H. E. WRIGHT
DAVID G. FREY

The Quaternary of the United States

A Review Volume for the VII Congress of the International Association for Quaternary Research

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THE QUATERNARY OF THE UNITED STATES

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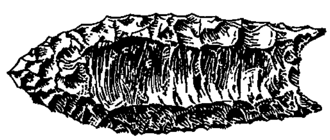
A REVIEW VOLUME FOR THE VII CONGRESS OF THE INTERNATIONAL ASSOCIATION FOR QUATERNARY RESEARCH

H. E. WRIGHT, JR. AND DAVID G. FREY, EDITORS



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PREFACE

THE QUATERNARY PERIOD of geologic time encompasses the last ice age proper (Pleistocene Epoch) and subsequent time (Holocene or Recent Epoch). The Quaternary is unique among the geologic periods for the relative perfection of its stratigraphic record—and thus for the unmatched opportunity it affords to decipher historical details with an accuracy impossible for earlier periods. The framework of the continents even in early-Quaternary time resembled that of the present so closely that major relations of land and sea and their effects on the general circulation of the atmosphere were probably very like those of the present, except perhaps when world-wide lowering of sea level exposed continental shelves or created significant land bridges between continents and arctic islands. Reconstructions of past conditions therefore may be controlled by a geographic framework relatively easy to visualize.

The dominating circumstance during the Quaternary in many parts of the world was climatic change. It resulted not only in the expansion and recession of glaciers, desert lakes, and coastal seas, but it also affected the composition and distribution of plant and animal communities, creating conditions potentially favorable for gene flow, extinctions, disjunctions, and other biogeographic processes. Thus not only the physical but also the biological events are complicated, and all have left undeciphered records which challenge the perspicacity and ingenuity of scientists in diverse fields. Investigators are repeatedly faced with the need to trade ideas and techniques in order to understand the larger environmental aspects of their particular problems. Not a few Quaternary scientists claim competence in at least two major disciplines, such as geomorphology and prehistory, glacial geology and pollen analysis, zoology and stratigraphic paleontology, geochemistry and paleolimnology, climatology and tree-ring analysis. Major advances in any one field are bound to be important in others because they modify disciplinary interpretations and lead to a progressively more satisfactory synergistic understanding of the entire period. Our knowledge of the Quaternary is advancing so rapidly that this book, even with the exciting findings it details, is scarcely more than a stocktaking or a progress report, although it should prove of great value in consolidating the field and in indicating critical areas for interdisciplinary attack.

Quaternary studies are best approached through the principle of uniformitarianism—that the present is the key to the past. This principle can guide the reconstruction of past conditions within the framework of physical, chemical, and biological laws, which seem constant. It can be studied in operation at the present time in many situations throughout the world, but difficulties of interpretation arise regarding Pleistocene conditions that have no analogues anywhere in the world today, such as the regime of a large ice sheet in temperate latitudes or the vegetation on a large, newly deglaciated area in a temperate region far from sources of seeds of the major forest trees.

The interplay between the present and the immediate past provides continuing fascination. A great many scientists who study modern natural processes are challenged by the opportunity to extrapolate their findings by applying them to a bit of history. At the same time, those concerned with historical sequences are always in quest of new and sharper methods for reconstructing past events. Thus the Quaternary provides a focus of cooperation among not only those with different competencies but also those with different approaches.

An example may be found in the field of oceanography, in which the recent development of deep-sea coring devices and of isotope analysis has provided the materials and techniques for the study of what may be the only continuous sediment record for the entire Quaternary. The geochemical, sedimentological, and paleontological variations in these sediments have implications for paleoclimatology. Concern for these implications has stimulated oceanographers to learn about the more diversified terrestrial record of Quaternary paleoclimates for the purpose of correlation. In the exchange of ideas, the horizons of the land-oriented Quaternary scientists have been enlarged, and overall understanding of the Quaternary has been increased.

Quaternary time saw the evolution of modern man and the development of human cultures. Interest in the chronology of these events and in the environmental conditions that accompanied and may have influenced them has led to an increasing number of joint projects in which biological, geological, and geochemical techniques are applied to archaeological problems. The most spectacular of these is radiocarbon dating, but detailed studies of the paleontology, sedimentology, and geochemistry of archaeological sites are also productive in elucidating the paleoecology of Early Man.

The present volume is an attempt to review the status of investigations into many facets of the Quaternary of the United States—a large area with diverse geographic regions and diverse Quaternary histories. In a certain sense Quaternary studies are of two major types. First are descriptive analyses of the morphology and stratigraphy of geologic or archaeologic materials, *e.g.* the mapping of moraines or terraces or caves and the study of the physical and paleontologic stratigraphy of the related deposits. These studies form the backbone of Quaternary work by providing the necessary maps and materials from which the environments can be reconstructed or the sequence of events worked out. Then there are investigations of current natural processes that have some known or potential historical implications or applications. Here lie a great group of studies ranging from glaciology and tererstrial magnetism to floristic phytogeography.

After an introductory chapter on early American studies, the book is organized in four principal sections: geology, biogeography, archaeology, and miscellany. The geologic chapters are arranged on a geographic basis; they are largely inventories of the present knowledge of the Quaternary history and geological processes, first in the area of continental glaciation, then in the unglaciated areas east of the Rocky Mountains, and finally in the western part of the country. They show the diverse methods of investigating the complex problems presented by the morphology, lithology, and stratigraphy of surficial sediments: field mapping, subsurface exploration, followed by detailed laboratory studies ranging from clay-mineral analysis to micropaleontology and isotope dating.

The chapters under the heading of biogeography (in its broad sense) include six on regional phytogeography and pollen analysis. Because these chiefly involve only the flowering plants, a separate chapter deals with the mosses. Another concerns the special problem of polyploidy as it relates to climatic history. The zoogeographic chapters are arranged on a taxonomic rather than a geographic basis, ranging from mammals to invertebrates. A special chapter treats recent adjustments in animal ranges. In a concluding essay, E. S. Deevey reflects on the nature of Quaternary biogeography.

Archaeology is treated principally in five regional chapters. These reviews emphasize the paleoecology rather than the typology of the cultures, or they deal with archaeological features that have some geological context. Most are concerned with the earlier phases of human activity in this country rather than with those of the last few millennia.

The miscellaneous chapters that conclude the volume concern diverse subjects that have new or interesting applications to Quaternary history. Although many Quaternary scientists are occupied with studying the record of geologic or biogeographic changes that record climatic fluctuations, processes other than climatic change left important records in Quaternary rocks and landforms. Thus various manifestations of crustal activity are described in chapters on tectonics, volcanic-ash deposits, and paleomagnetism. Other chapters in this section are concerned with processes whose historical implications have only recently been appreciated —isotope geochemistry, limnology, pedology, oceanography —and still others with familiar subjects in which newly developed knowledge encourages fresh approaches, such as theoretical paleoclimatology, glaciology, paleohydrology, and tree-ring analysis.

Many gaps remain in this coverage of the Quaternary of the United States, either because suitable authors were not available or because of limitations on the size of the book. Some authors found it convenient to include consideration of Canada and Mexico, but generally the coverage is limited to the continental United States (and its continental shelves). In any case, the diversity of authors and subject matter provides a broad sample of American scientists and disciplines involved in the study of the Quaternary in all of its manifestations.

Publication of this volume has been made possible by a grant from the National Science Foundation, made available for this purpose through the good offices of the National Academy of Sciences. The editors are grateful on behalf of the Organizing Committee of the 7th INQUA Congress for these contributions of assistance.

March 3, 1965

H. E. Wright, Jr.
David G. Frey
editors

CONTENTS

THE QUATERNARY OF THE UNITED STATES

INTRODUCTION: HISTORICAL PERSPECTIVES

RICHARD FOSTER FLINT[1]

WE HAVE before us a volume, the first of its kind ever to be published, in which the very considerable knowledge of the post-Tertiary geology of the United States has been condensed and interpreted, by regions, by a group of specialists. It is clearly the latest word on the subject, constituting a reference book that will be used widely and continuously. Eventually it will be superseded by a newer symposium of the same sort, which will distill the additions that are made each year to the description and interpretation of the multitude of stratigraphic units and other features that together constitute the Quaternary geology of the subcontinent.

The acceleration in the growth of knowledge since about 1945 has produced so much information about this geology that the literature has become bulky and in some respects poorly accessible. Hence this volume is timely. Probably something of its kind would have been put together under other auspices during the present decade, had not the VII INQUA Congress furnished the stimulus for a compendium in 1965.

Although the volume is a distillation of what is known today, it is interesting to examine it with reference to the history of thought about the surficial geology of the United States, and to realize when and by whom the major concepts inherent in the American Quaternary were put forth. This is the reason for beginning the volume with this introduction. It tells us nothing new, but it lends some perspective to modern research in the Quaternary System.

ORIGIN OF GLACIAL DRIFT

Although the Quaternary strata include sediments of many kinds, the kind that occupied by far the greatest attention during the 19th century, and that possibly still attracts major attention today, is glacial drift. The story of early interpretation of the origin of drift is mainly a European story. However, in the early 19th century, America too had its advocates of catastrophic floods, ocean currents, waves, winds, and other far-fetched agencies as explanations for the deposition of that mantle of diverse surficial sediment and included erratic boulders, which at first was known only in the eastern part of the country.

One of the earliest published statements about the drift in America is characteristic of its author, Benjamin Silliman (1821, p. 49), in being sturdy, objective, and cautious:

"The almost universal existence of rolled pebbles, and boulders of rock, not only on the margin of the oceans, seas, lakes, and rivers; but their existence, often in enormous quantities, in situations quite removed from large waters; inland,—in high banks, imbedded in strata, or scattered, occasionally, in profusion, on the face of almost every region, and sometimes on the tops and declivities of mountains, as well as in the vallies between them; their entire difference, in many cases, from the rocks in the country where they lie —rounded masses and pebbles of primitive rocks being deposited in secondary and alluvial regions, and vice versa; these and a multitude of similar facts have ever struck us as being among the most interesting of geological occurrences, and as being very inadequately accounted for by existing theories."

Undoubtedly much of the sediment described was glacial drift. Like nearly everyone else at that time, Silliman believed in the Deluge, but he did not attribute this sediment to the Deluge, as others did.

A long step forward is represented by the observation of Peter Dobson (1826, p. 217-218) of Vernon, Connecticut:

"I have had occasion to dig up a great number of bowlders, of red sandstone, and of the conglomerate kind, in erecting a cotton manufactory; and it was not uncommon to find them worn smooth on the under side, as if done by their having been dragged over rocks and gravelly earth, in one steady position. On examination, they exihibit scratches and furrows on the abraded part; and if among the minerals composing the rock, there happened to be pebbles of feldspar, or quartz, (which was not uncommon) they usually appeared not to be worn so much as the rest of the stone, preserving their more tender parts in a ridge, extending some inches. When several of these pebbles happen to be in one block, the preserved ridges were on the same side of the pebbles, so that it is easy to determine which part of the stone moved forward, in the act of wearing. . . .[2]

"These bowlders are found, not only on the surface, but I have discovered them a number of feet deep, in the earth, in the hard compound of clay, sand, and gravel. . . .

"I think we cannot account for these appearances, unless we call in the aid of ice along with water, and that they have been worn by being suspended and carried in ice, over rocks and earth, under water."

Although the material described is probably till, the description is accurate and the inference quite logical in view of the state of knowledge then existing.

The theory of continental glaciation originated in Europe, and its greatest exponent was Louis Agassiz. In 1837 Agassiz

[1] Department of Geology, Yale University, New Haven, Connecticut.

[2] Such features were described independently and correctly interpreted by Chamberlin (1877, p. 200).

read his epoch-making paper before the Helvetic Society in Luzern. Although the paper, entitled *Études sur les Glaciers* (Agassiz, 1840), was not published until three years later, the news spread to America sooner than that, for in 1839 we find Timothy Conrad, the paleontologist, writing in the *American Journal of Science* (Conrad, 1839, p. 241-242):

"M. Agassiz attributes the polished surfaces of the rocks in Switzerland to the agency of ice, and the 'diluvial scratches,' as they have been termed, to sand and pebbles which moving bodies of ice carried in their resistless course. In the same manner I would account for the polished surface of the rocks in Western New York. . . ."

After the publication of *Études sur les Glaciers,* Edward Hitchcock (1841, p. 247-258) likewise accepted the theory in principle and showed lucidly how it explained the observed characteristics and distribution of the American drift. Conrad's statement and Hitchcock's paper therefore mark the establishment, in America, of the concept of the origin of the drift through the agency of glacier ice. Notwithstanding, the following decades witnessed active debate, in which it was apparent that many people were still unconvinced. Agassiz himself arrived in America in 1846, and although his authoritative voice was added to those of Conrad and Hitchcock, the controversy went on and did not die out entirely until the end of the 19th century, a few years earlier than its disappearance in Europe.

Physical Features of Drift

Although "moraines" began to be spoken of by Americans soon after Agassiz' book was published, apparently the term was not clearly applied to a single well-described feature until 1871, when G. K. Gilbert (1871, p. 340-342) then geologist on the Ohio Geological Survey, wrote:

". . . the ridge which determines the courses of the St. Joseph and St. Marys rivers is a buried terminal moraine of the glacier that moved southwestward through the Maumee valley. . . . Its irregularly curved outline accords intimately with the configuration of the valley, and with the direction of the ice markings; its concavity is turned toward the source of motion. . . at every stage of its [the ice sheet's] existence—its margin must have been variously notched and lobed in conformity with the contour of the country, the higher lands being first laid bare by the encroaching secular summer."

This succinct and thoughtful description apparently refers to the Wabash and Fort Wayne Moraines of current nomenclature.

Six years later T. C. Chamberlin (1877, p. 204-215) published a description and true interpretation of an end moraine of a different kind. The moraine, then known as the Potash Kettle Range or Pots and Kettles Range, was a conspicuous ridge of lobate form in eastern Wisconsin. Characterized by an abundance of stratified drift and by kames, kettles, ice-channel fillings and other ice-contact features, it was accurately described by Chamberlin, who concluded: "The Kettle Range is evidently a gigantic moraine." It consists essentially of the Johnstown and Interlobate Moraines of current terminology. In his interpretation Chamberlin made use of the recognition by Whittlesey (1859) that glacial kettles are of ice-block origin. Whittlesey's interpretation was based on analogy with the termini of glaciers in the Alps, where ice masses mixed with drift had been reported.

Chamberlin recognized the moraine as compound, and correctly visualized the variations of glacial regimen that could create a nested sequence of end-moraine ridges.

It is noteworthy that at that early date Chamberlin suggested the piping of silt and fine sand as a possible origin for some of the small basins in the Kettle Range. The piping process, commonly related to loess, did not become widely recognized in surficial geology until the middle of the 20th century.

In this publication also, Chamberlin (1877, p. 218) first used the term *till,* taking it from the British literature. The term rapidly spread into general use in America.

In connection with the origin of kettles, we note that although their origin (as ice-contact features) was explained correctly in 1867, not until several years later did a correct explanation of the origin of kames appear. In 1873 N. H. Winchell (1873, p. 62) published a clear account of how such features were made, as a result of observations in Minnesota in the previous year, although he did not use the word kame. Similar interpretations were published in Sweden by Hummel (1874) and by Upham (1877) in New England, both apparently independently. Earlier, a common opinion had been that kames were of marine origin, at least in part (*cf.* Geikie, 1874, p. 246-252).

In his early treatise on glacial drift in New Hampshire, Hitchcock (1878, p. 181-215) gave an excellent description and interpretation of striations and other glacial marks on bedrock encountered in his survey. This anticipated by a decade the more detailed and comprehensive treatise by T. C. Chamberlin, mentioned on a later page.

By 1883 Chamberlin (1883, p. 295-309) had systematized and defined the chief kinds of glacial deposits and sediments of related origin. In this respect he was influenced, no doubt, by the publications on glacial geology by James Geikie in Britain. From this time onward the two scientists kept in touch with each other, as is evident from the mutual references and quotations in the publications of the two men.

Mapping the Drift Border

The decades of the 1860's and 1870's saw the launching of several systematic geological surveys, mostly under State authorities of various kinds, although some surveys began even earlier. Most of them dealt with the glacial drift in reconnaissance, and some dealt with it in detail. Some of them identified a border of the glacial drift and described its character and position. With these early studies the names of many men prominent in American geology were associated. Among them were G. H. Cook in New Jersey, W. W. Mather and Lardner Vanuxem in New York, G. F. Wright and H. C. Lewis in Pennsylvania, J. S. Newberry, Edward Orton, G. K. Gilbert, and Charles Whittlesey in

Ohio, John Collett in Indiana, L. C. Wooster and Alexander Winchell in Michigan, C. A. White in Iowa, T. C. Chamberlin in Wisconsin, Warren Upham in Minnesota, and J. E. Todd in South Dakota. As a compilation, the map published by Newberry (1874, p. 76) is outstanding. It shows a generalized drift border extending from New Jersey to Kansas and many striations in New England, east-central United States, and Canada. The accompanying text is informative and well reasoned.

In 1878 a compiled map with more extensive data was published by C. H. Hitchcock (1878, p. 323). In it the drift border is shown extending westward into Montana and eastward from Cape Cod along the continental shelf as far as Newfoundland. Hitchcock therefore understood that glaciation affected a region, now submerged, east of the Atlantic Coast. The text accompanying the map speaks of centers of glacial dispersion in Greenland, Labrador, and the Rocky Mountains.

Chamberlin, who had been appointed to the U.S. Geological Survey in 1881 and had spent much time in glacial studies, made a broader synthesis of a prominent drift border, extending from Cape Cod on the east to North Dakota on the west (Chamberlin, 1883, esp. Pl. 28). He had published a preliminary version of the map five years earlier (Chamberlin, 1878, p. 209). He called the border "a moraine" and identified it with a "Second Glacial Epoch" because in many sectors other drift lay south of it. Much though by no means all of Chamberlin's "moraine" coincides with the outer limit of the drift of Wisconsin age as we understand it today. The border of the region covered by older drift was later defined gradually, over a long period.

A few years later Chamberlin (1888, Pl. 8) published a compilation of glacial striations observed by many geologists at some 2,500 localities between the Atlantic Coast and Dakota Territory, and had extended the map of glacial-drift borders through the Cordilleran region right to the Pacific Coast. The map distinguishes between "Earlier Drift" and "Later Drift" more or less along the line separating the Wisconsin drift from pre-Wisconsin drifts of today's nomenclature. This publication, incidentally, was accompanied by an elegant treatise on striations and other glacially caused markings on bedrock.

Refinement of the outer limit of glaciation has continued up to the present time, but the 1888 map, for one of its small scale, is a good representation of the major glaciated areas of the United States.

Repeated Glaciation and Stratigraphic Subdivision

As early as 1874, J. S. Newberry (1874, p. 3) had identified in Ohio a "forest bed," a layer of plant matter, overlying glacial drift. Similar material was soon found to be widespread throughout the Middle West, where in places it lay between two sheets of till. McGee (1878, p. 341) correctly viewed this relationship as implying that the forest bed "must be of interglacial age," possibly in analogy with interglacial layers earlier recognized in Europe. At about the same time Chamberlin was recognizing the existence of two drift sheets differentiated by weathering of different degrees of intensity. Probably these observations constitute the beginnings of the concept of multiple glaciation in North America.

By the end of the 1880's many geologists were at work on the areal glacial geology and stratigraphy of a wide sector of northern United States. Among them were G. H. Stone (eskers of Maine), W. M. Davis (drumlins of southern New England), N. S. Shaler (Mt. Desert, Nantucket, and Martha's Vineyard Islands), G. F. Wright (drift border in Ohio, Indiana, and Illinois), J. C. Branner (areal studies in Ohio and Indiana), Frank Leverett (northeastern Illinois), R. D. Salisbury (Driftless Area), I. M. Buell (boulder trains in Wisconsin), and J. E. Todd (areal studies in the Dakotas, Nebraska, and Iowa). Chamberlin acted as official and unofficial coordinator and synthesizer of much of this work, as is reflected in his publications dating from around that time.

Shortly afterward the practice of applying geographic names to drift sheets was introduced in America. The Third Edition of Geikie's *Great Ice Age* (Geikie, 1894) contains two chapters on the glacial features of North America contributed by Chamberlin; in them the Kansan, East-Iowan, and East-Wisconsin stages of glaciation, separated by unnamed intervals of deglaciation, were introduced for the first time. The name Kansan was applied to the drift subsequently called Nebraskan, and the names East-Iowan and East-Wisconsin were soon shortened to Iowan and Wisconsin as used in subsequent literature. The general scheme, begun by Chamberlin, of classifying the American Pleistocene into glacial and interglacial stages endured for many decades without modification and still, with some changes, constitutes an important element in the classification generally favored today.

It is interesting to contrast the stratigraphic classification that developed in New England and adjacent regions with that which evolved in the Middle West. The two are different, and the differences are based on fundamental differences between the physical-geologic conditions in the two regions. As J. D. Dana (1873, p. 210) remarked, "No distinct terminal moraines . . . have been observed in New England." All the earlier New England reports emphasize (1) the presence of large quantities of stratified drift with diverse morphology, (2) a post-drift marine overlap in extensive coastal areas, and (3) the widespread existence of terraces in stream valleys. Two or more glacial sequences in stratigraphic superposition were not found in the region before the 20th century.

In contrast, the earlier Middle Western reports emphasize end moraines, till sheets in superposition, interglacial layers with fossil plants and mammals, loess, and sediments and shorelines of the glacial Great Lakes, altogether a more complex and more nearly complete sequence.

The evolution of the sequence in New England is well illustrated in the classifications shown in three editions of Dana's *Manual of Geology*, a standard textbook, between 1863 and 1895 (Table 1). Although stratigraphic names of Middle-Western origin have long since been introduced into the New England sequence, the names Champlain and Recent are still used, although with some differences of meaning.

TABLE 1

Stratigraphic Sequences Shown in Three Editions of the *Manual of Geology* by J. D. Dana

Dana, 1863, p. 535-586.	Dana, 1875, p. 527ff.	Dana, 1895, p. 940ff.
Age of Man, with deposits of modern origin: alluvium, deltas, beaches, sand dunes.		
Post-Tertiary: Recent or Terrace epoch, marked by upwarping, trenching, and terrace cutting by streams.	Quaternary: Recent or Terrace Period	Quaternary era or Era of Man: Recent Period
Champlain epoch, with alluvial, lacustrine, and marine sediments, the latter representing marine invasion of coastal regions.	Champlain Period (The earlier part was the era of melting of the great glacier.)	Champlain Period } = Pleistocene of Lyell
Glacial epoch, marked by drift and modified drift, mostly glacial but in part deposited by icebergs.	Glacial Period b. Alluvian epoch a. Diluvian epoch	Glacial Period } c. Final retreat b. First retreat a. Early advance

No comparable sequence of publications exists for the Middle-Western region, but the history of the major stratigraphic names used in that region can be extracted from the literature (see Table 2).

ORIGIN OF LOESS

Loess had been recognized in Europe for decades. Beginning with Lyell and his contemporaries, it had been thought by many to be of lacustrine or alluvial origin. It was the classic discussion by Richthofen (1877, p. 74-84) of loess in China that first clearly argued the case for deposition by wind. Pumpelly, who had earlier studied the Chinese loess and had subscribed to the widespread lacustrine hypothesis, and who had subsequently worked on loess in Missouri, rejected his own earlier explanation and adopted the eolian hypothesis for central North America, but with the added significant opinion that most of the American loess was derived from outwash sediments rather than from the products of rock weathering (Pumpelly, 1879, p. 135). There was much debate about the matter; apparently not until 1893 did Chamberlin accept the eolian origin for even a part of the loess, and even then he still thought that the bulk of the sediment had been deposited in water.

PLUVIAL LAKES

The concept of the expansion of lakes through lowered temperature during glacial ages originated in Europe (Jamieson, 1863). However, only two years later, and apparently independently, this concept was applied to the Basin-and-Range region by I. D. Whitney (1865, p. 452). Whitney suggested the probable connection between expanded lakes and former glaciers. A quarter century later the probability of this concept was established in America through the field demonstration of a close relationship between high strandlines of pluvial lakes and drift deposited by local glaciers.

In 1889, J. C. Russell (1889, p. 369) showed through the geologic relation of strandlines to end moraines that the lake in the Mono basin, on the east flank of the Sierra Nevada, was approximately contemporaneous with local glaciation. Almost at the same time Gilbert (1890, p. 318) demonstrated a similar time relation between pluvial Lake Bonneville and former glaciers in the Wasatch Mountains.

Apart from the specific relation between glaciers and lakes in the Basin-and-Range region, an important contribution to the climatic significance of the former lakes was made by Meinzer (1922). He plotted the distribution of existing lakes and former lakes then known, and he con-

TABLE 2

Origin of Names of Major Stratigraphic Units Currently Used in Middle-Western United States

Stage	Named by	Original reference	Source of name	Remarks
Wisconsin	T. C. Chamberlin	Geikie, 1894, p. 763 (named *East-Wisconsin*)	State of Wisconsin	Name shortened to *Wisconsin* (Chamberlin, 1895, p. 270).
Sangamon	Frank Leverett	Leverett, 1898a, p. 176	Sangamon County, Illinois	
Illinoian	Frank Leverett	Chamberlin, 1896, p. 874 (named *Illinois*)	State of Illinois	Name changed to *Illinoian* (Leverett, 1898a, p. 171).
Yarmouth	Frank Leverett	Leverett, 1898b, p. 239	Spoil of dug well, Yarmouth, Iowa	
Kansan	T. C. Chamberlin	Geikie, 1894, p. 755 (named *Kansan formation*)	Northeastern Kansas	The name was first applied to the drift now called Nebraskan, but was shifted to apply to post-Aftonian drift in Iowa (Chamberlin, 1896, p. 873).
Aftonian	T. C. Chamberlin	Chamberlin, 1895, p. 272	Peat exposed near Afton Junction, Iowa	Peat ("forest bed") described by W J McGee (1891, p. 486-496).
Nebraskan	Bohumil Shimek	Shimek, 1909, p. 408	State of Nebraska	Had earlier been called *sub-Aftonian* and *pre-Kansan*.

sidered that the least-dry parts of the region today are comparable with the parts that were driest during the last glacial age. On a basis of the difference of latitude between the two parts, he attempted to derive a value for the inferred change in mean annual temperature since glacial time. This was one of the earliest attempts at quantification of differences between pluvial and existing climatic parameters.

Although widespread usage now labels as *pluvial* the former expanded lakes in regions that are dry today, the label was not applied to the lakes at the time when they were first recognized, nor indeed was it originally applied to dry regions at all. Both the concept of greatly increased precipitation at times during the Pleistocene and the term *pluvial period* originated with Alfred Tylor, a British manufacturer and active amateur geologist. Tylor (1868) propounded the belief that there had been a time in the Quaternary when rainfall values reached many times their present values, and proposed for it the term *pluvial period.* Contrary to widespread present belief, the term was not originally applied to the dry regions or the tropical regions with which it is associated today. It was applied specifically to bodies of valley gravel, occurring as fills, terraces and mantles of outwash, coarse alluvium, and colluvium in England and northern France. Large runoff and hence heavy rainfall were inferred from the coarse grain size and topographic position of the gravel bodies. In other words the term *pluvial period* was first invoked to explain climatic conditions inferred from sediments in high-middle latitudes and nothing else. Only later (*cf.* Hull, 1885, p. 182) was the term transferred to dry regions and to lower latitudes.

Ice Sheets and Isostasy

The general concept of uplift of the crust by postglacial rebound, following isostatic depression under the weight of ice sheets, was first proposed in Europe as a theory (Jamieson, 1865, p. 178), but it was soon taken up in America. N. S. Shaler (1874, p. 338) applied it to explain observations of postglacial emergence on the coast of Maine. McGee (1881) deduced the existence of a time lag (since approximated by measurement) between the withdrawal of an ice sheet and crustal recovery. During his classic study of pluvial Lake Bonneville, Gilbert (1890, p. 362-383) tested the principle by measuring strandlines. He found that they are warped, and he attributed their domelike deformation to isostatic compensation following removal of the very considerable weight of the water when the lake disappeared.

Subsequently many American geologists, among them F. B. Taylor, J. W. Goldthwait, G. K. Gilbert, and Frank Leverett, contributed largely to the complex of measurements of altitude and observations of stratigraphic relations, by which the outline of the history and deformation of the Great Lakes was basically established. That complex is equalled in importance only by the data developed by Scandinavian and Finnish scientists on the similar features of the Baltic region.

Glacial Control of Sea Level

Although the first correct deduction of the relation between the building of the Pleistocene ice sheets and fluctuation of the level of the sea is a European contribution dating back to Charles Maclaren (1841), one of the early discussions of the problem is American, the work of Charles Whittlesey (1868). That discussion includes an admirably lucid statement of the hydrologic cycle, emphasizing the exchange among liquid, gaseous, and solid water-substance. Whittlesey deduced that lowered secular temperature would result in lowering the snowline. Also he calculated that, on assumptions as to the extent and thickness of former ice sheets, glacial-age sea level might have been lowered by as much as 350 to 400 ft.

Another theoretical discussion, by N. S. Shaler (1875), appeared shortly afterward. Shaler's calculation led him to believe that glacial-age sea level might have been lower by as much as 1,200 ft. A third discussion, by Warren Upham (*in* C. H. Hitchcock, 1878, p. 18, 329-333), appeared a few years later.

However, little information from direct field observation in America was brought to bear on the problem until near the end of the 19th century. Thereafter, study of the Pleistocene sediments and morphology of the Atlantic Coastal Plain led to the appearance of a series of publications through several decades. The early contributions, based on field studies in Virginia or the Chesapeake Bay region, were the work of McGee (1888a, 1888b), Shaler (1890), and Darton (1902). These workers recognized, in both the morphology of the terrain and the internal character of the surficial sediments, evidence of Pleistocene marine submergence of the outer part of the Coastal Plain. These publications were followed by a number of others, treating of various segments of the Coastal Plain from Maryland to Florida, and relying heavily on morphology with little attempt at detailed stratigraphic study. Typical of this group is a paper (Shattuck, 1901) setting forth the concept of several successive "terraces" at various altitudes, each covered with a thin blanket of marine sediments and each terminating landward in a wave-cut cliff. Most or all of the "terraces" were thought to be of Pleistocene age. The succession of benchlike forms was thought by some to have resulted from crustal movements, and by others to reflect eustatic movements of sea level. Although these views prevailed for several decades, it is now realized that they were rigid and unrealistic.

Direct field observation of coral reefs and morphology of islands in the tropical Pacific led R. A. Daly (1910) to propose what he later (Daly, 1915) called the glacial-control hypothesis of submergence of coral reefs, as an alternative to the hypothesis of crustal subsidence advocated earlier by Darwin and Dana. Today both subsidence and glacial control are recognized as having contributed in varying degrees to the submergence of Pacific islands.

Paleontology

The earliest large American contribution to the theory of distribution of organisms seems to have been that of Asa Gray (1878). Comparing the large collections of plants brought back from Japan by the voyages of Perry and Rogers with those from temperate latitudes elsewhere, Gray perceived strong resemblances between eastern North America and eastern Asia, and he inferred that the floras reflected a circumpolar dispersion of plants in Tertiary time. Based on the inferences of Oswald Heer and others in Europe in the 1860's, the sequence of events suggested by Gray was that the appearance of cold glacial-age climates pushed the widespread late-Tertiary floras toward the equator and introduced arctic plants into middle latitudes, around the margins of the ice sheets. Then, during nonglacial times, the arctic kinds returned toward the pole. Basically this concept has stood the test of nearly a hundred years of subsequent research.

Comparable results on shifts of populations among animals were reached much more slowly because of the scarcity and fragmentary nature of the fossil evidence. Although extinct mammals had been found as fossils early in the 19th century, their sporadic occurrence was insufficient as a basis for a sound theory. It was not until 1914 that O. P. Hay (1914) published a systematic account of the Pleistocene mammals of a large area, in the form of a monograph on Iowa. But even this did not include biogeographic inferences. The publication was followed, at long intervals, by systematic catalogs, by regions and localities, of Pleistocene mammals right across North America (Hay, 1923-1927). These works included maps and discussions of biogeography, stratigraphic ranges, extinctions, and post-Tertiary evolution. Since the appearance of this work, opinion has changed as to the stratigraphic positions of many of the occurrences, but the contribution remains basic.

At about the same time there appeared a comparable monograph on the life of the Pleistocene Period. The work of F. C. Baker (1920), this publication emphasized freshwater mollusks, but nevertheless dealt with other invertebrates and with vertebrates and plants as well. Although far less detailed than Hay's work as regards vertebrates, Baker's monograph discussed biogeography at greater length. These two monographs together mark the first long step toward cataloguing and synthesizing Pleistocene faunas and floras in North America.

Dating of Pleistocene Events

Apparently the earliest attempt in America to date an event in Pleistocene time was based on measurement made in 1789 by Andrew Ellicott, the first American surveyor to measure the Niagara Gorge and Falls. Ellicott told of his measurements to his friend William Maclay, Senator from Pennsylvania, who wrote in his journal on February 1, 1790, as follows (Maclay, 1927, p. 185):

"Mr. Ellicott's accounts of Niagara Falls are amazing indeed. I communicated to him my scheme of an attempt to account for the age of the world, or at least to fix the period when the water began to cut the ledge of rock over which it falls. The distance from the present pitch to where the falls originally were, is now seven miles. For this space a tremendous channel is cut in a solid limestone rock, in all parts one hundred and fifty feet deep, but near two hundred and fifty at the mouth or part where the attrition began. People who have known the place since Sir William Johnson took possession of it, about thirty years ago, give out that there is an attrition of twenty feet in that time. Now, if 20 feet = 30 years = 7 miles, or 36,960 feet; answer, 55,440 years."

Several later estimates were made, mostly during the 19th century; not until 1928, when it was shown by the Canadian geologist W. A. Johnston (1928) that geologic relations in the gorge invalidate any attempt at extrapolation on existing rates of retreat of the Falls, did this kind of estimate cease to be made. Attempts to date the retreat of the St. Anthony Falls of the Mississippi River encountered analogous difficulties.

Efforts to extrapolate into the past the estimated rates of activity of existing processes have been made repeatedly. An early attempt to use the rate of recession of wave-cut cliffs of Lake Michigan to date the Glenwood phase of glacial Lake Chicago, made by Edmund Andrews (1870), yielded a value too small by comparison with estimates from modern radiocarbon dates. An estimate of the time elapsed since deglaciation of the escarpment south of Lake Erie, based on rates of stream dissection (Wright, 1911, p. 565) proved later to be not far from C^{14} dates of contemporaneous events.

Various estimates have been based on the progress of processes of weathering, although similar attempts were made earlier in Europe. Notable among American efforts are the attempt by François Matthes (1930, p. 70-72) to measure the time elapsed since the El Portal glaciation in the Sierra Nevada, and calculations on the dates of Pleistocene soils (including "gumbotils") made in Iowa by G. F. Kay (1931) and in Bermuda by R. W. Sayles (1931). All these calculations involved assumptions recognized by their authors; perhaps their most useful result was the stimulation of thought about Pleistocene chronology.

At least two attempts have been made to date postglacial time by means of rates of accumulation of peat. In 1881, G. F. Wright (1881) measured the thickness of peat in a kettle near Andover, Massachusetts, as the equivalent of 8 ft. Applying a rate of accumulation of 1 in. per century from a measurement of post-Roman peat in northern France, he derived a value of approximately 10,000 years for the time elapsed since the kettle was formed during deglaciation. An interesting sidelight on this calculation is Wright's view of the bearing of this value on the then fashionable Croll hypothesis of climatic change. After deducing that under Croll's hypothesis the last glaciation should have ended, not 10,000 but about 80,000 years ago, Wright remarked: "These considerations have led me to look with increasing distrust upon the astronomical calculations which are made concerning the Glacial period. . . ." Some geologists today would express similar doubts about the more elaborate astronomical theory put forth by Milankovitch.

Another attempt at dating by means of peat accumulation was made in Alaska (Capps, 1915), using an ingenious combination of measurement of peat thickness and counting of the annual rings in related tree stumps.

Two other calculations, based on quite different processes, are worth mention. One was a calculation by H. S. Gale (1915, p. 260-264) of the "age" of Owens Lake, California.

The quantities of chlorine and sodium in solution in the lake in the year 1912 were divided by the respective annual inflow of these elements estimated from sampling water from Owens River. By "age" Gale meant the time since Owens Lake ceased to overflow and began to accumulate these elements. Like most such calculations this one involved assumptions that could not be proved, at least at that time. However, this was possibly the first use of this geochemical method to arrive at a date of a Pleistocene event.

The other calculation was made by A. C. Swinnerton (1925), who attempted to date the deglaciation of a locality in Ohio. He derived the volume of existing travertine at a large spring by the measured rate of accumulation, making a number of assumptions as to constancy of local climatic and geologic conditions that could actually well have been variable.

The foregoing examples are cited as an indication of the ingenuity that has been displayed by American scientists in trying to solve the problem of absolute ages in places where some means, however imperfect, seemed to present itself. The various methods used are seen to be second-class by comparison with direct radiometry.

An early application, in American science, of radiometry to a Pleistocene event was the attempt by Schlundt and Moore (1909, p. 33) to date the radium-bearing travertine of Terrace Mountain in Yellowstone National Park. The travertine underlies, and is older than, the latest glacial drift in the area and contains far less radium than does the travertine now being deposited. On the assumptions that (1) the older travertine had a comparable concentration of radium when first deposited and (2) no radium has been lost by leaching of the older deposit, and with correction of the half-life value for radium, the calculated age of the older travertine is 11,200 years. The actual age is now thought to be greater, perhaps because of errors in the assumptions.

The rapid advances in radiometry through the use of radiocarbon and potassium/argon constitute a part of modern science and are not properly history. The dimensions of Quaternary time are now seen in much better perspective, thanks to radiometry, than in the view that was available a bare quarter-century ago.

General Retrospect

To look through the literature of Quaternary geology, particularly the literature of the 19th century, is to perceive that advances in knowledge in that field in the United States proceeded almost step by step with advances in European countries. To some extent this was the result of intercommunication through the medium of scientific journals; citations and quotations show that such intercommunication took place in both directions. But this is not the whole explanation. From the dates of publications it is evident that scientists on both sides of the Atlantic Ocean often were thinking about the same problems at the same times, and not infrequently the same idea seems to have been put forward independently by people in two different countries. Whether this was stimulated by communication via newspapers or via private correspondence, or whether certain problems were strongly "in the air" at certain times could probably not be shown without very extensive and intensive search of the record.

Certainly the publication of syntheses must have helped. The most noteworthy 19th-century synthesis of the Quaternary was James Geikie's *The Great Ice Age* (1874, 1894). The subtitle, *Its Relation to the Antiquity of Man,* shows that the work was probably conceived originally during the scientific ferment that followed the publication of Darwin's *The Origin of Species* in 1859. That this was the case is borne out by the trend of Geikie's preface to the book, although he made no open statement of the fact. The book was intended not only for scientists but for general readers, and the demand among the latter group must have been great. During the period 1874 to 1894 the growth of knowledge about the Quaternary, in both Britain and America, was very rapid, as is apparent from comparison of the two editions of Geikie's work.

A somewhat similar American work was G. F. Wright's *The Ice Age in North America* (1890). It was surely patterned in part after Geikie's book, and indeed it carried a closely similar subtitle, reflecting the continued wide public interest in prehistoric man. Like its predecessor, it too appeared two decades later (in 1911) in a new edition with substantial improvements over the original.

The rate of increase in the output of American literature on the Quaternary seems to have been accelerated rather steadily throughout the century and a quarter of output. Probably this is a general reflection of the gradual increase in the number of American scientists interested in Quaternary problems. The rate of increase seems to have been less, however, in the early part of the 20th century than it was in the latter part of the 19th. The change coincides with T. C. Chamberlin's shift of attention from glacial geology to cosmogony and related problems. After 1899 Chamberlin published only two original titles (apart from a textbook) that had to do with glacial geology. Chamberlin (1843-1928) had been so dynamic and productive a leader in the field during the preceding 25 years and had influenced so many geologists who had worked under his direction, that his shift of interest could easily have caused a falling-off of output apart from his own, which ran to nearly 60 titles in glacial geology alone.

Probably the most noteworthy of Chamberlin's juniors was Frank Leverett (1859-1943), who unlike Chamberlin devoted his entire professional life to glacial geology. Leverett was single-minded from the very beginning. After graduating at Ames, Iowa, in 1885, he was advised by W J McGee to apply for a job to Chamberlin, who was then President of the University of Wisconsin and in charge of glacial investigations for the U.S. Geological Survey. Leverett walked the 250 miles from Ames to Madison, was employed, and remained a member of the Survey for 43 years. He mapped systematically the glacial or Pleistocene geology of parts or all of many States, including Illinois, Indiana, Iowa, Kentucky, Michigan, Minnesota, Ohio, Pennsylvania, and Florida. His bulky Professional Papers of the Geological Survey are classics. Much of Leverett's field work was done entirely on foot. After he retired he estimated that he had walked an aggregate of about 100,000 miles in the field.

Other students of Chamberlin were R. D. Salisbury (1858-1922) and W. C. Alden (1871-1959). Both contributed extensively to the literature of glacial geology, Salisbury in New Jersey and in the New York City area, and Alden in many areas from New England to Montana. Salisbury is probably best known for his collaboration with Chamberlin in textbooks of geology, which strongly stressed Pleistocene geology.

J. D. Dana (1813-1895) antedates Chamberlin by a full generation. His contributions to glacial geology were less numerous than those of Chamberlin because his research interest was spread over a wider spectrum of geology. His first publication with a bearing on glaciation was the first edition of his classic *Manual of Geology* (Dana, 1863), which, together with his textbook, went through eight editions. Thereafter he began to publish papers on Quaternary geology (mostly in New England), and his bibliography lists twenty-nine, out of a remarkable total of 201 titles.

Another strong contributor was R. S. Tarr (1864-1912), who published 38 titles in glacial geology and glaciology. Although both Chamberlin and Salisbury had visited and examined glaciers in Greenland, Tarr spent considerable time in the study of glaciers and glacial features in Alaska, and he applied his experience to the interpretation of glacial features in New York State, about which many of his publications were written.

The names mentioned by no means exhaust the list of workers who made significant contributions to American Quaternary during the 19th century and the early part of the 20th. No attempt has been made to present a complete list. The aim has been only to point out a few milestones in research in this field. There have been many significant contributions of much later date, and some milestones as well; but this discussion has been confined to earlier generations of scientists who are no longer living. There will be time enough in the future to pick up the thread of this brief narrative and review the many things that scientists now living have accomplished.

References

Agassiz, Louis, 1840, Études sur les glaciers: Neuchâtel, privately published, 346 p.

Andrews, Edmund, 1870, The North American lakes considered as chronometers of post-glacial time: Chicago Acad. Sci. Trans., v. 2, 1870, p. 1-23

Baker, F. C., 1920, The life of the Pleistocene or glacial period as recorded in the deposits laid down by the great ice sheets: Univ. Illinois Bull., v. 17, no. 41, 476 p.

Capps, S. R., 1915, An estimate of the age of the last great glaciation in Alaska: Washington Acad. Sci. J., v. 5, p. 108-115

Chamberlin, T. C., 1877, Geology of eastern Wisconsin, *in* Geology of Wisconsin, survey of 1873-1877, v. 2: Madison, Commissioners of Public Printing, p. 97-246

—— 1878, on the extent and significance of the Wisconsin kettle moraine: Wisconsin Acad. Sci., Arts Lett. Trans., v. 4 (1876-77), p. 201-234

—— 1883, Terminal moraine of the second glacial epoch: U.S. Geol. Surv. Ann. Rep. 3, p. 291-402

—— 1888, The rock-scorings of the great ice invasions: U.S. Geol. Surv., Ann. Rep. 7, p. 147-248

—— 1895, The classification of American glacial deposits: J. Geol., v. 3, p. 270-277

—— 1896, Nomenclature of glacial formations: J. Geol., v. 4, p. 872-876

Conrad, T. A., 1839, Notes on American geology: Amer. J. Sci., v. 35, p. 237-251

Daly, R. A., 1910, Pleistocene glaciation and the coral reef problem: Amer. J. Sci., v. 30, p. 297-308

—— 1915, The glacial-control theory of coral reefs: Amer. Acad. Arts Sci. Proc., v. 51, p. 157-251

Dana, J. D., 1863, Manual of geology: Philadelphia, Theodore Bliss & Co., 1st ed., 798 p.

—— 1873, On the Glacial and Champlain eras in New England: Amer. J. Sci., v. 5, p. 198-211

—— 1875, Manual of geology: New York, Ivison, Blakeman, Taylor & Co., 2d. ed., 828 p.

—— 1895, Manual of geology: New York, American Book Co., 4th ed., 1087 p.

Darton, N. H., 1902, Description of the Norfolk quadrangle: U.S. Geol. Surv. Geol. Atlas, Folio 80, 4 p.

Dobson, Peter, 1826, Remarks on bowlders: Amer. J. Sci., v. 10, p. 217-218

Gale, H. S., 1914, Salines in the Owens, Searles, and Panamint basins, southeastern California: U.S. Geol. Surv. Bull. 580, p. 251-323

Geikie, James, 1874, The great ice age and its relation to the antiquity of man: London, W. Isbister, 575 p.

—— 1894, The great ice age and its relation to the antiquity of man: London, Stanford, 3d. ed., 850 p.

Gilbert, G. K., 1871, On certain glacial and post-glacial phenomena of the Maumee Valley: Amer. J. Sci., v. 1, p. 339-345

—— 1890, Lake Bonneville: U.S. Geol. Surv. Monogr. 1, 438 p.

Gray, Asa, 1878, Forest geography and archaeology: Amer. J. Sci., v. 16, p. 85-94, 183-196

Hay, O. P., 1914, The Pleistocene mammals of Iowa: Iowa Geol. Surv., v. 42, Ann. Rep. for 1912, p. 1-662

—— 1923, 1924, 1927, The Pleistocene of North America and its vertebrated animals . . . : Carnegie Instn. Publ. 322, 322A, 322B (3 v.)

Hitchcock, C. H., 1878, Surface geology, *in* The geology of New Hampshire: Concord, v. 3, pt. 3, 340 p.

Hitchcock, Edward, 1841, First anniversary address before the Association of American Geologists . . . : Amer. J. Sci., v. 41, p. 232-275

Hull, Edward, 1885, Mount Seir, Sinai, and western Palestine: London, R. Bentley, 227 p.

Hummel, D., 1874, Om rullstenbildningar: K. Svenska Vetenskapsakad., Bihang til Handl., v. 2, no. 11, 36 p.

Jamieson, T. F., 1863, On the parallel roads of Glen Roy, and their place in the history of the glacial period: Geol. Soc. London Quart. J., v. 19, p. 235-259

—— 1865, On the history of the last geological changes in Scotland: Geol. Soc. London Quart. J., v. 21, p. 178

Johnston, W. A., 1928, The age of the upper great gorge of Niagara River: Roy. Soc. Can. Trans., Sec. 4, v. 22, p. 13-29
Kay, G. F., 1931, Classification and duration of the Pleistocene period: Geol. Soc. Amer. Bull., v. 43, p. 425-466
Leverett, Frank, 1898a, The weathered zone (Sangamon) between the Iowan loess and Illinoian till sheet: J. Geol., v. 6, p. 171-181
—— 1898b, The weathered zone (Yarmouth) between the Illinoian and Kansan till sheets: J. Geol., v. 6, p. 238-243
Maclaren, Charles, 1841, The glacial theory of Professor Agassiz of Neuchatel: Edinburgh, The Scotsman Office, 62 p.
Maclay, William, 1927, The journal of William Maclay, United States Senator from Pennsylvania, 1789-1791: New York, C. A. Boni, 429 p.
Matthes, F. E., 1930, Geologic history of the Yosemite Valley: U.S. Geol. Surv. Prof. Pap. 160, 137 p.
McGee, W J, 1878, On the relative positions of the Forest Bed and associated drift formations in northeastern Iowa: Amer. J. Sci., v. 15, p. 339-341
—— 1881, On local subsidence produced by an ice-sheet: Amer. J. Sci., v. 22, p. 368-369
—— 1888a, Three formations of the middle Atlantic slope: Amer. J. Sci., v. 35, p. 120-143, 328-330, 367-388, 448-466
—— 1888b, The geology of the head of Chesapeake Bay: U.S. Geol. Surv. Ann. Rep. 7, p. 537-646
—— 1891, The Pleistocene history of northeastern Iowa: U.S. Geol. Surv. Ann. Rep. 11, pt. 1, p. 189-577
Meinzer, O. E., 1922, Map of Pleistocene lakes of the Basin-and-Range province and its significance: Geol. Soc. Amer. Bull., v. 33, p. 541-552
Newberry, J. S., 1874, Geology of Ohio—surface geology: Ohio Geol. Surv. Rep. 2, p. 1-80
Pumpelly, Raphael, 1879, The relation of secular rock-disintegration to loess, glacial drift and rock basins: Amer. J. Sci., v. 17, p. 133-144
Richthofen, Ferdinand von, 1877, China: Berlin, D. Reimer, v. 1, 758 p.
Russell, I. C., 1889, Quaternary history of Mono Valley, California: U.S. Geol. Surv. Ann. Rep. 8, p. 261-394
Sayles, R. W., 1931, Bermuda during the Ice Age: Amer. Acad. Arts Sci. Proc., v. 66, p. 381-468
Schlundt, Herman, and Moore, R. B., 1909, Radioactivity of the thermal waters of Yellowstone National Park: U.S. Geol. Surv. Bull. 395, 35 p.
Shaler, N. S., 1874, Preliminary report on the recent changes of level on the coast of Maine. . . : Boston Soc. Nat. Hist. Mem. 2, p. 320-340
—— 1875, Notes on some of the phenomena of elevation and subsidence of the continents: Boston Soc. Nat. Hist. Proc., v. 17, p. 288-292
—— 1890, General account of the fresh-water morasses of the United States, with a description of the Dismal Swamp district of Virginia and North Carolina: U.S. Geol. Surv., Ann. Rep. 10, p. 255-339
Shattuck, G. B., 1901, The Pleistocene problem of the North Atlantic coastal plain: Johns Hopkins Univ. Circ. 20, p. 69-75
Shimek, Bohumil, 1909, Aftonian sands and gravels in western Iowa: Geol. Soc. Amer. Bull., v. 20, p. 399-408
Silliman, Benjamin, 1821, Notice of "Geological essays . . .": Amer. J. Sci., v. 3, p. 47-57
Swinnerton, A. C., 1925, A method of estimating post-glacial time: Science, v. 62, p. 566
Tylor, Alfred, 1868, On the Amiens gravel: Geol. Soc. London Quart. J., v. 24, p. 103-125
Upham, Warren, 1877, On the origin of kames or eskers in New Hampshire: Amer. Assoc. Adv. Sci. Proc., v. 25, p. 216-225
Whitney, J. D., 1865, Geological Survey of California: Geology, v. 1: Philadelphia, Sherman & Co., 498 p.
Whittlesey, Charles, 1860, On the drift cavities, or "potash kettles" of Wisconsin: Amer. Assoc. Adv. Sci. Proc., v. 13, p. 297-301
—— 1868, Depression of the ocean during the ice period: Amer. Assoc. Adv. Sci. Proc., v. 16, p. 92-97
Winchell, N. H., 1873, The geological and natural history survey of Minnesota: First Ann. Rep. (1872): 168 p.
Wright, G. F., 1881, An attempt to calculate approximately the date of the glacial era in eastern North America . . . : Amer. J. Sci., v. 21, p. 120-123
—— 1889, The ice age in North America: New York, Appleton, 622 p.
—— 1911, The ice age in North America and its bearings upon the antiquity of man: Oberlin, Ohio, Bibliotheca Sacra Co., 5th ed., 763 p.

