [Prairie du Chien, Ordovician, Fig. 3]; the other is formed by the abrupt front of the moraine. Through the next 1 1/2 miles the moraine front rises abruptly 60 to 80 feet from a flat terrace to a well-marked ridge crest, back of which a belt one-half to 2 miles in width, marked by gentle sags and swells and several ponds, extends to a very indefinite inner margin. Near the north line of Verona Township the moraine crosses the old valley and ascends 120 feet to the crest of the Trenton-capped [Platteville, Ordovician, Fig. 3] ridge beyond. Wells indicate thicknesses of 18 to 80 feet for the moraine in Verona Township, the average of 16 measurements being 46 feet. In sec. 5 the moraine is cut through by a narrow ravine 80 to 100 feet in depth, whose lower slopes and bottom expose the St. Peter sandstone and Lower Magnesian limestone.
>
> Outside the moraine near the town line the limestone crest of the ridge is covered only by thin clay soil and scattered boulders which probably came from the moraine or the ice front itself. Here for the first time, tracing it from the southeast, does the moraine reach the Driftless Area and mark the limit of glaciation for this part of the State. For 85 or 90 miles northward from this point no earlier glacier is known to have extended farther west in Wisconsin than the Green Bay Glacier of the later Wisconsin substage; and from this moraine front westward to the Mississippi, a distance of 75 to 80 miles, no unmodified glacial drift has ever been found. The relations of the Green Bay Glacier and its deposits to this thoroughly dissected erosion topography are instructive. For 2 miles in southwestern Middleton Township the ice front lay along the east slope

and crest of the rock ridge and deposited its moraine there and in the heads of the ravines which cut the western slope. It is remarkable that no outwash deposits lead down these ravines away from the moraine. Possibly most of the water drained backward down the east slope of the ridge beneath the ice. In sec. 30 the rock ridge swings westward about a mile across the line into Cross Plains Township. The ice did not press forward to the head of the valley thus extended but deposited its moraine across the valley in such a manner as to leave the upper part an enclosed basin 60 to 80 feet in depth. The west front of the morainal dam is bordered by a narrow flat terrace deposited in a temporary lake which occupied the basin. From this terrace the slope rises abruptly 30 to 40 feet to a narrow crest marked by parallel ridges and sharp kettles and many bowlders from which a long gentle slope drops down eastward to an indefinite inner margin. Mr. Voss's well near the top of this slope, at a point about 30 feet lower than the crest, reached rock at a depth of 130 feet.

North of this valley is a high limestone divide between the Sugar River basin and the Black Earth Valley. The rock ridge, before it was covered by the drift, had a relief of 140 to 200 feet on the south and of 300 feet or more on the north. In overriding this ridge obliquely a notch or reentrant three-fourths mile in depth was developed in the glacial margin. This indicates that the extreme frontal slope of the ice rose about 200 feet in the first mile from the edge. The moraine ascends the south slope, its crest reaching, in the SE¼ sec. 18, T. 7 N., R. 8 E., the highest elevation thus far attained, 1,239 feet above sea level. Just how much drift there is at this point is not known, but rock is exposed in the slope and reached in wells 80 to 100 feet lower beneath 20 feet of drift within half a mile. An average of the thicknesses of drift penetrated in the moraine in Middleton Township by 10 wells is about 60 feet. Continuing northwestward the front of the moraine runs for a mile along the crest of the south bluff of the Black Earth Valley as a small marginal ridge 150 to 200 feet above the bottom of the partly filled valley to the north, where it blocks the heads of two ravines. In the NW¼ sec. 13, T. 7 N., R. 7 E. (Cross Plains Township), it drops down the slope into the valley where, in crossing obliquely, its relief is lost in the general filling of moraine and outwash deposits.

In the NE¼ sec. 11 a narrow marginal ridge 15 to 20 feet high extends up the north slope of the valley and thence across the heads of ravines and intervening ridges one-fourth mile or less back of the crests of the bluffs, which rise abruptly 100 to 140 feet from the flat floor of the partly filled valley. The surface of this drift is thickly strewn with erratic bowlders, but outside its margin not a piece of drift is found in the thin clay soil. Within the marginal ridge, drift marked by slight sags and swells mantles the rock ridge, but the moraine is not bulky and its inner limit is poorly defined.

The converging of valleys from the east and northeast at Cross Plains led to the ice front crowding forward slightly into the narrow opening between the heads of opposite salients. Here the moraine is pitted with slight sags through a width of a mile and has a relief of 20 to 40 feet above the flat outwash plain to the west. The contrast between the craggy bluffs capped with Lower Magnesian limestone outside the moraine and the smoothly rounded slopes within it is very striking in the vicinity of Cross Plains.

Northwestward from Cross Plains to the Wisconsin River valley, a distance of about 10 miles, the limit of glaciation is generally plainly marked by a narrow marginal ridge with plentiful bowlders. This ridge is a few rods in width and rarely more than 20 feet high, but it is traceable continuously across a greatly dissected topography of five rock ridges 200 to 250 feet in height and four intervening valleys. After leaving the valley at Cross Plains the surface shows, for the most part, only the smooth undulating contours of ground-moraine topography. It looks as though the bulk of the morainal deposit was not formed at the limit of the advance in this part but is represented by the morainal belt which leaves the marginal ridge just north of Cross Plains and thence northward lies about 2 miles farther east.

For a mile north of Cross Plains the ice pressed against the east slope of the rock ridge fitting snuggly into the ravines, as shown by the little marginal ridge which encircles their heads. The crossing of this first ridge causes a reentrant of nearly three-fourths mile, the ice having pushed forward in the next valley. Two to three miles farther northwest a reentrant of one-half mile resulted from overriding a ridge 250 feet high, the ice pressing forward in the valley west of Marxville. So also on the broad low tract near Wisconsin River the ice extended about a mile farther west than on the narrow crest on the south, which rose 300 to 400 feet higher.

The deposit which separates from the marginal ridge north of Cross Plains merges with it again in the broad valley west of Roxbury. The limits of this morainal belt are very indefinite, its presence usually being indicated only by sags and kettles which pit the surface. A mile west of Martinville, however, in adjacent parts of secs. 11, 12, 13, and 14, T. 8 N., R. 7 E. (Berry Township), some of the most strongly marked morainal topography within the region occurs. One bulky ridge which appears to be drift has a relief of 160 feet on the west; and its crest stands 1,249 feet above sea level, or about 400 feet above the rock bottom of the partly filled valley at Marxville, 3 miles to the northwest. The average thickness of drift penetrated by 16 wells in this morainal belt in the towns of Cross Plains and Berry is about 55 feet. These depths range from 25 to 95 feet, several of them not reaching the base of the drift.

The above statement is a clear description of that part of the Johnstown Moraine in the vicinity of Cross Plains. When the conditions and methods of study imposed upon Alden (1918) in his reconnaissance of all southeastern Wisconsin are taken into account, his accomplishments and insight into the Pleistocene geology of the region are nothing short of remarkable. His map, published at a scale of 4 miles to the inch, could not show all the details portrayed by the new topographic quadrangles at about 0.4 mile to the inch and 10-ft contour interval nor by aerial photographs. To this day, only details of the story need be changed. Alden (1918:209-217) clearly recognized that not all of the drift in the Johnstown Moraine was deposited during the one substage, that the thickness of the drift varied markedly from segment to segment of the moraine, and that the outermost front of the Johnstown Moraine was not everywhere synchronous nor representative of equal periods of time. Alden (1918:220-222) also demonstrated that the bulk of the pebbles and stones in the moraine were derived from rocks that crop out in the vicinity and that only 5-20% were derived from Precambrian igneous and metamorphic rocks from northern Wisconsin, Upper Michigan, or Canada. Keewenaw copper nuggets from Upper Michigan and one diamond, presumably from Canada, are among the least common constituents. Dolomite, chert, and sandstone of the local formations (Fig. 3) are most abundant.

Details of the Recommended Areas

Two areas are recommended for acquisition—a northern and southern. The northern area is the north-central quarter section of sec. 13, T. 7 N., R. 7 E. (Fig. 47,48) and the southern contains portions of secs. 4 and 9, T. 6 N., R. 8 E. (Fig. 47). Several possible waysides can be found in Fig. 47, where excellent views of the Johnstown Moraine and other features may be seen. The northern area (Figs. 47, 48) contains an excellent and typical part of the terminal moraine (Figs. 49, 50), drained proglacial lakes, a marginal drainageway (Fig. 51), a subglacial channel, and weathered dolomitic bedrock with large erratics on it (Figs. 52-55). The southern area contains a representative portion of the terminal moraine breached by a drainageway that followed a preglacial bedrock valley, kettle ponds, and an outwash apron (Fig. 56).

A portion of the Cross Plains topographic map is reproduced in Fig. 47, showing the outer edge of the Johnstown Terminal Moraine, some of the fronts established during retreat (and a position occupied briefly beyond the main front), and some of the marginal lakes and outwash as interpreted by me largely from aerial photographs. The two ponds in secs. 24 and 25 are now separated by Mineral Point Road (County Highway S) (Fig. 57). They are the remnants of a former single proglacial lake that filled the basin to about 1155 ft in elevation. The lowest pass from that basin to the west into the headwaters of the Sugar River is about 1175 ft; no evidence that the former lake ever drained through it has been found. Instead it seems to have drained northward across a bedrock ridge of the Platteville limestone (Ord., formerly Trenton of Alden 1918) at about 1155 ft into the adjacent proglacial lake at the same elevation. That lake was short lived, being held in by ice that only temporarily filled the valley 0.4 mile east of the radio tower in the extreme northeast corner of sec. 24. (The gravel pit shown in Fig. 47, in the SE¼SW¼ sec. 18, is actually a small quarry in the Platteville-Galena group). Water from the two lakes to the south flowed northwestward marginal to the ice from the vicinity of that pit, past an outcrop of the St. Peter sandstone on the southwest side of the valley, into another small proglacial lake in

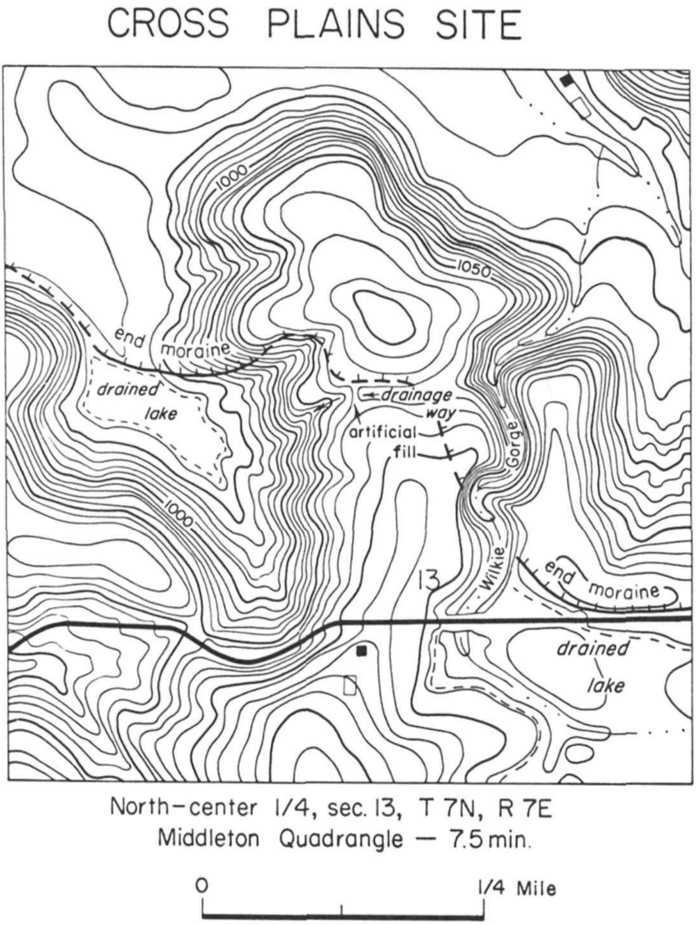

Fig. 48. North-central part of sec. 13, T. 7 N., R. 7 E., after U. S. Geological Survey Topgographic Quadrangle—Middleton.

sec. 13, at an elevation of about 1090 ft. The terminal moraine lies on the northeast side of that valley, although large foreign erratic boulders may be seen to the southwest of the intermittent stream. Water from the large proglacial lake in sec. 13 briefly flowed across the bedrock spur shown in the center of Fig. 48, through the drainageway indicated (Fig. 51) leaving bare weathered dolomite of the Prairie du Chien Group exposed in ridges (Figs. 53,54) between bifurcating distributaries as the water plunged westward from the steep face. The bare dolomite is solution etched into bizarre forms (Figs. 54,55). Large foreign

Fig. 49. End moraine on upland in southeast corner of Fig. 48, looking westward. An outwash area and drained lake are in the extreme left part.

Fig. 50. Cary end moraine in center blocking a small drained lake basin behind it in the west-central lowland of Fig. 48. View south-southeastward.

Fig. 51 View northeastward of drainageway in center of Fig. 48. Prairie du Chien dolomite crops out in foreground; the Johnstown End Moraine blankets the opposite side of the channel.

Fig. 52. Basic igneous rock erratic resting on Prairie du Chien dolomite at west edge of Wilkie Gorge, about where end moraine crosses (Fig. 48). View eastward.

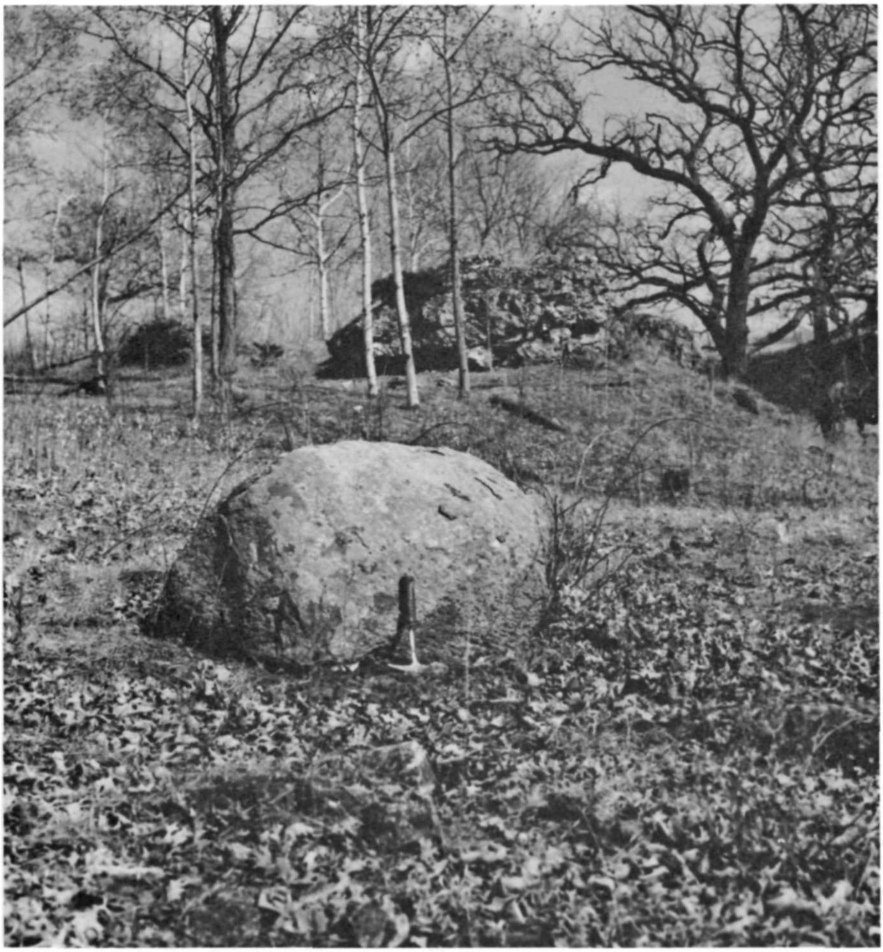

Fig. 53. Basic igneous rock erratic resting on eroded Prairie du Chien dolomite with weathered dolomite in the ridge behind and a distributary channel of the drainageway (Fig. 48) on the right. View eastward.

Fig. 54. Detail of Prairie du Chien dolomite ridge in Fig. 53.

Fig. 55. Solution effects of Prairie du Chien dolomite on top of ridge between distributary channels west of artificial fill (Fig. 48).

Fig. 56. Breached end moraine in background and outwash apron in foreground, looking east-southeastward from possible wayside shown south of County Highway PD in sec. 9 (Fig. 47).

Fig. 57. Proglacial pond on north side of County Highway S as viewed from the Johnstown Terminal Moraine (Fig. 47).

erratic boulders are scattered on the dolomite (Figs. 52, 53). That drainageway apparently was occupied only for a short time by the overflowing lake waters which soon began flowing down the Wilkie Gorge and under the ice. Water from about 2.3 miles along the front of the Cary glacier thus flowed northward along the front, from one proglacial lake to another, until finally cascading to the lowland. At first it flowed into the small drained lake basin shown in the western part of Fig. 48 and thence along the margin of the ice in Black Earth Creek Valley (Fig. 47). Shortly thereafter it cascaded down Wilkie Gorge beneath the ice.

The amount of material deposited directly by the ice in this part of the Johnstown Moraine varies markedly from point to point. Alden (1918:218) records the log of a well at the home of Mr. Voss (NW¼ sec. 30, T. 7 N., R. 8 E.) believed to be that which is 0.2 mile east of the road junction at 1166 ft on County Highway S east of the two ponds. The well penetrated 75 ft of clay and 55 ft of sand and gravel on top of the St. Peter sandstone. However, I found dolomite just 2 ft below the surface of the gently dipping slope of the outwash apron of the Johnstown Moraine, in the extreme southeast corner of sec. 24, and dolomite crops out 0.50 mile north, in the same ridge. Thus the thickness of till in the moraine seems to be no more than 40 ft at its crest which lies on the westerly rim of a preglacial valley. The moraine is even thinner to north and south from County Highway S. The amount of fill in the basin of the proglacial lake bisected by County Highway S probably is several tens of feet although no subsurface exploration has been attempted. Erratics have been found on the west side of the basin in the vicinity of the farm house. Whether carried there by ice rafting or by glacial ice directly is not known.

The gully crossing the drained lake basin in the southeast corner of Fig. 48 exposes 8 ft of silt on 7 ft of clean, poorly sorted sand and gravel. The base of the section was not seen. The upper silt resembles loess but contains more clay and sand and is believed to have been deposited in the former proglacial lake on top of deltaic sediments and outwash. The axis of the former valley occupied by the proglacial lake in the south part of sec. 13 lies to the east of Wilkie Gorge, about in the position of the town road that descends to the north along the east margin of Fig. 48. That axis is choked with glacial debris. Wilkie Gorge exposes the Prairie du Chien dolomite (formerly Lower Magnesian limestone of Alden 1918) up to the vicinity of the town road crossing the southern part of Fig. 48.

The end moraine to the east of the gorge is only about 20 ft thick. The moraine north of the drainageway (Fig. 48) is the same order of thickness. The upper part of the Prairie du Chien dolomite, as indicated by oolitic chert and sandstone layers in dolomite, crops out on the south side of the drainageway up to 1080 ft, and locally on the flanks of the spur at about the same elevation to the north.

Black Earth Creek Valley contains many tens of feet of glacial outwash (Dury 1964:11) whose bottom has not been reached in the vicinity of Cross Plains. The gravel pit operations 1 mile southeast of town expose at least 50 ft of coarse gravelly outwash. This outwash built up in the valley choking the mouths of tributaries downstream and forming lakes or swamps in them. Whether it is all late Woodfordian in age is not known.

The late Woodfordian or Cary ice quickly retreated slightly from its maximum position in several places north and south of Cross Plains. Only a few of the retreatal moraines are indicated in Fig. 47. One is crossed in the valley by the town road that trends north along the east margin of sec. 13 (Fig. 48). It forms a conspicuous ridge trending northwesterly on top of and west from the prominent bedrock-supported "island" in Black Earth Creek Valley, 3 miles southeast of Cross Plains. That "island" formed a distinct barrier to ice flow and to the later melt water which passed westward both on the north and on the south. Small kettles with ponds are conspicuous on its north side. Many tens of feet of lacustrine sediments are found in the south channel. Loess on top of outwash and on the moraine in Black Earth Creek Valley obviously means that part at least postdates the withdrawal of the Cary ice from its extreme position at the Johnstown Moraine (Dury 1964:13).

Glacial Lake Middleton (Alden 1918, Fig. 47) has had a long and complex history which is not clearly understood. Other shorter-lived lakes occupied parts of Black Earth Creek Valley, of which only one small one is shown in Fig. 47. The complex relationship of the Milton morainic system, thought by Alden (1918) to be a retreatal phase of the Cary, to the Johnstown Moraine will not be discussed here. Even though it affected Black Earth Creek Valley directly, its ice did not reach closer than 2 miles to the Johnstown front. Its effects on the history of the recommended sites are indirect, i.e., through loess formation or climatic modification.

The southern area recommended for inclusion in the Reserve includes the SW¼ sec. 4 and the northcentral ¼ sec. 9, T. 6 N., R. 8 E. Because of a bedrock ridge in the northern part of the area, the Cary ice left a prominent, but not thick, moraine with steep outer face, and small kettles in the drift behind the front (Fig. 47). Relief of the kettles is only a few feet. The steep outer face of the moraine rises 100 ft above the bedrock spurs on which it fronts, but only part is drift. A deep drainageway, cutting through the moraine in the center of sec. 5 exposes St. Peter sandstone and Prairie du Chien dolomite below the moraine, as does the marginal drainageway extending southward from it. Bedrock is also exposed in the road-cuts and in the gravel pits in the outwash area shown in Fig. 47. Drilling between the gravel pits disclosed bedrock at a depth of a few feet. Thus the moraine and outwash are only a relatively thin veneer mantling a stream-dissected bedrock topography. The top of the moraine rises over bedrock highs and drops in the preglacial valleys.

Dating of the outwash has not been done. Alden (1918, Pls. I, III) shows his boundary of older drift encompassing the outwash area and terminating against the Johnstown Moraine at about the center of sec. 5.

One of the deeper preglacial bedrock valleys may be traced from the gap in the moraine, south of County Highway PD, northeastward to Five Points and Morse Pond, and thence northeasterly to the marked steep-walled valley northwest of the WISC radio tower. Dolomite and sandstone of the St. Peter and Platteville-Galena Formations crop out on both sides of the valley which is occupied partly by kettle lakes, the largest being Morse Pond. Typically along the Johnstown front, areas of outwash are localized by preglacial topographic lows in the bedrock. There the glacier could maintain its thickness with a lower surface elevation, which in turn concentrated surface runoff at those points. The highest bedrock ridges have negligible outwash, as is seen between the two recommended areas or between the southern area and Verona where another preglacial valley is found. The one at Verona is larger and topographically lower than the one in the southern area (Fig. 47) and consequently contains far more outwash. The same principle applies to the Rock River Valley at Janesville where still lower topography concentrated many times more runoff and several hundred feet of outwash, not all of which can be attributed to the Cary ice.

The contrast of drift-mantled surfaces behind the Johnstown front with only a thin loess mantle beyond the front can be seen easily from a car. The topographic map (Fig. 47) shows somewhat more irregularity of topography in front of the moraine than behind, but the effect of glaciation close to the front has been one mainly of filling in the lower areas rather than of eroding the higher areas. Drift on many of the uplands behind the front is only a few feet thick, but in valleys it is tens of feet thick. The main valleys in front of the moraine have also been filled with tens of feet of outwash and loess-derived colluvium. Hence, only valley sides and tops of hills beyond the front show the paucity of cover. Erosion by frost action, gravity movements, and surface runoff apparently was far greater during glacial times than now, and temporary features like pinnacles and castelated spires of easily eroded sandstone are fairly commonplace within the Driftless

Area. Small sandstone pinnacles may be seen 1 mile west of Pine Bluff, on the south tip of a spur of St. Peter sandstone. A larger pinnacle is in a roadside park on Highway 92, 1.5 miles northwest of Mount Vernon. Devils Chimney 2.3 miles southeast of Mount Vernon (New Glarus quadrangle) is still larger (Fig. 58). Both of these pinnacles are thought to be related to a former glacial lake that occupied the West Branch of Mount Vernon Creek, but the details of these and other pinnacles in the area are beyond the scope of this discussion. It is hoped that some may ultimately find their place in the Reserve.

The problem of whether the Driftless Area was ever glaciated is not of immediate or direct concern to the two recommended areas. They provide no evidence either for or against the concept. The solution effects in the Prairie du Chien dolomite could have been accomplished in the 13,000 years the rock presumably has been exposed, but it has not been proved that a longer time was not utilized. I suspect that the solution phenomena have been exhumed by the glacial melt water and were formed during an earlier weathering cycle. The thin young loess cover and lack of residual materials on the bedrock outside the front prove that any older loess and residuum have been removed or were never deposited. The timing of this event cannot be dated except indirectly in other areas. No evidence of older drift outside the Johnstown Moraine between the town line north of Verona and the Wisconsin River has been recognized, yet no old accumulation of residuum nor old loess has been recognized either. Why? Their removal from the hills and completely out of the drainage network as well is not explicable by normal runoff such as is experienced today. This problem will be discussed further in Chapter 11.

Fig. 58. Devils Chimney a pinnacle of St. Peter sandstone as viewed northeasterly.

7
Devils Lake Park

Introduction

Devils Lake Park in the Baraboo Range, Sauk County, presently contains about 5 mile^2 of scenic cliffs, wooded hills, and Devils Lake itself (Figs. 1, 2, 3, 46, 59). Topographically the park is mostly a rolling upland, near 1400 ft above sea level, cut by a steep-walled, L-shaped gorge whose floor is generally 500 ft below the summit (Fig. 59). The north-trending part of the gorge is occupied by Devils Lake (Fig. 60). Ancient rocks of Cambrian and Precambrian age in the eastern part of the park (Fig. 3) largely are covered by glacial deposits and Recent wind-blown silt. The record in the rocks and in glacial and periglacial features is especially rich. To preserve unique and representative glacial features in the area the park must be tripled in size following approximately the boundaries shown in Fig. 59.

As a field laboratory in earth history, the Baraboo Range is one of the most valuable and fascinating in the Upper Mississippi Valley region. Besides being the most popular park in Wisconsin, the Devils Lake area has been the locus of field trips for thousands of students each year. In spite of the great amount of study given the area by scientists over the decades, new information continues to appear. Hopefully, there will be renewed interest for study of the park by this recommendation for its inclusion in the Ice Age National Scientific Reserve.

