EXPLANATORY TEXT TO ACCOMPANY THE GEOLOGIC MAP OF NORTH DAKOTA

by

Lee Clayton, S. R. Moran, and J. P. Bluemle

REPORT OF INVESTIGATION NO. 69

NORTH DAKOTA GEOLOGICAL SURVEY

Lee C. Gerhard, State Geologist

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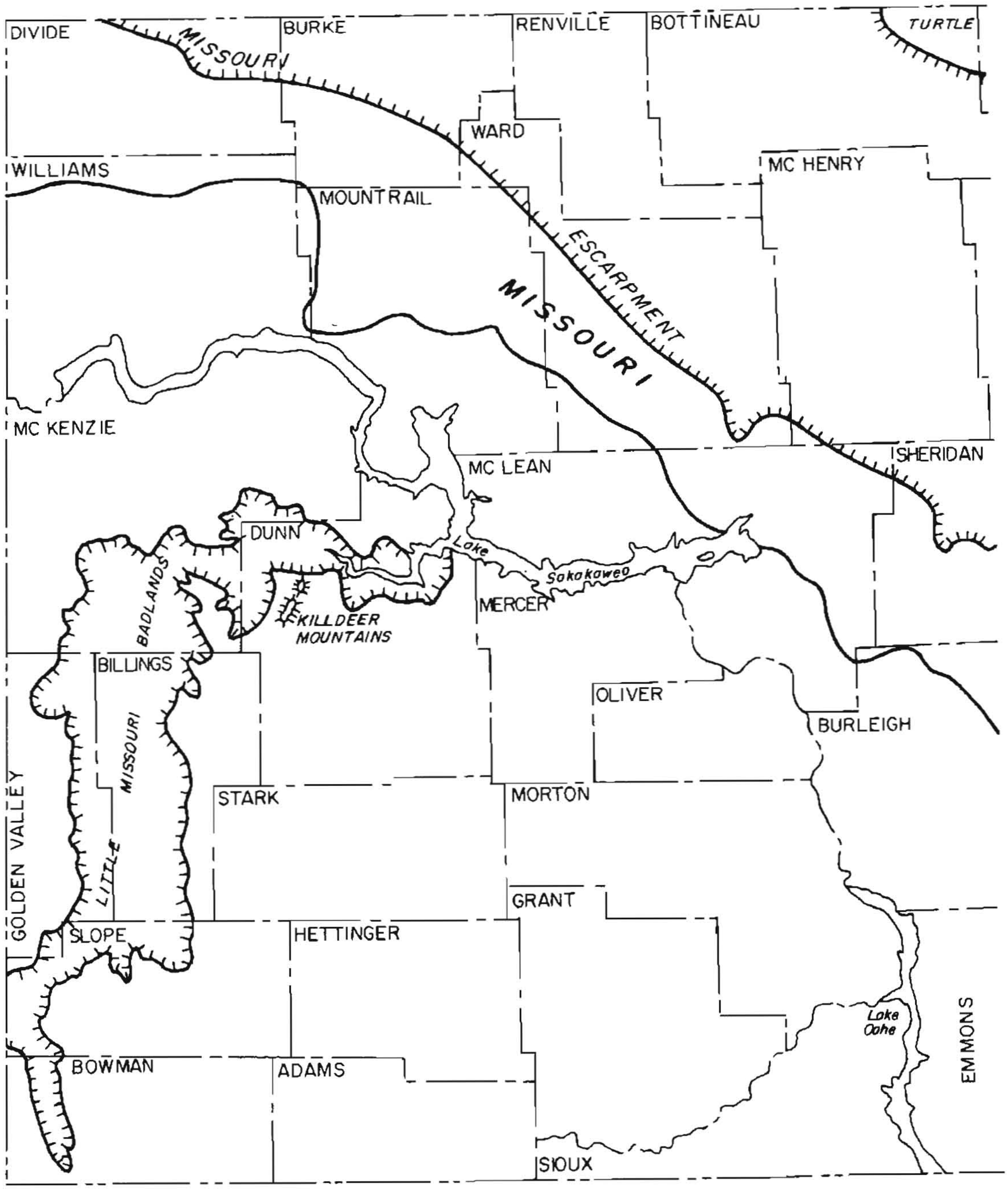
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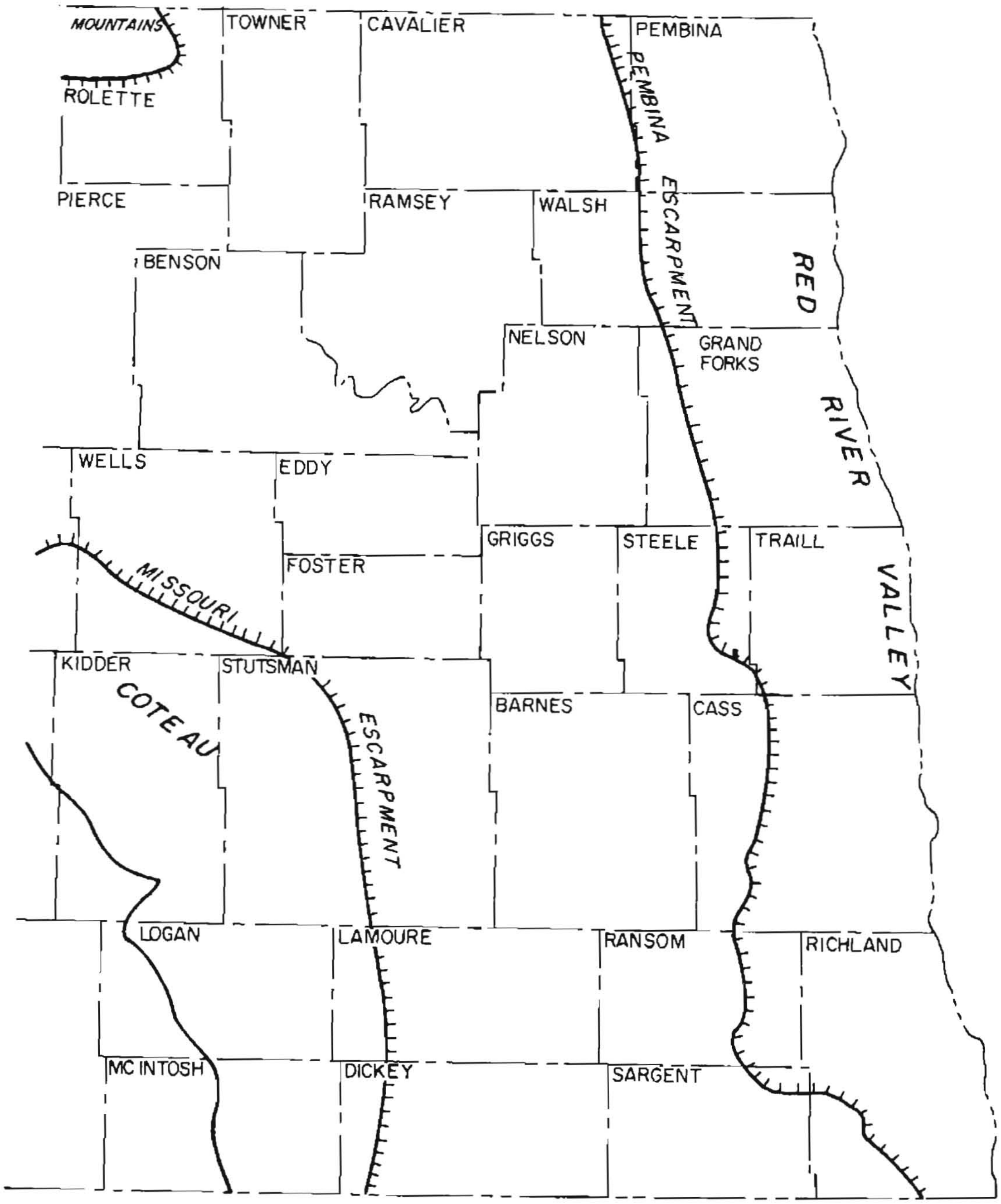

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INTRODUCTION

The U.S. Geological Survey has published a new 1:500,000 Geologic Map of North Dakota by Lee Clayton, with assistance from S. R. Moran, J. P. Bluemle, and C. G. Carlson (1980). This report outlines the background and history of the map, provides detailed explanations of the units shown on the map, gives the basis for differentiating the units, and justifies the approach used in preparing the map.

In 1963 the U.S. Geological Survey published the 1:500,000 Preliminary Glacial Map of North Dakota by Colton, Lemke, and Lindvall. It is a uniformly accurate map, compiled using 1:65,000 airphoto stereopairs, and it remains one of the most useful glacial-geology maps of any area of comparable size in North America. However, at that time, very little of North Dakota had been mapped in detail, and the map is now out of print.

In 1959 the North Dakota Geological Survey began mapping the surface geology of each county at a scale of 1:125,000. The first county report, by Rau and others, was published in 1962. Since then, the geology of all 53 counties has been mapped at a scale of 1:125,000.

This detailed county mapping and associated lithostratigraphic studies (Moran and others, 1976) resulted in a better understanding of the Pleistocene geology of the state. As a result, a new map was planned. In 1974, E. A. Noble (then State Geologist), Moran, and Clayton arranged for the publication of the map by the U.S. Geological Survey. They, at that time, decided that both Pleistocene and pre-Pleistocene material should be shown on the map--that it should be a map of all the surface geology of the state. Clayton compiled the map during the summers of 1975 and 1976, and the present text was largely written in 1978.

We thank the following people for reviewing this explanatory text: Clarence G. Carlson, Alan E. Kehew, David B. Johnson, Kenneth L. Harris, A. M. Cvancara, and F. D. Holland, Jr.

Types of Geologic Maps

Geologic maps may show (1) lithology, (2) stratigraphy, or (3) morphology. In addition, (a) descriptive, (b) genetic, or (c) chronologic aspects may be emphasized. They may be combined in nine different ways.

(1a) Descriptive-lithologic maps show the descriptive characteristics, such as grain size and mineralogy, of the sediment. Map units listed in the explanation might include sand, gravelly sand, or quartz sand.

(1b) Genetic-lithologic (lithogenetic) maps indicate the environment of deposition or the origin of the sediment. Map units might include fluvial sediment, fluvial overbank sediment, or loess.

(1c) Chronologic-lithologic (lithochronologic) maps indicate the age of the sediment. Map units might include Wisconsinan sediment (deposited during the Wisconsinan Age), Pleistocene sediment (deposited during the Pleistocene Epoch), or Cenozoic sediment (deposited during the Cenozoic Era).

(2a) Descriptive-stratigraphic maps show either of two descriptive kinds of stratigraphic units. Rock-stratigraphic (lithostratigraphic) units might include the Riverdale Member, the Oahe Formation, and the White River Group. Special-purpose units include various kinds of hydrostratigraphic units (named aquifers), heavy-mineral zones, and infiltration-potential units.

(2b) Genetic-stratigraphic (event-stratigraphic; ecostratigraphic) maps indicate the geologic event responsible for various bodies of sediment. Map units might include the Lostwood drift (resulting from the Lostwood Glaciation), the Leonard paleosol (resulting from the Leonard Soil-Forming Episode), or the Laramide deposits (resulting from the Laramide Orogeny).

(2c) Chronologic-stratigraphic (chronostratigraphic) maps indicate the age of the deposits. Map units might include the Wisconsinan Stage (deposited during the Wisconsinan Age), the Pleistocene Series (deposited during the Pleistocene Epoch), or the Cenozoic Erathem (deposited during the Cenozoic Era).

(3a) Descriptive-morphologic maps show the descriptive characteristics of the landscape. Map units might include plains, undulating topography, or ridges, or the character of the topography may be indicated by contours or shown pictorially.

(3b) Genetic-morphologic (morphogenetic) maps indicate the origin of landforms. Map units might include esker, beach, or meltwater channel.

(3c) Chronologic-morphologic (morphochronologic) maps show the age of the landforms. Map units might include Wisconsinan landforms (formed during the Wisconsinan Age), Pleistocene landforms (formed during the Pleistocene Epoch), or Cenozoic landforms (formed during the Cenozoic Era).

Lithochronologic and chronostratigraphic maps are essentially identical. Depositional, but not erosional, morphochronologic maps are essentially identical to lithochronologic and chronostratigraphic maps. There are therefore eight basic types of geologic map units.

Practical experience indicates that confusion results if too many different kinds of map units are combined on a single map. However, economic considerations generally prohibit the publishing of eight different maps for each area. A compromise between clarity and economy commonly results in two or three kinds of map units combined on a single geologic map.

In addition, different types of materials may be of interest for different reasons and may be distinguished by different means. Thus, a fluvial plain might not be distinguished from a lacustrine plain on a descriptive-morphologic map, but it would be distinguished on a descriptive-lithologic map; and eskers might not be distinguished from collapsed supraglacial fluvial sediment on a descriptive-lithologic map, but they would be distinguished on a descriptive-morphologic map. Therefore, a mixture of map-unit types is often needed to present all the information of interest.

For example, the Geologic Map of the United States (King and Beikman, 1974) shows chronostratigraphic units (for example, Oligocene), lithostratigraphic units (for example, Jackson Group), lithogenetic units (for example, continental deposits), and descriptive-lithologic units (for example, andesite). On map 1 of the Mountrail County report (Clayton, 1972), lithostratigraphic units (for example, Coleharbor Formation), descriptive-lithologic units (for example, sand and gravel), and descriptive-morphologic units (for example, hilly) have been combined; and on map 2, lithogenetic units (for example, stream sediment) and event-stratigraphic units (for example, deposits associated with the Lostwood Glaciation) have been combined.

Many maps, however, emphasize only one of these eight aspects. For example, the geologic map of Eddy and Foster County (Bluemle, 1965) shows primarily morphogenetic units (for example, dead-ice moraine).

Rodgers (1950, p. 298-301) has pointed out that confusion may result if descriptive and interpretive elements are combined in a single term before all aspects of the interpretation are understood. For clarity, descriptive and interpretive elements are sometimes also separated cartographically. In the Mountrail County report (Clayton, 1972, p. 43-46) descriptive aspects are shown on map 1, and genetic aspects are shown on map 2.

Commonly, however, a more useful geologic map results if both descriptive and interpretive elements are shown, providing that they are clearly separated conceptually. On lunar geologic maps produced by the U.S. Geological Survey, description and interpretation are given in separate paragraphs in the explanation (Wilhelms, 1970, p. 9-12).

Most geologic maps showing pre-Pleistocene units in North Dakota have emphasized descriptive rather than genetic aspects. The color or pattern on most of these maps indicates different lithostratigraphic units, each of which contains sediment of several different origins. The Cannonball Formation, for example, is easily mapped, but marine shoreline deposits and marine offshore deposits are not so easily mapped; however, the explanation of the map may indicate that the sand in the Cannonball Formation is largely marine shoreline sediment and the silt and clay is largely marine offshore sediment.

In contrast, most geologic maps showing Pleistocene deposits in North Dakota have emphasized genetic rather than descriptive aspects. The color or pattern on most of these maps indicates different lithogenetic or morphogenetic units. Lithostratigraphic units, such as the Coleharbor Formation or the Oahe Formation, could have been the main basis for the Geologic Map of North Dakota, but much useful lithologic detail within the units would have been lost. Similarly, purely descriptive-lithologic or descriptive-morphologic units could have been mapped, but the origin of many Pleistocene sediments and landforms is relatively obvious and most geologists prefer to differentiate eolian, fluvial, and shoreline sand.

For these reasons, several different kinds of units have been combined on the Geologic Map of North Dakota, and different kinds of units have been emphasized in different places. Although most materials have been included in some formation, lithostratigraphy has been emphasized only in the pre-Pleistocene units. For Pleistocene deposits, different map colors primarily indicate sediments of different origin. However, morphology is emphasized in areas with uniform lithology but with conspicuous differences in topography; for example, collapsed and uncollapsed river sediment are lithologically identical but have strikingly different topography. Similarly, in areas where the interpretation is unclear, description has been emphasized; the presence of ring-shaped hummocks is indicated even though the reasons for their presence or absence is unknown.

Sources of Information

Most of the sources listed in McIntosh and Eister's (1977) Geologic Map Index of North Dakota were consulted. The main sources of field data were the 1:125,000 maps of Rau and others (1962), Clayton (1962), Winters (1963), Kume and Hansen (1965), Kelly and Block (1967), Bluemle (1965), Hansen (1967), Baker (1967), Klausing (1968), Freers (1970), Bluemle (1967), Bluemle and others (1967), Hansen and Kume (1970), Freers (1973), Clayton (1972), Carlson (1973), Bluemle (1973), Deal (1971), Carlson and Freers (1975), Bluemle (1971), Arndt (1975), Bluemle (1975), Lemke (1960), Clayton (1969), Trapp and Croft (1975), Bluemle (1979a), Bluemle (1979b), and unpublished 1:125,000 maps by Clayton (McKenzie, Golden Valley, Billings, Adams, Grant, Sioux, Morton, Emmons, Towner, Ramsey, Cavalier, and Pembina Counties), Bluemle (Sheridan County), Hobbs (Ramsey County), Moran (McHenry, Bottineau, and Ramsey Counties), Carlson (Morton, Grant, Sioux, Adams, Slope, and Bowman Counties), W. B. Bickley, Jr. (Emmons County), and W. A. Pettyjohn (Ward and Renville Counties). The extent of the Golden Valley Formation is largely from an unpublished map by Hickey (1966); this map has since been published (Hickey, 1977).

Additional lithologic and morphologic information was supplied by 1:125,000 soils maps of every county (Patterson and others, 1968) and miscellaneous field notes of Clayton, Moran, Bluemle, Carlson, and Walter L. Moore (Geology Department, University of North Dakota). Geologic contacts were refined using 1:65,000 airphoto stereopairs taken in 1952 by the Army Map Service (eastern North Dakota) and in 1951, 1952, and 1953 by the Aero Service Corporation (western North Dakota).

Methods of Preparation

Few of the contacts or units shown

on the source maps are identical to the contacts or units shown on the final Geologic Map of North Dakota. The source maps were produced by a large number of different geologists; each geologist had a different approach to geologic mapping, and different maps were compiled for different purposes. As a result, many adjustments and compromises were required to produce a unified map. One geologist might place the boundary between two formations stratigraphically higher than another geologist in an adjacent area; if both contacts were stratigraphically justifiable, I have used the one most conspicuous on airphotos throughout the area. Similarly, in areas of patchy glacial sediment on Paleocene sediment, one geologist might map glacial sediment only where it is thicker than 2 metres, whereas another geologist in an adjacent area might map glacial sediment as thin as 1 metre; I have shifted the contacts so the same thickness is mapped throughout the area, giving preference to the thickness that is most easily interpreted on airphotos.

Some source maps were compiled before certain present-day geologic concepts or practices had evolved. For example, the contact between the Sentinel Butte Formation and the Bullion Creek Formation had not been mapped in most areas before 1967, and the contact between the Ludlow and Slope Formations in Bowman and Slope Counties had not been mapped before 1977. In these areas new contacts had to be located.

Some generalization has been necessary in changing from the 1:125,000 scale of most source maps to the 1:500,000 scale of the Geologic Map of North Dakota. An area of sediment is generally indicated only if it is more than about 0.8 kilometre across. But some conspicuous units (such as the caprock of isolated buttes) are shown even if they are less than 0.8 kilometre across, and some inconspicuous units (such as thin wind-blown silt) are not shown even if they are more than 0.8 kilometre across. An attempt was made to map surface units more than 1 metre thick, but some Holocene fluvial sediment less than 1 metre thick was mapped on the Lake Agassiz plain where the nature of the underlying sediment was unknown, and some wind-blown silt more than 1 metre thick southwest of the Missouri River could not be accurately mapped. Some features narrower than 0.8 kilometre (for example, eskers) are indicated by line symbols.

The 1:125,000 soil maps were used to provide objective unity in areas of conflicting lithologic information. Where adjacent or overlapping geologic maps showed different surface lithologies, the soil maps were used to resolve the conflict. The grain size of subsoil on most soil maps is based on considerably denser field observations than is the grain size of surface sediment on most geologic maps.

All contacts shown on the map were reinterpreted or newly interpreted using the 1:65,000 airphoto stereopairs. Every area was first viewed stereoscopically, and all obvious contacts were drawn on the photos. The contacts were then checked against existing geologic maps of the area and were revised as required by the evidence of the geologic maps. The contacts were then penciled onto a 1:500,000 transparent base map using a transparent 1-inch (1-mile) grid on the airphotos and a 1/8-inch (1-mile) grid under the base map. After the state was completely mapped, a unified explanation was devised and the contacts were revised and inked. These contacts were then scribed by a draftsman to produce the final map. Most lithologic and stratigraphic contacts are located within 0.4 kilometre of their actual position.

Map Colors

Traditionally, materials of different age have been represented by different hues. Cenozoic units are commonly shown as shades of yellow or gray, and Cretaceous units are commonly shown as shades of yellowish green. This scheme is unsatisfactory for the Geologic Map of North Dakota, however, because 32 shades of yellow and gray and 4 shades of yellowish green would be needed, but no reds or blues would be used. Instead, a color scheme was devised that reflects the multiple nature of the map explanation. Lithostratigraphy is emphasized for pre-Pleistocene units, which are indicated by drab colors, and lithogenesis is emphasized for the Pleistocene deposits, which are indicated by bright colors.

Where possible, the colors of pre-Pleistocene map units reflect the actual outcrop color of the formation. The Sentinel Butte Formation, for example, is dark drab brown, whereas the Bullion Creek Formation is light yellowish brown.

The color of the Pleistocene map units also reflects field colors. The generally finer grained units (offshore and glacial) are indicated by cooler colors (blue and green). The coarser grained units (eolian, shoreline, and fluvial) are indicated by warmer colors (yellow, orange, and red). Finer grained sediment more easily retains water and has more reduced colors (blue and green), whereas coarser grained sediment more easily drains and has more oxidized colors (yellow and red).

Units that are more conspicuous in the field are made more conspicuous on the map. For example, eolian sand, which is often in the form of conspicuous dune fields, is shown bright yellow. Eolian silt, which is generally an inconspicuous blanket draped over the pre-existing landscape, is shown light yellow.

Field units with more hilly topography (and therefore with abundant dark shadows) are shown darker than units with flatter topography (and therefore with few shadows). For example, collapsed fluvial sediment is shown darker than uncollapsed fluvial sediment. Badland areas (with abundant shadows) are indicated by a dark dotted pattern.

Stratigraphic Units

Lithostratigraphic (rock-stratigraphic) procedures used are in most cases those outlined by the American Commission on Stratigraphic Nomenclature (1970). Because "concepts based on inferred geologic history . . . properly play no part in the definition or differentiation of a rock-stratigraphic unit" (article 4c), all groups, formations, and members are considered to be strictly descriptive stratigraphic units. For example, the Coleharbor Group is defined on the basis of grain size, mineralogy, outcrop appearance, and stratigraphic position. The interpretation that it contains glacial sediment has no bearing on its definition.

The rest of this report is organized stratigraphically, starting with the oldest formations exposed in the state.

CARLILE AND NIOBRARA FORMATIONS

The Niobrara and Carlile Formations have too small an outcrop area to be shown separately on the Geologic Map of North Dakota.

The Carlile Formation consists of soft, black, noncalcareous shale deposited in an offshore marine environment in Late Cretaceous time. It is a little over 30 metres thick in northeastern North Dakota. It outcrops in the steep valley walls along the Pembina River from the Manitoba border to within 6 kilometres of Walhalla and along the Campbell Scarp northwest of Walhalla. The Carlile Formation is equivalent to the Morden Member of the Vermillion River Formation of Manitoba (Halstead, 1959, p. 8-9). Although the type section of the Morden is only 30 kilometres north of Walhalla and the type section of the Carlile is in Colorado, the name "Carlile" is used here because it has priority and has been used in nearly all previous North Dakota references. The correlations of previous geologists have been accepted, and no attempt was made to correlate it from its type section.

The Niobrara Formation has two members. The lower member consists of gray, calcareous shale deposited in an offshore marine environment in Late Cretaceous time. Both members are characterized by white specks of calcareous material about a millimetre in diameter. The upper member consists of marly clay that is light yellowish or orangish brown in outcrop. In areas of high relief, the upper member forms a cliff. The Niobrara Formation is about 45 metres thick in northeastern North Dakota. It outcrops along the Pembina River and its tributaries west of Walhalla, along the base of the Pembina Escarpment in southwestern Pembina County, in gullies near Niagara in western Grand Forks County, and in the banks of the Sheyenne meltwater channel in the Fort Ransom and Kathryn area. The Niobrara Formation is equivalent to the Boyne Member of the Vermillion River Formation in Manitoba (Halstead, 1959, p. 9). Although the type section of the Niobrara Formation is in Nebraska, the name "Niobrara" is used here because it has priority and because it has been used in nearly all previous North Dakota references.

PIERRE FORMATION

The Pierre Formation consists of a few hundred metres of gray shale deposited in an offshore marine environment in Late Cretaceous time (Gill and Cobban, 1965). It is easily distinguished from the Niobrara Formation in eastern North Dakota because the lowest part of the Pierre Formation is gray, noncalcareous shale with light-colored clay beds, whereas the upper member of the Niobrara is light-yellowish-brown, marly clay. In areas of high relief, such as along the Pembina River, the two formations can be distinguished on airphotos: the lowest part of the Pierre has gentle slopes and is commonly slumped, whereas the upper member of the Niobrara forms steep slopes.

The Pierre is exposed in the banks of meltwater channels and Holocene gullies cut into the Pembina Escarpment in northeastern North Dakota, in the banks of much of the Sheyenne meltwater channel north of Jamestown, in gullies east of the Missouri Escarpment south of Jamestown, along the Missouri River and its tributaries in southern Emmons County and southeastern Sioux County, and in western Bowman County.

FOX HILLS FORMATION

The Fox Hills Formation consists of as much as 120 metres of reddish-yellow sand and sandstone deposited in a shoreline environment and dark-reddish-brown shale deposited in an offshore environment in Late Cretaceous time (Cvancara, 1976a). In outcrop, it is easily distinguished from the Pierre Formation by the presence of sand and by its brown color. In western Bowman County, southeastern Sioux County, and southwestern Emmons County, it is easily distinguished from the Pierre Formation on airphotos: the Pierre has a more dense drainage network, has a more mottled soil pattern, has fewer cultivated fields, has more gentle slopes, and is more subject to slumping. However, where the base of the Fox Hills is poorly exposed in central Emmons County and southern Benson County, the contact is less conspicuous on airphotos, and the placement of the contact on the Geologic Map of North Dakota was more dependent on available outcrop information.

The features mapped as drumlins in northeast Emmons County by Colton, Lemke, and Lindvall (1963) are here interpreted to be exhumed beach deposits in the Fox Hills Formation.

HELL CREEK FORMATION

The Hell Creek Formation consists of as much as 150 metres of gray sand and sandstone generally interpreted to be fluvial channel sediment and dark-gray silt and clay generally interpreted to be fluvial overbank sediment deposited during Late Cretaceous time (Moore, 1976).

In the field, the Hell Creek is easily distinguished from the Fox Hills Formation. The sand in the Hell Creek has more conspicuous cross bedding than the sand in the Fox Hills. Hell

Creek outcrops are characterized by abundant dark-purple concretions. The Hell Creek has more dark-gray tones, whereas the Fox Hills has more reddish-yellow tones. The Hell Creek has steeper slopes and commonly forms badlands with rilled slopes, whereas the Fox Hills has few outcrops other than vertical cutbanks and rarely has rilled slopes. Soils developed on the Hell Creek Formation are poorer than on the Fox Hills Formation, and sagebrush is more abundant on the Hell Creek.

The Hell Creek can generally be distinguished from the Fox Hills Formation on airphotos. The Hell Creek has a denser drainage network and a more mottled soil pattern. The Fox Hills Formation has more uniform airphoto tones.

LUDLOW FORMATION

The Ludlow Formation consists of as much as 100 metres of silt, sand, clay, and sandstone, which are grayish brown and yellowish brown in outcrop, and lignite. They were deposited in fluvial, lacustrine, and paludal environments in Paleocene time (Moore, 1976).

The definition of the Ludlow used here is that of Clayton and others (1977). They accepted the long-used definition of the unit in its type area in South Dakota; the Ludlow is considered to be stratigraphically below the Cannonball Formation. Previous usage in southwestern North Dakota placed the top of the Ludlow at the contact between drab and bright sediment at the base of the Bullion Creek ("Tongue River") Formation. That is, the Ludlow Formation was considered to have an upper part (here considered to be the Slope Formation) and a lower part (equivalent to type Ludlow). Where the Cannonball Formation is absent, Clayton and others (1977) considered the top of the Ludlow Formation to coincide with the top of the T Cross Bed.

In the field, the Ludlow Formation is distinguished from the Hell Creek Formation by the great lateral persistence of beds in the Ludlow and the lack of lateral persistence in the Hell Creek. The Hell Creek contains more large-scale cross bedding than the Ludlow. The Ludlow contains several laterally persistent lignite beds a few metres thick, whereas the Hell Creek contains only a few lignite beds, which are generally less than 1 metre thick and can be traced laterally no more than 1 or 2 kilometres. The Ludlow lacks the dark-purple concretions and dinosaur bones present in the Hell Creek. The Ludlow and Hell Creek Formations are both generally drab colored, but the Ludlow contains some yellow beds, which are lacking in the Hell Creek. The Hell Creek has steeper slopes, more rilling on badland slopes, poorer soils, and more sagebrush than the Ludlow Formation. In many areas the contact is marked by a bed of black clay with a popcorn-like weathering surface at the top of the Hell Creek Formation.

On airphotos, the Ludlow Formation is distinguished from the Hell

Creek by its less dense drainage pattern and by its smoother, less mottled soil pattern.

The Ludlow thins eastward to about 5 metres in Morton County, where it becomes unmappable.

CANNONBALL FORMATION

The Cannonball Formation consists of as much as 120 metres of sand that is reddish olive in outcrop and shale that is dark brown in outcrop. The sand was deposited in a marine shoreline environment and the shale was deposited in a marine offshore environment in Paleocene time (Cvancara, 1976b). It outcrops in few places other than steep river cutbanks. It is lithologically and topographically similar to the Fox Hills Formation. It is redder or yellower than the Ludlow and Slope Formations, is better sorted, has laterally more persistent bedding, and has marine rather than fresh-water or terrestrial fossils. On airphotos, the Cannonball can generally be distinguished from the Ludlow and Slope Formations by its smoother, less irregular topography having more gentle slopes and laterally persistent benches held up by the shoreline sand.

SLOPE FORMATION

The Slope Formation is lithologically nearly identical to the Ludlow Formation (Moore, 1976, considered it to be the upper part of the Ludlow). In Morton, Grant, Adams, and eastern Bowman Counties, the Slope and Ludlow Formations are separated by the Cannonball Formation. In western Bowman and Slope Counties, where the Cannonball is absent, the contact is considered to be the top of the T Cross Bed (Clayton and others, 1977); on the Geologic Map the contact in this area was from an unpublished map by C. G. Carlson, which was in part derived from a map by Hares (1928). The top of the Slope Formation is in most places marked by the "Rhame Bed," which is interpreted to be a weathering zone on an unconformity between the Slope and Bullion Creek Formations (Wehrfritz, 1978). The Slope Formation is as thick as 100 metres and thins northeastward to a few metres in Oliver County. It cannot generally be distinguished from the Ludlow Formation on airphotos unless the Cannonball Formation intervenes, but in some badland areas the lignite of the T Cross Bed can be recognized on airphotos. The Slope is generally indistinguishable from the Bullion Creek Formation on airphotos, but in some areas the white "Rhame Bed" can be recognized.

BULLION CREEK FORMATION

The Bullion Creek Formation consists of as much as 200 metres of silt, sand, clay, and sandstone, which are yellowish brown in outcrop, and lignite. They were deposited in fluvial, lacustrine, and paludal environments in Paleocene time (Jacob, 1976). It is distinguished from the Slope Formation primarily by color: the Bullion Creek is predominantly yellowish brown in outcrop, whereas much of the Slope Formation is grayish brown. In most North Dakota Geological Survey reports published between 1967 and 1977 this unit was called the "Tongue River Formation" in the Little Missouri badlands. Between about 1948 and 1967 it was commonly considered to be an unnamed lower subdivision of the Tongue River Formation. In south-central North Dakota and northwestern South Dakota, it was considered to be an unnamed middle subdivision of the Tongue River Formation (the lower subdivision is now called "Slope Formation"). The relationship between the type Tongue River in Wyoming and this unit is unknown, and it therefore was named the "Bullion Creek Formation" by Clayton and others (1977).