PART I: GEOLOGY

QUATERNARY GEOLOGY OF NORTHERN GREAT PLAINS*

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THIS PAPER outlines the Quaternary geology of South Dakota, North Dakota, and the part of Montana that lies east of the Rocky Mountains (Fig. 1). This area, which comprises about 580,000 km^2 (225,000 sq miles) lies chiefly in the northern Great Plains physiographic province. The eastern parts of North Dakota and South Dakota, however, are in the Central Lowland province.

Most of the discussion concerns the area that was covered by continental glaciers (Fig. 1). Alpine glaciation along the east flanks of the Rocky Mountains in Montana is discussed briefly for purposes of correlation with the continental glaciations.

The glacial map of the United States east of the Rocky Mountains (Flint, 1959) at a scale of 1:1,750,000 depicts the overall area. The most recent State maps of glacial deposits (scale 1:500,000) are those of South Dakota by Flint (1955), of North Dakota by Colton *et al.* (1963), and of Montana east of the Rocky Mountains by Colton *et al.* (1961).

PREGLACIAL SETTING AND DRAINAGE

The bedrock underlying the continental glacial drift is chiefly poorly consolidated and easily eroded shales, siltstones, and sandstones of Cretaceous and Tertiary age, but in southeastern North Dakota and northeastern South Dakota rocks of Paleozoic and of Precambrian age directly underlie the drift. The preglacial land surface was more dissected and relief was greater than that of the present drift surface; it probably resembled the present unglaciated areas of these states.

The pattern and direction of the surface drainage also was different from that following glaciation. Figure 2 shows the known and inferred courses of this ancestral drainage. In preglacial time the Cheyenne River in South Dakota and all streams north of it in the report area drained into Hudson Bay in Canada (Alden, 1932, Pl. 1; Flint, 1949, p. 68; Benson, 1953, p. 165; Lemke, 1960, p. 108). Flint (1955, Pl. 7) indicated that all streams south of the Cheyenne River drained into the Gulf of Mexico. However, recent investigations by F. V. Steece and L. W. Howells and by L. S. Hedges (reports in preparation) in the central part of the James River Lowland indicate that the ancestral Bad River and possibly the ancestral White River also drained northward into the ancestral Cheyenne River instead of flowing south as shown on Figure 2. Thus, in preglacial time the present course of the Missouri River in South Dakota, North Dakota, and eastern Montana had not yet been established.

DESCRIPTION OF GLACIATED REGION

GREAT PLAINS AREA

Most of the drift surface of the northern Great Plains is gently rolling and slopes eastward from an altitude of about 1,200 m in Montana to approximately 300 m in eastern North and South Dakota. Drainage is into the Gulf of Mexico by way of the Missouri River, its tributaries, and the Mississippi River. Stream valleys together with a few mountain outliers and morainal hills and ridges break monotonous expanses of subdued topography. Much of the Missouri River valley is 2-3 km wide, having steep, locally dissected bedrock walls 100 to more than 200 m high. The floor of the valley is nearly flat and in most places is underlain by alluvium exceeding 30 m in thickness.

The Coteau du Missouri, the easternmost subdivision of the Great Plains province, is a topographically high belt of hummocky dead-ice morainal material and includes end moraines of several different ice advances. This feature, which is 30-120 km wide, extends from northeastern Montana to south-central South Dakota. It attains a maximum altitude of approximately 750 m in northwestern North Dakota. Its northeast-facing escarpment (Fig. 1), which forms the approximate boundary between the Great Plains province to the west and the Central Lowland province to the east, commonly is 60-90 m high in North Dakota but is generally somewhat less prominent in many parts of South Dakota. This escarpment acted as a buttress to advancing ice sheets and played an important role in deter-

* The authors have drawn freely upon many published and unpublished sources for information; all are gratefully acknowledged. We appreciate critical reading and suggestions by D. R. Crandell, J. T. McGill, R. D. Miller, R. Van Horn, F. V. Steece, S. J. Tuthill, and D. J. Varnes. However, the responsibility for the conclusions and interpretations drawn here rests with the authors.

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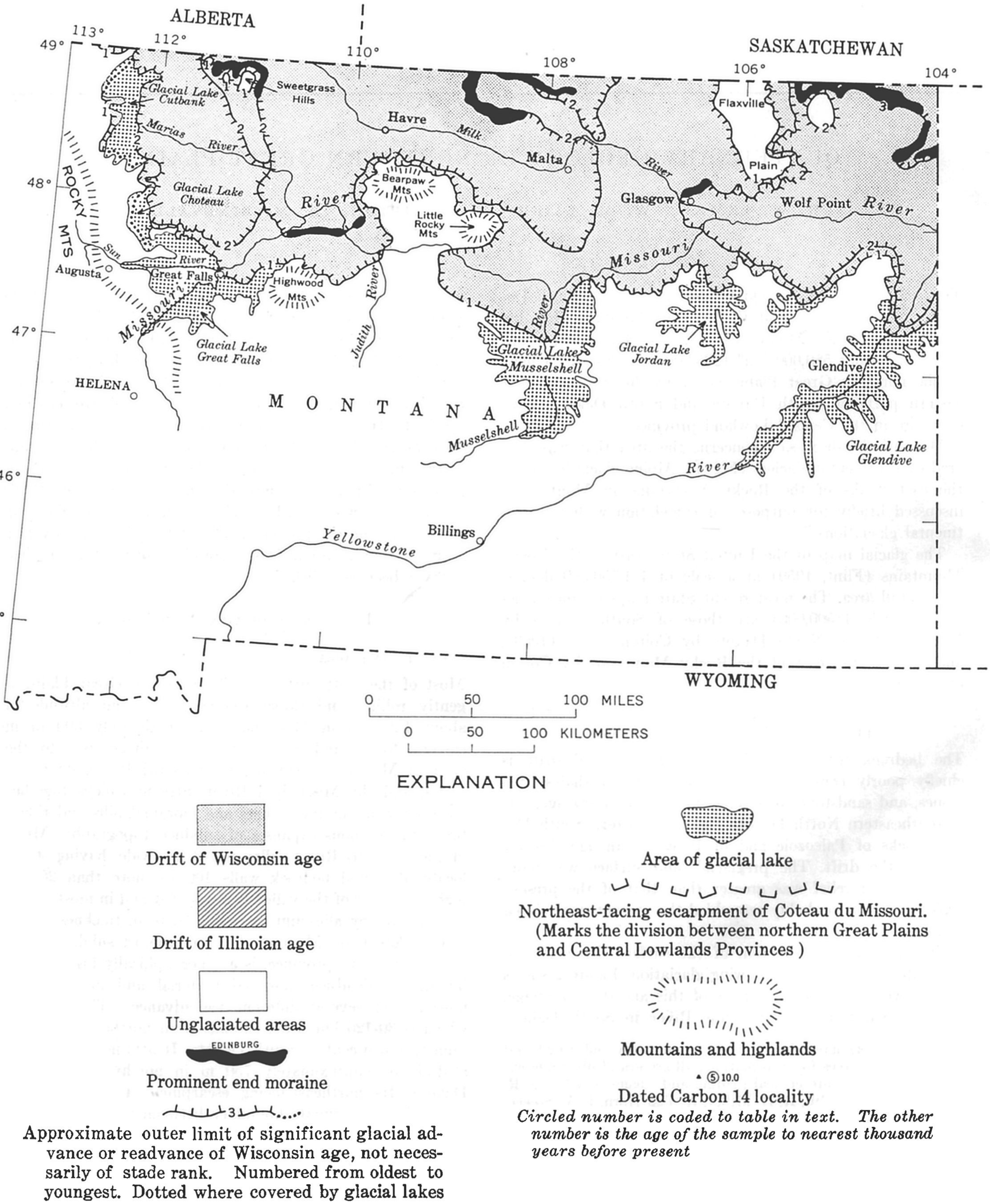
ALBERTA
SASKATCHEWAN
WYOMING
MONTANA
49°
48°
47°
46°
45°
113°
112°
110°
108°
106°
104°
Glacial Lake Cutbank
Sweetgrass Hills
Marias River
Glacial Lake Choteau
ROCKY MTS
Sun River
Augusta
Great Falls
Missouri
Glacial Lake Great Falls
HELENA
Highwood Mts
Judith River
Bearpaw Mts
Little Rocky Mts
Havre
Milk River
Malta
Flaxville
Plain
Glasgow
Wolf Point
Missouri River
Glacial Lake Musselshell
Musselshell River
Glacial Lake Jordan
Glendive
Glacial Lake Glendive
Yellowstone River
Billings
0 50 100 MILES
0 50 100 KILOMETERS
EXPLANATION
Drift of Wisconsin age
Drift of Illinoian age
Unglaciated areas
EDINBURG
Prominent end moraine
3
Approximate outer limit of significant glacial advance or readvance of Wisconsin age, not necessarily of stade rank. Numbered from oldest to youngest. Dotted where covered by glacial lakes
Area of glacial lake
Northeast-facing escarpment of Coteau du Missouri. (Marks the division between northern Great Plains and Central Lowlands Provinces)
Mountains and highlands
⑤10.0
Dated Carbon 14 locality
Circled number is coded to table in text. The other number is the age of the sample to nearest thousand years before present

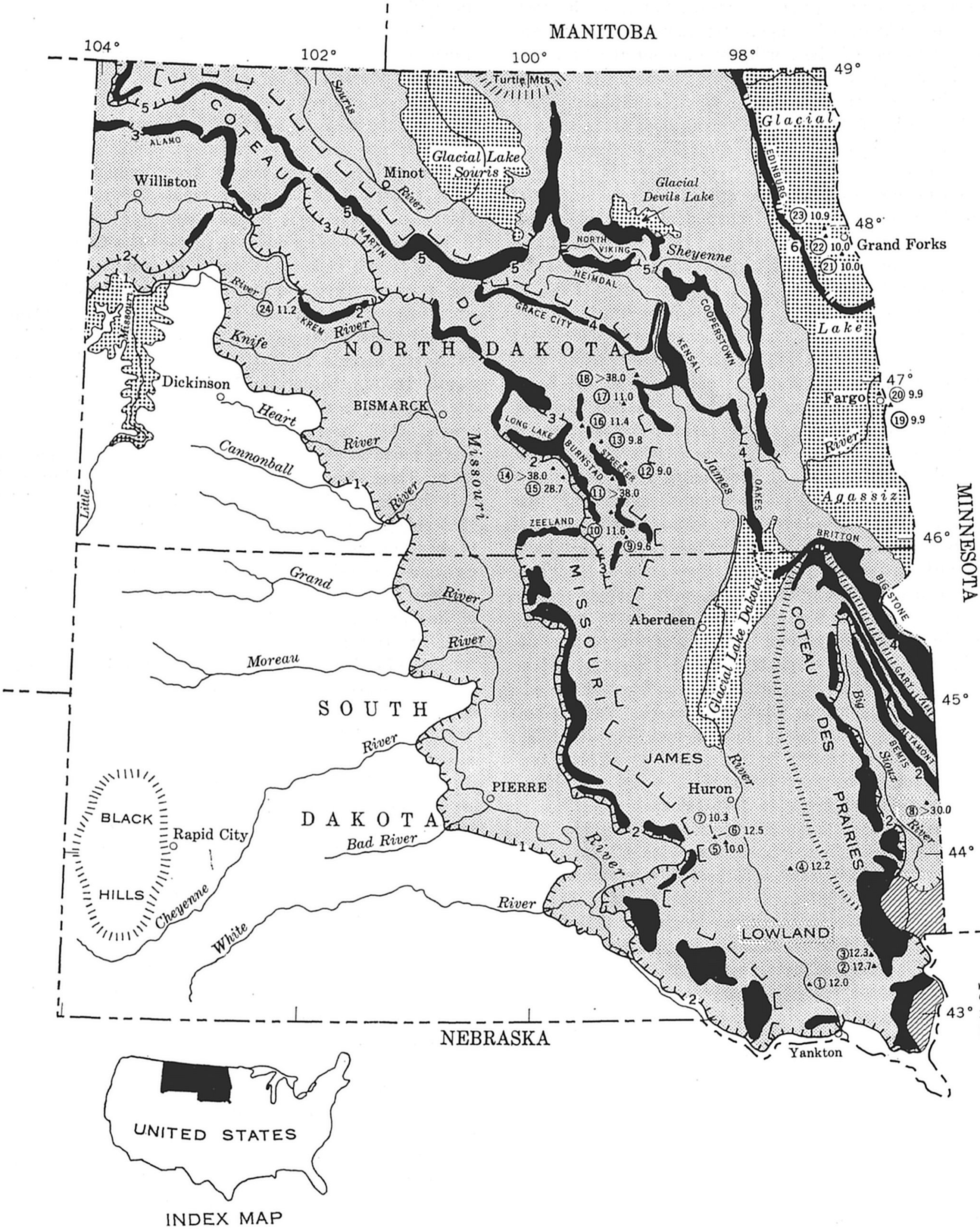

Figure 1. Map showing glaciated and unglaciated areas.

mining directions of ice advances and the positions of end moraines and other glacial features.

Drift on the Coteau du Missouri was formerly known as the Altamont moraine when it was believed that it represented a single large terminal moraine. Later, as the complex origin of the drift became better known, Townsend and Jenke (1951) named the northwest segment the Max moraine. In some places the highest and steepest parts of the Coteau du Missouri are underlain by topographically high bedrock, mostly of Tertiary age. These high areas are separated by several broad low sags underlain by Pierre Shale of Cretaceous age.

Several mountain outliers and other high areas are driftfree in the otherwise glaciated portion of Montana. These include the Sweetgrass Hills, Bearpaw, Highwood, and Little Rocky Mountains, and the Flaxville Plain (Fig. 1). Drift was deposited to an altitude of approximately 1,485 m (4,950 ft) along the north flank of West Butte in the Sweetgrass Hills at a time when glacier ice probably ranged in thickness from 300 to 500 m. Drift extends to an altitude of 1,275 m (4,250 ft) on the north side of the Bearpaw Mountains, indicating a minimum ice thickness of about 500 m on the surrounding plains. The continental glacier reached an altitude of 1,170 m (3,900 ft) along the north flank of the Highwood Mountains. If it is assumed that the ice sheet reached its maximum height against the north flanks of these three high areas at approximately the same time, the ice surface between these points must have sloped southward at less than 0.1%. Two areas of the Flaxville Plain in northeast Montana also were not covered by glacier ice. Ice reached a maximum altitude of approximately 870 m (2,900 ft) along the edges of this plain.

Proglacial lakes in the northern Great Plains formed in north-trending valleys when drainage was blocked by advancing glacier ice include Glacial Lakes Cut Bank, Choteau, Great Falls, Musselshell, Jordan, and Glendive (Fig. 1). The first three probably were dammed by at least two separate ice advances, resulting in the deposition of laminated silt and clay as much as 22 m thick.

Water reached a maximum altitude of approximately 1,170 m (3,900 ft) in Glacial Lakes Cut Bank, Choteau, and Great Falls, and these lakes were probably interconnected at their maximum heights. Glacial Lake Great Falls was partially lowered from its maximum height by the cutting of a deep spillway channel, Shonkin Sag, along the north side of the Highwood Mountains (Alden, 1932, p. 88). In part of the area covered by Glacial Lake Great Falls, lacustrine deposits both overlie and underlie a thick section of till.

The ephemeral nature of Glacial Lakes Musselshell, Jordan, and Glendive is suggested by the near absence of lacustrine deposits. The boundaries and existence of these lakes are based in large part upon the distribution of icerafted erratics, altitudes of supposed spillways, and topography.

CENTRAL LOWLAND AREA

In North Dakota the Central Lowland surface slopes northeastward from an altitude of about 570 m (1,900 ft) near the Coteau du Missouri to about 240 m (800 ft) on the floor of Glacial Lake Agassiz in the northeast corner of the State. Drainage is northward into Hudson Bay by way of the Souris and Red Rivers. The Turtle Mountains, which rise 90-120 m above the surrounding terrain, are a mesa-like high of Tertiary bedrock capped by dead-ice morainal material. Approximately half the Turtle Mountains lie in North Dakota and the rest in Canada.

Thick deposits of ground moraine, outwash deposits, well-defined end moraines, and glacial-lake deposits cover most of the Central Lowland in North Dakota. Bedrock outcrops are mostly confined to small exposures along stream valleys. In the Souris River loop north of Minot, drift (mostly till) commonly ranges from 30 to 75 m in thickness, and in some buried channels exceeds 150 m.

Three glacial lakes covered part of the Central Lowland area of North Dakota. Of these, Glacial Lake Agassiz was the most extensive and had the most complex history. The area covered by this lake is a remarkably flat plain that slopes imperceptibly toward the Red River. The lake floor in North Dakota is 50-75 km wide, extending from the South Dakota border northward into Canada, where its area exceeds that in the United States. The lacustrine deposits, which consist of silt, sand, and in places bedded clays, range in thickness from a few meters to more than 20 m. These deposits lie on till in some areas but on bedrock in others. Glacial Lake Souris was a proglacial lake that enlarged northward as the ice front receded. Its bed is very flat, local relief generally being less than 3 m. The associated deposits, which range in thickness from zero to at least 22 m, consist chiefly of sand with lesser quantities of fine gravel, silt, and clay. Deposits of Glacial Devils Lake are thin and patchy in most places (Aronow *et al.*, 1963, p. 67); boulders and cobbles, derived from reworking of till by wave action, are scattered over the former lake floor.

The Central Lowland in South Dakota extends over the eastern third of the State. The most conspicuous feature is the Coteau des Prairies, a widespread highland area reaching altitudes of over 630 m (2,100 ft), which extends southeastward from the North Dakota border into Minnesota and northwest Iowa. It is bounded on the east by the Minnesota–Red River Lowland (Flint, 1955, p. 5) and on the west by the James River Lowland. Unlike the Coteau du Missouri, its height of 180-240 m is mainly due to a thick accumulation of glacial drift rather than to bedrock. Test holes on this plateau penetrated as much as 210 m of drift before reaching the underlying Pierre Shale.

The broad, arcuate James River lowland lies mostly at altitudes ranging from 390 to 420 m (1,300 to 1,400 ft). In the center of its northern part is the Glacial Lake Dakota Plain, which is 145 km long in South Dakota and extends an additional 25-30 km into North Dakota. The plain is generally 40-50 km wide; local relief rarely exceeds 3 m and in most places is less than 1 m. Lake sediments, which consist mostly of silt with some fine sand and clay, have an average thickness of about 12 m and a maximum thickness of about 30 m.

The Minnesota–Red River Lowland is a broad, low depression drained to the southeast by the Minnesota River and to the north by the Red River. The extreme northeast

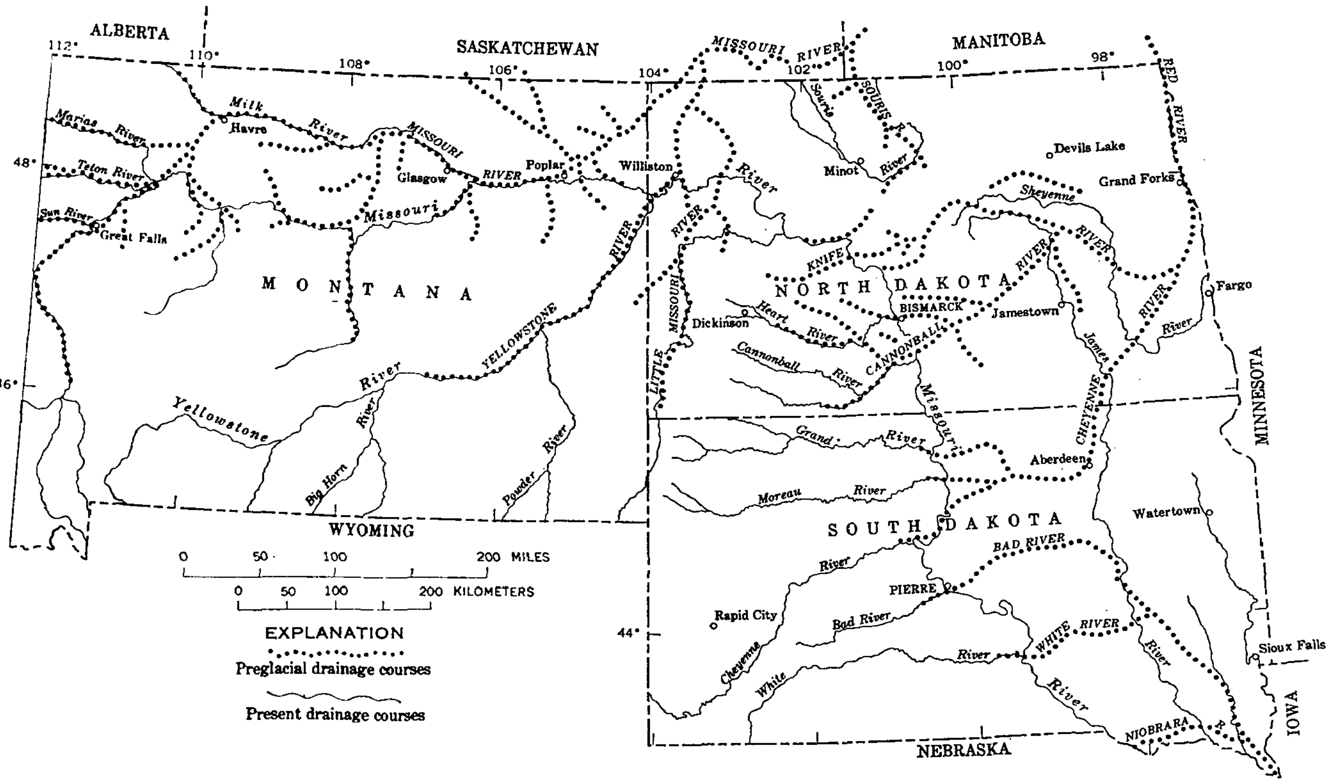

Figure 2. Map showing major preglacial drainage courses.

corner of South Dakota was occupied by the south end of Glacial Lake Agassiz, which was drained by a steep-sided trench, mostly 1-3 km wide and about 30 m deep. This trench now contains Lake Traverse and Big Stone Lake. The ridge between these lakes forms the present Continental Divide (altitude about 295 m or 980 ft) between northward drainage into Hudson Bay via the Red River and southward drainage into the Gulf of Mexico via the Minnesota and Mississippi Rivers.

Pre-Wisconsin Glaciations

Four glaciations have been recognized in the midwestern United States. These are, from oldest to youngest, the Nebraskan, Kansan, Illinoian, and Wisconsin, which are separated by the Aftonian, Yarmouth, and Sangamon Interglaciations, respectively. In the area of this report, Wisconsin glacial deposits predominate; therefore, the three older glaciations are discussed together.

Continental drift of pre-Wisconsin age has not been definitely recognized in either Montana or North Dakota but has been identified in a number of places in South Dakota (Flint, 1955, p. 30-41; Simpson, 1960, p. 74-77; Steece *et al.*, 1960).

The Nebraskan glacier probably covered most of South Dakota east of the Coteau du Missouri. Drift of Nebraskan age has been identified near Hartford 15 miles west of Sioux Falls (Steece *et al.*, 1960).

The Kansan glacier likewise probably covered all of South Dakota east of the Coteau du Missouri. Drift of Kansan age and sediments of the Yarmouth Interglaciation have been identified near Hartford (Steece *et al.*, 1960) and at a number of other exposures in the vicinity of Sioux Falls.

Illinoian drift, unlike drift of previous glaciations, is exposed in southeastern South Dakota (Fig. 1), according to Steece (1959) and Tipton (1959, 1960). The identification of Illinoian drift in northeastern Nebraska (Flint *et al.*, 1959; E. C. Reed, oral communication, 1959) supports this belief. Flint (1955, p. 30) suggested possible glaciation during the Illinoian in South Dakota on the basis of a few outcrops of till, the widespread presence of Loveland Loess in nearby states, and drainage relations. Warren (1952) believed that the part of the Missouri River near Chamberlain (about 100 km southeast of Pierre) was formed during the Illinoian. However, White (1964) has suggested that the Missouri River was formed during early Wisconsin time.

From the distribution of till of pre-Wisconsin age in South Dakota, it is evident that glaciers advanced southward chiefly via the James River and Red River lowlands. Whether the glaciers topped the Coteau du Missouri and extended farther westward is unknown.

Till older than Wisconsin age has not been definitely identified in North Dakota. Clayton (1962, p. 55) described a till found in several exposures about 90 km southeast of Bismarck that may be older than Wisconsin. This till, which is overlain by 0.3 m of gravel stained with iron oxide and about 6 m of younger till, is well consolidated. Unlike tills in the area that are believed by Clayton to be of Wisconsin age, it has numerous widely spaced joints as much as 3 m long. The joint surfaces are coated with iron and manganese oxide. No carbonate leaching was observed in this till, but in all other respects it differs little from tills

TABLE 1

Classification and Correlation of the Wisconsin Glaciations in Midwestern and North-Central United States[a]

YEARS[b]	LEIGHTON (1933)	THWAITES (1943, 1946)	FLINT (1955)	LEIGHTON (1956, 1957)	LEIGHTON (1960)	FRYE AND WILLMAN (1960)	LEMKE, LAIRD, TIPTON, AND LINDVALL (THIS PAPER) CONTINENTAL[c]	LEMKE, LAIRD, TIPTON, AND LINDVALL (THIS PAPER) ALPINE[d]
0			RECENT			RECENT	POSTGLACIAL	NEOGLACIATION
5000						VALDERAN		"ALTITHERMAL INTERVAL"
								Late stade, PINEDALE GLACIATION
10,000	MANKATO	5th-MANKATO (or VALDERS) 4th VALDERS	MANKATO	VALDERS	VALDERS GLACIAL			
	INTERSTADIAL	TWO CREEKS	INTERVAL	TWO CREEKS	TWO CREEKS INTERGLACIAL	TWOCREEKAN		INTERSTADE
	CARY	3d-CARY	CARY	MANKATO	MANKATO GLACIAL	WOODFORDIAN	ADVANCE 6 (Late Wisconsin)	Middle and early stades, PINEDALE GLACIATION
15,000				CARY	BOWMANVILLE INT.		ADVANCE 5	
	INTERSTADIAL		INTERVAL	INTERSTADIAL	CARY GLACIAL			
	TAZEWELL	2d-TAZEWELL	TAZEWELL	TAZEWELL	ST. CHARLES INT. TAZEWELL GLACIAL		ADVANCE 4	
	INTERSTADIAL		INTERVAL	INTERSTADIAL	GARDENA INT.			
20,000	IOWAN (OF ILL.)	1st IOWAN	IOWAN	IOWAN (OF ILL.)	IOWAN GLACIAL		ADVANCE 3 ADVANCE 2	
25,000					FARM CREEK INTERGLACIAL	FARMDALIAN	(Early Wisconsin)	INTER-GLACIATION
30,000	40,000 yrs. ---------- Type Iowan drift of Iowa in this age or older. ----------				FARMDALE GLACIAL	ALTONIAN	ADVANCE 1	BULL LAKE GLACIATION
70,000								

[a] From Wright (1957) and Frye and Willman (1963); modifications and additions by authors of this paper.

[b] Time scale in radiocarbon years before present; taken from Frye and Willman (1963). Time correlations of equivalent units of other classifications are approximate.

[c] Units used here restricted to discussion in this paper. Position of units in column do not indicate strict time correlation or necessarily direct correlation with other classifications.

[d] Tentative correlation based upon tracing of continental drift sheets across northern Great Plains--work of G. M. Richmond (this volume), and R. W. Lemke (this paper).

of Wisconsin age. Compact jointed till that has a manganese coating along joint surfaces underlies an upper Wisconsin till north of Minot (Lemke and Kaye, 1958, p. 93-98). Invertebrate fossils assigned to the Yarmouth Interglaciation have been found in terrace deposits in southwestern North Dakota (Tuthill *et al.*, in press).

Continental till of pre-Wisconsin age has not been positively identified in Montana. A till possibly older than Wisconsin has been reported in northeastern Montana along Smoke Creek (Howard, 1960, p. 23; Witkind, 1959, p. 18, 27). As discussed by Howard, the material in question may actually not be a till; however, it underlies a till of Wisconsin age.

The belief that glaciers of pre-Wisconsin age reached southward at least to a point about 90 km north of the northern boundary of Montana is supported by exposures of drift of pre-Wisconsin age along Oldman River in the vicinity of Lethbridge, Alberta. Stalker (1963) described measured sections in this area where, on the basis of stratigraphic relations and C^{14} dates, continental drifts of Nebraskan, Kansan, and Illinoian ages may be present.

Wisconsin Glaciations

Six distinct and separate glacial advances during Wisconsin time are postulated to have taken place in the area (Fig. 1). The lithological similarity of the tills, the near absence of loess sheets and fossil soils between the drift sheets, and irregularities in some of the C^{14} dates have precluded distinguishing drift sheets by the more conventional method of superposition. Therefore, it has been necessary to differentiate drift sheets chiefly by such features as the presence of prominent end moraines and ice-marginal channels, the trends of moraines, drumlins, and eskers, the relative development and modification of drainage systems, the relations of valley trains and gravel terraces, and the variations and ruggedness of the topography.

The tills of all the ice advances are very similar. Laboratory analyses and field observations indicate that they consist typically of nearly equal parts of clay, silt, and sand (Lemke, 1960, p. 46; Winters, 1963, p. 28). Gravel-sized materials generally constitute less than 5%, and cobbles and boulders less than 1%. The unoxidized tills are generally dark olive gray to bluish gray. Oxidation of iron results in a buff or drab yellow color, which commonly extends to depths of 10-15 m and which generally is the only discernible weathering effect. Partial cementation by caliche 0.5-1.5 m below the surface is common.

Nomenclature problems

Opinion differs as to the age and correlations of continental drift sheets of Wisconsin age in the United States. Moreover, recent changes in stratigraphic nomenclature and shifts in the stratigraphic position of certain glacial deposits have added to the problems (Table 1). Because of these problems and the long distances to type localities or dated localities, we have merely numbered the different ice advances. Each ice advance (numbered from oldest to youngest) will be described separately, together with a tentative correlation with the stratigraphic sequence of other workers. It should be emphasized that a particular advance may or may not be of stadial rank. In some instances it may represent an advance of only a few kilometers beyond a former glacier terminus, and thus it may have little regional significance.

Advance 1

The outer margin of advance 1, as shown on Figure 1, marks the maximum extent of glaciation during Wisconsin time. Its position is based upon the outer limit of glacial erratics, except that in the southeastern part of South Dakota, where erratics are scarce, it is marked by end-moraine remnants. The position of this margin in South Dakota is based largely on the work of Flint (1955, Pl. 1), as modified in the southeastern part of the State by Tipton (1959, 1960), Steece (1959), and Steece *et al.* (1960). In North Dakota the position of the margin is based chiefly on the work of Benson (1953, p. 184-194) and Howard (1960, Pl. 1). Laird (this paper) believes that the limits of glaciation near the southern boundary of the State should be several kilometers west of the position as now shown; however, accurate determination of this line must await more detailed investigation. In Montana the limits are based on the work of Colton *et al.* (1961, map) and an unpublished map by Richmond and Lemke.

In North and South Dakota the drift sheet of advance 1 forms a northwest-trending belt, 90-160 km wide, west of the Coteau du Missouri. It also occurs in the Big Sioux River valley of eastern South Dakota. The till is thin or missing in most places owing to erosion or non-deposition. The advance of ice over much of these two states, especially west of the Missouri River, is revealed only by erratic boulders and a few stratified ice-contact deposits. East of the Missouri River, in areas where the till is thicker and more continuous, surface drainage generally is well integrated. In Montana the belt of exposed drift is generally narrower than in the other two States but otherwise is similar.

Present evidence based on three radiocarbon dates (Table 2) suggests that advance 1 occurred during the latter part of early Wisconsin time. One sample (W-990) gave a date of more than 38,000 years B.P. (before the present), one (W-1045) of $28{,}700 \pm 800$ years B.P. in southern North Dakota, and one (W-115) of more than 30,000 years B.P. in South Dakota. The drift of this advance is correlated with the "Altonian substage" of Frye and William (1963, p. 503) and the "Farmdale glacial substage" of Leighton (1960).

Richmond and Lemke on their unpublished map have correlated the continental drifts in Montana with alpine drift units in the adjacent Rocky Mountains. The drift of advance 1 correlates with that of the late stade of the Bull Lake Glaciation in the Glacier National Park area, Montana. Richmond (1960, p. 223) also believes that this drift correlates with the Iowan Till of Iowa, which has been dated as >29,000 years B.P. (Ruhe and Scholtes, 1959).

Advance 2

The limits of advance 2 are not known with certainty and are particularly difficult to distinguish from the limits of advance 3, in both South Dakota and southern North Dakota and also in Montana, where individual lobes advanced into the State from Canada. In South Dakota the

TABLE 2

Radiocarbon Dates[a]

Map No.[b]	Sample No.[c]	Location	Material dated	Lithology of enclosing material	Depth below surface (meters)	Collector	Date (years B.P.)
1	W-1189	NW¼ sec. 29, T. 97 N., R. 57 W., Hutchinson County, S.D.	Wood fragments	Till above and gravel below	57.6	R. A. Schoon 1961	12,050 ± 400
2	Y-595	SW¼ sec. 15, T. 98 N., R. 53 W., Turner County, S.D.	*Picea* or *Larix*	Outwash	9.6	G. A. Avery	12,760 ± 120
3	Y-452	Parker, S.D. NW¼ sec. 28, T. 99 N., R. 53 W., Turner County, S.D.	Spruce wood	Till	7.8	H. A. Mateer; G. A. Avery	12,330 ± 180
4	W-801	NW¼ sec. 36, T. 106 N., R. 58 W., Miner County, S.D.	Wood fragments	Silt lens in till	3.6	M. J. Tipton 1958	12,200 ± 400
5	W-1033	SE¼ sec. 3, T. 107 N., R. 62 W., Sanborn, S.D.	Mollusk shells	Lake sediments in outwash	0.6	F. V. Steece 1959	10,060 ± 300
6	W-987	NE¼ sec. 26, T. 108 N., R. 63 W., Jerauld County, S.D.	Wood fragments	60-cm soil(?) between two tills	12.3	Wayne McDaniel 1951	12,530 ± 350
7	W-983	NE¼ sec. 26, T. 108 N., R. 63 W., Jerauld County, S.D. (Same well as sample W-987)	Wood fragments	Soil(?) horizon between upper unoxidized till and lower oxidized till	12.3	L. W. Howells 1960	10,350 ± 300
8	W-115	Sec. 26, T. 110 N., R. 48 W., Brookings County, S.D.	Spruce wood	"Drift"	41.4	G. A. Avery	>30,000
9	W-1149	SE¼ sec. 36, T. 130 N., R. 68 W., McIntosh County, N.D.	Shells	Sand	3+	S. J. Tuthill; Lee Clayton	9,620 ± 350
10	W-974	NW¼ sec. 20, T. 132 N., R. 68 W., McIntosh County, N.D.	Shells-molluscan	Perched lake silts	0.6-2.4	John Bonneville	11,650 ± 310
11	W-1021	SE¼ sec. 9, T. 134 N., R. 68 W., Logan County, N.D.	Peat	Pitted outwash gravel	13.5	Lee Clayton	>38,000
12	W-1019	NW¼ sec. 20, T. 135 N., R. 67 W., Logan County, N.D.	Clam shells *in situ*	Sand and silt of stagnation drift	0.52	Lee Clayton	9,000 ± 300
13	W-954	SE¼ sec. 29, T. 137 N., R. 69 W., Stutsman County, N.D.	Clam shells	In clay with till at base	0.3	Charles Huxel; H. C. Winters	9,870 ± 290
14	W-990	SW¼ sec. 32, T. 135 N., R. 72 W., Logan County, N.D.	Black peat	Outwash sand and gravel	1.5+	Lee Clayton	>38,000
15	W-1045	NW¼ sec. 24, T. 134 N., R. 72 W., Logan County, N.D.	Peat underlying iron-cemented till	Gravels	1.5-3.3	John Bonneville	28,700 ± 800
16	W-542	SW¼ sec. 25, T. 138 N., R. 71 W., Kidder County, N.D.	Wood	Sand overlying till	3.6-4.5	R. W. Lemke	11,480 ± 300
17	W-956	SE¼ sec. 17, T. 139 N., R. 67 W., Stutsman County, N.D.	Clam shells	Stagnation moraine outwash	Near surface	Charles Huxel; H. C. Winters	11,070 ± 300
18	W-1020	NE¼ sec. 21, T. 141 N., R. 66 W., Stutsman County, N.D.	Wood	Outwash under till	15	R. W. Schmidt	>38,000
19	W-388	Moorhead, Clay County, Minn.	Wood	Lake clay	?	W. F. Libby	9,930 ± 280
20	W-993	C sec. 20, T. 140 N., R. 48 W., Cass County, N.D.	Wood	Silt and clay	8.4	John Brophy	9,900 ± 400
21	W-1005	SW¼ sec. 14, T. 150 N., R. 51 W., N.D.	Wood	Beach gravel	2.1	F. D. Holland; W. M. Laird	10,050 ± 300
22	W-900	NW¼ sec. 31, T. 152 N., R. 52 W.; 15 miles west of Grand Forks, N.D.	Wood	Lacustrine sand over till	3	R. W. Lemke	10,080 ± 280
23	W-723	NE¼ sec. 25, T. 152 N., R. 53 W.; 15 miles west of Grand Forks, N.D.	Wood	Lacustrine sand	3	R. W. Lemke	10,960 ± 300
24	W-402	SE¼ sec. 30, T. 146 N., R. 89 W., Mercer County, N.D.	Gastropods	Marl lens in till	?	W. E. Benson	11,220 ± 300

[a] Tenuous nature of stratigraphic information on glacial deposits makes correlation of radiocarbon-dated localities with numbered ice advances impractical at present.
[b] Keyed to Figure 1.
[c] W = U.S. Geological Survey
Y = Yale University

placement of the limits of advance 2 is based largely on the work of Flint (1955). The outer margin of the Bemis moraine marks the maximum advance in the eastern part of the State, but in North Dakota the limit of the advance is not well defined. The Long Lake and Zeeland end moraines of Clayton (1962, p. 26-30) probably mark the distal edge in southern North Dakota; farther north, the limit is tentatively placed at the outer margin of the Krem moraine.

Three separate ice lobes moved into Montana during advance 2. The easternmost lobe advanced southwestward around the east side of the Flaxville Plain. A second lobe advanced around the east side of the Sweetgrass Hills and was split into two sublobes by the Bearpaw and Little Rocky Mountains. The direction of advance of one sublobe east of the Bearpaw Mountains is well marked by a southeast-trending boulder train, 90 km long, heading at an isolated intrusive rock mass known as Snake Butte (Knechtel, 1942). The southeast direction of advance of this sublobe is further indicated by the presence of elongate southeast-trending drumlins. In contrast, similar drumlins formed by the other sublobe trend southwestward (Colton *et al.*, 1961). The third major lobe advanced south and southeastward around the west side of the Sweetgrass Hills, but left no well-defined terminal moraine.

Surface drainage on this drift is considerably less well integrated than that on the drift of advance 1. The ground moraine surface is characterized by numerous undrained depressions, most of which are shallow and have gently sloping sides. Local relief on some end moraines is as much as 15 m. The till in most places is 6-15 m thick, but locally it is considerably thicker.

The age of the drift is uncertain. In Iowa, Ruhe and Scholtes (1959) assigned the Bemis and Altamont moraines to the Cary Stade. On this basis, at least the outer belt of drift of advance 2 as outlined in part by the Bemis moraine in South Dakota (Fig. 1) would be Cary in age. In the Big Sioux Valley of South Dakota, Tipton (1958a, 1958b) and Steece (1958a, 1958b, 1958c) were unable to locate the break between drifts of Iowan and Tazewell age shown by Flint (1955, Pl. 1) in front of the Bemis moraine. They therefore assigned the glacial deposits in front of the Bemis moraine to the Iowan Stade.

The drift of advance 2 correlates with the early stade of the Pinedale Glaciation (Richmond, 1960) in the area just east of the Rocky Mountains in Montana. This is roughly correlative with the early part of the "Woodfordian stage" of Frye and Willman (1963).

ADVANCE 3

In North Dakota advance 3 is more conspicuously represented than any other glacial advance; its drift covers the entire Coteau du Missouri. The prominent Burnstad moraine (Fig. 1), which overlaps part of the Long Lake moraine of advance 2, is the terminal moraine in southern North Dakota, and the Alamo moraine marks the outer margin of the advance in the northwestern part of the State. The conspicuous Streeter moraine is interpreted as a recessional moraine of this advance.

Much of the prominence of the drift of this advance in northwestern North Dakota and northeastern Montana results from buried bedrock hills. In the southern part of North Dakota, however, this prominence reflects thick accumulations of drift (Clayton, 1962, p. 27).

Larger moraines, such as the Burnstad, commonly are made up of several sub-parallel, steep-sided ridges having a local relief of about 6 m. The Streeter moraine, a steep-sided ridge on which are superimposed small, looping, parallel ridges, is as much as 90 m high. Dead-ice moraine, also called stagnation moraine (Colton *et al.*, 1963; Clayton, 1962, p. 34) forms the topography characteristic of much of the Coteau du Missouri. Here, drainage is almost entirely unintegrated, local relief is sharp, and kettles and other undrained depressions are innumerable. Ice-contact faces locally give the surface a terraced appearance (Clayton, 1962). This type of drift surface, we believe, resulted from large-scale glacial stagnation, in which superglacial till has collapsed into depressions left by isolated melting ice blocks.

The outer margin of the advance in South Dakota has not been determined. It seems likely, however, that an ice lobe moved southward into the James River lowland during this advance. Tentative correlations by Flint (1955, p. 118) suggest that either the Altamont moraine or the Gary moraine marks the terminus of advance 3 in northeastern South Dakota.

Radiocarbon dates confound attempts to date the drift sheet formed by advance 3. Material from two samples (W-1020 and W-1021) give dates greater than 38,000 years. These samples may have been contaminated by old carbon, possibly detrital lignite fragments of Tertiary age. The remainder of the dates fall largely between 10,000 and 12,000 years. Almost all of those about 10,000 years were obtained from organic matter in outwash overlying till and therefore are minimum ages. Also, several samples were obtained from areas of dead-ice moraine and may represent outwash from isolated, partially or wholly buried blocks of ice that survived melting through a considerable period of time—possibly well into the Two Creeks Interstade.

If the Altamont or Gary moraines mark the outer limits of advance 3, then the drift sheet should be assigned to the Cary Stade on the basis of work by Wright and Ruhe (this volume) and would correlate with the "Woodfordian stage" of Frye and Willman (1963).

ADVANCE 4

The terminus of advance 4 is marked by the Big Stone moraine in South Dakota and by the Britton, Oakes, Kensal, and Grace City moraines in North Dakota. The drift of this advance is exposed in a belt mostly 25 to 65 km wide. The orientation of drumlins, washboard moraines, and end moraines indicates that the ice advanced and retreated in two lobes, one from the northwest and the other from the northeast. A narrow interlobate area between the Grace City moraine and the Kensal moraine marks the junction of these lobes.

Leighton (1957, p. 1037-1038) suggested that the Big Stone moraine marks the maximum advance of ice of Valders age. If true, all the drift of advance 4 would belong to the Valders Stade. Elson (1957, p. 999-1002), however, believed that Valders ice did not reach northwestern Minnesota or North Dakota, an interpretation in accord with work done in Minnesota by Wright and Ruhe (this volume). Studies in Saskatchewan by several workers, as summarized by H. A. Roed (written communication), indicate that an ice lobe, which a radiocarbon date suggests may have been Valders ice, did not reach into North Dakota. W. M. Laird, however, believes that all drift sheets younger than that of advance 3 are recessional moraines from this main advance. Resolution of this problem must await additional regional studies and radiocarbon dates.

ADVANCE 5

Moraines of advance 5 sharply truncate the moraines and other drift of advance 4. The glacier of advance 5 moved into North Dakota as two distinct lobes split by the Turtle Mountains, although the mountains were probably at least partly overridden during the maximum stand of the ice. West of the Turtle Mountains, the Souris River lobe (Lemke, 1960, p. 111) advanced southeastward and formed the Martin terminal moraine. East of the Turtle Mountains, the Leeds lobe advanced south and southwestward and formed the Heimdal and Cooperstown terminal moraines. We believe that the North Viking moraine is a recessional moraine of this lobe. The trends of parallel linear drumlins (Lemke, 1958), some of which are several kilometers long, and of arcuate washboard moraines provide a good record of the direction of advance and retreat of the two lobes.

Drainage on drift of advance 5 is almost completely unintegrated. Kettle holes are sharply defined, and many have steep walls. All the till is calcareous to the surface and consists in most places of about 25% clay, 40% silt, 30% sand, and 5% gravel (Lemke, 1960, p. 48). Erratic boulders, commonly as much as 1 m long, are mostly of granite or limestone from Canada. The till of the Souris River lobe contains abundant lignite chips, whereas the till of the Leeds lobe generally does not.

During glacier recession, Glacial Lake Souris and Glacial Devils Lake came into existence. Drainage of Glacial Devils Lake was for a short time down the James River; later drainage was into Glacial Lake Agassiz via spillways and the Sheyenne River (Aronow, 1963). Glacial Lake Souris expanded northward into Saskatchewan as the ice front receded into Canada. Early drainage was down the Sheyenne River into the southern part of Glacial Lake Agassiz. Later, when the area north of the Turtle Mountains became ice-free, Glacial Lake Souris drained northward into Manitoba and entered Glacial Lake Agassiz at the northern border of North Dakota.

If Valders ice did not extend into North Dakota, the drift of advance 5 must have been deposited during late Mankato time. Although this drift represents a distinct advance, the glacier may have retreated only a few kilometers or tens of kilometers between advances 4 and 5.

ADVANCE 6

The discontinuous, looping Edinburg moraine in northeastern North Dakota defines the outer margin of the last glacier to occupy the State. The fact that the generally north-south trending moraine definitely truncates a series of northeast-southwest trending arcuate washboard moraines, which indicate ice recession toward the northwest, suggests that the Edinburg moraine does not represent merely a halt in the recession of the glacier of advance 5. The Holt moraine in Minnesota (Leverett, 1932, p. 117) and the Darlingford moraine in Manitoba seem to be continuations of the Edinburg moraine.

The Edinburg moraine is poorly defined because much of it is thinly veneered by lacustrine deposits of Glacial Lake Agassiz II of Elson (1938), who believed this stand of the lake was formed by melt waters draining from Valders ice in Manitoba. If so, advance 6 is probably of Mankato age. This implies that advances 3 to 6 occurred during a relatively short time interval and that succeeding advances probably were separated by a glacier recession of only a few kilometer or tens of kilometers. If, on the other hand, drift of advances 4 through 6 was deposited during all of post-Cary time, then deposition of drift during these ice advances might have occurred over a considerably longer period of time.

Postglacial History of Glaciated Area

The topography in the area covered by the last five glacial advances has been little modified since deglaciation. Nearly all glacial features are well preserved, and there has been little or no downcutting by present-day streams. Sand dunes have formed on some of the glacial-lake floors and outwash plains in North and South Dakota, but most of these probably formed soon after deglaciation of the respective area.

Periglacial features are nearly all confined to the drift sheet of advance 1 and adjacent nonglaciated areas. Frost-induced involutions in gravels and frost wedges in weathered bedrock form an irregular polygonal pattern in some areas in central Montana (Schafer, 1949). Similar features are sparingly present in drift of advance 1 in the same area but are lacking in drift of the younger advances.

Isolated remnants of two postglacial ash falls, described by Powers and Wilcox (1964), are preserved in the western part of the report area of Montana (see Wilcox, this volume). The older of these, derived from Glacier Peak volcano in Washington, is found on Pinedale drift (M. R. Mudge, oral communication) near Augusta, Montana. The younger ash, which came from the Mount Mazama (Crater Lake) eruption in Oregon about 6,600 years ago (W-858) is present at Galata, Montana (Horberg and Robie, 1955; Powers and Wilcox, 1964). In a roadcut 32 km southwest of Great Falls, Montana, Lemke discovered a well-developed paleosol that directly underlies the Mazama ash and in turn is underlain by alluvial deposits. This paleosol is believed to be of pre-Altithermal age. The pre-ash alluvium, as well as a post-ash alluvium, is widespread in the Great Falls area. These alluvial deposits once partially filled many minor tributaries incised into the older drifts but are now actively being removed by headward erosion.

Unglaciated Areas

Several mountain outliers and parts of the high plains in the glaciated area of Montana were too high to be covered by continental ice sheets. The mountain outliers, which reach altitudes of 1,800 to 2,250 m (6,000 to 7,500 ft), consist chiefly of volcanic and sedimentary rocks. The Flaxville Plain, a driftless area in northeastern Montana, is surfaced by the Flaxville Gravel of Miocene or Pliocene age underlain by strata of early Tertiary age.

Areas beyond the limits of glaciation in the three States include approximately the western half of South Dakota, the southwestern corner of North Dakota, and approximately the southern half of Montana (Fig. 1).

The unglaciated area in South Dakota lies west of the Missouri River and ranges in altitude from about 600 m (2,000 ft) in the east to 900 m (3,000 ft) in the west. Isolated to closely spaced hills and buttes rise to altitudes of more than 1,080 m (3,600 ft). Nearly all the streams have cut canyons 60 m or more in depth. Extensive badland topography has developed, particularly along the Cheyenne and White Rivers; the most spectacular exposures are east of the Black Hills in the Badlands National Monument. Three to four sets of discontinuous, nonglacial fluvial terraces, which range in age from middle Tertiary to Pleistocene, flank most of the main streams. The terraces are underlain by gravels derived in part from the Black Hills, which rise to an altitude of 2,172 m (7,241 ft) and consist of sedimentary and crystalline rocks.

Sand dunes derived from underlying sandy bedrock abound in the extreme south-central part of South Dakota, where they form an undulating surface of low relief. Springs flow from their northern margin.

The unglaciated upland of North Dakota is a continuation of the upland in South Dakota. It slopes gently eastward at altitudes mostly between 840 and 750 m (2,800 and 2,500 ft). Isolated erosion remnants, which reach a maximum altitude of 1,060 m (3,530 ft), surmount the upland. The gently rolling upland surface appears to be in a stage of late maturity. Drainage patterns are well developed, but at present there is relatively little surface runoff because of the semiarid climate. Extensive badlands have formed in many places adjacent to the Missouri and Little Missouri Rivers. The upland is modified by a number of wide, southeast-trending valleys, some of which, now devoid of streams, were glacial melt-water diversion channels utilized by the Missouri, Yellowstone, and Little Missouri Rivers and other streams. Some valleys were probably pre-Pleistocene stream courses, inasmuch as they contain extensive terrace gravels of lithology different from the gravels of glacial origin. These gravels may have been derived from the Black Hills or from outcrops of conglomerate in the Tertiary bedrock.

The unglaciated eastern part of Montana is much the same as the unglaciated parts of North Dakota and South Dakota. The Yellowstone River and its tributaries have cut trenches 1.5-6.5 km wide and a few hundred meters deep. Episodes of cutting and filling have produced several terraces. Broad troughs through the divides between adjacent streams were utilized by water draining from proglacial lakes.