To understand the deployment of ice in the park and the origin of many periglacial features it is necessary to understand the bedrock geology and geomorphology of the area. This note first summarizes the geomorphic development of Devils Lake Park and then comments on certain unusual or striking features in and adjoining it. Emphasis is given to description of the glacial and periglacial features. I have attempted to bring together the new information available since the last major summaries (Salisbury and Atwood 1900; Trowbridge 1917; Alden 1918). It is hoped that the features illustrated here can be seen and appreciated, but not destroyed, by the thousands of visitors each

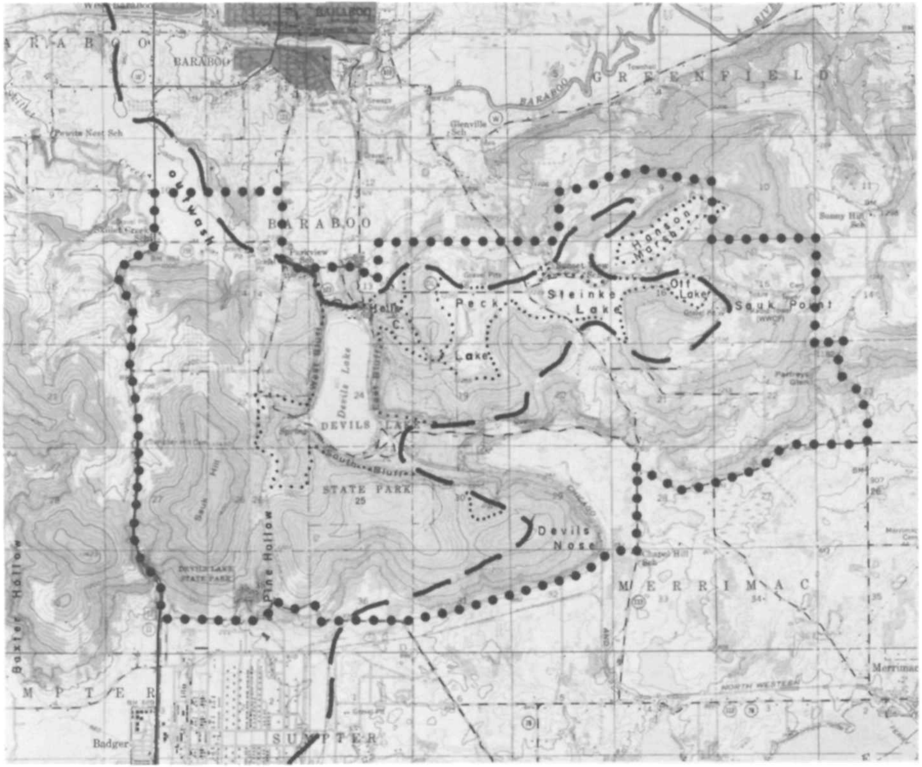

Fig. 59. Parts of U. S. Geological Survey Topographic maps—Baraboo and North Freedom, showing the position of the prominent Cary end moraine with heavy dashed line, the outlines of former glacial lakes with small dots, and the outline of the minimal recommended area with large dots.

Fig. 60. Air view looking southward of Devils Lake, its morainal plug in the foreground, the quartzite bluffs surrounding the lake but breached in the southeast corner, and the distant broad flat of the Wisconsin River.

year who come to Devils Lake. In their zeal, geology students particularly have contributed importantly to the natural attrition of certain exposures in the park. This must cease. Pressure of man's use has increased to the point where even the durable rocks now need protection.

Geomorphology

The story of Devils Lake Park must begin about a billion years ago, in latter Precambrian time, with the deposition in shallow seas of many hundreds of feet of very clean, well-winnowed quartz sand of medium-grain size. Subsequent burial in the earth's outermost crust and accompanying alteration during late Precambrian time lithified the rounded to subangular sand grains into the brittle Baraboo quartzite in which the gap containing Devils Lake has been cut. The lithification involved little or no crushing of the sand grains—only deposition of secondary silica cement in interstices. This makes the total rock very hard yet brittle so it breaks across grains.

Large joint blocks are commonplace and lead to the formation of extensive talus and jagged cliffs (Fig. 61). The characteristic pink, red, and lavender hues are attributed to finely disseminated iron oxides in very small amounts. Ripple marks, mud cracks, and cross-cutting stratification typical of the former marine environment are widespread in the park. The individual grains of sand and some pebbly and bouldery zones are still easily distinguished today.

Perhaps in part during metamorphism of the sand to quartzite and certainly afterwards the area was folded into a large basin by a mechanism not fully understood. Perhaps more than one episode of regional stress made itself felt in minor structures now visible in the

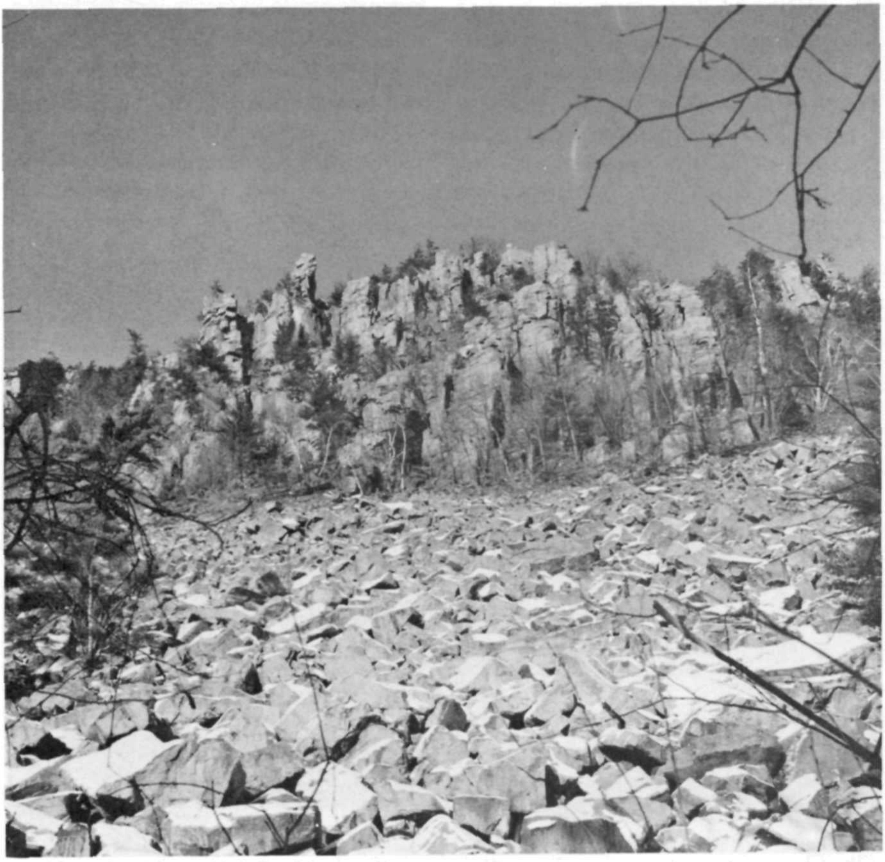

Fig. 61. Talus surmounted by castellated cliffs of Baraboo quartzite on the south face of East Bluff of Devils Lake.

Baraboo Range. Devils Lake lies in the center of the south limb of the syncline or basin (Fig. 3 shows the outcrop pattern of the south limb and of the east nose), where the gentle north dips and local gentle undulations of the quartzite are readily discerned in the cliffs of the West and East bluffs overlooking the lake (Fig. 62). Local, very gentle south dips of the quartzite are found in the cliffs 2-3 miles east of Devils Lake.

Fracture cleavage—a parallel splitting of the quartzite easily confused with bedding—at Devils Lake dips at angles greater than the bedding planes of the quartzite (Weidman 1904; Hendrix and Schaiowitz 1964). It too aids in the formation of joint blocks, talus, and jagged cliffs. Such fractures are considered "normal" in their orientation with respect to the stresses that are inferred to have produced the syncline (Hendrix and Schaiowitz 1964). So are minor drag folds in thin argillites (clayey zones) at Devils Lake, but other minor structures including small folds, slip cleavage, and shears are considered "reverse" by Hendrix and Schaiowitz (1964). The normal

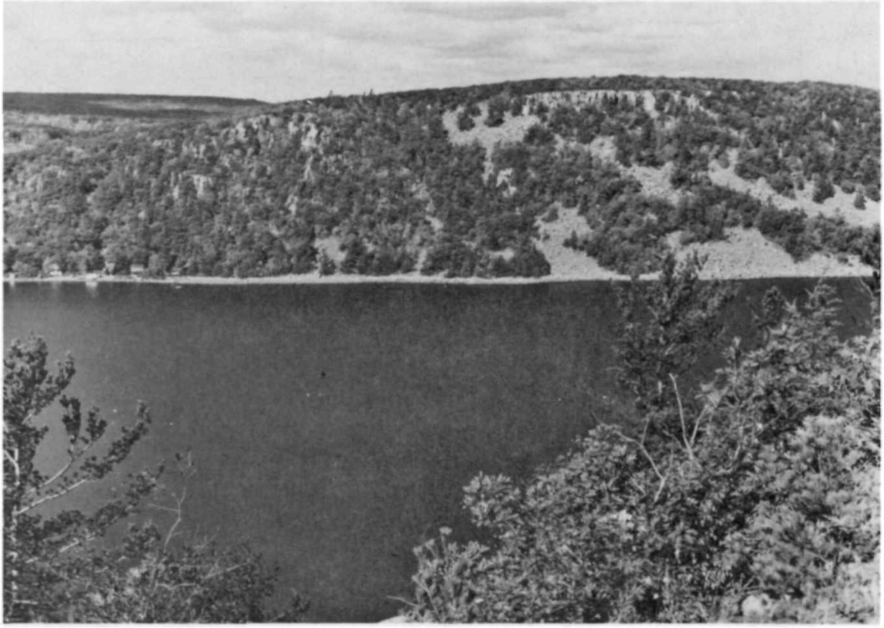

Fig. 62. View westward of West Bluff of Devils Lake, showing talus and jointed beds of the Baraboo quartzite dipping 10° northward.

minor features are confined to thin silty argillite layers interbedded with quartzite, whereas the reverse minor features are in thick argillite beds. Extensive exposures of the reverse structures are in Skillet Creek, about 1 mile northwest of Devils Lake; a small outcrop is just inside the present northwest entrance to the park (Hendrix and Schaiowitz 1964). (Both exposures are rapidly being destroyed by the promiscuous hammering of geology students who do not realize that more can be seen on a weathered surface than on a fresh one.)

The details of the folding mechanism of the quartzite are interesting but not especially germane to the problem of the present day surface features of the park even though their results are. As pointed out, the fracturing of the rock has made frost work especially effective in the formation of pinnacles, talus slopes, and bizarre forms (Figs. 61-64).

The younger Seeley Slate Formation and the Freedom Formation of iron-bearing slate, chert, and dolomite of Precambrian age (Fig. 3) are on the Baraboo quartzite southwest of North Freedom but have not been recognized in the park. They were the object of a flourishing iron ore exploration program in the late 1800s (Weidman 1904), but no mines are operating today. They do not produce any striking surface features.

Some time after the folding and uplift of the Baraboo Range, subaerial erosion (Trowbridge 1917) and probably marine shore erosion also (Thwaites 1958) developed relief of a thousand feet between the top of the range and the surrounding beveled Precambrian igneous and metamorphic rocks. Such relief was due almost entirely to the great resistance to weathering and erosion of the quartzite. Small bodies of igneous rocks crop out in Baxter Hollow, southwest of Devils Lake, and in isolated bodies farther west and on the north and east parts of the range. However, in most of the area surrounding the range, the igneous rocks are buried beneath thick accumulations of sand of Upper Cambrian age (500-550 million years old) (Fig. 3).

During the development of the relief, beveling of the upland quartzite obliquely across the bedding produced surfaces which look smooth to the eye and have long been called peneplains (Trowbridge 1917). The interpretation that the region was in the end-stage of one or more cycles of erosion is now discredited (Thwaites 1958, 1960). Nonetheless the mode of beveling of the resistant quartzite at such marked elevations above the surrounding plains is not truly understood. Certainly toward the close of the erosion cycle marine waters again inundated the quartzite.

Thick accumulations of sand were piled around the range which for a time stood as islands in the shallow seas, shedding their characteristic pinkish rocks into the surf zone to be transported downwind and along

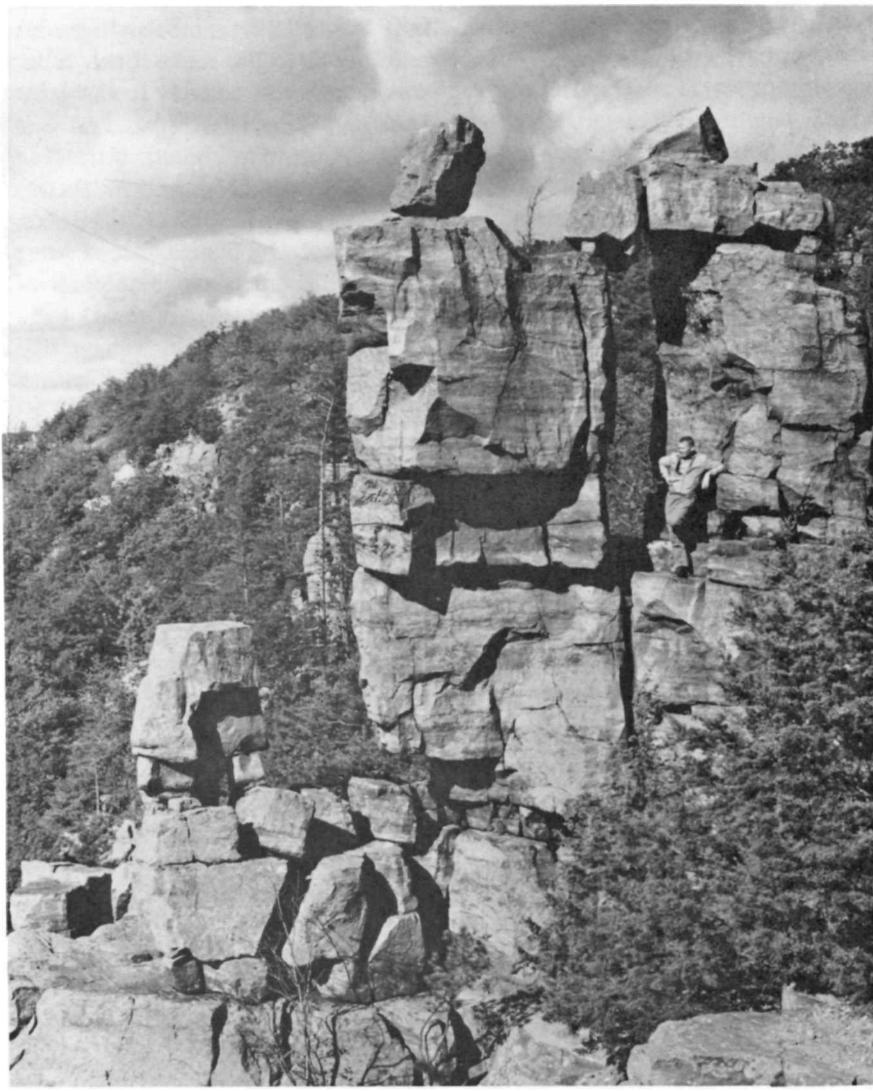

Fig. 63. Devils Doorway on the south-facing slope of East Bluff of Devils Lake, formed by frost action from the well-jointed Baraboo quartzite. See Fig. 65.

Fig. 64. Balanced Rock, a joint block of Baraboo quartzite, isolated by frost action and rock falls of adjacent material, on the west-facing slope of East Bluff of Devils Lake. See Fig. 65.

shore to inevitable burial (Raasch 1958). Thus we find pebbles and boulders of rounded quartzite to the south but only a short distance north. Quartzite pebbles are found locally from the Cambrian basal conglomerate up to the Platteville Formation of Ordovician age. The Cambrian sands not only lapped onto the flanks and filled the center of the syncline, but they also filled channels cut into the quartzite by supposedly ancient streams. The angular unconformity of the sands with respect to the beveled quartzite is striking in many places as is also the abrupt textural and compositional change to the basal conglomerate. A possible wave-cut terrace lies on the northeast part of Happy Hill, 6 miles west of Devils Lake (Thwaites 1958).

Gaps cut entirely through the range are common on the narrow, steeply dipping north flank (too narrow to show in Fig. 3). Only one is known, that with Devils Lake, on the broad south flank (Trowbridge 1917; Alden 1918:105-107). Some, such as at least part of Devils Lake Gap, are definitely Precambrian in age for they contain Cambrian sandstone; others probably are post-Paleozoic and still others were modified by streams as young as Pleistocene to Recent age. Hanging valleys in the quartzite of the south flank are anomalous also. They are broad and gently dipping in their upper reaches and plunge precipitously to the buried Precambrian surface hundreds of feet below. Some are filled partly with Cambrian sandstone, so date from the Precambrian erosion cycle; some also have narrow notches cut into them that may postdate the Paleozoic. The distribution of the hanging valleys in the Baraboo Range is not known nor is their origin. Pine Hollow in the southwest corner of the park, southwest of Devils Lake, is typical (Thwaites 1958).

Cambrian sandstone crops out near the northeast and northwest corners of Devils Lake, in the gorge east of Devils Lake and continuing eastward to Parfreys Glen, near Koshawago Springs and along Messenger Creek southwest of the Lake to the headwaters of Pine Hollow, and in a considerable area in Skillet Creek. It has not been found in the deep valley under Devils Lake itself which is filled with glacial sediments. Cambrian sandstone also is common along Highway 12 where it crosses the south limb, and continues westward to Baxter Hollow where it produces striking cliffs.

Paleozoic sediments continued to be deposited around and over the Baraboo Range probably until Silurian or possibly Devonian time (Fig. 3), with erosional intervals such as that below the St. Peter Formation (Wanenmacher et al. 1934; Thwaites 1961). However, in the park only the Upper Cambrian sandstone lies on the quartzite. The oldest unit exposed is the Galesville Member of the Dresbach Group. It is thickbedded and mostly white or very pale yellow. It is the unit that develops striking cliffs and steep slopes in Baxter Hollow west of the park. The next younger formation is the dolomitic, fine-grained Franconia sandstone that forms local benches, cliffs, and crags that are greenish grey in contrast to the Dresbach. Still younger rocks are more distant from the park today although they may have been present in the geologic past. Chert nodules and clay on top of the quartzite west of Devils Lake are thought to have been "let down" during weathering of the dolomitic formations of Ordovician age (Thwaites 1958).

Peneplaination of the upper quartzite surface also has been attributed to the erosion cycles that removed the Paleozoic strata from the top of the range. Thwaites (1958, 1960) discards those hypotheses in the same way as he discards that for the Precambrian.

Between the time of deposition of the post-Cambrian strata and their removal in post-Paleozoic time and then continuing to the Pleistocene, or Great Ice Age, geologic events in Devils Lake Park are obscured. The latter part of that interval, encompassing at least 300 million years, must have been largely a time for erosion as no rocks are left behind. If the interpretation is correct that the upland surface of quartzite is only the recently exhumed Precambrian surface protected during much of the time by a cover of Paleozoic sediments, then the topography of Devils Lake Park has changed considerably during the last 550 million years even though present day topography in the park may be essentially the same as it was 550 million years B.P. The remaining Cambrian sandstone in the present park does not make striking features as it does farther west, especially in Baxter Hollow, even though Parfreys Glen in the northeast corner of the recommended area is reasonably typical.

Although it seems clear that at least part of the Devils Lake gorge was cut in Precambrian times by an ancient stream, otherwise we would not see the Cambrian sandstone infilling it, perhaps not all was. Some writers attribute the north part of the gorge to the Paleozoic cycles of erosion (Thwaites 1958), and I do not believe that an early Pleistocene time for cutting part of the gorge can yet be ruled out.

Potholes on the East Bluff (Figs. 65, 66) are attributed by different people to the stream work associated with the cutting of the gorge during the Precambrian, the Paleozoic, the Cretaceous, or the Tertiary, yet they too may only be Pleistocene (Black 1964c). However, at the east end of the Baraboo Range one pothole in a group of about 25 in the quartzite has Cambrian sandstone firmly adhering to the inside walls so it was cut indisputedly in late Precambrian or Early Cambrian time. Different kinds of potholes are present at that site, and all may not be of the same age nor are they necessarily the same age as those at Devils Lake.

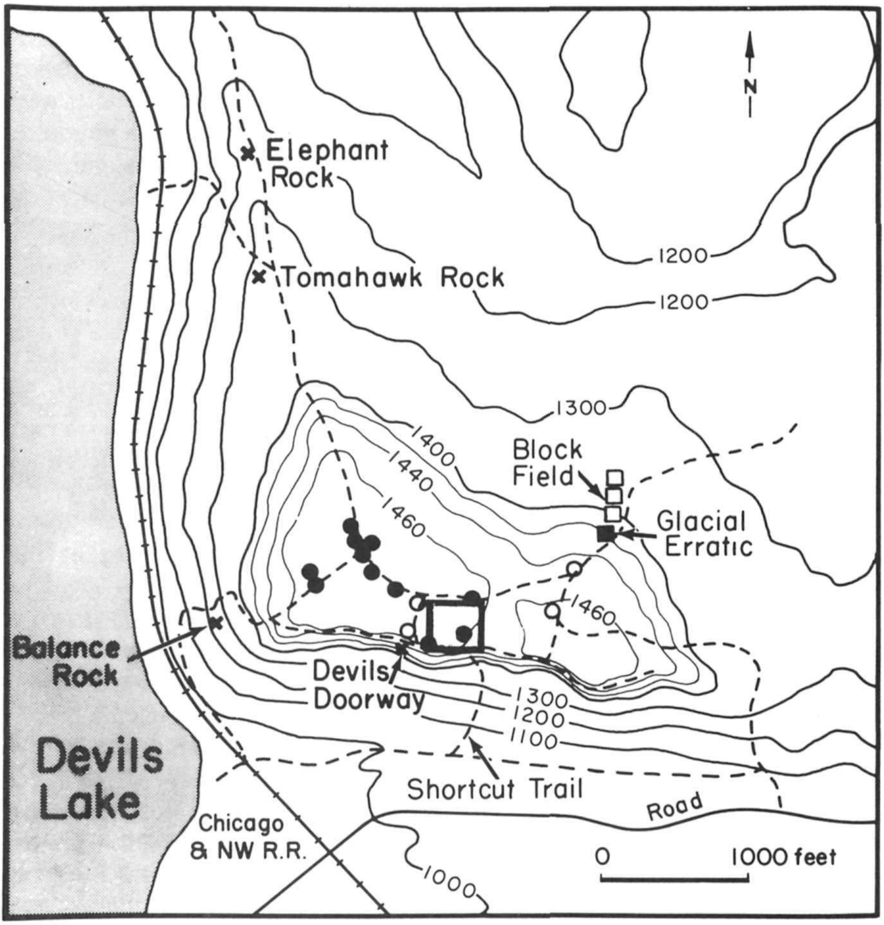

Fig. 65. Topographic map of East Bluff area of Devils Lake, showing location of various named features and area covered in Fig. 66. Solid black circles show locations of drill holes where the Windrow Formation was encountered below a broken rubble of Baraboo quartzite; open circles are locations of drill holes where the Windrow Formation was not encountered to bedrock.

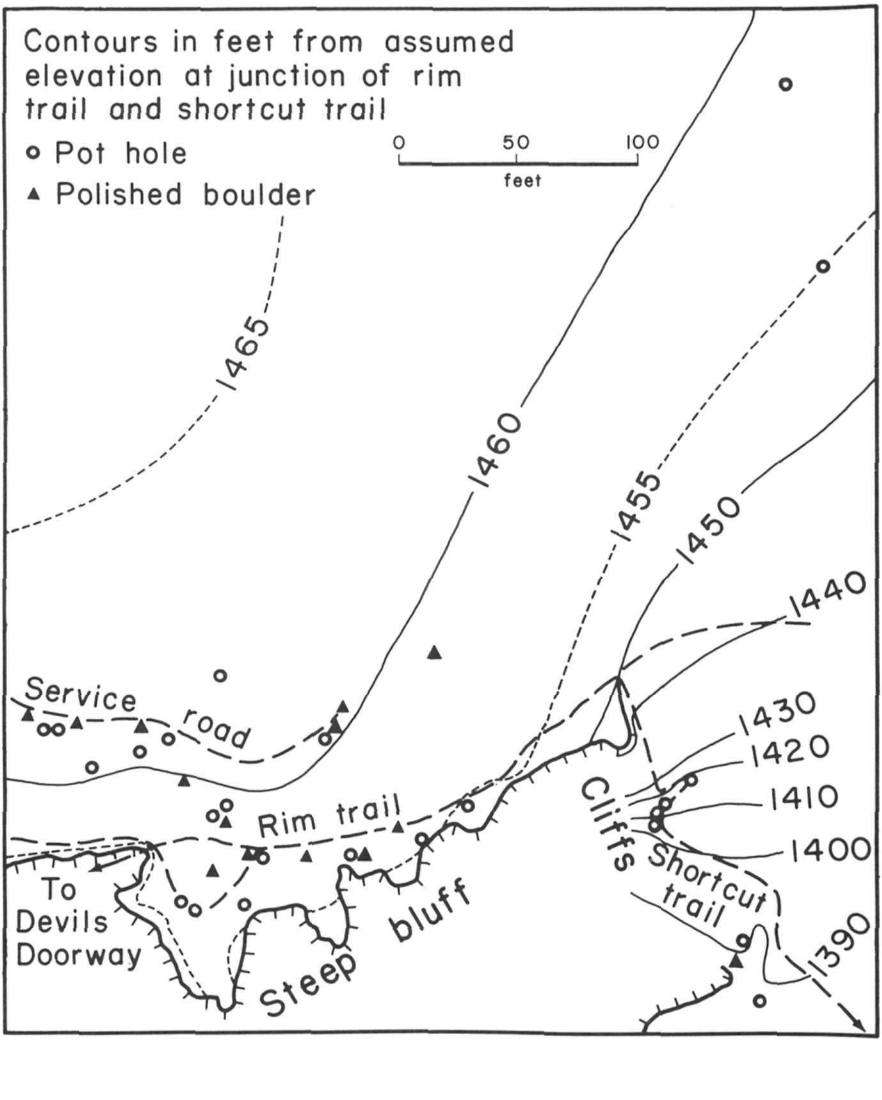

Fig. 66. Topographic map located by rectangle on Fig. 65, showing locations of potholes and water-polished boulders on East Bluff of Devils Lake.

Several are altered by glacial ice of Cary age, but if all were, the evidence is obscured by post-glacial weathering. Potholes in bedrock are thought to have formed under glacial ice in Norway (Gjessing 1965-66).

The pebbly loam with much expandable clay on top of East Bluff must be the source for the Windrow gravel (Fig. 65) (Andrews 1958) which is considered Cretaceous by him, but again an early Pleistocene deposit cannot yet be ruled out (Black 1964c). The gravel has been found in and around the potholes. No way has yet been found to date the deposits or cutting of potholes satisfactorily. Their place in the history of events must await new evidence. Regardless of their age, however, loose blocks with potholes have been moved about on the upland, and angular quartzite blocks lie on top of the pebbly clay. Glacial ice must have accomplished this, for blocks up to 85 tons seem to have moved upslope (Black 1964c). The area lies immediately west of the prominent Cary end moraine on the upland. This is correlative with the moraines that plug the southeast and north parts of the Devils Lake gorge (Fig. 59). These features are perhaps only 13,000-16,000 years old (Black et al. 1965:56-81). They do not prove that earlier ice went no farther into the Driftless area, and much evidence has now been amassed to indicate ice did go further west (Black 1960; Black et al. 1965:56-81; Frye et al. 1965:43-61).