SENTINEL BUTTE FORMATION

The Sentinel Butte Formation consists of as much as 200 metres of silt, sand, clay, and sandstone, which are grayish brown in outcrop, and lignite. They were deposited in fluvial, lacustrine, and paludal environments in Paleocene time (Jacob, 1976). The Sentinel Butte Formation is distinguished from the Bullion Creek Formation by color: the Sentinel Butte is predominantly grayish brown in outcrop, whereas the Bullion Creek Formation is predominantly yellowish brown. In areas with few outcrops and on airphotos, they can often be distinguished by the steeper, rougher topography of the Sentinel Butte Formation. However, in some areas with few outcrops around the southeastern side of the Williston Basin, the precise location of the contact is unknown.

GOLDEN VALLEY FORMATION

The Golden Valley Formation consists of two dissimilar members (Hickey, 1977). The lower one, the Bear Den Member, consists of as much as 20 metres of clay, silt, or sand, which is white or yellow in outcrop. In some areas, the Bear Den Member is capped with as much as 1 metre of chert. Except for its stratigraphic position, the Bear Den Member is nearly indistinguishable from the "Rhame Bed" at the top of the Slope Formation. Like the "Rhame Bed," it is interpreted to be a weathering zone on an unconformity. The weathering zone of the Bear Den developed on the top of the Paleocene Sentinel Butte Formation and is unconformably overlain by the Eocene Camels Butte Member. The Camels Butte Member consists of as much as 60 metres of micaceous sand, sandstone, silt, and clay, which are yellowish brown in outcrop. They were deposited in fluvial environments. The Camels Butte Member can generally be distinguished from the underlying Sentinel Butte Formation, which has more lignite, has less mica, and is darker colored. In McKenzie County the sandstone in the Camels Butte Member is well exposed where it caps buttes, but in most areas this member is poorly exposed and is difficult to distinguish from the

Sentinel Butte Formation on airphotos. In most areas, however, the contact between the Sentinel Butte and Golden Valley Formations has been located in the field with considerable accuracy because the Bear Den Member has been highly weathered, has few nutrients, is sparsely vegetated, is well exposed, and, therefore, is a good marker bed. In contrast to most other contacts on the Geologic Map, the contact between the Sentinel Butte and Golden Valley Formations was based almost entirely on pre-existing maps.

WHITE RIVER GROUP

The White River Group consists of two units. The lower one, the Chadron Formation, consists of as much as 30 metres of light-colored sand with pebbles and cobbles of quartzite and porphyry overlain by dark-colored clay with a popcorn-like surface on outcrops. The overlying Brule Formation consists of as much as 30 metres of pinkish-brown siltstone, clay, and sand. Both formations are interpreted to be fluvial and lacustrine sediment deposited during Oligocene time (Denson and Gill, 1965). They are well exposed in a few small areas of badlands in eastern Slope County and western Stark County and on isolated buttes (fig. 1).

MIDDLE AND LATE TERTIARY ROCK UNDIVIDED

Many of the buttes of southwestern North Dakota are capped with as much as 120 metres of sandstone or limestone deposited in a fluvial or lacustrine environment in Oligocene, Miocene, or Pliocene time. Although they are well exposed, many of these bodies of sediment are widely separated from each other, are lithologically quite different from each other, have been poorly studied, and as a result, cannot be confidently assigned to a specific formation. Many were once assigned to the White River Group (Hansen, 1956), but they have been included in the White River on the Geologic Map of North Dakota only if information was available indicating that they are lithologically similar to the Chadron or Brule Formations of western Stark County and eastern Slope County. Denson and Gill (1965) assigned many of these bodies of sediment, especially those with light-greenish-gray, tuffaceous, coarse sandstone, to the Miocene Arikaree Formation. The 120 metres of lacustrine limestone and sandstone on the Killdeer Mountains in Dunn County (Clayton, 1969) have been included in the "Killdeer Formation" by Delimata (1975).

LATE TERTIARY AND QUATERNARY SEDIMENT UNDIVIDED

The late Cenozoic sediment not included in the Coleharbor or Oahe Formations in southwestern North Dakota includes a variety of poorly studied deposits, mostly of fluvial origin. For example, the remnants of pediments flanking the Killdeer Mountains are blanketed with 2 or 3 metres of poorly sorted fluvial gravel

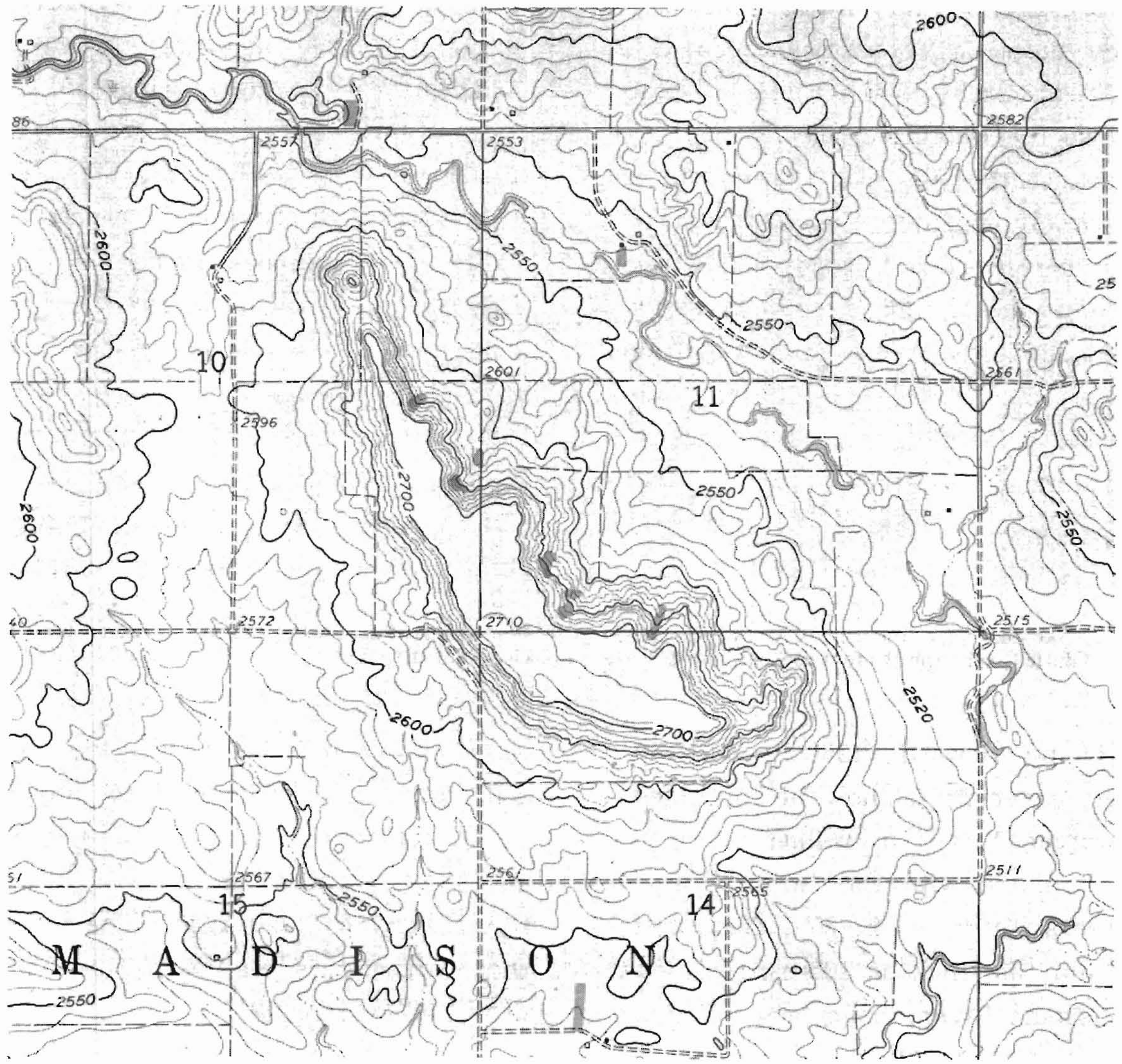

Figure 1. Butte capped with resistant rock of the White River Group in T136N, R94W, in northern Hettinger County. U.S. Geological Survey White Butte West Quadrangle (10-foot contour interval). Area shown is 3.7 kilometres wide.

composed of rubble derived from the middle (?) Cenozoic sandstone and limestone at the top of the Mountains; Clayton (1969) speculated that it was deposited in Early Wisconsinan time. Late Pleistocene terraces along the Cannonball, Heart, and Knife Rivers and their tributaries are capped by a few metres of sand and gravel derived from local Paleocene formations. The Pleistocene terraces along the Little Missouri River, Redwing Creek, and Tobacco Garden Creek are capped by sand and gravel derived from the Black Hills and from local early Tertiary formations. About 10 metres of gravel occurs beneath the glacial sediment at the top of the bluffs near the mouth of the Yellowstone River; it contains quartzite and porphyry derived from the Rocky Mountains and has been called the "Cartwright

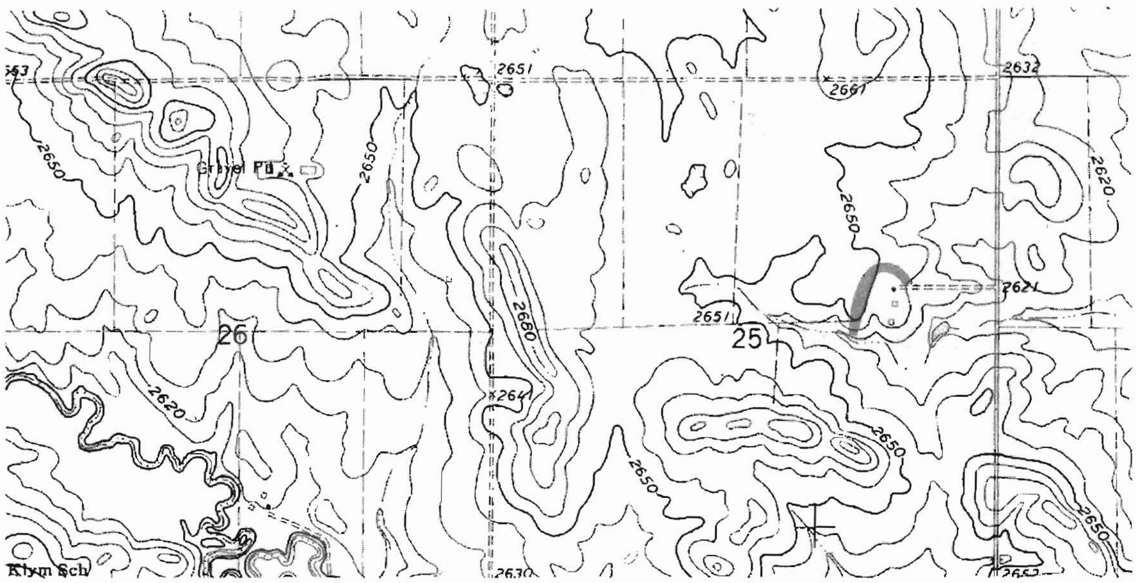

Figure 2. Esker-like ridge southeast of Gorham in T142N, R99W, in Billings County. U.S. Geological Survey Rattlesnake Butte Quadrangle (10-foot contour interval). Area shown is 3.7 kilometres wide.

gravel" by Howard (1960, p. 19-21). Similar material, along the preglacial Yellowstone Valley in Williams County, was correlated with the "Wiota Gravel" by Freers (1970, p. 22-25). Clay, silt, and sand, as thick as 100 metres along the Missouri River, have been included in the Charging Eagle Formation by Ulmer and Sackreiter (1973).

Esker-Like Ridges

In southwestern North Dakota, in an area that has never been glaciated, a number of esker-like ridges have been indicated on the Geologic Map of North Dakota, using the same red-line symbol used for eskers. The ridges consist of fluvial sand and gravel locally derived from older formations. They stand as ridges above adjacent areas because of topographic inversion resulting from differential erosion (fig. 2). Rainfall infiltrates into the sandy soil more easily than into adjacent clayey soil, resulting in less runoff and less slope-wash erosion on the fluvial channel deposits. The age of the sediment is unknown, but it is probably Pleistocene.

COLEHARBOR GROUP

Stratigraphy

The Coleharbor Group consists of as much as 200 metres of bedded, nonorganic clay, silt, sand, and gravel and unbedded sandy, silty clay with pebbles, cobbles, and boulders of granite, gneiss, and basalt derived from the Canadian Shield, limestone and dolomite derived from lower Paleozoic formations flanking the Canadian Shield, and shale derived from the Cretaceous formations in eastern North Dakota and southern Manitoba and Saskatchewan. The silt and clay were deposited mostly in offshore lacustrine environments. The sand and gravel were deposited in fluvial and shoreline environments. The unbedded pebbly, sandy, silty clay was deposited in glacial environments.

The top of the Coleharbor Group roughly corresponds to the Holocene/ Wisconsinan boundary. The Coleharbor in most outcrops can be distinguished from the overlying Oahe Formation (Holocene) by the absence of diffuse organic material.

The Coleharbor is distinguished from other Pleistocene deposits discussed in the previous section by the presence of pebbles derived from the Canadian Shield and the lower Paleozoic formations fringing the Canadian Shield; in contrast, the Pleistocene deposits of southwestern North Dakota contain pebbles derived from the Black Hills and the Rocky Mountains and from local Paleocene formations.

The oldest part of the Coleharbor Group is as old as the first glacier to bring rock from the Canadian Shield into North Dakota. It is distinguished from older late Cenozoic formations by the presence of pebbles from the northeast rather than pebbles from the southwest. The Coleharbor is distinguished from all other formations in North Dakota by the presence of unbedded, pebbly, sandy, silty clay.

The basis for subdivision of the Coleharbor Group has been stratigraphic position, outcrop appearance, grain-size distribution, and the petrology of the very coarse sand of the pebbly, sandy, silty clay units (Moran and others, 1976). The subdivisions are true lithostratigraphic units, not event-stratigraphic units, because inferred geologic history has had no bearing on their definitions. These subdivisions were not shown on the Geologic Map of North Dakota because they have not yet been mapped in much of North Dakota.

Most of the rest of this report is a discussion of the different materials contained within the Coleharbor Group.

Lacustrine Features

Shoreline Sediment

Shoreline sediment in the Coleharbor Group consists of as much as 5 metres of well-sorted sand and gravel. Mappable shoreline deposits (more than 0.8 kilometre wide) occur in only a few areas along the margin of Lake Agassiz (fig. 3), around Horsehead Lake in Kidder County, and between the lakes west of Napoleon in Logan County. Most of the beach deposits of Lake Agassiz are less than 0.8 kilometre across and are therefore indicated only by a line symbol. Only the beaches that are clearly recognizable on the 1:65,000 airphotos are indicated on the map. None of the high Agassiz beaches identified by Bluemle (1974, fig. 3; 1977, fig. 23) in Barnes County, for example, has been indicated on the Geologic Map; the one 2.5 kilometres west of Oriska (Bluemle, 1972, fig. 13)

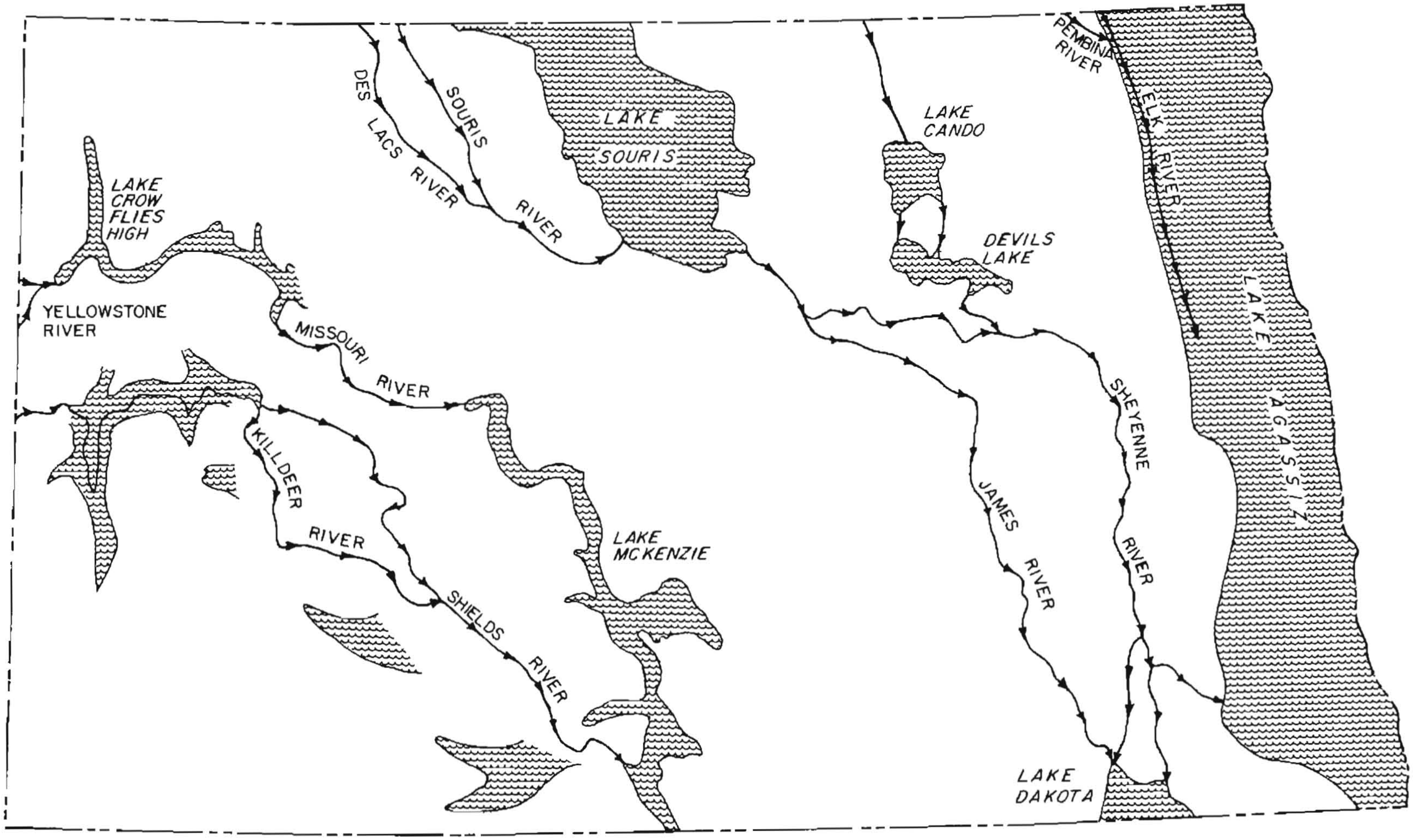

Figure 3. Glacial lakes (pattern) and major drainageways (arrows).

has been reinterpreted to be an esker. Few beaches have been identified around Lake Souris or Lake Dakota (fig. 3).

Offshore Sediment

Offshore sediment shown on the Geologic Map of North Dakota (Qcof, Qcoh, and Qcoe) consists of as much as 60 metres of laminated silt and clay deposited in glacier-dammed lakes. Most is irregularly laminated and is interpreted to be sediment deposited from underflows originating where meltwater rivers entered the lakes; unlike marine turbidity-current deposits, this sediment lacks graded rhythmites because it was deposited by continuous density currents originating from continuously flowing rivers, not by discontinuous turbidity currents originating from periodic submarine slumps. The clay occurs at the toes of the density-current fans in the deepest parts of the lake basins, and silt occurs farther up on the fans.

The upper parts of most density-current fans consist of well-sorted very fine to medium sand with flat bedding, small-scale cross bedding, and some large-scale cross bedding. This material has been included in the map unit "sand of the Oahe and older formations, undivided" (Qou) because it has everywhere been wind scoured and in many places is overlain by a thin layer of eolian sand deposited during Holocene time.

A small proportion of the offshore silt and clay is interpreted to be fallout from surface currents or interface currents above the lake bottom. In contrast to the meltwater deposits, it is predominantly clay (rather than silt or sand), it occurs draped over the lake-bottom topography in any part of the basin (rather than being restricted to fans plunging into the deepest parts of the basins), it has rhythmical graded bedding (rather than haphazard bedding), and it is typically only about a metre thick (rather than several metres or tens of metres thick). Much of it was probably deposited from density currents or wind-driven currents that were warmer, less turbid, and therefore less dense than the meltwater current because they started out as nonglacial rivers or originated in the shore zone. The rhythmites are probably varves; the currents stopped each winter because the rivers and lakes froze over, stopping the sediment supply. In contrast, large meltwater channels like the Pembina, Sheyenne, James, and Souris probably flowed throughout the year, producing cold, turbid, and heavy bottom currents.

On the Geologic Map of North Dakota, offshore deposits are subdivided into proglacial-lake deposits, ice-walled-lake deposits and collapsed supraglacial-lake deposits, and eroded lake deposits. Proglacial deposits consist of flat-bedded sediment of low, flat lake plains; they were deposited in lakes occupying valleys sloping toward the glacier. Ice-walled-lake deposits consist of flat-bedded sediment of flat lake plains elevated above the surrounding glacial topography; they were deposited in lakes bottomed on solid ground but surrounded by stagnant glacial ice. Collapsed supraglacial-lake deposits consist of folded and contorted sediment with hummocky topography; they were deposited in lakes bottomed on stagnant ice. Eroded lake deposits consist of sediment deposited, for the most part, in proglacial lakes and subsequently exposed in valley sides as a result of postdepositional erosion.

Proglacial-lake sediment (Qcof; "f" stands for "flat") has been mapped on the plains of Lake Agassiz, Lake Souris, Lake Dakota, and Lake Cando in southern Towner County and northwestern Ramsey County, and in a few other places around the state. All of these lakes, in their early stages, were supraglacial or ice walled. Collapsed supraglacial-lake deposits occur around all these lakes, and some of the higher beach ridges of Lake Agassiz were supraglacial and have been collapsed nearly beyond recognition. However, by the time the deposits indicated by the symbol Qcof were deposited, the lakes were essentially proglacial.

Ice-walled-lake deposits (fig. 4) and collapsed supraglacial-lake deposits have been grouped together on the Geologic Map (Qcoh; "h" stands for "high relief" or "hummocky") because they are closely associated and are gradational with each other. Their characteristics have been outlined by Clayton and Cherry (1967). Ice-walled-lake plains occur mostly on the Missouri Coteau and in the Turtle Mountains. Collapsed supraglacial-lake deposits also occur around the margins of Lake Agassiz, Lake Souris, Lake Dakota, and Lake Cando and probably in the valley of Apple Creek in Burleigh County.

Areas of eroded lake sediment (Qcoe) large enough to show on the Geologic Map occur near New Town in Mountrail County and along the Sheyenne River in Griggs County.

Ice-Drag Marks

The offshore sediment of Lake Agassiz is marked by low ridges and shallow grooves (figs. 5 and 6) that have been interpreted (Clayton and others, 1965) to be ice-drag marks. The grooves, which occur in only a few areas, are generally less than a metre deep and were probably gouged by the irregular bottoms of glacial icebergs. The ridges, which occur on most Agassiz offshore deposits, are generally less than a metre high, are typically about 100 metres wide, and are as much as 15 kilometres long. Where they cross, the earlier ridges are obliterated for a short distance on one side of the latest ridge, but are undisturbed on the other side, indicating that the ridge was plowed up at the side of a flat slab of lake ice where its diagonal edge touched the lake bottom. Only ridges and grooves longer than about 1 kilometre are shown on the map.

Compaction Ridges

Most of the compaction ridges (fig. 7) shown on the Geologic Map of North Dakota formed as a result of draping

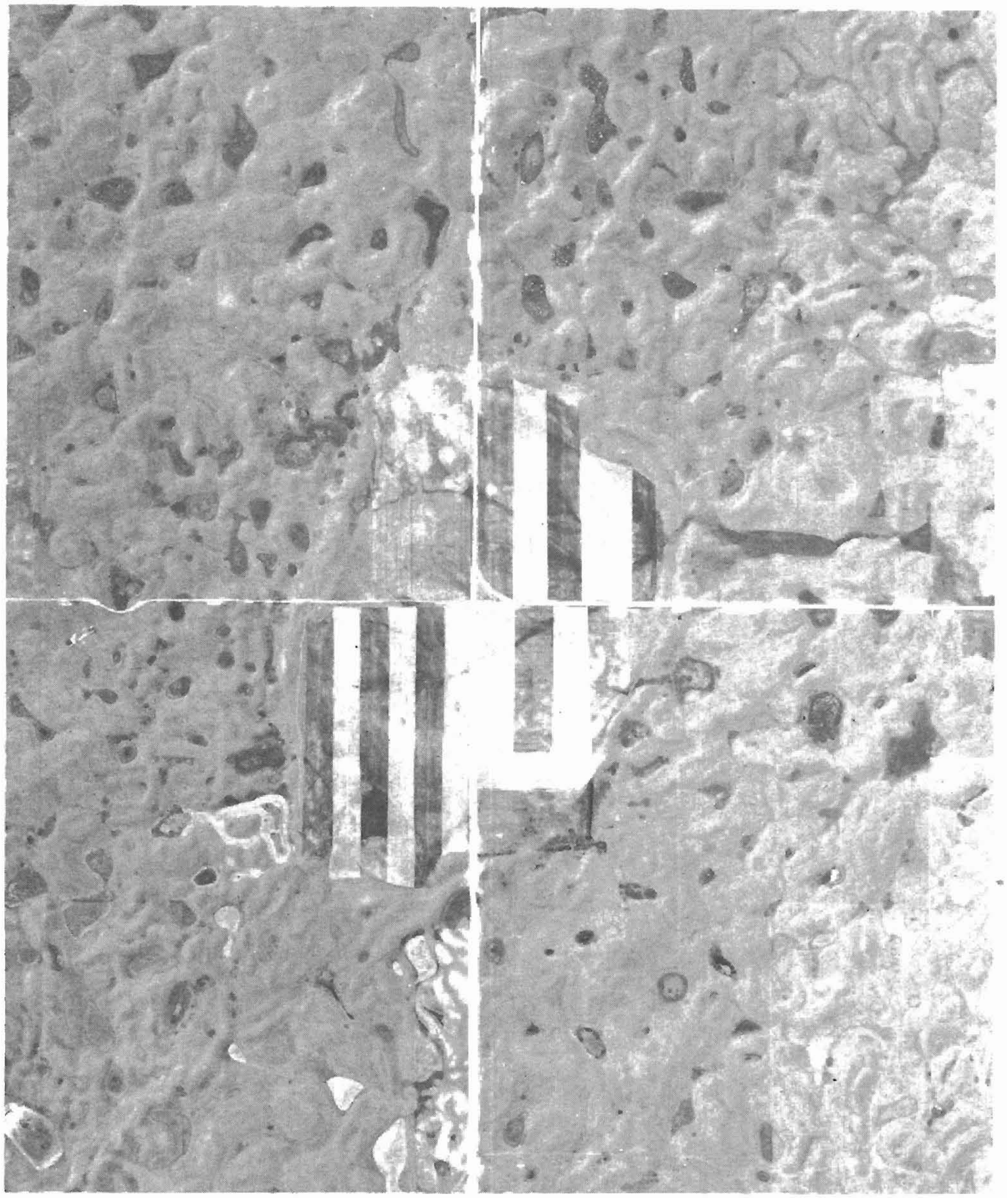

Figure 4. Ice-walled-lake plain (the smooth cultivated area) surrounded by collapsed glacial sediment (hummocky topography) in sections 1 and 12, T154N, R94W, and sections 6 and 7, T154N, R93W, Mountrail County. U.S. Department of Agriculture airphoto BAL-4V-45. Area shown is 2.6 kilometres wide. South is up.

and compaction of offshore sediment around underlying fluvial or beach sand and gravel bodies (Arndt, 1977, p. 7-8). The same symbol is used on the Geologic Map for low ridges in western and eastern Dickey County and northeastern LaMoure County that are interpreted to be palimpsest glacial ridges resulting from the draping of a thin layer of glacial sediment over older transverse glacial ridges. They are at right angles to the presumed last ice margin (fig. 32). Bluemle (1979b), however, has interpreted them to be transverse ridges of the last glacial advance and postulates a

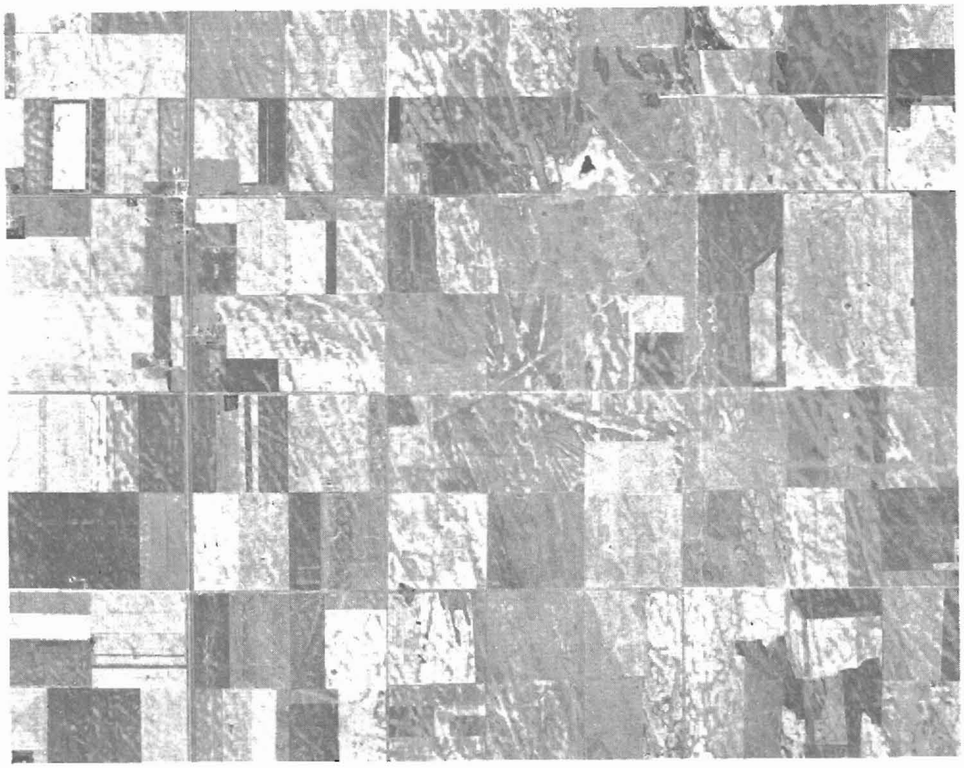

Figure 5. Ice-drag marks in T162N, Rs51 and 52W, Pembina County. Army Map Service airphoto vv-BE-M17-113-1976. Area shown is 8 kilometres wide. South is up.

readvance of the ice southeastward off the Missouri Coteau from a local ice cap.

Lake Agassiz

Lake Agassiz (Elson, 1967) occupied the eastern tenth of North Dakota (fig. 3), northwestern Minnesota, much of southern Manitoba, and parts of northwestern Ontario and central Saskatchewan for about 5,000 years in latest Wisconsinan time and earliest Holocene time (fig. 8). Evidence of Lake Agassiz indicated on the Geologic Map includes offshore sediment, ice-drag marks, shoreline sediment, beach ridges and scarps (figs. 9 to 11), and wave-eroded glacial sediment.

Map unit Qou includes sand of density-current fans with apexes at the mouths of the Sheyenne and Pembina meltwater channels. Deltas exist at the heads of the fans, but they are much smaller than mapped by Upham (1895). Horizontally bedded offshore sediment rather than slipface sediment is exposed in the supposed foreset faces, which are actually shore bluffs cut into the fans after the lake had dropped to the Campbell level. Lake Agassiz emptied south into the Minnesota and Mississippi Rivers when it was at higher levels (the Lockhart and Emerson Phases). It emptied east into Lake Nipigon and Lake Superior during lower levels (the Moorhead Phase and post-Emerson phases).

Figure 6. Ice-drag marks on plain of Lake Agassiz in T160N, R52W, in southeastern Pembina County. U.S. Geological Survey Glasston NE Quadrangle (5-foot contour interval). Area shown is 3.7 kilometres wide.