Gravel deposits of several ages from stream terraces along the Yellowstone River in southeastern Montana. Gravel deposits occur in isolated patches along the western side of the lower reaches of the Yellowstone valley, 200 to 240 m above the present valley bottom (Howard, 1960, p. 10). These gravels form the No. 1 bench of Alden (1932, map), who showed remnants of these gravels extending up the Yellowstone valley past Billings nearly to the southern border of Montana. They are now regarded as being equivalent to the Miocene or Pliocene Flaxville Gravel of northeastern Montana. Younger gravels of similar lithology occur in terraces at a lower level along the Yellowstone valley and some of its tributaries. These gravels, which form the No. 2 bench of Alden (1932, p. 44), also extend along the Yellowstone valley nearly to the southern border of the State. In the lower reaches of the Yellowstone valley they are called the Cartwright Gravel by Howard (1960, p. 19-21). Although adequate evidence of the age is lacking, they probably are of Pleistocene age and were derived from the southwest in pre-Wisconsin time. Still lower and younger terrace gravels in the Yellowstone valley, mapped as bench No. 3 by Alden, have been named the Crane Creek Gravel of Pleistocene age in the lower reaches of the Yellowstone by Howard. The Crane Creek Gravel, whose surface is about 7 m above the present floodplain of the Yellowstone, is superficially similar to the other nonglacial or largely nonglacial gravels of the region and does not contain glacial pebbles south of the glacial limit. Correlative gravel terraces along the Missouri River east of Glasgow are overlain by till that probably was deposited by glacial advance 2.

Mountain outliers in the northern Great Plains of west-central Montana include the Highwood, the Bearpaw, and the Little Rocky Mountains (Fig. 1) as well as the Judith, North and South Moccasin, and Big Snowy Mountains (not shown on map) which lie in the nonglaciated High Plains area to the south. Three well-developed pediments of different ages flank the mountain outliers in places and merge with alluvial fans and terraces in the adjacent plains. These pediments correspond to benches No. 1, 2, and 3 of Alden (1932, Pl. 1). We believe bench No. 1 is equivalent to the Flaxville Plain and benches No. 2 and 3 are Pleistocene in age.

Alden (1932) has correlated a terrace along the Sun River between Great Falls and Augusta with his bench No. 3. This terrace was deposited by melt-water from alpine ice during the Bull Lake Glaciation (M. R. Mudge, oral communication). Immediately west of Augusta the terrace is overlain by till deposited by alpine ice of the early stade of Pinedale Glaciation. The gravels of this terrace were traced by Mudge and Lemke down the Sun River valley and were found to overlie and partly interfinger with lacustrine deposits laid down in Glacial Lake Great Falls during the time of advance 1. If these correlations are correct, the Sun River terrace (Alden's bench No. 3) is correlative with the Iowan till of Ruhe and Scholtes (1959) in Iowa.

References

Alden, W. C., 1932, Physiography and glacial geology of eastern Montana and adjacent areas: U.S. Geol. Surv. Prof. Pap. 174, 133 p.

Aronow, Saul, 1963, Late Pleistocene glacial drainage in the Devils Lake region, North Dakota: Geol. Soc. Amer. Bull., v. 74, p. 859-873

Aronow, Saul, Dennis, P. E., and Akin, P. D., 1953, Geology and ground-water resources of the Minnewuakan area, Benson County, North Dakota: North Dakota Geol. Surv. Ground-Water Stud. 19, 125 p.

Benson, W. E., 1953, Geology of the Knife River area, North Dakota: U.S. Geol. Surv. Open-File Rep., 323 p.; also Yale Univ., Ph.D. thesis

Clayton, Lee, 1962, Glacial geology of Logan and McIntosh Counties, North Dakota: North Dakota Geol. Surv. Bull. 37, 84 p.

Colton, R. B., Lemke, R. W., and Lindvall, R. M., 1961,

Glacial map of Montana east of the Rocky Mountains: U.S. Geol. Surv. Misc. Geol. Inv. Map I-327
—— 1963, Preliminary glacial map of North Dakota: U.S. Geol. Surv. Misc. Geol. Inv. Map I-331
Elson, J. A., 1957, Lake Agassiz and the Mankato-Valders problem: Science, v. 126, p. 999-1002
—— 1958, Pleistocene history of southwestern Manitoba, *in* Guidebook 9th Ann. Field Conf., Midwestern Friends of the Pleistocene: North Dakota Geol. Surv. Misc. Ser. 10, p. 62-73
Flint, R. F., 1949, Pleistocene drainage diversions in South Dakota: Geogr. Annaler, v. 31, p. 56-74
—— 1955, Pleistocene geology of eastern South Dakota: U.S. Geol. Surv. Prof. Pap. 262, 173 p.
Flint, R. F., *et al.*, 1959, Glacial map of the United States east of the Rocky Mountains [scale 1:1,750,000]: Geol. Soc. America
Frye, J. C., and Willman, H. B., 1960, Classification of the Wisconsinan Stage in the Lake Michigan glacial lobe: Illinois State Geol. Surv. Circ. 285, 16 p.
—— 1963, Development of Wisconsinan classification in Illinois related to radiocarbon chronology: Geol. Soc. Amer. Bull., v. 74, p. 501-505
Hedges, L. S., in preparation, Pleistocene geology of Beadle County, South Dakota: South Dakota Geol. Surv. Rep. Inv.
Horberg, C. L., and Robie, R. A., 1955, Postglacial volcanic ash in the Rocky Mountain piedmont, Montana and Alberta: Geol. Soc. Amer. Bull., v. 66, p. 949-955
Howard, A. D., 1960, Cenozoic history of northeastern Montana and northwestern North Dakota with emphasis on the Pleistocene: U.S. Geol. Surv. Prof. Pap. 326, 107 p.
Knechtel, M. M., 1942, Snake Butte boulder train and related glacial phenomena, north-central Montana: Geol. Soc. Amer. Bull., v. 53, p. 917-935
Leighton, M. M., 1933, The naming of the subdivisions of the Wisconsin glacial age: Science, v. 77, p. 168
—— 1956, Radiocarbon dates and Pleistocene chronological problems in the Mississippi Valley region; a reply: J. Geol., v. 64, p. 193-194
—— 1957, Radiocarbon dates of Mankato drift in Minnesota: Science, v. 125, p. 1037-1039
—— 1960, The classification of the Wisconsin glacial stage of north-central United States: J. Geol., v. 68, p. 529-552
Lemke, R. W., 1958, Narrow linear drumlins near Velva, North Dakota: Amer. J. Sci., v. 256, p. 270-283
—— 1960, Geology of the Souris River area, North Dakota: U.S. Geol. Surv. Prof. Pap. 325, p. 138
Lemke, R. W., and Kaye, C. A., 1958, Two tills in the Donnybrook area, North Dakota, *in* Guidebook 9th Ann. Field Conf., Midwestern Friends of the Pleistocene: North Dakota Geol. Surv. Misc. Ser. 10, p. 93-98
Leverett, Frank, 1932, Quaternary geology of Minnesota and parts of adjacent states: U.S. Geol. Surv. Prof. Pap. 161, 149 p.
Powers, H. A., and Wilcox, R. E., 1964, Volcanic ash from Mount Mazama (Crater Lake) and from Glacier Peak: Science, v. 144, p. 1334-1336
Richmond, G. M., 1960, Correlation of alpine and continental glacial deposits of Glacier National Park and adjacent High Plains, Montana: U.S. Geol. Surv. Prof. Pap. 400-B, p. 223-224
——, this volume, Glaciation of the Northern Rocky Mountains
Ruhe, R. V., and Scholtes, W. H., 1959, Important elements in the classification of the Wisconsin glacial stage—a discussion: J. Geol., v. 67, p. 585-598
Schafer, J. P., 1949, Some periglacial features in central Montana: J. Geol., v. 57, p. 154-174
Simpson, H. E., 1960, Geology of the Yankton area, South Dakota and Nebraska: U.S. Geol. Surv. Prof. Pap. 328, 124 p.
Stalker, A. MacS., 1963, Quaternary stratigraphy in southern Alberta: Geol. Surv. Canada, Pap. 62-34, p. 1-52
Steece, F. V., 1958a, Geology of the Watertown quadrangle, South Dakota: South Dakota Geol. Surv., map and text
—— 1958b, Geology of the Hayti quadrangle, South Dakota: South Dakota Geol. Surv., map and text
—— 1958c, Geology of the Estelline quadrangle, South Dakota: South Dakota Geol. Surv., map and text
Steece, F. V., 1959, Geology of the Sioux Falls quadrangle, South Dakota: South Dakota Geol. Surv., map and text
Steece, F. V., and Howells, L. W., in preparation, Geology and ground water supplies in Sanborn County, South Dakota: South Dakota Geol. Surv. Bull. 17
Steece, F. V., Tipton, M. J., and Agnew, A. F., 1960, Glacial geology of the Coteau des Prairies, South Dakota: Guidebook 11th Ann. Field Conf. Midwestern Friends of the Pleistocene, South Dakota, 21 p.
Thwaites, F. T., 1943, Pleistocene of part of northeastern Wisconsin: Geol. Soc. Amer. Bull., v. 54, p. 87-144
—— 1946, Outlines of glacial geology: Ann Arbor, Mich., Edwards Bros., Inc., 119 p.
Tipton, M. J., 1958a, Geology of the Still Lake quadrangle, South Dakota: South Dakota Geol. Surv., map and text
—— 1958b, Geology of the South Shore quadrangle, South Dakota: South Dakota Geol. Surv., map and text
—— 1960, A new glacial drift sheet in South Dakota?: South Dakota Acad. Sci. Proc., v. 38, p. 45-48
—— 1959, Geology of Dell Rapids quadrangle, South Dakota: South Dakota Geol. Surv., map and text
Townsend, R. C., and Jenke, A. L., 1951, The problem of the origin of the Max moraine of North Dakota and Canada: Amer. J. Sci., v. 249, p. 842-858
Tuthill, S. J., Laird, W. M., and Frye, C. I., in press, Fossil molluscan fauna from the upper terrace of the Cannonball River, Grant County, North Dakota: North Dakota Acad. Sci. Proc., v. 18
Warren, C. R., 1952, Probable Illinoian age of part of the Missouri River, South Dakota: Geol. Soc. Amer. Bull., v. 63, p. 1143-1155
White, E. M., 1964, Post-Illinoian age for Missouri River in South Dakota proposed from relationship to a White River terrace: Amer. J. Sci., v. 262, p. 494-496
Wilcox, R. E., this volume, Volcanic-ash chronology
Witkind, I. J., 1959, Quaternary geology of the Smoke Creek-Medicine Lake-Grenora area, Montana and North Dakota: U.S. Geol. Surv. Bull. 1073, 80 p.

Winters, H. A., 1963, Geology and ground water resources of Stutsman County, North Dakota; Pt. I, Geology: North Dakota Geol. Surv. Bull. 41, 84 p.
Wright, H. E., Jr., 1957, Radiocarbon dates of Mankato drift in Minnesota: Science, v. 125, p. 1038-1039
Wright, H. E., Jr., and Ruhe, R. V., this volume, Glaciation of Minnesota and Iowa
Zumberge, J. H., and Wright, H. E., Jr., 1956, The Cary-Mankato-Valders problem: Geol. Soc. Amer. Guidebook Ser., Field Trip 3, Glacial geology, eastern Minnesota, p. 65-81

SUMMARY

The area discussed includes about 580,000 km^2 (225,000 sq. miles) of South Dakota, North Dakota, and Montana in the Central Lowland and northern Great Plains physiographic provinces of the United States. Most of the discussion concerns the area covered predominantly by Wisconsin continental glacial deposits. The drift surface is mostly gently rolling, and it slopes eastward from an altitude of about 1,200 m in northwestern Montana to about 300 m in eastern North and South Dakota. Present drainage is into the Gulf of Mexico by way of the Missouri River, its tributaries, and the Mississippi River. In preglacial times, however, all streams north of the Bad River in South Dakota drained northward into Hudson Bay. Bedrock consists chiefly of shales, siltstones, and sandstones of Cretaceous and Tertiary ages.

Six distinct advances of continental glaciers are believed to have occurred in the region during Wisconsin time. Because all the drift deposited is very similar in appearance and lithology, and because only a few radiocarbon age dates are available, the correlation of these advances with the Pleistocene stratigraphic sequence of the midwestern United States is uncertain. The advances are numbered arbitrarily from oldest to youngest and are so described and designated in this report. Many glacial lakes existed in the area during Pleistocene time, including Devils Lake and Lakes Agassiz, Souris, Dakota, Glendive, Jordan, Musselshell, Great Falls, Choteau, and Cut Bank. Pleistocene stream terraces, pediments, and deposits of windblown sand and silt occur in many places in the unglaciated portion of the report area.

GLACIATION OF MINNESOTA AND IOWA

H. E. WRIGHT, JR.,[1] R. V. RUHE[2]

MULTIPLE GLACIATION has dominated the development of the landscape of Minnesota and Iowa, leaving an almost continuous cover of till and such related sediments as glaciofluvial gravels, glacial-lake deposits, and upland loess. Bedrock is extensively exposed only in northeasternmost Minnesota near Lake Superior and the Canadian border, although the drift has been stripped from the deep postglacial valleys of the Mississippi and Missouri Rivers and their tributaries in Iowa and southern Minnesota.

Glacial features of Wisconsin age form the surface of most of Minnesota and northern Iowa, and their relations reveal a complex history of ice advance and retreat. In southern Iowa, on the other hand, all of the major pre-Wisconsin glacial and interglacial episodes are represented by glacial drifts, loesses, interglacial paleosols, and buried erosion surfaces. Together the two states thus record practically the entire range of activity of the main Pleistocene continental ice sheet.

The pioneer glacial studies in Minnesota were made by Winchell and Upham (Winchell, 1884, 1888) and in Iowa by McGee (1891) and Bain (1897). These men delineated the major moraines and ice lobes and demonstrated multiple glaciation. Among the early studies of special interest was the work on Glacial Lake Agassiz in northwestern Minnesota by Upham (1896) and on the retreat of St. Anthony Falls on the Mississippi River near Minneapolis by Winchell (1888, p. 313-341).

In the next generation, Leverett mapped the entire state of Minnesota (1929, 1932) with assistance locally of F. W. Sardeson. In Iowa the leading workers were Alden and Leighton (1917), Carman (1931), and Kay (Kay and Apfel, 1929; Kay and Graham, 1943). In the last two decades the glacial studies in Minnesota have been supported largely by the Minnesota Geological Survey, those in Iowa by the Soil Conservation Service and the Iowa Geological Survey.

PRE-WISCONSIN GLACIATIONS

In Minnesota, drifts presumably of pre-Wisconsin age are exposed in iron-ore pits and deep river cuts, but they cannot be traced with continuity or correlated easily with pre-Wisconsin drifts elsewhere. In Iowa, however, the older drifts are well exposed on ridge crests, hill slopes, river banks, and roadcuts in the southern part of the state (Fig. 1), and they have provided the basis for several of the major subdivisions of the Pleistocene sequence. Standard sections of the Nebraskan and Kansan glacial stages and the type locality of the Aftonian interglacial stage are all located near Afton Junction in southern Iowa. The type locality of the Yarmouth interglacial stage is at Yarmouth in southeastern Iowa.

NEBRASKAN

Nebraskan till nowhere forms the upland surface in Iowa and Minnesota, being covered by Kansan drift or younger drifts and loesses. It is a loam to clay-loam till with 45% carbonate rocks, 3% other sedimentary rocks, and 52% igneous and metamorphic rocks brought from Minnesota and Canada. It has been studied mainly in the southern half of Iowa, where its average thickness is estimated as 30-50 m (Kay and Apfel, 1929, p. 181).

AFTONIAN

In the gravel pits of the Afton-Thayer area, Union County (Kay and Apfel, 1929, p. 185), the gravels between the Kansan and Nebraskan tills were called Aftonian interglacial gravels for many years. They are now known to contain both Nebraskan and Kansan glacial gravels, but in the vicinity the Aftonian paleosol ("Nebraskan Gumbotil") separates the two tills. The Aftonian paleosol is either the B-horizon of a Planosol or Gray-Brown Podzolic soil or the A3- and B-horizons of a Humic-Gley soil or Grumosol (Ruhe, this volume). This "gumbotil" grades downward through yellowish-brown non-calcareous till ("oxidized and leached zone" of Kay and Apfel, 1929) to yellowish-brown calcareous till ("oxidized and unleached zone") and thus to the dark-gray calcareous parent material ("unoxidized and unleached zone").

Associated with the Aftonian paleosol at many localities is peat and muck. Lake clays occur above the paleosol and under Kansan till (Ruhe, 1954b). At other localities erosion surfaces separating the Nebraskan and Kansan tills represent the Aftonian interglacial age (Ruhe, 1954b).

KANSAN

Kansan till is indistinguishable lithologically from Nebraskan. Its average thickness has been estimated as 20 m (Kay and Apfel, 1929, p. 256). In south-central Iowa it is covered by thin Wisconsin loess on flat divides, but the till is exposed on valley-flanking pediments that were eroded in at least two post-Kansan cycles along dissecting streams (Ruhe, 1956).

YARMOUTH

The "Kansan gumbotil" is a Yarmouth paleosol with average thickness of 3.3 m where buried by Illinoian drift in southeasternmost Iowa (Kay and Apfel, 1929, p. 259). In

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Shelby County, southwestern Iowa, a similar paleosol, which is overlain by Loveland loess of Illinoian age, has an average thickness of 1.7 m (R. V. Ruhe, unpublished). Of the 50 major cuts along the new route of the Rock Island Railroad in southwestern Iowa, Yarmouth paleosol is found on Kansan till in 5 cuts and on lake clay in 17 cuts (Ruhe, 1954b), and Yarmouth interglacial deposits crop out in 13 cuts. In other cuts, lake clay occurs on top of Yarmouth paleosol at elevations between 1,266 and 1,279 ft above sea level (384 and 388 m) and is overlain by Loveland loess—apparently a late-Yarmouth lake on the Kansan till plain. Loveland loess is also found in the area at elevations below the lake clay, so the late Yarmouth was also a time of erosion, and the major drainage net of southeastern Iowa dates from this time. Similar pre-Illinoian erosion is known from southeastern Iowa (Kay and Apfel, 1929, p. 268).

ILLINOIAN

Illinoian till occurs only in a narrow belt close to the Mississippi River in southeastern Iowa (Kay and Graham, 1943, p. 15-44). It is similar in composition to the Kansan and Nebraskan tills, even though its eastward continuity into Illinois implies that it was deposited by ice from the Lake Michigan basin rather than by ice from western Minnesota. This ice from the northeast displaced the Mississippi River westward into the Goose Lake and Leverett Channels in Clinton and Lee Counties, Iowa, and dammed the Iowa and Cedar Rivers to form Glacial Lake Calvin (Schoewe, 1920). After ice retreat the Mississippi River shifted its course to the east.

Illinoian drift of the Superior lobe is reported from southeastern Minnesota (Leverett, 1932; Ruhe and Gould, 1954), but its correlation is not certain. In adjacent Wisconsin this

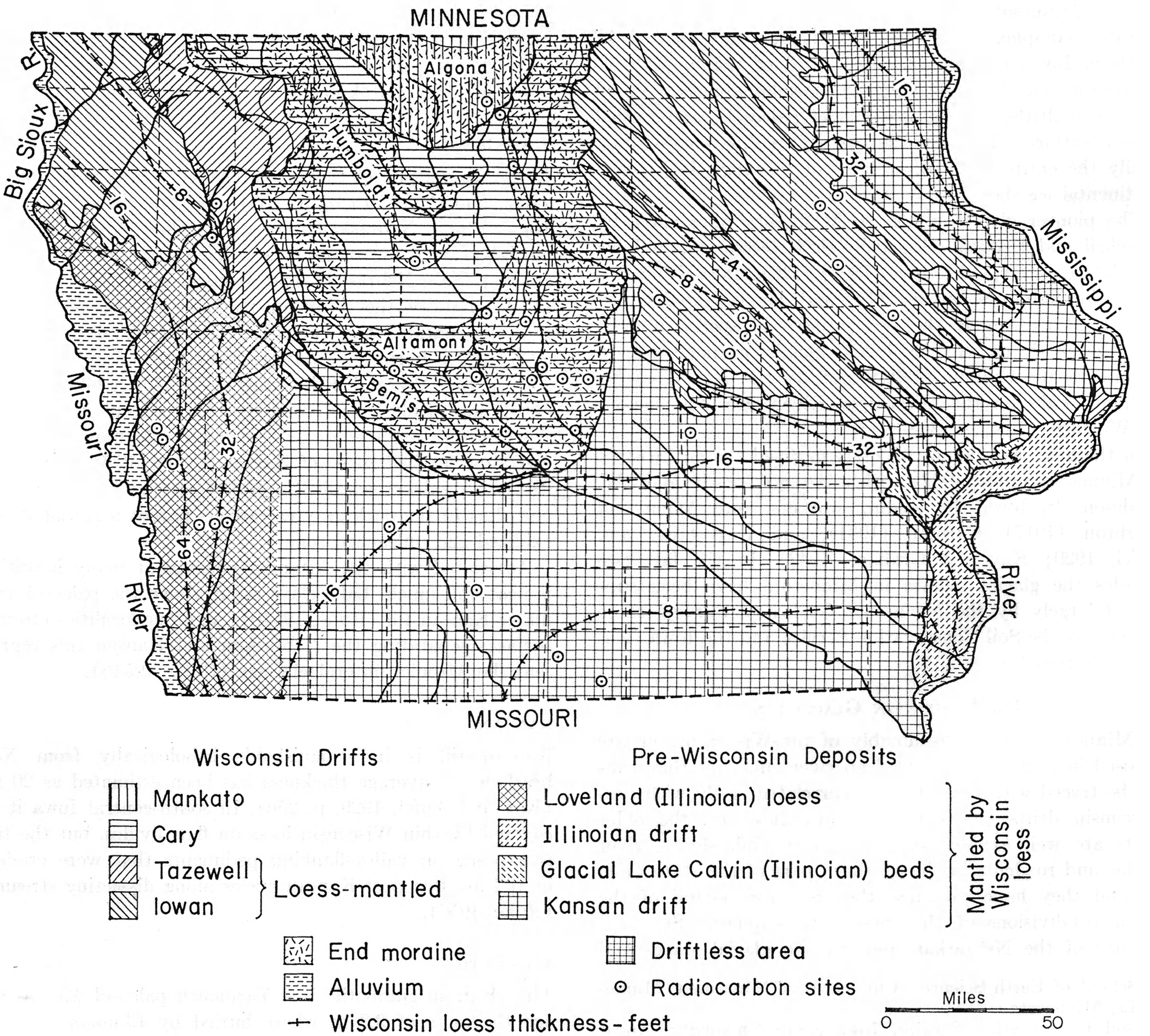

Figure 1. Pleistocene deposits in Iowa.

drift is correlated as Rockian (=Altonian = early Wisconsin) (Frye *et al.*, this volume).

Loveland loess, which is also a record of the Illinoian Glacial Age, has its type locality in southwestern Iowa. It was originally considered to be aqueous (Shimek, 1909), but it has since been demonstrated to be eolian (Kay, 1928). Along the new cuts of the Rock Island Railroad in southwestern Iowa, the Loveland loess decreases eastward in thickness (Fig. 2) and in coarse-silt content in the same manner as the overlying Wisconsin loess (Ruhe, 1954a, p. 665). Both sediments were deposited in many topographic positions through wide elevations.

The Loveland loess in Iowa is restricted to a band 50-125 km broad east of the Missouri River, but it extends southward along the river through Missouri to the Mississippi River, where it correlates with the pro-Illinoian loess of Illinois (Leighton and Willman, 1950, p. 601). It does not extend directly across Iowa to the area of Illinoian till, as believed by Kay and Graham (1943, p. 65), who erroneously interpreted the basal increment of the widespread Wisconsin loess as Loveland.

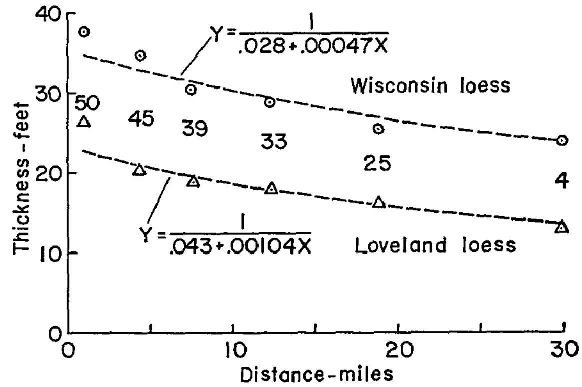

Figure 2. Thicknesses of Wisconsin and Loveland loesses from west (cut 50) to east (cut 4) along Rock Island Railroad, Pottawattamie County, southwest Iowa.

SANGAMON

Gumbotil and underlying weathering horizons on Illinoian till in southeastern Iowa are overlain by Wisconsin loess and thus record the Sangamon Interglacial Age. The Sangamon paleosol on Loveland loess exposed along the Rock Island Railroad cuts in southwestern Iowa increases in intensity of development as the loess thins from 8 to 5 m over a distance of 27 km (Ruhe, this volume). Comparable degree in intensification is developed in the soils on Wisconsin loess only with a thinning from 12 to 2.4 m over a distance of 203 km.

As the blanket of Loveland loess on the Yarmouth paleosol thins eastward across Iowa, its surficial Sangamon paleosol merges laterally with the emerging Yarmouth paleosol, which is formed on Kansan till. Thus in the large area of south-central Iowa between the Illinoian deposits on the west (Loveland loess) and on the east (till), the gumbotil on Kansan drift and under Wisconsin loess is actually a compound Yarmouth-Sangamon paleosol (Ruhe, 1956, p. 445).

This Yarmouth-Sangamon paleosol and the subjacent Kansan till in southern Iowa is cut by a widespread erosion surface in the form of valley-flanking pediments. The surface commonly has a stone-line and overlying pedi-sediment, which in turn has a paleosol that may extend through the stone-line and into the subjacent till. In southwestern Iowa this erosion surface bevels Loveland loess and its surficial Sangamon paleosol, which in turn is buried by basal Wisconsin loess. The widespread pediment is therefore late Sangamon in age. At places it descends to valley-slope fans and alluvial plains that stand well above present valley floors (Ruhe, 1960, p. 167). Elements of this late-Sangamon surface and its associated paleosols have been identified beneath younger Wisconsin drifts in northern Iowa.

THE IOWAN PROBLEM

The Iowan Stage or Substage has had a harried history. First described by McGee (1891) from a type locality at Doris Station in northeastern Iowa, it was considered for many years a separate stage of the Pleistocene (Leverett, 1926; Kay and Apfel, 1929, p. 17), being represented by till in eastern (Calvin, 1911; Alden and Leighton, 1917) and western (Carman, 1931) Iowa and by loess in Iowa and also in Illinois.

Another view, that the Iowan was the first substage of the Wisconsin, was championed by Leighton (1933), who thereupon introduced geographical names (Tazewell, Cary, Mankato) for Wisconsin subdivisions that had previously been termed Early, Middle, and Late Wisconsin. Leverett (1939), in one of his last publications, grudgingly yielded to the new classification, and for many years the Iowan stood as the earliest Wisconsin substage, although Leighton himself (Leighton and Willman, 1950, p. 602) later inserted the Farmdale substage beneath the Iowan on the basis of loess stratigraphy in Illinois.

Radiocarbon dates of coniferous wood extracted from presumed Iowan till in northeastern Iowa proved to be >29,000 to >37,000 years old, leading Ruhe and Scholtes (1959, p. 592) to suggest that the so-called Iowan till was pre-Wisconsin or at least that it was older than the Farmdale, which in its type area is 22,000-29,000 years old (Frye *et al.*, this volume). Current detailed geomorphic and stratigraphic studies, supported by drilling, by Ruhe and associates in the type area of Iowan till indicate that the landscape for the Iowan drift is a multi-leveled sequence of erosion surfaces and that many of these levels cut into Kansan and Nebraskan till. Thus the radiocarbon dates mentioned may refer to pre-Iowan rather than Iowan drift.

Drill holes on the summits of hills in areas of Kansan inliers and paha in the Iowan area (Scholtes, 1955) penetrate Wisconsin loess, Yarmouth-Sangamon paleosols, Kansan till, Aftonian paleosols, and Nebraskan till successively downward. In traverses drilled down the sloping flanks of the hills and paha, the same stratigraphic sequence occurs. On the adjacent, lower-lying so-called "typical Iowan drift plain" that surround the inliers and paha, the same deposits and horizons are identified in drill cores along traverses as long as 30 km. Such studies in Tama County and adjacent parts of Grundy and Blackhawk Counties fail to show the

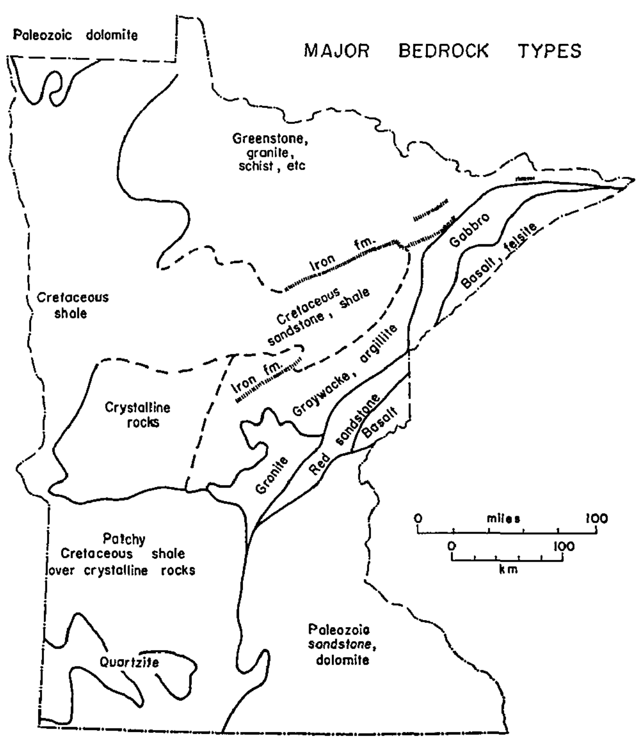

Figure 3. Generalized map of the bedrock of Minnesota.

presence of Iowan drift around the inliers and pahas. The paleosol and till cores of these features are not pre-Iowan topographic highs around which Iowan ice advanced but did not override, as previously proposed (Scholtes, 1955, p. 196-198). Instead the cores of inliers and pahas are erosion remnants on interstream divides (T. E. Fenton and R. V. Ruhe, *unpublished*).

The problem of the Iowan drift can only be resolved by detailed geomorphic and stratigraphic studies, the latter mainly in the subsurface. In the current work in the type area in northeastern Iowa, Iowan drift has yet to be identified.

The loess that continuously covers the Iowan till was formerly termed Peorian loess (Kay and Graham, 1943, p. 156) but is now called simply Wisconsin loess (Ruhe, 1954a). Its thickness decreases systematically eastward from its source areas (Fig. 2; Hutton, 1947). Parameters of particle size also change systematically with long distance of travel, although over short distances the changes are more complex (Ruhe, 1954a, p. 666). Radiocarbon dates from the basal part of the loess, which contains an A-C soil profile, indicate correlation with the Farmdale in many areas: coniferous wood from the A-horizon provides dates of 24,500 ± 880 (W-141) and 21,360 ± 850 (I-1023) in southwestern Iowa, 25,000 ± 2,500 (I-1267) and 29,000 ± 3,500 (I-1269) in northeastern Iowa, and 20,290 ± 1000 (I-1022) in southeastern Iowa.

Loess deposition was apparently interrupted in northwestern Iowa by advance of the ice. The till in this area that Carman (1931) had mapped as Iowan (Kay and Graham, 1943) contains wood at the base with a date of 20,000 ± 880 years (O-1325). This date confirms Ruhe's (1950, 1952) reclassification of much of this till to Tazewell, for the Tazewell (Shelbyville) till of Illinois is 19,200 ± 700 years old (W-187), and pre-Shelbyville loess in Illinois has dates of 20,340 ± 750 (W-349) and 20,300 ± 400 years (W-870). The reclassification in Iowa was based on the relatively abrupt decrease in thickness of the loess mantle at the outer limit of Tazewell drift, presumably because the late-Iowan, pre-Tazewell component of the loess was overridden. From this boundary eastward the loess thins systematically from 200 cm to 100-125 cm at the margin of the younger Wisconsin drift, which buries the entire loess and its weak paleosol and which has no continuous loess cover of its own. The Tazewell drift as thus identified has few undrained depressions; its topographic and stratigraphic contrast with the younger Wisconsin drifts (see below) is pronounced (Ruhe, 1952).

Wood dates from within the loess in central Iowa are 17,030 ±500 (I-1020), 16,720 ± 500 (W-126), 16,100 ± 1,000 (I-1270), and 16,100 ± 500 (I-1024). Dates from wood from the top of the loess, where it is buried by younger Wisconsin till in central Iowa, are about 14,000 years (see below). As no paleosols occur within the loess, except for the basal A-C paleosol, loess deposition must have been continuous from the time of the Farmdale, 29,000 years ago, through the Tazewell until the invasion of the ice into central Iowa 14,000 years ago.

Wisconsin

Introduction

Wisconsin glacial drift covers most of Minnesota and northern Iowa, and Wisconsin loess blankets much of the rest of the area. Whereas the pre-Wisconsin drifts in Iowa all have essentially the same lithology, implying a common source, *the Wisconsin drifts in Minnesota have a highly varied* lithology and complex stratigraphy, which reflect the configuration of the several lobes that protruded from the ice-sheet margin during various intervals of advance and retreat. This lobation of the ice margin was controlled by the major bedrock topography of the area over which the ice moved. An understanding of the relations therefore requires knowledge of both the lithology and topography of the bedrock.

Precambrian basic igneous rocks form the highlands north of Lake Superior (Fig. 3). Red sandstone occurs in a belt extending south from the head of Lake Superior. Paleozoic dolomite and sandstone are found in southeastern Minnesota and adjacent Iowa and also in the northwestern corner of Minnesota and adjacent Manitoba. Cretaceous shale occurs primarily in western Minnesota. The north-central part of Minnesota and adjacent Ontario are underlain by undiagnostic granitic and metamorphic rocks.

The bedrock topography (Fig. 4) is marked by (1) a deep lowland now occupied by Lake Superior, (2) a shallower lowland, here called the Minneapolis lowland, extending northeast through Minneapolis *en echelon* with the Superior lowland, (3) a long lowland in three segments now occupied by the Red, Minnesota, and Des Moines Rivers, and (4) a shallow lowland centering in the Red Lakes area in northwestern Minnesota. This topography probably dates in part from Cretaceous time, for Upper Cretaceous marine sediments are found in both the Red Lakes lowland and the Red-Minnesota lowland (Sloan, 1964).

Four major ice lobes are recognized for the glaciation of Minnesota and Iowa. On the east the Superior Lobe followed the Superior and Minneapolis lowlands, bringing generally red sandy till with pebbles of red sandstone and slate. Immediately to the northwest in northern Minnesota was the Rainy Lobe, which produced drift of variable color and composition according to the dominant rock types that it traversed. On the gabbro highland northwest of the Lake Superior basin the Rainy Lobe drift is gray or blue gray. Farther west the drift is light gray to light brown because of the abundance of granitic rocks from the terrane north of the Mesabi iron range. In central Minnesota the drift is brown because it is dominated by fragments of initially gray metamorphic rocks that apparently have oxidized readily in the fine fraction to a uniform brown color.

Leverett (1932) considered the Rainy and Superior Lobes as one, referring them to the Patrician ice center of Ontario 200 miles north of Minnesota. He did recognize, however, that during the late Wisconsin the Superior Lobe from the "Labradorian ice center" to the northeast moved quite separately into the area.

In western Minnesota the Des Moines Lobe followed the major lowland to central Iowa and sent sublobes eastward across northern Minnesota (St. Louis Sublobe) and across southern Minnesota (Grantsburg Sublobe), bringing gray (buff where oxidized) calcareous silty till with fragments of Cretaceous shale. Not so far west was the Wadena Lobe, which brought gray to buff sandy calcareous till to the Red Lakes lowland from the carbonate terrane of Manitoba that lies north of the limits of Cretaceous shale. The Des Moines and Wadena Lobes were not differentiated by Leverett (1932), who referred all the calcareous drift to the Keewatin ice center northwest of Hudson Bay.

TERMINOLOGY

The following description of the glacial geology of the Wisconsin Glacial Stage in Minnesota and Iowa is organized as a geologic history. Because of the uncertainties of correlation of the glacial events in this area with those in other parts of the Great Lakes region, the history is recounted as phases of ice advance of individual lobes; correlations between the phases of different ice lobes, as indicated by stratigraphic or drainage relations, are made where possible (Table 1). Phases are informal units of geologic time that are identified or defined on the basis of either morphologic or stratigraphic features and are given local geographic names. They thus are not necessarily bound by the stratigraphic strictures of the rock-stratigraphic, time-stratigraphic, or geologic-climate units of the Stratigraphic Code (American Commission of Stratigraphic Nomenclature, 1961) or even by those of the morphostratigraphic units of Frye and Willman (1960; see Wright, 1964).

HEWITT PHASE OF WADENA LOBE

The Hewitt phase (Fig. 5) is here named from till exposed in a Wadena Lobe drumlin on U.S. Highway 210 one mile east of Hewitt, Todd County (Wright, 1962, Table 2, Pl. 1). The Wadena Drumlin Field as exposed is about 110 km long and 65 km broad; it is overlapped on all sides by younger drift. It consists of about 1,200 drumlins formed by ice advancing toward the southwest and fanning to the west and south. The till of the drumlins is buff, sandy, calcareous, and without Cretaceous shale, implying a source in southeastern Manitoba north of the area of Cretaceous bedrock. The Wadena Lobe apparently entered the shallow Red Lakes lowland of northwestern Minnesota from the northwest but was blocked in its eastward progress by the contemporaneous Rainy Lobe advancing from the northeast. The Wadena Lobe was diverted thereby to the southwest to form the drumlin field. In the process of diversion it became contaminated by Rainy Lobe drift so that portions of the Wadena Lobe drift contain many indicator stones from the northeast (Wright, 1962).

The eastern extent of the Wadena Lobe in the Hewitt phase is not known because its drift is buried by younger drift of eastern source. On the south and west the Wadena Lobe is believed to have terminated in the Alexandria moraine complex, which forms a long arc of lake-dotted moraine bounding the drumlin field. This moraine, however, as well as the fringe of the drumlin field itself, was buried by later advance of the Des Moines Lobe from the west and southwest. On the north the drumlin field is buried by outwash leading from the Itasca Moraine, which was formed during the following phase of the Wadena Lobe.

The other ice lobes in Minnesota were probably active at the same time as the Hewitt phase of the Wadena Lobe but had not attained their maximum positions. Synchronous presence of the Rainy Lobe has already been postulated to explain the diversion of the Wadena Lobe from a south-

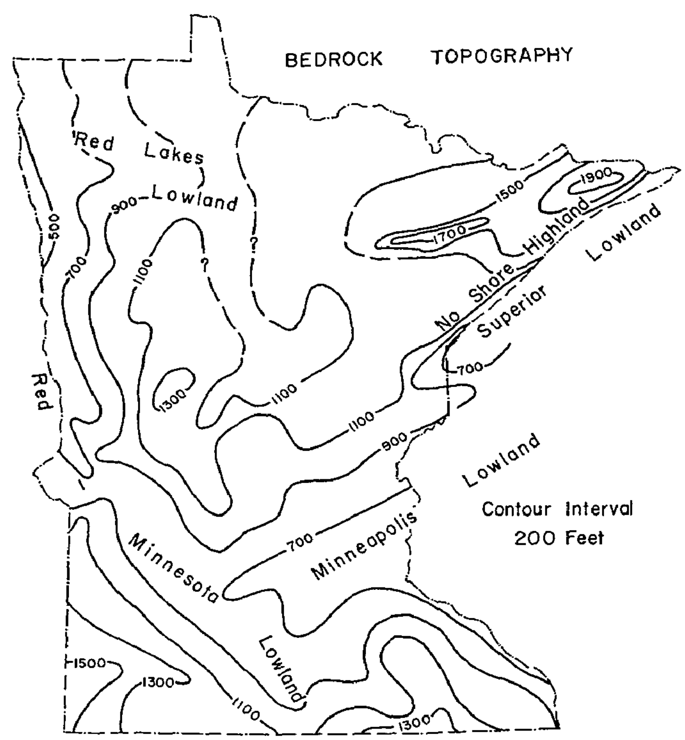

Figure 4. Generalized topographic map of the bedrock surface in Minnesota (compiled from well records by W. Vollendorf, 1950).

TABLE 1

Phases of Wisconsin Glaciation in Minnesota
(Figures indicate radiocarbon dates before the present. Read upward from bottom.)

DES MOINES LOBE					
MAIN (Red–Minnesota–Des Moines valleys)	GRANTSBURG SUBLOBE (Minneapolis lowland)	ST. LOUIS SUBLOBE (Red Lakes lowland)	WADENA LOBE (Red Lakes lowland)	RAINY LOBE (Upland)	SUPERIOR LOBE (Superior lowland; Minneapolis lowland in St. Croix phase)
9,200-7,000 L. Agassiz II 11,000-10,000 L. Agassiz I 11,740		L. Aitkin IIb Aitkin soil 11,635 L. Aitkin IIa and L. Upham II ALBORN PHASE		?	L. Duluth L. Nemadji NICKERSON PHASE Nickerson Mor.
MANKATO PHASE Algona Mor. 13,000	12,700-11,800 Anoka Sandplain PINE CITY PHASE L. Grantsburg Pine City Mor.	Advancing			11,500 SPLIT ROCK PHASE Hinckley Sandpl. Split Rock Dr.
BEMIS PHASE Bemis Mor. 14,000	?	Advancing?		VERMILION PHASE Vermilion Mor.	L. Aitkin I & L. Upham I AUTOMBA PHASE Highland-Mille Lacs Mor., Automba Dr.
?			ITASCA PHASE Itasca Mor.	ST. CROIX PHASE St. Croix Moraine 13,270 Brainerd Dr. Pierz Dr. *Toimi Dr.*	
?			>40,000 HEWITT PHASE Wadena Drumlins	Advancing	?

easterly to a southwesterly heading. The Des Moines Lobe must have occupied the Red River lowland at this time because this area is topographically lower than the shallow Red Lakes lowland of northwestern Minnesota into which the Wadena Lobe advanced; the Wadena Lobe must therefore have been an offshoot of the Des Moines Lobe but from a source farther north than its later counterpart (St. Louis Sublobe).

ITASCA PHASE OF WADENA LOBE, ST. CROIX PHASE OF SUPERIOR AND RAINY LOBES

The Wadena Lobe retreated to the Itasca Moraine, and the Superior and Rainy Lobes reached a common terminus at the St. Croix Moraine, which may be traced in a great loop from western Wisconsin on the St. Croix River southwestward to Minneapolis and thence to the northwest and north to a place where it becomes interlobate with the Itasca Moraine of the Wadena Lobe (Fig. 6). These two moraines give their names to the correlative phases of the respective ice lobes. The eastern and southern parts of the St. Croix Moraine are made of red sandy drift, diagnostic of the Superior Lobe. The northwestern part, however, is made of brown sandy till, characteristic of both the Pierz and Brainerd Sublobes of the Rainy Lobe. Here the drift rests on the Hewitt Till of the Wadena Lobe, with interlamination at the contact. These stratigraphic relations were interpreted by Schneider (1961) and Wright (1956) as indicating at least partial synchroneity of the Wadena and Pierz Lobes. As a result of the studies of two other Minnesota drifts by Cushing (1963), however, a more probable explanation is that erosion of Hewitt Till, perhaps still with dead ice, by active Pierz Lobe ice of the St. Croix phase, was followed by its deformation into a foliate structure.

Not only is the St. Croix Moraine west of the Mississippi River continuous as a topographic feature despite changing composition, but the fan-shaped Pierz Drumlin Field east of the moraine is also a continuous feature with generally gradational composition from red on the south to brown on the north. Although such gradations might imply that a single ice lobe was involved, perhaps in the sense of the Patrician lobe of Leverett (1932), it is possible to consider that two separate ice streams, one centered in the Superior basin and the other on the upland to the northwest, merged together in central Minnesota and fanned to the southwest and northwest to form a single Pierz Drumlin Field and the single St. Croix Moraine, in the same sense that two tributary valley glaciers from different sources may join in the main valley and then spread on the piedmont as a single ice mass. Lateral shifting of the two streams is implied by the occurrence of both red and brown till in superposition as major stratigraphic units in both the drumlin field and the moraine.

The relation of the Superior to the Rainy Lobe in the St. Croix phase is not completely clear, however, partly because their drifts were buried by younger ice advances in an area 130 km broad northeast of the Pierz Drumlin Field. North of the area of burial are Rainy Lobe drumlins with distinct southwest trend—a group called the Toimi Drumlin Field in St. Louis and Lake Counties (Wright, 1956), and a smaller group west of Hibbing on the western Mesabi iron range (Cotter *et al.*, 1964). As the history of glacier fluctuation is reconstructed for the Hewitt and St. Croix phases of the sequence, the broad Rainy Lobe moved slowly southwestward across the upland. On its right margin it blocked and diverted the Wadena Lobe, which was moving more rapidly into the Red Lakes lowland of northwestern Minnesota, as already recounted. The Wadena Lobe thereupon reached its maximum in the Hewitt phase and retreated. Meanwhile the Rainy Lobe pushed onward over part of the area just vacated. On its left flank it was confluent with the Superior Lobe, which was a thicker ice stream moving more rapidly in the deep Superior basin. This ice, advancing southwestward out of the basin, carried with it on its right flank, so to speak, the left portion of the Rainy Lobe, here called its Pierz Sublobe, and together they formed the Pierz Drumlin Field of red and brown drift.

The rest or western part of the Rainy Lobe, here called the Brainerd Sublobe, moved forward separately. Its drift is indistinguishable in lithology from that of the Pierz Sublobe, so the sequence must be reconstructed from the geomorphic relations. According to Schneider (1961), the Brainerd Drumlin Field, with southwest trend, crosscuts the northern edge of the Pierz field south of Brainerd. The Brainerd ice reached its terminus at the northernmost segment of the St. Croix Moraine, where it met at an angle the Wadena Lobe at the Itasca Moraine. Outwash from these two moraines formed the Park Rapids Outwash Plain and other extensive outwash fans and plains that partially bury the northeast portion of the Wadena Drumlin Field (Wright, 1962). Some of the outwash from the Brainerd Sublobe drained southward along the inner (eastern) margin of that segment of the St. Croix Moraine from which the Pierz Sublobe had already withdrawn (Schneider, 1961).

Wastage of the Superior and Rainy Lobes of the St. Croix phase from the St. Croix Moraine left a remarkable record of tunnel valleys and eskers, currently under study by E. J. Cushing. The tunnel valleys form a system trending southwest, parallel to the axis of the Superior Lobe, with only a slight tendency to fan to the west in the St. Cloud area and to the south near Minneapolis. The downstream ends of many of the tunnel valleys have been buried by younger drifts so are partially obscured. The tunnel valleys were formed in the ablation zone of the glacier when the ice was still thick enough to permit the development of closed, water-filled drainage tunnels at the base under enough hydrostatic pressure to provide the water velocity necessary to erode sizable gorges in the substratum. The sub-parallel pattern of tunnel valleys cuts obliquely across the Pierz Drumlin Field. It must reflect the surface contour in the ablation zone of the ice lobe and thus the hydrostatic gradient; it does not reflect the subglacier topography, which in fact in this area has a general southeastward slope into the Minneapolis lowland, at right angles to the southwesterly trend of the tunnel valleys.

Many of the tunnel valleys penetrate the St. Croix Moraine and must have been formed when the ice stood at its maximum. In fact the dissection of the moraine by tunnel valleys in the area northwest of Minneapolis may have been great enough to create gaps that later eased the eastward penetration of the Grantsburg Sublobe. As the ice thinned and stagnated during wastage, the hydrostatic pressure was reduced or lost by air intake through crevasses or other openings, and the streams changed from erosional to depositional, forming eskers in or beside the previously eroded channels. The system worked headward during ice wastage and extended for 130 km from the St. Croix Moraine to the low divide between the Minneapolis and Superior lowlands, beyond which it is buried or modified

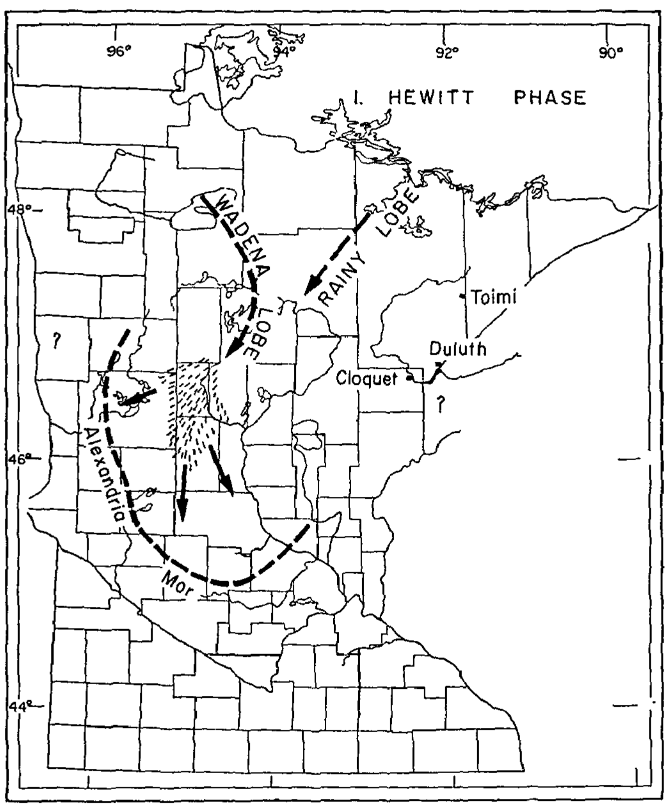

Figure 5. Hewitt phase of Wisconsin glaciation in Minnesota.

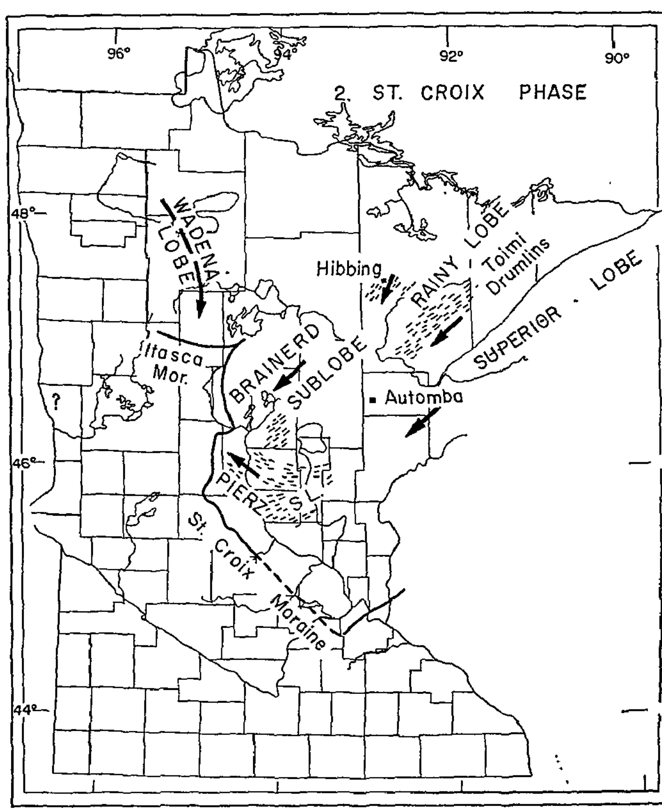

Figure 6. St. Croix phase of Wisconsin glaciation in Minnesota.