Much of the talus and the pinnacled cliffs around Devils Lake (Fig. 61) are associated with the late Wisconsinan Stage of glaciation (Smith 1949; Black 1964c; Black et al. 1965:56-81). Whether the area was glaciated more than once is not proved but is suggested (Weidman 1904; Alden 1918:177-178; Thwaites 1958; Black 1964c; Black et al. 1965:56-81). For example, organic matter from a depth of 130 ft in glacial deposits at Baraboo was submitted by Thwaites to Wilson (1936:43) for identification; he found leaves of several dicotyledonous plants, some probably *Vaccinium*, and one species of moss, identified by Cheney as *Campylium stellatum*. Thus the story of the geomorphic development of Devils Lake Park jumps quickly from the Paleozoic to the Pleistocene or even late Pleistocene.

Since glaciation, gravity and frost action have moved many large blocks of quartzite down slope although the present rate is very slow. Railroad and other construction and abortive attempts at farming in the last century have left their mark. Some pits for aggregates have been opened in glacial materials and in bedrock. Increasing pressure from tourists and students is showing, but relatively little gross modification of surface features has occurred since the last glacial ice disappeared from the area.

Description of Specific Features

Moraines

The most important glacial feature in Devils Lake Park is the end moraine of Cary age (late Woodfordian) depicted in Fig. 59. This end moraine can be traced with only minor breaks from the southwest boundary of the recommended area in an irregular looping course through the park to the northwest corner. It is part of the Johnstown Moraine traceable along the entire front of the Green Bay Lobe (Alden 1918). The detailed description of the moraine in the park, initially given by Salisbury and Atwood (1900:93, 94, 105-111), has stood the test of time. Because of its length, it will not be quoted here. The moraine marks the still stand of the outer edge of the ice sheet in part synchronous with the more massive moraines and outwash plugging the valley and enclosing Devils Lake.

Terminal moraine plugs such as occupy the gorge north and southeast of Devils Lake are unusual individually but together comprise a unique situation. Having such a prominent well-defined end moraine extending for so many miles from those plugs makes the situation even more astounding. The moraine outlined in Fig. 59, in the recommended area, if not the best, is certainly one of the few best such features to be found anywhere in the world. Having it so readily accessible to centers of population with so many other features nearby makes it especially attractive. As recognized by Salisbury and Atwood (1900), the striking loops show clearly the inability of the ice to surmount topographic obstacles of negligible relief because of restricted flow over and around buttresses up ice. Nowhere are similar features so well displayed amongst so many other phenomena of intriguing historical connotation.

The uniformity of height (15-50 ft) and width (100-200 ft) of the moraine on flat surfaces and the asymmetry of the moraine on hill sides (only 10- to 15-ft abrupt faces on the uphill side and 50- to 100-ft faces on the downslope side) are in themselves very unusual over such broad distances. Furthermore the position of the moraine from its high point on Devils Nose southwestward to the level of the plain records precisely the distal slope of the ice front during at least the latter part of the deposition of the moraine. Recording of such gradients is a rare occurrence almost anywhere because of concurrent or post-glacial destruction by flowing water and mass movements. Thus, in the recommended area of Devils Lake Park are textbook examples of glacial features hardly rivaled anywhere else in the world.

The moraine in the recommended area is but a small part, but unquestionably the best part, of the end moraine of essentially similar age that has been traced throughout Wisconsin, from the Minnesota border near Hudson to the Illinois border south of Lake Geneva, and also from the Great Plains to the Atlantic Ocean (Chamberlin 1883a). This moraine was designated as the Terminal Moraine of the Second Glacial Epoch (Chamberlin 1878, 1883a). It was considered by him to be the boundary between older and younger drift and as such to be the most important time break in the Pleistocene in the state. Recent field work has not supported this viewpoint (Frye et al. 1965:43-61). Unfortunately we do not have a single date recording the advance of ice to this end moraine in Wisconsin, but from evidence from Minnesota and Illinois, it probably was formed 13,000-16,000 years B.P. In many places outside of the recommended park area, the moraine appears more massive than it is within the park. Yet its massiveness commonly may be attributed to bedrock elevations on which it is found or to the overriding and pushing up of material from below (Alden 1918).

In the park the end moraine outside of the two plugs containing Devils Lake is generally only 15-20 ft high. Locally, as at the easternmost loop at Sauk Point on the very crest of the Baraboo Range, near the WWCF radio tower, the front is fully 80 ft high. It is accentuated there because of the high level of the Baraboo quartzite on which it is built and the low plain stretching to the west which was occupied by outwash and a former glacial lake (Ott Lake, Fig. 59). The more massive moraines containing Devils Lake rise 90-130 ft above Devils Lake and even higher above the valley floors north and east. Their massiveness is due mostly to outwash and deltaic deposits (Thwaites 1958:150) deposited in front of the advancing ice. Only a thin local veneer of till was deposited directly on these deposits by the ice. The deepest well in the gorge, 383 ft, did not reach bedrock.

From a car, views of the end moraine are particularly good along County Highway DL northeast of Devils Lake (Fig. 67) and at the extreme northwest corner of the park where Highway 159 crosses the moraine (Fig. 68). There especially, with the abrupt steep slope to the northeast formerly occupied by ice and the small moraine ridge with its smooth outwash plain to the west or front of the moraine, we have a classic example of the relationship of the ice sheet to its proglacial fluvial deposits. At the easternmost loop, by Sauk Point, the moraine and its relationship to the quartzite and proglacial lakes are accessible and readily discernible (Figs. 69-72). Stratification and texture of outwash, unassorted sandy till, and shear planes inclined steeply up ice are especially well displayed in the gravel pit shown in Fig. 71. Immediately below the pit is Ott Lake basin, a former proglacial lake, and weathered outcrops of the Baraboo

Fig. 67. The end moraine of Cary age is the topographic rise with trees forming the skyline. View northeastward from County Highway DL, about 1 mile east of Devils Lake.

Fig. 68. Outwash plain in the foreground from the Cary front in background as viewed northeastward from State Highway 159, about 0.5 mile east of State Highway 12.

Fig. 69. Moraine, outwash, and Ott Lake basin looking northwestward from the gravel pit at the eastern loop at Sauk Point. The north loop of the Cary moraine forms the skyline in the center of the photograph.

Fig. 70. View westward of Ott Lake basin from the gravel pit at the eastern loop of the end moraine at Sauk Point.

Fig. 71. Gravel pit at the eastern loop of the end moraine at Sauk Point, looking northeastward. Stratified outwash dips gently to the left; till and drift partly bedded are inclined steeply to the right, reflecting ice push and possible shear planes as the ice attempted to override its moraine. Copper nuggets from Keewenaw Peninsula of Upper Michigan, Lake Superior sandstone, and many igneous and metamorphic rocks reflect the northern source areas.

Fig. 72. Retreatal moraines of Cary age, looking southward from the gravel pit in the outermost moraine in the eastern loop at Sauk Point.

quartzite. Retreatal moraines are also common behind the outer moraine south of the gravel pit (Fig. 72). The abrupt interlobate junction of ice from the north and south sides of the south flank of the Baraboo Range is clearly portrayed in the moraines northeast of the pit.

For hikers the views of the moraine are particularly good near the Devils Nose on the South Bluff southeast of Devils Lake (Fig. 73), on the southern part of East Bluff (Fig. 74), extending eastward to the extreme east loop of the moraine at Sauk Point, and also at the north tip of the north loop (Figs. 75, 76). There is one of the most striking views to be had anywhere of the Baraboo Valley, the city of Baraboo, and the Lower Narrows gap of the Baraboo River through the North Range. On the south rim of the East Bluff views to the Wisconsin River Valley from the moraine above the quarry and farther east to the vicinity of Parfreys Glen are superb (Fig. 77). Excellent views of the plugs containing Devils Lake may be had from all of the bluffs rising above them.

Concentric moraines arc around the extreme north end of the north loop of the end moraine (Fig. 75), in sec. 9, north of Hanson Marsh. These show beautifully the building of ridges at the edge of the ice as it struggled to maintain its position around that high point. Probably at the initial advance the ice went over the inside of the loop, for erratics are to be found on it (Fig. 76).

Fig. 73. Outside or front of the end moraine on the north side of Devils Nose looking northwestward.

Fig. 74. Top of the end moraine on the East Bluff of Devils Lake, about 0.25 mile southwest of State Highway 113, looking northeastward.

Fig. 75. Apex of the end moraine in the north loop, showing the inside or supposedly unglaciated area on the right and a retreatal moraine on the left. View eastward.

Fig. 76. View southwestward from the apex of the inside of the north loop of the end moraine of Cary age. The line of trees is on the end moraine.

However, their presence can also be attributed to water transportation and even gravity movement from the steep face of the ice that must have developed there. As the terrain inside the loop is precipitous (Fig. 76), boulders could have bounced and rolled practically across the loop on a vegetation-free surface or on an ice-covered surface. At any rate the successive arcs are each slightly lower than their predecessor. The first two are separated by a gap only 60-100 ft across and 10-20 ft deep. The later ones are lower and less regular. The features at the nose of the arc are among the best developed anywhere. When coupled with the beautiful views of the Baraboo Valley to north and west and of the drained lakes and other features to the south, this can be considered truly one of the grand overlooks of the Reserve.

For additional details on the moraine, the reader is referred to the original works of Salisbury and Atwood (1900), Trowbridge (1917), and Alden (1918). All emphasize its uniqueness.

Fig. 77. The Wisconsin River Valley, looking southwestward past Devils Nose on the right.

Drained lakes

Proglacial lakes were formed immediately in front of the moraine in several places in the recommended park area (Fig. 59). All of these former lakes have been drained, but the sediments remain behind as mute testimony of their former existence. One unnamed lake formerly existed 1.3 miles southeast of Devils Lake on the north side of Devils Nose. Cary ice butted against the ridge leaving its end moraine which may be traced around and across the nose (Fig. 73). Where the moraine crosses the gully in the east half of sec. 30, it is a symmetrical ridge about 45 ft high and 100 ft wide, breached at the gully with a smooth plain to the southwest. That plain is underlain with 10-30 ft of silty sands and some clay and gravel. From the moraine down the gully to the north one sees numerous very large foreign erratics, but from the moraine up the gully to the west and southwest only the Baraboo quartzite blocks and small amounts of fine pea-sized foreign gravel are seen.

Similar but larger lakes were common in secs. 16, 17, and 18 in the northeastern part of the recommended area (Fig. 59). Peck and Steinke (Fig. 78) glacial lakes were named early (Salisbury and Atwood 1900; Trowbridge 1917). The name Glacial Ott Lake is assigned for convenience to the easternmost and smallest basin in the Sauk Point Loop. At one time those basins probably were merged into one lake which would have drained into Glacial Devils Lake. As the ice border retreated somewhat from the end moraine shown in Fig. 59, the lakes would have become separated from each other. Ott Lake in the southeastern part of sec. 16 was the first to be drained or to fill with outwash. Peck and Steinke lakes, farther west at lower levels, remained longer. Just how long they were able to survive is not known. However, in secs. 9 and 16, in the northeastern part of the recommended area, is the low swampy area known today as Hanson Marsh. It was a lake that survived for many centuries (Bachhuber 1966). Ice at its furthest extent at the position shown in Fig. 59, covered the area of the marsh but withdrew almost immediately thereafter to build an end moraine on the ridge to the west and another to the north, surrounding Hanson Marsh and forming a lake. While surrounding it, rhythmically-banded lacustrine sediments at least 25 ft thick were laid down in the lake along the ice margin. Bachhuber (1966) has counted representative samples of the supposed varves which represent at least 600 years of time. These are only part of that lacustrine sequence.

While the ice stood around the old lake, spruce forests to the west of the area were contributing pollen to the lake sediments. The pollen sequence throughout the deposits shows clearly the post-glacial climatic changes as reflected in the

Fig. 78. Steinke Lake basin, looking southward to the end moraine where it crosses State Highway 113.

local vegetation. In brief they are mainly a transition from spruce to pine to mixed hardwoods and other deciduous trees (Bachhuber 1966).

At its maximum, expanded Devils Lake may have reached an elevation of 1155 ft, enough to drain the lake to the northwest down Skillet Creek (Trowbridge 1917). (Elevations cited by Trowbridge differ from those cited here because of availability of more accurate maps today). Thwaites (1960) does not accept any available evidence that the lake actually overflowed through Skillet Creek even though Trowbridge (1917) found erratics in Messenger Creek on the lake side within 16 ft of the Skillet Creek divide (Fig. 79). The edge of the Cary moraine at the

Fig. 79. View westward of Messenger Creek valley, across Devils Lake, with the divide into Skillet Creek drainage at the cleared field on the upland surface in the right of the photograph.

north end of West Bluff has left its mark up to 1060 ft in elevation only. How much higher the late Woodfordian ice went is not known although I have found large igneous and dolomite boulders up to 1160 ft on the northwest side of that nose (SW¼ SE¼ NE¼ sec. 14, T. 11 N., R. 6 E.). Thus, I would agree with Trowbridge (1917) that Devils Lake overflowed into Skillet Creek. Furthermore, the cutting of lower Skillet Creek valley in quartzite must have been accomplished by far more discharge than is available from the present drainage area. In the lower part of the gorge, removal of the large blocks of quartzite from the gorge proper required torrential discharges. The monuments and jagged spurs now present in the gorge reflect post-glacial frost processes as do the bluffs around Devils Lake itself.

At its maximum advance, over 11 miles of the Cary ice front were contributing water to the lakes in the Devils Lake area (Trowbridge 1917:364). His argument is that ice was brought to the terminus at a rate of 6 inches a week or 26 ft a year. Assuming ice was melted along that entire front in a zone 100 ft wide by 26 ft deep we get a measure of the minimum amount of water that could reach the lakes. Surely melt waters farther back from the front would also in part reach the area. Trowbridge concludes that the annual discharge to the lakes would be at least on the order of 1.5 billion ft^3. The Devils Lake basin itself has a capacity, to the discharge level at Skillet Creek, of about 7.5 billion ft^3. Thus the main lake basin should have been filled to overflowing in only 5 years. The upper lakes would have held a relatively small amount before drainining into Devils Lake. With the ice standing in the area more than 600 years, surely Devils Lake overflowed for considerable time through Skillet Creek and later possibly past the north end of West Bluff, at the margin of the ice, even though no features or deposits there prove this unequivocally.

Trowbridge has supported these rough estimates of water volume by a check on the amount of material deposited from the glacial waters. Trowbridge calculated that 6 miles of the ice front drained into Steinke Lake, depositing over 2.5 billion ft^3 of debris. In Peck Basin its 0.5 mile of ice front contributed at least 142 million ft^3 of debris. The Devils Lake gap between the two morainal dams contains over 2 billion ft^3 of debris. Thus, it would seem clear that these lakes must have had more than enough water to drain through the headwaters of Skillet Creek, the lowest divide available if the late Woodfordian ice stood higher than 1155 ft against the north end of West Bluff. Unfortunately dating of the large foreign boulders buried in the soil up to at least 1160 ft elevation of the northwest side of that nose is not yet possible. Dolomite erratics are very little etched; gabbro and other coarse-textured boulders are not desegregated so it is assumed they were left by ice immediately preceding the building of the prominent end moraine.

A large fresh gravel kame (SE¼ SE¼ sec. 25, T. 12 N., R. 5 E.) 3.5 miles west of the same front at Baraboo and a deep kettle hole just off the southwest corner of Fig. 59 also attest to advance of the ice beyond the prominent end moraine.

In the Steinke Lake sediments Salisbury and Atwood (1900:120, Pl. 28, p. 108) noticed laminated silts and clays in which marked deformation of certain horizons were present. Locally more than 60 ft of these deposits were excavated. Salisbury and Atwood (1900:134) outline the history of that lake briefly. Because the basin is enclosed to the south, east, and west by quartzite, the basin was in a logical position to receive and hold water. The first lake thus formed against the ice to the north had no outlet and the water rose to the level of the lowest divide on the southwest side of Peck Lake where it overflowed to the west and northwest to Devils Lake via Hells Canyon. Sediments borne into the basin by the glacial drainage were deposited as deltas and outwash in the lake. The coarser particles were left near the ice; the finer ones were carried farther away. Continued melt water from the ice front brought more and more sediment into the lake until its delta front extended completely across the lake, filling it to the level of the outlet. Later drainage followed the retreating edge of the ice westward into Devils Lake.

Accompanying the lakes and following their destruction were other drainage modifications in the area. An example is that of Skillet Creek, the small tributary to the Baraboo River which flows northwesterly from the southwest divide of Devils Lake. Before glaciation of the district, Skillet Creek probably flowed in a general northeasterly direction to the Baraboo River (Salisbury and Atwood 1900:138). The Woodfordian ice blocked the stream forcing it to seek a new course. The only course open was to the north and west in front of the advancing ice. Drainage from the ice, depositing glacial fluvial and then glacial lacustrine materials, forced the stream farther westward, until finally it reached its position across the sandstone plain well to the north and west of its former route. In that position ancestral Skillet Creek began to downcut after deglaciation and drainage of Glacial Lake Baraboo that inundated the lowland up to 980 ft elevation where Baraboo is now located. The creek superimposed itself on the bedrock in a new gorge. Such superposition would have occurred after the cessation of overflow of water into Skillet Creek from the glacial lake occupying the gorge of Devils Lake. To drain Lake Baraboo it was necessary to clear the ice from the east nose of the Baraboo Range proper, near Portage (Bretz 1950). The position of the lower part of Skillet Creek well on the westward flank of the outwash apron of Cary age can be attributed to the initial topographic elevation left behind during the draining of Lake Baraboo.

Stagnant ice features

The retreat of the ice from Sauk Point at the crest of the Baraboo Range was by melting *in situ*, for it left behind typical ice-stagnation features with knob and swale topography (Fig. 80). Many knobs are small kames of poorly sorted but water worked sand and gravel; the depressions are almost invariably kettles produced by the melting out of buried ice blocks in the debris. The stagnant buried ice area formed at the junction of the advancing lobe from the north and another from the south, a kettle interlobate moraine of very small size when compared to the Kettle Interlobate Moraine of eastern Wisconsin. Yet its origin would have been similar. Section 15 contains the better features of this ice stagnation interlobate area. Relief is generally only 10-30 ft between the knobs and adjacent kettles. They are readily viewed from the east-west highway extension of County Highway DL.

Behind the end moraine as mapped through the recommended area (Fig. 59), numerous ice-stagnation features may be seen. These are particularly well-developed on the flanks of the Baraboo Range to the north toward the city of Baraboo and also to the south and east toward the Wisconsin River. Many knobs are kames in the sense that they were ice contact, water worked debris. Many of the swales are true kettles. Such ice-stagnation features of the steeper slopes of the Baraboo Range are generally nowhere as well developed or as large as those of the lowlands. It is in the lowlands that the larger ice blocks were buried more readily.

Potholes

Black (1964c) has described potholes on the East Bluff of Devils Lake overlooking the late Woodfordian moraine which plugs the southeast gorge (Figs. 65, 66). The potholes are carved in bedding plane surfaces of the Baraboo quartzite *in situ* and also in loose blocks of the quartzite that are scattered irregularly on the beveled upland surface. Polished chert-rich gravel of the Windrow Formation is associated with some potholes and has been found in them (Salisbury 1895:657). More than a dozen well-developed single potholes are known. They range from circular, polished depressions a few inches in diameter and only 1 or 2 inches deep to those more than a foot in diameter and twice as deep. Some are so beautifully symmetrical that they resemble artificially drilled holes with perfectly parallel smooth sides. Potholes above the sod are concentrated in an area 50 yards

Fig. 80. Knob and swale topography in the east-central part of sec. 15, about 0.5 mile east of Sauk Point.

along the bluff and 30 yards northwest from the rim. They also occur in a narrow cascade zone for 75 ft vertically below the rim, along the Short Cut Trail. Others are scattered in the woods to the north of the Short Cut Trail. Buried potholes may be more widespread (Salisbury 1895:655).

Through the years most writers have attributed the potholes and associated gravels to preglacial streams of Cretaceous or Tertiary age that flowed across a continuous upland surface at and above the level of the rim. No one seriously had considered them to be glacial yet Black (1964c) has suggested such an origin seems at least as plausible as others.

However, we know now that at least one of the potholes at the extreme east nose of the Baraboo Range, which contains Cambrian sandstone firmly adhering to its walls, must have been produced by water in late Precambrian or earliest Cambrian times. Not all of the other potholes of that locality can be ascribed necessarily to the same time of formation though several clearly were modified by the Cary ice that overrode them.

By analogy it would be logical to suspect that the potholes on the East Bluff of Devils Lake were produced at the same time, but this does not prove it. Regardless of when they might have been formed, it is clear that the loose blocks in which we do find potholes have been moved subsequent to the pothole drilling. Some blocks have been split and one side or bottom of its pothole is now gone. Others are turned on their sides or are upside down. These are scattered along with other loose blocks of the Baraboo quartzite over the pebble-rich, clay nearby. I have confirmed this relationship by drilling 12 holes through the quartzite rubble and into the clay below.

The splitting of the blocks and movement of the loose blocks to their present location is most easily explained on the basis of movement by glacial ice or possibly in part by melt waters associated with ice. The hundreds of blocks of Baraboo quartzite on top of the Windrow Formation cannot be explained by simply weathering down in place, as no quartzite nearby is higher. Such blocks of the quartzite on top of the Windrow must be considered true glacial erratics. The large erratic to the north of the pothole area (Fig. 66) is so described by Black (1964c). It weighs 85 tons and must have been moved upslope to its present resting place. This surely could only be accomplished by ice. Other smaller but impressive quartzite erratics may be seen on the South and West bluffs of Devils Lake as well (Fig. 81). No mechanism of erosion of the smoothly beveled upland surface that I know of can leave behind such large loose blocks to rise above the general level.

These various phenomena would imply that glaciation of the Devils

Fig. 81. "Erratic" of Baraboo quartzite on the highest part of the South Bluff of Devils Lake.

Lake area had occurred some time prior to the Cary advance. This I am certain has occurred from a variety of evidence that cannot be detailed here. Dolomite and igneous rock erratics are found well above the Cary moraine at the north end of West Bluff. Moreover, a kame deposit 3.5 miles west of the front near Baraboo has 40,000 yard^3 of gravel, and a deep kettle with till lies 1 mile west of the front at the Badger Ordinance Works south of the park. They attest to extensions of glacier ice beyond the end moraine of the Cary as recorded in Fig. 59. The freshness of dolomite and igneous erratics, the lack of erosion and filling of the kettle, the freshness of igneous outcrops to the west of Devils Lake, the youthfulness of the loess, and other criteria would suggest that the time of such glaciation did not long precede that of the Cary. In the Driftless Area, farther west of those phenomena, thin loess deposits in an area of a few square miles have buried soil profiles indicative of Altonian and Sangamonian age. Dating of the various deposits is a very perplexing problem, for which, unfortunately we have relatively little information and it cannot be discussed further here.

Erratics

For convenience, erratics within the recommended area of Devils Lake Park may be classified into two groups. One contains those rocks, such as igneous and highly metamorphosed materials, that could have originated only from a point far to the north, and the other those rocks of local derivation which are in anomalous situations. This section is concerned largely with the second group. The large mass of debris brought in by the Woodfordian ice and dumped inside the end moraine is clearly of glacial erratic origin. Within Devils Lake gap erratics have been washed out from the terminal areas of the ice that blocked the north and southeast ends. Erratics have been carried by drifting ice at least 90 ft above the present lake level (Salisbury and Atwood 1900:133), and Trowbridge (1917:366) in 1 hour found 103 erratic boulders in the valley of the north fork of Messenger Creek and one diabase cobble on the west slope of the divide in the drainage of Skillet Creek. He found igneous rock erratics 164 ft (202 ft in his paper reflects use of now outdated topographic maps) above the present level of Devils Lake and only 28 ft below the divide. Other glacial cobbles occurred within 16 vertical ft of the divide. Thus the origin of erratics behind the end moraine and those carried out from the terminus by outwash waters and floating icebergs in the proglacial lakes are readily explained. These are recognized easily because of their obvious foreign source.

In the second group of rocks, however, we find various local materials which are distributed in the area in such a way that it is far more difficult to prove that they are true erratics and even more difficult to prove that they obtained their present locations on the basis of glacial ice directly. In this group are placed the large Baraboo quartzite erratic blocks and fragments which occur on East Bluff on top of the Windrow Formation and also those which occur on the South and West bluffs on the Baraboo quartzite itself. To this group is added the Paleozoic cherts which lie outside the end moraine. These two categories require additional comment.

It is difficult not to accept as glacial erratics the angular rubble on top of the Windrow Formation on East Bluff. If one accepts the 85-ton Baraboo quartzite block near the block fields north of the pothole area (Fig. 66) on the East Bluff as a glacial erratic, then it would seem to me that we must also accept similar large angular blocks of the Baraboo quartzite on the South (Fig. 81) and West bluffs as well. On the rounded, relatively smooth upland surfaces of West Bluff, South Bluff, and also Sauk Hill to the west, isolated blocks of the Baraboo quartzite protrude through the loess cap which is a few inches to 2 or 3 ft thick locally. These blocks are loose and rest directly on the quartzite. Hence, they have not attracted the attention of previous workers in the area. However, no process of planation by sea or streams could leave such large angular fresh blocks behind to rise above the smoothly plained surfaces that are supposed to be exhumed from beneath hundreds of feet of Paleozoic sandstones and dolomites. These are the highest surfaces in the area. Obviously the material is related to that underlying it, and it cannot have been let down from a higher cover. How then can a dipping Baraboo quartzite be smoothly truncated with such loose blocks left behind to rise above the general surface by many feet? To me, it is far easier to explain such loose blocks as being brought in some time after the exhumation of the upper surfaces. The logical time is during the Pleistocene, by glacial action. Many blocks are angular with very sharp corners; relatively little pitting has taken place, and frost riving is minimal. No great antiquity can be given such indicators. A late Wisconsinan age for them would seem most logical, yet an earlier Pleistocene age is possible.

Associated with the erratic blocks of Baraboo quartzite on the South Bluff are distinct channels in the upland surface which are peculiar. One due south of the lake crosses through the crest of the range and has steep overhanging banks 10-15 ft high (Fig. 82). Corners of the blocks are very sharp. A few blocks, presumably derived by frost action, lie at the foot of the bank but hundreds of cubic yards of material have been removed from the largest channel. No accumulation of such debris is seen either to the north or to the south. Where has it gone? Are

Fig. 82. Overhanging bank of Baraboo quartzite on top of the South Bluff of Devils Lake, looking northward.

such features related to the Paleozoic or Mesozoic erosion cycles that have affected the area, or is this again something that may be attributable to a pre-Cary glacial event? We have really no basis for saying one way or the other except for the relative freshness of the edges and faces of the Baraboo quartzite exposed in these peculiar features. We have inherited at least one late Precambrian or Early Cambrian pothole, but it is a very small feature obviously protected by the Cambrian sandstone. No sandstone was seen anywhere in association with the loose angular blocks of the Baraboo quartzite on the upland or with the sharp channels. Hopefully, more detailed field work will provide additional clues to the perplexing origin of these features.