Lake Souris

Lake Souris occupied much of McHenry, Bottineau, and Pierce Counties (fig. 3) for no more than a few hundred years in Late Wisconsinan time (Lemke, 1960, p. 90-93). Evidence of Lake Souris indicated on the Geologic Map of North Dakota includes offshore sediment and wave-washed glacial sediment. Map unit Qou includes sand of density-current fans with apexes at the mouths of the Souris meltwater channel and several smaller

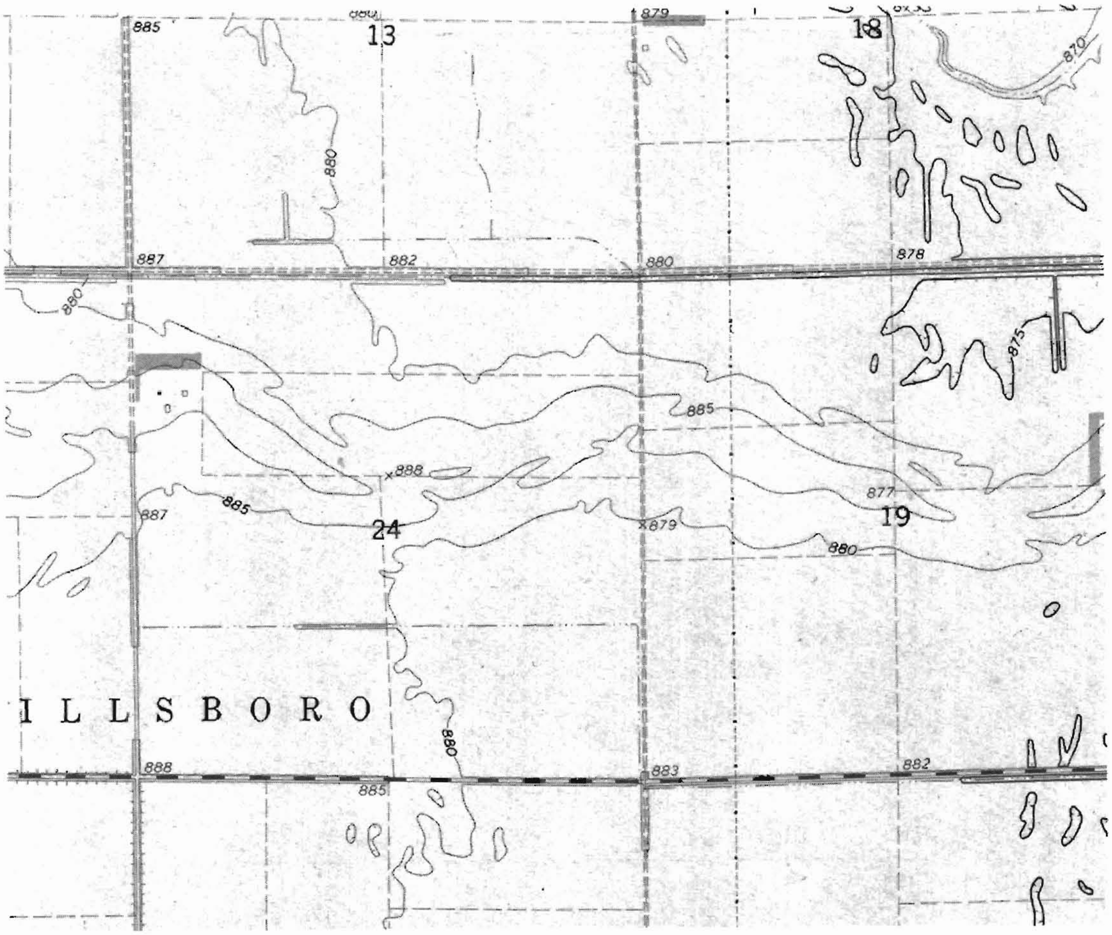

Figure 7. Compaction ridge, in T145N, Rs49 and 50W, in southeastern Traill County. U.S. Geological Survey Halstad SW Quadrangle (5-foot contour interval). Area shown is 3.6 kilometres wide.

meltwater channels with deltas on the west side of the lake. Lake Souris emptied into Lake Dakota through the James and Sheyenne meltwater channels and into Lake Agassiz through the Sheyenne and Pembina meltwater channels. It may also have briefly emptied around the north side of the Turtle Mountains through the Cando esker tunnel (fig. 3).

Lake Dakota

Lake Dakota (fig. 3) occupied parts of Dickey and Sargent Counties and much of the James valley in South Dakota for no more than a few hundred years in Late Wisconsinan time (Flint, 1955, p. 123-127). Evidence of Lake Dakota indicated on the Geologic Map of North Dakota includes offshore sediment (Qcof and Qcoh, which is largely coarse silt or sandy silt). Map unit Qou includes sand of density-current fans with their apexes at the mouths of the James and Sheyenne meltwater channels. Lake Dakota emptied southward into the Missouri River.

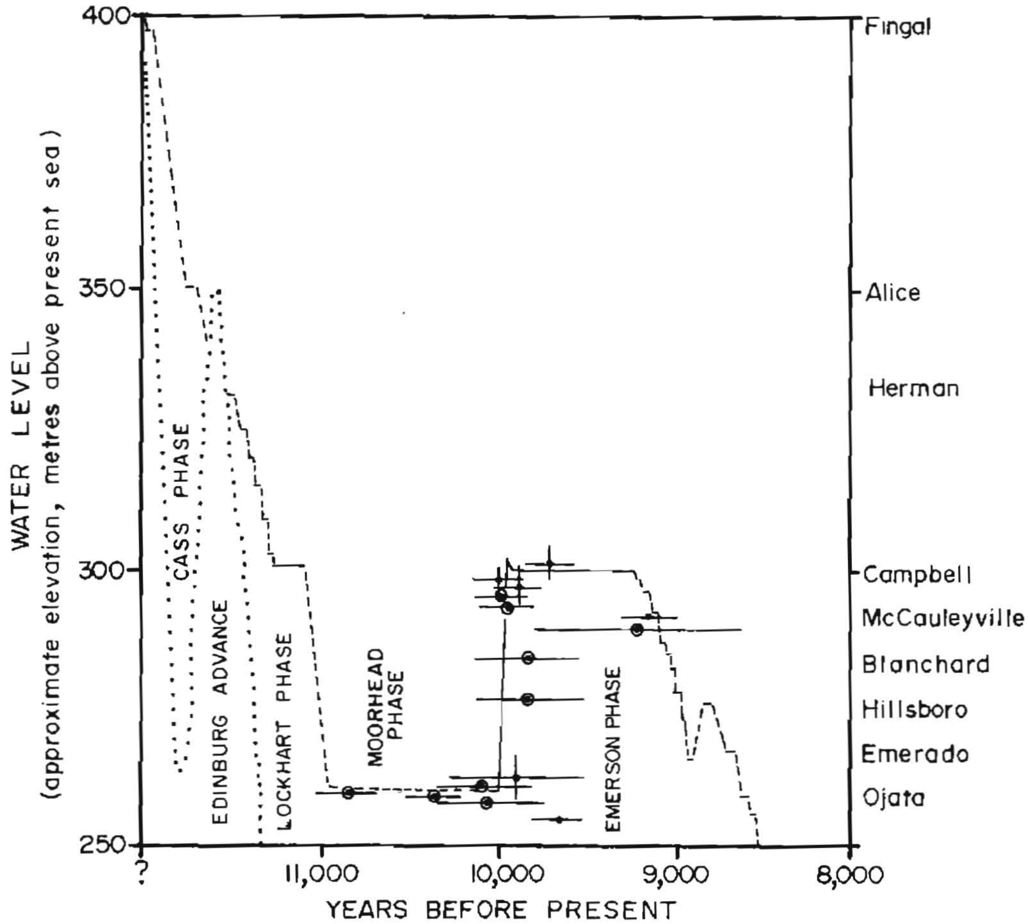

Figure 8. History of Lake Agassiz. From Elson (1967), Moran and others (1973), Bluemle (1974), and Arndt (1977). The heavy line represents the level of Lake Agassiz. The dotted line indicates active glaciation. The single dots indicate radiocarbon dates from wood, and the horizontal bars are standard deviations. Dates with a vertical bar are from nonlacustrine sediment. Those with a circle are from shore sediment. The others are from offshore sediment. On the right are named beaches. Correlation of the Herman, Campbell, and McCauleyville from their type areas in Minnesota is uncertain.

Lake Cando

Lake Cando (fig. 3) occupied parts of southern Towner County and northwestern Ramsey County for no more than a few hundred years in Late Wisconsinan time. Evidence of Lake Cando indicated on the map includes offshore sediment. The north end of the lake was occupied by a now-collapsed delta (map symbol Qcrh). During much of its life, the lake was bottomed on stagnant glacial ice. Lake Cando emptied southward, by an ice-contact channel, into Devils Lake.

Devils Lake

Devils Lake (fig. 3) has occupied parts of Ramsey and Benson Counties since the end of the Late Wisconsinan glaciation (Aronow, 1957). During that

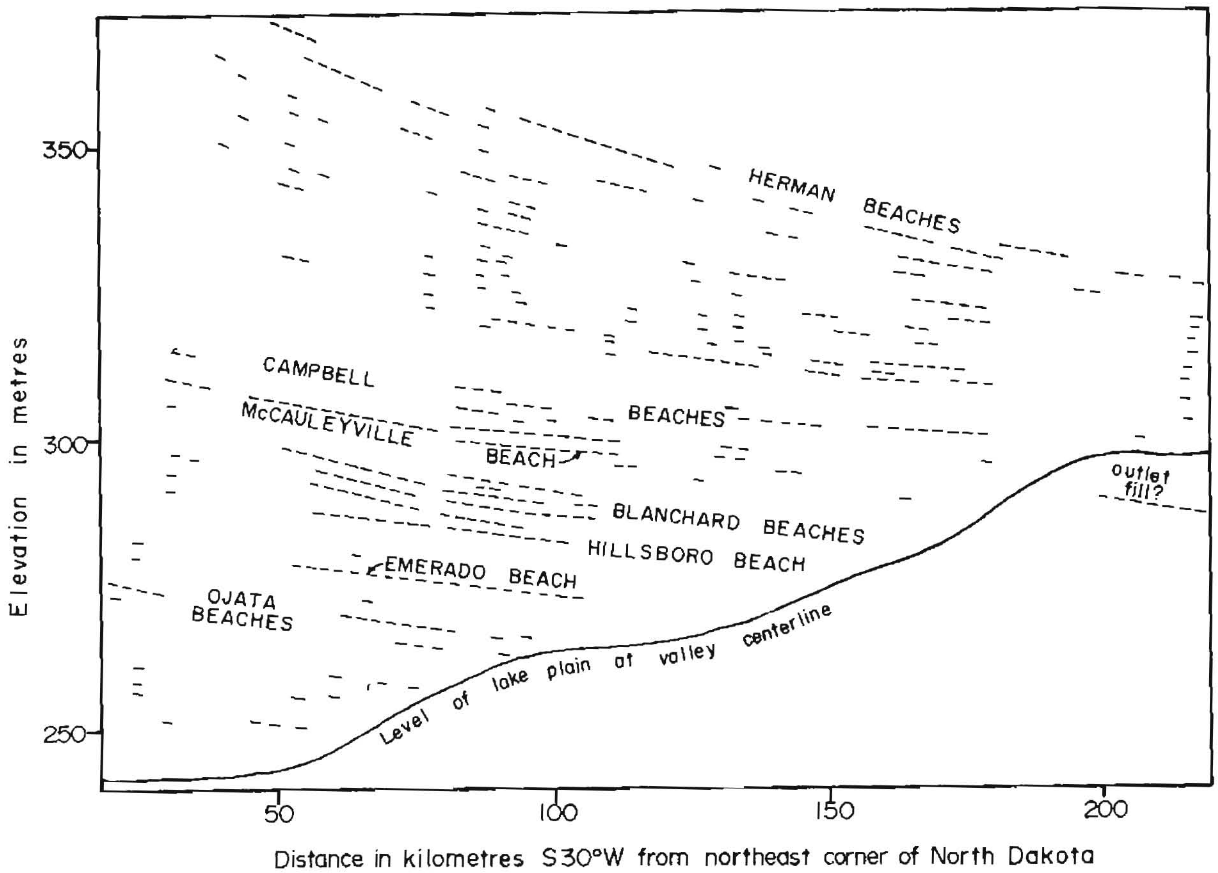

Figure 9. Elevation of beaches of Lake Agassiz in North Dakota. The line of the profile extends S30° W from the northeast corner of North Dakota, perpendicular to lines of equal glacial rebound. A Norcross Beach and a Tintah Beach are recognized between the Campbell and Herman Beaches in Minnesota, but they have not been definitely identified in North Dakota. The name "Hillsboro" is used even though the type "beach" is a fluvial deposit.

period, the lake level fluctuated about 16 metres. In Late Wisconsinan time, during its highest stages, the lake stood at a beach now at an elevation of 443 metres and emptied southward into the Sheyenne River through the spillway northeast of Warwick. During much of Holocene time the lake had no outlet. It reached a low of 427 metres in A.D. 1940 and has since been rising. Several stable periods during the Holocene are marked by beaches between 443 and 427 metres.

Aronow (1957, p. 415) and others have speculated that Devils Lake was connected in latest Wisconsinan time to Stump Lake to the east through an area now having elevations slightly below 443 metres. However, there seems to be no evidence of water flowing through this area, probably because surface elevations were then too high because of buried glacial ice.

Proglacial Lakes in the Southwest

Colton and others (1963) showed a proglacial lake in the Little Missouri Valley. Theoretically, lakes should have occurred in all the proglacial valleys in the area, including the Yellowstone, Little Missouri, Knife, Heart, and Cannonball Rivers and their tributaries (fig. 3), but no

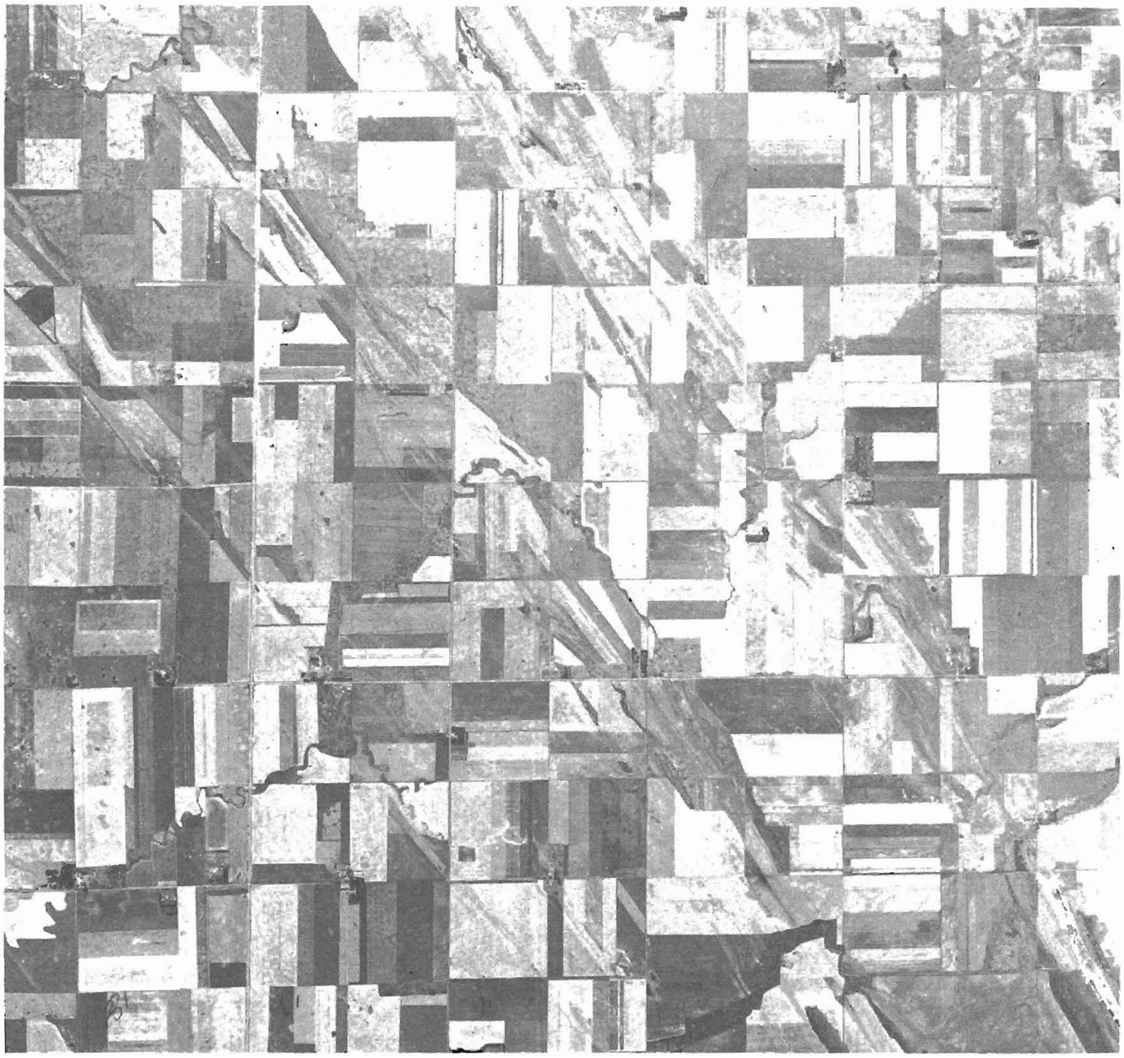

Figure 10. Beach ridges of Lake Agassiz in Ts150 and 151N, Rs52 and 53W, south-central Grand Forks County. Area shown is 9.5 kilometres wide (Army Map Service airphoto vv-BE-M9-113-1000). Starting at the lower-right corner are the closely spaced Campbell and McCauleyville Beaches, the three closely spaced Blanchard Beaches, the Hillsboro Beach, and in the lower-left corner the Emerado Beach. South is up.

beaches are known, and little offshore sediment has been found.

Evidence of two or more proglacial lakes along the Missouri River are shown on the Geologic Map of North Dakota. The offshore sediment in the New Town area of Mountrail County was deposited in Lake Crow Flies High when Late Wisconsinan ice dammed the part of the Missouri trench now occupied by the Van Hook arm of Lake Sakakawea (Clayton, 1972, p. 58); the narrow part of the present Missouri trench south of New Town was cut when the lake overtopped the lowest proglacial divide. The offshore sediment along Apple Creek in Burleigh County and between Beaver Creek and Horsehead Creek in Emmons County was probably deposited in Lake McKenzie when a Wisconsinan ice lobe blocked the Missouri River, perhaps

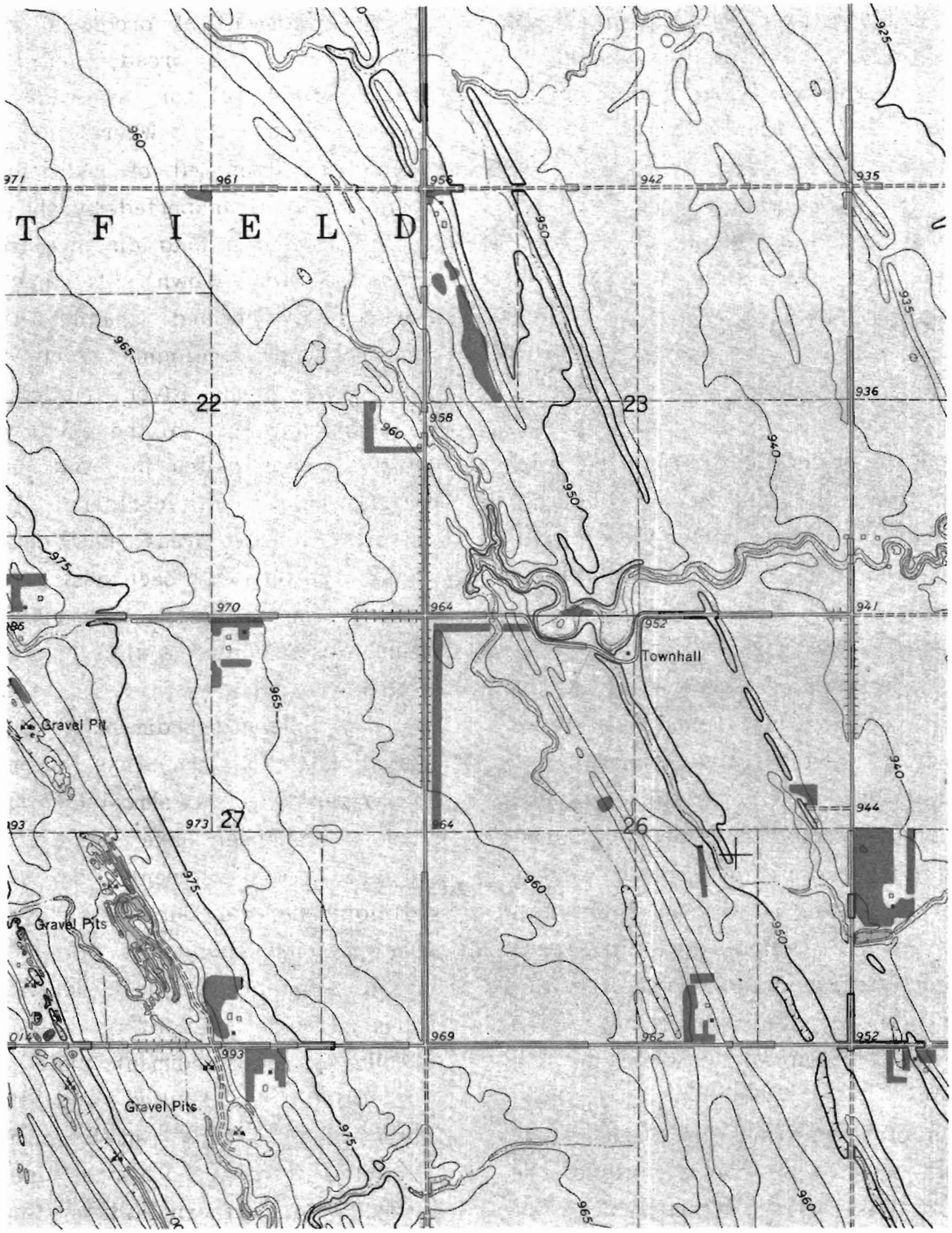

Figure 11. Beach ridges of Lake Agassiz in T153N, R54W, in northern Grand Forks County. U.S. Geological Survey Inkster SE Quadrangle (5-foot contour interval). Area shown is 3.7 kilometres wide. The gravel pits on the left are in the Campbell and McCauleyville Beaches; the Blanchard Beaches are between the 940 and 965 contours; and the Hillsboro Beach is at the 935 contour. This area is northwest of that shown in figure 10.

where the Missouri River flowed through the Strasburg area in Emmons County (Bickley, 1972, p. 117-120). The offshore sediment northwest of Fort Yates, Sioux County, was deposited in a proglacial lake resulting from a glacial dam to the south, probably in South Dakota.

Fluvial Features

River Sediment

River sediment in the Coleharbor Group consists of as much as 30 metres of moderately well-sorted sand with small-scale and large-scale cross bedding and poorly sorted gravel with plane bedding. It is largely channel sediment, with little or no overbank sediment.

The river sediment shown on the map includes the deposits of both meltwater rivers and nonmeltwater rivers. They were left undifferentiated because no consistent way of distinguishing them is known. Much of the material called "outwash" on previous maps was deposited by rivers consisting largely of normal runoff rather than meltwater. For example, the youngest "collapsed outwash" of the Missouri Coteau was deposited thousands of years after the glacier stagnated, when less than a tenth of the runoff was derived from melting ice (Clayton, 1967, p. 36, fig. A-7). Even the "outwash" deposited by some meltwater rivers is not really outwash. For example, the sand and gravel deposited by the Sheyenne meltwater river, after it left Lake Souris, was washed out of pre-existing glacial sediment in the river cutbanks, not out of the glacier.

The sediment of proglacial rivers (Qcrf) occurs as broad, flat fluvial plains (fig. 12) or as valley-side terraces (fig. 13). Where the flood plains are composed of sediment too coarse to be transported by the wind or are located a long distance from a source of wind-blown silt, they are marked by braided channel scars; examples can be found in the area where the Souris River emptied into Lake Souris and on the partly collapsed fluvial plains in front of the thrust ridges in northern Kidder County. In most areas, however, the channel pattern has been destroyed by wind erosion or, more commonly, has been hidden under a thin blanket of wind-blown silt.

The collapsed sediment of supraglacial rivers (Qcrh) has hummocky topography (fig. 14) almost identical to that of collapsed lake sediment and collapsed glacial sediment. They can be distinguished on airphotos by the sharper tonal boundaries on the collapsed sand and gravel. Lithologically, collapsed river sediment is identical to uncollapsed river sediment, except that its bedding is faulted and tilted. Normal fault scarps can be seen on airphotos of collapsed fluvial sediment in much of northern Kidder County.

Abandoned River Channels

Abandoned river channels have been indicated on the map only where discrete cutbanks are recognizable on

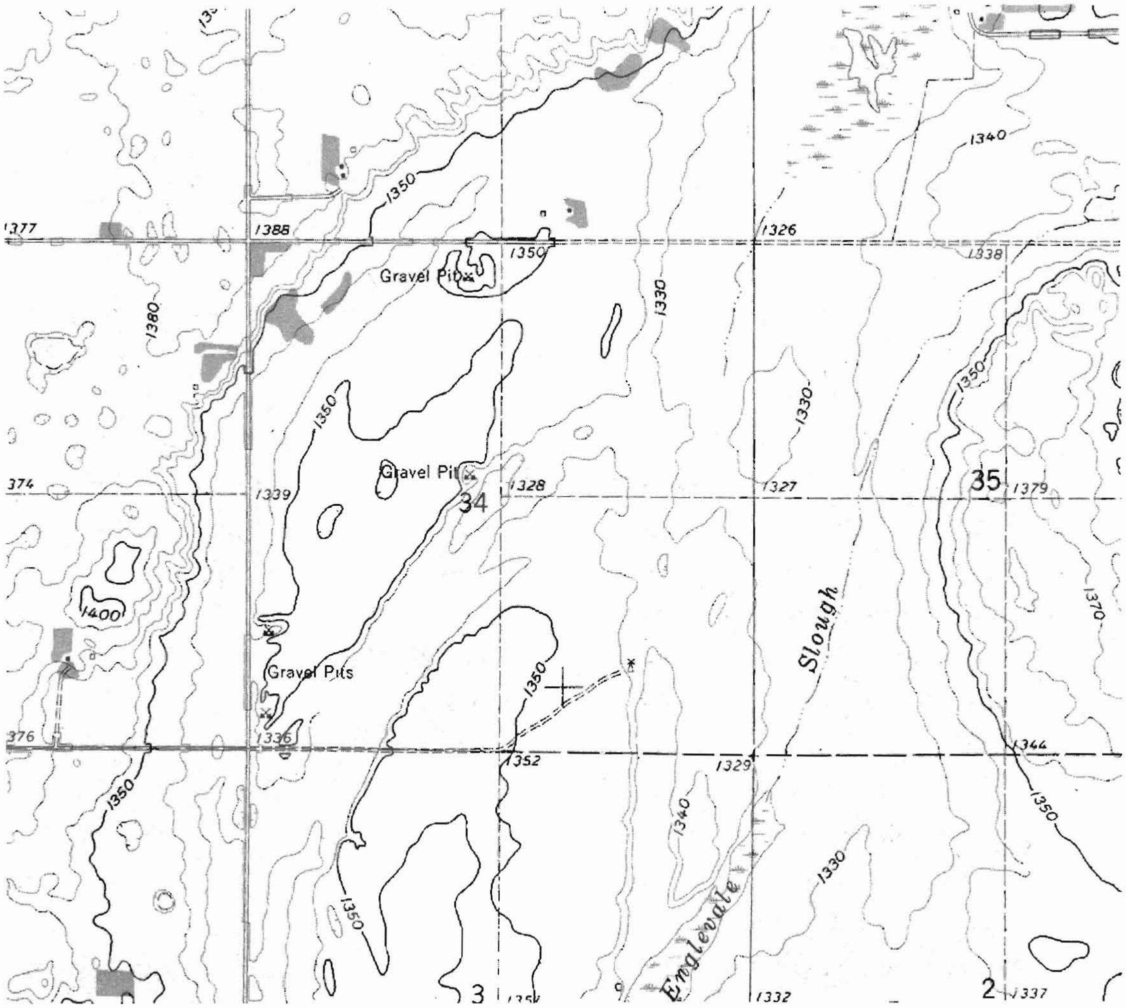

Figure 12. Uncollapsed river sediment (map unit Qcrf) bordered by glacial sediment above the 1355-foot contour in Ts134 and 135N, R58W, in western Ransom County. U.S. Geological Survey Englevale Quadrangle (10-foot contour interval). Area shown is 3.7 kilometres wide.

airphotos. For example, the symbol has been omitted along the Killdeer-Shields channel (fig. 3) where postglacial erosion near the major rivers has obliterated the cutbanks. Even though the Missouri River was a major meltwater channel in Wisconsinan time, no original cutbanks remain, probably because of postglacial erosion by the modern Missouri River. In contrast, the small modern James and Sheyenne Rivers (fig. 15) have had little effect on their meltwater channels.

The abandoned channels shown on the map are true channels, which were once filled bank to bank by flowing water. They are shaped like channels, with consistent bank heights, uniform channel widths, and regular meander shapes. They are unlike the Missouri trench and other nonchannels, which are merely valley bottoms, with sporadic cutbanks and irregular flood plains.