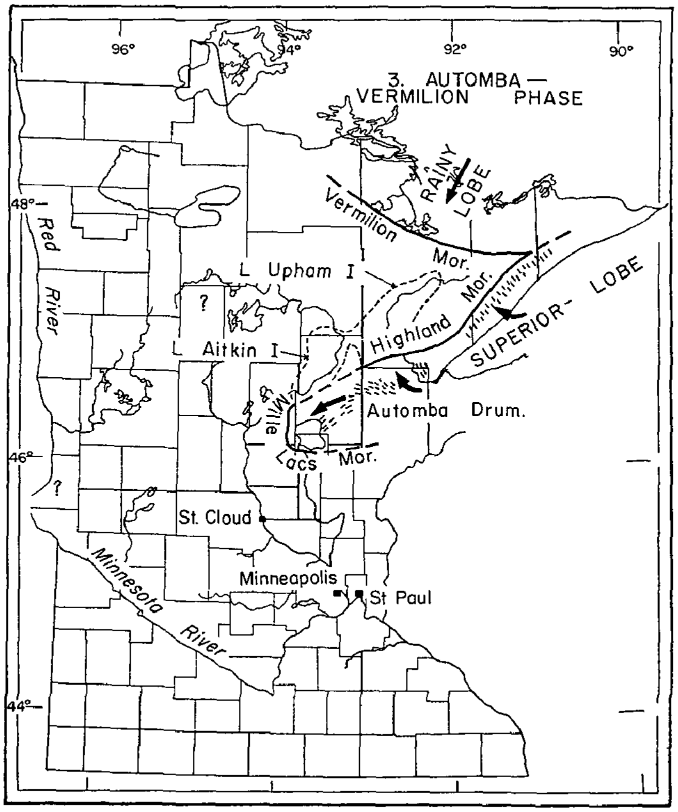

Figure 7. Automba-Vermilion phase of Wisconsin glaciation in Minnesota.

by younger drift. Similar tunnel valleys and eskers are found farther north in the Toimi Drumlins of the Rainy Lobe north of the area of burial (Baker, 1964).

AUTOMBA PHASE OF SUPERIOR LOBE,
VERMILION PHASE OF THE RAINY LOBE

The thinning and wastage of the ice first cleared the upland occupied by the Rainy Lobe, which by this time had retreated to or beyond the Vermilion Moraine close to the Canadian border (Fig. 7). The Deep Superior basin was still filled with ice, however, and the long, shallow Minneapolis lowland leading southwest to the St. Croix Moraine still had extensive areas of thin and stagnant ice. The active part of the Superior Lobe then readvanced out of its deep basin. On its right flank the ice rose up the steep slope of the North Shore Highland to build a strip of narrow drumlins terminating in the Highland Moraine about 15-25 km from the North Shore. This drift, although assigned to the Superior Lobe, lacks the characteristic red color and red sandstone pebbles because these materials were apparently less available this far north. The Highland Moraine truncates the Toimi Drumlins of the St. Croix-phase Rainy Lobe, and its outwash plains partially bury the drumlins.

In the Duluth area at the head of the Lake Superior basin, where the North Shore Highland ends, the ice fanned out to the west and northwest to build the Automba Drumlin Field of red sandy till, covering much of the area previously occupied by the Rainy Lobe. The ice probably terminated in a moraine that continues the southwest trend of the Highland Moraine to the Mille Lacs Moraine on the west side of Lake Mille Lacs in central Minnesota. The path of this lobe seems anomalous in the respect that the ice apparently did not follow southwestward down the axis of the Minneapolis lowland, although it had enough thickness to do this, but rather climbed out of the lowland and flowed westward to the Mille Lacs area. One possible but not completely satisfactory explanation is that the Minneapolis lowland was still plugged with stagnant ice remaining from the St. Croix phase. The stagnant ice was so drift-laden as a result of long wastage that it was not appreciably reactivated by the ice advance of the Automba phase. This active ice lobe thus was blocked from a southwesterly course and spilled instead westward onto the upland and around the northern margin of the dead-ice area, as recorded by the curving pattern of the Automba Drumlin Field. The ice left no recognizable moraine on its left flank, and it may have been dispersed as a thin overburden on the dead ice against which it impinged.

The Superior Lobe in this re-expanded position must have blocked the drainage of the St. Louis River, which could then have formed a large proglacial lake (Lake Upham I) north of the ice front, extending north to the Mesabi iron range. It may have formed a similar Glacial Lake Aitkin I farther to the west in the present Mississippi drainage. The two lakes were probably confluent and drained into the Mississippi system. They are believed to have received red lake clay and silt from the Superior Lobe meltwater. Lake Aitkin I must in addition have received brown lake sediments from ice to the northwest or north. These lake sediments were to be overridden in the next phase by ice from the west to produce red and brown clay tills.

The positions of the Wadena and Des Moines Lobes during the Automba phase are unknown. The former must have wasted completely at some time after its earlier stand at the Itasca Moraine because the Red Lakes lowland that it had occupied became filled subsequently with the St. Louis Sublobe of the Des Moines Lobe. Presumably this clearance coincided with the deep retreat of the Rainy Lobe and the broad stagnation of the Superior Lobe at the end of the St. Croix phase. Perhaps the St. Louis Sublobe then started its advance and contributed brown clay to Lake Aitkin I.

SPLIT ROCK PHASE OF SUPERIOR LOBE, PINE CITY PHASE
OF GRANTSBURG SUBLOBE

By the end of the Automba phase the active Superior Lobe had retreated across the low bedrock divide between the Minneapolis and Superior lowlands, although much stagnant ice was left in its wake. Red clay must have been deposited in one or more proglacial lakes of unknown dimensions, because when the ice readvanced southwestward in a narrow tongue as far as the bedrock divide it discontinuously covered eskers, tunnel valleys, and other older features with a veneer of red clay till (Fig. 8). In some cases the cap of clay till on eskers is underlain or replaced by a cap of sandy till. The sandy till can be interpreted as a product of Automba-phase till or till-laden dead ice redeposited or reactivated by the overriding ice of the Split Rock phase (E. J. Cushing, in preparation). Alternatively, it may have been deposited earlier in a separate movement by active ice of the Automba phase itself on top of the wastage features of the St. Croix phase.

A small bulge of ice extended westward over the edge of the Automba Drumlins to an elevation of 1,250 ft (380 m) up the basin of the Split Rock River, constructing a small field of drumlins of different form, here called the Split Rock Drumlins, from which the phase takes its name.

Elsewhere the right (northwest) side of this thin lobe formed two well-marked successive frontal outwash plains at 1,300 and 1,280 ft with ice-contact slopes and esker sources. Some of the western drainage continued into Lake Upham I. The rest of the western drainage and all the southern drainage from this narrow lobe moved southeastward toward the St. Croix River, but it was intercepted along the north edge of the Grantsburg Sublobe of the Des Moines Lobe by Lake Grantsburg and here formed the Hinckley Sundplain.

The Grantsburg Sublobe at its maximum built the Pine City Moraine (Leverett, 1932, p. 79), thus defining the Pine City phase. The ice had protruded northeastward up the Minneapolis lowland, which had previously been filled by the Superior Lobe in the St. Croix phase. It broke across the western arm of the St. Croix Moraine and buried the tunnel valleys, eskers, and other wastage features of the Superior Lobe. In the process it picked up red drift or drift-laden dead ice to produce the intricate interlaminated red and buff structures that mark the contact between the two drifts (Cushing, 1963). As the ice wasted, the Anoka Sandplain developed in its place. It received, as had Lake Grantsburg, the drainage of the Mississippi River as well as of the Superior Lobe, but its development was principally influenced by the pattern of local ice wastage (Cooper, 1935; E. J. Cushing, in preparation).

ALBORN PHASE OF ST. LOUIS SUBLOBE, NICKERSON PHASE OF SUPERIOR LOBE

The St. Louis Sublobe branched from the Des Moines Lobe far enough to the south to pick up Cretaceous shale, in contrast to the earlier Wadena Lobe, which reached the Red Lakes lowland from the non-shaly, carbonate terrane of the Lake Winnipeg area in Manitoba (Fig. 9). The St. Louis Sublobe was not blocked by the Rainy Lobe in its eastward advance into the Red Lake lowland as the Wadena Lobe had been. In fact on its left-hand side in northernmost Minnesota it buried the Vermilion Moraine of the Rainy Lobe. It crossed the low western end of the Mesabi iron range and overrode the red and brown clays of Glacial Lakes Aitkin I and Upham I, redepositing them in a complex of red or brown clay till interlaminated or intermixed with the light-brown shale-bearing silty till that is more characteristic of the up-glacier, uncontaminated portion of the St. Louis Sublobe drift. The ice advanced to an elevation of 1,550 ft (470 m) on the south flank of the Mesabi-range area, and eastward it buried the edge of the Toimi Drumlin Field and the dissecting tunnel valleys of the Rainy Lobe. It covered portions of the Highland Moraine and other Automba-phase features of the Superior Lobe and locally built its own morainic topography. This phase takes its name from the Alborn Till near its terminus in southwestern St. Louis County (Baker, 1964).

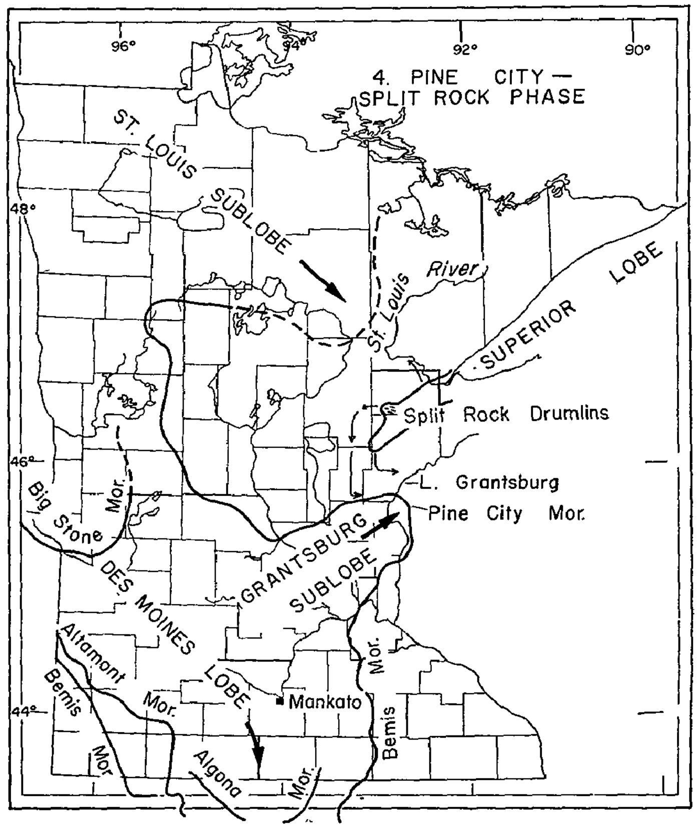

Figure 8. Pine City–Split Rock phase of Wisconsin glaciation in Minnesota.

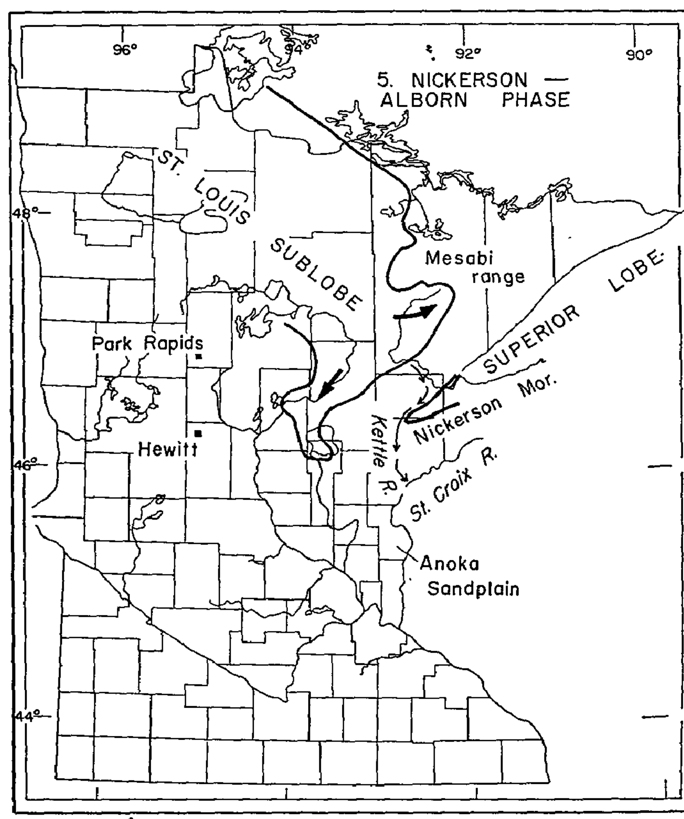

Figure 9. Nickerson-Alborn phase of Wisconsin glaciation in Minnesota.

Outwash from the St. Louis Sublobe extended in part eastward from the newly formed St. Louis River to Cloquet, whence it was diverted by the Superior Lobe southward across the Split Rock Drumlins and into the Kettle River, a tributary of the St. Croix. By this time the Grantsburg Sublobe had wasted almost completely and no longer diverted the Mississippi River.

The Superior Lobe had withdrawn from its Split Rock-phase position to the Nickerson Moraine, which is a striking feature of strong local relief in red clay till near the west end of the Superior lowland. During the waning portion of the phase the St. Louis Sublobe retreated. Glacial Lakes Upham II and Aitkin II formed at its front and supplied increasing amounts of water to the St. Louis River, which then flowed southward down the diversion channel around the Superior Lobe to the Kettle River. Progressive withdrawal of the blocking ice mass permitted the development of drainage channels at about 1,200, 1,170, 1,130, and 1,100 ft near the point of the Nickerson Moraine, until finally the Superior Lobe withdrew far enough into the basin to form Glacial Lake Nemadji, which had an outlet at 1,050 ft to the Kettle River. The clear outlet waters cut a gorge along the Kettle and St. Croix Rivers through the deposits of the Anoka Sandplain and Lake Grantsburg and into the bedrock. The dissection may have been speeded by the drop in base level induced by contemporaneous dissection of the Mississippi River, to which the St. Croix is tributary, by the outlet waters of Glacial Lake Agassiz (see below). This would imply complete wastage of the Des Moines Lobe to open at least the Minnesota portion of the Lake Agassiz basin while the Superior Lobe was still blocking the St. Louis River. This river in turn was still the outlet of Lake Upham II, which may still have been supplied by melt-water from the wasting St. Louis Sublobe. Glacial Lake Duluth, which drained directly to the St. Croix River in northwestern Wisconsin rather than first into the Kettle River, formed immediately after, and the St. Croix River gorge was further deepened.

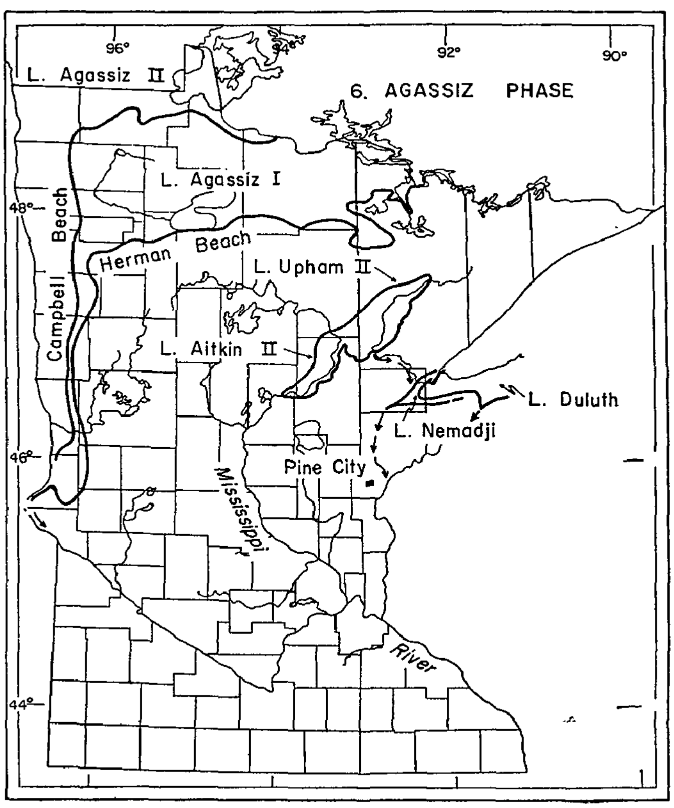

Figure 10. Agassiz phase of Wisconsin glaciation in Minnesota.

The relations of the St. Louis and Superior Sublobes as here recounted represent a revision in which several persons have had a part, notably Thomas (1959) and Baker (1964), as well as E. J. Cushing and John E. Foss in the course of their incidental field work in the area. The earlier interpretation, which had difficulties from its inception (Wright, 1955), is thereby discarded: it had assumed that the red clay till around much of the basins of Glacial Lakes Upham and Aitkin was deposited by the Superior Lobe climbing west and north out of the head of the Lake Superior basin. The revision was facilitated by the availability of new topographic maps, which now can be used to demonstrate the close control of elevation on the ice margins as well as on the drainage courses. Additional study revealed a persistent 25-km gap in the distribution of red clay till in the area west of the head of Lake Superior. This gap, previously discounted as an instance of non-deposition of the part of the Superior Lobe, is now considered to represent ground covered by neither the St. Louis Lobe nor Superior Lobe during their clay-till phases.

BEMIS AND MANKATO PHASES OF THE DES MOINES LOBE

The drift of the St. Louis Sublobe, as traced westward along its southern border, covers that of the Wadena Lobe north of the Itasca Moraine—the two gray calcareous drifts may be distinguished by the generally siltier texture and the shale content of the St. Louis Sublobe drift. The border of the younger drift may be traced thence southward to mark the east edge of the main Des Moines Lobe (Fig. 8). It buried the western and southern edges of the Wadena Drumlin Field. It forms the surface drift in the big Alexandria moraine complex, which, however, may have a buried core of Hewitt-phase Wadena Lobe moraine. The drift border thence may be traced eastward as the left-hand edge of the Grantsburg Sublobe.

The Des Moines Lobe proper at this time was extended southward to Mankato in southern Minnesota and onward to central Iowa. The attainment of its terminus at the Bemis Moraine is here called the Bemis phase. During retreat the ice formed the Altamont Moraine and then the broad Algona Moraine in northern Iowa (Kay and Graham, 1943, p. 239; Ruhe *et al.*, 1957; Ruhe and Scholtes, 1959). The Algona Moraine may be considered to mark the Mankato phase, for the ice soon retreated past the townsite of Mankato in southern Minnesota and thence northwest to the divide between the Minnesota and Red River valleys. Here at the Big Stone Moraine it dispatched great floods of outwash which coursed irregularly across the till plain to the south and eventually became channeled into the Minnesota River valley. Further retreat opened the basin of Glacial Lake Agassiz north of the Big Stone Moraine (Fig. 10), and as the lake expanded into Canada the spillway waters (Glacial River Warren) cut the outlet gorge as much as 100 m deep through late-Mankato outwash for hundreds of miles to the Mississippi River at Minneapolis and beyond to Illinois (Zumberge, 1952). As the outlet was eroded, successive beaches were formed at elevations ranging in the outlet region from 1,060 (Herman) to 980 ft (Campbell). The beaches may be traced northward into Canada, where the elevations are higher because of postglacial tilting (Upham, 1896; Leverett, 1932; Elson, 1957).

CHRONOLOGY AND CORRELATION OF WISCONSIN DRIFTS

The Hewitt Till of the Wadena Lobe is marked by undrained depressions, and the soils are no better developed than the soils on other Wisconsin drifts in Minnesota (Fig. 5). The drumlins were considered to be Iowan in correlation by Leverett (1932) but early Cary by Wright (1962). A radiocarbon date on wood from silts above this drift and beneath the Des Moines Lobe drift of the Bemis or Mankato phase is >40,000 years (W-1232). On the basis of the topographic expression and shallow weathering profile, however, the Hewitt phase should be younger than is implied by the date, so confirmation is desirable.

The St. Croix Moraine (Fig. 6) was correlated by Leverett (1932) with the Middle Wisconsin (Cary) of Illinois, and this correlation is retained by Black (Frye *et al.*, this volume). The only radiocarbon date available on the St. Croix Moraine is from the basal organic sediment at Kirchner Marsh south of Minneapolis: 13,270 ± 200 (Y-1326; Wright *et al.*, 1963). Recession of active ice from the moraine may have occurred at an earlier date, so the correlation of the St. Croix Moraine either with the Cary or with earlier moraines of Illinois and Indiana, which are at least 14,000 years old, is possible. No C^{14} dates are available for the Automba-Vermilion phases (Fig. 7).

The Split Rock phase of the Superior Lobe (Fig. 8) has a minimum date on a kettle-bottom sample in its outwash west of Cloquet of 11,500 (W-1059, Farnham *et al.*, 1964), implying that this phase is older than the Two Creeks interval. The Pine City phase of the Grantsburg Sublobe, whose maximum stand and early wastage is contemporaneous with the Split Rock and Nickerson phases of the Superior Lobe, fits this age assignment, for kettles in its outwash have bottom dates of 12,700-11,800 years old (Wright and Rubin, 1956).

The Alborn phase of the St. Louis Sublobe (Fig. 9) is older than 11,635 years (W-502 and W-1141, Farnham *et al.*, 1964), according to dates on a soil formed during a dry interval of Glacial Lake Aitkin II. An anomalous pair of dates on Alborn-phase till on the Mesabi iron range, however, implies a post-Two Creeks rather than pre-Two Creeks age for the Alborn phase: 11,330 (W-827) and 11,100 (W-1140) (Farnham *et al.*, 1964).

The Bemis and Mankato phases of the main Des Moines Lobe (Fig. 8) are also older than the Two Creeks interval. The ice reached its terminus at the Bemis Moraine in central Iowa at the time of the Cary phase of the Lake Michigan lobe, as is suggested by the well-controlled 14,000-year dates on wood (including rooted trees) buried by drift of the Bemis phase: 13,820 ± 400 (W-513), 13,910 ± 400 (W-517), 14,470 ± 400 (W-512), and 14,700 ± 400 (W-153) (Ruhe *et al.*, 1957). The next moraine to the north, the Altamont, was previously included in the Mankato phase (Ruhe, 1952b). However, better chronological control is available at the broad Algona Moraine still farther north, whose outwash buried trees and peat dated as 12,970 (W-626) and 13,030 (W-625) (Ruhe and Scholtes, 1959), and this moraine is now taken to mark the climax of the Mankato phase, equivalent in time to the Port Huron phase of the Lake Michigan Lobe (Wright, 1964). The retreating Mankato ice uncovered the area near Mankato itself before 12,650 years ago (W-824, Jelgersma, 1962).

The drift peripheral to the Bemis Moraine in eastern Iowa is the problematical Iowan till, deposited by an earlier Des Moines Lobe. In western Iowa, however, a portion of it is correlated with the Tazewell phase, a proper subdivision of the Wisconsin glaciation of the Lake Michigan Lobe in Illinois (see the above discussion of the Iowan problem).

The exact chronological relations between the main Des Moines Lobe and its two sublobes are not yet clear, for carbon dates on the sublobes all measure the times of wastage and furthermore are minimal dates. The two sublobes did not reach their maxima at exactly the same time, as shown by drainage relations with the Split Rock and Nickerson phases of the Superior Lobe. The maximum stand of the main lobe at the Bemis Moraine, 14,000 years ago, may have occurred either at about the same time as the sublobe maxima or before. The evidence indicates, however, that all these advances, as well as the associated red-clay-till advances of the Superior Lobe, predate the Two Creeks interval. The Valders phase of the Lake Michigan Lobe accordingly is not recorded in Minnesota.

Lake Agassiz was first formed in the Red River lowland about 12,000 years ago when the Des Moines Lobe retreated at the end of the Mankato phase, and it rapidly expanded into the Red Lakes lowland, into which the highest major beach (Herman) may be traced (Fig. 10). This level was abandoned by 11,740 years ago (Y-1327, Shay, 1965) as the southern outlet was eroded. Continued retreat of the ice permitted drainage of the lake to the east through some course in Ontario, as yet unidentified, and the lake dropped to low levels about 11,000-10,000 years ago (Johnston, 1946; Elson, 1957; Shay, 1965).

Readvance of the ice from the northeast (and/or crustal tilting) closed the eastern outlet and raised the lake again to the lowest level of southern outlet (Campbell Beach), forming Lake Agassiz II. Zoltai (1961) indicates that the ice (the Rainy Lobe of Minnesota) may at this time have been as far south as the Rainy Lake Moraine at the Minnesota border. North of the Rainy Lake Moraine are three younger moraines recording retreatal positions of the Rainy Lobe. The middle one of the three (Hartman Moraine) may extend eastward to the Dog Lake Moraine, which is interlobate with the red-clay-till Marks Moriane of a sublobe from the Superior basin (Zoltai, 1963). A proglacial lake in front of the Marks Moraine carried red clay westward to the Lake Agassiz basin, where it was deposited in the uppermost sediments of Lake Agassiz in the area between the Hartman Moraine on the north and the Minnesota border on the south. If these are the sediments of Lake Agassiz II, as is implied by Elson (1957, Fig. 2), then a means is available for correlating the Agassiz events with those of the Lake Superior basin in the following way:

Lake Agassiz II withdrew from the Campbell Beach 9,200 years ago (W-1057), as low outlets to the east were uncovered, presumably by retreat of the ice from the Hartman–Dog Lake–Marks Moraines. The red clay and the Marks Moraine are therefore not much older than 9,200 years. Zoltai (1963) indicates that Lake Duluth then was formed, but this Ontario portion of Lake Duluth (if it be a portion) must therefore be much younger than Lake Nemadji and its immediate successor Lake Duluth in the type area at the western end of Lake Superior, where it was first formed at the close of the Nickerson phase, perhaps 11,500 years ago. The Ontario lake may rather be compared with a similar ice-marginal lake on the south side of the Superior Lobe in northern Michigan—Lake Ontonagon, whose red lake clay bears wood dated at 10,220 years ago (M-359).

If Elson's (1957) projection of the Campbell beach to the Hartman Moraine is incorrect, then the red-clay band may refer to Lake Agassiz I (pre-Campbell) instead of II, and the Hartman Moraine could correlate with the Valders phase or even with the Nickerson phase of the Minnesota sequence. Such a correlation is supported by a recent C^{14} date of 9,380 years ago (GSC-287) from a Lake Superior beach much lower than Lake Duluth and associated with an ice front 120 miles northeast of the Hartman Moraine and separated from it by two more moraines, providing one considers that at least 2,000 years is necessary for 120 miles of ice retreat and moraine formation (S. C. Zoltai, written communication).

References

Alden, W. C., and Leighton, M. M., 1917, The Iowan drift, a review of the evidence of the Iowan stage of glaciation: Iowa Geol. Surv., v. 26, p. 49-212

American Commission on Stratigraphic Nomenclature, 1961, *Code of stratigraphic nomenclature: Amer. Assoc. Petroleum* Geologists Bull., v. 45, p. 645-665

Bain, H. F., 1897, Relations of the Wisconsin and Kansan drift sheets in central Iowa and related phenomena: Iowa Geol. Surv., v. 7, p. 433-487

Baker, R. G., 1964, Late-Wisconsin glacial geology and vegetation history of the Alborn area, St. Louis County, Minnesota: Univ. Minnesota M.S. thesis, 44 p.

Calvin, Samuel, 1911, The Iowan drift: J. Geol., v. 19, p. 577-602

Carman, J. E., 1931, Further studies on the Pleistocene geology of northwestern Iowa: Iowa Geol. Surv., v. 35, p. 15-194

Cooper, W. S., 1935, The history of the upper Mississippi River in late Wisconsin and postglacial time: Minnesota Geol. Surv. Bull. 26, 116 p.

Cotter, R. D., Young, H. L., and Winter, T. C., 1964, Preliminary surficial geologic map of the Mesabi-Vermilion iron range area, northeastern Minnesota: U.S. Geol. Surv. Misc. Geol. Inv. Map 1-403

Cushing, E. J., 1963, Origin of pseudostratification and interlayering in glacial tills (abst.): Geol. Soc. Amer. Spec. Pap. 76, p. 58

Elson, J. A., 1957, Lake Agassiz and the Mankato-Valders problem: Science, v. 126, p. 999-1002

Farnham, R. S., McAndrews, J. H., and Wright, H. E., Jr., 1964, A late-Wisconsin buried soil near Aitkin, Minnesota, and its paleobotanical setting: Amer. J. Sci., v. 262, p. 393-412

Frye, J. C., and Willman, H. B., 1960, Classification of the Wisconsinan Stage in the Lake Michigan glacial lobe: Illinois State Geol. Surv. Circ. 285, 16 p.

Frye, J. C., William, H. B., and Black, R. F., this volume, Outline of glacial geology of Illinois and Wisconsin

Hutton, C. E., 1947, Studies of loess-derived soils in southwestern Iowa: Soil. Sci. Soc. Amer. Proc., v. 12, p. 424-431

Jelgersma, Saskia, 1962, A late-glacial pollen diagram from Madelia, southcentral Minnesota: Amer. J. Sci., v. 260, p. 522-529

Johnston, W. A., 1946, Glacial Lake Agassiz, with special reference to the mode of deformation of the beaches: Geol. Surv. Canada Bull. 7, 29 p.

Kay, G. F., 1928, Loveland loess, post-Illinoian, pre-Iowan in age: Science, v. 68, p. 482-483

Kay, G. F., and Apfel, E. T., 1929, The pre-Illinoian Pleistocene geology of Iowa: Iowa Geol. Surv., v. 34, 304 p.

Kay, G. F., and Graham, J. B., 1943, The Illinoian and post-Illinoian Pleistocene geology of Iowa: Iowa Geol. Surv., v. 38, p. 1-262

Leighton, M. M., 1933, The naming of the subdivisions of the Wisconsin glacial age: Science, v. 77, p. 168

Leighton, M. M., and Willman, H. B., 1950, Loess formations of the Mississippi Valley: J. Geol., v. 58, p. 599-623

Leverett, Frank, 1926, The Pleistocene glacial stages: were there more than four? Amer. Philos. Soc. Proc., v. 45, p. 105-118

—— 1929, Moraines and shorelines of the Lake Superior region: U.S. Geol. Surv. Prof. Pap. 154-A, 72 p.

—— 1932, *Quarternary geology of Minnesota and parts of* adjacent states: U.S. Geol. Surv. Prof. Pap. 161, 149 p.

—— 1939, The place of the Iowan drift: J. Geol., v. 47, p. 398-407

McGee, W J, 1891, The Pleistocene history of northeastern Iowa: U.S. Geol. Surv. Ann. Rep. 11, p. 189-577

Rubin, M., and Alexander, C., 1960, U.S. Geological Survey radiocarbon dates v: Amer. J. Sci. Radiocarbon Suppl., v. 2, p. 129-185

Ruhe, R. V., 1950, Graphic analysis of drift topographies: Amer. J. Sci., v. 248, p. 435-443

—— 1952, Topographic discontinuities of the Des Moines lobe: Amer. J. Sci., v. 250, p. 46-56

—— 1954a, Relations of the properties of Wisconsin loess to topography in western Iowa: Amer. J. Sci., v. 252, p. 663-672

—— 1954b, Pleistocene soils along the Rock Island relocation in southwestern Iowa: Amer. Railway Engr. Assoc. Bull. 514, p. 639-645

—— 1956, Geomorphic surfaces and the nature of soils: Soil Sci., v. 82, 441-455

—— 1960, Elements of the soil landscape: 7th Intern. Congr. Soil Sci. Trans., v. 4, p. 165-170

Ruhe, R. V., this volume, Quaternary paleopedology
Ruhe, R. V., and Gould, L. M., 1954, Glacial geology of the Dakota County area, Minnesota: Geol. Soc. Amer. Bull., v. 65, p. 769-792
Ruhe, R. V., Rubin, Meyer, and Scholtes, W. H., 1957, Late Pleistocene radiocarbon chronology in Iowa: Amer. J. Sci., v. 255, p. 671-689
Ruhe, R. V., and Scholtes, W. H., 1959, Important elements in the classification of the Wisconsin glacial stage: a discussion: J. Geol., v. 67, p. 585-593
Schneider, A. F., 1961, Pleistocene geology of the Randall region, central Minnesota: Minnesota Geol. Surv. Bull. 40, 151 p.
Scholtes, W. H., 1955, Properties and classification of the paha loess-derived soils in northeastern Iowa: Iowa State Univ. J. Sci., v. 30, p. 163-209
Schoewe, W. H., 1920, the origin and history of extinct Lake Calvin: Iowa Geol. Surv., v. 29, p. 49-222
Shay, C. T., 1965, Postglacial vegetation development in northwestern Minnesota, and its implications for prehistoric man: Univ. Minnesota M.S. thesis
Shimek, Bohumil, 1909, Aftonian sand and gravels in western Iowa: Geol. Soc. Amer. Bull., v. 20, p. 399-408
Sloan, R. S., 1964, The Cretaceous System in Minnesota: Minnesota Geol. Surv. Rep. Inv. 5, 64 p.
Thomas, J. A., 1959, Geology of the Cloquet area, northeastern Minnesota: Univ. Minnesota M.S. thesis, 67 p.
Upham, Warren, 1896, The glacial Lake Agassiz: U.S. Geol. Survey Monogr. 25, 685 p.
Winchell, N. H., 1884, 1888, Geology of Minnesota: Minnesota Geol. Nat. Hist. Surv., Final Rep., v. 1, 697 p., v. 2, 695 p.
Wright, H. E., Jr., 1955, Valders drift in Minnesota: J. Geol., v. 63, p. 403-411
—— 1956, Sequence of glaciation in eastern Minnesota. Geol. Soc. Amer. Guidebook, Minneapolis Meeting, pt. 3, p. 1-24
—— 1962, Role of the Wadena lobe in the Wisconsin glaciation of Minnesota: Geol. Soc. Amer. Bull., v. 73, p. 73-100
—— 1964, Classification of the Wisconsin glacial stage: J. Geol., v. 72, p. 628-637
Wright, H. E., Jr., and Rubin, Meyer, 1956, Radiocarbon dates of Mankato drift in Minnesota: Science, v. 124, p. 625-626; Discussion: 1957, Science, v. 125, p. 1037-1039
Wright, H. E., Jr., Winter, T. C., and Patten, H. L., 1963, Two pollen diagrams from southeastern Minnesota: problems in the regional late-glacial and postglacial vegetational history: Geol. Soc. Amer. Bull., v. 74, p. 1371-1396
Zoltai, S. C., 1961, Glacial history of part of northwestern Ontario: Geol. Assoc. Canada Proc., v. 13, p. 61-83
—— 1963, Glacial features of the Canadian Lakehead area: Can. Geographer, v. 7, p. 101-115
Zumberge, J. H., 1952, Lakes of Minnesota—their origin and classification: Minnesota Geol. Surv. Bull. 35, 99 p.

Summary

Drifts of all four major glacial and interglacial stages of the Pleistocene occur in the area. Nebraskan drift and Aftonian paleosols and sediments are found buried by Kansan drift in southern Iowa. Kansan drift and Yarmouth paleosol and sediments are common under younger loess in south-central Iowa. Illinoian till is found only close to the Mississippi River in southeastern Iowa, but its correlative Loveland loess is widespread near the Missouri River in western Iowa, with well-developed Sangamon paleosol on top. Much of the so-called Iowan till may actually be older till from which the top has been eroded, as implied by C^{14} dates, but the Iowan problem is not completely solved. Loess was extensively deposited in Iowa during the period 29,000 to 20,000 years ago, when it was interrupted by the ice advance of the Tazewell phase of Wisconsin glaciation. Loess deposition continued in Iowa until about 14,000 years ago.

Wisconsin glaciation in Minnesota involved the interactions of four major ice lobes, from west to east the Des Moines, Wadena, Rainy, and Superior, localized by bedrock lowlands and characterized by distinctive rock types that reflect the bedrock geology of northern Minnesota and adjacent Canada. Only the Des Moines Lobe affected Iowa, reaching a maximum there about 14,000 years ago (=Cary). The Wisconsin glacial history in Minnesota is described in five phases of ice advance, correlated by C^{14} dates with the Cary (possibly Tazewell) to Port Huron advances of the Lake Michigan Lobe. Interactions of ice lobes are recorded by erosion and redeposition of drifts, stratigraphic superposition, overriding of moraines, drumlins, tunnel valleys, eskers, and other landforms, formation of ice-bordered lakes, and blocking of outwash valley trains and lake-outlet stream channels. Glacial Lakes Duluth, Upham, Aitkin, and Agassiz record late stages in ice recession. The active ice was apparently wasted completely from Minnesota by the time of the Two Creeks interval, and it did not re-enter the area during the Valders phase.

OUTLINE OF GLACIAL GEOLOGY OF ILLINOIS AND WISCONSIN

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THE GEOGRAPHIC position of Illinois and Wisconsin makes them unique in the glacial history of North America, because glacial lobes from both east and west of Hudson Bay, Canada, extensively invaded them and attained the southernmost limit of continental glaciation in the northern hemisphere (Fig. 1). Perhaps also because of their geographic setting these two states contain deposits (Fig. 2) from the greatest number of glacial advances so far documented for any region in the United States (Fig. 3).

The presently used North American standard classification of Pleistocene glacial deposits was developed during the last three-quarters of a century in the upper Mississippi Valley region, and type localities for the Illinoian, Sangamonian, and Wisconsinan occur within Illinois and Wisconsin. Type sequences for the Illinoian and Wisconsinan are drawn from deposits made by the Lake Michigan lobe, whereas type sequences for the Nebraskan and Kansan, west of the Mississippi River, are drawn from deposits made by the western (Keewatin) glacial advance. The stratigraphic relation of the younger Pleistocene deposits to the Kansan is based on their overlapping relations in the Mississippi Valley, particularly in western Illinois.

The early descriptions of the surficial drift did not recognize their glacial origin (Worthen, 1866). However, shortly thereafter the drift was acknowledged to be the product of continental glaciers. Chamberlin (1878, 1880) seems to have been the first to use topographic form to interpret glacial movements and weathering of the drift to infer age. Later the presence of prominent and widespread buried soils served as the primary criterion for establishment of major episodes of glaciation (Leverett, 1889, 1909). Erosion of bedrock valleys and differences of stream-terrace levels added support to the concept of great time intervals between glaciations (Chamberlin, 1890; Hershey, 1893). Salisbury (1893) detailed 12 criteria for the recognition of distinct glacial epochs. This was followed by the first attempt to give geographic names to drifts of different ages (Chamberlin, *in* Geikie, 1894). Minor subdivisions and units for mapping purposes continued to be based largely on topographic expression (Leverett, 1899; Leighton, 1917; Alden, 1909, 1918; Thwaites, 1928a; Leighton and Powers, 1934). The single classification that developed in this manner was a mixture of units based on different kinds of criteria.

The present policy utilizes multiple schemes of stratigraphic classification (Willman, Swann, and Frye, 1958; A.C.S.N., 1961), as described for use in the Pleistocene (Frye and Leonard, 1952; this volume; Frye and Willman, 1960). The major time-stratigraphic units (stages) are recognized throughout the Midwest, and subdivisions of these units (substages) are recognized in these two states for the Illinoian and Wisconsinan. The deposits are objectively classed as rock-stratigraphic units of local extent and morphostratigraphic units, which may be even more local

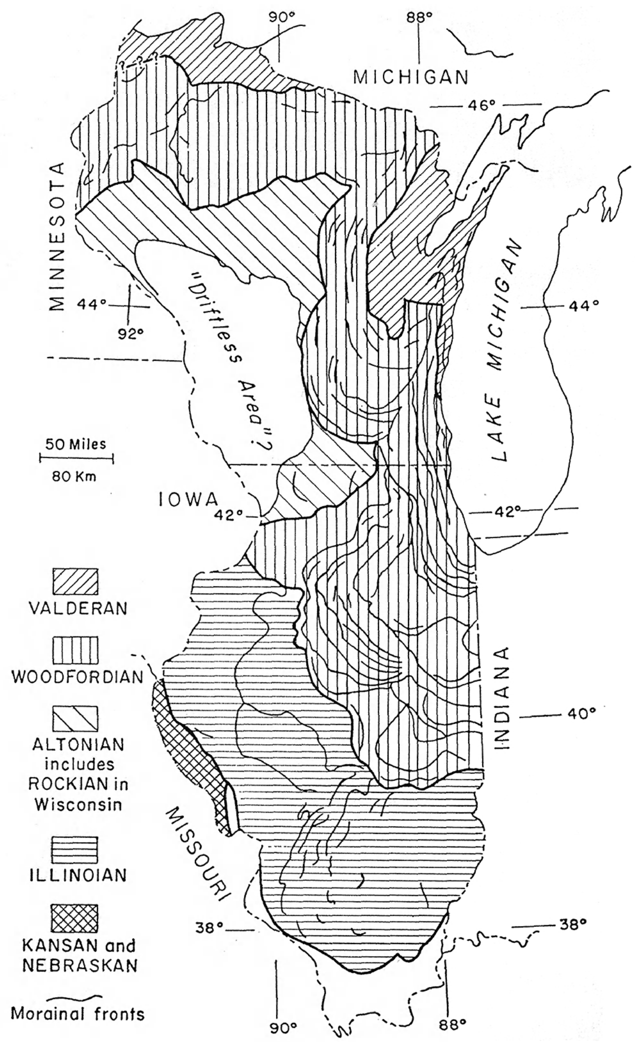

Figure 1. Map of Illinois and Wisconsin showing the distribution of surficial glacial deposits. The "Driftless Area" may have been partly or wholly covered by glacial ice during the Pleistocene. Modified from Thwaites (1956), Ekblaw (1960), and Flint *et al.* (1959).

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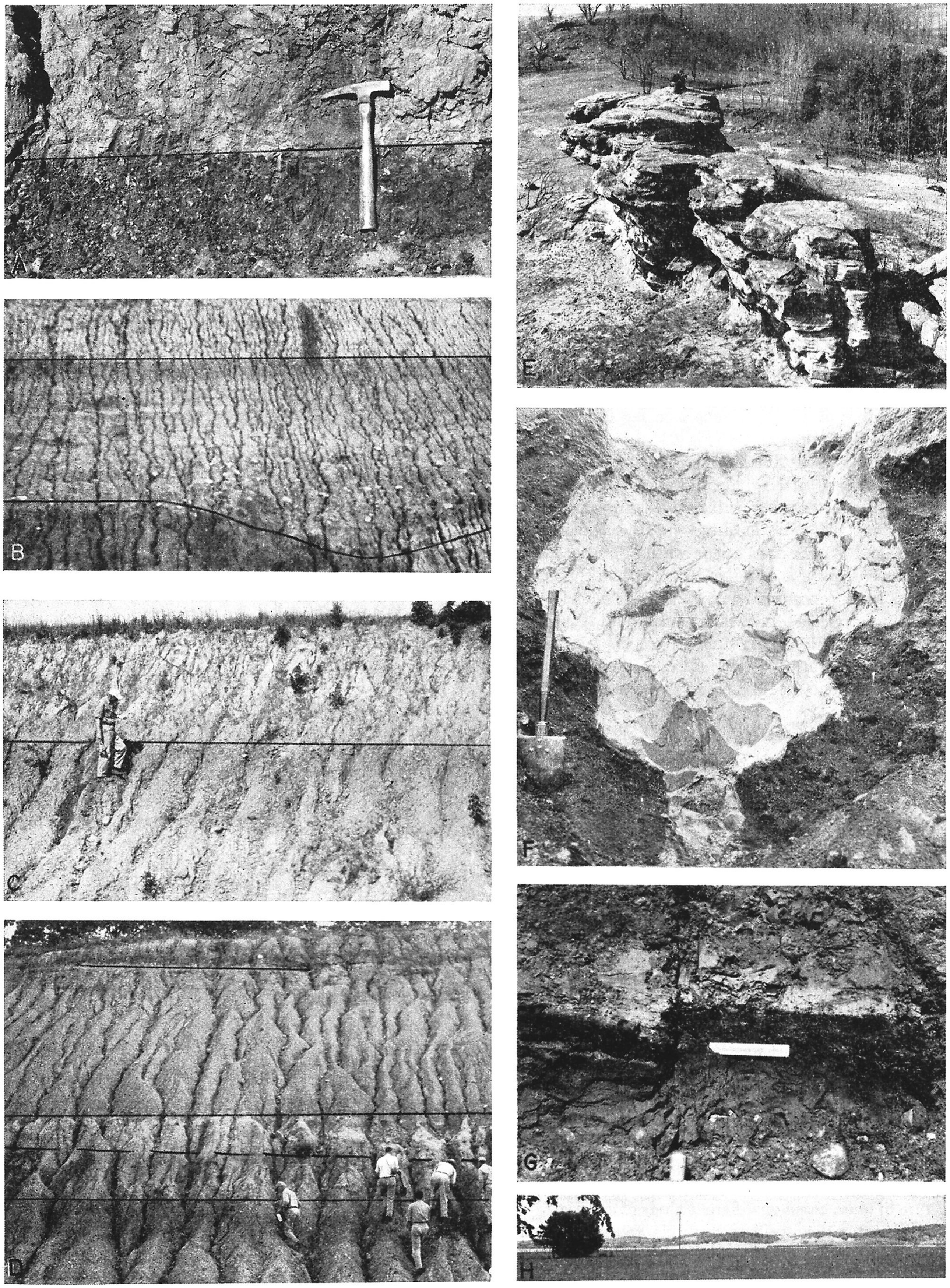

Figure 2. Pleistocene stratigraphy and surface features in Illinois and Wisconsin.

in extent. The development of these classifications is in a preliminary state. Soils are recognized as units in a separate stratigraphic classification.

Traditional methods of correlation are still of value in studying and correlating glacial deposits in the Illinois-Wisconsin area—intercalated buried soils, peat zones, gravel, and loess; fossils; general lithology; and topographic expression of end moraines, recessional moraines, and terraces. However, in recent years two relatively new techniques have greatly increased the precision and detail of correlation in this area. These are the extensive use of radiocarbon dates in the Wisconsinan (approximately 100 dates have been determined from the two states; Figure 5), and extensive studies of the mineral compositions of the fine sand, silt, and clay fractions of tills, loesses, and outwash by X-ray diffraction and optical-microscope analysis. This latter technique has been particularly useful in Illinois because of the diversity in source areas of the Pleistocene deposits, but it is applicable to all ages of Pleistocene deposits.

The rocks below the glacial deposits in the area range from Precambrian in northern Wisconsin upward through each system of Paleozoic rocks, including the Pennsylvanian (Weller *et al.*, 1945; Bean, 1939). Rocks of Cretaceous age occur to the northwest of these states and in small areas in extreme western Illinois (Frye, Willman, and Glass, 1964) and southwestern Wisconsin (Andrews, 1958). In general, however, the age of the bedrock becomes progressively younger southward through the region, and more than half of the glaciated area of Illinois is underlain by Pennsylvanian rocks.

Key to Figure 2:

A. Pro-Kansan silts (calcareous and sparsely fossiliferous) on Afton Soil developed in Nebraskan outwash. Zion Church geologic section, SE¼, SE¼, SW¼, sec. 9, T. 3 S., R. 8 W., Adams County, Illinois.

B. Peoria Loess (Woodfordian) and Roxana Loess (Altonian), on Sangamon Soil in Mendon till (Liman), on Yarmouth Soil in Kansan till. Rushville (4.5 W.) geologic section, NW¼, SW¼, SW¼, sec. 5, T. 1 N., R. 1 W., Schuyler County, Illinois.

C. Peoria Loess (Woodfordian) and Roxana Loess (Altonian) on Sangamon Soil in Buffalo Hart till (Illinoian). Hipple School North geologic section, NW¼, SW¼, SW¼, sec. 8, T. 7 N., R. 3 E., Fulton County, Illinois.

D. Richland Loess, on thick Shelbyville till, on Morton Loess (Woodfordian), on Farmdale Silt (Farmdalian), on accretion-gley of the Sangamon Soil on Buffalo Hart till (Illinoian). Farm Creek railroad-cut geologic section, cent. sec. 31, T. 26 N., R. 3 W., Tazewell County, Illinois.

E. Monument of St. Peter Sandstone (Ordovician) 5 km southeast of Monticello, Green County, Wisconsin (NW¼, SW¼, sec. 28, T. 3 N., R. 8 E.). The outcrop is at least 5 km inside the area covered by ice of late Altonian age.

F. Ice-wedge cast of clean yellow-brown sand in red-brown gravelly sand and sandy gravel of late Altonian (Rockian) age, River Falls, Wisconsin (SW¼, SW¼, sec. 36, T. 28 N., R. 19 W.).

G. Two Creeks paleosol on silts, overlain by Valders till, at depth of 365 cm, SW¼, SE¼, sec. 19, T. 23 N., R. 19 E., Outagamie County, Wisconsin.

H. Terminal moraine and outwash plain of the Chippewa lobe of middle Woodfordian "Cary" age in Chippewa County, Wisconsin. View northeast from a point 7 km north of Chippewa Falls, on highway 53.

The topography of Illinois and Wisconsin when the first continental glaciers advanced possessed significantly less relief than the present topography of the bedrock surface (Trowbridge, 1921; Frye, 1963). A plains topography may have existed at the position of the present uplands, and the major Nebraskan valleys were well above the present flood plains, which in turn are 150 to 250 ft (50 to 75 m) above the bedrock floors of the deeply filled valleys. Present data from Illinois indicate that the deep valleys were trenched in post-Nebraskan time, perhaps reaching their deepest incision during the Kansan, and that the maximum topographic relief of the Kansan has not been attained since then. Thwaites (1960) doubts that any erosional surfaces that formed before major valley incision have survived in southwestern Wisconsin.

Essential to a description of the glacial stratigraphy of these states is an understanding of the lobate configuration of the glaciers that advanced out of Canada into the region. The maximum extension of the several glacial lobes into Illinois is shown in Figure 4, and the arrows show, in a general way, the axial position of the several lobes. As these lobes crossed different parts of the Canadian Shield, they eroded different crystalline rocks and so are characterized by recognizable differences in their heavy-mineral suites. Because they crossed different belts of Paleozoic and Mesozoic rocks they are also characterized by differences in their assemblages of clay and carbonate minerals (Willman *et al.*, 1963). Distinctive rock types also characterize the pebbles and coarser materials found in the drift of some lobes (Anderson, 1957). The contrasting mineral compositions are shown by the averages for the several tills and loesses in Illinois given in Tables 1 and 2.

Five distinct glacial lobes advanced into the region, and one was complex within itself. Glaciers that entered Illinois and Wisconsin from the northwest dispersed from centers west of Hudson Bay that may be grouped as the Keewatin center. As glaciers from this dispersal center advanced on a broad front, the composition of the tills gradationally change across the area. In Illinois (Frye *et al.*, 1962; Willman *et al.*, 1963) the tills and loesses derived from this source are relatively rich in montmorillonite among the clay minerals and in hornblende and epidote among the heavy minerals, and they contain markedly more calcite than dolomite. In the tills of western Wisconsin, illite and montmorillonite prevail (Akers, 1961), and among the heavy minerals hornblende, epidote, pyroxene, magnetite, and garnet may dominate.