The chert erratics present another puzzling situation. Chert behind the end moraine of Cary age clearly can be explained on the basis of its having been brought in by ice. However, it has been customary to explain chert that is locally identifiable as Ordovician-Silurian in age to the west of the Cary terminus as having been let down during the weathering and removal of the Paleozoic formations that once overlay the Baraboo quartzite (Thwaites 1958, 1960). The abundance of such chert of Silurian age is puzzling. One would expect that the younger formations which would be removed first in the Paleozoic-Mesozoic-Tertiary weathering cycles would be essentially absent from the upland in comparison with chert of the older formations, all things but time being equal. No detailed studies have been attempted, yet we find considerable Silurian chert. This seems incongruous because there is no difference in size or weatherability. Is it possible that the chert has not been let down but has actually been brought in by ice of an earlier glaciation that did not have abundant igneous materials in it? Again we have no basis for discussion of such a problem, because the evidence is still too meager to constrain our thinking.

Periglacial features

Within the Baraboo area Smith (1949) lists three groups of features attributable to periglacial processes: 1) stabilized talus, 2) block concentrations and block-strewn slopes, and 3) choked valleys and block cascades. Within the recommended park area Smith (1949) discussed the talus deposits in the vicinity of Devils Lake and block concentrations and talus slopes northwest of Devils Lake and also on the south flank of the Baraboo Range south of the lake. To this list of features should be added the pinnacles and monuments on the cliffs of Devils Lake and wind-polished surfaces.

The talus accumulations around Devils Lake are among the most striking features of the park (Figs. 61, 62). They are better displayed there than anywhere else in the Baraboo Range. Other locations are in the gorge north of North Freedom (formerly Ableman's) (Salisbury and Atwood 1900:67), and also along the bluffs of the Lower Narrows northeast of Baraboo. Talus is best developed on the East, West, and South bluffs of the lake. Where the Cary ice stood in the southeast gap it presumably removed much of the talus that apparently was there before. On the bluffs above the lake the talus is almost continuous laterally, being interrupted locally by dipping ledges of the quartzite. It is partly covered by irregular discontinuous forest. The talus on the south-facing slope of East Bluff attains maximum height and continuity of exposure. On the north-facing slope of the South Bluff the talus is covered largely by trees, and the slope is slightly less steep.

The talus is composed of heterogenous, angular, irregular blocks of quartzite more or less firmly wedged together. The blocks commonly are more than 6 ft on a side. No marked vertical zoning of large blocks is apparent. Occasional igneous erratic boulders may be found in the talus up to 90 ft above the lake level (Salisbury and Atwood 1900:133). Especially around the trails built during the 1930s by the Civilian Conservation Corps, foreign material was brought in for surfacing. Erratic blocks of foreign debris should be found in talus up to the level of the divide between Messenger and Skillet creeks, if the interpretation of that divide as the outlet of Glacial Devils Lake is correct (Trowbridge 1917). The maximum height of talus is about 300 ft; the maximum inclination of the slope is about 36°. The hydrographic map of Devils Lake (Juday 1914, Map 8) suggests that the talus extends 30 ft out from shore below water level. According to data of Thwaites, the talus may extend to depths of as much as 285 ft below lake level.

Many of the talus blocks as well as the rock surfaces and ledges above them are partly covered with lichens and show some weathering stains. No clear indications of movement are available. The vegetation seems stabilized on the slopes. Few blocks are seen on snow surfaces in winter, and isolated loose blocks out from the foot of the bluffs in the forests are also relatively uncommon. The frost-rived bluffs and ledges above the talus show many loose blocks and pinnacles (Figs. 63,64) apparently in unstable situations, yet few seem to collapse. The angularity and weight of the blocks permit them to stand in relatively permanent features. Other signs of inactivity recorded by Smith (1949) indicate that the formation of talus blocks now is an exceedingly slow process.

How much of the talus originated prior to the advance of the Woodfordian ice into the north and southeast gorges is not known. If the talus does extend many tens of feet below the surface of the lake, it seems likely that it has been covered by outwash from the ice fronts. In the abandoned quarry on the northeast side of Devils Lake a thin veneer of talus is separated from the bedrock by about 5 ft of stony soil containing small blocks and rock fragments scattered through an earthy matrix (Smith 1949:202). The contrast between the talus and underlying material is striking and points, according to Smith, to a marked change in conditions of weathering when talus accumulation began.

Smith (1949) did not discuss the effects that the high lake level might have had on the formation of talus in the gorge. If Trowbridge (1917) is correct in that Devils Lake was up to the level of the divide between Messenger and Skillet creeks, then the bulk of the talus in the area would have been covered by the glacial lake waters, and the lake level would have been near the base of the present cliff in many places. Would frost action which is commonly more severe at the water level of a lake have been instrumental in the formation of some of the talus? This is a question that we cannot yet answer. However, the lack of erratics of obvious foreign sources among the talus blocks where they surely were covered by lake waters is difficult to explain unless the talus has come down on top of such material to hide it. The small particles could have been flushed through the coarse openings of the large talus blocks. Pinnacled slopes and jagged angular blocks are almost as common along the Baraboo bluffs to the east where the Cary ice definitely overrode them and also in the lower narrows to the northeast of Baraboo.

In the St. Croix Dalles area Cary ice clearly went through the gorge, and the pinnacled cliffs have developed subsequently. Time of formation can be brief. No unequivocable facts are available for dating the talus and pinnacles in Devils Lake Park. Some of the material may have been derived in pre-Cary times, some of the monuments such as Devils Doorway (Fig. 63), Elephant Rock and Balanced Rock (Fig. 64) possibly were produced after Cary glaciation.

The narrow depressions (Fig. 83) along the base of many talus slopes are peculiar. They are elongate, discontinuous and channel-like—15-25 ft wide and 5-15 ft deep. Thwaites (1935:395-404) attributed them to settling of the finer sediment into interstices of the talus, but their origin is conjectural.

The smaller block concentrations and block strewn slopes on the south-facing flank of the Baraboo Range south of Devils Lake are quite similar to those of the talus slopes of Devils Lake except for their elongated dimension. Locally many are covered with forests and interstices of the large blocks are filled with soil.

The locality less than a mile northwest of Devils Lake (NE¼ sec. 14) (Smith 1949:204) presents a problem. Smith records shattered blocks and boulders of quartzite,

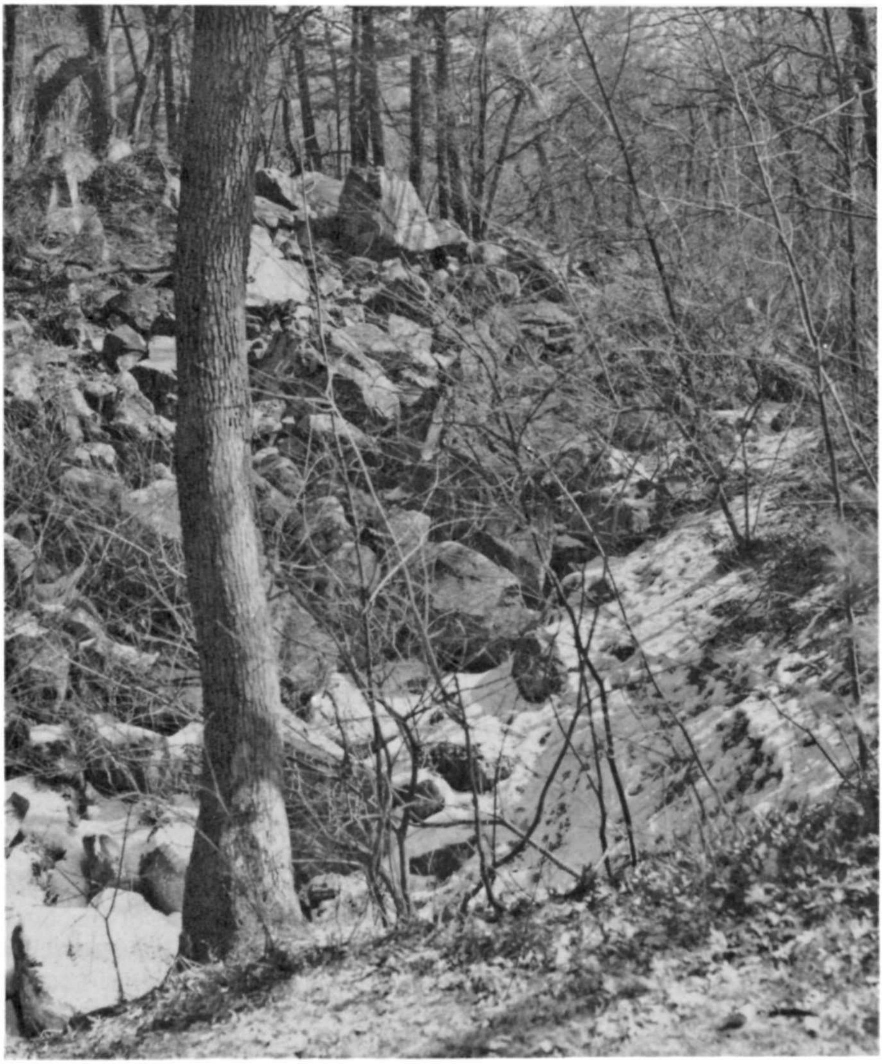

Fig. 83. Elongate depression at the base of talus along the south-facing slope of East Bluff, Devils Lake.

sandstone, and conglomerates occurring along a shallow valley and adjoining gentle slopes. Some of the blocks are almost buried in the soil, while others appear to be largely above the ground surface. Locally the blocks are jumbled together. This area is very close to the Cary ice front where it butted against the northeast corner of West Bluff. Some drainage went around the end of the bluff and may have affected this particular area. Smith concluded that this material was produced in the same way as that of the talus on the south flank of the Baraboo Range south of Devils Lake. However, the sandstone and conglomerate in this deposit call for different source rocks and the relative histories of the two locations are distinct. Some concentration by water working seems evident though surely not much rubble could have been produced in this way or the angular blocks would have been rounded.

In the extreme southeast corner of sec. 23, southwest of Devils Lake, a small ridge of Baraboo quartzite juts up above the level of the south fork of Messenger Creek. The relatively flat top of the ridge reaches an elevation of about 1100 ft but a large isolated pinnacle rises fully 20-30 ft above the level of the ridge and isolated rocks and smaller pinnacles are also present to the north. Removal of the upper part of the quartzite to leave the isolated monuments and pinnacles could have been accomplished possibly in Glacial Devils Lake if it had reached this general level, but the origin of some of these features obviously is conjectural.

Windwork is not common in the recommended park area. A thin accumulation of loess has been brought in by wind and deposited over the upland surface. This loess probably is latest Wisconsinan to Recent in age according to immaturity of weathering. Such deposition is common on uplands adjacent to abandoned lakes or glacial outwash such as have been so much in evidence around the Baraboo area. The sources of the loess could well have been Glacial Lake Baraboo to the northwest and the outwash apron in the Wisconsin River Valley to the south. That wind has been strong in the area is attested to by wind polished and fluted surfaces that may be seen outside the recommended park area south of Baraboo on quartzite knobs that rise above the early Paleozoic formations. There ventifacted, furrowed surfaces suggest strong winds from the west-northwest. Some polishing of the corners and faces of some of the upland cliffs of the Baraboo area have been attributed to windwork, but we cannot exclude waterwork and chemical action from such alteration.

Conclusion

The present boundaries of Devils Lake Park are too restricted to include a representative sample of the intricately looped moraine of the late Woodfordian or Cary ice advance and its associated glacial phenomena. To preserve a representative portion of this moraine and its associated phenomena, the park boundaries need to be expanded about three times. Particularly to the northeast, the recommended boundary barely includes the terminal moraine and very few of the retreatal moraines which were formed shortly after the farthest point was reached. Every geologist who has written extensively on the park has emphasized the uniqueness of the glacial and periglacial phenomena in the Baraboo Range. No other situation known to me has such a rich variety of unique features in so small an area near major centers of population. As a tourist area and scientists' field laboratory, it is unrivaled in the upper Midwest region.

Every effort should be made to achieve a park with boundaries at least out to those recommended. To do less would be tragic because the unconsolidated moraine is easily destroyed. No better moraine or combination of moraine and other features exists along this front from the Great Plains to the Atlantic Ocean. It is truly a unique area that should be dedicated to future generations of students and laymen alike. The story the Devils Lake area can tell should not be preserved only in print.

8 Mill Bluff Pinnacles

Mill Bluff (Fig. 84), currently a state park, is the locus of a number of "Bluffs" with appropriate names (Figs. 1, 2, 3, 46, 85). Each, whether called Bluff, Monument, or Rock (Fig. 85), generally has had a similar origin and is composed of the same Upper Cambrian sandstone (the Ironton Member of the Franconia Formation caps many and the Galesville Member of the Dresbach Group forms the base). They are striking features because they rise so abruptly from such a flat surface (Fig. 86), and because their yellowish sandstone contrasts with the green vegetation that surrounds and partly covers them. Some are pinnacles in the sense of being a tall, slender, pointed mass (Figs. 87, 88). More resemble the buttes of western United States (Figs. 89-91). Others are rounded (Figs. 92, 93) or irregular, like Ragged Rock and the unnamed ridge west of Round Bluff (Fig. 85). Genetically the geologist would call them "stacks" for their precipitous sides have been carved by wave and current action in Glacial Lake Wisconsin, where they stood as isolated islands during one or more ice advances of Woodfordian age into central Wisconsin. They are "outliers" of the Franconia cuesta, the Upper Cambrian sandstone that to the southwest and south comprises a continuous upland surface at and slightly above the level attained by the pinnacles. The sandstone lies unconformably on the Precambrian shield rocks from which weathering and erosion is stripping it away (Fig. 3).

The state park is a very restricted area encompassing Mill Bluff only (Fig. 84). In order to insure an uncluttered view and the preservation of a representative sample of the pinnacles the present park should be expanded to include all shown in Fig. 85. Although the sandstone comprising them is generally of little commercial value, it is soft and easily modified by man's use. Protection by inclusion in the Reserve is mandatory for their future best use.

Martin (1932:317) succinctly described some of the pinnacles and states their value:

If a traveler, on his way from eastern United States to the Pacific coast, be fortunate enough to cross central Wisconsin by daylight he will pass through the village of Camp Douglas or the village of Merrillan. At Camp Douglas and Merrillan, and for many miles nearby, he may see landscape features totally unlike those anywhere else in the United States east of the Mississippi River. The hills of the region near Camp Douglas are buttes and mesas. They have the straight lines, steep cliffs, and sharp angles of an arid country rather than the soft curves of a humid region. . . . The features to be seen . . . are (a) isolated, rocky hills which resemble ruined castles, (b) grostesque towers and crags of sandstone along a line of bold, irregular bluffs, and (c) an unusually-flat plain, which stretches away beyond the northern and eastern horizons. The bluffs and steep slopes on the west and south form the escarpment at the border of the Western Upland [Southwestern Upland of Black 1964b:171-177]. The level country is the Central Plain of Wisconsin. Not all of the Central Plain is exactly like the Camp Douglas Country. This, however, is a representative part, and one of the most beautiful and striking.

The flat-topped buttes in this area owe their angularity of outline to the selective cementation of certain horizons in the Dresbach Group and Franconia Formation of Upper Cambrian age. The cement commonly is ferric oxide, but silica seems almost equally abundant or important. These "case harden" the rock, making it more resistant to normal weathering and erosion which then must attack joints. By

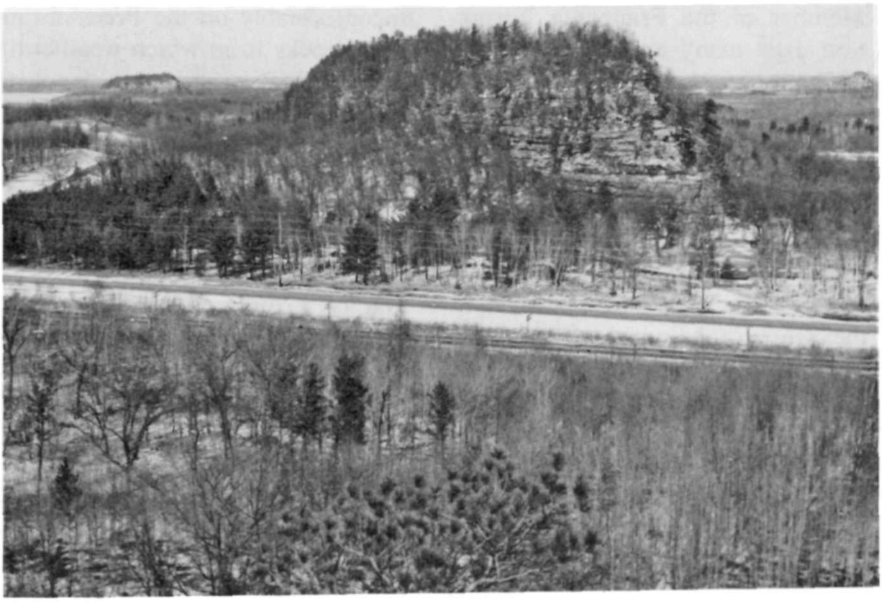

Fig. 84. Mill Bluff, looking northward from Round Bluff. See Fig. 85.

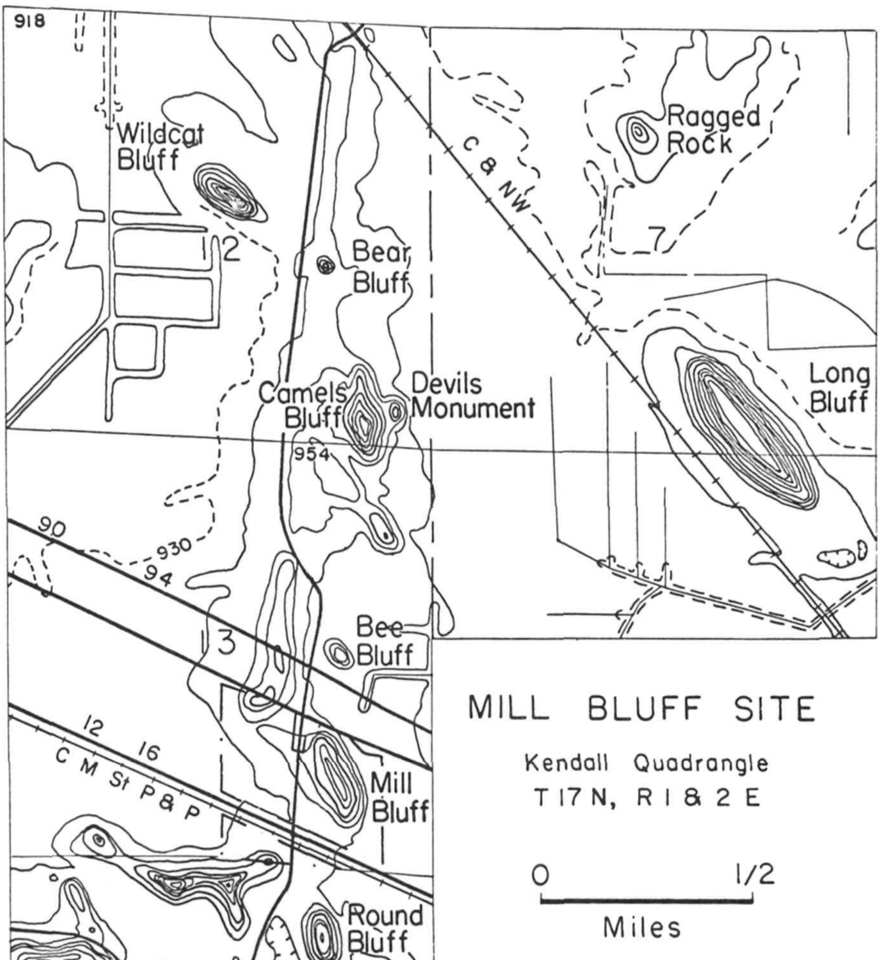

Fig. 85. Topographic map of the Mill Bluff area. After the U.S. Geological Survey Topographic Quadrangle—Kendall.

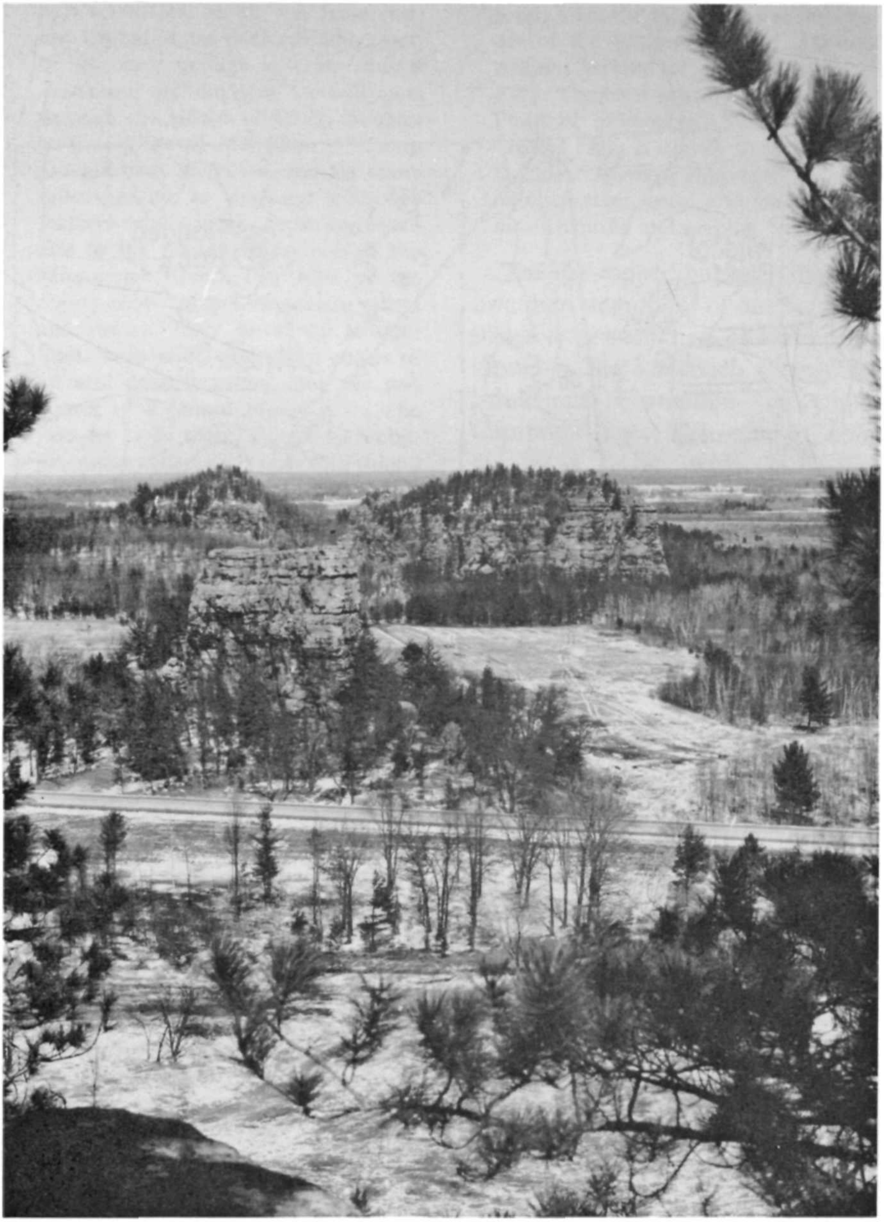

Fig. 86. Bee, Camels, and Wildcat Bluffs looking northward from Mill Bluff. See Fig. 85.

Fig. 87. View southeastward from Camels Bluff toward Camp Douglas. Spire, about 40 ft high, is here called Devils Needle.

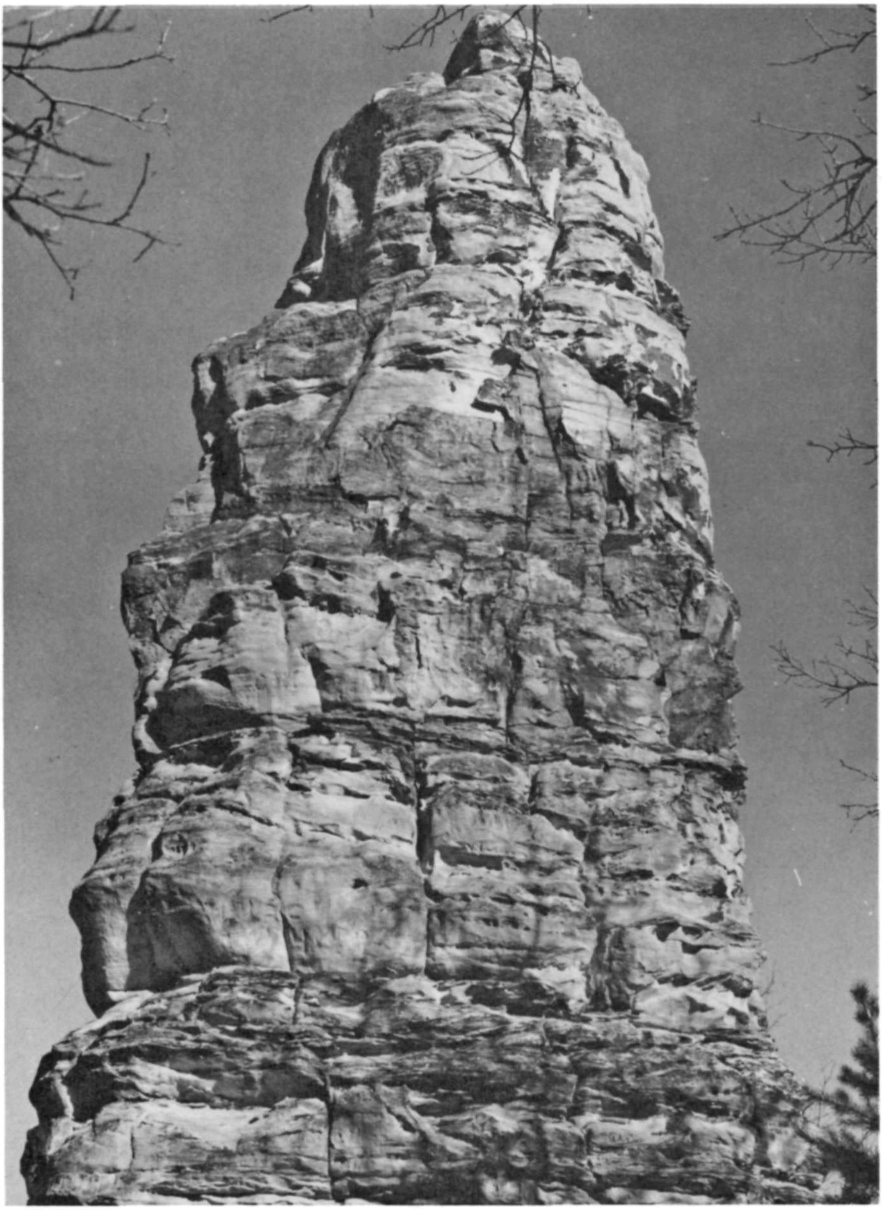

Fig. 88. Detail of Devils Needle, immediately south of Devils Monument.

Fig. 89. Bee Bluff, looking northeastward from County Highway W overpass at I-90. See Fig. 85.