The Killdeer-Shields, Souris, James, Sheyenne, and Pembina melt-

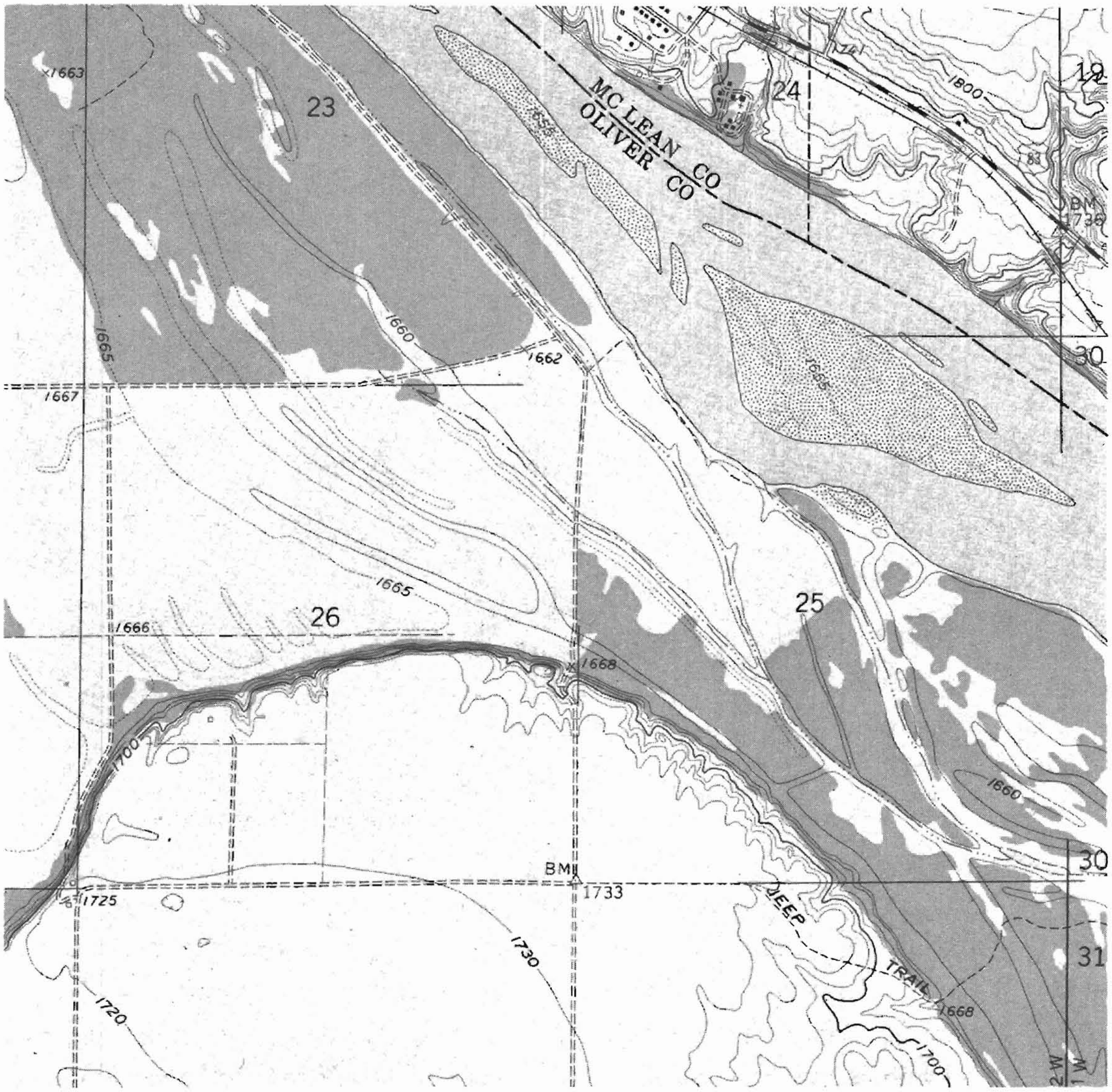

Figure 13. Uncollapsed river sediment (map unit Qcrf) of a terrace 22 metres above the modern flood plain of the Missouri River in T144N, Rs81 and 82W, in northeastern Oliver County. U.S. Geological Survey Washburn Quadrangle (10-foot contour interval). Area shown is 3.7 kilometres wide.

water channels are all about the size of the present lower Mississippi River channel, or about ten times as large as the present upper Missouri River channel. In Early Wisconsinan (?) time, the Killdeer-Shields River (fig. 3) drained nearly the same area as the present upper Missouri River, and in Late Wisconsinan time the Souris River drained an area of similar size in southern Alberta and Saskatchewan. Roughly a third or a half of the greater discharge during Late Wisconsinan time was the result of greater runoff during that time resulting from a mean annual temperature a few degrees cooler than present and a mean annual precipitation a few

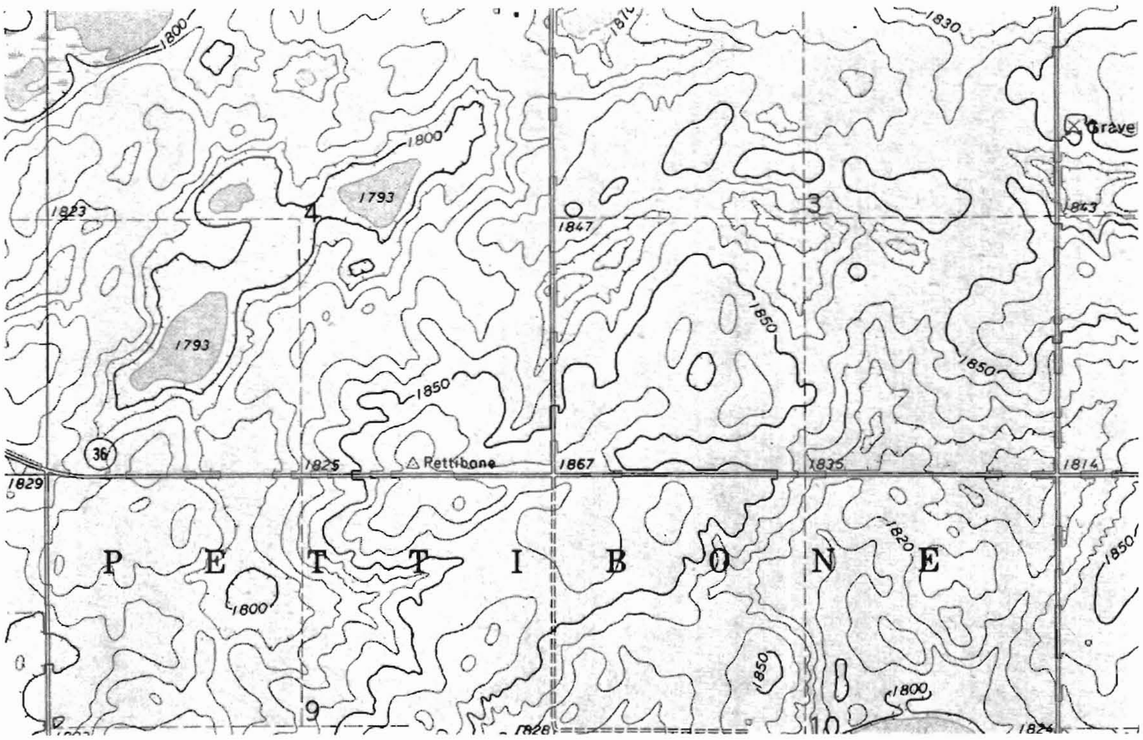

Figure 14. Collapsed river sediment (map unit Qcrh) in T143N, R70W, in northeastern Kidder County. U.S. Geological Survey Lake Williams Quadrangle (10-foot contour interval). Area shown is 3.7 kilometres wide.

hundred millimetres greater than present (Schumm, 1965, fig. 1). The remainder was glacial meltwater.

The abandoned channels indicated by blue line symbols on the Geologic Map of North Dakota were formed before the last glacial advance and were partly buried, but were not completely obliterated, by a thin layer of glacial sediment of the last advance.

Eskers

The eskers shown on the map were determined largely by airphoto interpretation. Few have been ground checked. However, because there are few other features with which they could be confused, the distribution of the more prominent eskers (figs. 16 and 17) is shown fairly accurately. Undoubtedly many of the less conspicuous eskers have been overlooked. Some of the eskers shown in northeastern Towner County and northeastern LaMoure County may have been misidentified: they may be transverse glacier-margin ridges or palimpsest features of an earlier advance.

Glacial Features

Glacial Sediment

Glacial sediment is the most widespread surface sediment in North Dakota (Moran and others, 1976). It consists of an unsorted mixture of

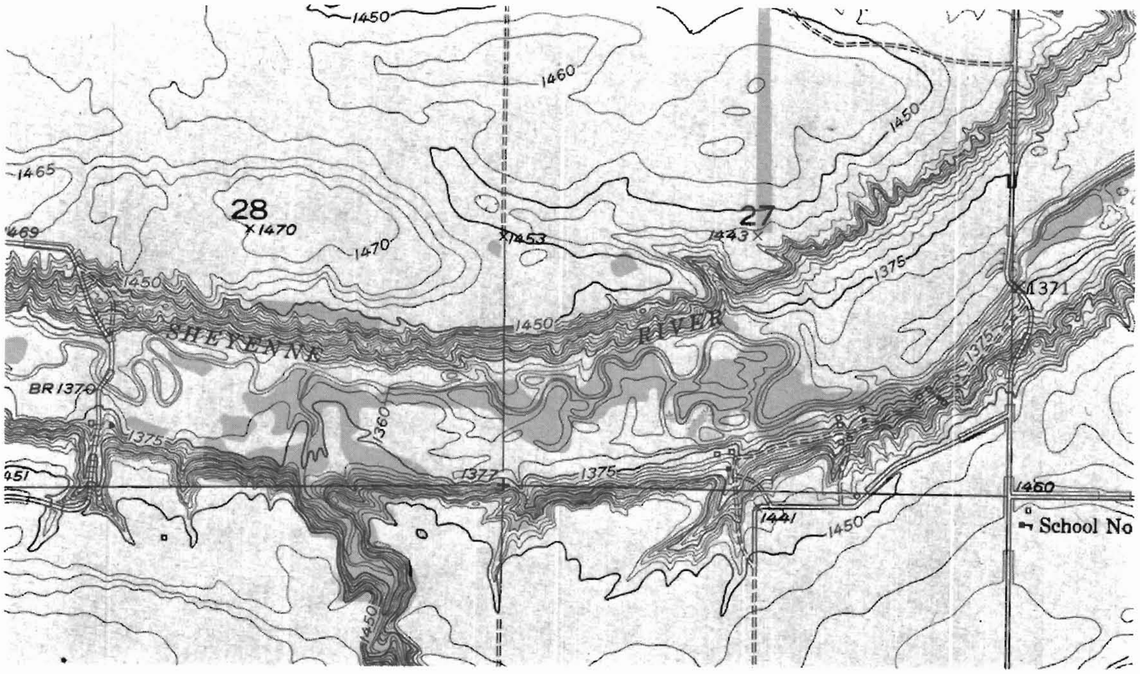

Figure 15. Large abandoned river channel (the Sheyenne spillway) in T150N, R61W, in southwestern Nelson County. U.S. Geological Survey Tolna Quadrangle (5-foot contour interval). Area shown is 3.7 kilometres wide.

clay, silt, sand, and gravel. It is typically 20% to 40% clay (grain diameter less than 0.004 millimetre), but some has as little as 15% and some has as much as 70% clay. It is typically 30% to 45% silt (0.004 to 0.063 millimetre), but some has as little as 20% and some has as much as 50% silt. It is typically 20% to 35% sand (0.063 to 2.0 millimetres), but some has as little as 5% and some has as much as 55% sand. It generally is roughly 5% pebbles (4.0 to 64 millimetres) and less than 1% cobbles (64 to 256 millimetres) and boulders (larger than 256 millimetres). The coarsest glacial sediment occurs closest to the Canadian Shield in northeastern North Dakota and where the glacier moved over fluvial sand and gravel. The finest occurs where the glacier moved over broad areas of lake sediment (in the Red River Valley, for example) or Cretaceous shale. Glacial sediment is one of the most homogeneous sediments; the proportion of any grain-size fraction commonly varies less than 5% along a 10-kilometre length of glacial flow line.

Mineralogically, the glacial sediment is similar to its source rock--the Precambrian igneous and metamorphic rock of the Canadian Shield; the lower Paleozoic limestone, dolomite, and sandstone fringing the Shield; the Cretaceous shale of eastern North Dakota, southwestern Manitoba, and southern Saskatchewan; and the Cretaceous and Paleocene continental and marine nearshore deposits of the Williston Basin in western North

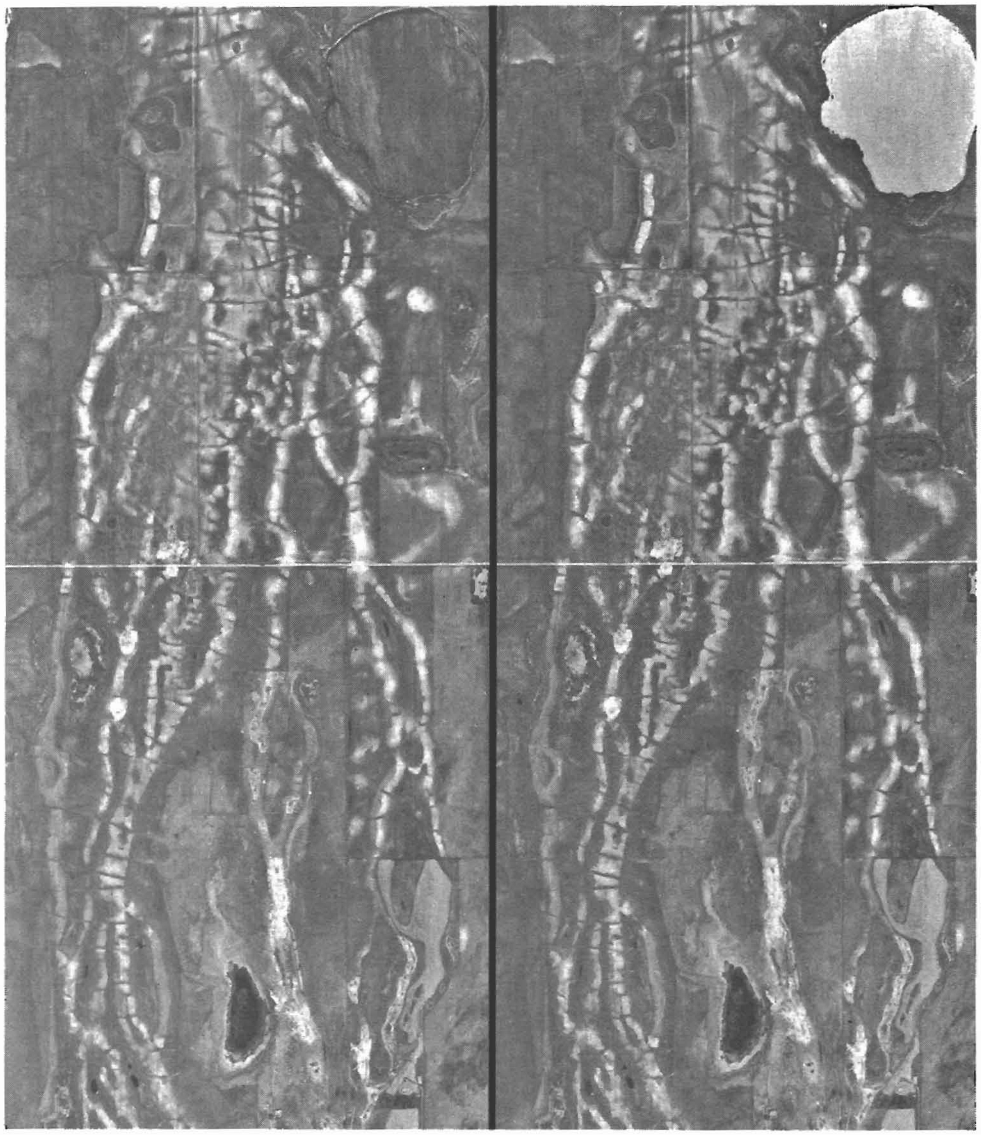

Figure 16. Esker in sections 28 and 33, T159N, R66W, Towner County. U.S. Department of Agriculture airphotos DOJ-1W-204 and -205. Area shown is 1.3 kilometres wide. The conspicuous narrow grooves are bison trails (Clayton, 1975). South is up.

Dakota. Nearly all the glacial sediment of North Dakota is calcareous and montmorillonitic. The silt and sand fraction consists of quartz, feldspar, and some limestone, dolomite, and shale. The gravel fragments consist in large part of limestone, dolomite, shale, granite, gneiss, and basalt.

Glacial Landforms

The Geologic Map of North Dakota differs most from the source maps in the way the glacial geology was handled. Our understanding of the glacial geology of North Dakota has evolved considerably during the past two

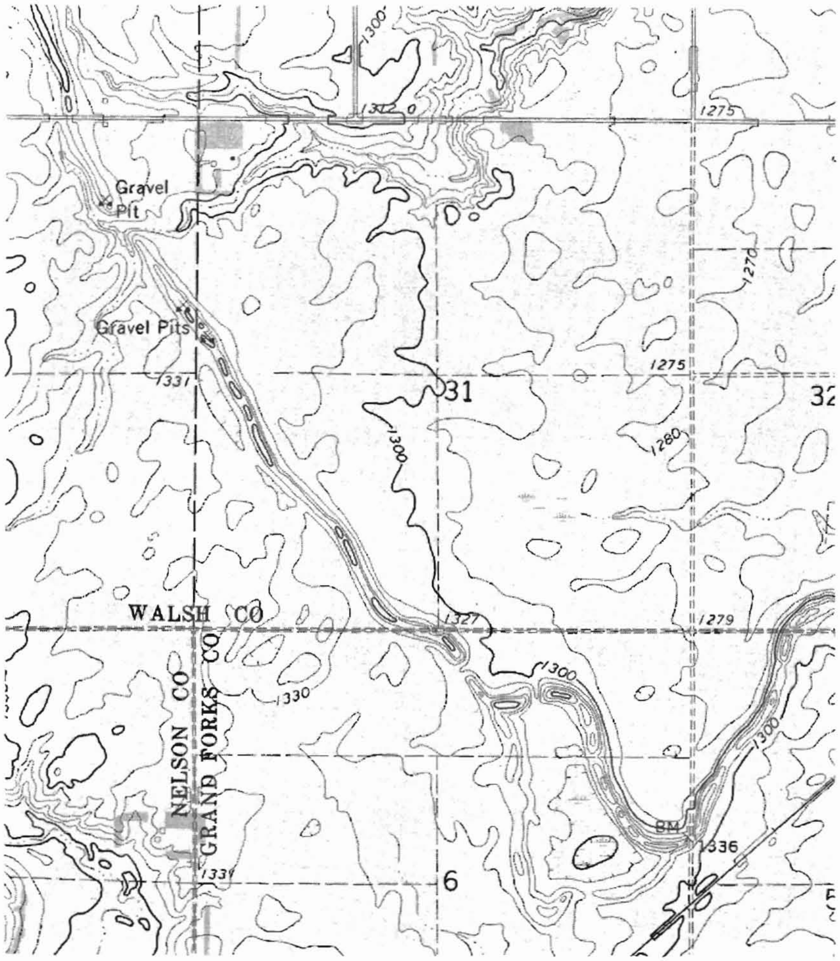

Figure 17. Esker in T155N, Rs56 and 57W, in the northwest corner of Grand Forks County and southern Walsh County. U.S. Geological Survey Dahlen Quadrangle (10-foot contour interval). Area shown is 2.7 kilometres wide. Surrounding material is gently undulating collapsed glacial sediment (map unit Qccg).

decades, and as a result the glacial geology shown on the map of one county commonly differs considerably from that shown on the map of an adjacent county. For this reason, the glacial landforms of the state had to be completely reevaluated.

After most of the 1:125,000 county maps had been completed, but before the map was begun, Clayton and Moran (1974) synthesized what was known about the glacial geology of the state, to serve as a foundation for the compilation of the map. During the first phase of compilation, however, an attempt was made to be as empirical as possible; we attempted to avoid superimposing a pre-existing genetic model on the legend.

Photos of every area with glacial sediment were first viewed stereoscopically (at a scale of 1:65,000). Contacts were drawn between all areas of glacial topography that looked different in some way. An attempt was made to disregard causes of these differences; a contact was drawn at every change in topography without regard to its significance.

A temporary name was assigned to each kind of topography. For example, a certain kind of glacial topography near the town of Bowdon was called "Bowdon moraine." This map unit was extended in all directions until a different kind of glacial topography was seen. It was identified in more distant areas by remembering what the type "Bowdon moraine" looked like; if it was poorly remembered, the memory was refreshed by returning to the "type" airphoto stereopair.

About fifty different glacial map units were identified. Some duplicated or partly duplicated other units, and the conceptual limits of some units drifted during the compilation period because the limits were not quantitatively defined. Quantification was avoided, however, because an attempt was being made to avoid superimposing a pre-existing model on the legend.

During the second phase of compilation, an attempt was made to isolate the visual criteria used to differentiate the map units. Some aspects were easily quantified, but others seemed impossible to quantify.

Finally, the fifty glacial map units were reduced to twelve units with three kinds of line-pattern overlay, the temporary names were eliminated, and a more permanent legend was constructed. The resulting legend is genetic where the interpretations seem fairly obvious, but purely descriptive elements that lack an interpretation were retained where the genetic element is obscure. For example, it seems obvious that hummocky topography resulted from the subsidence of supraglacial sediment, but the reasons for the presence or absence of ring-shaped hummocks are obscure.

The resulting legend has five major glacial subdivisions: (1) collapsed glacial sediment, (2) thin glacial sediment draped over topography existing before the last glacial advance, (3) thin glacial sediment draped over glacial thrust masses of the last advance, (4) thin glacial sediment draped over subglacially molded surfaces of the last advance, and (5) glacial sediment with nonglacial topography resulting from postglacial erosion, plus an additional subdivision that is transitional between the collapsed and draped subdivisions.

Collapsed Glacial Sediment

Hummocky glacial topography (fig. 18) is interpreted to be the result of the lateral movement of supraglacial sediment as it subsides (collapses; is let down) when the underlying ice melts out from under it (Clayton and Moran, 1974; Clayton, 1967). Although this is the generally accepted explanation, two additional alternatives have been suggested. Stalker (1960) has suggested that hummocks resulted from the squeezing of subglacial sediment into irregularities in the base of a stagnant glacier; however, hummocks composed of glacial sediment are essentially identical to hummocks composed of collapsed supraglacial fluvial and lacustrine sediment that lacks evidence of ever having been under a glacier. Bik (1967) has sug-

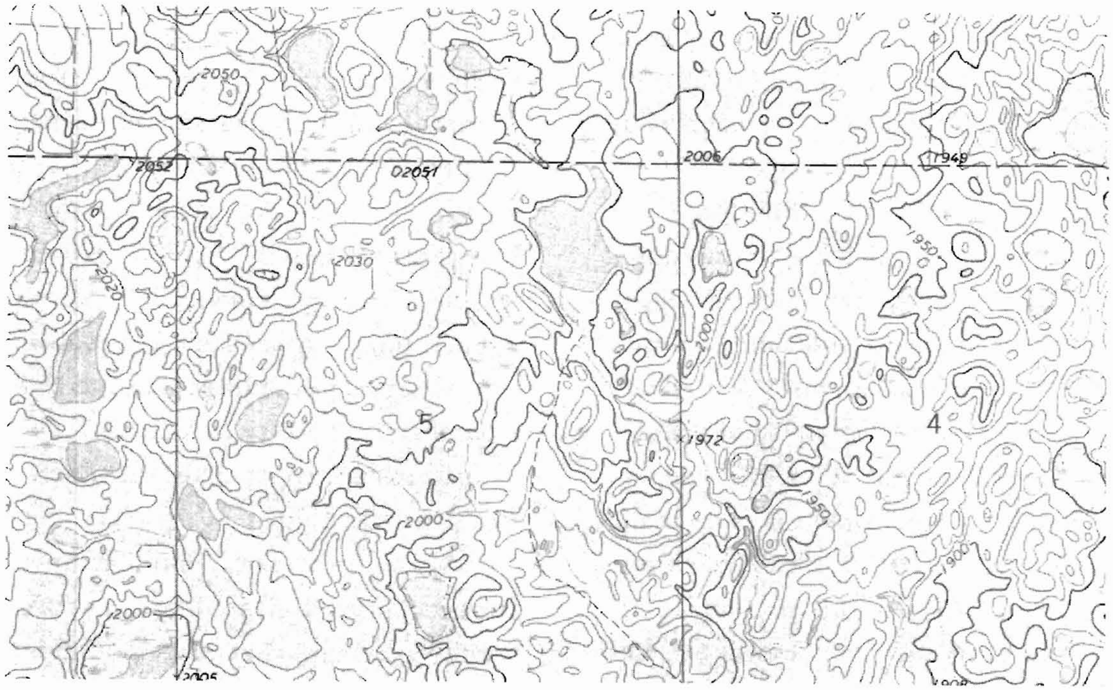

Figure 18. Hilly collapsed glacial sediment (map unit Qcch) in Ts131 and 132N, R65W, in southwestern Dickey County. U.S. Geological Survey Merricourt Quadrangle (10-foot contour interval). Area shown is 3.6 kilometres wide.

gested that hummocks resulted from the movement of sediment during the growth and decay of permafrost; however, in North Dakota, hummocks were generally formed at a time when paleoecologic evidence indicates a climate too warm for permafrost, and hummocks are generally absent in North Dakota in areas known to have had permafrost.

On the map, hummocky collapse topography (Qccg, Qccu, Qccr, and Qcch) has been subdivided on the basis of its three most conspicuous variables: slope angles, presence or absence of ring-shaped hummocks, and presence or absence of transverse ridges.

Four slope categories were recognized, based on the average maximum slope angle: Qccg, gently undulating (1° to 2°); Qccu, undulating (2° to 4°); Qccr, rolling (4° to 8°); and Qcch, hilly (8° to 15°). It was found that slope angle could be quickly estimated with fair precision either in the field or using airphoto stereopairs, and slope angles are shown on the 1:125,000 soil maps for every county (Patterson and others, 1968, p. 3). Slope angle was emphasized on the Geologic Map of North Dakota because it is one of the more obvious variables and because it has genetic significance--slope angle is related to the fluidity of the supraglacial sediment at the time of deposition. Fluidity was controlled in part by the grain-size distribution of the glacial sediment; hummocks composed of clayey glacial sediment or offshore lake sediment tend to have more gentle slopes than those composed of sandy glacial sediment or fluvial sand and gravel. Fluidity was probably controlled in part by the mineralogy of the clay-size fraction; the abundance of montmorillonite resulted in relatively fluid supraglacial sediment. Fluidity was most strongly controlled by the water content of the sediment; gentle slopes result from wet supraglacial sediment and steep slopes result from dry supraglacial sediment. The water content of the sediment was probably controlled by the rate of ice melt, which was controlled by the thickness of the insulating blanket of supraglacial sediment; gentle slopes are the result of thin supraglacial sediment and steep slopes are the result of thick supraglacial sediment. Although the thickness of the sediment of the last glacial advance is poorly known in most areas and the distinction between supraglacial and subglacial sediment is obscure, the little information that is available suggests that the thickness of the supraglacial sediment in metres roughly equals the maximum slope angles in degrees. For example, rolling hummocky topography, with 4° to 8° maximum slope angles, is the result of the collapse of supraglacial sediment between about 4 and 8 metres thick (Clayton and Moran, 1974, p. 114-115).

Similarly, local relief (hummock height) expressed in metres is nearly the same as the maximum slope angle expressed in degrees in most areas. For example, in areas of rolling collapse topography (Qccr), where the maximum slope angles are 4° to 8° (and where the supraglacial sediment was roughly 4 to 8 metres thick), the local relief is about 4 to 8 metres.

Three general hummock shapes were recognized. (1) Most commonly, hummocks are simple mounds (fig. 19). A profile in any direction across a series of adjacent mounds and their intervening swales tends to be the shape of a sine curve. (2) Many have their tops depressed, resulting in ring-shaped (doughnut-shaped) hummocks (fig. 20). (3) Many are elongated in a transverse direction (fig. 21).

The presence of ring-shaped hummocks is indicated on the map by a circle pattern. Because mound-shaped and ring-shaped hummocks are completely gradational with each other, a boundary between them is difficult to precisely determine; ring-shaped hummocks are indicated where they are conspicuous on 1:65,000 stereopairs. Ring-shaped hummocks are most abundant in undulating collapse topography (Qccu) but are also common in gently undulating (Qccg) and rolling (Qccr) topography. Although ring-shaped hummocks are generally thought to be the result of the incomplete filling of supraglacial sinkholes by mudflow (Clayton and Moran, 1974, p. 107, fig. 6G), the reasons for their presence or absence in any area are unknown.

The crests of transverse glacial ridges are indicated by a line pattern on the Geologic Map. Those shown on collapsed glacial sediment (Qccg, Qccu, Qccr, Qcch, and Qcdc) and most of those shown on glacial sediment draped over pre-existing topography (Qcdg and Qcdn) are transversely elongated

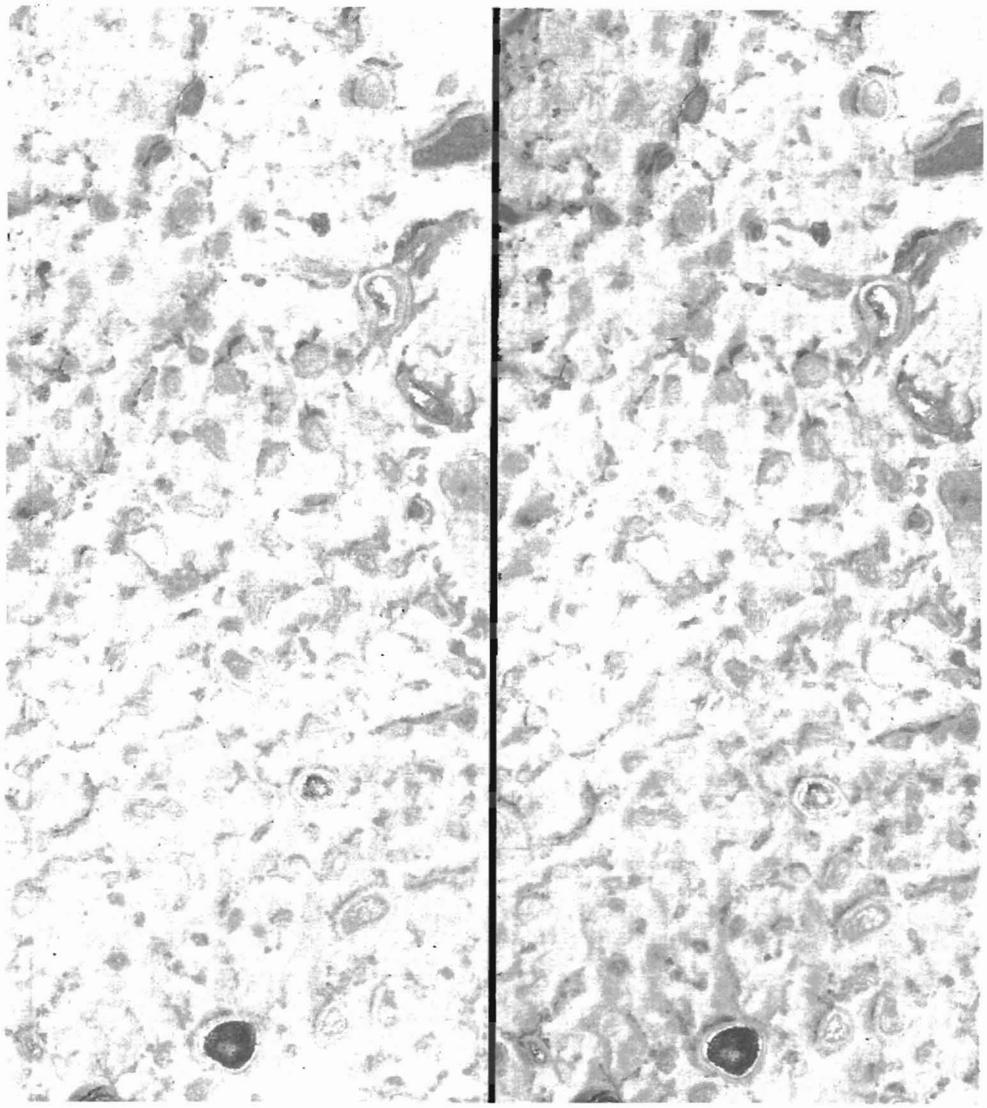

Figure 19. Simple mound-like hummocks (map unit Qcch) in sections 6 and 7, T157N, R89W, Mountrail County. U.S. Department of Agriculture airphotos BAL-6V-90 and -91. Area shown is 1.3 kilometres wide. South is up.

hummocks. Collapsed glacial sediment with transversely elongated hummocks has in the past been called "washboard moraine," "washboard moraines," "washboard ridges," and "minor moraines" (Clayton and Moran, 1974, p. 105, fig. 5; Elson, 1968, type B; Gwynne, 1951). Transversely elongated hummocks are generally considered to have formed where transversely elongated concentrations of englacial sediment were lowered to the ground as the underlying ice melted; they are of the same origin as mound-like and ring-shaped hummocks except that the glacial sediment was anisotropically distributed. Although it is generally agreed that the transverse concentrations of englacial sediment resulted from the shearing of subglacial sedi-

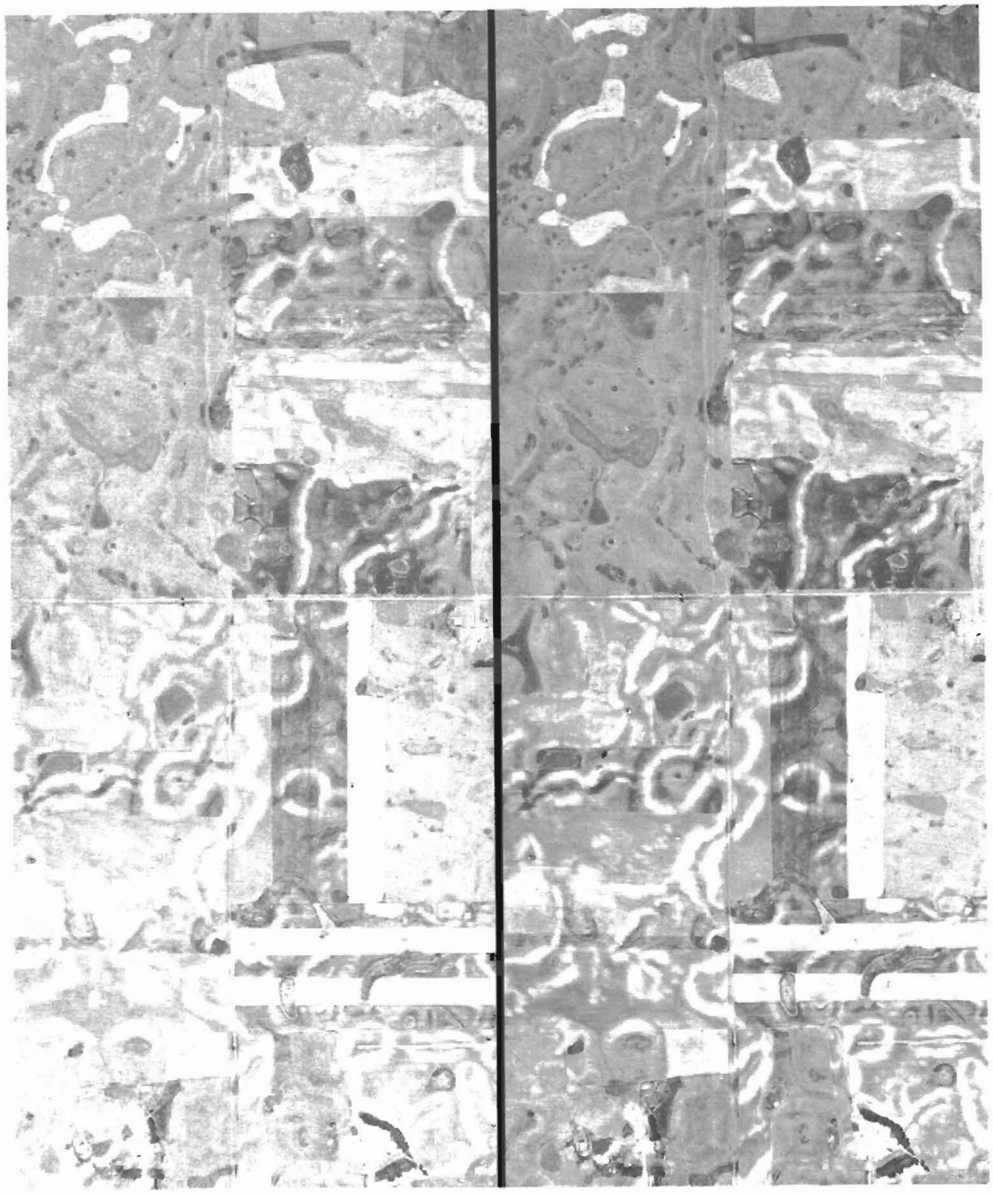

Figure 20. Ring-shaped hummocks (map unit Qccu) in section 25, T155N, R91W, Mountrail County. U.S. Department of Agriculture airphotos BAL-4V-107 and -108. Area shown is 1.3 kilometres wide. South is up.