Lake Michigan and Green Bay define the position of two nearly parallel lobes that were particularly significant during the Illinoian and Wisconsinan. The dispersal center for glaciers in these lobes was in Ontario, slightly north-northeast of these two basins. Ice from Lake Superior also entered Green Bay. Tills from the Lake Michigan lobe are rich in illite among the clay minerals and in hornblende, garnet, and epidote among the transparent heavy minerals; they contain more dolomite than calcite. Tills from Green Bay are similar but contain less garnet and illite and more vermiculite and epidote. Precambrian pebbles are distinct between part of the Green Bay lobe and the Lake Michigan lobe (Anderson, 1957). Tills deposited by the

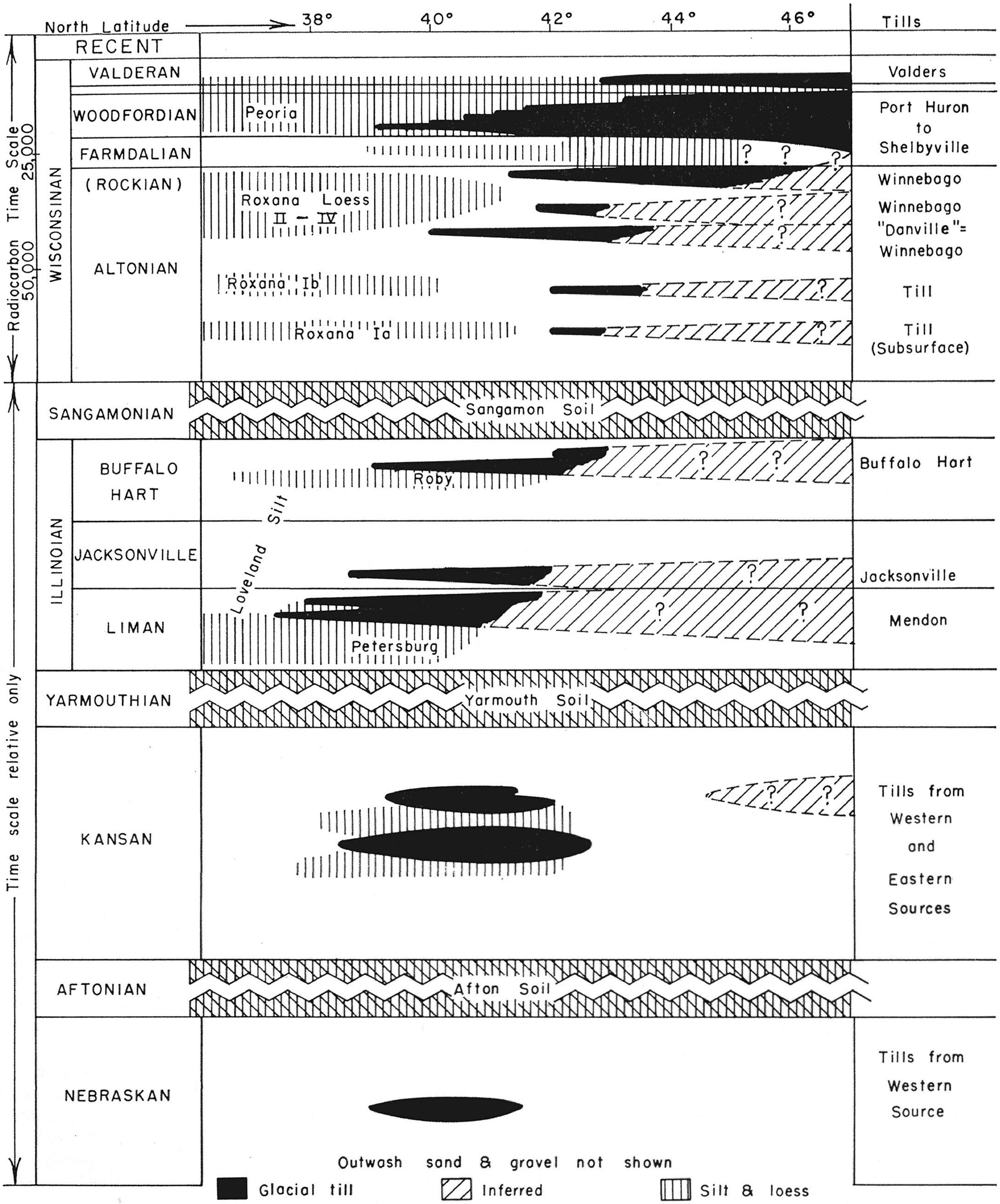

Figure 3. Time-space diagram showing the stratigraphic relations of Pleistocene glacial tills and loesses in Illinois and Wisconsin. The southernmost extent of each glacial advance is plotted without regard to east-west position within the states. Kansan and Nebraskan glaciers that invaded Illinois from the northeast or northwest did not reach into extreme northern Illinois or southern Wisconsin.

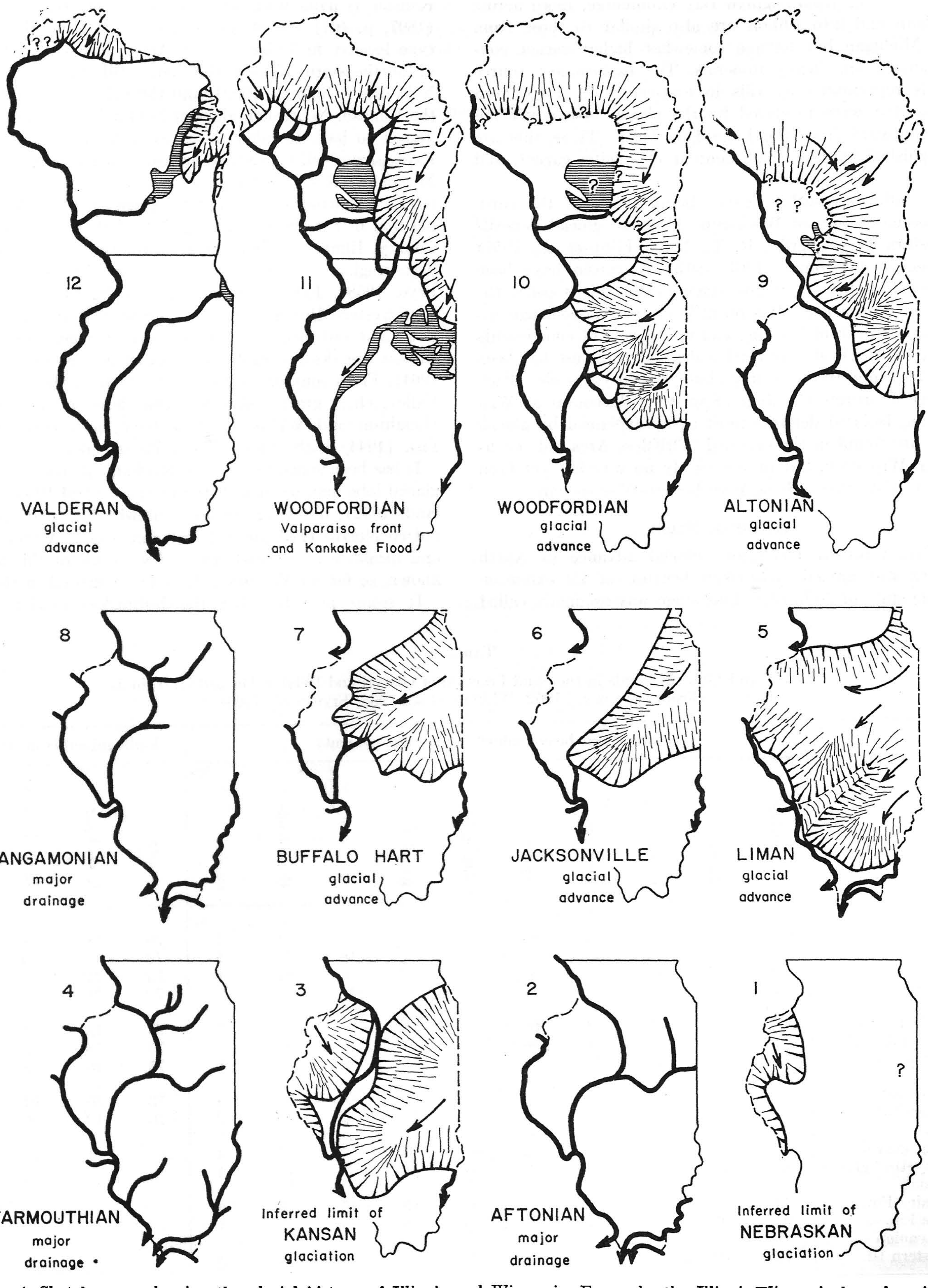

Figure 4. Sketch maps showing the glacial history of Illinois and Wisconsin. For scale, the Illinois-Wisconsin boundary is 150 miles (240 km) long.

glaciers moving from Saginaw Bay (Zumberge, 1960) across Michigan and into Illinois are also similar to those from Lake Michigan but have a somewhat higher garnet content among the heavy minerals. The easternmost source area is represented by tills in east-central and southern Illinois that were produced by the glacial lobe advancing southwestward from the Lake Erie basin. These tills are distinguished by their high content of calcite, garnet, and illite.

The following outline leaves blank much of the early Pleistocene history of Wisconsin. Although glacial deposits in western (Chamberlin, R. T., 1910; Flint *et al.*, 1959) and central (Weidman, 1907, 1913) Wisconsin have been correlated with each major stage of the Pleistocene, the oldest drift in central Wisconsin is clearly Wisconsinan according to depth of leaching and weathering of constituents (Hole, 1943). As its correlative drift to the west has been dated by radiocarbon as only about 30,000 years old, Black (1959a) interprets the drift of western Wisconsin as Wisconsinan. Isolated deposits most easily explained by glacial action are found in the classical "Driftless Area" of southwestern Wisconsin, but unfortunately no way has yet been found to date them. Some may be pre-Wisconsinan.

Nebraskan Stage

The first recorded Pleistocene glacial advance in North America was named Nebraskan because of its extension into the state of Nebraska. This stage was originally called Kansan (Chamberlin, 1894, p. 753-764), but after Bain (1897, p. 464) traced the overlying till from the original type locality in Union County, Iowa, into the surficial till of northeastern Kansas the lower till was renamed the Nebraskan (Shimek, 1909), and the till above was renamed the Kansan. Nebraskan glaciers from a Keewatin center are known to have invaded this area only in western Illinois. Exposures of till of this age are exceedingly rare east of the Mississippi River, but its presence at a few places (Wanless, 1957) demonstrates that the glacier crossed the present position of the river (Fig. 4). Upland deposits in Calhoun County, Illinois (Rubey, 1952), formerly supposed to have been unglaciated, may be Nebraskan till (Willman and Frye, 1958). Erratic boulders in the "Driftless Area" of northwestern Illinois may also represent Nebraskan glaciation, and outwash deposits at several places in western Illinois are demonstrably of Nebraskan age (Frye *et al.*, 1964). From southern Illinois southward in the Mississippi Valley, chert gravels generally considered of Tertiary age (Leighton and Willman, 1950) have been correlated by Fisk (1944) with stages of the Pleistocene.

It has been suggested that Nebraskan till from an eastern glacial lobe may occur in extreme east-central Illinois (Eveland, 1952), but evidence to establish this relationship is not conclusive (Ekblaw and Willman, 1957). Although several localities of outwash and a few of till of this age are known, so far no Nebraskan loess is recognized in the area.

It seems probable that the Nebraskan glacier largely

TABLE 1

Heavy and Light Minerals in the Sand Fraction of Glacial and Related Deposits in Illinois
(From Frye *et al.*, 1962; Willman *et al.*, 1963; Frye *et al.*, 1964)

Stratigraphic unit	Transparent heavy minerals (average percent)										Light minerals (average)			
	Tourmaline	Zircon	Garnet	Epidote	Rutile	Sillimanite	Kyanite	Staurolite	Hornblende	Others	Quartz	K Feldspar	Na-Ca Feldspar	Others
Wisconsinan														
Woodfordian till	1	2	12	14	tr.	—	tr.	—	64	7	70	19	8	3
Peoria Loess	2	9	8	22	—	—	—	tr.	51	8	72	15	9	4
Winnebago till	2	1	12	17	tr.	—	—	tr.	62	6	72	17	8	3
Roxana Loess	2	11	11	24	—	—	—	tr.	46	6	72	16	9	3
Illinoian														
Buffalo Hart till	3	3	12	11	tr.	—	tr.	—	66	5	73	13	8	6
Jacksonville till	2	2	16	16	—	—	—	—	57	7	70	18	7	5
Mendon till	3	2	16	15	tr.	—	tr.	—	55	9	73	13	9	5
Kansan														
Eastern till	1	1	21	5	—	—	—	—	58	14	76	10	10	4
Western till	4	4	10	23	1	—	tr.	2	49	7	76	12	8	4
Tertiary														
Grover gravel	19	60	2	5	2	—	2	5	1	4				
"Lafayette" gravel	11	36	3	2	2	6	16	32	—	1				
Cretaceous														
McNairy Fm.	11	8	tr.	1	6	8	45	15	—	6				
Baylis Fm.	31	45	1	1	3	—	3	14	1	1				
Pennsylvanian in western Ill.	22	60	3	tr.	11	1	—	—	—	3				

tr. = trace
Fm. = formation

TABLE 2

Clay and Carbonate Minerals in Glacial Deposits in Illinois: by Stratigraphic Unit (Analyses by H. D. Glass in Frye *et al.*, 1962; Willman *et al.*, 1963; Frye *et al.*, 1964)

Stratigraphic unit	Clay minerals (average percent)			Carbonate minerals
	Montmorillonite	Illite	Kaolinite and chlorite	(Ca = calcite; Do = dolomite)
Wisconsinan				
Woodfordian till	4	79	17	Do > Ca
Peoria Loess	65	24	11	Do > Ca
Winnebago till	25	59	16	Do > Ca
Roxana Loess	63	17	20	Do > Ca
Illinoian				
Buffalo Hart till	5	69	26	Do > Ca
Jacksonville till	21	68	11	Do > Ca
Mendon till	36	45	19	Do > Ca
Petersburg Silt	66	19	15	Do > Ca
Kansan				
Western till (west of about 90°30′ W. Long.)	66	13	21	Do < Ca
Western till (east of about 90°30′ W. Long.)	43	35	22	Do = Ca
Eastern till	8	65	27	Do < Ca

determined the position of the Ancient Mississippi River throughout its course above the mouth of the Missouri River and, less directly, the position of the river southward to the head of the Embayment region.

AFTONIAN STAGE

The Aftonian Stage, defined from water-laid sediments at Afton Junction, Iowa (Chamberlin, 1894), is known with certainty in Illinois only as the Afton Soil, which is exposed at a few localities. The Afton Soil has recently been described in Adams County (Frye *et al.*, 1964), where it occurs in Nebraskan outwash gravels (Fig. 2A) overlain by calcareous pro-Kansan silts and Kansan till. In the same part of western Illinois, peaty deposits formerly classed as Aftonian or older (Horberg, 1956) are now judged to be early Kansan in age, but compact, greenish gray, noncalcareous silt present locally beneath the Kansan till may be Aftonian or older. At several places in western Illinois, weathering profiles below known Kansan deposits and on bedrock, Cretaceous sand, or lag gravel from Nebraskan deposits could represent an Afton Soil, if it is assumed that Nebraskan glaciers eroded the preglacial soil.

KANSAN STAGE

The Kansan Stage, redefined from deposits in Union County, Iowa, is named from a type area in northeastern Kansas (Frye and Leonard, 1952). Type Kansan is based on till deposited by the Keewatin (western) glacier that invaded western Illinois from Iowa and Missouri (Fig. 4). A till in eastern Illinois (MacClintock, 1929) that was deposited by an eastern (Erie) glacial lobe is correlated with the Kansan of western Illinois, because they have the same relation to the overlying Illinoian till and because their associated water-laid deposits contain a similar fauna of fossil mollusks, but the physical stratigraphic relations of these tills from eastern and western sources have not been as yet firmly established (Fig. 4).

Although Kansan till of eastern source occurs nowhere in Illinois as the surface till, western (Keewatin) Kansan till in an area in Adams and Pike Counties (Frye *et al.*, 1964) is the uppermost glacial deposit (Figs. 1 and 4). Also, at several places west of the Illinois River Valley Kansan till is well exposed below Illinoian till (Willman *et al.*, 1963), which overrode it from the northeast (Fig. 2B). Water-laid silt, sand, and gravel within the western Kansan drift may in the future serve as a basis for subdivision of Kansan drift.

Till and associated glacial-fluvial deposits on high terraces of the Wisconsin River have been correlated with the Kansan Stage (MacClintock, 1922). They were partly derived from the northwest and may temporarily have reversed the river. Thwaites (1928b) disagreed and traced some other possibly equivalent terraces of the "Driftless Area" to the drifts in central Wisconsin that are now considered Wisconsinan (Hole, 1943).

In both eastern and western Illinois, water-laid and loess-like silts are associated with the lower part of the Kansan till or occur conformably below it. Such silts are generally calcareous and at several places have yielded fossil snail faunas. At no place has evidence of weathering been observed between such silts and the overlying till, but at several places these calcareous Kansan silts rest on an Afton Soil. Water-laid or loessial deposits of Kansan age have not been observed resting on Kansan till. The upper surface of the Kansan till, except where dissected by subsequent erosion, generally is deeply weathered.

Deposits of Kansan age have not as yet been identified in the subsurface of northernmost Illinois or in eastern Wisconsin, but Kansan drift in north-central Illinois has a mineral composition more like the Illinoian of the Lake Michigan lobe than the eastern Kansan of the Erie lobe and may indicate the existence of a Lake Michigan glacial lobe during Kansan time (Johnson, 1964).

Yarmouthian Stage

The name Yarmouth was proposed by Leverett (1898a) to apply to the soil that separates the tills of Kansan and Illinoian age in east-central Iowa. In extreme western Illinois the Yarmouth Soil is overlain by Illinoian Loveland Silt. Along the Illinois River Valley (Fig. 2B) and farther east in central Illinois it is overlain by tills of Illinoian age. The soil commonly shows a poorly to moderately well-drained profile, developed *in situ,* but at several places it consists of accretion-gley deposits as much as 5 to 6 feet thick. In southern Illinois the Yarmouth Soil has been developed in Paleozoic rocks and is directly overlain by Illinoian till and by Loveland Silt (Frye *et al.,* 1962), but this soil has not as yet been identified in extreme northern Illinois or in Wisconsin.

Great lake trout and other fossils in northern Wisconsin, formerly correlated (Hussakof, 1916) with the Yarmouthian Stage, are now considered Wisconsinan.

Illinoian Stage

The name Illinois till sheet was first proposed by Leverett (*in* Chamberlin, 1896), and shortly thereafter he (Leverett, 1899) documented the extent of Illinoian glaciation in central and southern Illinois. The stage was defined as the deposits lying above the Yarmouth Soil and below the Sangamon Soil (Fig. 2B, C) in western, central, and southern Illinois. Much of the area within North America in which deposits of this age constitute the surficial glacial deposits lies in the state of Illinois (Fig. 1), and it is here that its stratigraphic position above Kansan tills and below Wisconsinan tills has been demonstrated. The drift formerly called Illinoian in northern Illinois and southern Wisconsin (Alden, 1918; Leighton and Brophy, 1961) is here considered Wisconsinan, as shown by Shaffer (1956).

Glacial lobes in the position of the Lake Michigan Basin have not been identified for the Nebraskan nor with certainty for the Kansan, but in Illinoian time a lobe in this position attained particular prominence (Fig. 4). Whereas western (Keewatin) glaciers had invaded the western part of Illinois during the two previous advances, Illinoian glaciers did not reach Illinois from the west. A glacial advance from the northeast (Saginaw or Erie lobe), however, extended into southeastern Illinois, deflecting the Lake Michigan lobe glacier to the west, and attained the southernmost extent of any Pleistocene continental glacier in the northern hemisphere.

Three substages (time-stratigraphic units) are recognized within the Illinoian of Illinois (Fig. 3). The oldest (lowest) of these, formerly called Payson, has recently been redescribed and renamed Liman (Frye *et al.,* 1964), the second is called Jacksonville, and the youngest Buffalo Hart.

Beyond the limit of Illinoian glaciation, Illinoian loess and water-laid silts and sands are commonly deeply weathered and have the Sangamon Soil in their uppermost part; these silts are in turn overlain by Wisconsinan loesses. Because of the correlation of these deposits with loess and sandy loess called Loveland in western Iowa and the Great Plains region (Frye and Leonard, 1951), they are classed as Loveland Silt in Illinois (Leighton and Willman, 1950). As the largest and most extensive Illinoian silt unit is the Petersburg Silt of early Illinoian age, it is presumed that a major part of the Loveland is of the same age. However, the Loveland includes silt and sand deposits representing the entire Illinoian. At some places in western Illinois at least one minor soil occurs within the Loveland Silt.

Liman Substage

The Liman Substage was defined from the Pryor School section in Lima Township, Adams County, Illinois (Frye *et al.,* 1964). At this locality, and generally throughout western and south-central Illinois, it includes two rock-stratigraphic units—the Petersburg Silt at the base and the Mendon and equivalent tills above.

The Petersburg contains both loess and water-laid silts and fine sands, it attains a maximum thickness of more than 30 ft (9 m), and locally it contains an abundant fauna of fossil snails. At many places it rests on a well-devoloped Yarmouth Soil developed in Kansan till or in bedrock, but a weathered zone separating it from the overlying till has nowhere been observed. The Petersburg is interpreted as loess, outwash, and lacustrine sediments deposited during the period of advance of the earliest Illinoian glacier. The mineralogy of this unit in the Illinois and Mississippi Valley area shows a striking change from bottom to top, reflecting the composition of the underlying western Kansan till in the base and the Illinoian till of the advancing Lake Michigan lobe in its upper part.

Whereas tills of Liman age in western and central Illinois were deposited by the glacial lobe that advanced through the basin of Lake Michigan, tills of this age in southern and east-central Illinois were deposited by a glacial lobe from the northeast, probably the Saginaw lobe (Figs. 1, 4). The two areas are separated by an interlobate belt of ridged drift. These earliest Illinoian tills are distinguishable, on the basis of their mineral composition (Tables 1, 2), from each other and from the Kansan tills below (Willman *et al.,* 1963). In southwestern Illinois a thin zone of silt, locally containing sparse fossil snail shells, occurs within the tills of Liman age.

The most striking characteristic of Liman deposits is the vast expanse of essentially featureless till plain in south-central and west-central Illinois. In no other area of North America is found a region underlain by glacial till that so completely lacks both topographic features of glacial deposition and erosional dissection. In fact, it resembles the topography of the High Plains except for the sharp ridges and hillocks that locally demonstrate the glacial origin of the topography. Although many of these ridges have been ascribed to crevasse fillings in stagnant ice (Leighton and Brophy, 1961; Leighton, 1959), their preponderant composition of glacial till and their topographic shape and large size prompt us to interpret them as interlobate ridges rising above a till plain that has been smoothed by local erosion and deposition, as indicated by abundant accretion gleys.

Glaciers of Liman age overrode the Ancient Mississippi

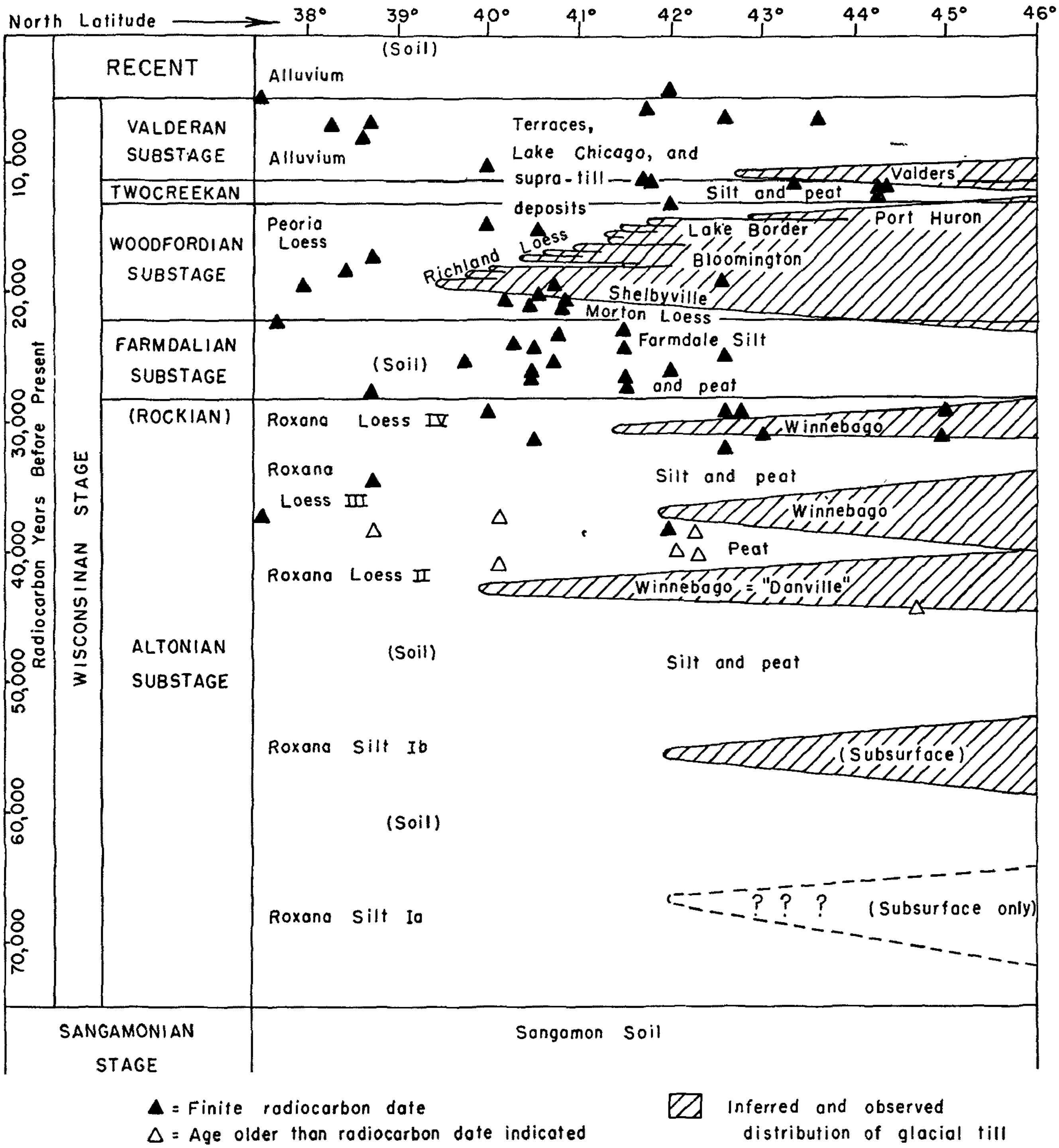

Figure 5. Time-space diagram showing the relation of available radiocarbon dates to the Wisconsinan glaciers in Illinois and Wisconsin. The southernmost extent of each glacial advance is plotted without regard to east-west position within the states. Each radiocarbon date symbol indicates a single determination except for the Two Creeks locality in Wisconsin where the symbols represent averages of many dates. The majority of the dates shown were determined by the U.S. Geological Survey in the Washington Laboratory (W), but some dates are included from Chicago (C), Isotopes, Inc. (I), Lamont (L), and Yale (Y).

Valley and forced the river into a position across eastern Iowa (Schoewe, 1920). When the Illinoian glacier retreated, the river returned to the position of the major valley through Illinois (Fig. 4), from which it was later permanently deflected by the early Woodfordian glacier (Shaffer, 1954).

JACKSONVILLE SUBSTAGE

The name Jacksonville was first applied by Ball (1937) to a moraine that occurs just west of Jacksonville, Illinois, and the till of this area was later used as the type for the substage (Leighton and Willman, 1950). The substage includes *the till and outwash that occur stratigraphically* above the Mendon till and below the Roby Silt (Johnson, 1964). As neither a soil profile nor extensive deposits of fossiliferous silts have been observed separating the Jacksonville till from the Mendon till below, it is judged that this glacial advance was not separated from the Mendon by a long interval of time.

West of Illinois Valley, the Buffalo Hart glacier extended beyond the limits of the Jacksonville, and in this area the Jacksonville is represented by extensive deposits of outwash gravel, sand, and varved clays between the Mendon till below and Buffalo Hart till above.

BUFFALO HART SUBSTAGE

The Buffalo Hart Moraine was described by Leverett (1899) from northeastern Sangamon County, Illinois, and the deposits of that area were used as the type for the Buffalo Hart Substage by Leighton and Willman (1950). As now used, the substage includes the Roby Silt (Johnson, 1964) and the overlying Buffalo Hart till. This till, which is recognized in a relatively small area in central Illinois, is distinguished from the earlier Illinoian by its surface topography and by its position above the fossiliferous Roby Silt. The Buffalo Hart till plain topographically resembles the early Woodfordian more than it does the earlier Illinoian, but it is sharply distinguished from the overlying younger tills and loesses by the presence of the Sangamon Soil in its top.

Sangamonian Stage

The Sangamon Soil, first named by Worthen (1873) from exposures in Sangamon County, Illinois, was used by Leverett (1898b) as a basis for an interglacial interval; the Sangamonian Stage is based on the Sangamon Soil. Tills of both Buffalo Hart and Jacksonville age occur in Sangamon County, but the name Sangamon Soil, as a soil-stratigraphic unit, is applied to the soil where developed on Liman, Kansan, or older deposits. The stage is named from the Sangamon Soil as developed in Buffalo Hart till, and therefore it encompasses only the span of time from the retreat of the youngest Illinoian glacier to the deposition of the earliest Roxana (Wisconsinan) Loess (Fig. 3).

Throughout most of central Illinois the Sangamon Soil, where formed on Illinoian till (Fig. 2B, C), is generally a poorly drained profile with a gray-brown to dark gray-brown B-zone. However, in western and extreme southern Illinois where the soil is commonly developed on silt, sand, or bedrock, it has a relatively well-drained profile with red-brown to dark red B-zones. The Sangamon Soil has been identified in southwestern Wisconsin (Hogan and Beatty, 1963).

At many places in Illinois the Sangamon Soil (Fig. 2D) is not a profile formed *in situ* but consists of a deposit called accretion-gley (Frye, Shaffer, *et al.*, 1960; Frye, Willman, and Glass, 1960; Frye and Willman, 1963b). This material (formerly called gumbotil: Leighton and MacClintock, 1930, 1962) accumulated slowly in the initial shallow swales on the till-plain surface as a result of sheet wash from the slightly higher parts of the till-plain and perhaps the minor addition of eolian silt. Such deposits show less mineral decomposition than the B-zones of the profiles formed *in situ* and are gray in color and rich in montmorillonite. At places they contain organic zones and streaks, and locally they are inconspicuously bedded. Deposits of Sangamon accretion-gley attain a maximum thickness of more than 6 ft (2 m), and they are essentially the only deposits in the region that can be demonstrated to be of Sangamonian age.

Wisconsinan Stage

In 1894 Chamberlin proposed the name East Wisconsin Stage of Glaciation to include the glacial deposits extensively exposed in the southeastern part of that state and extending into central Illinois, where the relation to the underlying Sangamon Soil and Illinoian drift could be demonstrated. In 1895 Chamberlin shortened the name to Wisconsin, and since then this region has been generally accepted as the type area for the youngest of the major episodes of continental glaciation in the United States. In conformance with modern stratigraphic practice, the adjectival form, Wisconsinan, is now widely used to apply to the time-stratigraphic unit.

Wisconsinan glaciers covered most if not all of Wisconsin and most of north and east-central Illinois (Fig. 4). Loess deposits of Wisconsinan age form a surficial mantle over much of the area (Fig. 6). Black believes that Wisconsinan glaciers covered the "Driftless Area" of Wisconsin, but Frye and Willman consider the scattered erratic cobbles in the Illinois part of the "Driftless Area" as more likely to be remnants of an old glaciation, probably Nebraskan.

As early as 1878, Chamberlin recognized differences in the Wisconsinan drifts, and he postulated a two-fold glacial advance (Chamberlin, 1883a, b). The major framework of Wisconsinan classification was established by Leverett in 1899 when he recognized an Early Wisconsin with four substages and a Late Wisconsin with three substages, distinguished mainly by topography and direction of ice flow. Naming and correlation of the various substages was largely developed on the basis of topographic expression of moraines with only partial agreement (Leighton, 1960; Flint, 1963; Frye and Willman, 1963a). The subdivision followed here is based primarily on the stratigraphy of the till sheets and related loess sequences and on radiocarbon dates (Frye and Willman, 1960, 1963b; Frye *et al.*, 1962) with some supporting data from paleontology (Leonard and Frye, 1960) and from subsurface studies (Kempton, 1963; Kempton and Hackett, 1964). It uses the sequence of deposits related to the Lake Michigan lobe in Wisconsin and Illinois as a type section, and it recognizes the widely traced and radiocarbon-dated zones of weathering and peat accumulation as being better time planes for differentiation of time-stratigraphic substages than the morainal fronts. Subdivisions based on morainal fronts are useful for local but not for interlobe correlation, which reduces such units to subdivisions of substages.

Sub-substages are currently not recognized as formal units in time-stratigraphic classification, but for comparison with older literature the following informal subdivisions are used:

Early Woodfordian "Tazewell" refers to the deposits previously called Iowan (of Illinois), all of the Tazewell, and the earliest Cary moraines as far as the morainal front represented by the Valparaiso and West Chicago moraines in Illinois and the Johnstown and correlative moraines in Wisconsin.

Middle Woodfordian "Cary" refers to the deposits related to the Valparaiso, West Chicago, and Johnstown moraines as far as the front of the unnamed moraine at Sheboygan, Wisconsin (Thwaites and Bertrand, 1957, Fig. 1), correlated with the Port Huron moraine in Michigan but omits the youngest moraines that were generally included in the Cary before 1957.

Late Woodfordian "Mankato" refers to the deposits related to the unnamed moraine at Sheboygan, Wisconsin, and the younger deposits older than Two Creeks. Before 1957 these deposits were included in the Cary but were renamed Mankato when type Mankato was found to be older than the Two Creeks deposits. In the pre-1957 literature, the term Mankato refers generally to the deposits younger than the Two Creeks deposits; they are called Valderan in this report.

The relation of tills and loess deposits to radiocarbon dates and to the present system of classification is shown in Figure 5. Radiocarbon ages older than those shown in Figure 5 are inferred by extrapolation from older finite dates of deposits in Ontario and Ohio. Five substages (Altonian, Farmdalian, Woodfordian, Twocreekan, and Valderan) are now recognized within the Wisconsinan Stage, and of these the earliest, the Altonian, includes approximately half of the time span of the stage (Fig. 5). Topographic expression, or the form of the glacial deposits, is still used and was formalized by the recognition of morphostratigraphic units as a basis for local mapping, particularly within the Woodfordian of Illinois, as shown in Figure 7 (Ekblaw, 1960; Flint *et al.*, 1959).

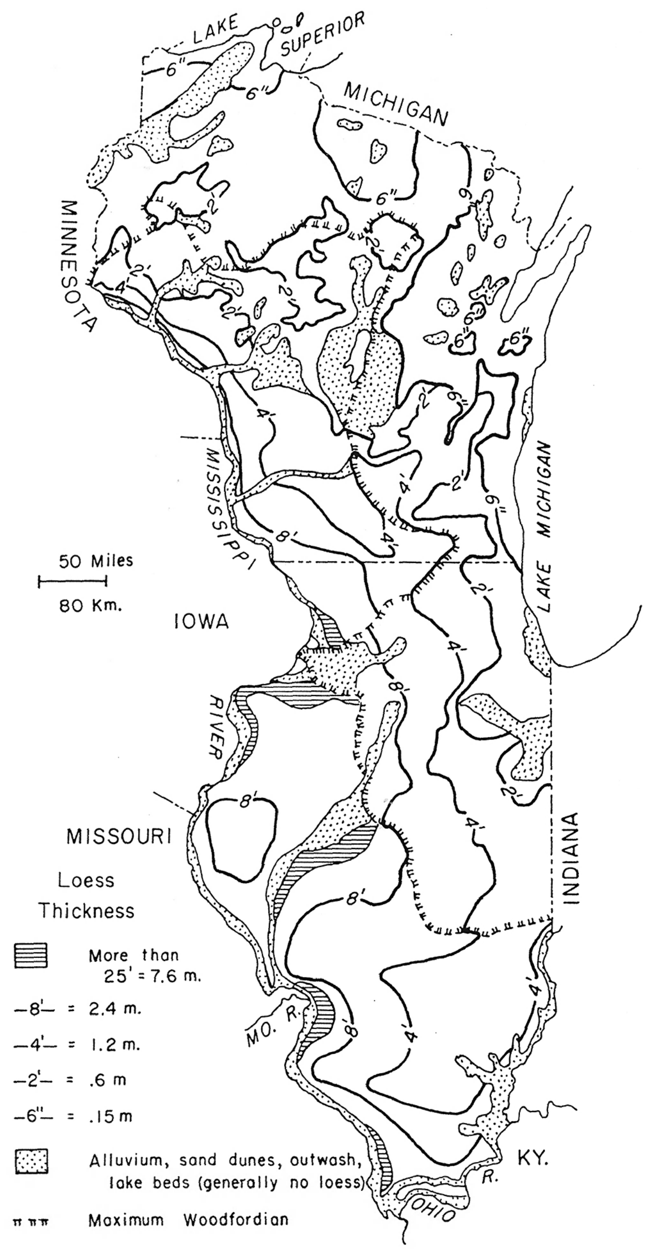

Figure 6. Map of Illinois and Wisconsin showing the generalized thickness of loess. Beyond the limit of Wisconsinan glaciation the thickness includes Roxana (Altonian) and Peoria (Woodfordian and younger) Loesses, in the area of Altonian tills it consists of Peoria Loess, and inside the Woodfordian maximum it consists of Richland Loess (middle to late Woodfordian and younger). Modified and generalized from Smith (1942), Leighton and Willman (1950), and Hole (1950).

ALTONIAN SUBSTAGE

The Altonian Substage is named from Alton, Illinois (Frye and Willman, 1960) and is based on the succession of silts and loesses called the Roxana Silt. Many of the deposits formerly called Late Sangamon Loess (Leverett, 1899; Leighton, 1926; Smith, 1942; Wascher *et al.*, 1948) and Farmdale Loess (Leighton and Willman, 1950; Wanless, 1957; Leighton, 1960) are now assigned to the Roxana.

The Roxana Silt, which is extensively exposed along the Illinois and Mississippi Valleys of central and southern Illinois, was largely deposited as loess derived from the outwash valley trains of the Ancient Mississippi and Ancient Iowa Rivers (Frye *et al.*, 1962). It commonly rests on Sangamon Soil, and five zones have been recognized within it (Fig. 5). Zone Ia at the base consists of silt and sand and locally contains small pebbles; it is strongly weathered and at many places is colluvium. Zone Ib is loess, sandy loess, or sand; it is locally calcareous, and a few snail shells have been found within it, but it commonly is weathered in at least the upper part and locally contains a humic zone at the top. Zone II is calcareous and fossiliferous in the thick sections along the valley bluffs; shells from the top of this zone gave radiocarbon dates 35,200 ± 1,000 (W-729) and 37,000 ± 1,500 (W-869) from localities nearly 200 miles apart. These dates indicate that this zone is contemporaneous with silts between Winnebago tills in northern Illinois dated 38,000 ± 3,000 (I-847). Zone III also is calcareous and fossiliferous in the thick sequences but is distinguishable from Zones II and IV by its color—gray in contrast to the pink below and above. Zone IV is generally leached of carbonates, at least in its upper part. Away from the valley bluffs (Frye *et al.*, 1963; Frye *et al.*, 1962; Smith,

1942) the Roxana becomes thin and is leached throughout, but the zonation is distinguishable by degree of weathering, texture, and color.

Tills now classed as Altonian were recognized below the surface drifts of northeastern Illinois (Horberg, 1953), but it was not until the work of Shaffer (1956) that the surface tills of northwestern Illinois and southern Wisconsin were demonstrated to be of early Wisconsinan age. Subsequently, subsurface studies (Kempton, 1963; Kempton and Hackett, 1964) in northern Illinois have shown the complexity of the Altonian glacial deposits, and it now appears (Fig. 5) that glacial advances in northern Illinois were approximately contemporaneous with each of the five zones of the Roxana Silt of the central and southern part of the state. The surface drift of southern Wisconsin seems equivalent to the latest Altonian only.

In Wisconsin, simultaneous advances of ice, one southeastward from the Des Moines and Superior lobes and the other westward from the Lake Michigan and Green Bay lobes, took place about 29,000 to 32,000 years ago, according to radiocarbon dates of spruce logs overrun or incorporated in the drift. The lobes overrode a residual soil rich in illite and chert and joined in the center of the state. This latest Altonian advance (locally termed Rockian) covered much if not all of the "Driftless Area" but did little work there (Black, 1962).

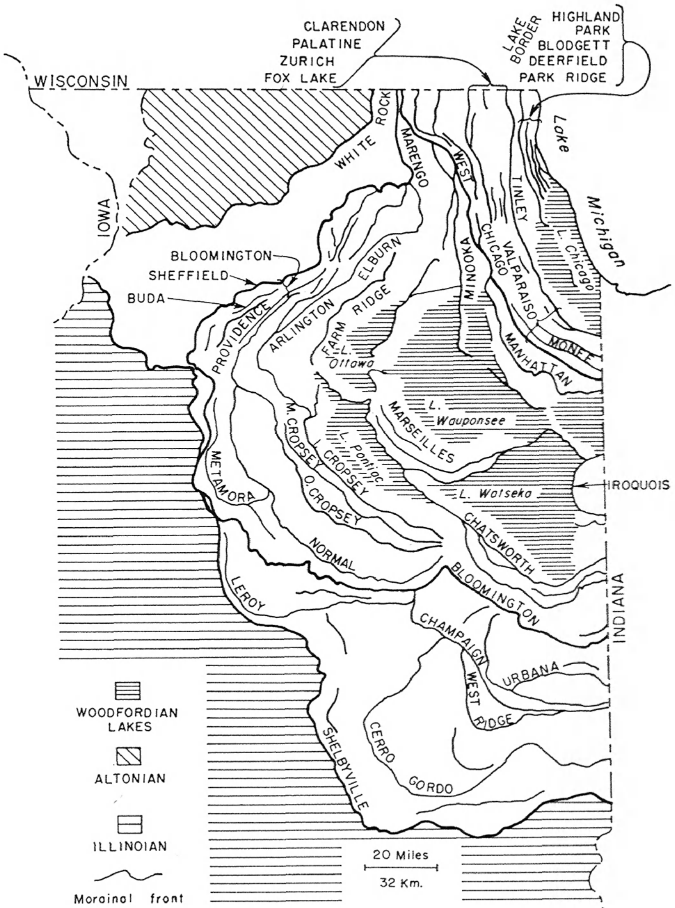

Figure 7. Woodfordian moraines and lakes in Illinois. Modified from Ekblaw (1960), Flint *et al.* (1959), and other sources.

Positive evidence of glaciation in the "Driftless Area" of Wisconsin comes from abundant fragments of Precambrian igneous rocks and Paleozoic chert and sandstone that rest on younger formations. Erratics of sedimentary rocks are especially abundant in the central and northern parts of the area. 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 erratics are also found on bedrock terraces along the Wisconsin River near Muscoda. Large angular to subangular blocks of dolomite occur in sand and gravel in foreset beds that indicate a flow of the Wisconsin River reverse to the present. This gravel of apparently western source contains unusual concentrations of agates similar to deposits of Rockian age in west-central Wisconsin. The dolomitic gravel is covered with chert-rich gravel, sand, and silt that contains 1-2% fresh igneous rocks and also blocks of Baraboo quartzite up to 1.5 m, indicating a northeasterly source.

Rubble deposits on the uplands of the Wisconsin "Driftless Area" are mostly angular chert and sandstone, but locally they contain well-rounded and polished pebbles thought to be derived from the Windrow Formation of Tertiary or Cretaceous age. These deposits commonly have anomalous clay-mineral suites (Akers, 1961); they also contain fresh glauconite topographically above known sources. The stratification and form of the deposits suggest kames.

Direction of ice flow may be inferred from boulder trains. For example, southeast of a sandstone pinnacle near Bosstown in the central part dozens of angular blocks of sandstone up to 5 m in length extend about 200 m obliquely across a gentle slope underlain by dolomite with clay residuum. A few fresh blocks have accumulated at the base of the pinnacle on the west, north, and east sides, but the disparity in concentration is so striking that it suggests glaciation. Overturned folds in residuum in several places on flat uplands also suggest a southeastward ice movement.

The absence of preserved loess older than 30,000 years in the area, the paucity of clay and chert residuum on flat uplands or in all valleys but one (South Fork of Baraboo River at Hillsboro), and lack of weathering of exposed igneous rocks suggest ice cover of much if not all the Wisconsin "Driftless Area" as recently as Rockian time. The Rockian drift in places resembles the younger Woodfordian and Valderan drifts in texture and in content of clay minerals, heavy minerals, and pebble types (Akers, 1961; Ellsworth, 1932; Oakes, 1960). A pre-Wisconsinan age for some of it can neither be confirmed nor denied.

Rock monuments and pinnacles have been widely interrupted to indicate lack of glaciation (Martin, 1932, p. 91) or antiquity of glaciation (Leighton and Brophy, 1961, p. 29) in the "Driftless Area" of Wisconsin, or they have been ignored where striae on outcrops of easily weathered basic-igneous rocks and other evidence attest to the recency of glaciation (Martin, 1932, Pl. 28). It is clear that equally large monuments were produced in northwest Wisconsin after definite glaciation during Woodfordian times. Most monuments in the "Driftless Area" are locally cemented sandstone (Fig. 2E) in the very friable St. Peter Formation, and most are at the ends of spurs or on the southeast sides of large high knobs. In general only minimal amounts of loose sand need to be removed from cliffs to isolate the cemented zones. For example, Footville monument in southern Wisconsin (Leighton and Brophy, 1961, p. 29), which is 19 ft (5.7 m) high, was exposed about 30,000 years ago. The loose sand needs to be washed or blown off the surface only at a maximum rate of 0.2 mm per year to round the cliff and isolate the monument. Abundant ice-wedge casts in southwest Wisconsin (Fig. 2F; Black, 1964) indicate permafrost conditions some time during the period from about 28,000 to 12,500 years ago. Strong mass movements, the inferred absence of trees, and presence of ventifacts suggest that rapid erosion under strong periglacial conditions was commonplace.

FARMDALIAN SUBSTAGE

The Farmdale Substage was defined by Leighton and Willman (1950) from the Farmdale silt and peat as it was exposed along Farm Creek, east of Peoria, Illinois. The name Farmdale Substage was used to encompass all Wisconsinan deposits occurring below the Morton (then called Iowan) Loess and above the Sangamon Soil (Fig. 2D). With the discovery of a complex sequence of loesses and tills stratigraphically below the type Farmdale but above Sangamon Soil, Frye and Willman (1960) redefined the Farmdalian Substage as including only the span of time represented by the Farmdale silt and peat in its type area, a zone that has been widely dated by radiocarbon throughout the Midwest; it comprises those deposits between the Roxana Loess and the Morton Loess (Fig. 5) with a radiocarbon age of about 22,000 B.P. to 28,000 B.P. (before the present) or perhaps slightly more.

The Farmdale Silt is known primarily from central and north-central Illinois where it contains eolian, colluvial, and lacustrine deposits, peat, and organic silts. Although found largely in the areas where the peat was buried by younger tills, Farmdale peaty silt is known at a few places capping terrace remnants in the Illinois Valley. Isolated small remnants of paleosols in southwestern Wisconsin may be Farmdalian (Hogan and Beatty, 1963). The Farmdalian Substage is an interval of major glacial withdrawal, but, as the sediments and fauna indicate cool moist conditions, it is judged that continental deglaciation did not occur. In Wisconsin, permafrost and buried glacial ice survived through the substage.

WOODFORDIAN SUBSTAGE

The Woodfordian Substage was named (Frye and Willman, 1960) from Woodford County, Illinois, and defined as representing the span of time from the base of the Morton Loess (top of the Farmdale Silt) upward through the overlying succession of tills to the base of the Two Creeks deposits of Wisconsin. Within this type succession are included the original types for the Tazewell and Cary Substages of former usage (Leighton, 1933), the Iowan loess of Illinois (Morton Loess), and the redefined Mankato (Leighton, 1957). An exceptionally large number of radiocarbon dates from the Woodfordian and from the beds immediately above and below it (Fig. 6) serve to establish its age as from about 22,000 radiocarbon years B.P. to about 12,500 radiocarbon years B.P.

The maximum spread of ice during Wisconsinan glaciation came in the Woodfordian Substage (Fig. 4), when ice from the Lake Michigan and Erie lobes advanced as far south as central Illinois (Fig. 1). Locally, cryoturbations formed beyond the limit of the ice (Frye and Willman, 1958). The ice appears to have advanced persistently through an interval of 2,000 to 3,000 years and reached its maximum between 19,000 and 20,000 years ago (Fig. 5).

The two major lobes met in eastern Illinois in a major re-entrant along which the moraines diverge at a sharp angle. The Erie drift is more bouldery and contains more limestone, and its heavy minerals contain more garnet than Lake Michigan drift, whereas the latter is richer in dolomite, black shale, and epidote. Erie drift lacks the pink color found in some of the Lake Michigan drifts.

In Illinois, Woodfordian tills are dominantly pebbly clayey silts or silty clays. Boulders and cobbles are generally scarce, but the youngest drifts are more bouldery. The tills of the Lake Michigan lobe can be separated into major units on the basis of gross differences in lithology. Although there are lateral variations, the Shelbyville-Champaign till is commonly gray and sandy, the Bloomington-Normal till is largely pink and very sandy, the Cropsey-Farm Ridge till is largely yellow-gray and very silty, the Marseilles-Minooka till is largely greenish gray and very clayey with an abundance of small pebbles, the Valparaiso till is largely gray, sandy, and commonly gravelly, and the Tinley-Lake Border till is gray and clayey and contains much black shale. Variations in the drift are reflected in the soils (Wascher *et al.*, 1960).

In Wisconsin, the Woodfordian tills are mostly sandy and bouldery, but these generally represent the middle Woodfordian "Cary" portion. The oldest and youngest tills tend to be finer textured. At any one locality drifts laid down by ice of different ages but traversing the same route are similar.

In Illinois a sequence of more than 30 end moraines (Fig. 7) was deposited during the retreating stage, an interval of less than 8,000 years. To accomplish this, the ice must have been continuously active, and stagnation features are rare. Nearly all the end moraines show evidence of readvance before deposition, although some probably readvanced only a few miles. The changes in realignment of the ice front, overridden moraines, differences in composition of the successive drift sheets, interbedded water-laid deposits, and many exposures showing superposition of several till sheets all suggest that some of the Woodfordian readvances covered as much as 50 to 100 miles (80 to 160 km) (Willman and Payne, 1942). The high rate of ice-front fluctuation required to form the sequence of moraines is consistent with the absence of leaching of carbonates between the till sheets and the absence of breaks within the loess deposited during the interval of moraine building.

The moraines are remarkably uniform in size, many about 50 feet (15 m) high and 1 to 2 miles (1.5 to 3 km) wide, suggesting a roughly rhythmic fluctuation of the ice front. Many of the larger moraines show evidence of a complex structure and have closely spaced, parallel crests that suggest superposition of several episodes of moraine building. Many moraines show thickening and widening to the apex of minor lobes, probably resulting from greater debris transport along the axis of most rapid movement.