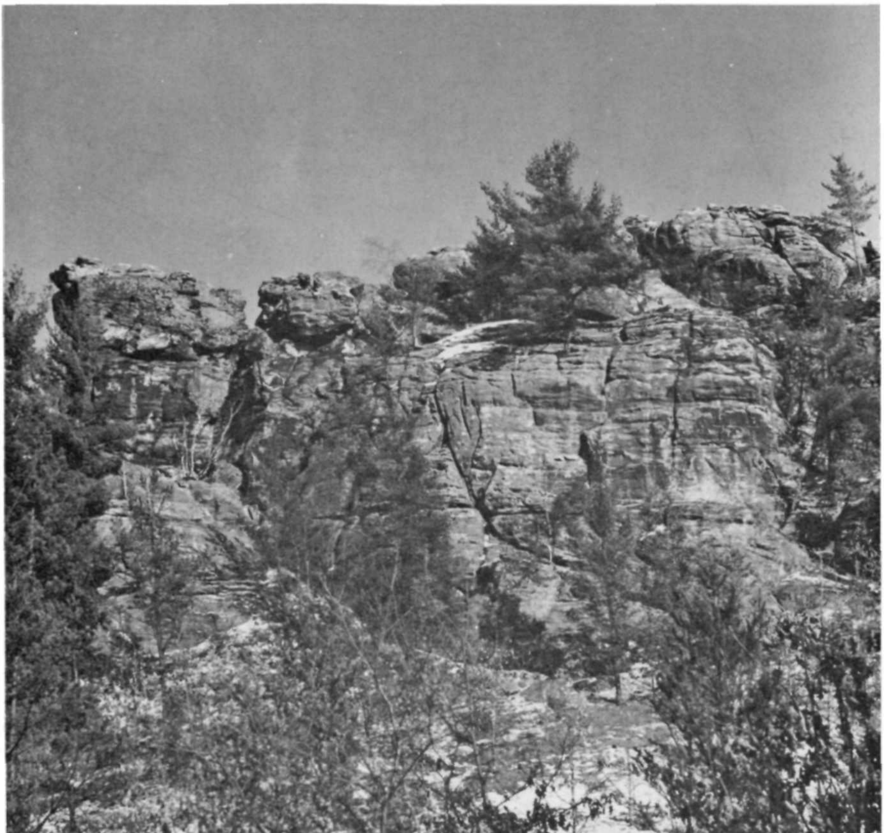

Fig. 90. Camels Bluff, looking eastward toward the northern part. See Fig. 85.

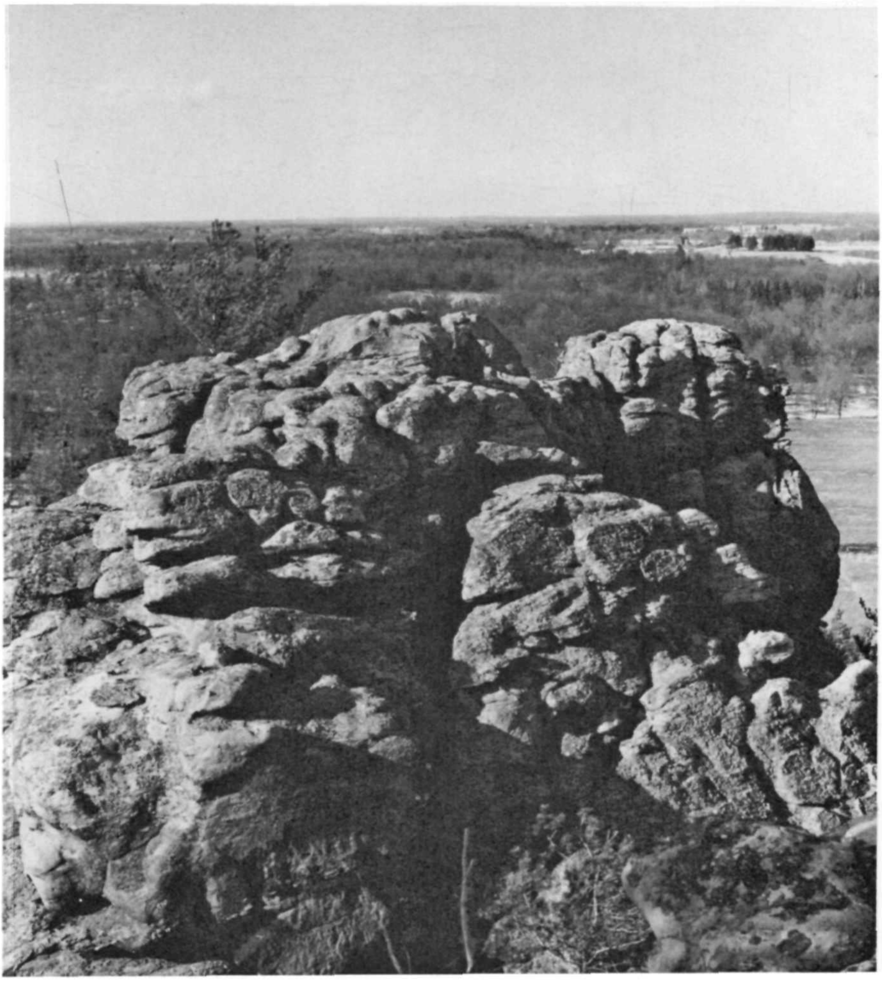

Fig. 91. Camels Bluff, at the north end, looking north-northeastward at part of the upper Cambrian sandstone and the plain of Glacial Lake Wisconsin.

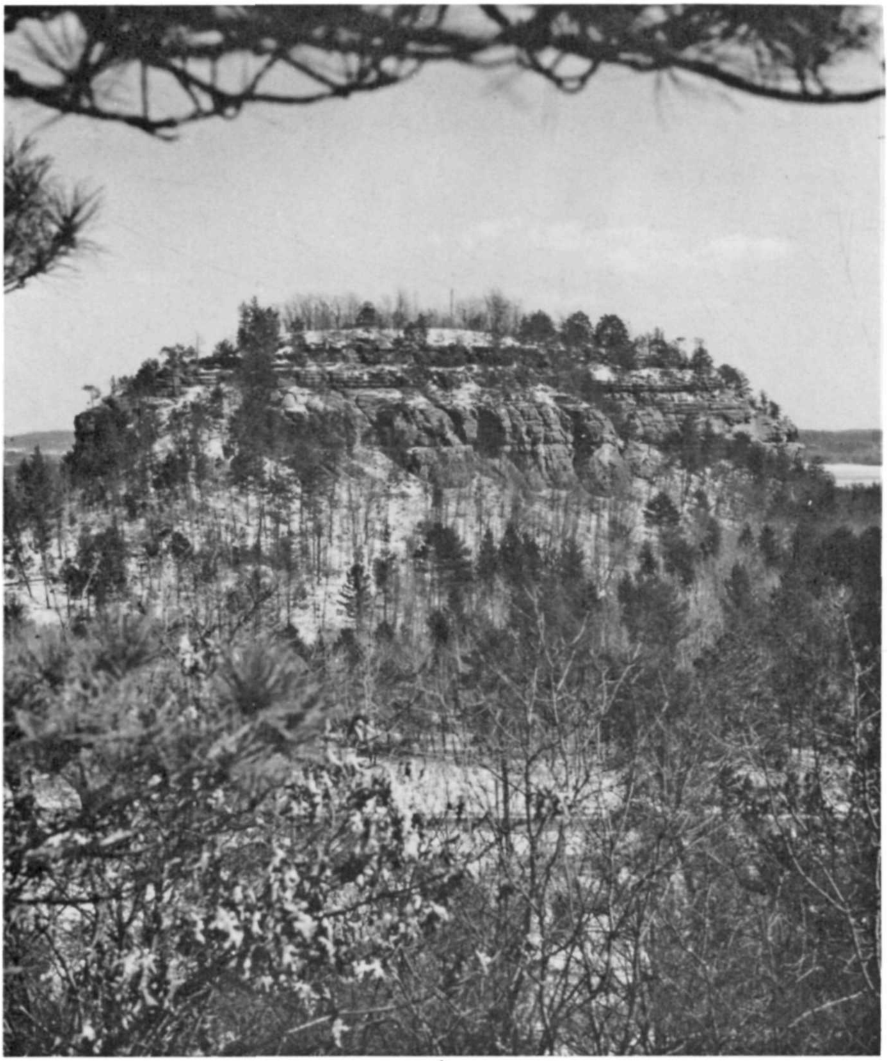

Fig. 92. Round Bluff, looking east-southeastward from the ridge to the west. See Fig. 85.

Fig. 93. Wildcat Bluff, looking northwestward from the north end of Camels Bluff. See Fig. 85.

breaking the rock along joints and bedding planes, especially through frost action, the angularity of the buttes is preserved. As soon as the cemented horizons are removed, however, and particularly in the Galesville Member of the Dresbach Group, the lack of cement shows up by a general rounding of outcrops and the transition of the angular buttes to conical hills.

The selective cementation has not been studied in detail. Presumably it came about by ground water activity and is not restricted to any particular horizon over a broad area. Banding of zones of dark red iron oxide contrasts markedly with white to yellow quartz sandstone locally in what is called "zebra rock." The iron-cemented zones in pipes, layers, bands, and the like, in places only a fraction of an inch in size, stand out on weathered surfaces as resistant nodes. In places the zones emphasize cross stratification, ripple marks, and fossils; in other places the iron oxide zones cut across them. Where the sandstone is free of iron oxides, it can usually be removed easily with one's fingers.

The grotesquely weathered forms are considered by Martin (1932: 329) to have originated prior to glaciation and elsewhere to have been removed by the ice except in the Driftless Area. However, Weidman (1907:602) stated flatly that their picturesque forms clearly were acquired after early ice invasions. I recognize that the various pinnacles, monuments, and buttes are not common in areas covered by the late Woodfordian ice, but they are common and well-developed in areas covered by the much earlier Rockian (late Altonian) ice. However, local conditions may be more important than time in the formation of such features. Certainly wave action in Glacial Lake Wisconsin accelerated erosion there, and similar features are seen today being produced along the south shore of Lake Superior.

The area of Footville Monument (Frye et al. 1965:43-61) in southern Wisconsin and many other pinnacles (Fig. 94) of sandstone are known to have been covered with ice as late as 30,000 years B.P. , and the bizarre forms evolved since. This is not a surprising or unusual rate of erosion when one sees how soft the rock actually is. One castellated spur in southern Wisconsin (Frye et al. 1965:43-61) is reported to have "grown" from a sandstone ridge during the lifetime of the local residents. It too was formerly glaciated. The larger buttes and mesas generally require more time than the smaller pinnacles and monuments.

Locally the sandstone is sufficiently cemented to retain glacial polish and striations such as may be seen on one of the stone steps in the trail to the top of Mill Bluff. (The possibility that the surface was produced by faulting cannot be ruled out entirely.) Its quarry site has not been located, but surely is nearby. To the north the nearest prominent moraine of the older drift lies near the north boundary of Jackson County, and that of the late Woodfordian ice lies to the east in Adams and Waushara counties. However, Akers (1964) reports in detail on glacial drift and other phenomena related to glaciation west and southwest of Tomah. That study, done in part by me and under my supervision, is considered indisputable proof of glaciation of the area 30,000 or more years B.P. Much of that story still remains to be told.

Glacial Lake Wisconsin at one or more times probably was joined with Glacial Lake Baraboo in a huge lake covering most of Juneau County and large parts of the adjoining counties (Figs. 1, 2). It formed in front of the Cary ice, and probably other advances during Woodfordian and earlier times, being trapped by the higher land to south, west, and north. So long as the present valley of the Wisconsin River around the east nose of the Baraboo Range was filled with ice, water in the Central Plains was backed up to the level of the lowest divide across the Southwestern Uplands—about 1000 ft in the vast swamp between the Black River drainage to the Mississippi River and that of the Lemonweir and Yellow rivers, now tributary to the Wisconsin River. The divide is

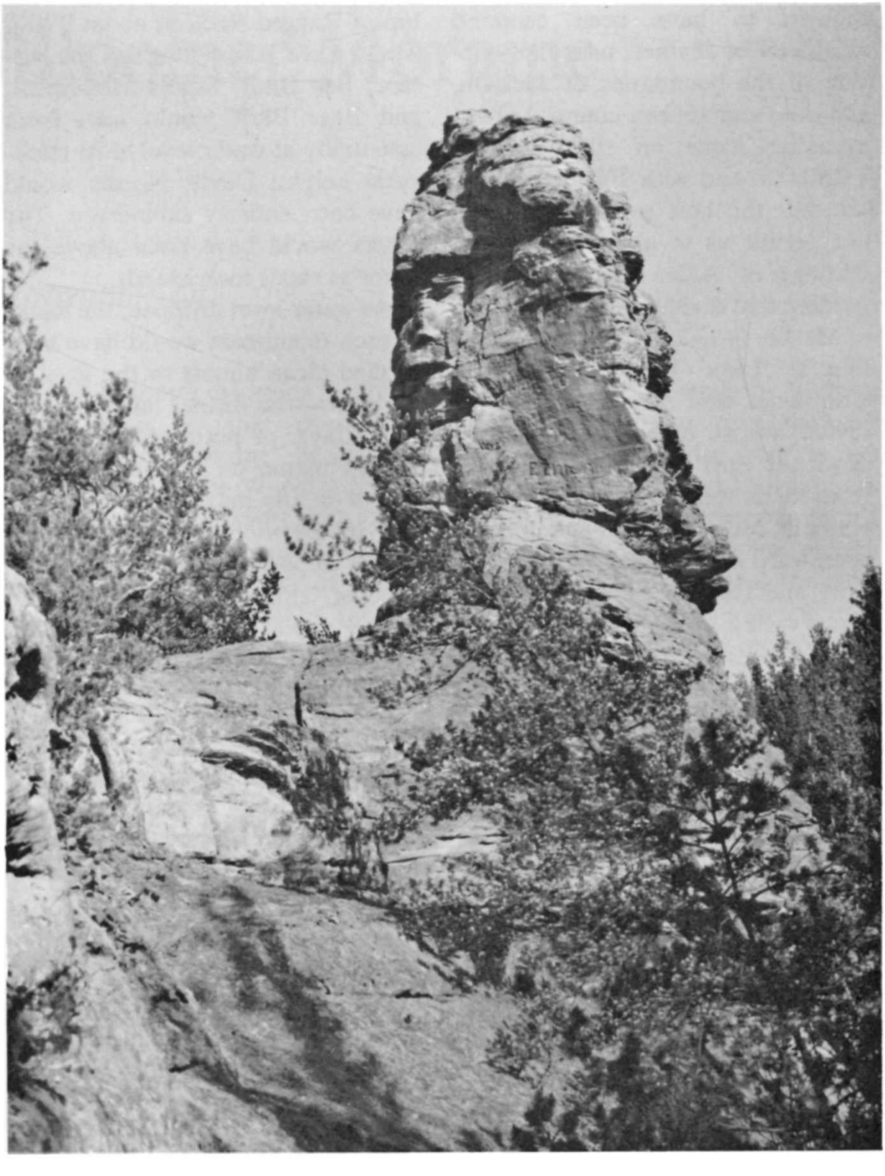

Fig. 94. Pinnacle, 30 ft high above left shoulder, at the north end of Bruce Mound, 2 miles east of Merrillan, Clark County. The pinnacle is at an elevation of about 1240 ft, well above the level of Glacial Lake Wisconsin but several miles within the boundary of the older drift. It is believed to have formed since glaciation of the mound.

thought to have been centered northwest of Mather, near the junction of the boundaries of Jackson, Monroe, and Juneau counties. Topographic maps on the scale of 1:250,000 and with 100-ft contours are still the best available and do not permit us to alter the original findings of Alden (1918:223) regarding that divide.

Martin (1932:340) believed that Glacial Lake Wisconsin merged with water west of the Black River divide, whose outlet was westward down the East Fork of Black River from Scranton, Wood County, to Hatfield, Jackson County and thence southward down Black River. Near Pray and Hatfield the stream has a broad valley—presumably oversized for the present stream. Martin attributes its granite and greenstone boulders to be erratics from Glacial Lake Wisconsin, but I have found moraine with many igneous erratics south of that stream.

Alden (1918:223) found erratics in the Lake Baraboo basin to an elevation of 980 ft. The moraine at the north end of the West Bluff of Devils Lake is firmly plastered against the rock at an elevation of 1060 ft, leaving little doubt that the level of Black River divide must have been reached. The level of the plain shown in Fig. 85 is about 920-940 ft.

Assuming Glacial Lake Wisconsin attained a level of about 1000 ft, and ignoring any isostatic effects of the weight of water and ice on the land, the Mill Bluff area should have been a lake 60-80 ft deep. The top of Ragged Rock at about 990 ft would have almost reached the surface; Bee Bluff, Devils Monument, and Bear Bluff would have been essentially at water level at its maximum height; Devils Needle would have been entirely submerged. The others would have risen above the water as small rock islands.

As water level dropped, the flanks of each monument would have been washed clean almost to the level of the plain—the former lake bed with a thin layer of postglacial sand and organic matter on it. The upper few feet of debris may be seen in road cuts in the vicinity. Many tens of feet of deposits are known over an area of almost 2000 mile^2. The coarser deposits are in the north and east, close to the ice fronts; red silts characterize the southern part. Only brief study of the sediments has been made (Harloff 1942).

Alden (1918:225-226) reports on sediments deposited in Glacial Lake Baraboo that suggest more than one stage of filling separated by an accumulation of marsh muck. Squier (1916) reported crystalline rocks in rounded gravel and a buried log at a depth of 100 ft in the lacustrine deposits 6 miles northeast of Tomah. We know a lake occupied much of the Central Plains, at least once. The tremendous volume of material, breaks in stratigraphy, and our understanding of glacial fluctuations in Wisconsin suggest it existed several times. However, much study is needed to define the extent of the former lake, its sediments, and former history.

9
Bloomer Moraine

Introduction

The Bloomer Moraine, as used here, refers to that portion of the "dead-ice moraine" or ice-stagnation area of late Woodfordian or Cary age in the vicinity of the town of Bloomer, Chippewa County (Figs. 1, 2, 95, 96). This is part of the terminal moraine that Chamberlin (1883a:381) referred to as the Moraine of the Chippewa Valley Glacier of the Second Glacial Epoch. This usage was with reference to the lobate nature of the ice sheet in northwestern Wisconsin rather than to a distinctly separate glacier apart from the continuous ice sheet. His map covering that part of the moraine (Chamberlin 1883a:35) showed the generalized flow pattern of the ice and the general distribution of the moraine. His verbal description of the characteristic features of the moraine of the Second Glacial Epoch is quoted in Chapter 4 and need not be repeated. Leverett (1929, Pl. 1) shows the position of the outer edge of the moraine at a somewhat larger scale, but includes no verbal description of the moraine in the Bloomer area. Reference of the Cary and older moraines to the north and west is made by Mathiesen (1940) whose map includes about 2 mile^2 of the front in the area of Fig. 96. Published data are singularly lacking in the area, and I made only brief trips to certain portions. It is obvious that we know less of the details of this area than of any of the others being recommended for inclusion in the Reserve.

However, just because we know less about the area is not reason to downgrade its importance for inclusion in the Reserve. Actually I consider it one of the four major areas that will receive most use—on a par with the Northern Kettle Interlobate Moraine—particularly if sufficient area and variety of features can be included. Because the Chippewa County Park must for practical reasons provide the nucleus, attention has been focused on it and its immediate surroundings. This cannot be helped, but is regrettable because a number of superb features are several miles removed from it. Consequently, even though funds will not permit a larger area to be recommended at this time, I am in effect ignoring

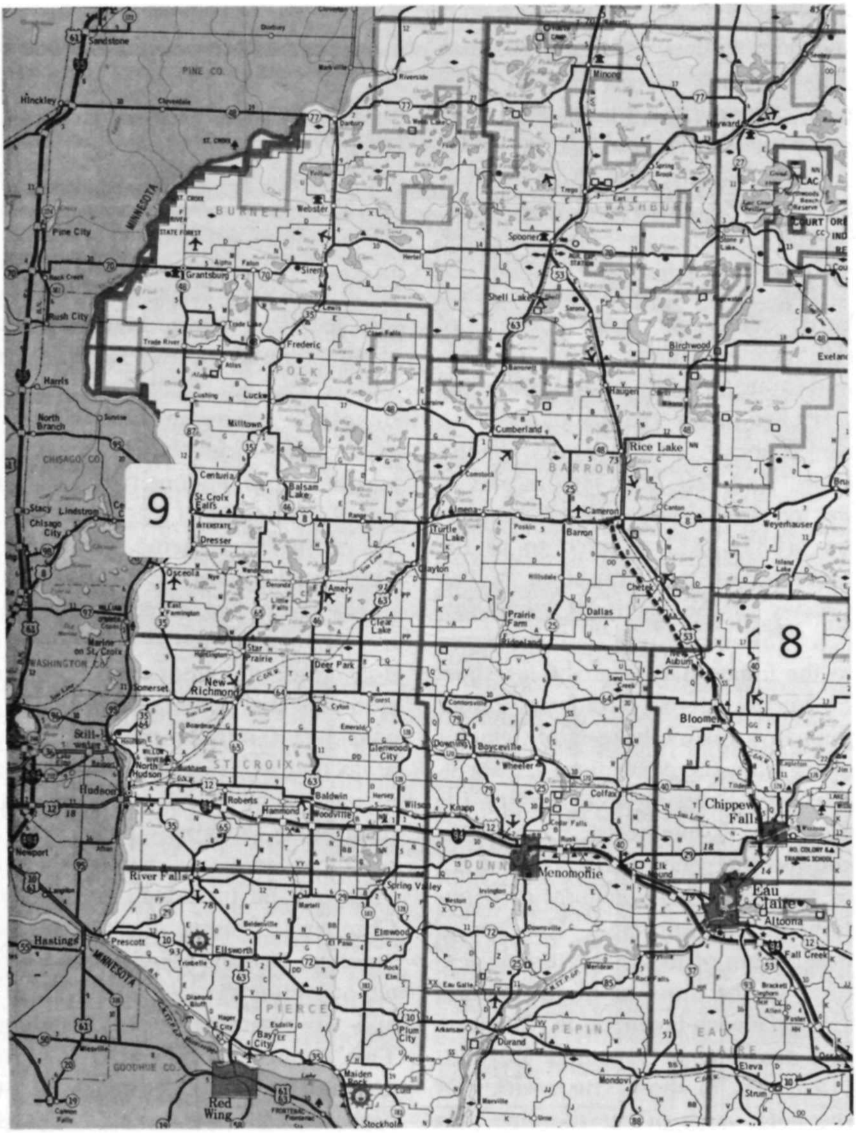

Fig. 95. Part of Wisconsin State Highway map, showing locations 8 and 9.

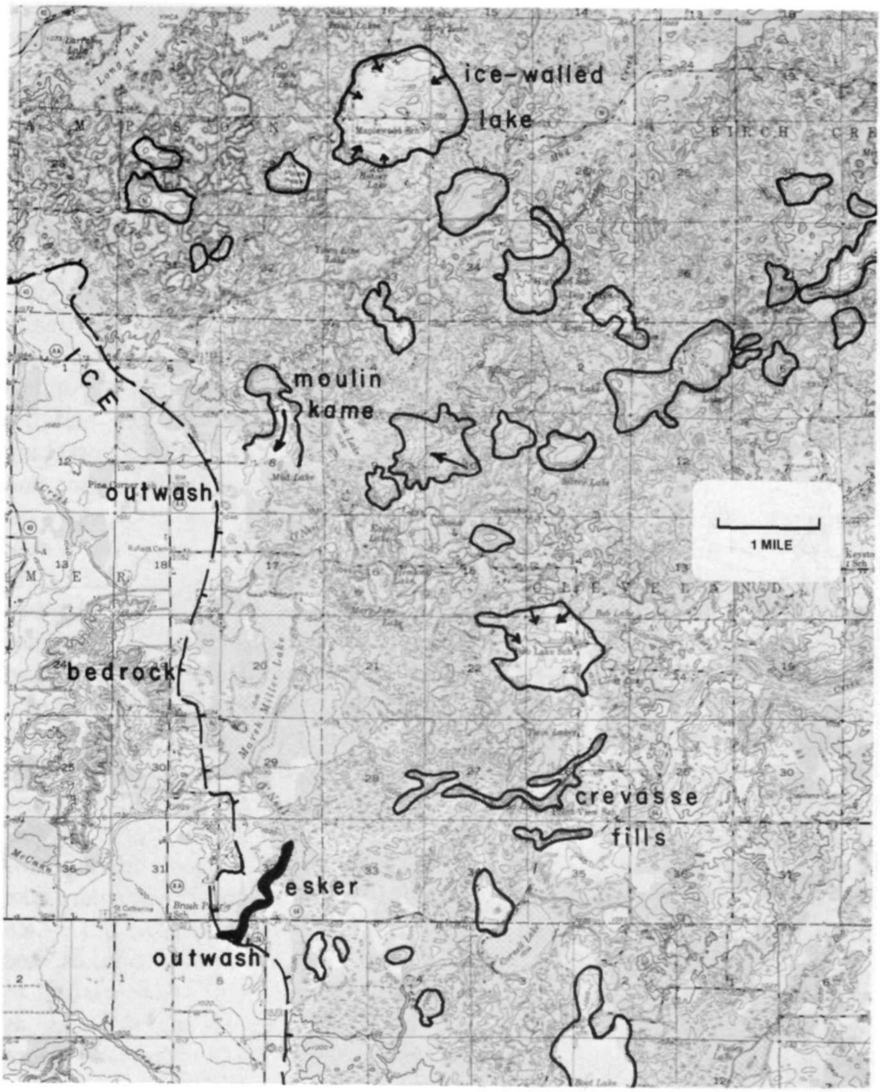

Fig. 96. Part of U.S. Geological Survey Topographic Quadrangle—Bloomer, showing by heavy, ticked line the prominent front of the dead-ice moraine, by solid line the outlines of former ice-walled lakes and other named features, and by arrows the direction of former water flow.

that recommendation in order to illustrate and present a case for representative features outside the recommended area. Therefore, Fig. 96 includes the recommended boundary and some of the more unusual features in it, but also many which will be illustrated or described near it so the reader too may judge their importance for inclusion in the Reserve.

Description of Glacial Phenomena

Figure 96 contains a representative area of the dead-ice moraine of the Chippewa Lobe of Cary age, its associated outwash, and a representative area of Cambrian sandstone-supported, subdued hills with thin drift cover whose age seems only slightly more than that of the dead-ice moraine. Bedrock outcrops within the dead-ice moraine are rare, but they are common in the hills westward outside it.

The dead-ice moraine is one of the best developed in the state even though its features are smaller and more compact than in most other dead-ice moraines in the northwestern or northern part of the state. Because most knobs and swales are smaller, they are more readily seen and appreciated.