ment into the ice just behind the glacial terminus, the cause of the 200-metre spacing of the concentrations is unknown; the suggestion that each concentration represents a winter readvance of the terminus seems unlikely because the radiocarbon chronology suggests that the active-ice terminus in these areas retreated faster than 200 metres a year. Because transversely elongated hummocks are completely gradational with mound-like and ring-shaped hummocks, a boundary between them is difficult to precisely determine; transversely elongated hummocks are indicated on the map

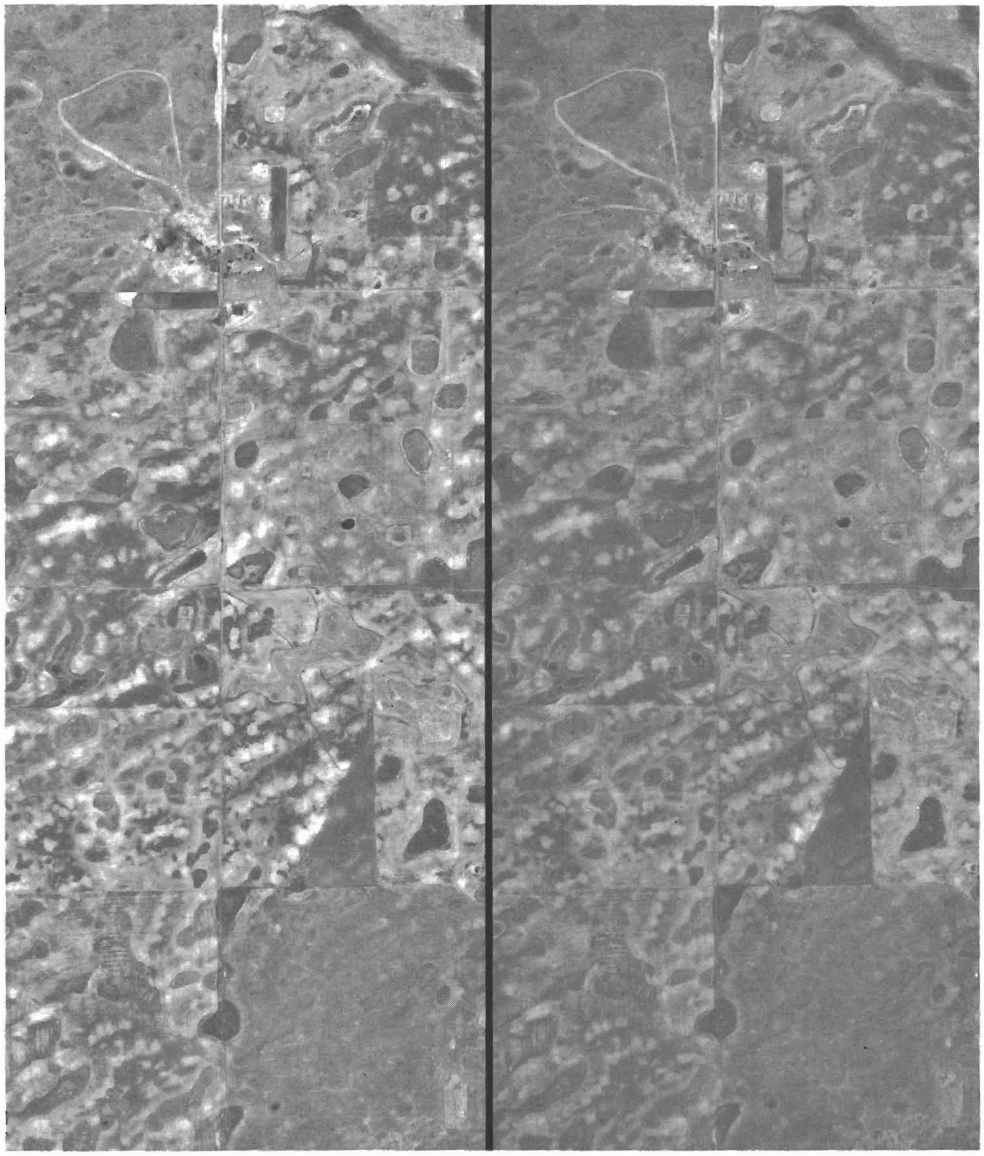

Figure 21. Transversely elongated hummocks (map unit Qccu) in sections 4, 5, 8, and 9, T153N, R60W, Nelson County. U.S. Department of Agriculture airphotos CWM-3W-108 and -109. Area shown is 1.3 kilometres wide. Glacier came from northwest (lower right).

where they can be distinguished on 1:65,000 stereopairs. Transversely elongated hummocks commonly occur in gently undulating (Qccg) and undulating (Qccu) collapse topography, and they rarely occur in rolling (Qccr) collapse topography.

The transverse glacial ridges indicated on the glacial sediment draped over nonglacial topography (Qcdn) in northwestern McKenzie County (Colton, 1958) are probably push ridges (Elson, 1968, type A) rather than transversely elongated hummocks. Unlike the hummocks, which have cross profiles shaped like a sine curve, the push ridges are widely spaced, with flat areas between them.

Two other variables were noted but were not directly incorporated into the legend: hummock spacing (diameter) and local relief.

The centers of adjacent hummocks are typically about 200 metres apart versely elongated hummocks (the scale of the Geologic Map is too small to show every transversely elongated hummock, and as a result the spacing shown on the map is considerably greater than 200 metres). Hummock spacing seems to increase slightly with increasing slope angle, but this has not been proven quantitatively. Clayton (1967, p. 31) and Clayton and Moran (1974, p. 107-108, fig. 7) argued that the 200-metre spacing of hummocks was inherited from the 200-metre spacing typical of sinkholes in stagnant glaciers.

In many areas, smaller hummocks spaced about 50 metres apart are superimposed on top of the larger hummocks. In a few areas this seemed to be the predominant hummock spacing. No explanation for this spacing is known.

Collapsed/Draped Transition

Map unit Qcdc is transitional between map units Qccg, Qccu, Qccr, and Qcch (collapsed glacial sediment) and map units Qcdg and Qcdn (glacial sediment draped over pre-existing topography). It has hummocky topography, interpreted to be collapse topography formed during the last glacial advance into the area; and it also has other topographic elements, larger in scale than the hummocks, interpreted to be topography formed before the last advance but not completely obliterated during the last advance (fig. 22). Because this interpretation depends on the recognition of the pre-existing topography, which is sometimes difficult, and because this map unit is completely gradational with collapse topography and draped topography, the contact between these units is imprecisely placed in many areas.

Draped Topography

Map units Qcdg and Qcdn consist of thin glacial sediment draped over and only slightly modifying the topography, either glacial (Qcdg) or nonglacial (Qcdn), existing before the last glacial advance into the area. In many areas, outcrops show that the glacial sediment of the last advance is less than 1 metre thick, and in many areas it is probably as thin as 0.1 metre and is nearly unrecognizable because of modification by soil-forming processes. In some areas, where it is thicker than 1 metre and where the pre-existing topography was rugged, draped topography (Qcdg or Qcdn) has been mapped rather than collapse topography (Qccg, Qccn, Qccr, Qcch, or Qcdg) because the pre-existing topography has been only slightly modified. In other areas, where the glacial sediment is thinner than 1 metre and where the pre-existing topography was nearly flat, collapse topography rather than draped topography has been mapped because the pre-existing

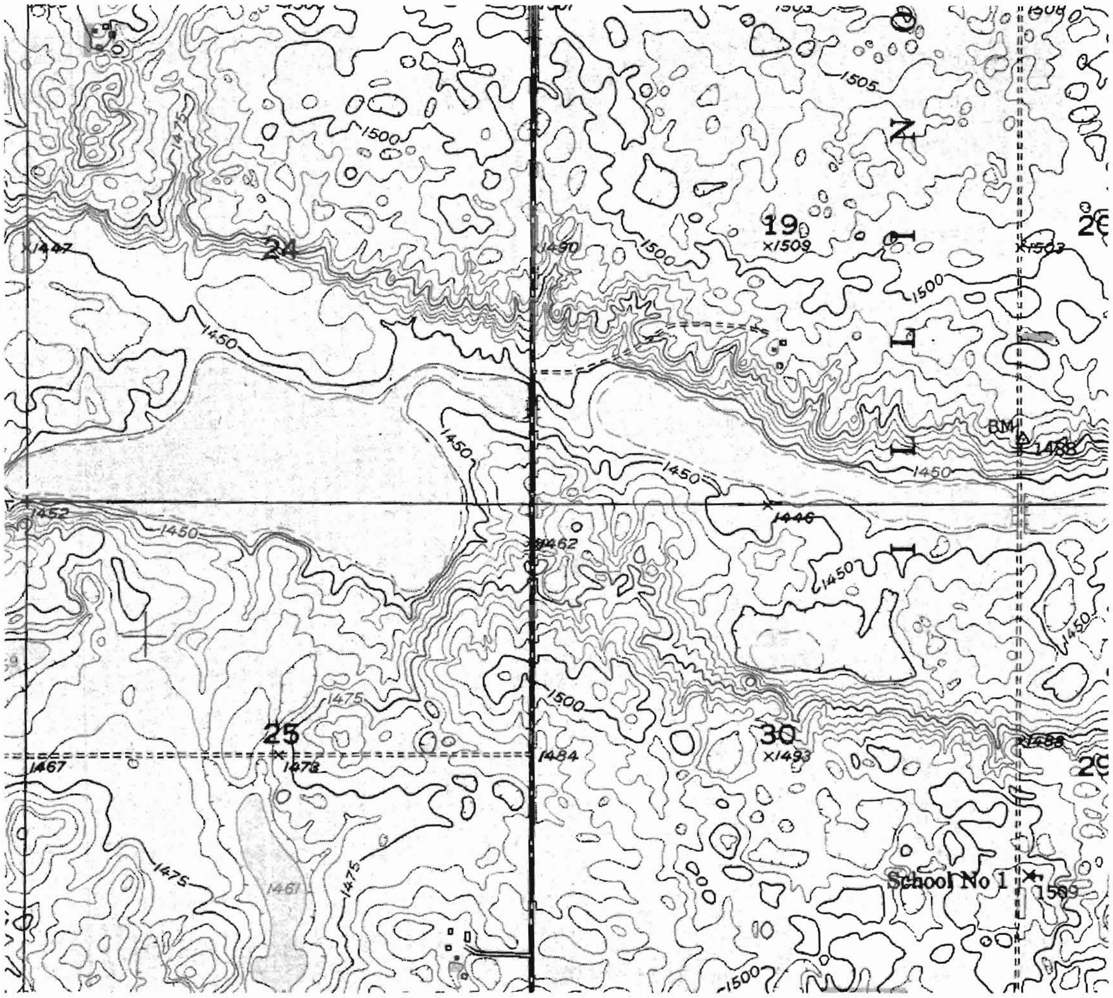

Figure 22. Hummocky collapsed glacial sediment draped over pre-existing channel (map unit Qcdc) between Devils Lake and Stump Lake (T152N, Rs61 and 62W). U.S. Geological Survey Devils Lake Mountain Quadrangle (5-foot contour interval). Area shown is 3.6 kilometres wide.

topography has been largely obliterated. Previously, draped topography has been called "sheet moraine" (Kume and Hansen, 1965, p. 15), "palimpsest moraine," or "palimpsest topography."

Glacial sediment draped over pre-existing nonglacial topography (Qcdn) has been mapped in areas where there was clearly a nonglacial topography with an integrated drainage pattern, for the most part southwest of the Missouri Coteau (fig. 23). On the ground, this topography looks much like the nonglacial topography found in gently undulating to rolling upland areas beyond the limit of glaciation. Much of the area mapped as Qcdn has undergone enough postglacial erosion to largely obliterate any hummocky topography that was originally present. The topography, however, is known to be largely the result of preglacial rather than postglacial processes because a uniformly thin layer of glacial sediment of the last advance is draped over the land-

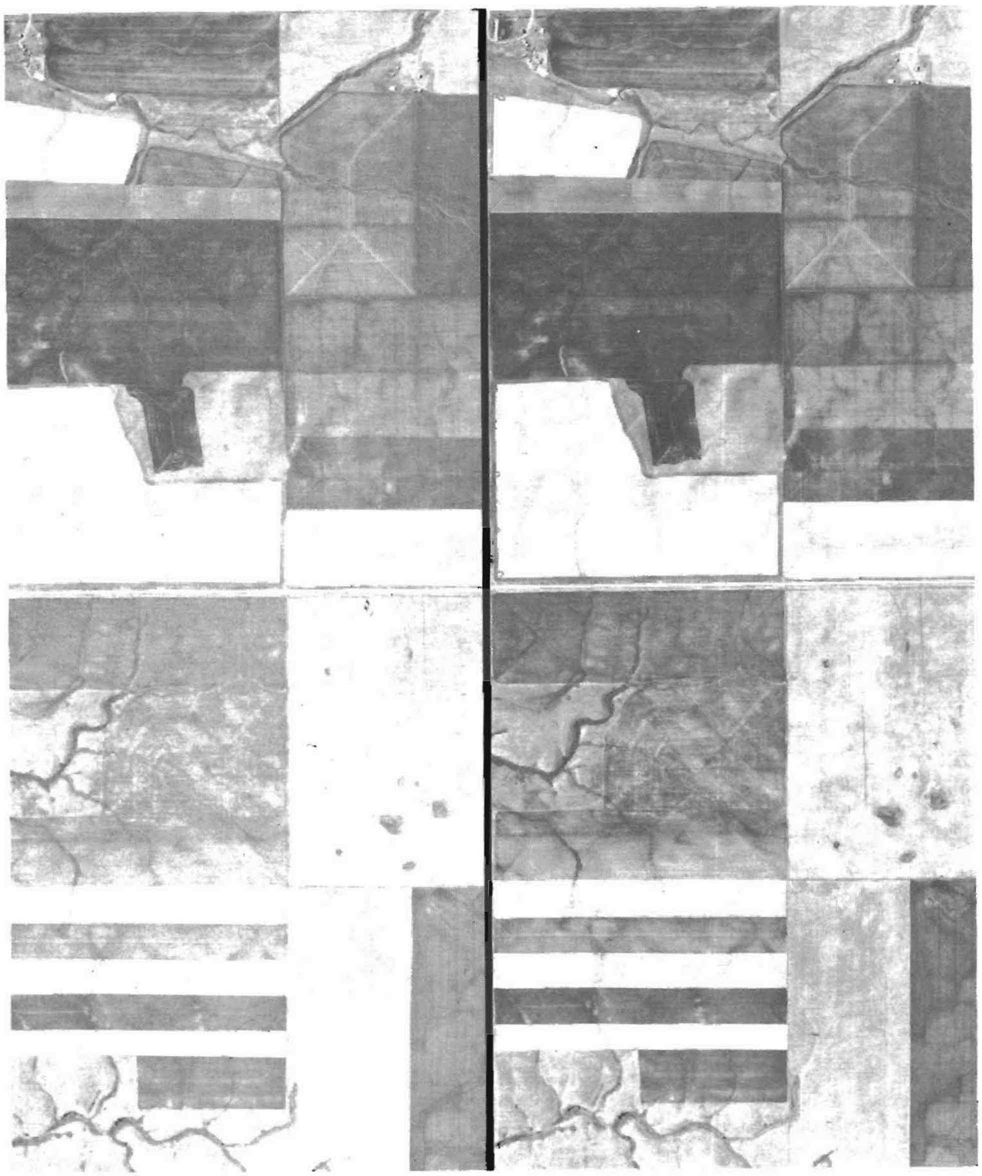

Figure 23. Thin glacial sediment draped over pre-existing nonglacial topography (map unit Qcdn) in Ts151 and 152N, Rs92 and 93W, in southern Mountrail County. U.S. Department of Agriculture airphotos CWM-3W-108 and -109. Area shown is 1.3 kilometres wide. South is up.

scape.

Glacial sediment draped over pre-existing glacial topography (Qcdg) has been mapped in areas where pre-existing collapsed glacial sediment, collapsed fluvial sediment, collapsed lacustrine sediment, or glacial thrust masses (fig. 24) are interpreted to have been only slightly modified during the last glacial advance. Because of the lack of good stratigraphic information in many areas, this was generally a difficult interpretation to make.

In some areas, where previous maps showed collapsed fluvial sand and gravel or lacustrine silt and clay at the surface, Qcdg has been mapped

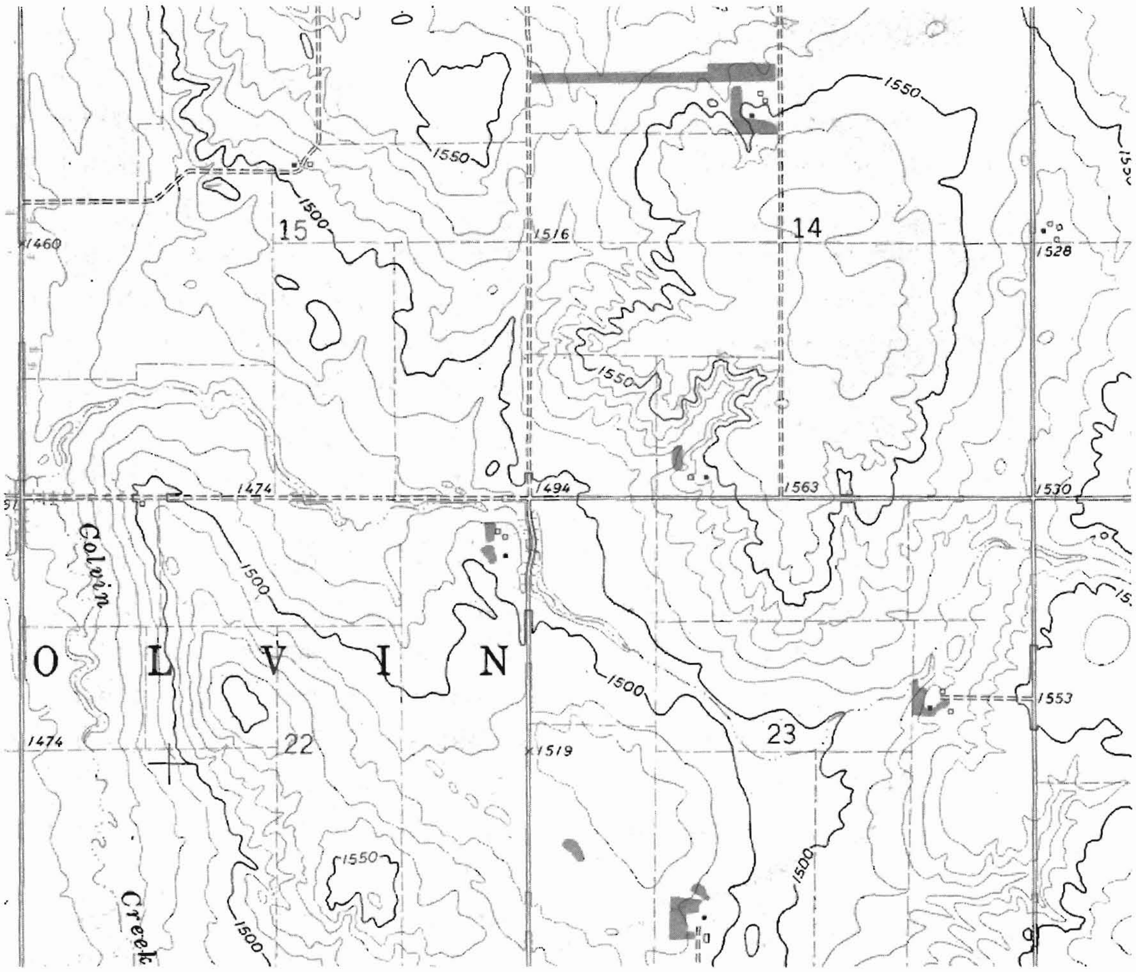

Figure 24. Thin glacial sediment draped over pre-existing glacial thrust masses (map unit Qcdg) in T149N, R62W, in eastern Eddy County. U.S. Geological Survey Johnson Lake Quadrangle (10-foot contour interval). Area shown is 3.6 kilometres wide. The last glacier came from the north-northwest.

because on airphotos the area looks like it has undergone slight subglacial remolding. The symbol for longitudinal glacial ridges is shown on the Geologic Map in many of these areas. (In some areas, however, postglacial eolian scour might have been confused with subglacial molding on the airphotos.) Map unit Qcm (discussed below) rather than Qcdg is shown where the effects of subglacial molding are judged to predominate over the effects of processes before the last advance. None of the longitudinal glacial ridges shown on the map are judged to be palimpsest in the sense that they formed before the last advance.

Thrust masses (discussed below) well behind the limit of an advance generally have evidence of subglacial molding. It is generally difficult to determine whether the thrusting occurred during the last advance or during a previous advance. Where the thrust masses appear to have been only slightly modified (north of

Cooperstown in Griggs County, for example), they have been interpreted to have formed during the last advance and are therefore mapped as thrust masses (Qct, discussed below). Where the thrust masses appear to have been highly modified, but have not been molded into longitudinal ridges (south of Pekin in Nelson and Griggs Counties, for example), they have been interpreted to have formed before the last advance and are therefore mapped as draped topography (Qcdg).

Thrust Masses

The thrust masses (Qct) shown on the Geologic Map of North Dakota are of three general forms: transverse ridges above overturned folds or at the ends of imbricate thrust slabs; roughly equidimensional hills containing thrust masses downglacier from a source depression of similar size and shape; and irregular forms.

The type with transverse ridges (Clayton and Moran, 1974, p. 103-104) is indicated on the Geologic Map by map unit Qct with the same line symbols used for transversely elongated hummocks in areas of collapse topography. It generally consists of a composite ridge about 2 kilometres wide, as high as 100 metres, and as long as 50 kilometres. The composite ridge generally contains 10 or 20 smaller ridges, each of which is several metres to a few tens of metres high and several tens of metres wide. In map view, the crests of the ridges are concave upglacier, with a radius of curvature of 2 to 10 kilometres. The individual ridges and the composite ridge are generally steepest on their downglacier side.

The small ridges are the crests of imbricate overturned folds or the ends of imbricate thrust slabs. The depth of folding or thrusting is generally unknown, but Kupsch (1962) suggested that similar features in Saskatchewan are generally deformed to a depth of less than 180 metres. Individual thrust sheets are typically 30 to 100 metres thick, a few hundred metres long (parallel to ice flow), and hundreds or thousands of metres wide (perpendicular to ice flow), and they typically dip upglacier between 30° and 80°.

Thrust masses consist of any sediment or rock that happened to be present below the glacier snout at the time of thrusting. Older glacial sediment is perhaps the most common, but it lacks the bedding needed to reconstruct the structure of the mass. Many thrust masses are composed of preexisting lacustrine or fluvial sediment of Pleistocene age, and several are known to be composed of Cretaceous or Paleocene sediment or rock. Sibley Buttes in Kidder County, for example (Clayton and Freers, 1967, p. 20, fig. R-20), consists of a ridge about 2 kilometres wide, with perhaps two dozen parallel hogbacks consisting of Fox Hills sandstone dipping upglacier about 60°. Much of the Antelope Hills in Pierce County (fig. 25) consists of steeply-dipping Cretaceous or Paleocene sediment (Carlson and Freers, 1975, p. 18). Dogden Butte in northeastern McLean County contains at least 65 metres of the Bullion Creek Formation

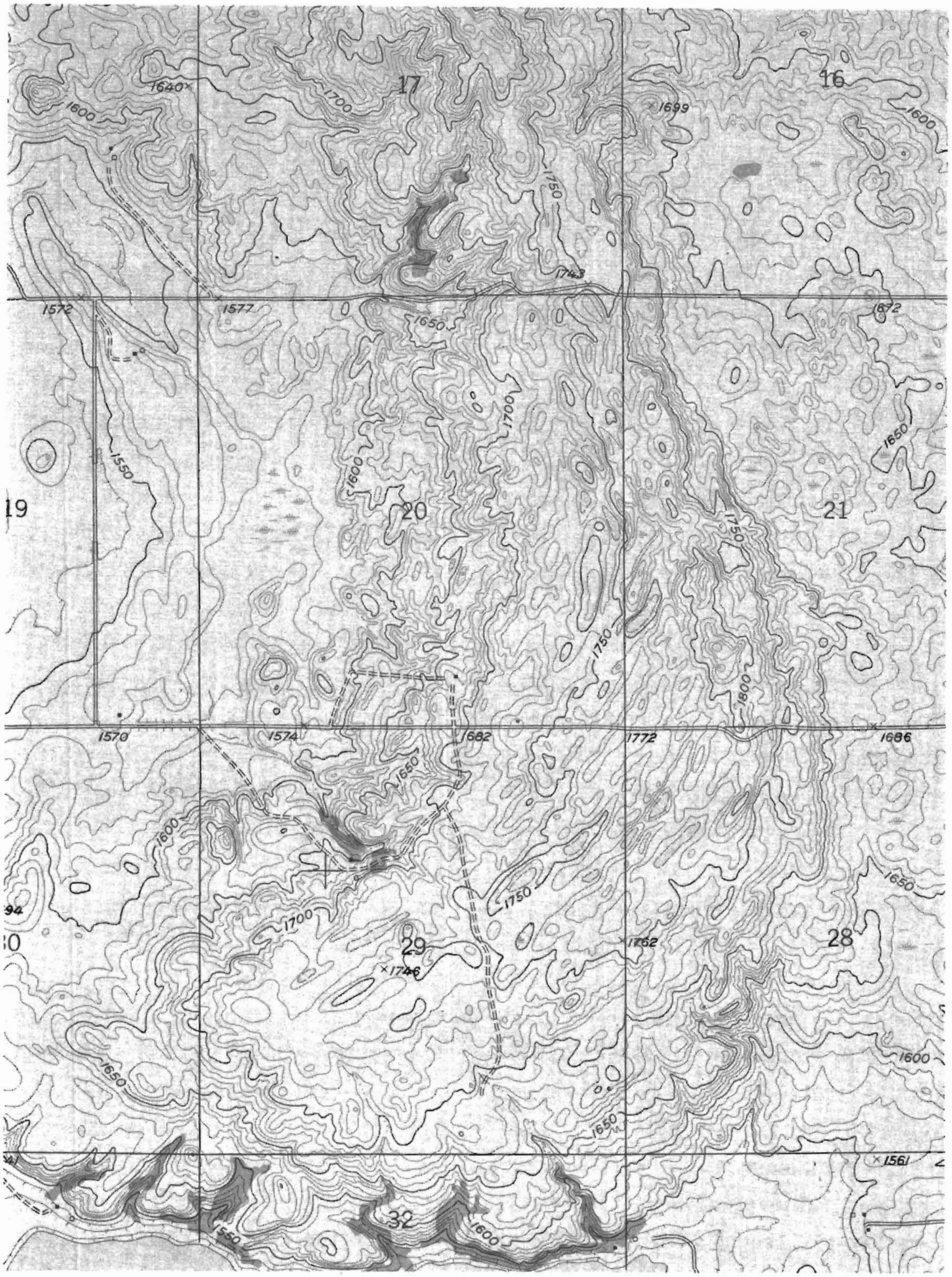

Figure 25. Glacial thrust masses (map unit Qct) in T152N, R73W, in southern Pierce County. U.S. Geological Survey Clifton Quadrangle (10-foot contour interval). Area shown is 3.6 kilometres wide. Glacier came from northwest, removing material from depression in section 19. The ridges are anticlines or the ends of thrust sheets.

thrust on top of Pleistocene sediment (Bluemle, 1971, p. 29).

Other examples of imbricate thrust masses include the "Streeter moraine" in Logan and McIntosh Counties; McPhails Buttes, "Woodhouse Lake loop," "Lake Williams loop," "Crystal Springs loop," and "Lake George loop" in Kidder County; Prophets Mountains in Sheridan County; "Alkabo moraine" in Divide County; Hawks Nest in Wells County; Binford Hills ("Cooperstown moraine") in Griggs County; and most other thrust masses on the Missouri Coteau and Turtle Mountains. Many of them have been described in detail in North Dakota Geological Survey county reports. Imbricate thrust ridges can be seen on the following U.S. Geological Survey 7.5-minute quadrangles: Pelican Lake, Tuttle SW, Steele NW, Steele NE, Horsehead Lake, Binford, Tappen NE, Tappen N, Goodrich SW, Goodrich SE, Hurdsfield SW, and Clifton (fig. 25).

A second general type of glacial thrust mass consists of a roughly equidimensional hill containing a horizontal, slightly crumpled, thrust sheet lying downglacier from a source depression of similar size and shape (Bluemle, 1970; Clayton and Moran, 1974, p. 104). The thrust sheet is typically about 30 metres thick and roughly 1 kilometre in diameter. The thrust mass is typically 1 or 2 kilometres downglacier from its source depression. An esker may originate in the source depression and wander downglacier around one side of the thrust mass. Many thrust slabs have been somewhat broken apart by imbricate thrust faults, making this type of thrust mass gradational with the type described previously.

Examples of thrust masses downglacier from source depressions include Egg Lake Hill and Egg Lake southeast of Harvey in Wells County, Butte de Morale and Goose Lake north of Harvey, the hill south of Anamoose and Steel Lake in McHenry County, Grasshopper Hills and Medicine Lake north of Jamestown in Stutsman County (fig. 26), Rugh Lake and the adjacent hill in eastern Nelson County, Blue Mountain and the adjacent depression west of Stump Lake in Nelson County, and Devils Lake Mountain and the adjacent depression in southern Ramsey County. Examples of thrust masses downglacier from source depressions can be seen on the following U.S. Geological Survey 7.5-minute quadrangles: Anamoose, Anamoose SW, Crete, Drake, Drake SE, Jim Lake (fig. 26), Manfred, Manfred NW, Wellsburg, and Harvey. A large proportion of the equidimensional lakes between 1 and 5 kilometres in diameter in North Dakota are thrust depressions.