Lakes are common in the younger drift but rare in the drift older than about 14,000 years. The older drift contains peat and marl deposits laid down in lakes, but the difference in abundance of lakes is perhaps due to more than age, because the older drift is more gently undulating and generally lacks the sharp hummocky surface that distinguishes the areas with abundant lakes. The change to the very youthful-appearing drift occurs at the Valparaiso-West Chicago front.

When the advancing Woodfordian ice reached the headwaters of the valleys draining to the Mississippi and Illinois Rivers, aggradation and widespread silting of the valleys resulted in initiation of the greatest episode of loess deposition on the continent. Advancing over the initial loess deposits, which are called Morton Loess (Frye and Willman, 1960), the ice reached the big bend in the Ancient Mississippi Valley in north-central Illinois about 21,000 years ago, as shown by a change in mineral composition in the Morton Loess (Glass *et al.*, 1964).

When the ice retreated from its maximum spread, loess continued to be deposited, and the name Richland Loess is applied to the loess overlying the Woodfordian drift. Outside the area covered by the Woodfordian glaciers and their outwash, the Morton and Richland Loesses are not differentiated and are combined in the Peoria Loess, defined from exposures near Peoria, Illinois. Beyond the limit of Woodfordian glaciation, the Peoria Loess mantles earlier Pleistocene deposits and bedrock and, with the single exception of the Sangamon Soil, is the most widely traceable Pleistocene unit in the interior United States. It has been identified from eastern Colorado to Ohio and from Wisconsin to Mississippi.

Outwash from the Woodfordian ice aggraded the major valleys 50 ft (15 m) and locally more. The outwash is largely sand and sandy gravel except near the ice border, where the deposits grade sharply into bouldery gravel. In most valleys these deposits form the most extensive terraces; the older terraces commonly have a cover of loess and sand dunes on unweathered outwash, but the younger have little loess. Along the Mississippi Valley these deposits have been almost entirely swept away by later floods. Remnants are found along the tributary valleys where the up-valley slope of the terrace surfaces attests the rapidity of aggradation of the major valleys.

When the ice built the Valparaiso and West Chicago end moraines, it discharged large volumes of meltwater into the Rock, Fox, DuPage, Des Plaines, and Kankakee Valleys. Drainage from the Lake Michigan, Saginaw, and Erie lobes was concentrated into the Kankakee Valley and formed the Kankakee Flood (Ekblaw and Athy, 1925). At its maximum the flood spread over the uplands between the end moraines, forming Lakes Watseka, Wauponsee, Pontiac, and Ottawa (Fig. 7), all short-lived lakes that produced few shoreline features except where discharges were concentrated through the moraines. Bars of angular boulder rubble show the violence of the discharge in the Kankakee region. Laminated silt and sand deposits locally show periglacial involutions (Sharp, 1942). Passage of the flood down the valley resulted in deep erosion of earlier valley-train deposits, and high-water stages were perhaps 50 ft (15 m) above normal flood stages. Sand, silt, and clay deposited on the bluffs are locally preserved within the Peoria Loess. In southern Illinois the high water is believed to be responsible for diversion of the Mississippi River across the headwaters of a tributary valley into the Ohio Valley and permanent establishment of the river through the Thebes Gorge (Leighton and Willman, 1950).

When ice withdrew from the massive Valparaiso Moraine, Lake Chicago formed between the ice and the moraine (Bretz, 1955, 1959). Earlier stages of a lake in the Michigan basin probably had an outlet to the Des Plaines Valley at Joliet along the buried Hadley Valley, which is more deeply entrenched in bedrock than the present valley (Horberg, 1950). Valparaiso and older drift completely buried Hadley Valley, and, with retreat of the ice from the Valparaiso Moraine, drainage was established at the position of the present Des Plaines Valley, several miles north of the old channel.

Early stages of the post-Valparaiso Lake Chicago were controlled by a buried ridge of Niagaran dolomite that crossed the outlet river at Lemont at an elevation of 640 ft (195 m). By the time the ice advanced to the backslope of the Valparaiso end moraine and deposited the Tinley end moraine, the outlet had been lowered to 620 ft (189 m) or lower, and the 640-ft (195-m) levels attained by the first major stages of Lake Chicago probably result from the large volume and depth of water passing through the outlet. The three major shorelines of Lake Chicago truncate the Tinley drift, but the earliest stages occurred during the stand of the ice at the Lake Border Moraines. The shorelines at levels of 640 ft (195 m) (Glenwood), 620 ft (189 m) (Calumet), and 605 ft (184 m) (Toleston), were all reoccupied several times, and the history of the lake is complex and controversial (Bretz, 1955; Hough, 1958). The low-level lake stage represented by the Two Creeks deposits, which mark the end of Woodfordian time, has been variously related to the Calumet and Glenwood lake stages.

Discharge from the Calumet lake stage required deepening of the outlet to about 600 ft (183 m), and this process required a great volume of water, which apparently came from the discharge of eastern lakes to Lake Chicago through the Grand River across Michigan. The outlet channel was deepened into dolomite for about 25 miles (37 km), and below the area where the bedrock was swept clean the channel was filled with gravel consisting of well-rounded cobbles of dolomite. Cobbles as large as 6 in. (15 cm) are abundant in the gravel near Ottawa, 40 miles (65 km) down the valley from the parent dolomite ledges (Willman and Payne, 1942), and a well-defined low terrace of gravel, grading to pebbly sand, marks the level of the Lake Chicago discharge down the valley until it passes below floodplain level near Beardstown.

In Wisconsin, an early Woodfordian "Tazewell" moraine was deposited beyond the limit reached by ice of middle Woodfordian time in only a few places. The middle Woodfordian "Cary" moraines are multiple yet form a distinct front or end moraine (including the Kettle interlobate moraine) (Alden, 1918) that is traceable with only minor breaks across the state from Illinois to Minnesota. It is the most striking glacial discontinuity in the state (Chamberlin, 1878). Outwash aprons of gravel and sand up to 500 ft (150 m) thick occur along portions of the front (Fig. 2H). Weidman (1907) called attention to the paucity of outwash from the front in central Wisconsin, and it now seems clear that early Woodfordian and Altonian outwash have been overridden for many kilometers in southern Wisconsin (Oakes, 1960). Only a small fraction of the outwash can definitely be attributed to middle and late Woodfordian times.

Following the establishment of the prominent end moraine of middle Woodfordian "Cary" time, ice stagnation occurred throughout Wisconsin. The famous drumlin field of southeastern Wisconsin (Alden, 1918) displays typical kames and associated glaciofluvial deposits, many of which have been incorrectly mapped as recessional end moraines. Such features are particularly well developed and widespread in the lake country of northern Wisconsin. Pitted outwash (Thwaites, 1926) is especially abundant in the Woodfordian drift. Latest Woodfordian end moraines bordering southern Lake Michigan reflect the resurgence of clay- and silt-rich ice from the basin.

Many lakes came into existence in central Wisconsin with the retreat of the Woodfordian ice (Trowbridge, 1917; Bretz, 1950). Glacial Lake Wisconsin (Alden, 1918; Martin, 1932) probably formed at least twice. Early Lake Oshkosh (Thwaites and Bertrand, 1957) appeared in the Green Bay–Fox River lowland. Drainage of Glacial Lake Keweenaw (Hough, 1958) and later Lake Duluth (Leverett, 1929, 1932) and other waters associated with retreat of Woodfordian and Valderan ice from the Lake Superior basin cut the large gorge and giant potholes of the St. Croix Dalles (Chamberlin, 1905). The famous Wisconsin River Dells (Martin, 1932) also were cut at the close of the Woodfordian Substage.

Loess up to 20 ft (6 m) thick in southwestern Wisconsin (Fig. 6) shows no marked irregularities of physical and chemical pedogenic properties that might reflect multiple deposition (Hogan and Beatty, 1963; Glenn *et al.*, 1900). The entire sequence is correlated with the Peoria Loess of Illinois.

The Late Woodfordian and Valderan varved clays of Wisconsin are more uniform in their gross composition than in their heavy minerals, which cannot be used for correlation even within the same lake (Ellsworth, 1932).

TWOCREEKAN SUBSTAGE

The Twocreekan Substage (Frye and Willman, 1960) is based on the Two Creeks Forest Bed and associated silt deposits of eastern Wisconsin, which were discovered by Goldthwait (1907) and most recently described by Thwaites and Bertrand (1957). The oldest tree found had 142 years of growth (Wilson, 1936). The initial phase is characterized by aquatic and semiaquatic mollusks on early Lake Chicago clays (Wilson, 1932). Fossils of woodland mosses, sedges, mollusks, mites, wood-boring beetles, fungi, and trees such as spruce, birch, and jackpine are found in the overlying organic sediments. Pollen analyses imply that the forest was dominated by *Picea* (West, 1961). These are the first records of trees in Wisconsin in deposits younger than latest Altonian (Rockian) (Black, 1962). Later, aquatic mosses and mollusks came in along with sediment from the advancing Valders ice. Although Antevs (1962) believes the substage was about 19,000 years B.P., radiocarbon dates of the wood average 11,850 years (Broecker and Farrand, 1963).

At many places in the Green Bay–Fox River lowland, pieces of Two Creeks wood occur in the overlying Valders till, but organic deposits formed in place are scarce (Fig. 2G). However, weakly developed buried soils interpreted as Twocreekan are relatively common in the same area. In the Duck Creek Ridges southwest of Green Bay, red clayey till above the dated organic sediments cannot be distinguished lithologically or texturally from that below (Piette, 1963).

During Twocreekan time Lake Chicago was at a low level, and discharge of the major rivers was reduced as the ice generally retreated beyond their headwaters.

A zone of caliche nodules locally recognized in the upper part of the Peoria Loess may indicate an interruption in loess deposition and partial solution of carbonates. At other places dark organic streaks may likewise indicate a slowdown in loess deposition. With the migration of the forests into Wisconsin during this substage came ancient man (Black, 1959b; Black and Wittry, 1959) and many animals, including bison, mammoth, and mastodon (Parmalee, 1959; Potzger, 1951).

VALDERAN SUBSTAGE

The Valderan Substage, which draws its name from the Valders Till of eastern Wisconsin (Thwaites, 1943; Thwaites and Bertrand, 1957), is defined (Frye and Willman, 1960) as including that till and subsequent glacial and alluvial deposits that accumulated before the dissipation of the continental glacier and the return of sea level to approximately its present position at about 5,000 years B.P. Presently available radiocarbon dates suggest that the upper 50 ft (15 m) of alluvial fill of the Mississippi Valley in southwestern Illinois was deposited subsequent to 7,000 radiocarbon years B.P. The upper 23 ft (7 m) of fill of the Wisconsin River Valley near Portage was deposited rapidly about 6,000 radiocarbon years B.P.

During Valderan time, ice of the Lake Michigan lobe readvanced to the vicinity of Milwaukee, Wisconsin, but did not reach Illinois. Ice entered the Green Bay lowland (Thwaites, 1943). It may have come out of the Lake Superior lowland onto the highlands of northern Wisconsin (Leverett, 1929), although recent field work by Black suggests that it did not reach the highlands. The location of the ice terminus is mapped with difficulty, as the ice is commonly ended in proglacial lakes, and the lithology of the drift is not diagnostic. Although the drift was defined as red clay till, it is texturally more complicated (Lee *et al.*, 1962). All such drift is not necessarily Valderan, and other kinds of drift may be Valderan. Murray (1953) showed that texturally-unlike Valders and "Cary" (middle Woodfordian) tills contained the same heavy minerals. Other criteria must be used in mapping the distribution of the

Valders till. Prominent moraines, kames, eskers, and other features are now considered characteristic of the Valders drift (Suttner, 1963).

Lake Chicago and associated Later Lake Oshkosh (Thwaites and Bertrand, 1957) re-established the Des Plaines outlet when the water level rose to the Calumet beach (Bretz, 1959). At a lower lake stage, the prominent Toleston beach was constructed. By the end of Valderan time the lake level was at an unusually low stage (Chippewa), about 350 ft (107 m) below the present lake of 580 ft (177 m) (Hough, 1958).

The *Picea* forest that came into Wisconsin during Twocreekan time was overrun by Valderan ice. At the edge of the ice a *Picea* woodland with openings existed. On retreat of the ice, the *Picea* forest reoccupied the region outside the limits of the front, and a *Picea* woodland with openings invaded the areas vacated by the ice (West, 1961).

A widely traceable valley train of Valderan or latest Woodfordian age is represented in the lowest major terrace of Mississippi Valley. The material in this terrace is generally distinguishable from that of older valley trains because the ice front was largely north of the margin of Paleozoic limestone and dolomite formations, so that much of the material is noncalcareous, even though not weathered.

Some loess was probably added to the Peoria Loess during this interval, but no large body of loess comparable to the Bignell of the Missouri Valley appears to have been deposited in the Mississippi Valley. The terraces of this age have little loess and are almost without sand dunes.

Recent Stage

The Recent Stage as used in Illinois and Wisconsin (Frye and Willman, 1960) lacks the basis for definition in stratigraphy that has been used for all previous stages and substages. This time-stratigraphic unit is arbitrarily defined as starting at 5,000 radiocarbon years B.P., since it is judged that at about that point in time sea level had risen to about its present position, and all direct influence of continental glaciers had ceased. Great Lakes drainage, in response to isostatic movements, may have continued through the Des Plaines outlet as recently as about 3,000 years ago (Hough, 1958). Deposits generally classified as Recent include much older deposits and, in general, represent alluvium, dune sands, colluvium, and lacustrine sediments that were deposited after the last recognizable influence of glaciers. The regimen of the Mississippi and Illinois Rivers and the morphology of their floodplains have been described in detail in western Illinois by Rubey (1952).

Undated but presumably post-Valderan pollen sequences from bogs associated with the Wisconsin River (Hansen, 1939) and with the last remnant of Glacial Lake Wisconsin (Hansen, 1937) show typical changes from spruce to pine and finally to hardwood forests with some prairie (West, 1961).

References

Akers, R. H., 1961, Clay minerals of glacial deposits of west-central Wisconsin: Univ. Wisconsin M.S. thesis, 82 p.

Alden, W. C., 1909, Concerning certain criteria for discrimination of the age of glacial drift sheets as modified by topographic situation and drainage relations: J. Geol., v. 17, p. 694-709

—— 1918, Quaternary geology of southeastern Wisconsin: U.S. Geol. Surv. Prof. Pap. 106, 356 p.

American Commission on Stratigraphic Nomenclature, 1961, Code of stratigraphic nomenclature: Amer. Assoc. Petroleum Geologists, v. 45, p. 645-665

Anderson, R. C., 1957, Pebble and sand lithology of the major Wisconsin glacial lobes of the central lowland: Geol. Soc. Amer. Bull., v. 68, p. 1415-1449

Andrews, G. W., 1958, Windrow Formation of the upper Mississippi Valley region—A sedimentary and stratigraphic study: J. Geol., v. 66, p. 597-624

Antevs, Ernst, 1962, Transatlantic climatic agreement versus C^{14} dates: J. Geol., v. 70, p. 194-205

Bain, H. F., 1897, Relations of the Wisconsin and Kansan drift sheets in central Iowa, and related phenomena: Iowa Geol. Surv., v. 6, p. 429-476

Ball, J. R., 1937, The physiography and surficial geology of the Carlinville quadrangle, Illinois: Illinois State Acad. Sci. Trans., v. 30, p. 219-223

Bean, E. F., 1939, Geologic map of Wisconsin: Madison, Univ. of Wisconsin, Wisconsin Geol. Nat. Hist. Surv.

Black, R. F., 1959a, Friends of the Pleistocene: Science, v. 130, p. 172-173

—— 1959b, Geology of Raddatz rockshelter, Sk5, Wisconsin: Wisconsin Arch., v. 40, p. 69-82

—— 1962, Pleistocene chronology of Wisconsin (abst.): Geol. Soc. Amer. Spec. Pap. 68, p. 137

—— 1964, Periglacial phenomena of Wisconsin, north-central United States: Biuletyn Peryglacjalny (in press)

Black, R. F., and Wittry, W. L., 1959, Pleistocene man in south-central Wisconsin (abst.): Geol. Soc. Amer. Bull., v. 70, p. 1570-1571

Bretz, J H., 1950, Glacial Lake Merrimac: Illinois State Acad. Sci. Trans., v. 43, p. 132-136

—— 1955, Geology of the Chicago region. Part II. The Pleistocene: Illinois Geol. Surv. Bull. 65, 132 p.

—— 1959, The double Calumet stage of Lake Chicago: J. Geol., v. 67, p. 675-684

Broecker, W. S., and Farrand, W. R., 1963, Radiocarbon age of the Two Creeks forest bed, Wisconsin: Geol. Soc. Amer. Bull., v. 74, p. 795-802

Chamberlin, R. T., 1905, The glacial features of the St. Croix Dalles region: J. Geol., v. 13, p. 238-256

—— 1910, Older drifts in the St. Croix region: J. Geol., v. 18, p. 542-548

Chamberlin, T. C., 1878, On the extent and significance of the Wisconsin kettle moraine: Wisconsin Acad. Sci. Trans., v. 4, p. 201-234

—— 1880, Le Kettle moraine et les mouvements glaciaires qui lui ont donne naissance: Intern. Geol. Congr. Paris Compt. Rend., p. 254-268

—— 1883a, Geology of Wisconsin: Wisconsin Geol. Nat. Hist. Surv., v. 1, p. 1-300

—— 1883b, Terminal moraine of the second glacial epoch: U.S. Geol. Surv. Ann. Rep. 3, p. 291-402

—— 1890, Some additional evidences bearing on the interval between the glacial epochs: Geol. Soc. Amer. Bull., v. 1, p. 469-480

—— 1894, Glacial phenomena of North America, *in* James Geikie, The great ice age: New York, D. Appleton & Co., 3d ed., p. 724-774

—— 1895, The classification of American glacial deposits: J. Geol., v. 3, p. 270-277

—— 1896, *Editorial:* J. Geol., v. 4, p. 872-876

Ekblaw, G. E., 1960, Map—Glacial geology of northeastern Illinois: Illinois Geol. Surv.

Ekblaw, G. E., and Athy, L. F., 1925, Glacial Kankakee Torrent in northeastern Illinois: Geol. Soc. Amer. Bull., v. 36, p. 417-428

Ekblaw, G. E., and Willman, H. B., 1957, Farmdale drift near Danville, Illinois: Illinois State Acad. Sci. Trans., v. 47, p. 129-138

Ellsworth, E. W., 1932, Varved clays of Wisconsin: Wisconsin Acad. Sci. Trans., v. 27, p. 47-58

Eveland, H. E., 1952, Pleistocene geology of the Danville region: Illinois Geol. Surv. Rep. Inv. 159, 32 p.

Fisk, H. N., 1944, Geological investigation of the alluvial valley of the lower Mississippi River: Mississippi River Comm., War Dept., Corps of Engineers, U.S. Army, 63 p.

Flint, R. F., 1963, Status of the Pleistocene Wisconsin Stage in Central North America: Science, v. 139, p. 402-404

Flint, R. F., Colton, R. B., Golthwait, R. P., and Willman, H. B., 1959, Glacial map of the United States east of the Rocky Mountains: Geol. Soc. Amer.

Frye, J. C., 1963, Problems of interpreting the bedrock surface of Illinois: Illinois State Acad. Sci. Trans., v. 56, p. 3-11

Frye, J. C., Glass, H. D., Leonard, A. B., and Willman, H.B., 1963, Late Pleistocene loesses of Midwestern United States of America: Biuletyn Peryglacjalny, Nr. 12, p. 111-118

Frye, J. C., Glass, H. D., and Willman, H. B., 1962, Stratigraphy and mineralogy of the Wisconsinan loesses of Illinois: Illinois Geol. Surv. Circ. 334, 55 p.

Frye, J. C., and Leonard, A. B., 1951, Stratigraphy of the late Pleistocene loesses of Kansas: J. Geol., v. 59, p. 287-305

—— 1952, Pleistocene geology of Kansas: Kansas Geol. Surv. Bull. 99, 230 p.

Frye, J. C., Shaffer, P. R., Willman, H. B., and Ekblaw, G. E., 1960, Accretion-gley and the gumbotil dilemma: Amer. J. Sci., v. 258, p. 185-190

Frye, J. C., and Willman, H. B., 1958, Permafrost features near the Wisconsin glacial margin in Illinois: Amer. J. Sci., v. 256, p. 518-524

—— 1960, Classification of the Wisconsinan Stage in the Lake Michigan glacial lobe: Illinois Geol. Surv. Circ. 285, 16 p.

—— 1963a, Development of Wisconsinan classification in Illinois related to radiocarbon chronology: Geol. Soc. Amer. Bull., v. 74, p. 501-506

—— 1963b, Loess stratigraphy, Wisconsinan classification and accretion-gleys in central-western Illinois: Illinois Geol. Surv. Guidebook series no. 5, 37 p.

Frye, J. C., Willman, H. B., and Glass, H. D., 1960, Gumbotil, accretion-gley, and the weathering profile: Illinois Geol. Surv. Circ. 295, 39 p.

—— 1964, Cretaceous deposits and the Illinoian glacial boundary in western Illinois: Illinois Geol. Surv. Circ. 364, p. 28

Glass, H. D., Frye, J. C., and Willman, H. B., 1964, Record of Mississippi River diversion in the Morton Loess of Illinois: Illinois State Acad. Sci. Trans., v. 57, p. 24-27

Glenn, R. C., Jackson, M. L., Hole, F. D., and Lee, G. B., 1960, Chemical weathering of layer silicate clays in loess-derived Tama silt loam of southwestern Wisconsin: Eighth Natl. Conf. on Clays and Clay Minerals, Intern. Earth Sci. Ser. Monogr. 9, New York, Pergamon Press, p. 68-83

Goldthwait, J. W., 1907, The abandoned shore lines of eastern Wisconsin: Wisconsin Geol. Surv. Bull. 17, 134 p.

Hansen, H. P., 1937, Pollen analysis of two Wisconsin bogs of different age: Ecology, v. 18, p. 136-148

—— 1939, Postglacial vegetation of the Driftless Area of Wisconsin: Amer. Midl. Nat., v. 21, p. 752-762

Hershey, O. H., 1893, The Pleistocene rock gorges of northwestern Illinois: Amer. Geol., v. 12, p. 314-323

Hogan, J. D., and Beatty, M. T., 1963, Age and properties of Peorian loess and buried paleosols in southwestern Wisconsin: Soil Sci. Soc. Amer. Proc., v. 27, p. 345-350

Hole, F. D., 1943, Correlation of the glacial border drift of north central Wisconsin: Amer. J. Sci., v. 241, p. 498-516

—— 1950, Areas having aeolian silt and sand deposits in Wisconsin: Map. Soils Div., Wisconsin Geol. Nat. Hist. Surv.

Horberg, Leland, 1950, Bedrock topography of Illinois: Illinois Geol. Surv. Bull. 73, 111 p.

—— 1953, Pleistocene deposits below the Wisconsin drift in northeastern Illinois: Illinois Geol. Surv. Rep. Inv. 165, 61 p.

—— 1956, Pleistocene deposits along the Mississippi Valley in central-western Illinois: Illinois Geol. Surv. Rep. Inv. 192, 39 p.

Hough, J. L., 1958, Geology of the Great Lakes: Urbana, Univ. of Illinois Press, 313 p.

Hussakof, Louis, 1916, Discovery of the great lake trout, *Cristivomer namaycush*, in the Pleistocene of Wisconsin: J. Geol., v. 24, p. 685-689

Johnson, W. H., 1964, Stratigraphy and petrography of Illinoian and Kansan drift in central Illinois: Illinois Geol. Surv. Circ. 378

Kempton, J. P., 1963, Subsurface stratigraphy of the Pleistocene deposits of central-northern Illinois: Illinois Geol. Surv. Circ. 356, 43 p.

Kempton, J. P., and Hackett, J. E., 1964, Radiocarbon dates from the pre-Woodfordian Wisconsinan of northern Illinois (abst.): Geol. Soc. Amer. Spec. Pap. 76, Abstracts for 1963, p. 91

Lee, G. B., Janke, W. E., and Beaver, A. J., 1962, Particle-size analysis of Valders drift in eastern Wisconsin: Science, v. 138, p. 154-155

Leighton, M. M., 1917, The Iowan glaciation and the so-called Iowan loess deposits: Iowa Acad. Sci. Proc., v. 24, p. 87-92

—— 1926, A notable type Pleistocene section: the Farm

Creek exposure near Peoria, Illinois: J. Geol., v. 34, p. 167-174

—— 1933, The naming of the subdivisions of the Wisconsin glacial age: Science, v. 77, p. 168

—— 1957, The Cary-Mankato-Valders problem: J. Geol., v. 65, p. 108-111

—— 1959, Stagnancy of the Illinoian glacial lobe east of the Illinois and Mississippi Rivers: J. Geol., v. 67, p. 337-344

—— 1960, The classification of the Wisconsin glacial stage of North Central United States: J. Geol., v. 68, p. 529-552

Leighton, M. M., and Brophy, J. A., 1961, Illinoian glaciation in Illinois: J. Geol., v. 69, p. 1-31

Leighton, M. M., and MacClintock, Paul, 1930, Weathered zones of the drift sheets of Illinois: J. Geol., v. 38, p. 28-53

—— 1962, The weathered mantle of glacial tills beneath original surface in north-central United States: J. Geol., v. 70, p. 267-293

Leighton, M. M., and Powers, W. E., 1934, Evaluation of boundaries in the mapping of glaciated areas: J. Geol., v. 42, p. 77-87

Leighton, M. M., and Willman, H. B., 1950, Loess formations of the Mississippi Valley: J. Geol., v. 58, p. 599-623

Leonard, A. B., and Frye, J. C., 1960, Wisconsinan molluscan faunas of the Illinois Valley region: Illinois Geol. Surv. Circ. 304, 32 p.

Leverett, Frank, 1889, On the occurrence of the "forest bed" beneath intramorainic drift: Amer. Assoc. Adv. Sci. Proc. v. 37, p. 183-184

—— 1898a, The weathered zone (Yarmouth) between the Illinoian and Kansan till sheets: J. Geol., v. 6, 238-243

—— 1898b, The weathered zone (Sangamon), between the Iowan loess and the Illinoian till sheet: J. Geol., v. 6, p. 171-181

—— 1899, The Illinois glacial lobe: U.S. Geol. Surv. Monogr. 38, 817 p.

—— 1909, Weathering and erosion as time measures: Amer. J. Sci., v. 27, p. 349-368

—— 1929, Moraines and shore lines of the Lake Superior region: U.S. Geol. Surv. Prof. Pap. 154, 72 p.

—— 1932, Quaternary geology of Minnesota and parts of adjacent states: U.S. Geol. Surv. Prof. Pap. 161, 149 p.

MacClintock, Paul, 1922, The Pleistocene history of the lower Wisconsin River: J. Geol., v. 30, p. 673-689

—— 1929, Recent discoveries of pre-Illinoian drift in southern Illinois: Illinois State Geol. Surv. Rep. Inv. 19, p. 26-57

Martin, Lawrence, 1932, The physical geography of Winconsin: Wisconsin Geol. Surv. Bull. 36, 609 p.

Murray, R. C., 1953, The petrology of the Cary and Valders tills of northeastern Wisconsin: Amer. J. Sci., v. 251, p. 140-155

Oakes, E. L., 1960, The Woodfordian moraines of Rock County, Wisconsin: Univ. Wisconsin M.S. thesis, 61 p.

Parmalee, P. W., 1959, Animal remains from the Raddatz rockshelter, Sk5, Wisconsin: Wisconsin Arch., v. 40, p. 83-90

Piette, C. R., 1963, Geology of Duck Creek Ridges, east-central Wisconsin: Univ. Wisconsin M.S. thesis, 86 p.

Potzger, J. E., 1951, The fossil record near the glacial border: Ohio J. Sci., v. 51, p. 126-133

Rubey, W. W., 1952, Geology and mineral resources of the Hardin and Brussels Quadrangles (in Illinois): U.S. Geol. Surv. Prof. Pap. 218, 179 p.

Salisbury, R. D., 1893, Distinct glacial epochs and the criteria for their recognition: J. Geol., v. 1, p. 61-84

Schoewe, W. H., 1920, The origin and history of extinct Lake Calvin: Iowa Geol. Surv., v. 29, p. 49-22

Shaffer, P. R., 1954, Extension of Tazewell glacial substage of western Illinois into eastern Iowa: Geol. Soc. Amer. Bull. 65, p. 443-456

—— 1956, Farmdale drift into northwestern Illinois: Illinois Geol. Surv. Rep. Inv. 198, 25 p.

Sharp, R. P., 1942, Periglacial involutions in northeastern Illinois: J. Geol., v. 50, p. 113-133

Shimek, Bohumil, 1909, Aftonian sands and gravel in western Iowa: Geol. Soc. Amer. Bull., v. 20, p. 399-408

Smith, G. D., 1942, Illinois loess—variations in its properties and distribution: A pedologic interpretation: Univ. Illinois Agr. Exp. Stat. Bull. 490, p. 139-184.

Suttner, L. J., 1963, Geology of Brillion Ridge, east-central Wisconsin: Univ. Wisconsin M.S. thesis, 99 p.

Thwaites, F. T., 1926, The origin and significance of pitted outwash: J. Geol., v. 34, p. 308-319

—— 1928a, The development of the theory of multiple glaciation in North America: Wisconsin Acad. Sci. Trans., v. 23, p. 41-164

—— 1928b, Pre-Wisconsin terraces of the Driftless Area of Wisconsin: Geol. Soc. Amer. Bull., v. 39, p. 621-641

—— 1943, Pleistocene of part of northeastern Wisconsin: Geol. Soc. Amer. Bull., v. 54, p. 87-144

—— 1956, Wisconsin glacial deposits—Map: Wisconsin Geol. Nat. Hist. Surv.

—— 1960, Evidences of dissected erosion surfaces in the Driftless Area: Wisconsin Acad. Sci. Trans., v. 49, p. 17-49

Thwaites, F. T., and Bertrand, Kenneth, 1957, Pleistocene geology of the Door Peninsula, Wisconsin: Geol. Soc. Amer. Bull., v. 68, p. 831-879

Trowbridge, A. C., 1917, The history of Devil's Lake, Wisconsin: J. Geol., v. 25, p. 344-372

—— 1921, The erosional history of the Driftless Area: Univ. Iowa Stud. Nat. Hist., v. 9, 127 p.

Wanless, H. R., 1957, Geology and mineral resources of the Beardstown, Glasford, Havana, and Vermont Quadrangles: Illinois Geol. Surv. Bull. 82, 233 p.

Wascher, H. L., Alexander, J. D., Ray, B. W., Beavers, A. H., and Odell, R. T., 1960, Characteristics of soils associated with glacial tills in northeastern Illinois: Univ. Illinois Agr. Exp. Stat. Bull. 665, 155 p.

Wascher, H. L., Humbert, R. P., and Cady, J. G., 1948, Loess in the southern Mississippi Valley—Identification and distribution of the loess sheets: Soil Sci. Soc. Amer. Proc. 1947, v. 12, p. 389-399

Weidman, Samuel, 1907, The geology of north-central Wisconsin: Wisconsin Geol. Surv. Bull. 16, 697 p.

—— 1913, The Pleistocene succession in Wisconsin: Science, v. 37, p. 456-457

Weller, J. M., *et al.*, 1945, Geologic map of Illinois: Illinois Geol. Surv.

West, R. G., 1961, Late and postglacial vegetational history in Wisconsin, particularly changes associated with the Valders readvance: Amer. J. Sci., v. 259, p. 766-783

Willman, H. B., and Frye, J. C., 1958, Problems of Pleistocene geology in the greater St. Louis area: *in* Geol. Soc. Amer. Guidebook, St. Louis Meetings, p. 9-19

Willman, H. B., Glass, H. D., and Frye, J. C., 1963, Mineralogy of glacial tills and their weathering profiles in Illinois. Part I, Glacial tills: Illinois Geol. Surv. Circ. 347, 55 p.

Willman, H. B., and Payne, J. N., 1942, Geology and mineral resources of the Marseilles, Ottawa, and Streator Quadrangles: Illinois Geol. Surv. Bull. 66, 388 p.

Willman, H. B., Swann, D. H., and Frye, J. C., 1958, Stratigraphic policy of the Illinois State Geological Survey: Illinois State Geol. Surv. Circ. 249, 14 p.

Wilson, L. R., 1932, The Two Creeks forest bed, Manitowoc County, Wisconsin: Wisconsin Acad. Sci. Trans., v. 27, p. 31-46

—— 1936, Further fossil studies of the Two Creeks forest bed, Manitowoc County, Wisconsin: Torrey Bot. Club Bull., v. 63, p. 317-325

Worthen, A. H., 1866, Physical features, general principles, and surface geology, *in* Geology of Illinois: Geol. Surv. Illinois, v. 1, p. 1-39

—— 1873, Geology of Sangamon County: Illinois Geol. Surv., v. 5, p. 306-319

Zumberge, J. H., 1960, Correlation of Wisconsin drifts in Illinois, Indiana, Michigan, and Ohio: Geol. Soc. Amer. Bull. 71, p. 1177-1188

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SUMMARY

Deposits of each of the four glacial stages occur in Illinois and Wisconsin, as do soils formed during each of the three major interglacial stages. Wisconsin and northeastern Illinois is the type region for the Wisconsinan Stage, and central Illinois is the type region for the Illinoian Stage. The Sangamonian Stage is named for buried soil in the Sangamon River basin in Sangamon County, Illinois, and the substages used here for the Wisconsinan and Illinoian are based on type sequences in these two states.

The geographic position of Illinois in the area of overlap of tills deposited by glaciers from both eastern and western centers of dispersal makes it the only region in North America where the physical relations of the pre-Illinoian tills from these two principal centers can be examined. It also is a focal point of drainage diversions produced by the several glacial advances. These drainage diversions resulted in changes in mineral composition of outwash and loess as the tills from different sources have recognizably different compositions.

The stratigraphic succession from Nebraskan to Recent as it occurs in these two states is briefly outlined and methods of correlation are described. Recent work on the mineralogy of Pleistocene deposits, the subdivision and classification of the Illinoian and Wisconsinan Stages, and the evidence for glaciation of the "Driftless Area" is summarized.

PLEISTOCENE GEOLOGY OF INDIANA AND MICHIGAN *

WILLIAM J. WAYNE,[1] JAMES H. ZUMBERGE[2]

THE FIRST comprehensive studies of the Pleistocene geology of Indiana and Michigan were those of Leverett (1899, 1902; Leverett and Taylor, 1915), although earlier workers had mapped the drift in many areas (Hubbard, 1840; Thompson, 1886, 1889; Dryer, 1894). Subsequent statewide treatment has been provided in Indiana by Malott (1922), Thornbury (1958), and Wayne (1956b, 1963). In addition, considerably more detailed knowledge has come from the work of Gooding (1957, 1963), Harrison (1959, 1960, 1963), Kapp and Gooding (1964b), Thornbury (1937), Wayne (1958a, 1959), and Zumberge (1960). Surficial geologic maps have been prepared for Michigan by Martin (1955, 1957) and for both Indiana and Michigan by Flint *et al.* (1959).

About 80% of Indiana was covered with ice during one or more Pleistocene glaciations (Leverett and Taylor, 1915, Pl. 6; Wayne, 1958b; Flint *et al.*, 1959). In the two-thirds of the state that lies north of the Wisconsin glacial boundary, most of the landscape was formed by glacial construction (Fig. 1). Between the Wisconsin boundary and the maximum extent of the earlier glaciations, glacially formed features are interspersed with topography that was clearly developed on the underlying bedrock. South of the limit reached by glacial ice, valleys partly filled with outwash and lacustrine sediments dominate the topography.

The physiography of southern Indiana is characterized by southward-trending areas of alternating uplands and lowlands (Fig. 1; Malott, 1922; Wayne, 1956b). The broad Wabash Lowland on the west side of the state, underlain by Pennsylvanian shales and coals, and the Scottsburg Lowland in the southeastern part of Indiana, underlain by Devonian and lower Mississippian shales, provided avenues between adjacent uplands along which the lobes of the earlier glaciers extended nearly to the 38th parallel. Resistant Mississippian and lower Pennsylvanian rocks formed an upland barrier between these two lobes that was too high for the ice to cover (Fig. 1).

Michigan consists of two tracts of land called the Northern and Southern Peninsulas. The surface is dominated by a youthful topography developed on glacial drift of Wisconsin age. The surface drainage lines are poorly developed, and inland lakes form an important part of the landscape throughout most of Michigan and northern Indiana (Scott, 1921).

Michigan and Indiana are bordered by more than 2,300 miles of shoreline of the Great Lakes of Superior, Michigan, Huron, and Erie. The shorelines vary in character from long stretches of sandy beaches along Lake Michigan to the steep sandstone cliffs of Lake Superior. The coasts are attacked by wave erosion during periods of high water levels, such as occurred in 1952 when considerable damage to property and loss of shore installations were sustained. During low-water phases, such as in 1934 and 1964, considerable dredging in harbors and connecting channels is required (St. Clair River, Detroit River).

The levels of the Great Lakes follow a pattern of highs in the summer and lows in the winter. Seasonal fluctuations are normally about 30 cm on Lake Superior, 30 to 45 cm on Lakes Michigan and Huron, and 45 to 60 cm on Lakes Erie and Ontario. Table 1 gives some information on the morphometry of the lakes.

The bedrock geology of Michigan consists of two distinct provinces. The Michigan Basin, containing Cambrian to Pennsylvanian shales, sandstones, carbonate rocks, and evaporites, occupies all the Southern Peninsula and the eastern half of the Northern Peninsula. The Precambrian shield, which occupies the western half of the Northern Peninsula, contains chiefly Proterozoic metamorphosed lava flows, Huronian iron formations, Killarney granite intrusives, and Keweenawan lava flows, conglomerates, and sandstones.

The glacial deposits of Indiana, Michigan, and adjacent states and the history of their deposition have been the subject of several schemes of classification by the various workers who have studied them during the past 75 years. The early classifications of Leverett and Malott and of Leighton (1933) were reviewed and compared by Thornbury (1937) and by Wayne (1963). Thornbury (1937) introduced into Indiana the substage terminology that Leighton (1933) had recently developed for the classification of the Wisconsin Stage in Illinois. This classification has undergone many changes in recent years, but the use of stages and substages remains the primary basis of time-stratigraphic classification of the Pleistocene Series.

Correlation problems inherent in this time-rock classification system, when it is used alone, have been a major factor in encouraging the development of a rock-stratigraphic classification for glacial sediments in Indiana (Wayne, 1963). Both sets of terminology have been used in this summary (Fig. 2).

All surficial deposits in Michigan are products of glaciation during the Wisconsin Age. Their classification is based on morphologic features of various drift sheets rather than on stratigraphic type sections. The Wisconsin drift sheets

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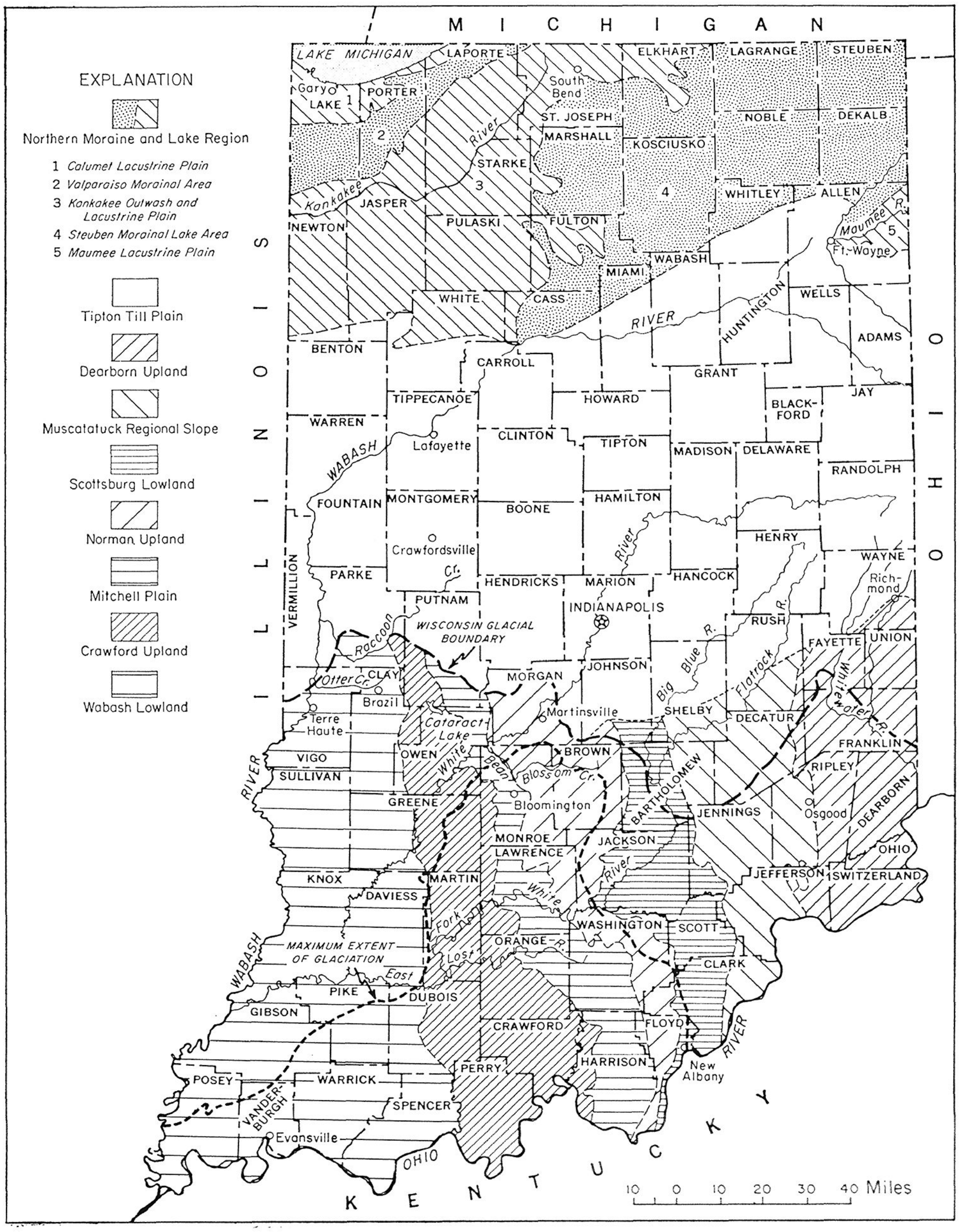

Figure 1. Map of Indiana showing physiographic units and glacial boundaries.

and their associated moraines in Michigan have been assigned to the Cary, Port Huron (Mankato), and Valders Substages. This system is built on the original classification of Leverett and Taylor (1915), Leverett (1929b), and Leighton (1933) and on important modifications by Bretz (1951a), Melhorn (1954), Zumberge *et al.* (1956), Hough (1958, p. 94), Leighton (1960), and Zumberge (1960).

In addition to the three Wisconsin drift sheets and their associated end moraines, the late-Pleistocene chronology of Michigan and parts of northern Indiana is built around the phases of proglacial lakes that began to develop during the retreat of the Cary ice from the basins of Lake Michigan, Lake Huron, and Lake Erie.

Late-Tertiary Geomorphic History and Pre-Pleistocene Topography

Except for deposition of alluvial and colluvial deposits, nearly all of Indiana and Michigan probably was eroded continuously from the time that late Paleozoic sedimentation ceased until the area was covered by Pleistocene glaciers. Direct evidence of Mesozoic physiographic history is lost, but the Ohio River Formation in southern Indiana probably represents a depositional record from the early part of the Tertiary Period (Wayne, 1960, p. 29). Evidently, thick residual soils existed at the time that unit was deposited; the basal part of the Ohio River Formation is in many places a thick cherty clay, now compacted and altered by burial, but otherwise similar to the surficial soils on the adjacent limestone plain.

In the other parts of Indiana and Michigan, subaerial erosion and weathering continued throughout Tertiary and early Quaternary time. Reddish-brown cherty clay soils 12 m or more thick are preserved in broad areas of the Mitchell Plain (Fig. 1), where development of karst has reduced the amount of water available to erode the surface.

Scattered deposits of siliceous gravels at high levels, called the Lafayette Gravel (Malott, 1922, p. 132), are the residue of old stream deposits and are considered to be middle to

TABLE 1

Morphometry of the Great Lakes
(From U.S. Lake Survey)

	Superior	Michigan	Huron	Erie	Ontario	Total[a]
Area (sq. miles)						
Drainage basin[b]	80,000	67,860	72,620	32,490	34,800	295,200
Water surface	31,820	22,400	23,010	9,930	7,520	95,170
Shoreline (miles)						
United States	1,427	1,661	769	490	331	5,506
Canada	1,549	0	2,416	366	395	5,471
Total	2,976	1,661	3,185	856	726	10,977
Depth (ft)						
Average	487	276	195	58	264	
Maximum	1,333	923	750	210	778	
Mean surface elevation						
(ft above sea level)	602	580	580	573	246	

[a] Includes Lake St. Clair and its drainage basin.
[b] Including lake area.

late Tertiary in age. They probably represent deposits of more than one episode of deposition and degradation during and shortly following a late-Tertiary erosion cycle, during which an erosion surface known as the Lexington Peneplain was formed. Uplift of this surface is now generally considered to have begun during the Miocene or Pliocene Epoch (Thornbury, 1958, p. 454). Remnants of this peneplain are preserved on the dissected uplands of the unglaciated part of Indiana (Malott, 1922) and on the buried uplands (Wayne, 1956b). The buried bedrock physiography of Indiana was developed on flat-lying limestones, shales, and sandstones and is the northward continuation of the physiographic units recognized in the unglaciated part of the state. Bedrock physiography of Michigan is still not well known, but studies to date (Mozola, 1962) indicate that similar features underlie the thick drift cover of that state.

Most of Indiana was drained during Tertiary time by the Teays River and its tributaries. The Teays (Tight, 1903) headed in the Appalachian Mountains and flowed northwestward through West Virginia, Ohio, Indiana, and Illinois, where it joined the ancestral Mississippi (Horberg, 1950; Wayne, 1952). Before the earliest glaciers reached Indiana, the Teays had eroded a gorge as much as 75 m below the adjacent limestone plains in north-central Indiana. Major tributary streams had cut their valleys to accordance. Glacial erosion was slight, so that details of the buried topographic features remain essentially those of erosion by running water. Differences of interpretation place the time of deep valley erosion of the Teays as preglacial (Horberg, 1950; Wayne, 1956b) or interglacial (Stout *et al.*, 1943; Stout, 1953).

The drainage history of the Teays Valley has been inferred from studies of the sediments that partly fill its abandoned strath in southern Ohio and West Virginia, from the physiographic details of that region, and from bedrock topographic studies in western Ohio, Indiana, and Illinois. Most students of the Teays have considered it to have been deranged by an early glacier (Stout and Shaaf, 1931; Stout *et al.*, 1943). Conclusions reached from a recent study of the sediments in West Virginia (Rhodehamel and Carlston, 1963, p. 271) do not agree with this interpretation. These authors believed that the Teays was diverted before glacial ice reached it as a result of nonglacial processes of stream capture. Although their data and arguments seem reasonable, they are not altogether convincing, and it is difficult to see how their hypothesis could be applied to the creation of the Ohio River. Each abandoned segment of a major valley would have to be regarded as a separate situation of normal stream capture. Glacial diver-

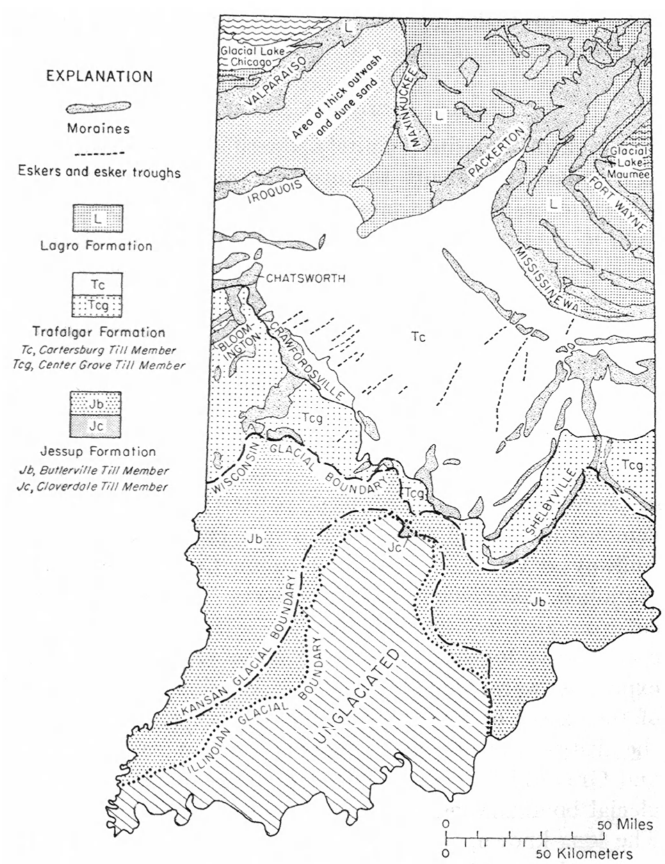

Figure 2. Map of Indiana showing extent of glacial sediments and moraines.

sion during the Nebraskan and Kansan Ages probably was the cause of derangement of the Teays and the creation of the Ohio River (Wayne, 1956b, p. 46).

Recognizable former floodplain sediments can be found on straths that stand only a few meters above the modern drainage levels outside the limits of glaciation. These sediments, included by Wayne (1963, p. 38) in the Prospect Formation, are particularly noticeable along streams that have been diverted to subsurface routes, such as Lost River. Age of the Prospect Formation may be as great as late Pliocene in some drainage basins and as young as early to middle Pleistocene in others.

Nebraskan Stage

Deposits referable to the Nebraskan glaciers have not been recognized in either Indiana or Michigan. Thwaites (1946, Pl. 3) suggested that the scattered erratics of northern Kentucky (Leverett, 1929a, p. 34) might be Nebraskan in age, and Eveland (1952) interpreted an exposure at Danville, Illinois, as gumbotil of Nebraskan and Aftonian age embedded in till of Kansan age. Reinterpretation of the Danville exposures (Ekblaw and Willman, 1955, p. 134; Thornbury, 1958, p. 455) suggests that no deposits referable to Nebraskan glaciation exist there.

The most likely places for a record of Nebraskan glaciation to have remained in northern Indiana and southern Michigan are over the bedrock lowlands and in the Teays Valley and its tributaries. Unfortunately, glacial deposits there are very thick, and few logs of wells are sufficiently detailed to make possible the recognition of pre-Kansan sediments.

Aftonian Stage

The lack of deposits definitely identifiable as Nebraskan in age makes it difficult to interpret any sediments in the two states as Aftonian in age. Some of the Prospect Formation in Indiana may be Aftonian in age, however, and buried alluvial materials beneath lacustrine sediments of Kansan age undoubtedly are Aftonian. There is little positive evidence about the nature of such buried deposits, because few of them have been found.

Kansan Stage

The presence in Indiana of deposits of Kansan age was suspected by Thornbury (1937, p. 100), but none had been demonstrated until 1954 when a new cut for an emergency spillway at Cataract Lake in Putnam County (Fig. 1) was examined. Discovery of well-preserved Kansan till in the cut provided an impetus for the careful examination of other areas for exposures of drift of Kansan age. Many exposures were found (Wayne, 1958a, Fig. 1), and a limit of the extent of Kansan ice into Indiana was inferred from the distribution of the exposures. Recent mapping (Wier and Gray, 1961) indicates that in a few places the Kansan glacial boundary may extend beyond the Illinoian ice limit. The long-known, scattered, large erratics in northern Kentucky (Leverett, 1929a) are more likely Kansan in age (Wayne, 1956b, p. 48; Thornbury, 1958, p. 456) than Nebraskan, as was suggested by Thwaites (1946). Ray (1963) reported Kansan till in Kentucky and probable Kansan loess along the Ohio River east of Evansville, Indiana. Kansan drift has not been recognized in Michigan.