This part of the Bloomer Moraine is characterized by small kettle lakes —commonly less than 0.25 mile across—although a few larger ones are present (Figs. 97-102), by kettle swamps (Fig. 103), or by dry kettles (Fig. 104). Most of the lakes are shallow with very soft to soft water (Sather and Threinen 1963). Generally, relief is only 20-60 ft in the knob and swale topography in till. The till is mostly dark red, sandy, and stony. However, knobs and irregular hills of washed drift with relief of more than 200 ft are present. These are part of an interesting family of features genetically related but markedly different in size and appearance. All are due to the stagnation of the ice and the consequent melting and cracking through to the base of moulins and crevasses. Some such openings were favorably located to receive vast floods of melt water and great quantities of debris. In Fig. 96 they are located especially on two lines —one trending southwesterly and another at right angles or northeasterly—parallel and perpendicular to former ice-flow directions. The flood of water enlarged the openings and deposits grew from simple conical moulin kames or small crevasse fills to large and more irregular lake beds and complexes bounded by ice on all sides. Into those lakes vast quantities of debris were carried to build ice-walled lake deposits as much as 1 mile across and more than 200 ft high. The deposits now resemble small rounded buttes and mesas.

Characteristically the melt waters of glaciers varies drastically in volume depending on diurnal and seasonal temperature cycles and other factors. Mere trickles of water with

Fig. 97. View southward of kettle lake in the south-center, sec. 15, T. 31 N., R. 8 W.

Fig. 98. View eastward of kettle lake in the west-center, sec. 27, T. 32 N., R. 8 W.

Fig. 99. View north-northeastward of kettle lake in the NE¼NW¼ sec. 8, T. 31 N., R. 8 W.

Fig. 100. View southwestward of kettle lake in the NW¼ SW¼ sec. 5 and the NE¼ SE¼ sec. 6, T. 31 N., R. 8 W.

Fig. 101. View southward of kettle lake in the NW¼NW¼ sec. 26, T. 32 N., R. 8 W.

Fig. 102. View southwestward of kettle pond in the NW¼ SE¼ sec. 6, T. 31 N., R. 8 W.

Fig. 103. Kettle swamp in the SW¼NW¼ sec. 32, T. 32 N., R. 8 W.

Fig. 104. View southeastward of large dry kettle from ice-walled lake deposit in the SE¼ NW¼ sec. 32, T. 32 N., R. 7 W.

little debris undoubtedly changed literally overnight to great torrents of water capable of transporting almost any rock fragment the glacier provided. Debris slides and slumps off the ice banks would have been commonplace; undercut ice masses would fall in the lake or be planed off on top. Ice masses loaded with debris would be unable to float and would be buried in the lake sediments to later produce kettles. Deltas around the margins were especially common. The coarser material would tend to remain at the periphery; the fines would be carried to the lower centers. The large ice-walled lake deposit in secs. 21 and 28 is a beautiful and very representative example (Fig. 96).

These ice-walled lake deposits are far too large to have been simple cavities under the ice—they were unquestionably open to the sky, they are relatively simple in concept, but exceedingly complicated in the details of their construction. The surface shows only the partial recording of the last events. The buried deltas, slumps, slides, flows, and layers of sediments have little direct manifestation at the surface except through artificial exposures. However, the former buried ice blocks, also characteristic of such complexes, leave their telltale depressions on melting. The larger deposits probably started as moulin kames or crevasse fills. They now rise well above the general level of their surroundings in reverse relief. They stand today as mute testimony of their former history in a wide range of sizes and shapes (Figs. 105-109) of which only some of the larger or higher ones are shown in Fig. 96. The smaller ones cannot always be distinguished by surface inspection from normal till knobs. Little subsurface exploration has been done on them. Yellow brown, brown, and red silty sand and local thin layers of gravel were seen most commonly in natural or artificial exposures. In many places the flat-lying undisturbed stratification of relatively quiet waters was seen. The poorly sorted silty sand is particularly subject to slump and flow in road cuts, and exposures do not remain open long.

In the recommended area, secs. 27-30 contain all or parts of five characteristic ice-walled lake deposits. Pikes Peak in sec. 29 has been named although it is neither the highest nor sharpest of these features. It probably started as a simple moulin kame which opened to a circular lake. The elongate irregular mound in the southern part of sec. 30 may have been two separate moulin kames which in later life merged into a single lake. The elongate mound north of it seems to have been a single moulin that later grew westward as an elongate lake. The largest deposit in secs. 28 and 21 shows beautifully the higher rim of debris, generally coarser than the center, that was washed in from all sides. The mound in sec. 27 is favorably situated to show its abrupt outer walls (Figs. 107, 108) with lake sediments exposed in the road cuts.

Fig. 105. View southward of mound and kettle in the SE¼ SE¼ sec. 23, T. 32 N., R. 8 W.

Fig. 106. View southward of large conical ice-walled lake deposit with elevation of 1289 ft, in the NW corner, sec. 5, and the SW corner, sec. 32, T. 31 and 32 N., R. 7 W.

Fig. 107. View eastward of the outer wall of the ice-walled lake deposit in sec. 27, from the NE corner, sec. 33, T. 32 N., R. 8 W.

The most striking of these features is the conical peak with elevation of 1289 ft in the SW corner sec. 32 and the NW corner sec. 5, T. 32 and 31 N., R. 7 W. (Fig. 106). It is 4 miles east of the recommended area and immediately west of County Highway E which exposes lacustrine sediments in the lowermost flank of the mound. County Highway E to the north and east rises across a large complex area from which striking views of the adjacent kettles can be had (Fig. 104).

Fig. 108. View west-northwestward of the outer wall of the ice-walled lake deposit in sec. 27, from the SE corner, sec. 27, T. 32 N., R. 8 W

Fig. 109. View eastward of the top of the ice-walled lake deposit in sec. 28 from the NE¼ NW¼ sec. 28, T. 32 N., R. 8 W.

Views from the tops of the ice-walled lake deposits are especially attractive where slopes are steep and overlooks to present kettles are available (Figs. 110-113). Unquestionably the best one for a view is the sharp conical peak with elevation of 1289 ft (Figs. 106, 110, 111). The view from Pikes Peak is disappointing because the peak is too broad and lacking in lakes immediately at its base. However, views from those features east and west of it, in secs. 27 and 30 are good (Figs. 112, 113).

In secs. 5 and 8 west of Rock Lake and in secs. 9 and 10 east of Rock Lake are two interesting areas of washed drift. From an inspection of aerial photographs their surficial form and surface material suggests that they are moulin kames with associated drift laid down as outwash or eskers. No detailed investigation has been made, so the hypothetical reconstruction of their history is subject to modification. Nonetheless, the feature west of Rock Lake, which I visited briefly, shows washed sand and gravel in the southern part whose surface form and near surface structure demonstrate that it came from the conical hill to the north. Moulin kames during formation must have an outlet for the water that drops into the moulin. If water coming in exceeds outflow, a ponded water body forms. The outflow under the ice is commonly in the form of a subglacial stream in addition to ground-water flow through the porous material underlying the ice. Such streams leave eskers most commonly recognized as long sinuous ridges. However, if occupied long enough for the subglacial stream to meander or cut laterally, the deposit is broadened accordingly and often not recognized as an esker even though genetically it was formed in the same way. As the subglacial stream migrates laterally, the ice above being weak collapses behind it. No broad tunnel exists at any one time even though the stream deposits may be very broad. The two areas cited appear to be of this type wherein the subglacial streams originating at the base of the two moulins fanned out or migrated laterally to leave broad deposits of fluvial-glacial material. It is not known how much of the sand and gravel so laid down was actually deposited under the ice in a small stream or how much was deposited (particularly as in the case of the feature west of Rock Lake) subsequently in an area open to the sky between ice walls as the ice sheet thinned, broke up into blocks, and melted away. This moulin and associated outwash is especially close to the front. During the later part of its life it seems logical to have it go from a subglacial situation to one at least partly exposed. Kettles and ridges from the stagnant ice surround this moulin and esker on all sides.

Other ice-stagnation features well displayed to the south of the recommended area include crevasse fills and true eskers. They are difficult to

Fig. 110. View eastward from top of the ice-walled lake deposit at 1289 ft, in Fig. 106.

Fig. 111 View westward from top of the ice-walled lake deposit at 1289 ft, in Fig. 106.

Fig. 112. View southeastward to Plummer Lake from southeast corner of the ice-walled lake deposit, sec. 34, T. 32 N., R. 8 W.

Fig. 113 View southwestward from the ice-walled lake deposit in the south-center, sec. 30, T. 32 N., R. 8 W.

distinguish from each other on aerial photographs. The long sinuous ridge in secs. 26 and 27, T. 31 N., R. 8 W. (Fig. 114) is composed of red sandy till according to one cross section seen along the road in sec. 26. Its right-angle bends show beautifully the control by fracture patterns in the ice. Another higher line of ridges and knobs (Fig. 115) in sec. 35, T. 31 N., R. 8 W. borders an ice-walled lake area northeast of Cornell Lake; its composition is not known. A true esker (Fig. 116), part of which is in the southwest corner of Fig. 96, is being used as a source for sand and gravel and is exposed in pits along Highway 64. Many other interesting features may be seen throughout the area but are not described here.

One of the best places to see the front of the dead-ice moraine, its associated outwash, and relation to the subdued bedrock hills to the west is 0.25 mile east of the Tillinghast School in sec. 35, T. 32, N., R. 9 W. From the moraine bordering the outwash there is an excellent view southwestward (Fig. 117). The moraine itself is dark red, sandy, stony till. Knob and swale topography in the till along the front is shown in Fig. 118. Good views of the front across its outwash plain can be had from County Highway AA, east of Highway 40 (Fig. 119).

Fig. 114 View northwestward of kettle pond and crevasse fill in the SW¼ NE¼ sec. 26, T. 31 N., R. 8 W.

Fig. 115. Crevasse fills or line of kames in the north-center, sec. 35, T. 31 N., R. 8 W.

Fig. 116. Esker, looking southeastward, in the NW¼ NE¼ sec. 6, T. 30 N., R. 8 W.

Fig. 117. View southwestward from the Cary front in the NW¼SE¼ sec. 35, T. 32 N., R. 9 W. This drift mantles Cambrian sandstone hills.

Fig. 118. Knob and swale topography in the Carey end moraine, County Highway AA, 0.9 mile east of State Highway 40.

Fig. 119. View northeastward across outwash to the Cary front, from County Highway AA, 0.4 mile east of State Highway 40.

The outwash plain locally is pitted showing that ice formerly went beyond the front as indicated in Fig. 96. The high concentration of igneous and metamorphic rocks from the Precambrian shield areas is well displayed in a borrow pit at the front on County Highway AA. Lake Superior sandstone is common but no dolomite was seen. The local Upper Cambrian sandstone is too weak to survive long in glacial streams and is recorded in the outwash mainly as sand.

The age relationships of the dead-ice moraine, its outwash, and the thin drift cover or sparse erratics on the bedrock hills westward are interesting. Leverett (1929, Pl. 1) and Mathiesen (1940:262) in turn advanced the front of the younger drift beyond that of the earlier workers, reducing considerably the actual area of pre-Cary drift in the reentrant between the Superior and Chippewa Lobes (Fig. 2). Based on soils, weathering of erratics, appearance and filling of kettles, and erosional phenomena, I would agree that little time difference seems to exist between the Cary and pre-Cary areas. One single ice pulsation in Woodfordian time may account for the younger drift in both lobes. Nonetheless the precise front of the Bloomer Moraine does have outwash from it showing that a time difference existed at least during deglaciation. Insufficient data are available to pursue this problem here, and further speculation seems unwarranted at this time.

Conclusion and Recommendation

In conclusion I wish to emphasize the marked difference between the large ice-walled lake features of the Bloomer Moraine and the sharply conical moulin kames of the Northern Kettle Interlobate Moraine. Although related in origin, the lake deposits clearly demonstrate the effects of large ponded areas open to the sky. The knob and swale topography and associated lakes and swamps of the Bloomer Moraine are representative small features more readily appreciated than larger features of similar origin to the north and northeast—farther removed from centers of population. This area is less disturbed by men than many comparable-size areas in the state. Hundreds of small lakes, ponds, and swamps offer a maximum variety of conditions for eutrophication studies of lake history. Unless controlled in a Reserve they probably will be lost because pressure from summer tourists and cottage owners is increasing rapidly.

The Bloomer Moraine presumably is approximately equivalent in age to the end moraine of Devils Lake, Cross Plains, and Kettle Interlobate areas. A range in time of many centuries may exist, but no quantitative data are available. Although little in detail is known of the area, it is a vital and different part of the Ice Age phenomena of

Wisconsin. If sufficient area and variety of features can be incorporated, such as illustrated or pointed out here, this area would become one of major importance rivaling the Northern Kettle Interlobate Moraine. It is hoped that the initially recommended area, expanding the Chippewa County Park, can be extended outward and southward to include many of the features shown in Fig. 96.

10
St. Croix Dalles Interstate Park

Introduction

"The most beautiful gem of scenery in the states of Minnesota and Wisconsin is the Upper Dalles of the St. Croix River, which forms a part of the boundary between these states." (Upham 1905:347.) Tourists come to the St. Croix Dalles mainly to see the gorge (Figs. 95, 120-122) and its supersize potholes —mute testimony to the accomplishments of vast torrents of running water during glacial times. Potholes are circular and irregular depressions in rock, produced by the grinding action of silt, sand, and small stones swirled in initially minor depressions by strong currents of water. They are not to be confused with "potholes" in earlier colloquial usage or even today, which are kettle ponds in moraine. The potholes of the St. Croix Dalles ". . . are unsurpassed by any other known locality in respect to their variety of forms and grouping, their great number, the extraordinary irregularity of contour and the much jointed diabase in which they are eroded, and the difficulty of explanation of the conditions of their origin." (Upham 1900:26).

Potholes are much larger and more numerous on the Minnesota side of the St. Croix River, but those in Wisconsin are good (Figs. 123-127). The Wisconsin potholes are as much as 6 by 10 ft across and 15 ft deep. Those in Minnesota are many tens of feet deep. Potholes of such tremendous size cut into hard basalt up to 100 ft above the river are unusual features. This was recognized decades ago by the governments of the states of Wisconsin and Minnesota, and in 1900 each state set aside areas on both banks of the St. Croix River in order to preserve them. Although each park operates independently, together they are known as Interstate Park.

No definitive detailed account of the geology of the park has been brought together since that of Berkey (1897). Upham (1900) provided detailed descriptions of the potholes

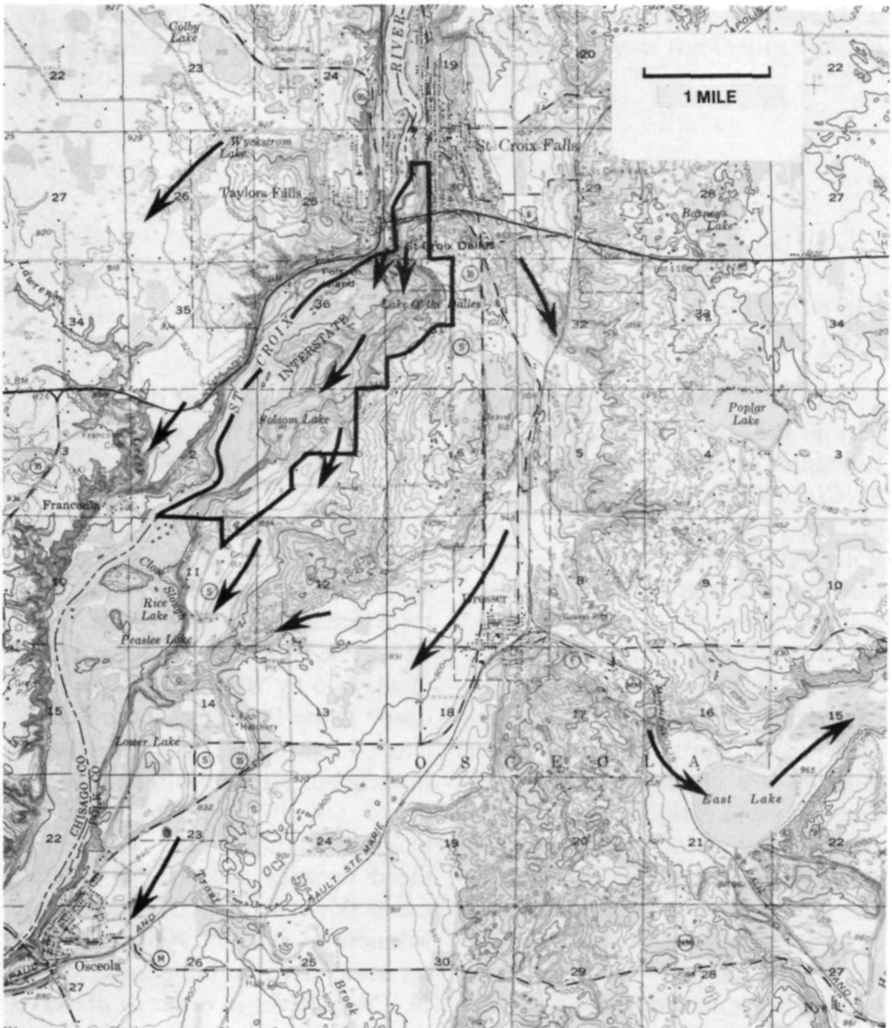

Fig. 120. Part of U.S. Geological Survey Topographic Quadrangle—St. Croix Falls, showing boundary of the recommended area and some arrows indicating directions of former water flow, both glacial and post-glacial undifferentiated.

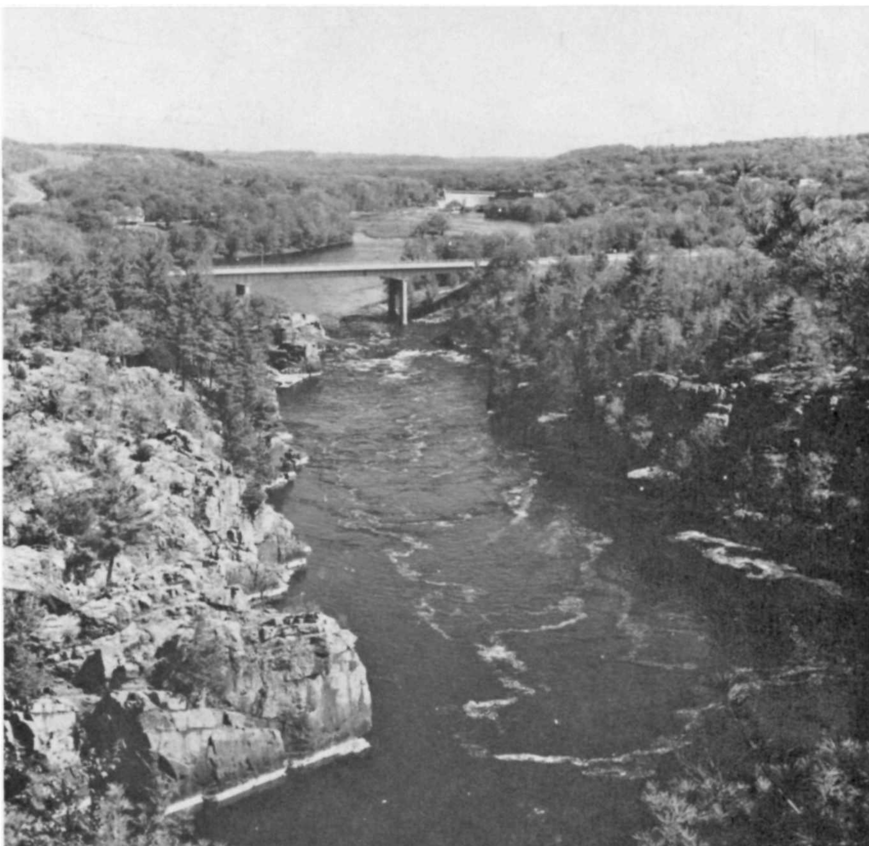

Fig. 121 Gorge of the St. Croix River at the Dalles, looking northward from Summit Rock.

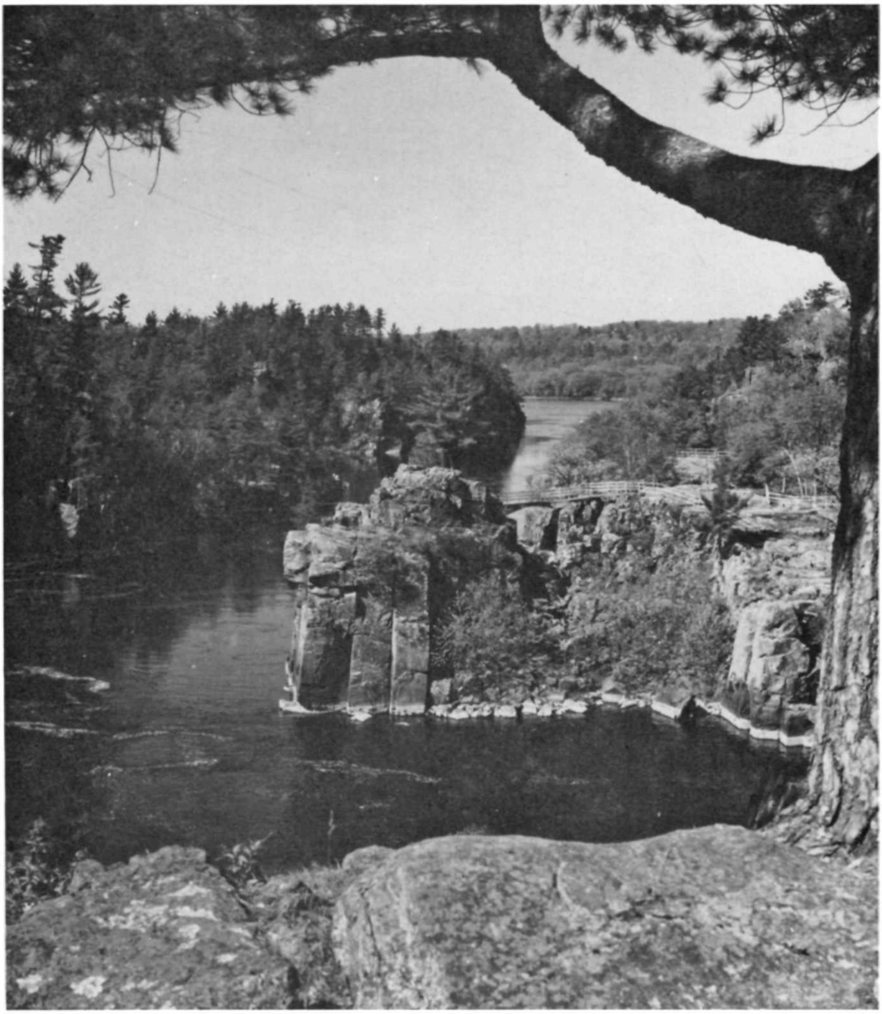

Fig. 122 Gorge of the St. Croix River at the Dalles, looking southwestward from the "elbow."

Fig. 123 Large pothole near the Old Man O' the Dalles. Geologic hammer for scale.

Fig. 124 Pothole, showing large crystals of feldspar in basalt.

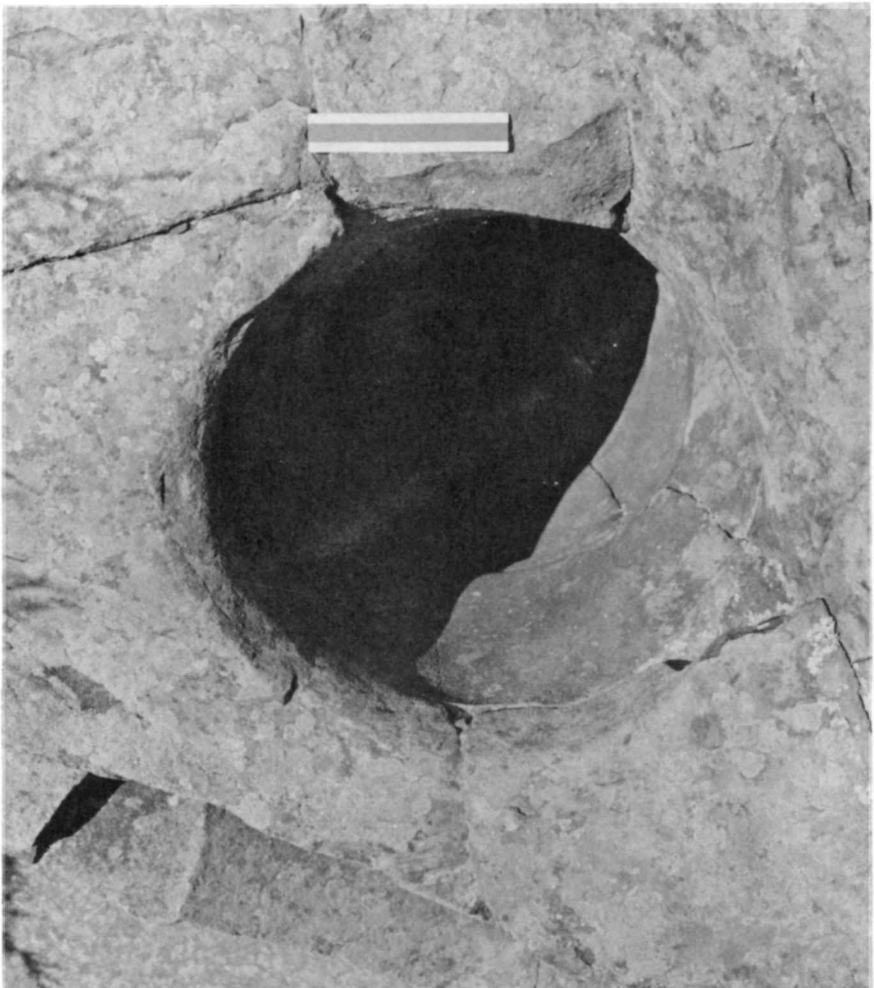

Fig. 125. Circular pothole.

Fig. 126. Small shallow pothole.