A third type of glacial thrust mass, which is gradational with the two previously discussed types, consists of an irregular jumble of hills with no obvious transverse ridges and no obvious source depression. It is more difficult to identify than the other two types. Some have been recognized by the presence of displaced or deformed Pleistocene fluvial or lacustrine or Paleocene or Cretaceous sediment in outcrop or in

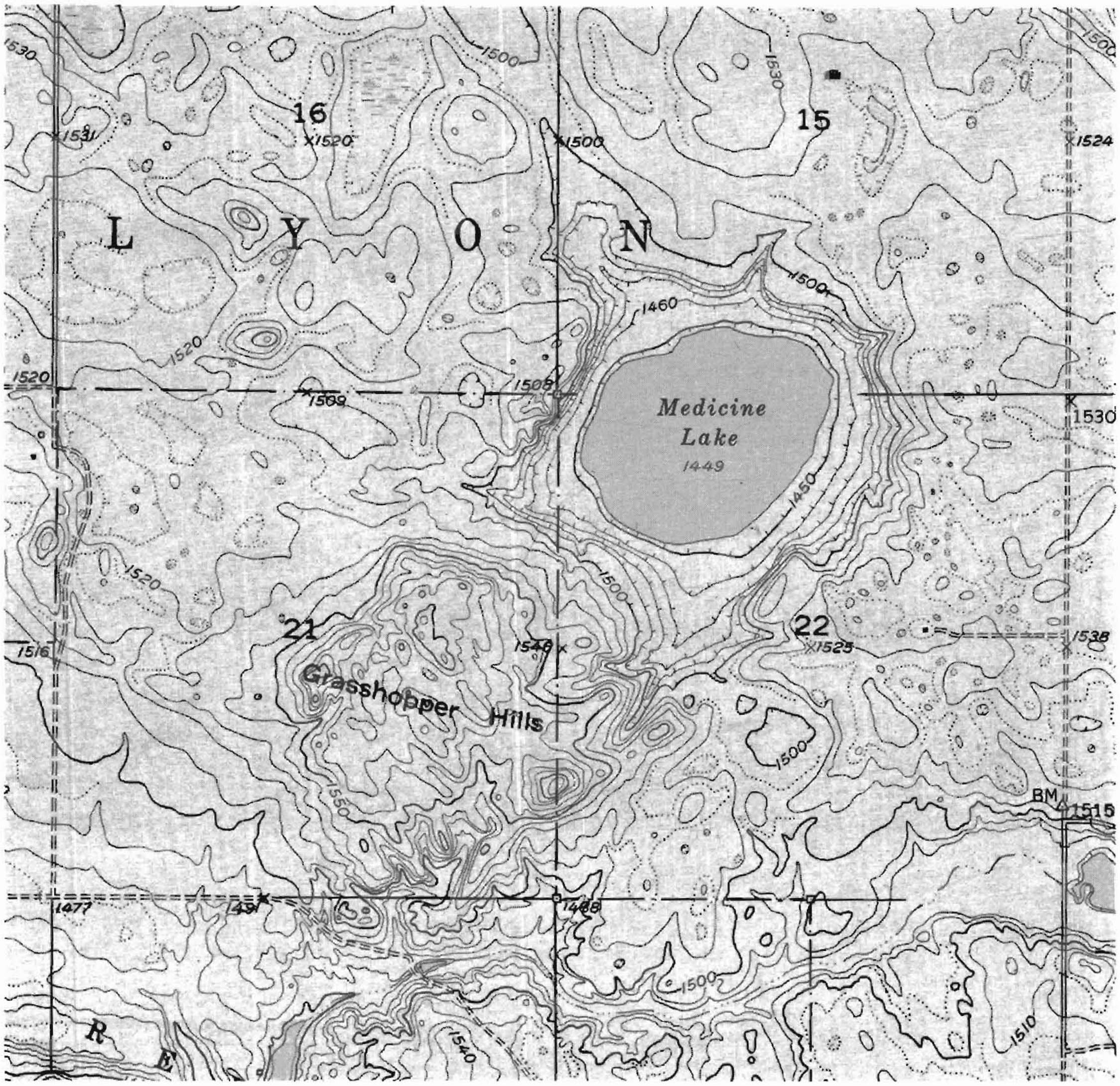

Figure 26. Glacial thrust mass (map unit Qct) in T143N, R64W, in northern Stutsman County. U.S. Geological Survey Jim Lake Quadrangle (10-foot contour interval). Area shown is 3.6 kilometres wide. Glacier came from northeast, thrusting Grasshopper Hills out of Medicine Lake.

drill holes. Most, however, were identified on airphotos. Conspicuous hills, having the general character of features known to be thrust masses, with no other known origin, were interpreted to be thrust masses on the Geologic Map. Examples include the areas mapped as Qct in northern Pierce County, northwestern Benson County, eastern Eddy County, and northwestern Griggs County.

All thrust masses shown on the Geologic Map are interpreted to have formed during the last glacial advance. Those that seem indistinct on airphotos are interpreted to be palimpsest thrust masses formed during an earlier advance and modified by hillslope processes before the last advance and by glacial processes during the last

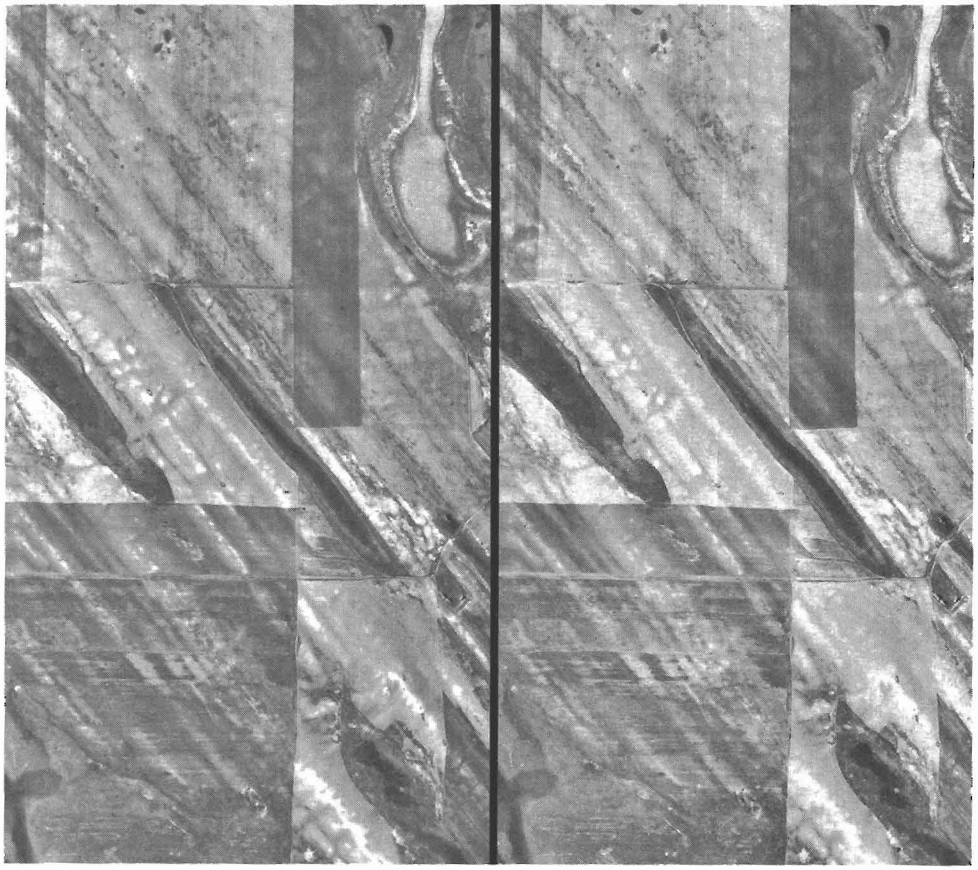

Figure 27. Subglacially molded topography (map unit Qcm) in sections 9, 10, 15, 16, 21, and 22, T152N, R77W, McHenry County. U.S. Department of Agriculture airphotos BA1-7BB-191 and -192. Area shown is 1.3 kilometres wide. Glacier came from northwest (lower right).

advance; they have been mapped as draped topography (Qcdg). Some of the irregular thrust masses discussed in the previous paragraph may have been misinterpreted and may in fact also be palimpsest thrust masses.

Subglacially Molded Surfaces

The areas mapped as Qcm have a thin layer of glacial sediment draped over Quaternary or pre-Quaternary sediment or rock that has been molded into streamlined longitudinal ridges and grooves beneath a sliding glacier. The ridges are indicated by a line symbol on the Geologic Map of North Dakota. This landform was recognized primarily by the presence of the streamlined forms on airphotos (fig. 27).

The streamlined ridges that are big enough to be indicated by a line symbol on the map are typically 1 to 5 kilometres long, about 50 metres wide, and a few metres high. The largest, Hogback Ridge in southern McHenry County, is 25 kilometres long, nearly 100 metres wide, and 15 metres high at its upglacier end (fig. 28). Only the larger ridges show up on topographic maps; see for example the following U.S. Geological Survey 7.5-minute

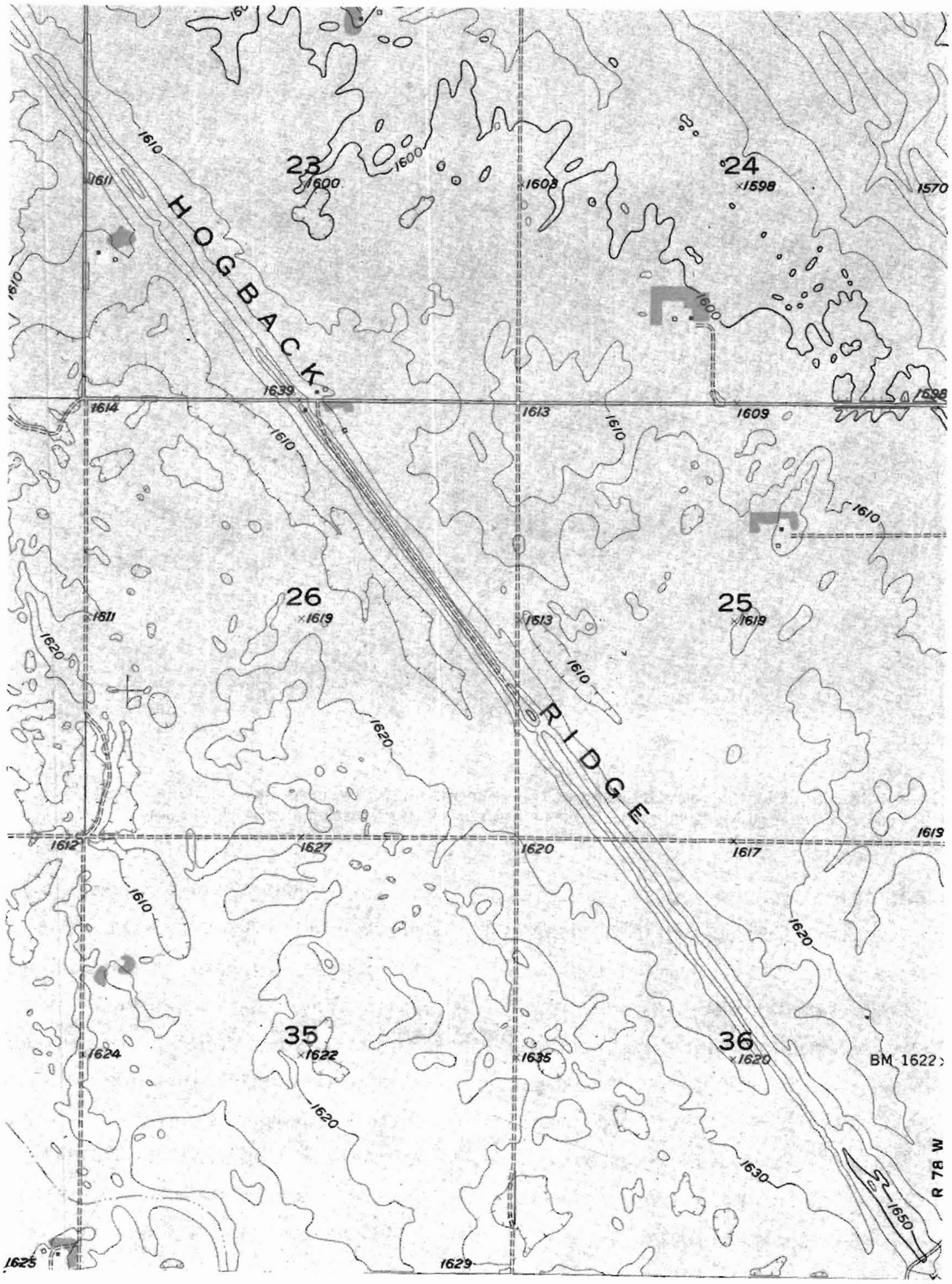

Figure 28. Subglacially molded topography (map unit Qcm) in T153N, R78W, in southern McHenry County. U.S. Geological Survey Bergen Quadrangle (10-foot contour interval). Area shown is 3.7 kilometres wide. Glacier came from northwest. The areas on either side of the ridge are gently undulating collapsed glacial sediment (map unit Qccg) with obscure transversely elongated hummocks.

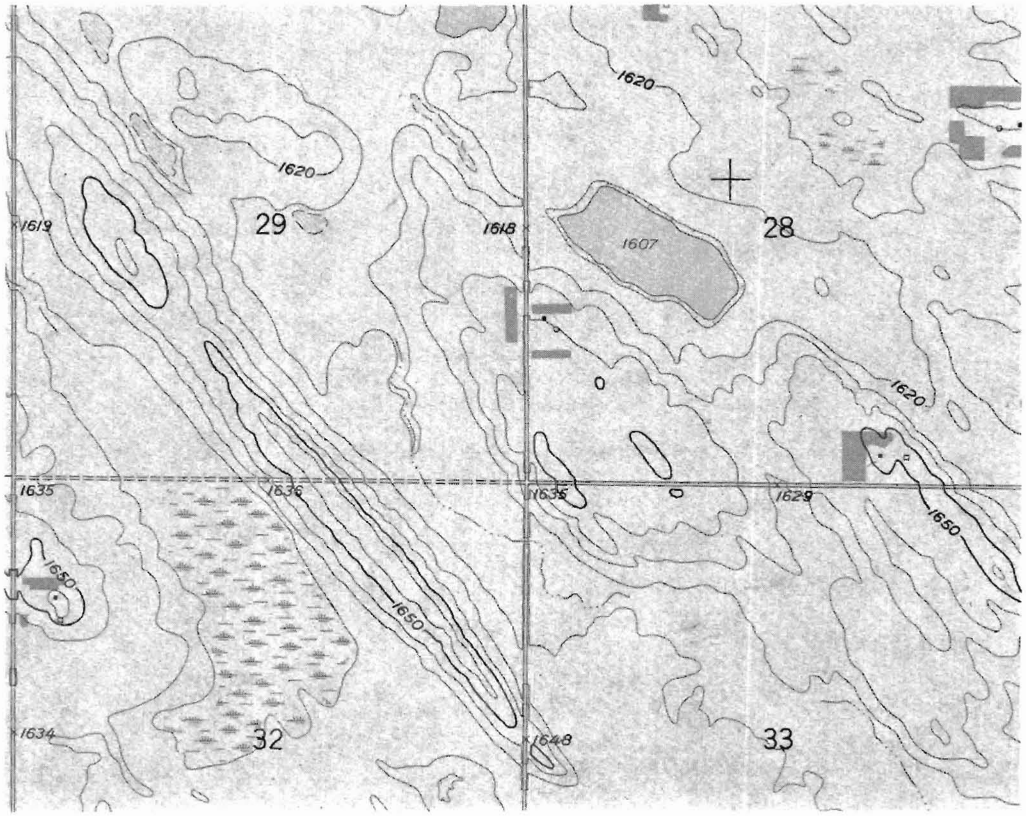

Figure 29. Subglacially molded topography (map unit Qcm) in T152N, R79W, in southern McHenry County. U.S. Geological Survey Kongsberg NE Quadrangle (10-foot contour interval). Area shown is 3.2 kilometres wide. Glacier came from northwest.

quadrangles: Voltaire (both conventional and shaded-relief maps), Bergen, Karlsruhe, Kongsberg NE (fig. 29), Balfour NW, and Balfour.

Many streamlined ridges contain some fine-grained glacial or lacustrine sediment, but fluvial sand and gravel seems to be the most common constituent. Those in central Cavalier County are composed of Pierre shale.

Any glacial sediment present in the cores of the ridges is generally interpreted to have been deposited before the last glacial advance. The glacial sediment of the last advance in most areas is a thin layer, generally less than 1 metre thick, draped over streamlined ridges and grooves. Where the glacial sediment of the last advance is thicker than about 1 metre but thinner than about 4 metres, collapse hummocks partly obscure the streamlined ridges; these areas are mapped as collapsed glacial sediment with line symbols for the longitudinal ridges.

Because most streamlined ridges in North Dakota are composed of sediment deposited before the last advance, they have generally been interpreted to be erosional features (Clayton and Moran, 1974, p. 100-101, fig. 3). However, many of the thrust masses

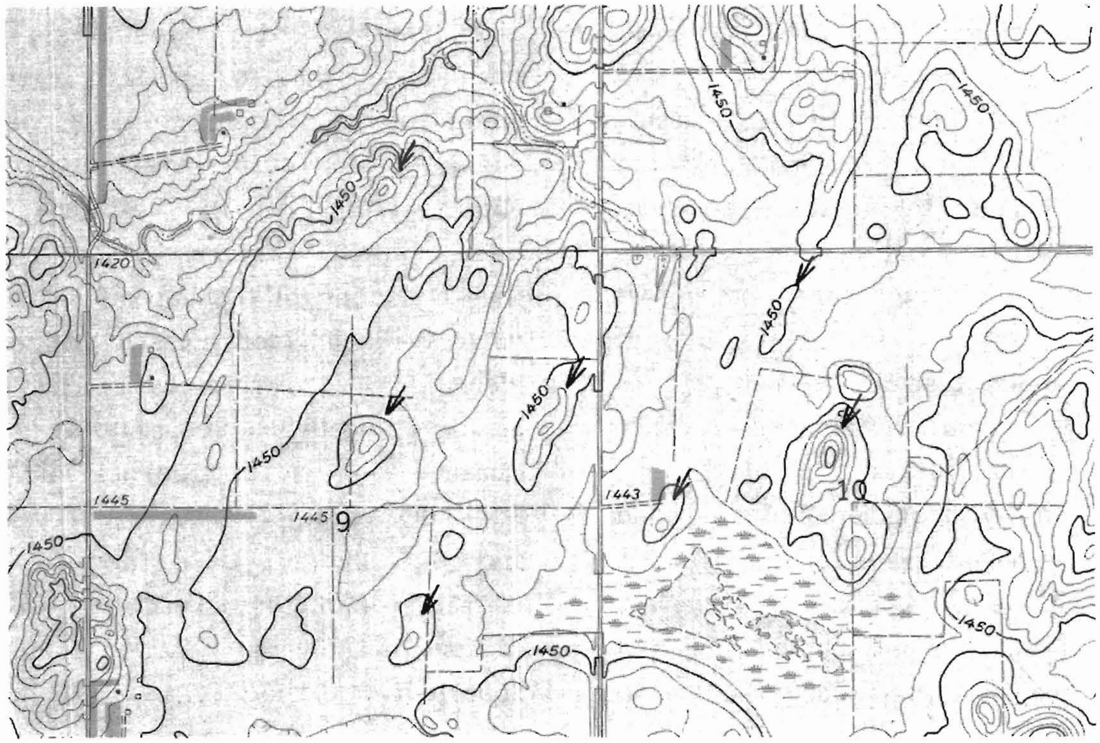

Figure 30. Glacial thrust masses somewhat streamlined by subglacial molding (arrows) in T147N, R59W, in northeastern Griggs County. U.S. Geological Survey Jessie Quadrangle (10-foot contour interval). Area shown is 3.6 kilometres wide. Glacier came from northeast.

discussed in the previous section have been somewhat streamlined by subglacial molding (fig. 30), and the upglacier end of many streamlined ridges is an irregular knob that was originally a thrust mass, suggesting that streamlined ridges and thrust masses are intergradational; many streamlined ridges consist of sediment sheared out of the grooves and piled behind thrust masses.

In the past this landform has generally been mapped as "ground moraine." The ridges have commonly been called "drumlins," and the grooves have commonly been called "flutes." The drumlins mapped in northeastern Emmons County by Colton and others (1963) are probably actually exhumed beaches in the Fox Hills Formation.

Eroded Glacial Sediment

Glacial sediment that lacks glacial topography because of postglacial erosion has been mapped as Qcew, Qcer, and Qces.

Wave-eroded glacial sediment (Qcew) occurs along the western part of the Lake Agassiz plain, the northern part of the Lake Souris plain, and the southern part of the Lake Cando plain. It is flat, and in some areas is overlain by thin unmappable patches of lag gravel, shoreline sand and gravel, and offshore silt and clay of the Coleharbor Group and fluvial, eolian, or lacustrine sediment of the Oahe Formation.

River-eroded glacial sediment (Qcer) has been mapped in areas that geomorphically appear to be the bottoms of large meltwater channels but are underlain by glacial sediment rather than fluvial sediment. It is possible that they are palimpsest channels with the fluvial sediment buried under glacial sediment of the last advance, but the channels seem to be too fresh looking to have been overridden by a glacier. River-eroded glacial sediment generally has flat topography and is commonly overlain by thin unmappable patches of lag gravel or sediment of the Oahe Formation.

Slopewash-eroded glacial sediment (Qces) has been mapped along the steepest parts of the escarpments at the edge of the Missouri Coteau (fig. 31), Prairie Coteau, and Turtle Mountains. In detail, these escarpments are made up of numerous small valleys. Although each valley has a stream in its bottom, the main process on the valley slopes, which makes up most of the area, is erosion by sheet flow. The valleys were cut in Holocene time, and the escarpments themselves were formed earlier, probably by glacial erosion and deposition.

Glacial Terminology

The terms "end moraine," "ground moraine," and "dead-ice moraine" have appeared on many previous North Dakota maps, but have not been shown on the Geologic Map of North Dakota for a variety of reasons.

The fundamental element in most definitions of "end moraine" is the requirement that it represent a thickening of the deposits of a glacial advance near (or parallel to) the outermost limit of the advance. This interpretation requires evidence for the thickening of the deposit, which is not available for most previously identified "end moraines." The only stratigraphically documented end moraine in North Dakota is the Edinburg moraine. Arndt (1977, pls. 1 and 2) showed that the Falconer Formation and Huot Formation, which consist largely of glacial sediment deposited during the Edinburg Advance, thicken from about 10 metres in southern Grand Forks County to about 30 metres in the Edinburg moraine in northern Traill County. The Huot Formation is included in map unit Qcew in northeastern Traill County, and the Falconer Formation is included in map unit Qcew in northern Traill County, Grand Forks County, Walsh County, and Pembina County and in map unit Qccu east of Elk Valley. The Kensal moraine in western Barnes County, eastern Stutsman County, and eastern Foster County, has not been stratigraphically documented, but probably is a true end moraine: truncated transverse lineations and overridden meltwater channels indicate that a readvance did occur, and steeper hummocks near the limit of the advance suggest that the glacial sediment is thicker there. Similar but less conclusive geomorphic evidence suggests

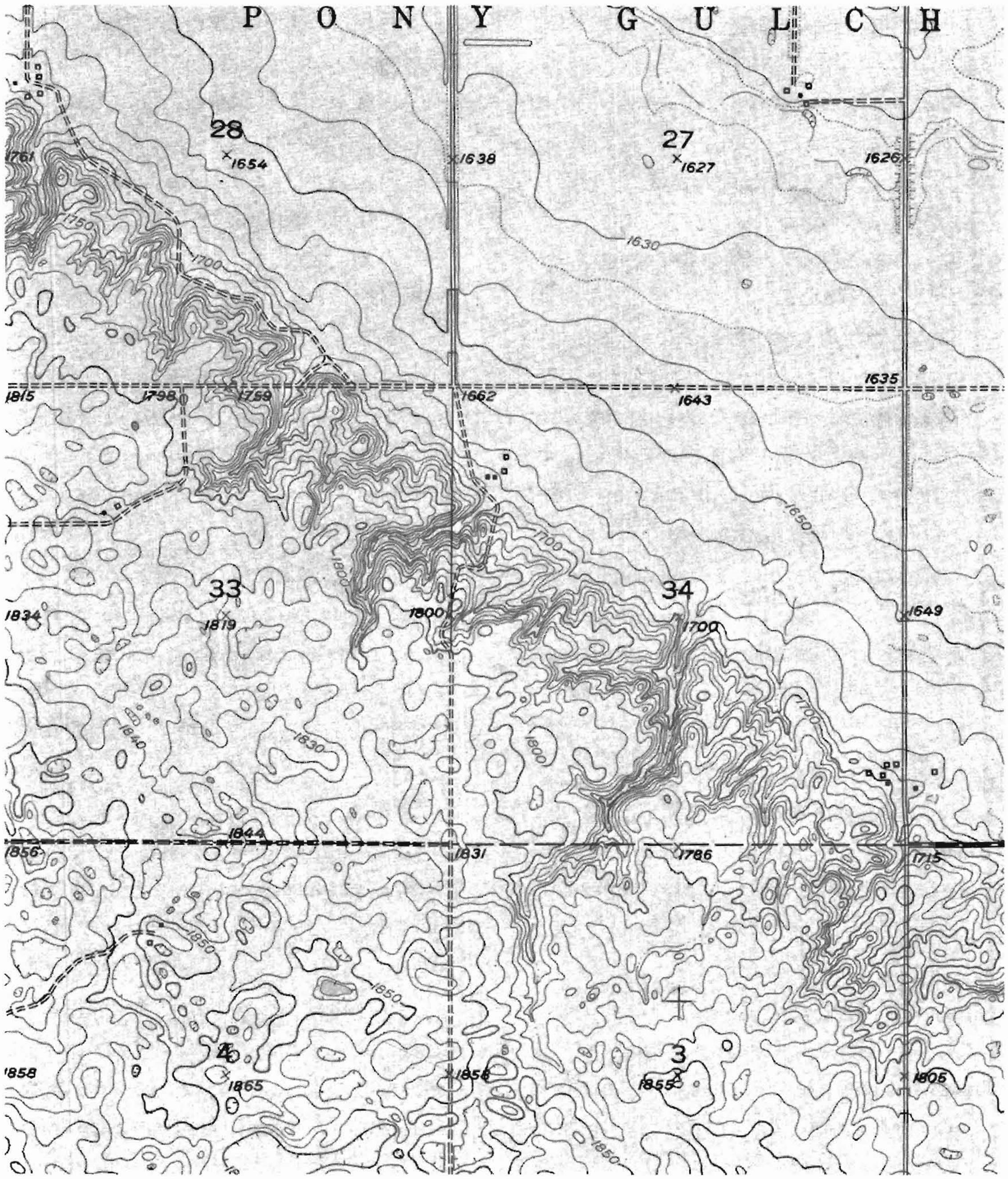

Figure 31. The Missouri Escarpment in Ts147 and 148N, R73W, in northeastern Wells County. U.S. Geological Survey Manfred SW Quadrangle (10-foot contour interval). Area shown is 3.7 kilometres wide. The southwestern part of the area is hummocky collapsed glacial sediment (map unit Qcch). The detailed topography of the escarpment is the result of postglacial slopewash erosion (map unit Qces). The northeastern part of the area is postglacial alluvial fans (map unit Qor). The origin of the escarpment is poorly understood, but in most places it is probably the result of glacial erosion and deposition.

that the Heimdal moraine in northeastern Wells County, the New Town moraine in southern Mountrail County, and the Charlson moraine in northeastern McKenzie County also are true end moraines. Little evidence is available to indicate that other "end moraines" previously mapped in North Dakota really are end moraines.

The well-documented end moraines are not shown on the Geologic Map because "end moraine" represents a different level of generalization than is represented by the units shown on the map. "Collapsed river sediment" or "collapsed glacial sediment," for example, indicate the processes involved in the deposition and modification of the sediment, and "subglacially molded" and "wave-eroded" indicate the processes involved in forming the topography. "End moraine," in contrast, indicates the event involved, not the process. For example, "Kensal moraine" indicates the extent of the Kensal Advance, but it indicates little about the specific glacial processes involved. The extent of an advance is indicated on the Geologic Map by a blue line (see next section).

"Ground moraine" has two basically different meanings. In practice, it has often referred to areas of glacial sediment that are not end moraine, making it an event-based rather than a process-based term. However, written definitions of the term commonly emphasize process; ground moraine is commonly considered to consist of subglacial tractive-load sediment (lodgment till). The only landforms known with some confidence to be the result of subglacial molding are longitudinal streamlined ridges and grooves (Boulton, 1976, p. 77), yet Flint (1971, p. 200) stated that they "are not ground moraine." Instead, Flint (1971, p. 199) stated that the characteristic topographic elements of ground moraine are low relief "swells, sags, and depressions"; however, these features are here interpreted to be collapse features (hummocks), not subglacial features. Flint (1971, p. 199) also stated that ground moraine is, by definition, "devoid of transverse linear elements"; however, transverse elements are one of the characteristic features of the "ground moraine" mapped by Colton and others (1963). Because of this confusion, the term has not been used on the Geologic Map of North Dakota.

"Dead-ice moraine" (or "stagnation moraine" or "collapse moraine") appears on many previous North Dakota maps (for example, Colton and others, 1963; and Clayton, 1962). The term has generally referred to hilly or rolling collapsed glacial sediment that is not in a transverse band behind an important ice-margin position (hilly or rolling collapsed glacial sediment in a transverse band behind an important ice-margin position was generally called "end moraine"). The terms "dead-ice" or "stagnation" are not used here because they put the emphasis in the wrong place. Active ice, not dead or stagnant ice, is required to produce supraglacial glacial sediment; areas of large-scale stagnation probably have little collapse topography because they lacked the thin marginal ice with active shearing required to produce thick supraglacial sediment. However, once an active glacier has produced englacial sediment, which turns into an insulating blanket of supraglacial sediment, the underlying stagnant ice may persist for hundreds or thousands of years. That is, large-scale (en masse) stagnation, which usually comes to mind when the terms "dead-ice" or "stagnation" are mentioned, may result in active-ice landforms, whereas small-scale (progressive marginal) stagnation may result in collapse landforms.

"Collapse moraine" is a better term because it emphasizes the process that produced the hummocky topography; on the Geologic Map, the term would not only be applied to hilly and rolling topography, as in the past, but also to undulating and gently undulating topography, which was generally called "ground moraine." However, the word "moraine" has not been used because there seems to be little general agreement about what the term includes. Instead, an attempt has been made to use terms that are in the general English vocabulary and are therefore less subject to divergent local usage than are technical terms.

Ice-Margin Positions

Significant ice-margin positions (fig. 32) are indicated on the Geologic Map of North Dakota by a blue line symbol. Many probably mark the outer limit of glacial readvances of at least several kilometres, and some mark readvances of hundreds of kilometres.

The ice-margin positions southwest of the Missouri River are based largely on boulder abundance. The limit of the Verone (?) Glaciation in Dunn County is marked by a band of abundant boulders; beyond it, an average of only about one boulder is seen along a kilometre of road (Clayton, 1969). The area of Verone glaciation in south-central North Dakota (Bickley, 1972, p. 97) has about ten boulders along a kilometre of road, whereas the area of Napoleon glaciation has hundreds.

Most of the other ice-margin positions shown on the Geologic Map of North Dakota are based on lithologic and geomorphic evidence. For example, the Charlson, New Town, Burnstad, Long Lake, and Zeeland ice-margin positions mark the boundary between glacial sediment with little or no constructional topography and glacial sediment with some areas of relatively fresh-looking constructional topography. The Kensal and Heimdal ice-margin positions are marked by truncated transverse lineations. The Zeeland, Burnstad, Long Lake, Martin, Heimdal, McHenry, Oakes, Cooperstown, and Pekin ice-margin positions are marked by outwash fans. The Mylo ice-margin position is the outer edge of a zone of abundant transverse lineations. The Luverne ice-margin position is marked by a slight change in hummock steepness and by outwash fans. The Minot ice-margin position is marked by proglacial meltwater channels, thrust masses, and a general change in topography, with subglacially molded topography and transversely lineated collapse topog-

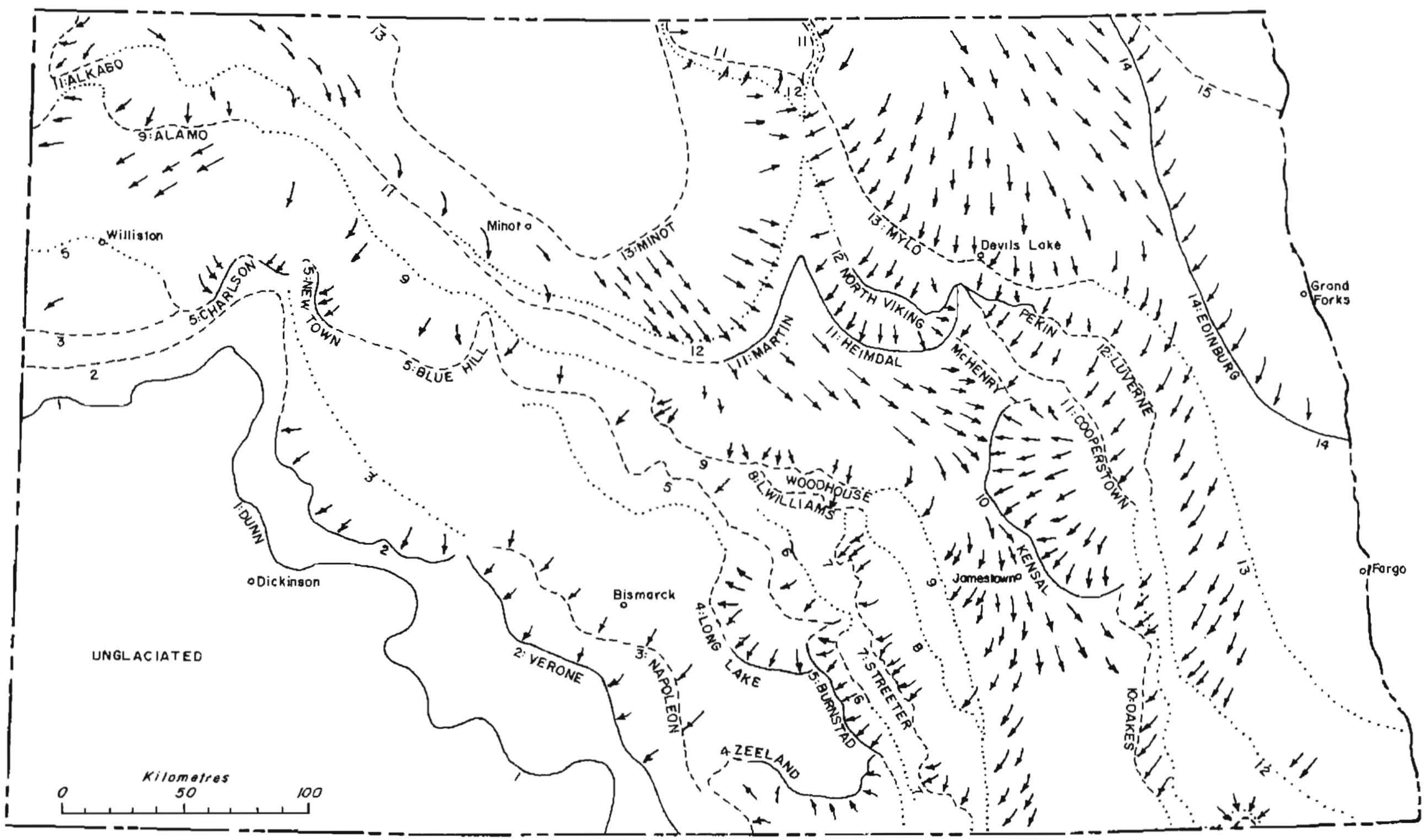

Figure 32. Ice-movement directions (arrows) and outer margins of glacial advances.

raphy beyond the margin and collapse topography and ring-shaped hummocks behind the margin. The minor ice-margin positions shown on the Geologic Map in Divide, western Sheridan, northern and eastern Kidder, eastern Logan, eastern McIntosh, and north-eastern Eddy Counties are marked by thrust masses.