Till of Kansan age, which makes up the bulk of the Cloverdale Till Member of the Jessup Formation of Wayne (1963, p. 52), is generally pebbly, sandy, and silty and is very compact. Its lithologic features, the distinctive paleosol that caps it, and its position in an exposure generally make identification of Cloverdale till reasonably certain.

One exposure of fossiliferous proglacial loess of Kansan age has been found in Indiana; it is at the Cataract Lake spillway. This silt contains wood fragments and 20 species of snails (Wayne, 1958a, p. 15; 1963, Table 2) that suggest an ice-marginal climate and a forest cover similar to that in central Ontario today.

The Kansan glaciation of central Indiana consisted of at least two pulsations of the ice separated by a brief ice-free time (Fig. 2). A thin bed of fossiliferous silt has been found intercalated between two tills of the Cloverdale member in western Indiana (Wayne 1963, Table 2). Three exposures of this bed are known, all in the same area; its fauna of 16 snail species gives the only record so far of the land mollusks that lived around the Kansan ice margin during deglaciation.

Valley-fill sediments deposited in Indiana during the Kansan glaciation were never wholly removed during the Yarmouth Age and later erosional episodes. Cloverdale till was recognized in an auger boring west of Brazil, Indiana, where, at a depth of 12 m below the floodplain of Otter Creek, a paleosol on compact till was found. Where no identifying paleosols occur, till units are difficult to distinguish in subsurface studies. Gravel of Kansan age fills the valley of Raccoon Creek, where it can be recognized beneath a thick accumulation of younger drift. The lowest gravels, sands, and silts that fill major pre-Kansan valleys, such as the lower Wabash, White, and Ohio, probably are also of Kansan age. These glaciofluvial sediments, along with similar younger sediments, make up the Atherton Formation (Wayne, 1963, p. 31).

Yarmouth Stage

After the Kansan glacier had melted from Indiana, it left over most of the state a relatively thin cover of drift that probably did not greatly obscure the major features of the pre-Kansan bedrock topography. Stream erosion soon cleaned out the main valleys to a new base level, this time not so deep as the bottom of the partly filled bedrock valleys.

Floodplain sediments of Yarmouth age have been observed in several of the valleys of central Indiana. Along Raccoon Creek in Parke County, exposures of peat and old floodplain deposits can be traced for several miles along the edge of the valley and indicate that the Yarmouth floodplain was slightly lower than the modern floodplain of the same stream. The same is true of the lower part of Bean Blossom Creek, a tributary of White River (Wayne, 1958a, p. 13); farther upstream the top of the Yarmouth floodplain sediment is 3 to 4 m higher than the modern floodplain. West of Brazil in Otter Creek, Yarmouth sediments lie 12 m below the modern floodplain.

The soil profile developed during Yarmouth time on till of Kansan age is well displayed in exposures in western Indiana and has been observed in the eastern part of the state. This soil and associated interglacial sediments mark the top of the Cloverdale Till Member of the Jessup Formation. In the type section of the Cloverdale till, carbonates have been removed to a depth of 3.7 m. Bhattacharya (1962, p. 1016) found kaolinite in each of the Yarmouth paleosols he examined but not in Sangamon paleosols and Recent soils. He related this fact to a longer period of weathering.

Even though many exposures of Yarmouth sediments have been discovered, fossils other than wood fragments are scarce. Englehardt (1962) studied the pollen in the upper parts of three buried Yarmouth alluvial deposits. Two of these, near Osgood in Ripley County and on Bean Blossom Creek in Monroe County, probably were Illinoian proglacial accumulations on a Yarmouth surface; the pollen in both, though sparse, was dominated by *Picea* and *Pinus* (Englehardt, 1962, p. 92, 103). The third, a buried alluvial sediment that contains layers of compressed peat, shows a strong dominance of *Pinus* and *Picea* but also contains a small percentage of *Quercus* pollen. All mollusks recovered from Yarmouth sediments suggest a cool climate (Wayne, 1963, p. 56). Inasmuch as most floodplain sediments probably are reworked every few thousand years, absence of any but very late Yarmouth deposits is not surprising.

Illinoian Stage

Only a few exposures of pre-Wisconsin drift are known in Michigan. These were assigned to the Illinoian Stage by Leverett because of their occurrence immediately below drifts of Wisconsin age. The position of the Illinoian glacial boundary, the southernmost of the three in Indiana (see Fig. 3), is well known from the mapping by several workers (Leverett, 1899, 1902; Leverett and Taylor, 1915; Malott, 1922; Thornbury, 1937).

At least three advances of Illinoian ice reached into southwestern Indiana (Wayne, 1963, p. 53) and left tills that are separated by thin beds of fossiliferous silt. Two such tills and fossiliferous silt were reported in the Whitewater basin of southeastern Indiana by Gooding (1963, p. 669), who named the ice advances and retreats they represent the Centerville Stade, the Abington Interstade, and the Richmond Stade, respectively, from oldest to youngest (Fig. 2).

The Butlerville Till Member of the Jessup Formation of Wayne (1963, p. 14) includes all the tills of Illinoian age. The unit lies between the top of the paleosol on the Cloverdale and the top of the paleosol on the Butlerville. Deposition of the till was accompanied by outwash and loess deposition, but the loess did not escape subsequent erosion in many places. Outwash gravels, lacustrine sands, silts, and clays, and loessal silts of Illinoian age are included as part of the Atherton Formation.

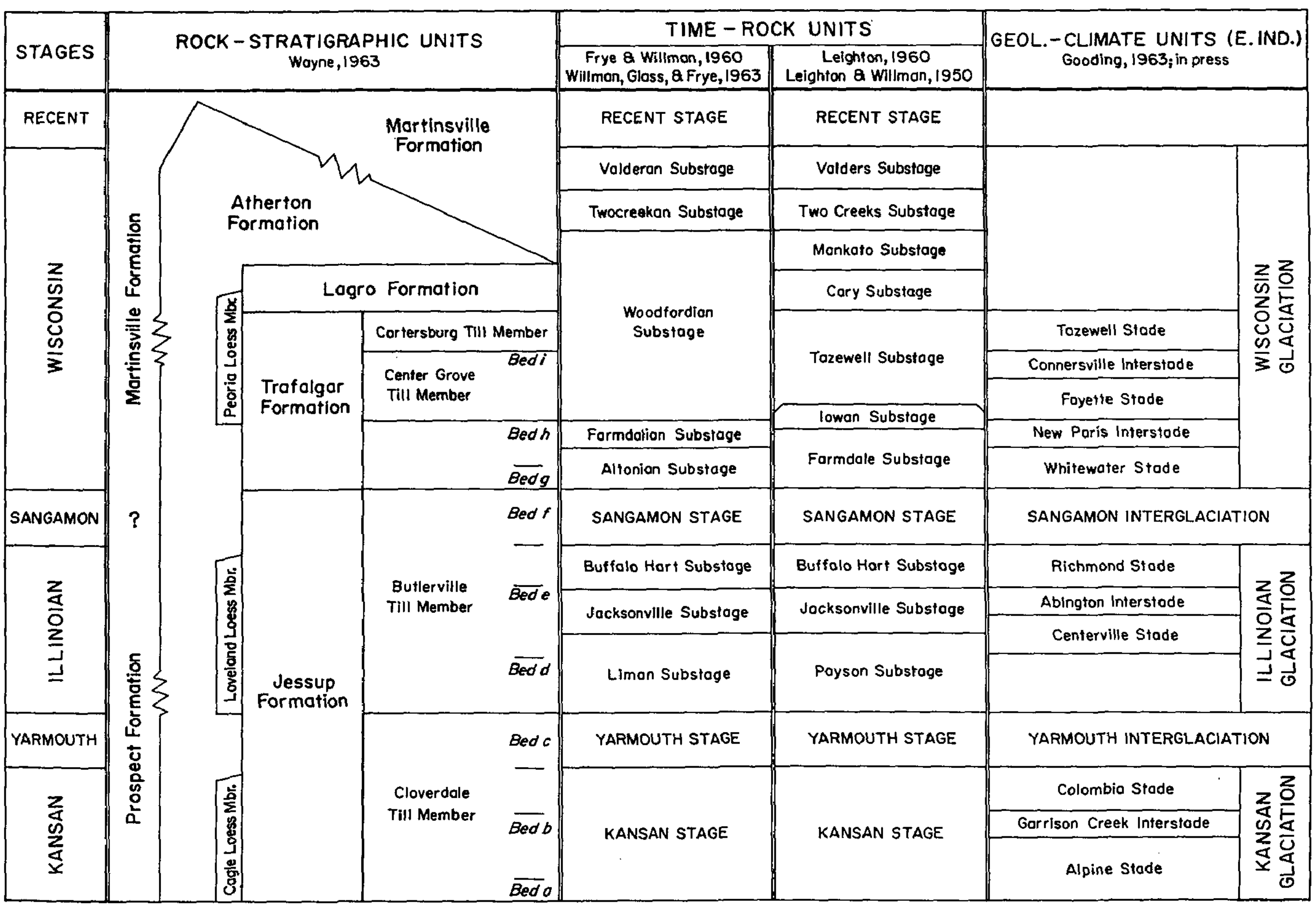

Figure 3. Correlation chart of Pleistocene stratigraphic terms and geologic-climate units used in Indiana.

Only two exposures of the lowest till of the Butlerville are known, both in Parke County. The till has a distinctive grayish-brown color not found in other tills of the Butlerville member. The overlying fossiliferous silt contains only species of very small snails.

The second advance of Illinoian ice evidently extended farther south in western Indiana than any other glacier; its deposits probably correlate with the Liman (Frye *et al.*, 1964) and Jacksonville Substages of Illinois (Fig. 2). The ice blocked nearly all westward-flowing streams, creating extensive ice-marginal lakes (Thornbury, 1950, p. 4-8). Sedimentation in many of the lakes was extensive, and some of the broader lacustrine plains, such as those of glacial Lakes Flatwoods, Quincy, and Patoka, remain essentially undissected.

The middle till of the Butlerville member is separated from a younger till in west-central Indiana by a thin fossiliferous silt bed. Exposures of the silt bed have been found in Vigo and Parke Counties (Wayne, 1963, p. 53), and a bed reported by Harrison (1963, Fig. 6) in Marion County may be the same unit. The bed has not been traced across the state, but on the basis of position and similarity of lithology and fauna it might be referable to the Abington Interstade of eastern Indiana described by Gooding (1963, p. 672). The snail fauna from this bed is dominated by small, cool-climate species, particularly such forms as *Catinella gelida* and *Columella alticola.* The till that overlies it is almost coextensive with the Wisconsin glacial boundary in west-central Indiana and may correlate with the drift of the Buffalo Hart Moraine of Illinois.

Moraines are rare on the Illinoian till plain in Indiana. A few ridges of till as much as 10 m high are associated with the Illinoian drift border in Daviess and Pike Counties, but generally the margin is not marked by a recognizable moraine. Other segments of ridged drift exist but have not been traced far enough to allow correlation. Only one, Chestnut Ridge in Jackson County, has been named (Leverett, 1902, p. 255).

Leverett (1917, p. 114) asserted that some of the more massive morainic ridges in southeastern Michigan contain cores of Illinoian till with a veneer of younger Wisconsin drift on top. An older till, presumably Illinoian in age, crops out in the wave-cut cliffs along the coast of Lake Huron between Port Huron and the head of Saginaw Bay (the Thumb area). Bergquist and MacLachlan (1951, p. 15) suggested that the blue-gray tills described by Leverett in the Thumb are unoxidized parts of a till sheet of Wisconsin age. This explanation is not likely, however, if peats and soils of undoubted interglacial origin occur between the lower indurated till and the upper Wisconsin till, as Leverett (1917, p. 113) claimed. Leverett and Taylor (1915, p. 72) suggested that an indurated blue-gray till occurring beneath a Wisconsin drift in the Huron valley near Ann Arbor is Illinoian in age. Leverett (1917, p. 114) believed that most of the Southern Peninsula was covered by the Illinoian ice, as indeed it must have been if Illinois and Indiana were invaded by ice from the northeast (Leverett, 1899; Leighton and Brophy, 1961), but D. F. Eschman (personal communication) is of the opinion that no Illinoian till crops out in eastern Michigan.

Leighton (1959) interpreted the lack of moraines and the abundance of kames in Illinois as indicating that the Illinoian glaciers became stagnant ice masses. Only a few kames have been noted in the Illinoian drift in Indiana, but several extensive outwash plains have been mapped. This evidence suggests that the main advance of Illinoian ice probably did not disappear from Indiana by stagnation.

Broad outwash plains in Indiana were laid down by melt-water from the Illinoian glacier, particularly in Morgan, Owen, Greene, and Daviess Counties. Valley-train outwash has been largely destroyed by later erosion and obscured by sedimentation, but remnants of it can be found along those valleys that never carried melt-water from the Wisconsin ice sheet as well as along most of the major valleys. These coarse sediments and the finer-grained deposits associated with them are included in the Atherton Formation of Wayne (1963, p. 31).

Loess was blown from the outwash deposits along the Wabash and Ohio Valleys and deposited as a thin blanket over the adjacent uplands. The volume of the loess (Loveland Loess Member) seems not to have been great, and it has been observed below the Butlerville Till Member only near the edges of valleys that served as major sluiceways during the Illinoian Age.

Lakes were ponded not only in the valleys blocked by the Illinoian ice sheet but also in valleys tributary to the major sluiceways. Most of these lacustrine sediments of Illinoian age are now buried beneath younger materials (Fidlar, 1948, p. 18).

Sangamon Stage

The record of deposition during the Sangamon Age is much better preserved and exposed for study than is that of the earlier Yarmouth Age. Many exposures of buried Sangamon soils and deposits were recorded by Leverett and Taylor (1915) and by Thornbury (1937, p. 100-123). Thinness of overlying Wisconsin deposits in central Indiana has permitted post-Wisconsin streams to erode valleys in many places below the level that was the surface during Sangamon time. Even so, the record of the entire interglacial age is far from complete.

Most of the interglacial record consists of soils that developed on the glacially derived sediments of Illinoian age and the colluvial, alluvial, and paludal sediments that accumulated in depressions and on floodplains on the post-Illinoian surface. By far the greatest number of exposures display materials that had accumulated just prior to the onslaught of Wisconsin ice.

Probably the most complete record of Sangamon vegetation in Indiana is in a buried floodplain accumulation near Richmond. The pollen studies of Kapp and Gooding (1964b, p. 312) and Englehardt (1962, p. 126) show a progression of vegetation from coniferous to deciduous forest that took place at the close of the Illinoian glaciation and a return to coniferous forest dominance that heralded the advance of Wisconsin ice.

One other fossiliferous deposit dates from the early part of the Sangamon; near Martinsville, Indiana, a small creek has trenched a marl-filled depression in an outwash plain of Illinoian age. The upper part of the accumulation has been destroyed by weathering, but the lower part contains a record of the cool-climate, aquatic snails and clams that lived in a small early post-Illinoian lake.

The main part of the Sangamon Age is poorly represented except by the products of weathering. The soil profile on the Butlerville member, where it is not buried beneath a thick layer of younger sediments, is 3 to 4 m thick, but the upper part (half a meter or more) is a veneer of loessial silt. Where the Sangamon soil has been protected from further weathering by overlying Wisconsin drift, the thickness of till that has been leached of carbonates is generally less than 2 m (Thornbury, 1937, 1940), but gravels, sands, and silts are noncalcareous to depths of 3 m or more. The soil profile seems to have been developed under conditions at least as warm as those of the present, perhaps slightly warmer. Fossil records of the warm part of Sangamon time are scarce, but the pollen records from deposits near Richmond cited earlier (Kapp and Gooding, 1964b, p. 312) represent the full vegetational record of the Sangamon and include a period warmer and drier than the present. In an exposure of Sangamon alluvium in Johnson County, Indiana, Englehardt (1962, p. 120) found a decrease of *Quercus* and a corresponding increase in conifer pollen from base to top of the section, which he interpreted as indicating the approach of Wisconsin ice.

Most of the exposures of sediments that lie directly beneath Wisconsin deposits in Indiana contain almost wholly conifer wood, pollen, and mollusks characteristic of cool climates; they have yielded radiocarbon dates of 21,000 to 23,000 years B.P. (before the present), well within the time span of the Wisconsin Age (Wayne, 1963, p. 80).

The shallow depth of oxidation of till in the Sangamon weathering profile contrasted with the much greater depth of oxidation during post-Wisconsin time suggests that ground-water levels may have been higher than at present. Local base levels of erosion may have differed, too, although the top of a late Sangamon floodplain sediment exposed along the Ohio River at Owensboro, Kentucky (L. L. Ray, 1955, oral communication), and dated as 23,150 years B.P. (W-270) is almost identical in elevation to the modern floodplain surface. Sangamon floodplain sediments buried beneath stratified silts in tributary valleys of the Whitewater and Wabash also are similar in elevation to the modern floodplains in those valleys.

Buried peats and soils in southeastern Michigan are cited by Leverett and Taylor (1915, p. 72) and Leverett (1917, p. 113) as evidence of an interglacial stage of pre-Wisconsin and post-Illinoian age. Because none of these occurrences has been studied in detail in recent times, it is impossible to evaluate Leverett's interpretation of their age. This is a problem in Michigan that still awaits elucidation.

Wisconsin Stage

At its greatest extent, Wisconsin ice spread over all of Michigan and about two-thirds of Indiana (Figs. 2, 5). The southern limit of Wisconsin drift was mapped by Leverett and Taylor (1915, Pl. 4) and later in greater detail by Thornbury (1937, p. 40). More detailed mapping has provided further refinements during the past decade (Wayne, 1956b, Pl. 1; Gooding, 1957, Pl. 1; Flint *et al.*, 1959; Wier and Gray, 1961).

Since about 1950 the stratigraphy of the Wisconsin Stage has been extensively re-examined. A more critical study of available streambank exposures and many new roadcuts has provided the basis for major alterations in some existing concepts. Dreimanis' (1958, 1960) work in southern Ontario laid much of the groundwork for recognizing that sediments of Wisconsin age considerably older than those of the recognized morainal sequences in Illinois might in fact exist in North America.

Classifications of the Wisconsin Stage have been reviewed recently by Leighton (1958a, 1958b, 1959, 1960), by Frye and Willman (1960), and by Wright (1964). These and recent studies in Indiana by Wayne (1963) and Gooding (1963) are summarized in Figure 2.

Tazewell Substage and Older Wisconsin Deposits

In most parts of central Indiana where streams have eroded through the deposits of Wisconsin age, both physical stratigraphy and radiocarbon dates indicate little material of glacial origin that is greater than 20,000 to 23,000 years old. In the Whitewater basin, however, Gooding (1961, p. 102; 1963, p. 674) found sediments that he interpreted as indicating an extension of glacial ice into southeastern Indiana during the Wisconsin Age prior to deposition of the bulk of Wisconsin drift in the state.

Gooding (1963, p. 674) gave names to the several advances and retreats of the Wisconsin glacier represented by the sediments he had examined in the Whitewater valley (Fig. 2). The Whitewater Stade is named for a glacial advance that left a basal till containing inclusions of reddish-brown till that directly overlies Sangamon sediments and soil. Deposits representing the Whitewater Stade are overlain by fossiliferous silt left during a brief glacial retreat, which Gooding named the New Paris Interstade. Possibly these two intervals correspond in time with the Altonian and Farmdalian Substages of Frye and Willman (1960, p. 4), although a discrepancy seems to exist in the radiocarbon dates available, and the degree of weathering seems inconsistent with this interpretation. Frye and Willman placed the Altonian Subage between 50,000 and 28,000 years B.P.; all dates on the Whitewater Stade indicate an age greater than 40,000 years B.P. In spite of the long time evidently involved between the time of the Whitewater Stade and the next younger glacial advance, which Gooding called the Fayette Stade, the sediments from which the Whitewater Stade is inferred are reported to show little effect of weathering.

Gooding's Fayette Stade probably is the glacial advance that left the Center Grove Till Member of the Trafalgar Formation of Wayne (1963, p. 45). The Center Grove, like the till representing the Fayette Stade, can be observed in many exposures directly overlying a Sangamon soil profile. Both are separated from the next younger till by a fossiliferous silt bed about 30 cm thick and dated at about 20,000 years B.P. Wayne referred to the fossiliferous unit as the *Vertigo alpestris oughtoni* bed; Gooding named the deglaciation it represents the Connersville Interstade.

This silt bed has been observed in many exposures in a

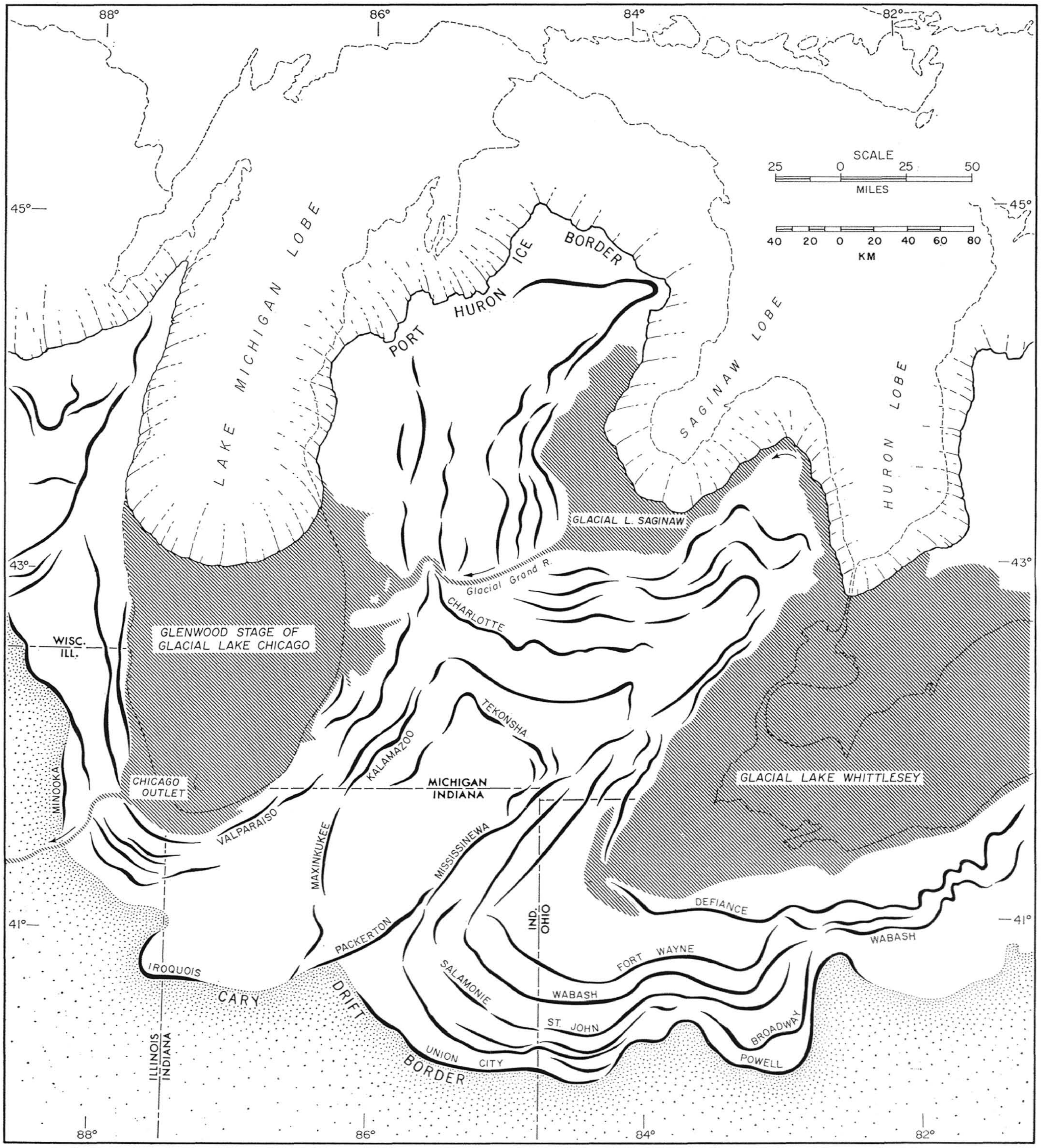

Figure 4. Paleogeographic map of the upper Great Lakes region during the Port Huron (Mankato) maximum *ca.* 13,000 B.P. Heavy black lines show trends of Cary recessional moraines. The glacial Grand River carried the overflow of glacial Lakes Whittlesey and Saginaw to Lake Chicago, where the Allendale delta was constructed. Morainic trends based on Flint *et al.* (1959); Cary drift border after Zumberge (1960). (Outline of modern Great Lakes is shown by a dashed line.)

band about 30 km wide between Crawfordsville and Richmond. It represents a brief but distinct ice-free phase during the Wisconsin Age when a subarctic fauna and flora began to repopulate the newly deglaciated land surface. Extensive outwash plains in Fountain, Montgomery, Marion, Shelby, and Wayne Counties and buried outwash gravels elsewhere in central Indiana were deposited during the next glacial advance and probably were the sources of loessial silts that preserved the fossil material. Till overlying the fossil bed was named the Cartersburg Till Member of the Trafalgar Formation by Wayne (1963, p. 48), and the glacial advance was referred to the Tazewell Stade by Gooding (1963, p. 678). Retreat of the ice margin probably did not exceed 100 to 200 km between deposition of the Center Grove and the Cartersburg members.

Recent mapping in west-central Indiana indicates that the outer limit of the Cartersburg till overlaps the deposits related to the Bloomington Moraine of the Lake Michigan Lobe and is in turn overlapped by a younger drift, perhaps that associated with the Chatsworth Moraine (Fig. 3). If this interpretation is valid, the Center Grove till would be the central-Indiana correlative of the drift sheets in Illinois related to the Shelbyville, Champaign, and Bloomington Moraines, and the Cartersburg would correlate with the Chatsworth and younger moraines. The segmented moraine along the edge of the Cartersburg till has been named the Crawfordsville Moraine (Wayne, in preparation).

Thickness of the ice sheet that deposited the Cartersburg member in central Indiana was determined by Harrison (1958) by means of preconsolidation studies of the underlying silt bed. He found that the ice sheet thickened abruptly from the margin and was 510 m thick 44 km from the edge.

Harrison's studies (1959, p. 19; 1963, p. 31) of the petrographic characteristics of the tills in Marion County, Indiana, showed that all the units he included in the Wisconsin drift, some of which conceivably could be Illinoian in age, were so nearly identical that he could not separate them statistically on petrographically defined parameters. Reconstitution of the major rock fractions in the till permitted Harrison (1960, Fig. 3) to determine the probable path taken by the ice from southeastern Canada through the Lake Erie basin to central Indiana.

During the Wisconsin Age melt-water poured down the valleys of the Wabash, both forks of White River, the Whitewater, the Ohio, and many lesser sluiceways. Gravel and sand accumulated in the sluiceways, slack-water lakes were created in all their tributaries, and loess was blown onto the adjacent uplands (Thornbury, 1950). These sediments are environmental facies of a single depositional unit, the Atherton Formation (Wayne, 1963).

The Peoria Loess Member of the Atherton Formation is fossiliferous in many localities along the Wabash and Ohio Rivers, and the vertical sequence of fossil snails can be used to determine climatic changes as the ice approached. At the ice margin the loess splits into two tongues, one of which underlies and the other overlies the Center Grove till. Fossils from the two tongues are distinctive (Wayne, 1963, Table 2) and have been studied extensively from exposures in Hendricks and Johnson Counties.

The surface of the Cartersburg till in central Indiana indicates deglaciation by stagnation. Few moraines are present, but eskers, kames, and esker troughs are abundant (Fig. 3). These features in central Indiana, together with the pattern produced by streams and by very shallow, nearly parallel flutings on the till plain just south of the Wabash River in west-central Indiana (Schneider *et al.*, 1963, p. 172), outline well the direction of ice movement in all parts of the White River Sublobe (Horberg and Anderson, 1956, p. 105).

CARY SUBSTAGE

The glaciation of Michigan and northern Indiana during the Cary Subage was accomplished by the nearly simultaneous advance of the Lake Michigan Lobe, Saginaw Lobe, and Erie Lobe. The lobate pattern of ice retreat is inferred from the festooned morainic pattern (Fig. 4; Martin, 1955; Flint *et al.*, 1959). Interpretation of the history of deposition of the complex intermingling of tills, outwash sediments, and lake deposits in the northern part of Indiana and in southern Michigan posed problems for Leverett, which have not been worked out adequately to this day. Typical tills from the three lobes can be distinguished without difficulty, but in the interlobate areas where they intertongue with each other, their identification becomes increasingly difficult. Because of this, Wayne (1963, p. 43) grouped all of them together into the Lagro Formation but suggested means to separate the tills into members wherever they could be distinguished. Tills of the Lake Michigan and Erie Lobes typically are very clayey and silty; till of the Saginaw Lobe is sandy and bouldery by contrast.

Two nearly parallel morainic ridges that lie in western Indiana have provided a basis for two fundamentally opposing interpretations regarding the maximum advance of ice during the Cary Subage. Leverett (1899, p. 258, Pl. 6) first considered the Nebo-Gilboa ridge in northern Benton County as part of the Iroquois Moraine (Figs. 3, 4), which nearly parallels it in Newton and Jasper Counties. Later he (Leverett and Taylor, 1915, p. 124) altered this interpretation and treated the two as separate moraines. Many of the field relationships observable in Indiana, including the sharp south margin of the Iroquois Moraine and the pattern of outwash deposition in it, seem to support the interpretation that ice from the north deposited both moraines, but some evidence (Zumberge, 1960, p. 1180), indicates that ice from the east may have extended in an elongated tongue and deposited these ridges and the intervening till. Zumberge (1960) started with the Minooka Moraine in Illinois, which was deposited by the Lake Michigan Lobe, and which was defined by Leighton (1960, p. 547) as the outermost moraine of Cary age. Zumberge's resultant Cary maximum advance of the three lobes is shown in Figure 4. According to his interpretation the Iroquois and Packerton Moraines of the Saginaw Lobe and the Union City Moraine of the Erie Lobe are equivalent in age and define the maximum Cary advance of the three lobes. This also is the interpretation currently accepted in Illinois. (See Horberg, 1955, Fig. 1; Frye *et al.*, this volume).

Recent mapping along the west end of the Packerton Moraine (Wayne *et al.*, in preparation) and along the Maxinkuckee Moraine (Schneider and Keller, in preparation) indicates that these moraines are not exactly as they have long been mapped. Rather than a long protruding tip at the Wabash River, the drift sheet probably had a much more rounded margin. Intensely pitted outwash plains and kames form the margin of the Maxinkuckee toward the north; later trenching of the outwash has removed much of the direct physiographic evidence that would allow a better correlation of deposits of the Saginaw Lobe on the east with the Lake Michigan Lobe on the west. That the two were nearly contemporaneous is certain.

During the retreatal phases of the Cary Subage, the Saginaw Lobe dissipated more rapidly than either the Lake Michigan or Erie Lobes. The Erie Lobe had extended into Indiana from the east during the Cary Subage, depositing the clay-rich New Holland Till Member of the Lagro Formation (Wayne, 1963, p. 44), while some ice, perhaps stagnant masses only, from the Saginaw Lobe still covered the area northwest of it. Outwash sand and gravel from the Erie Lobe occur in the area previously occupied by the Saginaw Lobe in northeastern Indiana and south-central Michigan. Additional support for this interpretation is to be found 45 km southwest of Ann Arbor, Michigan, and in Steuben and DeKalb Counties, Indiana, where till of the Erie Lobe overlies till of the Saginaw Lobe (Zumberge, 1960, p. 1185; Wayne and Thornbury, 1955, p. 9). The west edge of the clay-rich till can be traced fairly readily, but it becomes less recognizable to the northeast where kames and pitted outwash plains form the interlobate zone. Channels carried melt-water from the ice through the Mississinewa Moraine of the Erie Lobe and westward across the jumbled mass of moraines, till plains, and outwash plains left by the Saginaw Lobe.

One of the major features of the post-Cary deglaciation in Michigan was the coalescence of several melt-water channels in south-central Michigan and northeastern Indiana to produce the large discharge of the so-called Kankakee Torrent (Ekblaw and Athy, 1925), which drained southwestward across the northwest corner of Indiana to Illinois. The heads of these channels start at the Tekonsha Moraine of the Lake Michigan and Saginaw Lobes and the Mississinewa Moraine of the Erie Lobe (Fig. 4) and thus provide a basis for correlating these moraines (Zumberge, 1960, p. 1181).

Much of the correlation problem exists because, during the withdrawal of ice from northern Indiana and southern Michigan, a large amount of outwash was deposited in the Kankakee valley. Wind whipped the sand from this outwash and from the aprons of outwash south of the Valparaiso Moraine and west of the Maxinkuckee Moraine into dunes that have obscured physiographic relationships of glacial landforms. The ubiquitous sand blanket has reduced the likelihood that exposures of stratigraphic significance will be found.

During the retreat of the Cary ice, proglacial lakes formed along the south margins of all three lobes. The history of these lakes is based on identifiable changes in level recorded in abandoned shorelines that exist peripheral to the modern Great Lakes and in the character of sediments as determined from cores. A résumé of the history of these ancestral lakes is in a later section of this paper.

PORT HURON (MANKATO) SUBSTAGE

The Cary ice retreated an unknown distance to the north, possibly north of the Straits of Mackinac and then readvanced to the position of the Port Huron Moraine (Fig. 4). This feature was described by Taylor (Leverett and Taylor 1915, p. 293) as ". . . one of the best developed and most clearly defined moraines in the Great Lakes region." The *Glacial Map of the United States East of the Rocky Mountains* (Flint *et al.*, 1959) shows the Port Huron Moraine as the outer limit of a significant glacial advance, a status that has been accorded to it by every glacial geologist who has worked in Michigan since the time of Leverett. The Port Huron Moraine was dated by Hough (1958, p. 278) at 13,000 years ago.

Bretz (1951a), Melhorn (1954, p. 31), and Zumberge *et al.* (1956) believed that the Port Huron advance was an event of late Cary time, but Leighton (1957, p. 109) argued that if the Port Huron Moraine recorded an ice advance of substage rank, it could not naturally be part of another substage. Zumberge (1960) yielded to this viewpoint but retained the name Port Huron for this substage rather than the term Mankato, which Leighton (1960, p. 547) continues to use.

The Port Huron till is generally gray blue and in places is quite sandy, characteristics that distinguish it from the red Valders till, which overlies parts of the Port Huron Moraine in Otsego and Montmorency Counties, Michigan.

TWO CREEKS SUBSTAGE

The retreat of the Port Huron ice was of sufficient magnitude to allow the draining of a proglacial lake that occupied the Lake Michigan basin. These newly exposed lacustrine deposits eventually supported a forest, which was then drowned and overridden by the readvance of the Valders ice. The buried forest near the village of Two Creeks, Wisconsin, was first described by Goldthwait (1907, p. 61) and later by Wilson (1932), but its significance was placed in fuller perspective by Thwaites (1943) and others, notably Bretz (1951a).

Recent redating of the Two Creeks forest bed has increased its age from 11,400 (Thwaites and Bertrand, 1957) to 11,850 years B.P. (Broecker and Farrand, 1963, p. 796). A buried organic layer on the eastern shore of Lake Michigan near South Haven was correlated by Zumberge and Potzger (1956, p. 277) with the Two Creeks bed in Wisconsin, but its date (10,860 years B.P., W-167) may be too young. If so, then the Two Creeks stratigraphic horizon in Michigan is represented only by a disconformity between Valders and Port Huron tills in the northern part of the Southern Peninsula of Michigan (Melhorn, 1954).

VALDERS SUBSTAGE

The ice that readvanced over the forest at Two Creeks, Wisconsin, was first recognized in Wisconsin as a separate glacial advance of substage rank by Thwaites (1943). Melhorn (1954) showed that the red Valders till was restricted to the proximal slope of the Port Huron Moraine in the northern part of the Southern Peninsula of Michigan, and Bretz (1951a) found it overlapping the Port Huron Moraine along the Lake Michigan coast between Muskegon and Grand Traverse Bay.

The most distinguishing feature of the Valders drift in both Wisconsin and Michigan is its red or pink color, which contrasts strongly with the gray-blue (unoxidized) or brown (oxidized) till of the Port Huron Substage. The color intensity of the Valders till increases with the percentage of clay in the till matrix.

No end moraine was built by the Valders ice, but a swarm of drumlins on Valders ground moraine near the southwest end of Grand Traverse Bay of Lake Michigan trends northwestward—an indication of a radial flow from the central Lake Superior region. The Port Huron ice advance, in contrast, came from the northeast, a fact that accounts for the different lithology of the two drifts.

A reddish till forming the surface drift of the Northern Peninsula of Michigan was mapped by Bergquist (1933). This drift is undoubtedly of Valders age and indicates that the Valders ice invasion was the last glacial event in the Pleistocene history of Michigan. Broecker and Farrand (1963, p. 800) concluded that all of Michigan was free of Valders ice 10,000 years ago.

Recent Stage

Deposition of nonglacial sediments began as soon as the ice had melted from the area. Except for their fossil content, the earliest postglacial sediments in Indiana and Michigan are indistinguishable from those now being laid down in streams and lakes. Nevertheless, for mapping purposes these sediments generally are regarded as Recent in age even though the lower parts of basin-fill deposits were laid down before the Wisconsin Age came to a close.

Sediments of the Recent Stage include alluvial deposits along streams, colluvial accumulations along the bases of steep slopes, swamp and lake sediments, such as peat and marl, and sand-dune accumulations around the edge of Lake Michigan. Alluvial, paludal, and colluvial sediments in Indiana were treated by Wayne (1963, p. 28) as environmental facies of the Martinsville Formation; the modern dunes are mapped as dune-sand facies of the Atherton Formation.

Floodplain sediments, which make up the greatest volume and area of Recent deposits, have been under continuous deposition, erosion, and redeposition since retreat of glacial ice from the individual valleys. Radiocarbon dates available (W-59, W-254, W-666, W-832, LJ-290) of wood at the base of the Martinsville indicate that few of these alluvial sediments probably exceed 7,000 to 8,000 years in age. Downstream migration of meanders in most streams of any size is a process that brings about regular reworking of older floodplain deposits; it seems likely that few floodplain deposits escape being eroded in 5,000 to 7,000 years. Although the most common kind of fossil in Indiana alluvial sediments is woody debris, mollusk shells have been found, and some vertebrate remains probably exist in the sediments, even though they are not often reported.

Paludal and lacustrine sediments, composed dominantly of fine-grained materials and deposited in quiet water, are primarily peat and marl. Thickness of sediments in some of the filled basins in central Indiana exceeds 12 m; similar thicknesses exist in the lakes and bogs of northern Indiana and throughout Michigan, but most of the deposits are much thinner.

Paleobotanists have long searched for evidence of postglacial tundra in Indiana (Dillon, 1956, p. 174), but so far data from pollen profiles of bogs have shown a forest cover of spruce when the oldest sediments accumulated (Potzger and Wilson, 1941; Prettyman, 1937; Otto, 1938; Swickard, 1941; Guennel, 1950). In recent studies of Indiana bogs (Frey, 1959; Englehardt, 1960; Kapp and Gooding, 1964a) a Two Creeks equivalent has been interpreted, and although the cores penetrated inorganic sediments below the peat, no zone was found that indicated a postglacial tundra vegetation in Indiana.

Pollen and mollusks from intertill silt beds indicate that a narrow zone around the ice margin had tundra-like vegetation (Dillon, 1956, p. 174), but it probably never existed as a climax cover in the state (Wayne, 1956a, p. 164; Martin, 1958, p. 384). Features ascribable to intensive frost action likewise are scarce, although polygonal features have been noted near some ice marginal positions (Wayne, 1964, p. 179) in west-central Indiana.

Many filled and partly filled depressions in central and northern Indiana and Michigan contain marl, a soft fine-grained lime mud. Cores through some of the marl beds show that few are more than 5 to 7 m thick. The basal marl beds in many of the deposits are very fine-grained and medium to dark gray; the molluscan fauna consists dominantly of species of *Gyraulus, Valvata, Physa,* small lymnaeids, and *Pisidium.* The upper parts of the deposits commonly are light gray to yellowish white and contain a much larger molluscan fauna, particularly species of *Helisoma, Amnicola, Physa, Campeloma,* and a few of the large lymnaeids, as well as *Valvata* and *Gyraulus.* No correlation has yet been made on these depositional and faunal changes, but they may reflect post-Wisconsin climatic changes. The marl lakes are in ice-block depressions, and the lower beds may have accumulated while an ice block still existed below the basin. Nearly all the remains of extinct fossil vertebrates that have been found in Indiana and Michigan have come from the lower parts of the paludal sediments.

Geologic History of the Great Lakes

PHYSIOGRAPHY OF THE LAKE BASINS

The area of the five Great Lakes and their drainage basins is 445,000 km^2, of which 144,000 km^2 is accounted for by the lakes themselves. Lake Superior is the largest and deepest, Lake Ontario is the smallest, and Lake Erie is the shallowest. Important facts about the morphometry of the lakes are given in Table 1. General details of the bathymetry of the lakes can be found in Hough (1958), which is the latest and most detailed work in a single volume; large-scale charts of all segments of the lakes, their harbors, and connecting waterways are available from the Lake Survey Division of the U.S. Corps of Engineers, Detroit, Michigan.

The basins of the Great Lakes all lie in bedrock, and their shorelines are more or less parallel to the regional strike of the bedrock formations surrounding them. Paleozoic sediments surround most of the four lower lakes, but Precambrian rocks bound the north shore of Georgian Bay of Lake Huron and all of Lake Superior except the south shore between Marquette, Michigan, and Sault Sainte Marie, Canada. The shorelines of the lakes, except for Lake Superior, which is mostly rockbound, are developed mainly in glacial drift.

ORIGIN OF THE BASINS OF THE GREAT LAKES

The geologic processes that can produce large depressions on the earth's crust are few. Yet geomorphologists and geologists who have studied the Great Lakes do not agree completely on the origin of their basins. Paucity of information on the geology of the lake basins has been one of the main reasons for disagreement. A reasonably accurate hydrographic chart of the eastern basin of Lake Superior was not available until early in this decade (Zumberge and Gast, 1961), and almost nothing is known of the configuration of the bedrock surface beneath the lake bottoms. Thwaites (1949), Webb and Smith (1961), and Cuancara and Melik (1961) attempted a correlation of bedrock features across the basins of Lakes Michigan and Huron by using hydrographic charts, and Thwaites (1935) produced a tectonic map of the Lake Superior bottom, but these efforts will remain highly speculative until direct supporting evidence is made available.

During the last 75 years, geologists have proposed four processes to account for the origin of the basins of the Great Lakes: (1) stream erosion during the Tertiary Period, (2) glacial erosion during the Pleistocene Epoch, (3) damming by glacial drift, and (4) subsidence or warping of the earth's crust.

Newberry (1874, 1882) argued that the basins of the Great Lakes were originally stream valleys but that glacial erosion was of prime importance in the final shaping of the basins. Spencer (1891) believed that glacial modification of the ancient stream valleys ". . . could have been only slight, and does not appear to have been more than the sweeping of loose geological dust into the valleys, or on to the highlands to the south" (Spencer, 1896, p. 163). Leverett and Taylor (1915, p. 319) asserted that all the basins except that of Lake Superior had distinct characteristics of stream-eroded valleys and that all the changes produced by successive glacial invasions did not destroy these characteristics to any great extent. Martin (1916, p. 407) claimed that the Lake Superior basin ". . . owes a notable part of its exhumation, and all, or nearly all, of its present depth below lake level to erosion by the Superior lobe . . ." Shepard (1937) and Thwaites (1949) joined Martin in support of the glacial-erosion hypothesis. Schwartz (1949, p. 82) suggested faulting as a factor in the origin of the Lake Superior basin but dismissed it because of lack of supporting evidence and concluded that glacial erosion was the most important agent in the making of the basin.

Horberg and Anderson (1956, p. 103) believed that the preglacial topography of the Great Lakes region was extensively modified by glacial scour in the uplands and ". . . by profound deepening of preglacial lowlands along the axes of the Great Lakes." Hough (1958, p. 113) believed that both preglacial stream erosion and glacial scour were important in the origins of the lake basins, and he dismissed diastrophic action as a primary agent in their origin. Laidly (1961, p. 282) suggested that narrow north-south submerged troughs in eastern Lake Superior were produced by glacial erosion concentrated along pre-existing fracture zones.

The most definite statement regarding the origin of these lake basins is that they are glacially modified river valleys. The glacial modification may have been in the form of glacial erosion or glacial deposition. Zumberge and Gast (1961) and Zumberge (1962a) showed that at least 670 ft of glacial till lies below the bottom of western Lake Superior near the Minnesota shore in 938 ft of water, and Hough (1955) reported glacial till at a depth of 35 ft below the bottom of Lake Michigan in its deepest part (923 ft). In addition to glacial deposits in the lake basins, tens of feet of lacustrine materials have accumulated in various parts since the withdrawal of glacier ice.

The most important information necessary to formulate an hypothesis on the origin of the Great Lakes basins—details of the configuration of the bedrock floors of the lakes—is lacking. Hydrographic charts do not reveal the character or shape of the bedrock floors and thus are insufficient. Modern geophysical methods, such as seismic profiling (Hersey, 1963) and shipboard drilling techniques (Zumberge, 1962b), are expensive, but until more information of this kind is forthcoming, substantial glacial erosion in any of the basins of the Great Lakes cannot be proved beyond a reasonable doubt.

EVIDENCE AND CAUSES OF FORMER LAKE STAGES

Whatever the origin of the basins of the Great Lakes, there is abundant evidence that in the past they all contained greater or lesser volumes of water than they do today. Evidence for higher water planes exists in the form of ancient shoreline features and associated lake deposits. Strandlines and wave-cut cliffs have been traced for many miles in areas peripheral to the modern lakes. In addition, the outlet channels of some of the ancestral lakes have been discovered, and deltas of former river mouths have been mapped.

Water planes below the present levels of the lakes are known from the presence of shallow-water sediments in places where normal deep-water sediments are now accumulating (Hough, 1955). A submerged river channel beneath the Straits of Mackinac led Stanley (1938) to the discovery of low-water stages in the Lake Michigan and Lake Huron basins, which were connected by the submerged valley.

The pronounced changes in the level of water planes were caused by (1) advance and retreat of the ice margin, making the proglacial lakes rise and fall, (2) erosion of lake outlets, resulting in falling water levels, and (3) warping of the crust in response to glacial unloading, causing submergence of the coastal region in some areas and emergence in others.

CRUSTAL WARPING DETERMINED BY NONHORIZONTAL STRANDLINES

The strandlines of many of the ancient lakes rise in elevation toward the north, a fact that, circumstantially, is related to late-glacial and postglacial isostatic adjustment of the earth's crust following glacial unloading. The mapping of elevations on abandoned beaches permits the drawing of isobases on a given water plane. The zero isobase for any water plane defines the hinge line, north of which the strandlines rise in elevation and south of which they have a horizontal attitude.

The hinge lines of the Great Lakes region (Fig. 5), when plotted on a map, form a family of nearly parallel lines trending in a northwesterly direction; the hinge lines defined by the oldest beaches are farthest south, and the

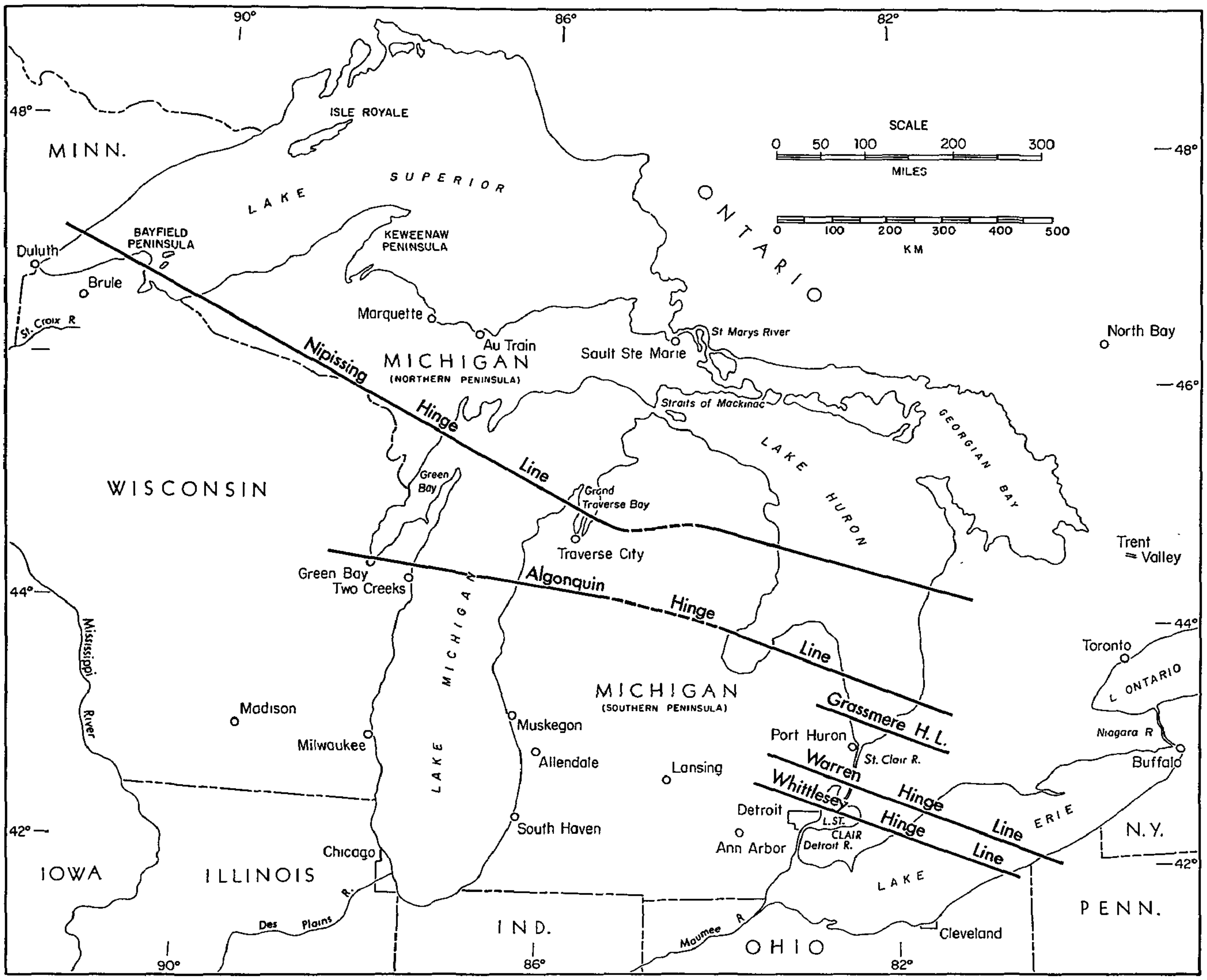

Figure 5. Index map of the upper Great Lakes region showing hinge lines (zero isobases) based on the elevations of the beaches of some of the ancestral stages of the Great Lakes. (Data for hinge lines from Leverett and Taylor, 1915, p. 505, and Hough, 1958, p. 136.)

youngest are farthest north (Leverett and Taylor, 1915, p. 505; Flint, 1957, p. 251; Gutenberg, 1941, p. 743; Hough, 1955, p. 136).