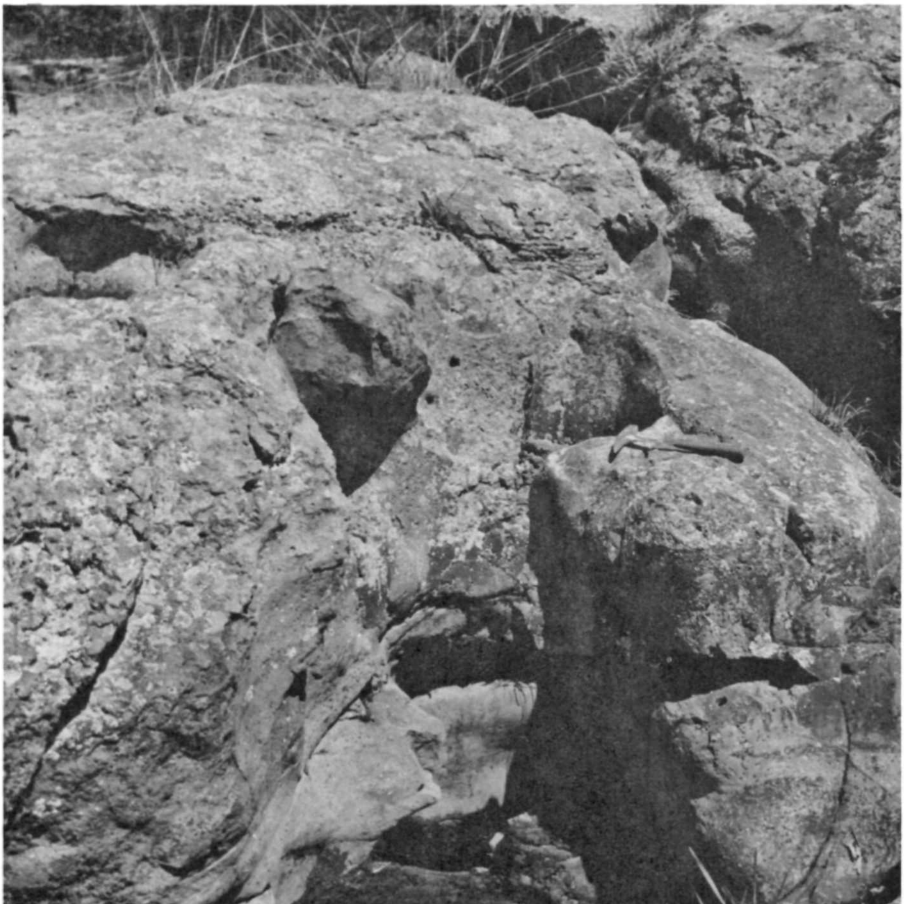

Fig. 127 Compound potholes.

and a possible origin. Upham (1905), Chamberlin (1905), and many others contributed to the knowledge of the history of the area, but strong differences in interpretation continued. Martin (1932:364-366) provides a simplified outline of part of the story. Leverett (1932) expanded the glacial history considerably. Cooper (1935) outlines in some detail the general history of the Upper Mississippi River. Sardeson (1936) attempted to summarize his own interpretation of the Pleistocene of the St. Croix River, but the story continues to be changed (Wright and Ruhe 1965:29-41).

I will not attempt a detailed review of the extensive literature available nor report on all aspects of the geology of the park and its surroundings. Reevaluation of the older works can only be done satisfactorily in the field, and for purposes of this report is not necessary. All who have ever seen the gorge of the St. Croix Dalles Interstate Park and its huge potholes are outspoken in their desire to preserve them uncommercialized. Inclusion in the Reserve of one of the most unusual collections of glacial phenomena in Wisconsin, or the world, is only fitting. Only slight enlargement of the existing park—extending it southward along the bluffs of the St. Croix River—is needed now to provide more room for the vast number of tourists who each year make use of the park's facilities.

General Description

The St. Croix Dalles is a 100-ft deep gorge cut in Precambrian basalt (Fig. 3) (several areas of Upper Cambrian sandstone lie uncomformably on the basalt in and adjacent to the park) of which seven individual flows have been recognized, rising like giant steps above the river (Martin 1932:364). The lava is well-jointed (Fig. 122). This aids in its removal by water, frost, and gravity and in formation of unusual pinnacled forms (Fig. 128). Large white or light gray feldspar crystals in the rock stand out against the black background but are ground smooth as is their host by pothole drilling (Fig. 124). Composition and texture of the flows in part, but particularly the structures of the rock, have controlled the location and irregularity of some potholes. They are particularly common along joint cracks and are elongated in them. Irregular sides at various levels in the potholes presumably also reflect variations in the rock, but no detailed studies have been made. Some small potholes are on smooth rounded shoulders of rock with no clue visible in the rock reflecting a weak point. I wonder (as did Upham 1900) whether some of them may have originated subglacially, for they are in anomalous positions for surface streams to start them. Such have been recorded in Norway (Gjessing 1965-66).

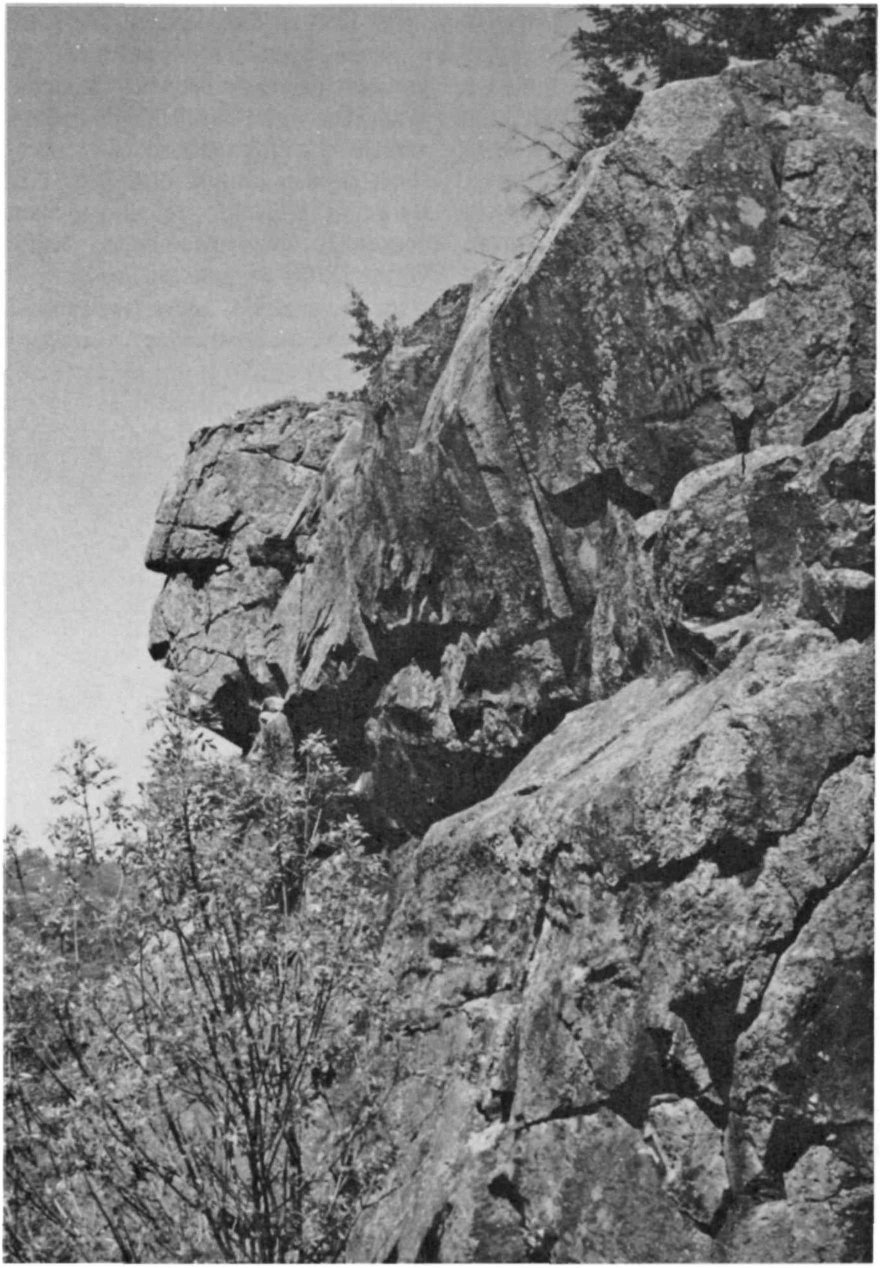

Fig. 128. Old Man O' the Dalles, resulting from fracturing and frost action.

The main gorge obviously was cut and is being cut by running water. Former torrential discharges during glacial wastage and glacial lake drainage sought the weakest route through the lava and partly excavated many other weak points as well. Excavation seems in large part to have been accomplished by physical removal of large joint blocks by the rushing waters. Lake O' the Dalles (Fig. 129) was at the receiving end of water that passed southward on the east and west sides of Summit Rock, the high point 0.2 mile northeast of the lake between the two short arrows shown on Fig. 120. Berkey (1897:352) considered the depression now occupied by Lake O' the Dalles to have been part of a gigantic whirlpool and as such to be one of the more striking features in

Fig. 129. Lake O' the Dalles, looking downstream.

the park. The west arrow is in Echo Valley (Fig. 130) where water excavated fractures in the rock, leaving broken cliffs (Fig. 131) and talus (Fig. 132) to develop in post-late Woodfordian (Cary) times. The east arrow is in Canyon Valley, a similar weak zone extending southward from the north-trending gorge of the Dalles. Both Echo and Canyon valleys were vacated when the southwest-trending part of the gorge was cut below their level. The fracture zone on which that part of the gorge developed may be traced northeastward on the Wisconsin side of the river as a distinct trough. Part of the deepening and removal of rock in the park area can be attributed also to ice action for striae are found at upper levels, but the quantitative effects of ice versus water are not known.

Lake O' the Dalles was formerly much larger, extending both eastward and westward. It is reported (Howard S. Kunsman 1937, unpublished report) that filling on the east side of 32 ft of clay, marl, and vegetable matter included wood with beaver teeth marks, plant seeds, leaves, and branches with the bark on. These came from excavations at what is now the parking lot for the lake. In addition, in the upper part of the deposit a copper pike or awl, 10.75 inches long, was found in peat with two flint weapon points and bones and horns of extinct bison (Palmer 1954). Lake O' the Dalles lies in part of only one of the earlier drainageways. Other routes of temporary water flow are shown by arrows in Fig. 120. Not all were occupied at once; some show by later development of kettle holes that water was flowing over buried ice blocks and are glacial rather than postglacial.

The washed areas along the St. Croix River are part of an extensive system of terraces of which five have been distinguished (Berkey 1897:352-354). Some are determined by rock ledges. The main street of the town of St. Croix Falls is on the next to the highest terrace. None of the terraces up to 920 ft elevation continue below the Dalles (Chamberlin 1905:256). Now that more detailed and accurate topographic maps and aerial photographs are available, these terraces need to be restudied along with the other drainage changes in the area.

The original map of Strong (1880:365-428) showed one set of glacial striae at the Dalles, with southeasterly trends and one major end moraine, and the inner part of the St. Croix or Kettle Moraine which originated from the Superior Lobe. Later Chamberlin (1905:249-250) increased the number of sets of striae to three in order of age—southerly, southeasterly, and east-northeasterly—and the number of drift sheets to four (Chamberlin 1910). The oldest drift was correlated with the Kansan Stage, but the same drift to the south is radiocarbon dated at about 30,000 years (Black 1959a). Obviously the Pleistocene history of the

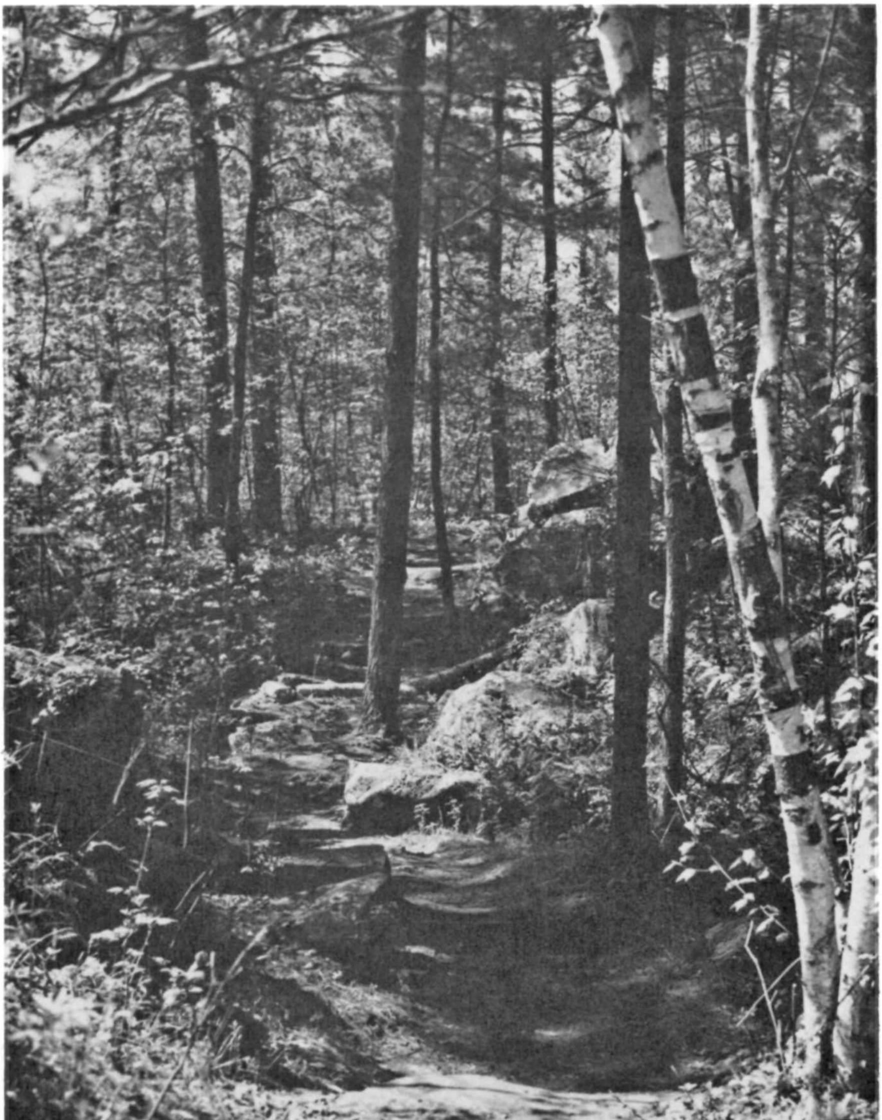

Fig. 130. Echo Valley trail.

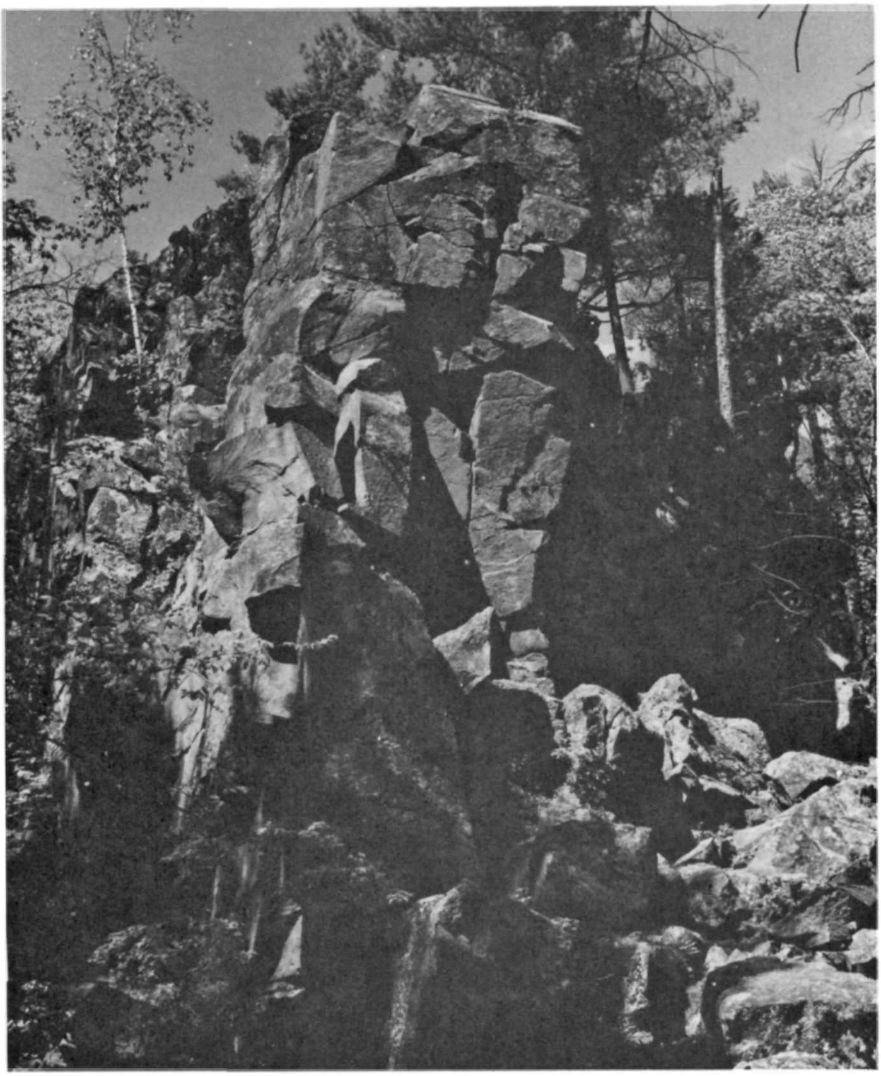

Fig. 131. Cliff of jointed basalt in Echo Valley.

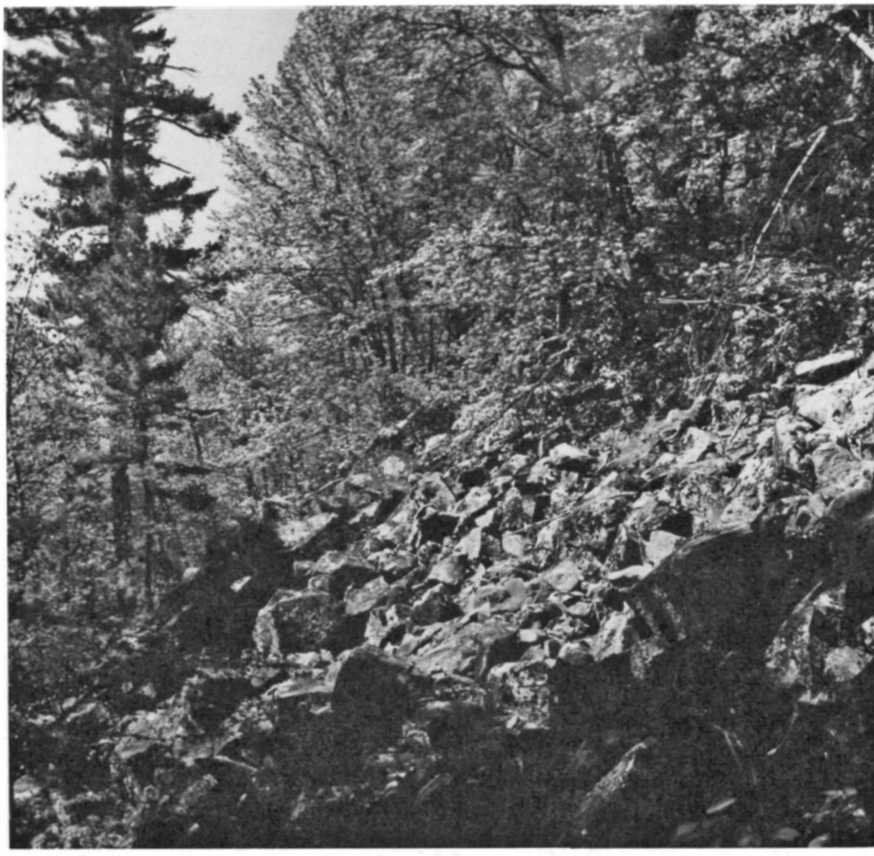

Fig. 132. Talus from well-jointed basalt in Echo Valley.

area is complicated (Wright and Ruhe 1965:29-41; Frye et al. 1965:43-61), and I have not done sufficient field work in the area to work out details or adjudicate differences of opinion. For purposes of this report this does not seem vital.

However, for full appreciation of the formation of the main features in the gorge an understanding of the sequence of events is vital. Unfortunately that sequence is not understood or at least agreed upon. Many geologists believe that the potholes can be attributed to torrential drainage from glacial lakes to the north. Recent work (Elson 1957; Zoltai 1965) now suggests that Glacial Lake Agassiz existed more than once and that it has drained into the

Lake Superior basin to add its enormous discharge to the glacial waters which at different times have gone down the St. Croix River. It is not clear to me just how many times nor when such discharges have come down the St Croix River, nor the effect each discharge has had in the formation of the particular features remaining in the Dalles region.

During late Woodfordian time the Grantsburg Glacial Sublobe, named for the town of Grantsburg in Burnett County, flowed northeasterly across the St. Croix River and established a moraine along a front from that town to the Dalles. It dammed Glacial Lake Grantsburg on its north side (Cooper 1935:23-38; Wright and Ruhe 1965:29-41) which also received the drainage of the Mississippi River from the area to the north and that of the Superior glacial lobe as well (Cooper 1935:34). Water from Glacial Lake Grantsburg flowed around the edge of the ice and through the Dalles, shifting its course as the ice retreated. Wright and Rubin (1956) date organic matter in the bottom of kettles and in the Anoka sand plain associated with the Grantsburg sublobe as 11,800-12,700 years old. Zoltai (1965:268) has Lake Agassiz draining into the Superior basin as recently as 8610-9530 years B.P. I suspect that some of the potholes are related to these late glacial events. However, how many times during the Woodfordian or Altonian substages glacial drainage affected the Dalles region seems to be speculation at this time.

Thus, much of the chronology of events for the Dalles must await detailed field studies in the area and the correlation of late Pleistocene events from Wisconsin to Lake Agassiz basin of North Dakota and Manitoba.

Because of the greater degree of development of potholes (Figs. 133, 134) on the Minnesota side of the Dalles, it would seem appropriate to include that area in the Reserve as well. There, individual potholes attain depths as much as 60-80 ft and are 15-25 ft in diameter. Compound potholes also are much larger and more varied than their counterparts on the Wisconsin side. In addition, Devils Chair (Fig. 135) is a rock monument produced since late Cary times by frost action and gravity movements in the well-jointed basalt. It may be seen on the cliff between the two highway overlooks to the south of the pothole area. It is analogous to the Devils Doorway (Fig. 63) at Devils Lake.

Conclusion

Although this note by its brevity may seem to do an injustice to the Interstate Park, this is purely coincidental. To me an Ice Age Reserve in Wisconsin would axiomatically include the St. Croix Dalles gorge and potholes. No other comparable area is known in the world.

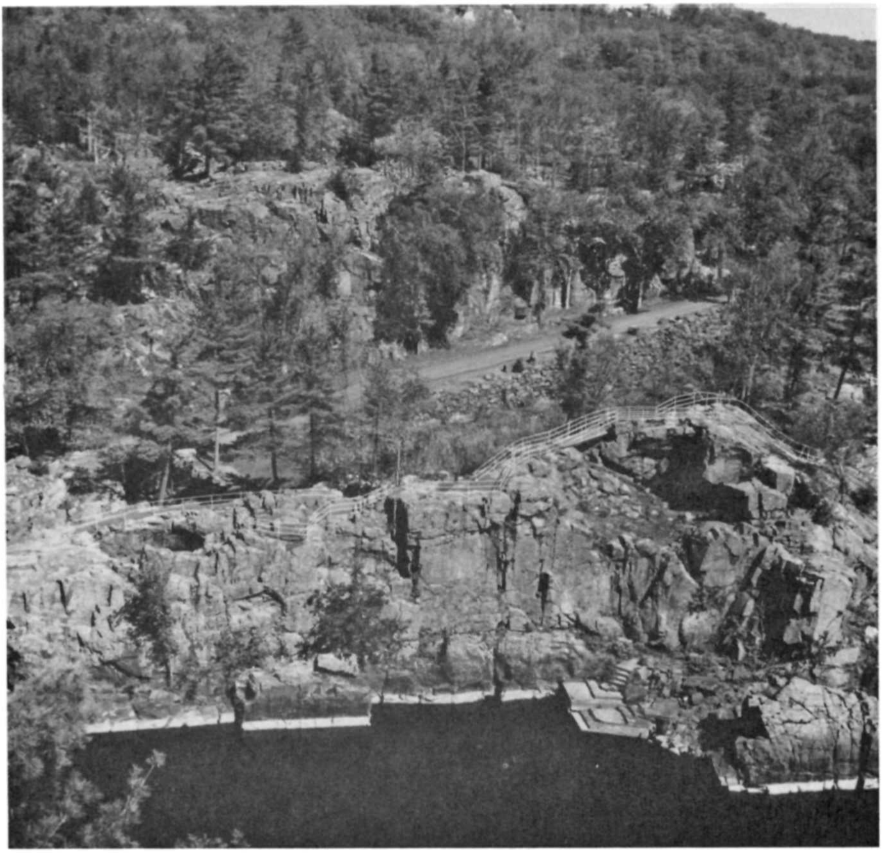

Fig. 133. Potholes on the Minnesota side of the Dalles, looking down from Summit Rock.

Fig. 134. Large pothole on the Minnesota side of the Dalles.

Fig. 135. Devils Chair, a post-glacial stack of loose joint blocks of the basalt on the Minnesota side of the Dalles.

11 Discussion of the Pleistocene of Wisconsin

For our purposes with respect to specific areas recommended for inclusion in the Reserve, we can do little with the history of events that preceded the Wisconsinan Stage. When our partial record is compared with the record of events in Illinois (Frye et al. 1965:43-61) for the Pleistocene as a whole and for the early Wisconsinan specifically, the differences are striking. We simply do not have recorded or have not recognized many events well displayed in Illinois. It seems likely that many fluctuations of ice margins in Illinois did not take place in Wisconsin or were so slight as to be inseparable; in other instances events recorded only by deposits had the record erased by subsequent glaciations. As Wisconsin does not have the complete record, we must turn to adjoining states for theirs in order to work out a composite sequence to which Wisconsin's record can be fitted. This has been done frequently since the concept of former widespread glaciation was established many decades ago. Unfortunately, in spite of the multiplicity of chronologies proposed, we cannot yet agree on a standard one for the Upper Mississippi Valley (Black and Reed 1965). The official chronology for Illinois by Frye and Willman (1960) (Table 1) differs markedly from the classical or standard rendition as recently presented in detail by Leighton (1960) (Table 1). The former attempts to bring order out of a proliferation of names and to correct discrepancies. that developed as new data were acquired. The latter retains the classical names, introduces some more, and perpetuates some correlations and usages that should be changed. Names from both have been used as needed for historical reasons or emphasis on the local situation. For additional comments on those classifications see Wright (1964).

Part of the difficulty of correlating events from one state to another is our inability to trace deposits or other indicators of events from one point to another across major breaks. We must then turn to other means of correlation such as similarity of features, materials, or pattern of events or to precise means of dating. Fossils are locally helpful, as in loess deposits, and widespread paleosols are traditionally one of the most important, but Wisconsin lacks them in any mappable deposits of consequence. Radioactive carbon has been especially useful in recent years (Libby 1961). Although not without problems, results from carbon dating can be duplicated and generally are consistent when newer techniques are used. Surely it has done more in recent years to point out discrepancies in our former chronologies and to place a truer time span on events than was remotely possible before its development.

When we examine the dates now available from Wisconsin, we see they fall into natural groups (Table 2) that may be interpreted in different ways. For example, of the four available dates greater than 33,000 radiocarbon years, the two from the vicinity of Marshfield in Wood County are of finely disseminated organic matter in silty clay on bedrock and beneath a single drift sheet that is surely Wisconsinan in age (Hole 1943). The date of more than 45,000 radiocarbon years may be interpreted to mean that the fluctuations of the Altonian ice in Illinois (Table 1) (Frye et al. 1965:43-61) were not represented in central Wisconsin from the time of existence of the pond to the advance of the ice that left the overlying till. The same interpretation is possible for the situation in St. Croix County. There the basal till with erratic wood dated at 29,000 and 30,650 radiocarbon years also seems to have incorporated peat in former ponds that is dated at greater than 45,000 radiocarbon years. The wood is thought to date the time the ice advance destroyed the spruce forest of the area; the older peat which now overlies the younger wood is thought to represent pond fillings on the surface overrun by the ice. If the different kinds of organic matter were transported by the ice only once, this would imply the area was ice-free from more than 45,000 years ago until about 31,000 years B.P. Obviously other interpretations are possible, pointing up the kinds of situations one encounters. Nonetheless, a similar situation in the same general time span seems to have existed in Ontario (Dreimanis and Vogel 1965:782-791) but apparently not in Illinois (Kempton 1963). Similarly, the spruce and willow fragments from Polk County are more than 38,000 radiocarbon years, but they tell us little about the chronology of the area. The 175-180 ft of drift overlying is so poorly recorded in the well records that almost any interpretation is possible.