In addition, the transverse glacial ridges indicated on the Geologic Map of North Dakota each mark an ice-margin position of minor significance.

Some of the major contributions to the Quaternary geology of North Dakota in the past decade have come from lithostratigraphic studies (Moran and others, 1976). Only a few ice-margin positions can be definitely related to known stratigraphic units, but several can be tentatively correlated with known units.

Ulmer and Sackreiter (1973), Salomon (1974), and Fulton (1976) have identified a sequence of three units of glacial sediment across much of western North Dakota. Various interpretations seem possible for this sequence of units. The Snow School Formation may be the sediment of the Napoleon Glaciation (margin 3), the Horseshoe Valley Formation may be the sediment of the Verone Glaciation (margin 2), and the Medicine Hill Formation may be the sediment of the Dunn Glaciation (margin 1). Alternatively, the Snow School and Horseshoe Valley Formations, which are similar in lithology, may both be sediment of the Napoleon Glaciation, and the Medicine Hill Formation, which is considerably more weathered looking than the overlying units, may be the sediment of the Verone Glaciation. Moran and others, 1976 (fig. 7), suggested that the Snow School and Horseshoe Valley are Late Wisconsinan in age and the Medicine Hill is the result of the Napoleon Glaciation.

In the south-central part of the state, the limit of the Zeeland Advance (margin 4) is marked by the outer limit of an unnamed lithostratigraphic unit consisting largely of dark-gray, clayey glacial sediment, which strikingly contrasts with the unnamed underlying unit containing yellowish, sandy glacial sediment deposited during the Napoleon Advance (Clayton, 1962, p. 62; Bickley, 1972).

In the eastern part of North Dakota a similar relationship appears to exist across the Kensal ice-margin position (margin 10), which marks the limit of gray, clayey, shale-laden glacial sediment of the Dahlen Formation (Salomon, 1975, p. 40-68; Hobbs, 1975). The underlying unnamed unit, which forms the surface to the west, contains glacial sediment that is olive brown, more sandy, and contains very little shale (Camara, 1977).

In the Red River Valley, two ice-margin positions are clearly the outer limit of lithostratigraphic units. The limit of the Edinburg Advance (margin 14) is marked by outer limit of the Falconer and Huot Formations, which are lateral facies equivalents (Moran and others, 1976, p. 138-140). Ice-margin position 15 marks the southernmost extent of a sandy unit of glacial sediment that is correlated with the Marchand Formation of southeastern Manitoba (Teller and Fenton, 1979). Extensive deposits of proglacial gravel underlie the sediments of Lake Agassiz along this boundary.

Some of the ice-margin positions indicated by Colton and others (1963) have been omitted from the Geologic Map because of lack of evidence or because of evidence against their existence. For example, ice movement was parallel rather than perpendicular to the "Beldon" ice-margin position (Clayton, 1972, p. 64).

Glacial Chronology

The glacial chronology of North Dakota has recently been discussed by Moran and others (1976), emphasizing the lithostratigraphy. The following discussion emphasizes the ice-margin positions and radiocarbon dates.

Figures 33 and 34 are correlations of the ice-margin positions of adjacent areas. Correlations with Saskatchewan (Christiansen, 1979), Montana (Colton and others, 1961), Manitoba (Klassen, 1972 and 1975), South Dakota (Anonymous, 1971), Iowa (Ruhe, 1969), and Minnesota (Wright, 1972; Wright and others, 1973; Flint and others, 1959; Moran and others, 1976; Harris, 1975; Sackreiter, 1975; Anderson, 1976; and Perkins, 1977) were made, for the most part, by direct matching at area boundaries.

As can be seen in figure 34, there has been little agreement on the age of the ice margins. Each geologist has selected a different set of radiocarbon dates, and these are in conflict because many of the dates are unreliable as the result of contamination.

Dead carbon in organic detritus, peat, gyttja, and muck results in dates that are too old. Many, perhaps most, of the dates from these materials in the midcontinent area are too old because of contamination with dead carbon. Coal-bearing rock is widespread throughout this region. Coal, especially the lignite that underlies western North Dakota, southwestern Manitoba, and southern Saskatchewan, has a low density and is brittle. As a result, small fragments are easily transported and are widespread in glacial, fluvial, and lacustrine sediment. The inclusion of even small amounts of dead carbon can alter the age of a sample by several hundred to several thousand years.

Numerous examples of samples that are contaminated with dead carbon can be cited from the midcontinent region. Two samples collected from the same outcrop in overbank alluvium along the Sheyenne River in Richland County permit an estimate of the degree of such contamination. The lower sample, a piece of wood that had a date of 2540±300 B.P. (W-1185), was collected about 0.3 m beneath a sample of "charcoal" that gave a date of 23,400±800 B.P. (W-1184) (table 1). Calculations made using the radioactive-decay law indicate that the "charcoal" consisted of 92 percent lignite.

Another example is given by two basal dates from organic muck in Hebron Bog in northern Iowa: $30{,}300\,^{+1500}_{-1300}$ B.P. (I-1859) and $27{,}990\,^{+1100}_{-1000}$ B.P. (I-1858). On the basis of various lines of evidence,

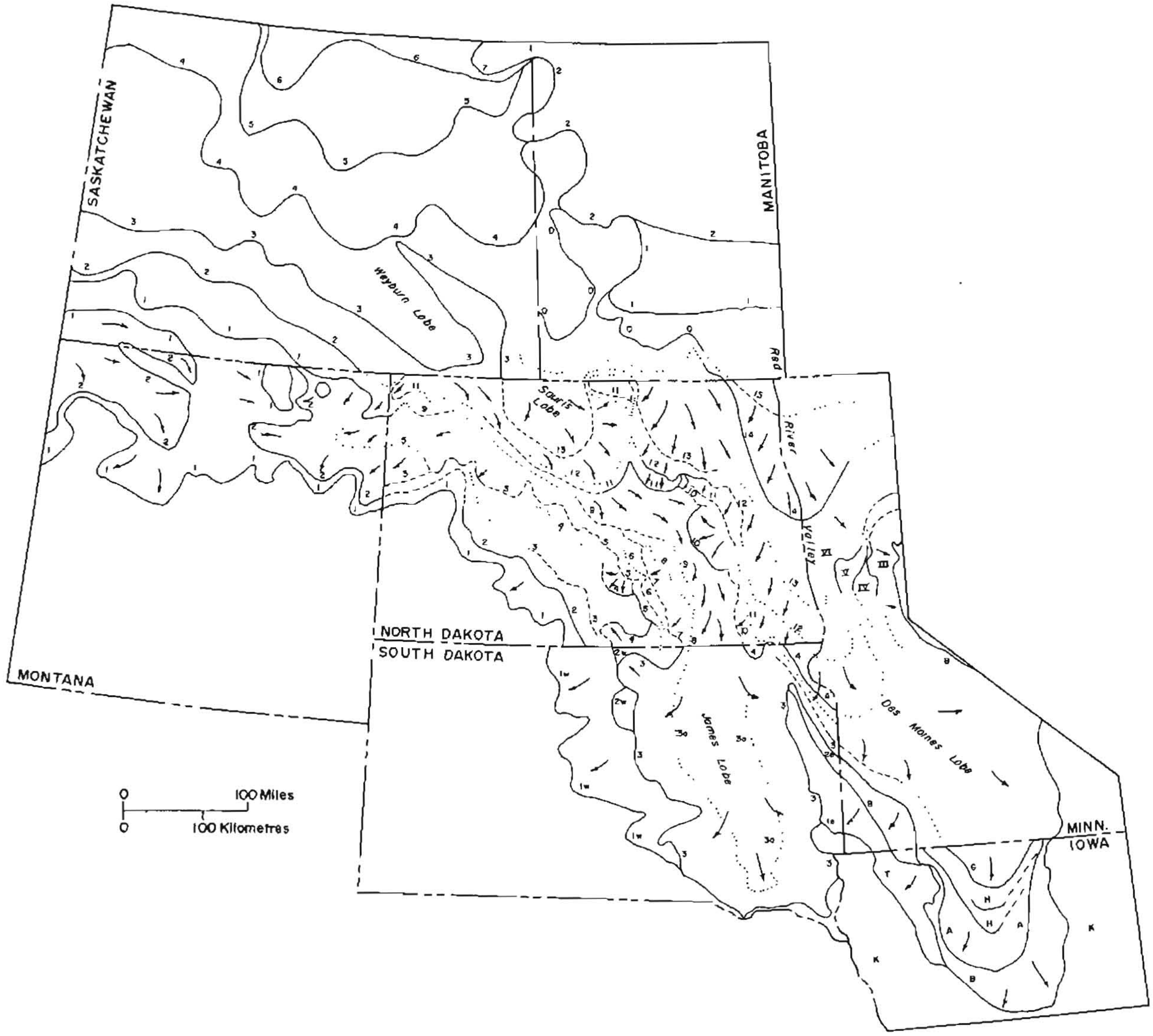

Figure 33. Ice-margin positions in the midcontinent region. Sources of information are given in figure 34.

Ruhe (1969, p. 223-231) concluded that the base of the bog could be no older than about 13,000 B.P. and that the samples were contaminated.

The samples just described were obviously contaminated. Problems in interpretation arise, however, when the contamination is much less than in the above examples. Inclusion of 5 percent dead carbon in a sample results in an artificial aging of 424 years, and inclusion of 20 percent results in an age that is too old by 1,845 years. Errors this small often go undetected, and some dates have been accepted as uncontaminated because they agreed with the proposed chronology, whereas others were rejected because they disagreed. Frequently a certain sequence of dates has been accepted because the series is internally consistent. However, because

Saskatchewan Christiansen, 1979	Manitoba Klassen, 1972	North Dakota: Moran and others, 1976		North Dakota: Clayton, 1966	South Dakota Anonymous, 1971	Iowa Ruhe, 1969	N.W. Minnesota Perkins, 1977
6: 12 000							
5: 12 500	2: 14 000						
4: 14 000	1: 15 000						
	0		15				
		13 000	14	11 900			
3: 15 500			13				
			12	12 000			
2: 16 500			11	12 100	4e		VI
			10	12 200	4w		? V
			9			G: 13 000	? IV
			8			H	
1: 17 000			7	12 400	3	A	
			6	12 700			
			5	12 900	? 2e	B: 14000	? III
			4	15 000	2w ? 1e	T: 20 000	
		Early Wisc.	3	40 000			
			2				
		Pre-Wisc.	1	Pre-Wisc.	1w	K: Pre-Wisc.	

Figure 34. Correlation of ice-margin positions in the midcontinent region. The large numbers are years before present, as given in the original reference.

such sequences of dates are commonly based on several dates from fine-grained organic sediment, there is no way to determine which, if any, of the dates are uncontaminated. In many areas it is possible to build two rather different internally consistent chronologies depending on which dates are accepted and rejected. This is not to say that all dates from organic detritus, peat, muck, organic silt and clay, and charcoal are contaminated, but rather that there exists a reasonable probability of such contamination. Since there seems to be no way to distinguish contaminated from uncontaminated samples, we have adopted the conservative position of rejecting all these dates.

Dates from marl and mollusk shells,

TABLE 1. References to radiocarbon dates discussed in section on glacial chronology

Sample Designation	Date	Reference
C-528	16,367±1000	Ruhe, 1969, p. 200
C-596	11,952±500	Ruhe, 1969, p. 200
C-653	12,200±500	Ruhe, 1969, p. 200
C-664	14,042±1000	Ruhe, 1969, p. 201
C-912	12,120±530	Ruhe, 1969, p. 201
C-913	13,300±900	Ruhe, 1969, p. 201
GSC-205	24,490±200	Dyck, Fyles, and Blake, 1965, p. 31
GSC-383	10,600±150	Lowdon, Fyles, and Blake, 1967, p. 165
GSC-391	9990±160	Lowdon, Fyles, and Blake, 1967, p. 165
GSC-492	10,670±160	Klassen, 1972, p. 550
GSC-618	10,710±250	Klassen, 1972, p. 551
GSC-653	37,700±150	Klassen, 1969, p. 5
GSC-677	10,690±190	Klassen, 1972, p. 550
GSC-689	10,920±150	Lowdon and Blake, 1970, p. 65
GSC-711	28,220±380	Lowdon and Blake, 1968, p. 217
GSC-797	9700±140	Klassen, 1972, p. 550
GSC-859	10,900±160	Lowdon, Robertson, and Blake, 1971, p. 291-292
GSC-870	10,000±150	Klassen, 1972, p. 550
GSC-902	10,600±150	Klassen, 1972, p. 550
GSC-903	12,400±600	Lowdon, Robertson, and Blake, 1971, p. 291-292
GSC-1081	11,600±430	Klassen, 1972, p. 550
GSC-1129	21,700±840	Lowdon, Robertson, and Blake, 1971, p. 290-291
GSC-1279	23,700±290	Lowdon and Blake, 1973, p. 22
GSC-1319	12,100±160	Christiansen, 1979, table 1
GSC-1332	10,500±180	Lowdon, Robertson, and Blake, 1971, p. 286
GSC-1342	30,000±490	Lowdon and Blake, 1973, p. 23
GSC-1369	14,300±320	Christiansen, 1979, table 1
GSC-1428	10,000±280	Klassen, 1972, p. 551
Gx-2741	$20{,}670^{+1500}_{-1000}$	Matsch, Rutford, and Tipton, 1972, p. 10
Gx-2864	$26{,}150^{+3000}_{-2000}$	M. J. Tipton (SDGS), personal communication
Gx-3530	$22{,}260^{+1000}_{-900}$	Teller and Fenton, 1979
I-1015	13,775±300	Ruhe, 1969, p. 203
I-1019	11,640±400	Ruhe, 1969, p. 203
I-1024	16,100±500	Ruhe, 1969, p. 204-205

TABLE 1. References to radiocarbon dates discussed in section on glacial chronology--Continued

Sample Designation	Date	Reference
I-1268	13,900±400	Ruhe, 1969, p. 206
I-1270	16,100±1000	Ruhe, 1969, p. 206-207
I-1402	14,200±500	Ruhe, 1969, p. 207
I-1414	14,500±340	Ruhe, 1969, p. 209
I-1416	11,570±330	Buckley, Trautman, and Willis, 1968, p. 259
I-1682	12,800±350	Klassen, 1972, p. 550
I-1858	$27{,}990^{+1100}_{-1000}$	Ruhe, 1969, p. 211
I-1859	$30{,}300^{+1500}_{-1300}$	Ruhe, 1969, p. 211
I-1864A	20,500±400	Ruhe, 1969, p. 212
I-1982	9130±150	Moran and others, 1973, p. 5
I-2106	10,250±140	Buckley, Trautman, and Willis, 1968, p. 264
I-2289	13,500±200	Moran and others, 1973, p. 7
I-2537	12,000±250	Moran and others, 1973, p. 11
I-3157	9430±160	Buckley and Willis, 1969, p. 63
I-3475	11,140±200	Klassen, 1972, p. 550
I-3476	13,900±240	Ritchie, 1976
I-3880	9940±160	Ashworth, Clayton, and Bickley, 1972
I-4878	27,400±850	Lichti-Federovich, 1975
I-5123	9650±150	Moran and others, 1973, p. 13
I-5123C	9730±160	Moran and others, 1973, p. 14
I-5213	10,340±170	Moran and others, 1973, p. 14
I-6361	14,190±220	Christensen, 1977, p. 22
O-1325	20,000±800	Ruhe, 1969, p. 201
S-41	10,000±300	McCallum and Dyck, 1960, p. 74
S-83	11,700±300	Parizek, 1964
S-96	27,750±1200	McCallum and Wittenberg, 1962, p. 72
S-107	7350±100	McCallum and Wittenberg, 1962, p. 74
S-110	10,300±400	McCallum and Wittenberg, 1962, p. 72
S-123	10,900±700	McCallum and Wittenberg, 1962, p. 73
S-128	10,800±300	McCallum and Wittenberg, 1962, p. 73
S-129	9570±130	McCallum and Wittenberg, 1962, p. 75
S-140	10,600±300	McCallum and Wittenberg, 1962, p. 74
S-173	13,000±200	McCallum and Wittenberg, 1965, p. 230
S-174	10,250±150	McCallum and Wittenberg, 1965, p. 230-231
S-176	20,000±850	McCallum and Wittenberg, 1965, p. 231

TABLE 1. References to radiocarbon dates discussed in section on glacial chronology--Continued

Sample Designation	Date	Reference
S-190	11,650±150	Klassen, 1972, p. 551
S-198	12,140±240	McCallum and Wittenberg, 1965, p. 231
S-206	21,700±1400	McCallum and Wittenberg, 1968, p. 365
S-227	10,800±160	McCallum and Wittenberg, 1965, p. 231-232
S-228	21,000±800	McCallum and Wittenberg, 1968, p. 366
S-228A	18,000±450	McCallum and Wittenberg, 1968, p. 366
S-228B	19,200±400	McCallum and Wittenberg, 1968, p. 366
S-235	12,000±180	McCallum and Wittenberg, 1968, p. 367
S-236	9400±160	McCallum and Wittenberg, 1968, p. 367
S-241	15,200±260	McCallum and Wittenberg, 1968, p. 367-368
S-246	12,000±200	McCallum and Wittenberg, 1968, p. 369
S-248	11,620±130	McCallum and Wittenberg, 1968, p. 369
S-457	11,400±190	Rutherford, Wittenberg, and McCallum, 1975, p. 329
S-461	9500±150	Rutherford, Wittenberg, and McCallum, 1975, p. 330
S-494	22,100±465	Rutherford, Wittenberg, and McCallum, 1975, p. 330
S-527	22,410±485	Rutherford, Wittenberg, and McCallum, 1975, p. 330-331
S-553	12,025±205	Christiansen, 1979, table 1
S-685	14,040±465	Christiansen, 1979, table 1
TAM-1	10,820±190	Moran and others, 1973, p. 19
W-126	16,720±500	Ruhe, 1969, p. 217
W-153	14,700±400	Ruhe, 1969, p. 217
W-512	14,470±400	Ruhe, 1969, p. 218
W-513	13,820±400	Ruhe, 1969, p. 218
W-517	13,910±400	Ruhe, 1969, p. 218
W-542	11,480±300	Moran and others, 1973, p. 21
W-560	11,540±200	Rubin and Alexander, 1960, p. 148
W-625	13,030±250	Ruhe, 1969, p. 220
W-626	12,970±250	Ruhe, 1969, p. 220
W-635	12,090±300	Rubin and Alexander, 1960, p. 149
W-723	10,960±300	Moran and others, 1973, p. 22
W-771	9800±250	Rubin and Alexander, 1960, p. 149
W-801	12,200±400	Lemke and others, 1965, p. 22
W-824	12,650±350	Rubin and Alexander, 1960, p. 149
W-900	10,080±280	Moran and others, 1973, p. 22

TABLE 1. References to radiocarbon dates discussed in section on glacial chronology--Continued

Sample Designation	Date	Reference
W-954	9870±290	Moran and others, 1973, p. 23
W-956	11,070±300	Moran and others, 1973, p. 23
W-974	11,650±310	Moran and others, 1973, p. 24
W-987	12,530±350	Lemke and others, 1965, p. 22
W-993	9900±400	Moran and others, 1973, p. 24
W-1005	10,050±300	Moran and others, 1973, p. 25
W-1019	9000±300	Moran and others, 1973, p. 25
W-1033	10,060±300	Lemke and others, 1965, p. 22
W-1057	9200±600	Ives and others, 1964, p. 44
W-1149	9620±350	Moran and others, 1973, p. 27
W-1184	23,400±800	Moran and others, 1973, p. 27
W-1185	2540±300	Moran and others, 1973, p. 28
W-1189	12,050±400	Lemke and others, 1965, p. 22
W-1360	9810±300	Moran and others, 1973, p. 29
W-1361	9820±300	Moran and others, 1973, p. 29
W-1369	9860±400	Moran and others, 1973, p. 30
W-1372	12,200±400	Levin and others, 1965, p. 378
W-1436	9990±300	Moran and others, 1973, p. 31
W-1755	11,770±500	Ives and others, 1967, p. 509-510
W-1756	12,350±300	Ives and others, 1967, p. 510
W-1757	12,680±300	Ives and others, 1967, p. 511
W-1817	10,350±300	Moran and others, 1973, p. 34
W-1818	10,330±300	Moran and others, 1973, p. 34
W-2201	10,880±330	Sullivan, Spiker, and Rubin, 1970, p. 322
W-2305	9220±300	Christensen, 1977, p. 22
W-2450	28,340±1000	Moran and others, 1973, p. 36
Y-165	12,400±420	Klassen, 1972, p. 550
Y-166	11,230±480	Preston, Person, and Deevey, 1955, p. 957
Y-411	10,550±200	Barendsen and others, 1957, p. 912
Y-452	12,330±180	Lemke and others, 1965, p. 22
Y-595	12,760±120	Lemke and others, 1965, p. 22
Y-925	12,520±100	Stuiver, 1969, p. 578
Y-1327	11,740±200	Stuiver, 1969, p. 576
Y-1922	12,000±160	Stuiver, 1969, p. 573

which make up a large proportion of the dates in the drier areas of the western interior, are also suspect. As a result of isotopic fractionation within biological systems and the incorporation of dead carbon from limestone (Bowen, 1978, p. 117-120), both marl and shells appear to give dates that are several thousand years older than equivalent wood dates. A pair of dates (GSC-859 and GSC-903) from wood and shells in Alberta and another (GSC-797 and GSC-689) from Manitoba have differences of 1500 and 1220 years, respectively. Similarly paired dates from wood and marl in Alberta (S-107 and S-140) and Saskatchewan (S-236 and S-235) have differences of 3250 and 2600 years, respectively. For this reason, dates from carbonate materials are also rejected.

The earliest glaciations for which evidence exists in southwestern North Dakota are beyond the range of radiocarbon dating (fig. 35). Margin 1 (fig. 32) is pre-Wisconsinan. Glacial landforms are totally absent, and glacial sediment other than scattered lag boulders is known in only one place between margins 1 and 2 (Moran and others, 1976, p. 147). At least 60 metres of erosion has occurred in many areas since advance 1 (Clayton, 1969).

Advance 2 (fig. 32) may have occurred in pre-Wisconsinan or Early Wisconsinan time. Lag boulders are more abundant behind margin 2 than behind margin 1 (Bickley, 1972, p. 97).

During advance 2, water from most of Montana and northern Wyoming flowed just beyond margin 2, entering the state through the Bennie Pierre channel in southwestern McKenzie County. It then emptied into a hypothetical lake in the Little Missouri Valley and continued southeastward through the Killdeer-Shields channel (fig. 3).

Lag boulders are even more abundant behind margin 3, and glacial sediment is widely preserved. In some places, constructional glacial topography is still present (Moran and others, 1976, p. 149-150). Glaciation 3 could have occurred in pre-Wisconsinan, Early Wisconsinan, or Late Wisconsinan time. Early Wisconsinan seems most likely.

The Late Wisconsinan glacier advanced across North Dakota between about 25,000 and 20,000 B.P. Only one radiocarbon date in North Dakota relates to this advance, but dates in surrounding areas confirm this interpretation. Near Minot, a date of 28,340±1000 B.P. (W-2450) was obtained from wood beneath a thick sequence of fluvial sediment overlain by at least two units containing glacial sediment (references to all dates are listed in table 1). Wood from a depth of 52 metres in sand under two units containing glacial sediment near Outram, Saskatchewan, gave an age of 27,750±1200 B.P. (S-96). Wood near the top of a layer of fluvial sediment overlain by 18 metres of glacial sediment at Medicine Hat, Alberta, was dated at 24,490±200 B.P. (GSC-205). Wood from fluvial sediment overlain by lake clay near Watino, Alberta, was dated 27,400±850 B.P. (I-4878). Wood from a depth of 27.5 metres in a well

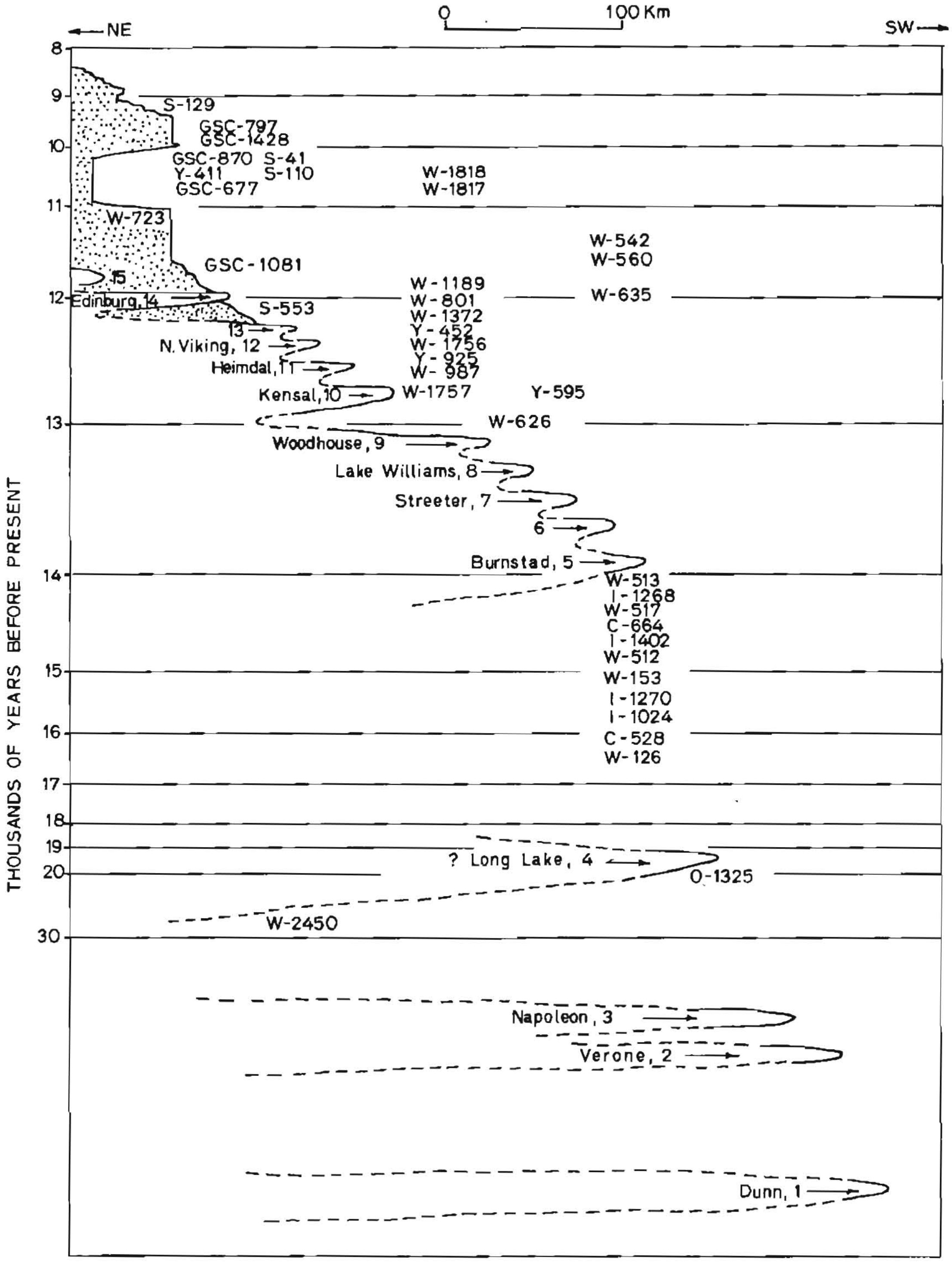

Figure 35. North Dakota glacial chronology, along a line from northeastern to southwestern North Dakota. The dotted area represents Lake Agassiz (fig. 8).

in eastern South Dakota was dated $26{,}150^{+3000}_{-2000}$ B.P. (Gx-2864). Wood under a layer of pinkish-gray, sandy glacial sediment in the Big Stone trench in northeastern South Dakota was dated $20{,}670^{+1500}_{-1000}$ B.P. (Gx-2741). Wood from the base of a layer of glacial sediment in northwestern Iowa was dated 20,000±800 B.P. (O-1325).

Fourteen additional dates from the

midcontinent region may be related to this interval, but they are here rejected because they are from dispersed organic material rather than wood and are therefore subject to contamination. They include GSC-653, GSC-711, GSC-1129, GSC-1279, GSC-1342, Gx-3530, I-1864A, S-176, S-206, S-228, S-228A, S-228B, S-494, and S-527 (table 1).

Material associated with ice margin 4 (fig. 32) has never been radiocarbon dated, but it seems most likely that advance 4 occurred during the earliest part of Late Wisconsinan time. Fresh constructional glacial topography occurs most places behind ice margin 4, suggesting that it is much younger than ice margin 3.

Advance 4 probably occurred at least several thousand years before advance 5. Interlocking supraglacial collapse features form a continuous network between ice margins 5 and 15 in North Dakota, suggesting that advances 5 to 15 all occurred during no more than a few thousand years. However, there is no evidence of stagnant ice existing behind ice margin 4 during advance 5 in North Dakota. This relationship seems to be duplicated in Iowa, where the Tazewell Advance (T in fig. 33), which occurred about 20,000 B.P. (Ruhe, 1969), was separated from the Bemis Advance (B in fig. 33) by a nonglacial period of at least 3,000 years. If North Dakota margin 4 does correlate with Iowa margin T, South Dakota margin 2 west of the James Lobe must be older than South Dakota margin 2 east of the James Lobe (Anonymous, 1971), as suggested in figure 34.

Ice margin 5 in North Dakota (fig. 32) has not been radiocarbon dated, so a chronologic interpretation depends on correlation with adjacent areas. No relevant radiocarbon dates from wood are known from the area to the northwest of North Dakota. In Iowa, eleven wood dates are from loess or gravel overlain by glacial sediment of the Bemis Advance (ice margin B in fig. 33): 16,720±500 B.P. (W-126), 16,367±1000 B.P. (C-528), 16,100±500 B.P. (I-1024), 16,100±1,000 B.P. (I-1270), 14,700±400 B.P. (W-153), 14,470±400 B.P. (W-512), 14,200±500 B.P. (I-1402), 14,042±1000 B.P. (C-664), 13,910±400 B.P. (W-517), 13,900±400 B.P. (I-1268), and 13,820±400 B.P. (W-513). The cluster of the six youngest dates around 14,000 B.P. indicates that this is a good estimate for the age of the Bemis Advance. Two dates (C-596 and C-653) suggest that this advance occurred even later, but they have been rejected by Ruhe (1969, p. 60-62) because they were determined using the "carbon black method." For the same reason, two dates (C-912 and C-913) from glacial sediment behind the Humboldt margin (H in figs. 34 and 35) were also rejected by Ruhe (1969, pp. 61 and 65). Because of the distance involved, correlation of ice margins from Iowa, through South Dakota, to North Dakota is uncertain. However, the most likely correlative of the Bemis ice margin in North Dakota is the Burnstad margin (margin 5, fig. 33). Therefore, the Burnstad Advance is interpreted to have occurred about 14,000 B.P.