Gutenberg (1933) agreed with Leverett and Taylor (1915) that the cause of crustal uplift in the Great Lakes region was the isostatic rebound that accompanied deglaciation. Moore (1948) concluded from a study of apparent changes in the elevation of gauging stations on the shores of the present Great Lakes that the upwarping was related to earlier tectonic movement rather than isostatic adjustment of the crust during and after the disappearance of glacial ice, but MacLean (1963) showed that Moore's interpretation may have been in error because of a failure to take into account the influence of wind on the pairs of gauges used in his analysis.

Farrand (1962) showed that a uniform pattern of strongly decreasing rate of uplift from the time of deglaciation to the present must be causally related to glacial unloading of the crust.

ANCESTRAL GREAT LAKES

Lake Superior basin. The most recent study of the ancestral stages of Lake Superior is Farrand's (1960), which is here summarized. Earlier work was by Leverett (1929b), Stanley (1932), and Sharp (1953).

The Superior basin was filled by Valders ice about 11,500 years B.P. Murray (1953) postulated a proglacial Lake Keweenaw in at least part of the Lake Superior basin during Two Creeks time (Port Huron–Valders interval), but all shorelines of this water body were destroyed by the advance of the Valders ice (Fig. 6).

As the Valders ice began its withdrawal from the western part of the basin a group of small proglacial lakes formed marginal to the ice border. These lakes, known as glacial Lakes Nemadji, Brule, Ashland, and Ontonagon, formed along the southern border of the retreating ice and discharged westward to the St. Croix River along the Wisconsin-Minnesota border to the Mississippi River. Farrand (1960) referred to these lakes collectively as the epi-Duluth stage, which had original elevations of more than 1,100 ft A.T. (above sea level).

Further retreat of the Valders ice permitted the lakes of the epi-Duluth stage to merge into a single water body peripheral to the ice border, Lake Duluth, the trend of which was from the midpoint of the Minnesota shore southeastward across the lake to the base of the Keweenaw Peninsula in Michigan. Lake Duluth stood at 1,085 ft A.T. before the region was uplifted; it drained westward, first through an outlet at Duluth and later through the Brule–St. Croix River outlet in Wisconsin. The erosion of the latter outlet allowed the water surface of Lake Duluth to fall to three lower stages at 1,070 ft, 1,060 ft, and 1,035 ft A.T. Glacial Lake Duluth terminated a little more than 10,000 years ago when further retreat of the Valders ice opened a lower outlet across the Huron Mountains northwest of Marquette,

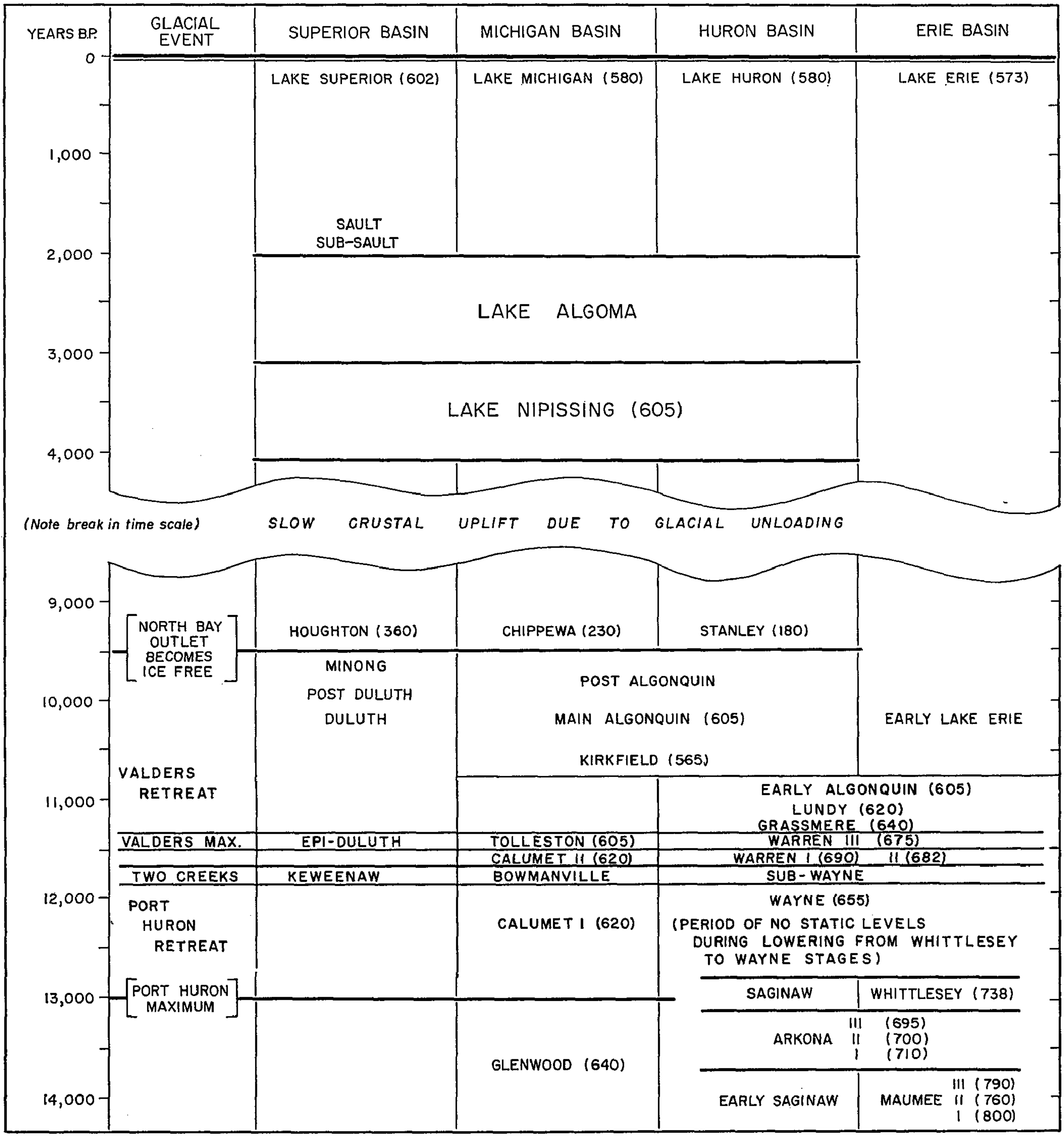

Figure 6. Correlation chart of ancestral lakes of the upper Great Lakes basins. Solid horizontal lines indicate reasonably well-established correlative events. Elevations (in feet) of lake stages from Hough (1958). Correlations based in part on Bretz (1959), Broecker and Farrand (1963), Farrand (1960), Hough (1958), and Leverett and Taylor (1915).

Michigan, and thereby initiated a series of lower lake levels of the post-Duluth stages.

Six post-Duluth glacial lakes, ranging from 1,007 ft A.T. (sub-Duluth stage) to 650 ft A.T. (Beaver Bay stage), discharged southward across the Northern Peninsula of Michigan to one of the ancestral lakes in the Lake Michigan basin. The Valders ice, which was retreating rapidly northeastward along its axis, formed a part of the shore of each of the post-Duluth lakes.

When the edge of the Valders ice finally reached the present north shore of Lake Superior in Canada, Lake Minong at 470 ft A.T. came into existence. It was the first of the lakes in the Superior basin to be free of an ice border along any part of its shoreline. Lake Minong shared a common water plane with one of the ancestral lakes in the Lake Michigan basin, Lake Sheguiandah of the post-Algonquin group. The two water bodies were connected by straits across the Northern Peninsula at Au Train and the St. Marys River, the modern outlet of Lake Superior. Lake Minong ended about 9,500 years ago.

When the ice border had retreated to North Bay, Ontario, a very low outlet to the east was opened for the discharge of the ancestral lakes in the Lake Michigan and Lake Huron basins, into which Lake Minong drained through the St. Marys River. Lake Houghton, whose water level stood at 360 ft A.T., had the lowest stage of any of the ancestral lakes in the Lake Superior basin; it is correlated with the extreme low-water stages of Lakes Chippewa and Stanley in the Lake Michigan and Lake Huron basins, respectively. The Houghton shoreline is now above the present level of Lake Superior (602 ft A.T.) on the Canadian shore because of crustal uplift, but its south shore in Wisconsin and Michigan is submerged. Farrand (1962, p. 187) puts the date of Lake Houghton at 9,600 B.P. Although no part of the Houghton shoreline was bordered by glacier ice, melt-waters were discharging into Lake Houghton (Farrand, 1960, p. 121). Outwash terraces in southward-trending valleys of the Lake Superior watershed in Ontario are graded to the Houghton level.

Lake Michigan basin. Glacial Lake Chicago was formed at the southern extremity of the Lake Michigan Lobe when the latter began its retreat from the Tinley Moraine of Cary age (Fig. 4). Three stages of Lake Chicago are represented by shore features above the modern lake: the Glenwood stage (oldest), 60 ft above Lake Michigan; the Calumet stage, 40 ft; and the Toleston stage, 20 ft (Fig. 6). Leverett (1897) assumed that the three stages of Lake Chicago were caused by successive abrupt lowerings of the water level caused by the erosion of the outlet at Chicago. Wright (1918) proposed that the three stages were not sequential in descending order, the Toleston following the Glenwood and succeeded by the Calumet.

The Glenwood stage of Lake Chicago came into being about 14,000 years B.P. (Bretz, 1959; Hough, 1958). Bretz (1959, p. 680) believed that it ended with the increase in discharge from the glacial Grand River fed by glacial lakes in the Huron and Erie basins; this increased the outflow at Chicago so that a boulder pavement in the drift dam was destroyed by erosion. Ice retreat from the Port Huron maximum opened an eastern outlet that replaced the Chicago outlet, but not until the latter had been deepened by 20 ft of erosion to 620 ft. By Two Creeks time a low-water stage in the Michigan, Huron, and Erie basins prevailed. These levels were below the head of the Grand River outlet.

Readvance of the Valders ice closed the eastern outlet and raised the water level in the Lake Michigan basin to the 620-ft level. This was the Calumet stage. It was not a static level, however, because the glacial Grand River was again swollen by discharge from the Huron and Erie basins (Lake Warren), which caused deepening of the Chicago outlet another 15 ft to the bedrock surface. This bedrock still maintained the level of the Toleston stage at 605 ft A.T., until subsequent events caused it to merge with early Lake Algonquin in the Huron basin (Bretz, 1959).

Hough (1958, p. 166) believed that the Glenwood stage persisted well into the retreatal phase of the Valders ice about 10,000 years B.P. If so, the Glenwood beaches should have been developed on the Valders till near Lake Michigan, but none has been discovered (Bretz, 1959).

Lake Huron and Lake Erie basins. The ancestral water bodies of Lakes Huron and Erie are so intimately related that they must be considered together for a coherent picture. When the Erie and Saginaw Lobes advanced, separate proglacial lakes were formed in front of them; when the lobes retreated, their glacial lakes merged to form a single continuous lake, generally at a lower level. The details of the sequence synthesized here are based on the work of Taylor (Leverett and Taylor, 1915) with modifications by Bretz (1951b, 1953, 1959, 1964) and Hough (1958).

The earliest water body in the Erie basin was glacial Lake Maumee, which was formed perhaps 14,000 years B.P. (W-198) when ice of the Erie Lobe retreated from the Fort Wayne Moraine, and whose outlet at first (highest Maumee or Maumee I) was through the Wabash Valley to the Ohio and Mississippi River systems (Figs. 6 and 7). The lowest level of Lake Maumee (Maumee II) used a lower outlet to the north along the edge of the Saginaw Lobe to the Grand River, which flowed westward across Michigan to Lake Chicago. A subsequent ice advance closed the northern outlet and caused the waters of Lake Maumee to rise to a level somewhat below that of Maumee I. This rise in level initiated Maumee III, the outlet of which was also the Wabash River. The erosion surface along the Wabash Valley planed by the overflow waters of the two high phases of Lake Maumee stands 16 to 22 ft above the modern floodplain and is called the Maumee Terrace (Fidlar, 1948).

About the time of glacial Lake Maumee, a proglacial lake in front of the Saginaw Lobe came into existence. It too discharged to Lake Chicago via the trans-Michigan Grand River valley. Further retreat of the Saginaw Lobe, which was really a sublobe of the Huron Lobe, caused the waters of glacial Lake Saginaw and glacial Lake Maumee to merge into a single water body, Lake Arkona. This lake had three stages, I, II, and III, each of which was lower than the lowest of the three Maumee levels; its outlet was the glacial Grand River. Lake Arkona covered all but the extreme east end of what is now Lake Erie as well as a major part of the Saginaw lowland.

Advance of the ice border separated Lake Arkona into two separate parts at different elevations. That part representing the Saginaw Lobe remained at the Arkona level, but the waters of the Erie basin rose above the Arkona level to a new position intermediate in elevation between the Maumee and Arkona levels. This lake is known as Lake Whittlesey and is particularly important in the sequence because it is correlated with the Port Huron Moraine, the feature on which the Port Huron (Mankato) Substage is defined (Fig. 5).

Subsequent to the Port Huron maximum, the glacial border retreated. The retreat exposed an eastern spillway lower than the head of the Grand River valley and thereby caused the water surface to drop to the level of those in the Saginaw Bay area. This level is known as Warren I (or highest Lake Warren). A less well developed Warren beach, Warren II, at 682 ft A.T., was caused by further retreat. Ice withdrawal continued until the waters in the combined Huron and Erie basins dropped to 655 ft A.T., the Lake Wayne level, and probably lower to a sub-Wayne stage. This event is correlated with the Two Creeks interval (Bretz, 1959, p. 682; Hough, 1958, p. 150). The Grand River valley carried no discharge at this time. Hough (1963, Fig. 7) later correlated the Two Creeks interval with the Kirkfield stage of Lake Algonquin (see the section on the Ontario basin in this paper).

The readvance of the ice to the Valders maximum raised the waters in the Huron and Erie basins from the Wayne stage at 655 ft A.T. to 675 ft A.T., Lake Warren III (lowest Lake Warren); the readvance also closed the eastern outlet and reactivated the Grand River spillway. Bretz (1959, p. 681) believed that the Calumet stage of Lake Chicago occurred between the times of Lake Whittlesey and Lake Warren and that Lake Warren discharged via Grand River to the Toleston stage of Lake Chicago, but Hough (1958, p. 151) argued that Lake Warren was contemporaneous with a third Glenwood stage of Lake Chicago.

The last two stages in the Huron and Erie basins were glacial Lakes Grassmere (640 ft A.T.) and Lundy (620 ft A.T.). Hough (1958, p. 154) thought that Grassmere waters were merged with Glenwood waters of the Lake Michigan basin (both water planes had the same elevation) across the northern part of the Southern Peninsula of Michigan, but Bretz (1959, p. 682) believed that both Grassmere and Lundy discharged eastward near Syracuse, New York, and were never connected with any lake stages in the Lake Michigan basin.

Lake Ontario basin. According to Spencer (1890), Fairchild (1909), and Leverett and Taylor (1915), the retreat of the ice (later identified as of Port Huron age by Mason, 1960, and Karrow *et al.*, 1961) from the Ontario basin was accompanied by the development of ice-marginal lakes around the western and southern borders of the ice (Fig. 7). Lake waters along the western margin were merged with waters in the Erie basin from the time of the Warren stage to the Lundy stage. The opening of the Rome, New York, outlet, which discharged via the Mohawk Valley to the Hudson River, brought an end to the Lundy stage and initiated Lake Iroquois. Lake Lundy was thus reduced to a lake in the Erie basin (Early Lake Erie) which spilled over the Niagara Escarpment to Lake Iroquois. The cutting of the present gorge of the Niagara River began at that time.

Lake Iroquois was 330 ft A.T., according to Hough (1958, p. 202). This figure was derived from an estimate of postglacial uplift, because there are no horizontal Lake Iroquois strands. Glacial retreat from the St. Lawrence lowland brought Lake Iroquois to an end. Leverett and Taylor (1915, p. 445) postulated a short-lived lake stage, glacial Lake Frontenac, following Lake Iroquois; Karrow *et al.* (1961, p. 666) stressed the very short duration of such a stage, because no strong shore features have been found below the Iroquois beaches.

As the St. Lawrence Valley became free of glacier ice, it was invaded by marine waters. Fairchild (1907) believed that this marine embayment extended into the Ontario basin and gave it the name Gilbert Gulf. Mather (1917, p. 542) called it the Champlain Sea, and Hough (1958, p. 203) referred to it as the St. Lawrence Sea. On the basis of stratigraphic sections at Hamilton, Ontario, Karrow *et al.* (1961, p. 665) deduced that marine waters never extended into the Ontario basin, although the early phase of Lake Ontario may have been very close to sea level. Postglacial uplift raised Lake Ontario to its present level.

The correlation of events in the Ontario basin with the late-Pleistocene chronology of the upper Great Lakes poses a serious problem. Hough (1958, Table 22) considered Lake Iroquois to have endured from about 9,300 years to 8,600 years ago and to have been related to the retreat of Valders ice. Mason (1960), however, argued for an older date of Lake Iroquois, making it a correlative of the waning Port Huron (Mankato) ice. Karrow *et al.* (1961, p. 665) presented radiocarbon evidence from the Hamilton, Ontario, sediments that brackets Lake Iroquois from 12,500 to 10,500 years ago, a date that supports Mason's interpretation. Dreimanis (1964, p. 248) pointed out that the acceptance of the Karrow *et al.* dates placed all lake stages from lowest Lake Warren to Lake Lundy (Fig. 7) in pre-Two Creeks time because they are all older than Lake Iroquois. Hough (1963, Fig. 7) attempted to reconcile the Karrow *et al.* dates by correlating the Kirkfield stage of Lake Algonquin with the Two Creeks interval, thereby allowing less than a thousand years for Lakes Whittlesey, Warren, Wayne, Grassmere, and Lundy. Acceptance of this revised correlation conflicts with the thesis held by Bretz (1964, p. 625) that the lowest Warren stage (Warren III at 675 ft A.T.) in the Huron and Erie basins is correlative with the Toleston stage in the Lake Michigan basin, an event that he believed was caused by the advance of the Valders ice after Two Creeks time. In order for the Warren through Lundy stages to have persisted above 600 ft A.T., ice had to block the Ontario basin.

Figure 7 shows the sequence of lake stages in the Ontario basin according to Karrow *et al.* (1961) and Dreimanis (1964) and a possible correlation with the Huron and Erie basin chronology (Hough, 1963, Fig. 7). Comparison of Figures 6 and 7 reveals the incompatibility of the two chronologies between the time of Lake Whittlesey and the Early Algonquin stage.

The lowering of Lake Lundy to Early Lake Erie caused

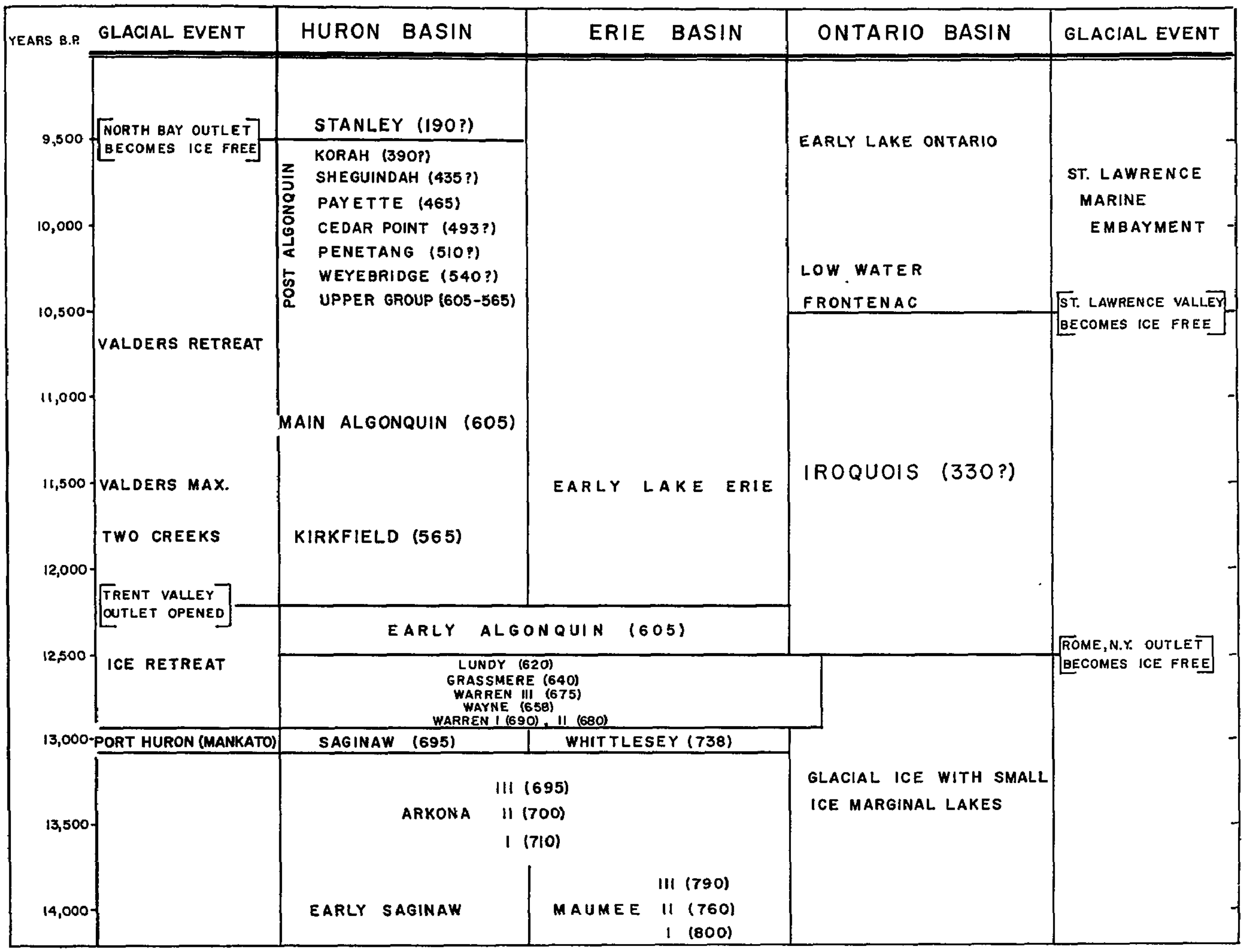

Figure 7. Correlation chart of lake stages in the Ontario, Erie, and Huron basins, based on Hough (1963), Dreimanis (1964), and Karrow *et al.* (1961).

the isolation of the waters in the Huron basin at the 605-ft A.T. level; this lake is called Early Lake Algonquin. It discharged through the newly formed St. Clair and Detroit River systems to Lake Erie and eventually merged with the waters of the Lake Michigan basin to form Lake Algonquin. The Chicago outlet, which had functioned for the Toleston stage, continued to function after the merger with Early Lake Algonquin in the Huron basin, and it gave Lake Algonquin a second outlet.

Lake Algonquin. At its highest level Lake Algonquin covered all the Lake Michigan basin and all the Huron basin except the north shore of Georgian Bay in Ontario, where a post-Valders glacial border stood. Leverett (1929b) believed that Lake Algonquin covered part of the Superior basin, but Hough (1958) and Farrand (1960) showed that the eastern part of the Superior basin was still ice-blocked. Farrand (1960, p. 114) correlated Lake Algonquin with Lake Duluth, believing that the former ". . . might be somewhat older than 10,500 B.P." (Broecker and Farrand, 1963, p. 800).

Actually, the history of Lake Algonquin is quite complex. An early Algonquin stage was restricted to the Huron basin at a time when glacial Lake Chicago stood at the Toleston stage of 605 ft A.T. Merger of these two bodies of water culminated in the Kirkfield stage of Lake Algonquin, 40 to 60 ft lower than the main Algonquin level. The Chicago and St. Clair outlets were abandoned at this time in favor of the lower Trent valley outlet (east of Georgian Bay), which debauched to Lake Iroquois in the Ontario basin. The main Algonquin stage, which followed the Kirkfield stage, was caused either by uplift of the Trent valley or by a readvance of the glacial border. After the main Algonquin level the water level was progressively lowered to levels recorded by an upper group of beaches on Mackinac Island (Stanley, 1938) and elsewhere in Ontario north of the Algonquin hinge line. Below these beaches are four prominent strandlines studied by Stanley (1936, 1937) in the Georgian Bay area. Stanley named these the Wyebridge, Penetang, Cedar Point, and Payette stages. They all used eastern outlets, the most important one of which was at North Bay, Ontario. The original elevation of the last of these low stages, the Payette, was 465 ft A.T., as compared to 605 ft A.T. of the main Lake Algonquin.

Hough (1958, p. 234) suggested two stages below the Payette stage—the Sheguiandah and Korah. During these and previous stages below the main Algonquin level, the Chicago and St. Clair outlets were inoperative.

EXTREME LOW-WATER STAGES IN THE HURON AND MICHIGAN BASINS

Lake Algonquin waters were lowered through successive stages as lower eastern outlets were opened by ice retreat (Chapman, 1954) and eroded by outlet waters. The lowering finally culminated in low-water lakes in the Huron and Michigan basins. Stanley (1936, p. 1958; 1937, p. 1681) inferred the existence of these stages from the submerged channel through the Straits of Mackinac, and Hough (1955) added evidence from cores taken from the bottom of Lake Michigan. Hough gave the names Lake Chippewa and Lake Stanley to the two low-level lakes in the Michigan and Huron basins, respectively. Lake Chippewa was at 235 ft A.T., or about 350 ft below the present level of Lake Michigan, and Lake Stanley stood at 180 ft A.T. Zumberge and Potzger (1956) placed the date of Lakes Chippewa and Stanley at about 5,000 B.P., Hough (1958, p. 282) placed it at 6,000 years B.P., and Farrand (1960, p. 114) put it at 8,500 years B.P. According to Terasmae and Hughes (1960) the North Bay outlet was deglaciated prior to 9,500 years B.P., an event that initiated the Chippewa and Stanley stages. Hough (1963, p. 106) used this date in a later interpretation of the chronology of the Lake Huron basin.

Nipissing Great Lakes. A period of crustal uplift between 9,500 B.P. and 4,000 B.P. raised the North Bay outlet enough to cause the waters in the Michigan and Huron basins to rise until they once again spilled through the Chicago and St. Clair outlets. For a while, all three outlets were in use. Waters in the Lake Superior basin merged with the rising waters of the two lower basins so that the Nipissing Great Lakes encompassed the area of all three basins, Superior, Huron, and Michigan, and were the largest of all postglacial Great Lakes. The ice had retreated far to the north by this time to a position beyond the watersheds of any of the present Great Lakes.

The shoreline features of the Nipissing stage are among the strongest of any found in the Great Lakes region. The mouths of many stream valleys that had been graded to the Chippewa and Stanley levels were drowned by the rising Nipissing waters, and some were cut off from the main lake by construction of bay-mouth bars and spits. Eolian activity along the Michigan coast of Lake Michigan was reestablished during the later part of the rise from the Chippewa to the Nipissing level (Zumberge and Potzger, 1956, p. 279).

The Nipissing Great Lakes endured for 1,000 years between 4,000 and 3,000 B.P. The end of the Nipissing stage came when the St. Clair River at Port Huron was eroded to a lower level. The Chicago outlet had been abandoned earlier because it was floored in bedrock, and downcutting could not keep pace with the St. Clair outlet, which was floored in till.

Algoma stage. The Algoma stage is represented by shoreline features on Lakes Superior, Michigan, and Huron that occur about 10 ft below the Nipissing beaches south of the Nipissing zero isobase (Fig. 5). Before Hough (1953) recognized that the Nipissing water plane in the southern parts of the Michigan and Huron basins was coincident with the main Algonquin stage at 605 ft A.T., the beaches now assigned to the Algoma stage had been considered Nipissing-stage features by earlier workers. The Algoma stage ended about 2,000 years ago when continued downcutting of the outlet at Port Huron brought the levels of Lakes Huron and Michigan to the present elevation of 580 ft A.T.

Lake Superior remained at a higher level, because its outlet, the St. Marys River, is floored in bedrock. Farrand (1960, p. 59) confirmed an earlier suspicion of Taylor (1895, p. 312) that at least one intermediate level between the Algoma stage and the modern Lake Superior water plane existed. This is known as the Sault stage; it is represented by beaches on the north shore of Lake Superior that have been preserved by the crustal uplift still in progress. Elsewhere around the Superior basin the Sault beaches are submerged (Farrand, 1960, p. 61).

The absolute chronology and correlations (Fig. 6) differ from previous ones in that they allow a period of 5,000 years for the transition from Lakes Chippewa and Stanley to the Nipissing Great Lakes. Hough (1958, Table 22) referred to this period as a transition phase and allowed less than 2,000 years for it. Zumberge's interpretation (Zumberge and Potzger, 1956, p. 277), which was followed by Flint (1957, p. 347), did not include any transition period, a view that is not acceptable in light of the radiocarbon date for the deglaciation of the North Bay outlet (Terasmae and Hughes, 1960) and the age of Lake Algonquin given by Broecker and Farrand (1963, p. 800). The time scale presented in Figure 6 therefore is more in accord with the facts than previous absolute chronologies, and concurs with Hough's (1963, p. 106) most recent interpretation.

References

Bergquist, S. G., 1933, The Pleistocene history of the Tahquamenon and Manistique drainage region of the Northern Peninsula of Michigan: Michigan Dept. Conserv., Geol. Surv. Div. Publ. 40, Geol. Ser. 34, Pt. I, 148 p.

Bergquist, S. G., and MacLachlan, D. C., 1951, Pleistocene features of the Huron-Saginaw ice lobes in Michigan: Geol. Soc. Amer. Guidebook, Detroit meeting, Glacial field trip, 36 p.

Bhattacharya, Nityananda, 1962, Weathering of glacial tills in Indiana. I, Clay minerals: Geol. Soc. Amer. Bull., v. 73, p. 1007-1020

Bretz, J H., 1951a, The stages of Lake Chicago, their causes and correlations: Amer. J. Sci., v. 249, p. 401-429

—— 1951b, Causes of the glacial lake stages in Saginaw basin, Michigan: J. Geol., v. 59, p. 244-258

—— 1953, Glacial Grand River, Michigan: Michigan Acad. Sci. Pap., v. 38, p. 359-382

—— 1959, The double Calumet stage of Lake Chicago: J. Geol., v. 67, p. 675-684

—— 1964, Correlation of glacial lake stages in the Huron-Erie and Michigan basins: J. Geol., v. 72, p. 618-627

Broecker, W. S., and Farrand, W. R., 1963, Radiocarbon age of the Two Creeks forest bed, Wisconsin: Geol. Soc. Amer. Bull., v. 74, p. 795-802

Chapman, L. J., 1954, An outlet of Lake Algonquin at Fossmill, Ontario: Geol. Assoc. Canada Proc., v. 6, p. 61-68

Cuancara, A. M., and Melik, J. C., 1961, Bedrock geology of Lake Huron: Univ. Michigan, Great Lakes Res. Div., Fourth Conf. Great Lakes Res. Proc., Publ. 7, p. 116-125

Dillon, L. S., 1956, Wisconsin climate and life zones in North America: Science, v. 123, p. 167-176

Dreimanis, Aleksis, 1958, Wisconsin stratigraphy at Port Talbot on the north shore of Lake Erie, Ontario: Ohio J. Sci., v. 58, p. 65-84

—— 1960, Pre-classical Wisconsin in the eastern portion of the Great Lakes region, North America: Intern. Geol. Congr., 21st Session, Pt. 4, p. 108-119

—— 1964, Lake Warren and the Two Creeks interval: J. Geol., v. 72, p. 247-250

Dryer, C. R., 1894, The drift of the Wabash-Erie region—a summary of results: Indiana Dept. Geol. Nat. Resources Ann. Rep. 18, p. 38-90

Ekblaw, G. E., and Athy, L. F., 1925, Glacial Kankakee Torrent in northeastern Illinois: Geol. Soc. Amer. Bull., v. 36, p. 417-428

Ekblaw, G. E., and Willman, H. B., 1955, Farmdale drift near Danville, Illinois: Illinois Acad. Sci. Trans., v. 47, p. 129-138

Englehardt, D. W., 1960, A comparative pollen study of two early Wisconsin bogs in Indiana: Indiana Acad. Sci. Proc., v. 69, p. 110-118

—— 1962, A palynological study of post-glacial and interglacial deposits in Indiana: Indiana Univ. Ph.D. thesis, 148 p.

Eveland, H. E., 1952, Pleistocene geology of the Danville region: Illinois State Geol. Surv. Rep. Inv. 159, 32 p.

Fairchild, H. L., 1907, Gilbert Gulf (marine waters in the Ontario basin): Geol. Soc. Amer. Bull., v. 17, p. 112

—— 1909, Glacial waters in central New York: New York State Mus. Bull. 127, p. 5-66

Farrand, W. R., 1960, Former shorelines in western and northern Lake Superior basin: Univ. Michigan Ph.D. thesis, 226 p.

—— 1962, Postglacial uplift in North America: Amer. J. Sci., v. 260, p. 181-199

Fidlar, M. M., 1948, Physiography of the lower Wabash Valley: Indiana Div. Geol. Bull. 2, 112 p.

Flint, R. F., 1957, Glacial and Pleistocene geology: New York, John Wiley & Sons, Inc., 553 p.

Flint, R. F., Colton, R. B., Goldthwait, R. P., and Willman, H. B., 1959, Glacial map of the United States east of the Rocky Mountains (Scale 1/750,000): Geol. Soc. Amer.

Frey, D. G., 1959, The Two Creeks Interval in Indiana pollen diagrams: Inv. Indiana Lakes and Streams, v. 5, p. 131-139

Frye, J. C., and Willman, H. B., 1960, Classification of the Wisconsinan Stage in the Lake Michigan glacial lobe: Illinois State Geol. Surv. Circ. 285, 16 p.

Frye, J. C., Willman, H. B., and Glass, H. D., 1964, Cretaceous deposits and the Illinoian glacial boundary in western Illinois: Illinois State Geol. Surv. Circ. 364, 28 p.

Frye, J. C., Willman, H. B., and Black, R. F., this volume, Outline of glacial geology of Illinois and Wisconsin

Goldthwait, J. W., 1907, The abandoned shore-lines of eastern Wisconsin: Wisconsin Geol. Nat. Hist. Surv. Bull. 17, 134 p.

Gooding, Ansel, 1957, Pleistocene terraces in the upper Whitewater drainage basin, southeastern Indiana: Earlham Coll. Sci. Bull. 2, 65 p.

—— 1961, Illinoian and Wisconsin history in southeastern Indiana: Geol. Soc. Amer. Guidebook, Cincinnati meeting, p. 99-106

—— 1963, Illinoian and Wisconsin glaciations in the Whitewater basin, southeastern Indiana, and adjacent areas: J. Geol., v. 71, p. 665-682

—— in press, The Kansan glaciation in southeastern Indiana: Ohio J. Sci.

Guennel, G. K., 1950, History of forests in the Lake Chicago area: Butler Univ. Bot. Stud., v. 9, p. 140-158

Gutenberg, Beno, 1933, Tilting due to glacial melting: J. Geol., v. 41, p. 449-467

—— 1941, Changes in sea level, postglacial uplift, and mobility of the earth's interior: Geol. Soc. Amer. Bull., v. 52, p. 721-772

Harrison, Wyman, 1958, Marginal zones of vanished glaciers reconstructed from the preconsolidation pressures of overridden silts: J. Geol., v. 66, p. 72-95

—— 1959, Petrographic similarity of Wisconsin tills in Marion County, Indiana: Indiana Geol. Surv. Rep. Prog. 15, 39 p.

—— 1960, Original bedrock composition of Wisconsin till in central Indiana: J. Sed. Petrology, v. 30, p. 432-446

—— 1963, Geology of Marion County, Indiana: Indiana Geol. Surv. Bull. 28, 78 p.

Hersey, J. B., 1963, Continuous reflection profiling, *in* Hill, M. N., (ed.), The sea: New York, Interscience Publishers, v. 3, p. 47-72

Horberg, Leland, 1950, Bedrock topography of Illinois: Illinois State Geol. Surv. Bull. 73, 111 p.

—— 1955, Radiocarbon dates and Pleistocene chronological problems in the Mississippi Valley region: J. Geol., v. 63, p. 278-286

Horberg, Leland, and Anderson, R. C., 1956, Bedrock topography and Pleistocene glacial lobes in central United States: J. Geol., v. 64, p. 101-116

Hough, J. L., 1953, Revision of the Nipissing stage of the Great Lakes: Illinois State Acad. Sci. Trans., v. 46, p. 133-141

—— 1955, Lake Chippewa, a low stage of Lake Michigan indicated by bottom sediments: Geol. Soc. Amer. Bull., v. 66, p. 957-968

—— 1958, Geology of the Great Lakes: Urbana, Univ. Illinois Press, 313 p.

—— 1963, The prehistoric Great Lakes of North America: Amer. Scientist, v. 51, p. 84-109

Hubbard, Bela, 1840, Report on Lenawee, Hillsdale, Branch, St. Joseph, Cass, Berrien, Washtenaw, Oakland, and Livingston Counties, with notes on lake ridges and Great Lakes: Michigan State Geol. Surv. Ann. Rep., v. 3, p. 77-111

Kapp, Ronald, and Gooding, Ansel, 1964a, A radiocarbon dated pollen profile from Sunbeam Prairie bog, Darke County, Ohio: Amer. J. Sci., v. 262, p. 259-266

—— 1964b, Pleistocene vegetational studies in the Whitewater basin, southeastern Indiana: J. Geol., v. 72, p. 307-326

Karrow, P. F., Clarke, J. R., and Terasmae, Jaan, 1961, The age of Lake Iroquois and Lake Ontario: J. Geol., v. 69, p. 659-667

Laidly, W. T., 1961, Submarine valleys in Lake Superior: Geogr. Rev., v. 51, p. 277-283

Leighton, M. M., 1933, The naming of the subdivisions of the Wisconsin glacial age: Science, v. 77, p. 168

—— 1957, The Cary-Mankato-Valders problem: J. Geol., v. 65, p. 108-111

—— 1958a, Important elements in the classification of the Wisconsin glacial stage: J. Geol., v. 66, p. 288-309

—— 1958b, Principles and viewpoints in formulating the stratigraphic classifications of the Pleistocene: J. Geol., v. 66, p. 700-709

—— 1959, Important elements in the classification of the Wisconsin glacial stage; a reply: J. Geol., v. 67, p. 594-598

—— 1960, The classification of the Wisconsin glacial stage of north central United States: J. Geol., v. 68, p. 529-552

Leighton, M. M., and Brophy, J. A., 1961, Illinoian glaciation in Illinois: J. Geol., v. 69, p. 1-31

Leighton, M. M., and Willman, H. B., 1950, Loess formations of the Mississippi Valley: J. Geol., v. 58, p. 599-623

Leverett, Frank, 1897, The Pleistocene features and deposits of the Chicago area: Chicago Acad. Sci. Bull., v. 2, 86 p.

—— 1899, The Illinois glacial lobe: U.S. Geol. Surv. Monogr. 38, 817 p.

—— 1902, Glacial formations and drainage features of the Erie and Ohio basins: U.S. Geol. Surv. Monogr. 41, 802 p.

—— 1917, Surface geology of Michigan: Michigan Geol. Biol. Surv. Publ. 25, Geol. Ser. 21, 223 p.

—— 1929a, Pleistocene of northern Kentucky: Kentucky Geol. Surv., Ser. 6, v. 31, p. 1-80

—— 1929b, Moraines and shorelines of the Lake Superior region: U.S. Geol. Surv. Prof. Pap. 154-A, 72 p.

Leverett, Frank, and Taylor, F. B., 1915, Pleistocene of Indiana and Michigan and the history of the Great Lakes: U.S. Geol. Surv. Monogr. 53, 529 p.

MacLean, W. F., 1963, Modern pseudo-upwarping around Lake Erie: Univ. Michigan, Great Lakes Res. Div., Sixth Conf. Great Lakes Res. Proc., Publ. 10, p. 158-168

Malott, C. A., 1922, Physiography of Indiana, *in* Logan, W. N., Handbook of Indiana Geology: Indiana Dept. Conserv. Publ. 21, Pt. 2, p. 59-256

Martin, H. M., 1955, Map of the surface formations of the Southern Peninsula of Michigan, Scale 1/500,000: Michigan Dept. Conserv., Geol. Surv. Div. Publ. 49

—— 1957, Map of the surface formations of the Northern Peninsula of Michigan, Scale 1/500,000: Michigan Dept. Conserv., Geol. Surv. Div. Publ. 49

Martin, L. M., 1916, The physical geology of Wisconsin: Wisconsin Geol. Nat. Hist. Surv. Bull. 36, 547 p.

Martin, P. S., 1958, Pleistocene ecology and biogeography of North America, *in* Hubbs, C. L. (ed.), Zoogeography: Amer. Assoc. Adv. Sci. Publ. 51, p. 375-420

Mason, R. J., 1960, Early Man and the age of the Champlain Sea: J. Geol., v. 68, p. 366-376

Mather, K. F., 1917, The Champlain Sea in the Ontario basin: J. Geol., v. 25, p. 542-554

Melhorn, W. N., 1954, Valders glaciation of the Southern Peninsula of Michigan: Univ. Michigan Ph. D. thesis, 174 p.

Moore, Sherman, 1948, Crustal movement in the Great Lakes area: Geol. Soc. Amer. Bull., v. 59, p. 697-710

Mozola, A. J., 1962, The bedrock topography of Wayne County, Michigan: Michigan Acad. Sci. Arts Lett., v. 47, p. 19-27

Murray, R. C., 1953, The petrology of the Cary and Valders tills of northeastern Wisconsin: Amer. J. Sci., v. 251, p. 140-155

Newberry, J. S., 1874, On the structure and origin of the Great Lakes: New York Lyceum Nat. Hist. Proc., v. 2, p. 136-138

—— 1882, On the origin and drainage of the basins of the Great Lakes: Amer. Philos. Soc., v. 20, p. 91-95

Otto, J. H., 1938, Forest succession of the southern limits of early Wisconsin glaciation as indicated by a pollen spectrum for Bacon's Swamp, Marion County, Indiana: Butler Univ. Bot. Stud., v. 4, p. 93-116

Potzger, J. E., and Wilson, I. T., 1941, Post-Pleistocene forest migration as indicated by sediments from three deep inland lakes: Amer. Midl. Nat., v. 25, p. 270-289

Prettyman, R. L., 1937, Fossil pollen analysis of Fox Prairie bog, Hamilton County, Indiana: Butler Univ. Bot. Stud., v. 4, p. 33-42

Ray, L. L., 1963, Quaternary events along the unglaciated lower Ohio River valley: U.S. Geol. Surv. Prof. Pap. 475-B, p. 125-128

Rhodehamel, E. C., and Carlston, C. W., 1963, Geologic history of the Teays Valley in West Virginia: Geol. Soc. Amer. Bull., v. 74, p. 251-274

Schneider, A. F., and Keller, Stanley, in preparation, Geologic map of the $1° \times 2°$ Chicago Quadrangle, Indiana, Illinois, and Michigan: Indiana Geol. Surv. Regional Geol. Map, Chicago Sheet

Schneider, A. F., Johnson, G. H., and Wayne, W. J., 1963, Some linear glacial features in west-central Indiana (abst.): Indiana Acad. Sci. Proc., v. 72, p. 172-173

Schwartz, G. M., 1949, Geology of the Duluth metropolitan area: Minnesota Geol. Surv. Bull. 33, 136 p.

Scott, I. D., 1921, Inland lakes of Michigan: Michigan Geol. Biol. Surv. Publ. 30, Geol. ser. 25, 383 p.

Sharp, R. P., 1953, Shorelines of the glacial Great Lakes in Cook County, Minnesota: Amer. J. Sci., v. 251, p. 109-139

Shepard, F. P., 1937, Origin of the Great Lakes basins: J. Geol., v. 45, p. 76-88

Spencer, J. W., 1890, The deformation of Iroquois beach and birth of Lake Ontario: Amer. J. Sci., 3rd ser., v. 40, p. 443-451

—— 1891, Origin of the basins of the Great Lakes of America: Amer. Geologist, v. 7, p. 86-97

—— 1896, How the Great Lakes were built: Popular Sci. Monthly, v. 49, p. 157-172

Stanley, G. M., 1932, Abandoned strands of Isle Royale and northeastern Lake Superior: Univ. Michigan Ph.D. thesis, 158 p.

—— 1936, Lower Algonquin beaches of Penetanguishene Peninsula: Geol. Soc. Amer. Bull., v. 47, p. 1933-1960

—— 1937, Lower Algonquin beaches of Cape Rich, Georgian Bay: Geol. Soc. Amer. Bull., v. 48, p. 1665-1686

—— 1938, The submerged valley through Mackinac Straits: J. Geol., v. 46, p. 966-974

—— 1945, Pre-historic Mackinac Island: Michigan Dept. Conserv. Geol. Surv. Div. Publ. 43, Geol. Ser. 36, 74 p.

Stout, Wilbur, 1953, Age of the fringe drift in eastern Ohio: Ohio J. Sci., v. 53, p. 183-189

Stout, Wilbur, and Shaaf, Downs, 1931, Minford silts of southern Ohio: Geol. Soc. Amer. Bull., v. 42, p. 663-672

Stout, Wilbur, VerSteeg, Karl, and Lamb, G. F., 1943, Geology of water in Ohio: Ohio Geol. Surv., 4th ser., Bull. 44, 694 p.

Swickard, D. A., 1941, Comparison of pollen spectra from bogs of early and late Wisconsin glaciation in Indiana: Butler Univ. Bot. Stud., v. 5, p. 67-84

Taylor, F. B., 1895, The Nipissing beach on the north Superior shore: Amer. Geologist, v. 7, p. 86-97

Terasmae, Jaan, and Hughes, O. L., 1960, Glacial retreat in the North Bay area, Ontario: Science, v. 131, p. 1444-1446

Thompson, Maurice, 1886, Glacial deposits of Indiana: Indiana Dept. Geol. Nat. Hist. Ann. Rep. 15, p. 44-56

—— 1889, The drift beds of Indiana: Indiana Dept. Geol. Nat. Hist. Ann. Rep. 16, p. 20-40

Thornbury, W. D., 1937, Glacial geology of southern and south-central Indiana: Indiana Div. Geol., 138 p.

—— 1940, Weathered zones and glacial chronology in southern Indiana: J. Geol., v. 48, p. 449-475

—— 1950, Glacial sluiceways and lacustrine plains of southern Indiana: Indiana Div. Geol. Bull. 4, 21 p.

—— 1958, The geomorphic history of the upper Wabash Valley: Amer. J. Sci., v. 256, p. 449-469

Thwaites, F. T., 1935, Sublacustrine topographic map of the bottom of Lake Superior: Kansas Geol. Soc. Guidebook, Ninth Ann. Field Conf., p. 226-228

—— 1943, Pleistocene of part of northeastern Wisconsin: Geol. Soc. Amer. Bull., v. 54, p. 87-144

—— 1946, Outline of glacial geology: Ann Arbor, Michigan, Edwards Bros., Inc., 129 p.

—— 1949, Geomorphology of the basin of Lake Michigan: Michigan Acad. Sci. Arts Lett., v. 33, p. 243-251

Thwaites, F. T., and Bertrand, Kenneth, 1957, Pleistocene geology of the Door Peninsula, Wisconsin: Geol. Soc. Amer. Bull., v. 68, p. 831-880

Tight, W. G., 1903, Drainage modifications in southeastern Ohio and adjacent parts of West Virginia and Kentucky: U.S. Geol. Surv. Prof. Pap. 13, 111 p.

Wayne, W. J., 1952, Pleistocene evolution of the Ohio and Wabash Valleys: J. Geol., v. 60, p. 575-585

—— 1956a, Pleistocene periglacial environment in Indiana (abst.): Indiana Acad. Sci. Proc., v. 65, p. 164

—— 1956b, Thickness of drift and bedrock physiography of Indiana north of the Wisconsin glacial boundary: Indiana Geol. Surv. Rep. Prog. 7, 70 p.

—— 1958a, Early Pleistocene sediments in Indiana: J. Geol., v. 66, p. 8-15

—— 1958b, Glacial geology of Indiana: Indiana Geol. Surv. Atlas Mineral Resources of Indiana Map. 10

—— 1959, Stratigraphic distribution of Pleistocene land snails in Indiana: Sterkiana, v. 1, p. 9-12

—— 1960, Stratigraphy of the Ohio River Formation: Indiana Geol. Surv. Bull. 21, 44 p.

—— 1963, Pleistocene formations of Indiana: Indiana Geol. Surv. Bull. 25, 85 p.

—— 1964, Pleistocene patterned ground and periglacial temperatures in Indiana (abst.): Geol. Soc. Amer. Spec. Pap. 76, p. 176-177

—— in preparation, The Crawfordsville and Knightstown Moraines in Indiana: Indiana Geol. Surv. Rep. Prog.

Wayne, W. J., and Thornbury, W. D., 1955, Wisconsin stratigraphy of northern and eastern Indiana: Indiana and Ohio Geol. Surveys, 5th Biennial Pleistocene Field Conf. Guidebook, p. 1-34

Wayne, W. J., Johnson, G. H., and Keller, Stanley (in preparation) Geologic map of the $1° \times 2°$ Danville Quadrangle, Indiana and Illinois: Indiana Geol. Surv. Regional Geol. Map, Danville Sheet

Webb, W. M., and Smith, R., 1961, The bedrock geology of Lake Michigan (abst.): Univ. Michigan, Great Lakes Res. Div., Fourth Conf. Great Lakes Res. Proc., Publ. 7, p. 146

Wier, C. E., and Gray, H. H., 1961, Geologic map of the Indianapolis $1° \times 2°$ Quadrangle, Indiana and Illinois: Indiana Geol. Surv. Regional Geol. Map, Indianapolis Sheet

Willman, H. B., Glass, H. D., and Frye, J. C., 1963, Mineralogy of glacial tills and their weathering profiles in Illinois: Illinois Geol. Surv. Circ. 347, 55 p.

Wilson, L. R., 1932, The Two Creeks forest bed, Manitowoc Co., Wisconsin: Wisconsin Acad. Sci. Trans., v. 27, p. 31-46

Wright, G. F., 1918, Explanation of the abandoned beaches about the south end of Lake Michigan: Geol. Soc. Amer. Bull., v. 29, p. 235-244

Wright, H. E., Jr., 1964, The classification of the Wisconsin glacial stage: J. Geol., v. 72, p. 628-637

Zumberge, J. H., 1960, Correlation of Wisconsin drifts in Illinois, Indiana, Michigan, and Ohio: Geol. Soc. Amer. Bull., v. 71, p. 1177-1188

—— 1962a, Problems on the origin of Lake Superior: Metropolitan Detroit Sci. Rev., v. 23, p. 57-59

—— 1962b, A new shipboard coring technique: J. Geophys. Res., v. 67, p. 2529-2536

Zumberge, J. H., and Gast, Paul, 1961, Geological investigations in Lake Superior: Geotimes, v. 6, p. 10-13

Zumberge, J. H., and Potzger, J. E., 1956, Late Wisconsin chronology of the Lake Michigan basin correlated with pollen studies: Geol. Soc. Amer. Bull., v. 67, p. 271-288

Zumberge, J. H., Spurr, S. H., and Melhorn, W. H., 1956, The northwestern part of the Southern Peninsula of Michigan: Univ. Michigan Dept. Geology, Friends of the Pleistocene, Midwest Section, Guidebook, 36 p.