Table 2. Some radiocarbon dates from Wisconsin.[a]

Age	Laboratory number	County	Location	Remarks
Post-Valderan				
4800 ± 150	L-605	Bayfield	46°57′N 91°00′W	Sand Island. Compressed peat under 14 ft of sand in water depth of 40 ft. Equivalent to C-504 at 3656 yrs. Dates low water stage of Lake Superior.
4880 ± 190	Y-238	Marinette	47°5′N 87°38′W	Sewer trench, 600 ft or less in elevation. White oak. Dates maximum or recession of Lake Nipissing.
6040 ± 350	W-1139	Columbia	SW¼ SW¼ sec. 7, T. 12 N., R. 8 E.	Driftwood at 7 ft in Wisconsin River alluvium. With W-1138, dates rapid filling of alluvium of the Wisconsin River.
6070 ± 320	W-1138	Columbia	SW¼ SE¼ sec. 7, T. 12 N., R. 8 E.	Beaver--cut stump *in situ* at 20 ft in Wisconsin River alluvium. See W-1139.
6340 ± 300	W-1017	Kenosha	42°33′N 87°49′W	Red oak. Dates paleosol at present lake level, under dune sand.
7650 ± 250	SM-16	Vilas	46°9′N 89°37′W	Grassy Lake. Yel-brown-black gyttja in south center in 6-7 ft of water; lower part of 30 ft of sediment on sand. Total organic content is minimal date for organic accumulation in lake.

Table 2. (continued) Some radiocarbon dates from Wisconsin.

Age	Laboratory number	County	Location	Remarks
Twocreekan		Manitowoc		
10,400 ± 600	M-343		Type area	Inner part of log.
10,700 ± 600	M-342		Type area	Outer part of log.
10,877 ± 740	C-308		Type area	Spruce.
Twocreekan		Manitowoc		
11,097 ± 600	C-366		Type area	Peat from soil horizon.
11,130 ± 350	Y-227		Type area	Spruce from soil horizon. Equivalent to C-308, 365, 366, 536, 537, and W-42 and 83.
11,350 ± 120	W-42		Type area	Wood from soil horizon. Equivalent to Y-141.
11,410 ± 180	W-83		Type area	Wood from soil horizon. Equivalent to Y-227.
11,437 ± 770	C-365		Type area	Spruce root from soil horizon.
11,442 ± 640	C-537		Type area	Peat from soil horizon.
11,500 ± 300	W-698		Type area	Wood from soil horizon.
11,840	L-698C		Type area	Log in sediment below soil. Average of cellulose and lignin.
11,850 ± 100	L-607A		Type area	Wood in soil horizon. Cellulose age.
12,000 ± 400	A-79B		Type area	Wood.
12,150 ± 400	A-79A		Type area	Wood from soil horizon.
12,168 ± 1500	C-536		Type area	Spruce from soil horizon.
12,200 ± 400	W-670		Type area	Wood from soil horizon.
		Outagamie		
10,676 ± 750	C-630		At Kimberly Clark Paper Mill.	Tree stump at depth of 10 ft in varved deposits 25 ft thick, formed in front of Valders ice. Reworked.
Twocreekan		Outagamie		
10,700 ± 210	Tx-44		SE¼ sec. 28, T. 21 N., R. 17 E.	Spruce at depth of 14 ft below Lake Oskosh sediments. Equivalent to C-800.
10,856 ± 410	C-800		SE¼, sec. 28, T. 21 N., R. 17E.	Spruce at depth of 14 ft below Lake Oskosh sediments in Valders till. Reworked (average of 10,241 ± 650 and 11,471 ± 500).
11,640 ± 350	W-1110		SW¼ SE¼ sec. 19, T. 23 N., R. 19 E.	Tamarack in soil horizon.
11,790	L-607B		Menasha. 44°12′N 88°27′W	Wood in till. Average of cellulose and lignin.
11,790	L-698D		44°17′N 88°25′W	Spruce in Valders till. Average of cellulose and lignin.
11,840	L-698B		NE¼ SE¼ sec. 19, T. 23 N., R. 19 E.	Spruce in soil horizon. Average of cellulose and lignin.

Table 2. (continued) Some radiocarbon dates from Wisconsin.

Age	Laboratory number	County	Location	Remarks
11,900	L-698A		SW¼ SE¼ sec. 19, T. 23 N., R. 19 E.	Spruce in soil horizon. Average of cellulose and lignin.
		Brown		
11,140 ± 300	W-590		sec. 23, T. 24 N., R. 21 E.	Wood in Valders till.
11,940 ± 390	Y-147X		SW¼ SW¼ sec. 23, T. 24 N., R. 21 E.	Wood in Valders till.
		Winnebago		
11,280 ± 100	Y-488		44°14′N 88°27′W	Wood from varved clay of Greater Lake Oshkosh. Equivalent to C-419 at 6401 ± 230.
11,690 ± 370	Y-237		NE¼ SW¼ sec. 15, T. 20 N., R. 17 E.	Wood in red clay.
Twocreekan		Winnebago		
12,060 ± 700	W-1183		SW¼ NE¼ sec. 11, T. 20 N., R. 17 E.	Peat and spruce under till.
		Waushara		
10,420 ± 300	W-820		SE¼ NE¼ sec. 31, T. 19 N., R. 10 E.	Peat at base of kettle.
11,000 ± 400	UCLA-633		NE¼ SE¼ sec. 31, T. 20 N., R. 10 E.	Carbonate in gyttja at base of kettle (organic age estimated at 9900).
11,600 ± 300	UCLA-631		NE¼ SE¼ sec. 31, T. 20 N., R. 10 E.	Wood at base of kettle.
12,000 ± 500	W-641		SE¼ SW¼ sec. 25, T. 19 N., R. 13 E.	Peat at 2-4 ft under silt in last stage of Later Lake Oshkosh.
12,220 ± 250	W-762		NW¼ SE¼ sec. 10, T. 18 N., R. 12 E.	Peat at 4 ft under lake clays.
13,700 ± 300	UCLA-632		NE¼ SE¼ sec. 31, T. 20 N., R. 10 E.	Carbonate in marl, 4 ft above base of kettle (organic = 12,800 ± 400).
		Other Locations		
11,130 ± 600	W-1391	Jackson	SW¼ SE¼ sec. 17, T. 22 N., R. 5 W.	Wood in meander scar of Trempealeau Valley.
11,611 ± 600	M-812	Sauk	Raddatz rockshelter Sk5. 43°21′N 89°56′W	Charcoal in fire bed in stratum "R".

Table 2. (continued) Some radiocarbon dates from Wisconsin.

Age	Laboratory number	County	Location	Remarks
12,800 ± 220	WIS-48	Jefferson	NW¼ SW¼ sec. 9, T. 5 N., R. 15 E.	Spruce at base of peat mound that formed shortly after deglaciation.
Woodfordian 17,250 ± 300	GX-0457	Grant	NW¼ NE¼ sec. 26, T. 5 N., R. 2 W.	Caribou bone at base of 7 ft of loess in cave and on sandy gravel. H. Palmer, pers. comm.
19, 250 ± 350	GrN-3624	Grant	SE¼ SW¼ sec. 16, T. 1 N., R. 2 W.	Alkali soluble organic matter 3.7 m. below top of loess and 2.8 m. above base.
Farmdalian 24,800 ± 1100	Gro-2114	Grant	NW¼ NE¼ sec. 35, T. 2 N., R. 2 W.	Bulk soil sample at base of loess.
Rockian 29,000 ± 900	W-903	Walworth	NW¼ SE¼ sec. 1, T. 3 N., R. 16 E.	Spruce in overridden outwash.
29,000 ± 1000	W-747	St. Croix	NW¼ SW¼ sec. 6, T. 28 N., R. 17 W.	Spruce in dark-gray clayey till on bedrock.
29,300 ± 700	GrN-2907	Grant	SE¼ SW¼ sec. 16, T. 1 N., R. 2 W.	Spruce charcoal in paleosol at base of loess, on bedrock.
30,650 ± 1640	Y-572	St. Croix	Woodville. 44° 57′N 92° 18′W	Spruce in dark gray clayey till on bedrock. See W-1758.
30,800 ± 1000	W-901	Waukesha	NW¼ SE¼ sec. 36, T. 7 N., R. 19 E.	Spruce in overridden outwash.
Rockian 31,800 ± 1200	W-638	Walworth	SW¼ NE¼ sec. 17, T. 17 N., R. 1 E.	Spruce in till. Equivalent to M-936 at >30,000.
Pre-Rockian				
>33,000	W-1370	Wood	NW¼ sec. 2 T. 25 N., R. 3 E.	Organic matter in fine mud below till.
>38,000	W-1598	Polk	SW¼ SW¼ sec. 3, T. 36 N., R. 15 W.	Spruce and willow at depth of 175-180 ft.
>45,000	Nucl. Sci. & Eng. Corp.	Wood	NW¼ NE¼ sec. 22, T. 25 N., R. 3 E.	Organic matter in fine mud below till. M. T. Beatty and F. D. Hole, pers. comm.
>45,000	W-1758	St. Croix	Woodville. 44° 57′N 92° 18′W	Peat in transported pond filling above Y-572.

[a] Excludes younger archeologic dates, those samples well above the bottom of lake deposits, and some solid carbon dates of doubtful validity. Letter prefix denotes laboratory where sample was run. See various issues of *Radiocarbon* for further details on individual samples.

However, the three dates (Table 2) of 29,000, 30,800, and 31,800 radiocarbon years from spruce in drift of Walworth and Waukesha counties, the two comparable dates of spruce from St. Croix County, and the comparable date of spruce of the basal loess in Grant County are believed to represent the time of a brief ice advance, called Rockian by Black (1960, 1962) (Frye et al. 1965:43-61; Black et al. 1965:56-81) that occurred simultaneously from the Des Moines Lobe on the west and from the Lake Michigan Lobe on the east. This time is latest Altonian (Table 1) and is recorded also outside Wisconsin (White and Totten 1965). The wood in St. Croix County is in the basal till which is rich in disseminated organic matter, clay, and residual chert (Black 1959a); the wood in Walworth and Waukesha counties is in oxidized sandy till and overridden gravelly outwash. All the wood is erratic and could have been picked up and transported more than once by ice or water, so clearly other interpretations than mine are possible. Nonetheless, if the dates are correct, the deposits can only be younger—not older.

Rockian ice from the two lobes joined in the center of the state. How far the Rockian ice went into the Driftless Area of southwest Wisconsin is not known for certain; it may have covered all of it. At least the Driftless Area has been glaciated (Black 1960), and disagreement is concerned more with timing—Frye and Willman (Frye et al. 1965:43-61) suggest the igneous erratics in the Driftless Area of northwest Illinois may be Nebraskan in age. Positive evidence of glaciation (Frye et al. 1965:43-61) comes from some fragments of Precambrian igneous and metamorphic rocks and particularly Paleozoic chert and sandstone (Akers 1964) that rest on younger formations. Erratics of sedimentary rocks are especially abundant in the central and northern parts of the area (Akers 1964). Sparse igneous erratics occur in isolated kame-like deposits south of Taylor in the northern part of the area and in fresh gravel on the upland beneath thick loess at Hazel Green. Igneous and metamorphic rock erratics are also found tens of feet above the Wisconsin River, as near Muscoda, and are associated with large blocks of dolomite transported several miles from possible sources. These deposits have structures and sorting typical of ice-contact deposits. Large sand bodies in the Kickapoo River Valley have come off dolomite uplands and have glauconite (complex iron-magnesium silicate) above any known source. Anomalous rubble deposits on the upland (Akers 1964) also have anomalous clay minerals (Akers 1961). Thus, in an area of 10,000 mile^2 in southwest Wisconsin, we see an absence or paucity of chert and clay residuum on bedrock, an absence or paucity of loess older than 29,000-30,000 years, and an almost complete absence of older paleosols (Hogan and Beatty 1963).[1] Moreover, shale with thin seams of unweathered dolomite (Maquoketa Formation) caps East Blue Mound, with only small fragments of the silicified Niagaran dolomite scattered on the broad flat upland (Black et al. 1965:56-81). This is an incongruous situation as supposed peneplains lie below it. Ice seems the only logical agent capable of removing this material out of the area. These suggest that most of the Driftless Area was covered by the Rockian ice (Black et al. 1965:56-81). A pre-Wisconsinan age for some of the peculiar features or deposits in the area can neither be confirmed nor denied (Akers 1964; Black et al. 1965:56-81; Frye et al. 1965:43-61; Palmquist 1965); it is suggested by weathering phenomena.[2]

Dates within the time of the Farmdalian deglaciation, which is recorded so well in Illinois (Frye et al. 1965:43-61), are limited in Wisconsin to one at the base of the loess of the Driftless Area. It does not record a break in the loess sequence —in fact none of significance from the dated paleosol at the base to the present surface has been found (Glenn et al. 1960:63-83; Hogan and Beatty 1963). Farmdalian and Woodfordian time in Wisconsin was at least partly a time of very cold climates and accompanying permafrost and periglacial phenomena (Black 1964a, 1965). No trace of trees has been found in Wisconsin from Rockian time to Twocreekan time which is dated about 11,000-12,500 years B.P.

Woodfordian time is represented in Wisconsin only by two dates in the Driftless Area. One, of caribou bone, is 17,250 radiocarbon years (H. Palmer pers. comm.). The other is a bulk sample of loess. Their significance and relationship to the prominent late Woodfordian (Cary) front or the chronology of glacial events are not known. Drift of middle and late Woodfordian age makes up the surface of more of the state than any other, yet isochronous boundaries (Alden 1918) at the front or within the drift sheet are exceedingly tenuous. Woodfordian time in Illinois is represented by tens of moraines and numerous radiocarbon dates (Frye et al. 1965:43-61). Clearly the Woodfordian is multiple, that is, it is composed of many pulsations of the ice front some with only limited movement but others with retreats or advances (in Illinois) up to 100 miles. The outermost Cary of presumed late Woodfordian age is not represented everywhere in either Wisconsin or Illinois by the same pulse. Although its border from the Plains to the Atlantic Ocean has been described and mapped for decades as the break between deposits of the First

[1] I have recently found Altonian and Sangamonian paleosols in a small area of loess near Hillsboro.

[2] See Univ. Wis. Geol. and Nat. Hist. Surv. Info. Circ. No. 15 for more recent information.

and Second Glacial Epochs (Chamberlin 1878, 1883a, 1883b:261-298), we still have much to learn about it. Without a single radiocarbon date related to the advance of that 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 lobe (Oakes 1960); lithology helps to distinguish major lobes (Anderson 1957).

Post-Cary or latest Woodfordian events which are pre-Twocreekan are much less well known in Wisconsin than elsewhere. Moraines assigned to Mankato and Port Huron in Minnesota and Michigan, for example, are presumed to be present in Wisconsin behind the Cary front. However, the correlation of moraines in Wisconsin with type localities has not been done, and deployment of such ice in the state is conjectural.

Deglaciation of the Woodfordian ice in Wisconsin may be time transgressive—being slightly earlier in the south than in the north. However, the available radiocarbon dates tend to negate this. A peat mound on Cary drift in Jefferson County has spruce at the base dated at 12,800 ± 200 radiocarbon years (Ciolkosz 1965). In Waushara County a date of 12,800 ± 400 years B.P. was obtained on organic matter in marly gyttja (fine compact, organic-rich detritus) 4 ft above the base of undisturbed marsh deposits (Park 1965:8). Spruce at the base of the same deposit and higher on the flank of the kettle was dated at 11,600 ± 300 years B.P. Three other dates on peat in basal pond deposits in Waushara County are 10,420 ± 300, 12,000 ± 500, and 12,220 ± 250 years B. P. One in Winnebago County is 12,060 ± 700 years B. P. The main evidence for the time transgressive deglaciation is morphologic—that is the widespread evidence of ice stagnation in the north particularly and the more youthful lakes and other features in the north. However, the time difference may be several thousand years for all buried ice to melt out (Black 1962).

The Twocreekan interval is named from Two Creeks, where a buried soil and organic remains were recognized in lacustrine deposits along the exposed bluff of Lake Michigan (Goldthwait 1907). This is the best dated interval in Wisconsin, the latest dates yielding an average of 11,850 years B.P. (Broecker and Farrand 1963). A number of dates (Thwaites and Bertrand 1957) derived by the original solid-carbon method were as much as several thousands of years in error according to reruns by better methods. The general range of Two Creeks time from 11,000 to 12,500 years proposed by Frye and Willman (1960) seems distinctly longer than the interval represented at Two

Creeks. There, only an incipient soil profile was formed under trees in which the oldest by tree-ring count was only 142 years (Wilson 1932, 1936). Several other localities in east-central Wisconsin contain the Two Creeks horizon *in situ* , and logs from it are incorporated in the overlying Valderan till. These also tend to cluster close to 11,800 years B. P. so the span of Twocreekan time in central and northern Wisconsin probably is considerably less than in Illinois. This is to be expected, for deglaciation through several hundred miles of latitude of an ice lobe the size of that which occupied Lake Michigan cannot be accomplished over night.

The land surface of Wisconsin generally was exposed during Twocreekan time and was covered by the subsequent Valderan ice advance only in northeastern Wisconsin (Black 1966). It was named from the Valders quarry where relations were established by Thwaites (1943). Consequently, over most of the state the effects of Twocreekan soil formation and geomorphic processes were merged and obliterated by the same processes that continued down to the present day in all but rare situations. One such is the deposits in the rock shelter under the Natural Bridge in Sauk County (Black 1959b; Wittry 1959). Twocreekan deposits were recognized in the shelter from an interpretation of the geology and confirmed by radiocarbon dating. Man was associated with the shelter, leaving his wood fires for dating purposes. The climate in northeastern Wisconsin at the time was perhaps similar to that of today in northern Minnesota (Roy 1964).

Distribution of the Valderan ice is only now being reevaluated (Black 1966). Whereas it was formerly thought to extend across northern Wisconsin (Leverett 1929, 1932) and correlate with red clayey till in eastern Minnesota, this is clearly incorrect (Wright and Ruhe 1965:29-41). Unfortunately we have no radiocarbon dates in Wisconsin directly reflecting either its rate of advance or retreat. Valderan ice occupied the eastern part of Lake Superior and the northern part of Lake Michigan. Parts of both those lakes must have been open water from the latter part of Woodfordian time to the present. This interpretation differs somewhat from that of Hough (1958), but other differences are also appearing (Bretz 1964, 1966; Hough 1966). Unquestionably the full history of the Great Lakes is complicated and beyond the scope of this book. It seems that Valderan ice advanced to the vicinity of Milwaukee[3] after having retreated north of the Straits of Mackinac during Twocreekan time. However, only the Valderan ice of the Lake Michigan Lobe entered the State. It possibly reached its climax about 10,500 years B.P. and by 9500 years B.P. was largely gone from the state.

[3]*This is now being questioned.*

The close of the Valderan substage has been accepted as the close of the Pleistocene and assigned arbitrarily at about 5000 years B.P. when sea level returned to approximately its present level (Frye and Willman 1960), but others draw the boundary at 10,000 years ago and elsewhere. Wright (1964) would prefer to use the pollen-zone boundary that marks the end of the boreal spruce forest. He would define the boundary in dated pollen-bearing lake sediments resting directly on Valders drift near the type locality of that drift in Wisconsin. Such studies unfortunately have not been made at the type locality although some have been made nearby (West 1961; Schumacher 1966). Broecker et al. (1960) would use the abrupt change in several climatic indicators at about 11,000 years B.P. as the boundary between the Pleistocene and Recent, but this is obviously complicated because of the subsequent Valderan glaciation in the north-central United States. Clearly we have as little agreement on the termination of the Pleistocene as on its beginning although the differences in time involved are markedly less. Agreement on the time of termination is not likely to be reached immediately. It is further complicated, for example in Canada, by the splitting up of the continental ice sheet into several masses that continued to fluctuate independently of each other (Zoltai 1965) for some time after the Valderan ice disappeared from Wisconsin.

After the Valderan glaciation fluctuating climates are recorded in pollen sequences in Wisconsin (West 1961; Schumacher 1966), in physical changes such as the rapid filling of part of the Wisconsin River Valley near Portage about 6000 radiocarbon years B.P. (Frye et al. 1965:43-61), or in other drainage changes (Dury 1964, 1965; Palmquist 1965). These and other facets like faunal and floral changes are beyond the scope of the present discussion.

The full story of the Pleistocene is a fascinating one, yet so long and involved that a single textbook can no longer do justice to it. We have come a long way since the last summary of the glacial geology of the central states was presented (Alden 1932). The new INQUA volume (Wright and Frey 1965) will stand for years as the best summation for the entire United States yet many portions of it are already outdated.

12 Other Outstanding Areas in Wisconsin

This book would not be complete without reference to a few other areas in Wisconsin that are worthy of being included in the Ice Age Reserve. Because of the limit on available funds particularly, but in part because of the bounds of practicality in administering and using some remote areas, not all desirable features are included in the nine areas previously discussed. Occasional reference to similar nearby features has been made in the text in connection with some of the recommended areas. These are not repeated here. Only brief mention of some of the other most important areas and the kinds of things they show or contain is given.

Without question by far the most important locality not now being included in the Reserve, as agreed upon, is the Natural Bridge and underlying Rock Shelter, Sk5, in Sauk County (Wittry 1959; Black 1959b; Parmalee 1959). I believe very strongly that this site should be in a scientific reserve now, and certainly should be in the proposed Ice Age Reserve. It is a unique locality that is in imminent danger of being irreparably damaged or destroyed. The bridge is part of a narrow spur of Upper Cambrian sandstone which rises 35 ft above a small picturesque glen. The opening under the arch is 15 ft high and is directly above the sandstone capping the shelter. Archeologic excavations and geologic study of the area establish man's presence at the close of the Valderan substage and suggest his presence during Twocreekan time. The rockshelter and bridge were carved by a combination of weathering and erosional processes shortly before occupancy by man. In the upper part of the deposits were hundreds of artifacts of chipped and ground chert, basalt, bone, teeth, and shell. Several hundred pounds of split deer bones and other faunal remains, hundreds of firebeds, and several artificial pits attest to man's presence. Three loess horizons and involutions are common to older strata in the shelter.

Because we still lack a firm chronology to which to correlate the strata in the shelter and it by itself is not sufficient to establish a detailed chronology, half of the deposits in the shelter were left undisturbed. These are in constant danger of being destroyed. This would be tragic, for this is the oldest authenticated site for man in the Upper Mississippi Valley. Inclusion of this site in the Reserve would preserve it and provide a fitting location to bring man into the story of the Pleistocene. The bridge and surroundings are a beautiful natural attraction; the shelter and its story are a scientific must.

Nowhere in the nine recommended areas are there good examples of glacial striae on bedrock. Weathered striae are present on the higher bedrock ridges at the St. Croix Dalles. Some unweathered examples might be buried there that could be included in that park. The deepest and commonly most striking glacial striae and polish are those on dolomite or limestone, but to be preserved they must have been under several feet of soil since they were made. This means that only recent excavations will show them, for solution by rain destroys them in a few years. Once exposed they should be protected from the elements if they are to be preserved with all their original gloss. The Silurian dolomite (Fig. 3) is a logical place to look for them, and the escarpment in the vicinity of the Northern Kettle Interlobate Moraine should provide an excellent location. The Valders quarry, the type locality of the Valders drift, lies only 20 miles north-northeast of the present Northern Kettle Moraine State Forest and displays striae at right angles which can be correlated with two distinct tills, Valders and Cary. Regrettably, quarrying operations are fast removing them. Sites with two sets of striae probably are available closer to the forest, but I made no search for them.

Fossil periglacial phenomena are widespread in Wisconsin, but good sites are rare and almost inevitably are destroyed quickly. Most are found in artificial excavations, such as road cuts, borrow pits, and the like and are quickly covered over. Of the various kinds of features (Black 1964a) the most striking and important as paleoclimatologic indicators are ice-wedge casts (Black 1965). Sites relatively close to Mill Bluff or Bloomer have been found, and hopefully one at least could be made suitable for presentation to the public. These are in unconsolidated material and require protection from the elements to survive. Because they are so scarce and difficult to preserve, at least one site should be set aside for the future.

The Great Lakes history is a complicated and fascinating part of the Pleistocene. Part of the evidence is demonstrated at Two Creeks in the form of deposits, but no actual beaches are preserved there. Shorelines and shoreline phenomena are recorded at several places in southeastern Wisconsin, near or on the present shoreline of Lake Michigan. A swamp deposit with many kinds of wood dating about 6340 years B.P. is found at water level under dune deposits at the south city limits of Kenosha. The Glenwood and Calumet beaches can be traced almost continuously along that shore from the Illinois State line northward to Windy Point, north of Racine (Goldthwait 1907). Superb beaches may be found from present lake level to a height of several hundred feet above Lake Superior, between Superior and the Michigan border. Some can be correlated across or around the Bayfield Peninsula; higher, older beaches cannot be traced across, suggesting that individual lakes were established on east and west sides that were independent of each other. The Lake Superior beaches are independent of the Lake Michigan beaches and a site on either or both might be desirable.

Although occasional fragments of bone or teeth of Pleistocene mastodon or mammoth continue to appear, no complete skeletons have been found for many decades in Wisconsin. A reconstructed skeleton of a mastodon is mounted in the small museum of the Geology Department of the University of Wisconsin, Madison. That specimen, almost completely intact, came from the Wisconsin River deposits near Boaz, Richland County. Such a skeleton at a museum at the Natural Bridge and Rock Shelter would be most appropriate. Specimens of beaver-cut wood, such as have been found in the Wisconsin River deposits near Portage and dated at 6070 radiocarbon years B.P., could be added to such a museum along with other "finds."

West Blue Mound is already a state park, and East Blue Mound has a small county park on its north side. These two mounds provide an excellent view of the "Driftless Area", contain abundant block fields of chert rubble considered to have moved under periglacial climates, and provide us with the problem of explaining how soft shale and thin seams of dolomite of the Silurian formations (Fig. 3) can cap an upland surface which is above several supposed peneplains (Black et al. 1965:56-81). Here is a natural setup for examining some of our basic tenets in geomorphology as affected by the various forms of weathering and erosion during the Pleistocene.[1]

Last but not least of interesting places especially singled out for mention here is the Wisconsin Dells. Now almost 100% commercialized, it is out of the question to purchase it. Nonetheless, as an example of drainage diversions produced during the Ice Age it is excellent (Martin 1932:345-353; Powers 1946).

[1]See Univ. Wis. Geol. and Nat. Hist. Sur. Info. Circ. No. 15.

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