Extending this correlation to the northwest, advance 1 in Saskatchewan (fig. 33) must also have occurred about 14,000 B.P. or perhaps later. This conclusion is at odds with the chronology worked out by Klassen (1972, 1975) and Christiansen (1979), which will be discussed below.

Two dates, 12,090±300 B.P. (W-635) and 11,540±200 B.P. (W-560), from wood in postglacial depressions in southwestern Minnesota indicate that the Bemis Advance (ice margin B, fig. 33) was over by about 12,000 B.P. This is backed up by a date of 11,480±300 B.P. (W-542) from wood in postglacial fluvial sediment behind ice margin 5 in south-central North Dakota.

The Missouri River assumed its present path during advance 5. Before that time it had flowed through the depression now occupied by the Van Hook arm of Lake Sakakawea at New Town in Mountrail County. The glacier dammed the Van Hook channel, forming Lake Crow Flies High (fig. 3; Clayton, 1972) at the maximum of advance 5. The lake overtopped a divide southwest of New Town, and the resulting spillway was from that time on occupied by the Missouri River. This, the narrowest segment of the present Missouri River trench, was the last part to form in North Dakota.

Although there are no radiocarbon dates for the early stages of deglaciation from margin 5 in North Dakota, correlations with South Dakota permit a tentative age to be assigned. Margins 8 and 9 are traced to margin 3a in South Dakota, which is traced around the James Lobe and into the Des Moines Lobe in Minnesota and Iowa, where it is correlated with margin G (Algona Advance, fig. 33). Seven dates from central South Dakota suggest an age of about 12,000 B.P. for margin 9 in North Dakota (fig. 35). These dates, which are between 12,600 and 12,000 B.P., are from wood in unoxidized sediment at depths of at least several metres. Most of these samples were within or beneath glacial sediment. A log dated 12,530±350 B.P. (W-987) and another dated 12,200±400 B.P. (W-1372) were from a depth of 12 metres in soil above oxidized glacial sediment and beneath unoxidized glacial sediment. Wood dated 12,050±400 B.P. (W-1189) was above gravel and beneath 58 metres of glacial sediment. Wood dated 12,340±300 B.P. (W-1756) was from the base of a layer of glacial sediment. Wood dated 12,200±400 B.P. (W-801), 12,330±180 B.P. (Y-452), and 12,520±100 B.P. (Y-925) was from within or beneath the surface glacial sediment. In addition, date 12,760±120 B.P. (Y-595) is from wood at a depth of 10 metres in surface fluvial sediment beyond margin 3a. These dates suggest that the James Lobe was advancing in southeastern South Dakota after about 12,500 B.P., and that advance 9 occurred about the same time in North Dakota.

However, two dates conflict with this interpretation. A log dated 12,680±300 B.P. (W-1757) was at a depth of 3 metres in surface fluvial sediment well behind margin 3a. This suggests that the dates mentioned in the previous paragraph are from wood buried by mass movement during melting of buried stagnant glacial ice and that advance 3a occurred before 12,700 B.P. In addition, a date of 12,970±250 B.P. (W-626) from wood in fluvial sediment immediately in front of ice margin G in northern Iowa was cited by Ruhe (1969, p. 62-65) as evidence that the Algona Advance occurred about 13,000 B.P.

The following dates are from postglacial sediment related to margins 5 through 9 but are not used in the glacial chronology because they are not from wood: I-1015, I-1019, I-1414, I-1416, I-1858, I-1859, I-6361, W-625, W-771, W-824, W-954, W-956, W-974, W-1019, W-1033, W-1149, W-1436, W-1755, W-2201, W-2305, and Y-1922.

Date 9,860±400 B.P. (W-1369) from wood under glacial sediment behind margin 9 seems to conflict with the suggested chronology. This date has been rejected because the sample was excavated several years before it was collected and dated and was therefore probably contaminated.

Before margin 9 was abandoned, meltwater drained southwestward across the Missouri Coteau to the Missouri River. On the Coteau, the meltwater rivers flowed across stagnant ice; the fluvial sediment later collapsed to form the gravel hills (Qcrh) characteristic of the Coteau. As the active-ice margin retreated northeastward off the Coteau, proglacial drainage shifted southeastward into the James Valley.

Both stratigraphic and geomorphic evidence indicates that margin 10 represents a significant glacial readvance. Transverse marginal ridges, drumlins, striations, eskers, and meltwater channels all indicate a major shift from southeasterly ice flow beyond margin 10 to southwesterly flow behind margin 10 in central North Dakota. The Kensal moraine, at margin 10, marks the western margin of the glacial sediment of the Dahlen Formation. Concentrations of glacial thrust masses along margins 10, 11, and 12 also reflect renewed glacial advance.

Advances 10, 11, and 12 have not been radiocarbon dated, but a series of dates from collapsed supraglacial sediment gives a minimum date for deglaciation. North Dakota date 10,330±300 B.P. (W-1818) is from a log at a depth of 4 metres in sand under glacial sediment, and 10,350±300 B.P. (W-1817) is from wood at a depth of 43 metres in clay under glacial sediment; these dates record the slumping of supraglacial sediment during the melting of buried stagnant ice long after active ice was at margin 11.

Eight other dates from postglacial sediment behind margin 2 in Saskatchewan are of questionable validity because they are not from wood: S-83, S-128, S-173, S-198, S-227, S-235, S-236, and S-457.

After the ice melted back from margin 9 on the Missouri Coteau, meltwater was channeled southeastward into the James River. The James channel in southeastern North Dakota is nearly as large as the later Sheyenne channel, indicating that, like the Sheyenne, it received meltwater from western Canada. The area south of the Turtle Mountains was apparently largely free of ice and accumulated a large amount of fluvial sand, which was overridden during advances 10, 11, and 12 (map unit Qcdg; see also Deal, 1971, p. 36-38). During the maximum of advance 10, the James River flowed against the ice margin in Foster and northern Stutsman Counties.

During wastage from margin 10 in North Dakota and margin 2 in Saskatchewan, meltwater drained from near Moose Mountain in southeastern Saskatchewan between the Missouri Coteau and the Souris Lobe by way of the Des Lacs and Souris spillways. Initially this drainage spilled by a series of channels in southern McHenry County across buried ice behind margin 11 to the Heimdal and James spillways. Until the stagnant ice behind margins 10 and 11 in Eddy County was breached, this water deposited the fluvial plain of Tiffany Flats. As the Souris Lobe wasted, Lake Souris came into existence as a series of discontinuous small lakes between the ice and the higher land behind margin 11. Drainage continued through the North Fork Sheyenne spillway (present elevation 478 metres) and Heimdal spillways until the Heimdal moraine (margin 11) was breached. Later, when the Cooperstown moraine was breached, drainage shifted from the Tiffany Flats and the James channel to the main Sheyenne channel in northeastern Eddy County.

In southeastern North Dakota, the James spillway initially emptied into Lake Dakota west of ice margin 10. As the ice margin retreated, drainage shifted from the James spillway to the Sheyenne spillway in front of ice margin 11. As the ice disintegrated behind margin 11, the Sheyenne spilled eastward at Fort Ransom between walls of stagnant ice to enter Lake Agassiz, which was in large part supraglacial at this time. At first the water flowed southeastward to the Minnesota River by way of the Milnor channel in northeastern Sargent County. Continued wastage of the ice opened a lower inlet to Lake Agassiz east of Lisbon, and the Sheyenne density-current fan began to form.

During this period of successively eastward shifts to lower channels at the southern end of the Sheyenne spillway, wastage of ice continued from margin 12 in the Lake Souris basin. Exposure of the Girard Lake spillway (present elevation 472 metres) resulted in the abandonment of the North Fork Sheyenne spillway. By this time, Lake Souris had expanded until it was a continuous lake stretching as far north as the west side of the Turtle Mountains in Manitoba and as far west as the present west bank of the Souris River near Verendrye. The west edge of the lake was bounded by stagnating ice.

There is little evidence that margin 13 represents anything more than a temporary halt or minor readvance of the ice. It is an obscúre margin compared to margins 10, 11, 12, and 14. Thrust blocks and overridden lake sediment (perhaps several hundred square kilometres in area) north of Devils Lake suggest at least minor readvance. Similarly, the thrust blocks in McHenry County in Black Butte, the Velva-Simcoe Hills, Henderson Hills, the hills associated with Buffalo Lodge Lake and Rock Lake near Riga, the bedrock hills covered with dunes south of Denbigh, and the small hills at Denbigh indicate at least minor readvance to margin 13. Much of margin 3 in Saskatchewan also appears obscure (Christiansen, 1956, 1960). However, where it impinges against the Missouri Coteau south of Moose Jaw, the extensive glacial thrust ridges of the Dirt Hills and Cactus Hills suggests a significant readvance in the area.

Two dates, 10,300±140 B.P. (S-110) and 10,000±300 B.P. (S-41) from wood in postglacial lake sediment, provide minimum ages for wastage of ice from margin 3 in Saskatchewan. Eight others are from bone, marl, organic mud, or shells and are therefore suspect: GSC-618, GSC-1332, S-123, S-174, S-190, S-246, S-248, and S-461.

As ice wasted from margin 3 in Saskatchewan, Lake Regina expanded to fill the depression that had been occupied by the Weyburn Lobe. Water from Lake Regina flowed down the Souris and Des Lacs spillways to Lake Souris, through the Girard Lake spillway, the Sheyenne spillway to Lake Agassiz, and down the Minnesota River trench to the Mississippi River.

As ice in the Red River Valley wasted, Lake Agassiz expanded northward. The offshore silt and clay in the lower part of the Argusville Formation was deposited in the southern part of the basin, and the offshore sediment in the Wylie Formation was deposited in the northern part of the basin in North Dakota and Minnesota (Arndt, 1977).

Reactivation of the glacier resulted in a readvance of more than 100 kilometres to margin 14, depositing the glacial sediment of the Huot and Falconer Formations over the Wylie Formation (Arndt, 1977). Lake Souris continued to drain through the Girard Lake spillway, possibly eroding its floor to an elevation of about 463 metres.

Wastage of the glacier from margin 14 resulted in the exposure of an outlet for Lake Souris into the Pembina spillway. This produced a drop in level of Lake Souris to about 457 metres and the abandonment of the Girard Lake spillway. The accompanying decrease in discharge and sediment load in the Sheyenne spillway resulted in deposition of a fine-grained unit overlying the coarser grained basal sediment in the Sheyenne River density-current fan.

The Pembina spillway entered the Red River Valley, where it flowed southward down Elk Valley (fig. 3) between the Pembina Escarpment and the Edinburg moraine at the west edge of the glacier (margin 14). As much as 30 metres of fluvial sediment was deposited in Elk Valley, and a fan of density-current sediment was deposited southward into the main body of the lake. As the ice wasted back from margin 14, the river breached the Edinburg moraine just east of Fordville to cut the trench later occupied by the Forest River. Continued wastage caused the abandonment of the rest of the Elk channel when the Pembina spillway emptied directly into Lake Agassiz, depositing the density-current fan south of Walhalla. Throughout this period, deposition of the Argusville Formation continued south of the Edinburg moraine. Following the wastage of the ice, the Brenna Formation was deposited north of the Edinburg moraine.

In the Lake Souris basin, the level of the lake dropped to an elevation of about 450 metres as the floor of the Pembina spillway was eroded. This drop resulted in the deposition of nearshore sand and river sand and gravel over previously deposited offshore sediment. This sequence can be seen in numerous bore holes around Upham in northern McHenry County.

The glacier again reactivated and advanced to margin 15. The glacial sediment of the Marchand Formation, in southern Manitoba, northeastern North Dakota, and northwestern Minnesota, was deposited during this advance (Moran and others, 1976). This advance probably blocked the Pembina spillway north of the Turtle Mountains, causing the level of Lake Souris to return to about 463 metres, the level of the Girard Lake spillway. Fine-grained offshore sediment was deposited over nearshore sand in northern McHenry County and over fluvial gravel in parts of northern Bottineau County, such as near Landa. Lake Souris drained through the Sheyenne spillway to Lake Agassiz, depositing coarser sediment once again on the Sheyenne density-current fan. The floor of the Girard Lake spillway was lowered to about 460 metres during this period. Because the Pembina spillway was blocked, the sediment load supplied to the Pembina density-current fan decreased abruptly, resulting in the deposition of a finer grained unit. The northeast edge of the fan, called First Mountain, was banked against the glacier at this time; some beds of fluvial gravel derived from the ice are interbedded with finer grained density-current sediment in this area.

Wastage of ice from margin 15 reopened the Pembina spillway and Lake Souris dropped again to 450 metres. Coarser sediment was briefly delivered to the Pembina density-current fan of Lake Agassiz before continued wastage of the glacier near Brandon, Manitoba, exposed a lower outlet for Lake Souris into Lake Agassiz, and flow down the Pembina spillway ceased. The combined drainage of the Souris and Assiniboine spillways began to build the Assiniboine density-current fan in Lake Agassiz east of Brandon.

Since its beginning, Lake Agassiz had drained southward through the Minnesota River trench. The ice continued to waste northward into Manitoba and northeastward into northern Ontario, until a series of lower outlets were exposed north of Lake Superior. Once these outlets were reached, the level of Lake Agassiz dropped from the Campbell level until most of the Red River Valley was dry land. The Sheyenne and Assiniboine Rivers cut trenches through their density-current fans. The Sheyenne River flowed across the exposed floor of the lake to be joined by numerous tributaries before reaching the lake and forming a series of muddy deltas similar to the present Red River delta north of Winnipeg. The organic sand, silt, and clay of the Poplar River Formation was deposited on the Brenna and Argusville Formations throughout the Red River Valley by these streams (Arndt, 1977).

Renewed advance of the glacier in northern Ontario blocked the low outlets, causing Lake Agassiz to return to the Campbell level and again drain through its southern outlet into the Minnesota River trench. During this phase of the lake, the offshore silt of the Sherack Formation was deposited throughout the Red River Valley. Following wastage of ice back from the low-level eastern outlets, Lake Agassiz once again drained out of the Red River Valley of North Dakota for the last time.

The chronology of wastage from margin 14 is based on a group of minimum dates from Manitoba, Saskatchewan, and North Dakota. The oldest reliable dates are 12,025±205 B.P. (S-553) from wood at a depth of 50 metres beneath Qu'Appelle River alluvium in Saskatchewan and 11,600±430 B.P. (GSC-1081) from wood fragments at a depth of 18 metres below the present flood plain of the Assiniboine River near Virden, Manitoba. On the basis of elevation relationships with respect to the present surface of the Assiniboine density-current fan, the Virden date is probably related to the building of the fan. Related dates are 10,690±190 B.P. (GSC-677) from wood in Assiniboine River alluvium north of Virden; 10,550±200 B.P. (Y-411) from wood in alluvium near Lavenham, Manitoba; 10,000±280 B.P. (GSC-1428) from wood in Assiniboine alluvium between Virden and Brandon; and 10,000±150 B.P. (GSC-870) and 9,700±140 B.P. (GSC-797) from wood in Assiniboine alluvium near Rossendale, Manitoba. A series of related dates are rejected because they are not from wood and are therefore subject to contamination. One, 14,300±320 B.P. (GSC-1369), is from organic detritus in overbank sediment in southern Manitoba; because this date is about 300 years older than margin B in Iowa and margin 5 in North Dakota, the date is clearly too old if the overbank sediment was deposited after advance 0 in Manitoba. Other rejected dates include GSC-383, GSC-492, GSC-689, GSC-902, GSC-1319, I-1682, I-2106, I-3157, I-3475, I-3476, Y-165, and Y-166.

A date of 9570±130 B.P. (S-129) from wood in postglacial pond sediment on Riding Mountain in Manitoba gives a minimum age for wastage from margin 0 in Manitoba (fig. 33).

A series of dates from wood in shoreline and offshore sediment deposited in North Dakota during a period of drainage of Lake Agassiz through low-level eastern outlets indicates that the ice wasted from margin 15 considerably before 10,900 B.P.: 10,960±300 B.P. (W-723), 10,820±190 B.P. (TAM-1), 10,340±170

B.P. (I-5213), 10,080±280 B.P. (W-900), 10,050±300 B.P. (W-1005) and 9900±400 B.P. (W-993) (fig. 8). Dates of 9820±300 B.P. (W-1361) and 9810±300 B.P. (W-1360) from wood in shoreline sediment in North Dakota, 9730±160 B.P. (I-5123C) and 9650±150 B.P. (I-5123) from a single piece of wood in offshore sediment in North Dakota, 9940±160 B.P. (I-3880) from wood in shoreline sediment in Minnesota, and 9990±160 B.P. (GSC-391) from wood in shoreline sediment in Manitoba indicate the beginning of the Emerson Phase, when the eastern outlets were blocked again and Lake Agassiz returned to the Campbell level (fig. 8). Dates 9130±150 B.P. (I-1982) from wood in offshore sediment in North Dakota and 9200±600 B.P. (W-1057) from wood in shoreline sediment in Minnesota mark the end of the Emerson Phase of Lake Agassiz (fig. 8).

Wright (1972, p. 543) stated that Lake Agassiz began just before 11,740±200 (Y-1327), the age of organic sediment deposited in a pond on one of the highest beaches of Lake Agassiz. "Such a pond is not subject to problems in many lakes caused by persistence of buried ice, because the lake water had caused the melting of all buried ice along the shores." However, the upper beaches of Lake Agassiz, like those of other large glacial lakes in the midcontinent area, were deposited at least in part on stagnant glacial ice. The pond is an ice-block depression, and the deposition of the pond sediment began long after the beaches formed. In addition, the date is from fine-grained organic sediment, which was probably contaminated with older carbon.

Chronologic Summary

Using only those dates that are believed to be valid because of an unambiguous stratigraphic setting and a slight probability of contamination, a chronologic summary has been constructed (fig. 35). This chronology is based almost entirely on wood dates.

Advances 1, 2, and 3 have not been radiometrically dated. The Late Wisconsinan advance, sometime between 28,000 B.P. (W-2450) and 20,000 B.P. (O-1325) may correspond to the Long Lake Advance (margin 4, figs. 32 and 33). After a period of roughly 6,000 years, for which there is no evidence of events in North Dakota, the glacier advanced to margin 5, at about 14,000 B.P. (based on 11 dates in Iowa). During the next several hundred years, the ice wasted back about 50 kilometres, with minor readvances to margins 6, 7, 8, and 9 (W-626).

There are no dates relating directly to any of the remaining glacial events in the region. Dates associated with Lake Agassiz and with rivers draining into the lake require that advance 15 occurred before 11,600 B.P. (GSC-1081), or at very latest, 11,000 B.P. (W-723). The remaining glacial events are guessed to have occurred at about the following times.

Wastage of ice from margin 9 after 13,000 B.P. was interrupted by a readvance of at least several tens of kilometres to the Kensal margin (margin 10), followed by three readvances to margins 11, 12, and 13 over a period of a few hundred years. Continued wastage nearly to the international border was interrupted by a readvance of roughly 100 kilometres to the Edinburg moraine (margin 14) about 12,000 B.P. This was followed by rapid wastage and a brief readvance to margin 15.

Alternate Chronologies

The preceding chronology is based on wood dates because of a conservative assumption that any material except wood was probably contaminated. Starting with the less conservative assumption that all dates are valid without direct evidence of contamination, Moran and Clayton (1972) presented an alternative chronology. They argued that wastage of the glacier was extremely rapid following its maximum advance into Iowa about 14,000 B.P. Dates of 13,500±220 B.P. (I-2289) and 12,000±250 B.P. (I-2537) from shells in Sheyenne River sediment at the west edge of the Agassiz plain in southeastern North Dakota, 12,800±350 B.P. (I-1682) from gyttja at the Glenboro site in Manitoba, and 12,400±420 B.P. (Y-165) and 12,100±140 B.P. (GSC-1319) from peat in alluvium at the west edge of the Agassiz plain in southern Manitoba require ice-free conditions, at least along the major streams, as early as 13,500 B.P. in southeastern North Dakota and 12,500 B.P. in southern Manitoba. This requires marginal retreat of at least 1 kilometre a year. Rapid deglaciation was also required by the chronologies of Moran and others (1976) and Teller (1976). However, as pointed out at the beginning of the section on chronology, we now conclude that this interpretation was based on samples that can be shown to be contaminated or are likely to be contaminated.

Similarly, Klassen (1972, 1975) and Christiansen (1979) have constructed chronologies (fig. 34) with events 3,000 to 4,000 years earlier than suggested here (fig. 35). Klassen (1972, p. 552) stated that "scattered radiocarbon dates from valley fills and interfluve areas (table 1) indicate that the western part of the Assiniboine Basin was largely deglaciated more than 14,000 years ago and most parts of the Assiniboine drainage were ice-free some 13,000 years ago." Klassen (1972, p. 553; 1975, p. 47) dated margin 0 (fig. 33 of this report) at 15,000 to 14,500 B.P. because dates GSC-1369 and I-1682 "suggest that vegetation had invaded this region by 14,000 and 13,000 years ago." These conclusions are based on samples that were probably contaminated with dead carbon.

Christiansen (1979) concluded that advance 4 (fig. 33 of this report) occurred about 14,000 B.P., based on three radiocarbon dates: 14,300±350 B.P. (GSC-1369) from organic debris in alluvium in southern Manitoba (mentioned above), 14,040±465 B.P. (S-685) from bone under 1 metre of boulders near Saskatoon, and 13,600 B.P. (extrapolated from dates S-522 through S-526) from gyttja in Crater Lake near Yorkton. All are from materials that are commonly subject to contamination. We here suggest that they have in fact been contaminated with older carbon because the dates are in conflict with the soundest group of dates in the midcontinent region, which indicate that the Des Moines Lobe was advancing into central Iowa in 14,000 B.P. The Des Moines Lobe has generally been considered to have been supplied by ice moving down the Red River Valley from Manitoba (see, for example, Prest, 1969), and glacial accumulation has been assumed to have begun many hundreds of kilometres back from the ice margin (Sugden, 1977). If this were true, advance 1 in Saskatchewan (figs. 33 and 34) must have occurred after 14,000 B.P.

However, Hooke and others (1976) have suggested that near-margin accumulation was significant in the Des Moines Lobe in Late Wisconsinan time. If accumulation were great enough on the Des Moines and James Lobes, they could have functioned as separate ice caps that were advancing southward while Lake Agassiz was in existence. The Des Moines Ice Cap and the James Ice Cap would also have flowed northward toward Lake Agassiz. Ice-margin indicators should be present to support this hypothesis, but none are known.

Christiansen (1979) has listed no radiocarbon dates associated with margins 1, 2, and 3 (fig. 33). His suggested ages (fig. 34) are apparently extrapolations. Margin 5 is apparently also undated, and the age of margin 6 (fig. 34) is based on date 12,025±205 B.P. (S-553). This date is from wood at a depth of 50 metres in Qu'Appelle River sediment. Christiansen suggested that this sediment was deposited after the Qu'Appelle spillway was abandoned shortly after advance 6, but it seems as likely that it was deposited at the beginning of meltwater flow during phase 4, for example.

OAHE FORMATION

Stratigraphy

The Oahe Formation was defined by Clayton, Moran, and Bickley (1976) to include all the well-sorted silt above the Coleharbor Group near Riverdale in McLean County. It has been redefined by Clayton and Moran (1979) to include material of all grain sizes above the Coleharbor Group. The Oahe differs from the Coleharbor in lacking glacial sediment. In general, the nonglacial sediment at the top of the Coleharbor can be distinguished from the sediment of the Oahe by its better sorting and lack of dispersed organic material.

The Oahe in many places can be subdivided into four members (Clayton, Moran, and Bickley, 1976). The lowest, the Mallard Island Member consists of coarse sediment with little organic material. It is by definition part of the Oahe, because it overlies type Coleharbor (Bluemle, 1971) and is in the Oahe type section and was included in the Oahe in the type description (Clayton, Moran, and

Bickley, 1976) and redefinition (Clayton and Moran, 1979). In most places it seems more appropriate to include it in the Oahe than in the Coleharbor, because it is generally more similar in grain size to the overlying sediment than to the underlying sediment. Nevertheless, it could be argued that its lack of organic material makes it more similar to the Coleharbor than the Oahe; no satisfactory criteria have been determined for distinguishing sediment of the Mallard Island Member from nonglacial sediment not overlain by glacial sediment in the Coleharbor Group.

The Aggie Brown Member overlies the Mallard Island. It is characterized by abundant organic material. The overlying Pick City Member is coarser grained and has less organic material than the Aggie Brown Member. The uppermost member, the Riverdale, is finer grained and has more organic material than the Pick City Member.

Much of the Mallard Island Member was deposited in Late Wisconsinan time. The Holocene/Wisconsinan boundary (here arbitrarily placed at 10,000 B.P.) is probably near the base of the upper submember of the Aggie Brown Member. The Pick City Member was probably deposited in Middle Holocene time (8500 to 5000 B.P.), and the Riverdale Member was deposited in Late Holocene time (after 5000 B.P.). These lithostratigraphic boundaries are all time transgressive, however. In general, the contact between the upper two members becomes younger from the northeast to the southwest, and the contacts between the Pick City Member, Aggie Brown Member, Mallard Island Member, and Coleharbor Group become older from the northeast to the southwest, but there are significant exceptions to this generalization; for example, the gravel in the lower beaches of Lake Agassiz is Early Holocene (10,000 to 8500 B.P.) but is clearly part of the Coleharbor rather than the Oahe because of its excellent sorting and lack of organic material. The Riverdale Member is the surface unit in most places where the Oahe Formation is mapped.

In most places these members are too thin to be mapped separately at the scale of the Geologic Map of North Dakota or at the scale of existing county maps. However, lithogenetic subdivisions (river, pond, and windblown sediment) can be mapped. The members can be recognized within these subdivisions.

River Sediment

River sediment in the Oahe Formation consists of as much as 10 metres of channel and overbank sediment. The overbank sediment is well exposed in cutbanks. It consists of poorly sorted, obscurely bedded clay and silt with some thin layers of sand, weak paleosols, scattered mammal bones and teeth, terrestrial snail shells, and fragments of wood. The underlying poorly exposed channel sediment in most places consists of cross-bedded sand, but plane-bedded gravel does occur in alluvial fans at the base of the steepest scarps. Most of the exposed river sediment of the Oahe is in the Riverdale Member.

Pond Sediment

Most ponds (sloughs) in northeastern North Dakota contain a few metres of silt and clay (Bickley and Clayton, 1971). The Mallard Island Member generally consists of sandy silt with little organic material. The lower submember of the Aggie Brown Member contains more organic material than other units in the Oahe Formation. In much of northeastern North Dakota, the Aggie Brown Member contains finely bedded clay with abundant fossil fish, algae, mollusks, and insects; in the Turtle Mountains it contains peat and marl. The upper submember of the Aggie Brown Member and the Riverdale Member contain obscurely bedded, dark-colored silty clay with fossil snails. The Pick City Member contains light-colored sandy silt. The stratigraphy of the ponds of southwestern North Dakota has never been investigated.

Wind-Blown Silt

The Oahe Formation contains as much as 6 metres of wind-blown silt (loess) with obscure bedding on level surfaces near the Missouri River, in McLean, Mercer, Oliver, Burleigh, and Emmons Counties (Clayton and others, 1976). The four members of the Oahe are of roughly equal thickness in these areas. The Mallard Island Member consists of yellowish coarse silt; it is thickest on the oldest surfaces and is thin or absent on the latest Wisconsinan or younger surfaces. The lower submember of the Aggie Brown Member consists of reddish silt (possibly a fossil forest soil). The upper submember consists of black fine silt (possibly a fossil prairie soil). The Pick City Member consists of light-gray coarse silt. The Riverdale Member consists of dark-gray fine silt or dark-gray fine silt alternating with light-gray coarse silt.

In the northeastern part of the state most surfaces formed in Late Wisconsinan time are blanketed with 0.3 to 1.0 metre of wind-blown sediment; and in the southwestern part of the state level upland surfaces are blanketed with as much as 2 metres of wind-blown sediment, which has never been mapped in these areas. Where the sediment is thickest, silt predominates, and the members can be distinguished. Where it is thinnest, silt, clay, and sand are equally abundant and no members can be distinguished. On steeper slopes the wind-blown sediment is mixed with slopewash sediment.

Wind-Blown Sand

The largest areas of wind-blown sand in North Dakota occur where well-sorted sand has accumulated in glacier-dammed lakes in the upper parts of density-current fans. The wind-blown sand in the southwestern part of the state is derived from Paleocene formations or from fluvial sediment derived from Paleocene formations.

Map unit Qod includes areas of thick wind-blown sand with knobby

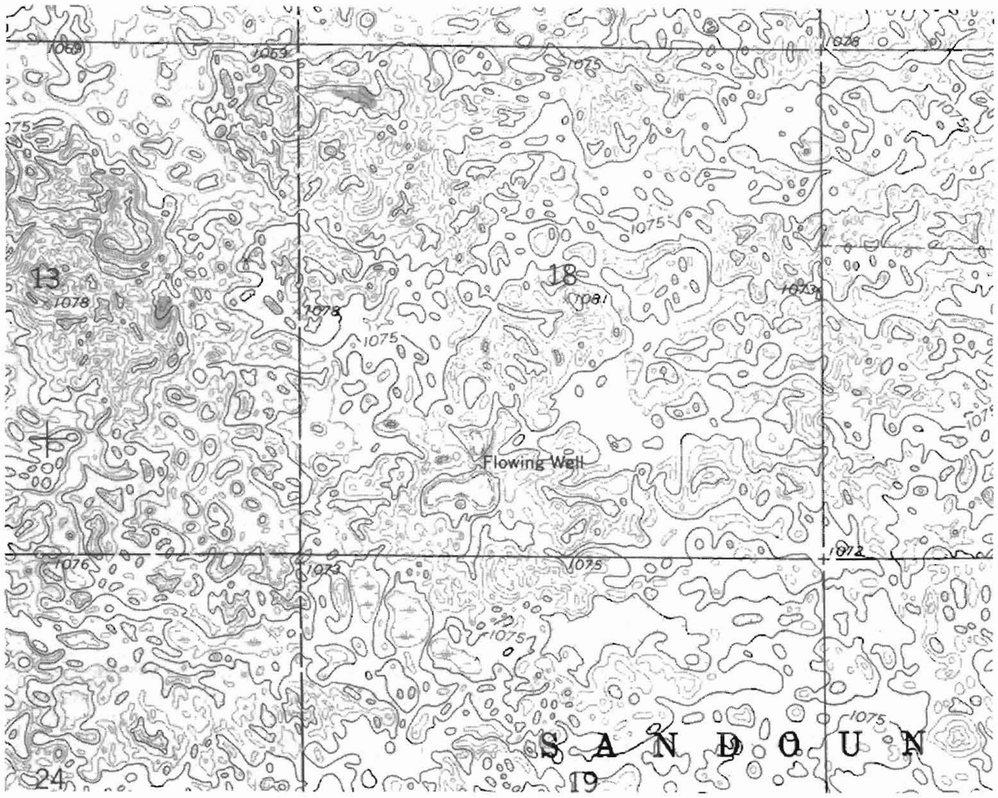

Figure 36. Blowout dunes (map unit Qod) in T134N, Rs53 and 54W, in eastern Ransom County. U.S. Geological Survey Venlo Quadrangle (5-foot contour interval). Area shown is 3.2 kilometres wide.

dune topography (fig. 36). In some areas transverse or longitudinal dunes, probably formed in Middle Holocene time, can be distinguished. In most areas, however, these dunes have been nearly obliterated by blowouts, probably during Late Holocene time.

Most outcrops expose well-sorted, medium sand, with obscure bedding and poorly developed paleosols. This sand, which is part of the Riverdale Member, is blowout sediment deposited during Late Holocene time. Similar sediment in southern Manitoba has been studied by David (1971).

Cross-bedded sand without paleosols is rarely exposed. It is part of the Pick City Member or Coleharbor Group and is sediment of transverse or longitudinal eolian dunes deposited in Middle Holocene time.

Map unit Qou includes areas of thin wind-blown sand with undulating to flat topography. In contrast to areas mapped as Qod, much of this area has been cultivated. The wind-blown sand is probably generally less than 3 metres thick. In many areas it is lacking, and the sand at the surface consists of density-current or fluvial sand of the Coleharbor Group or sand of pre-Pleistocene formations. Sand of the Oahe Formation and sand of older formations have been left undifferentiated in map unit Qou because they have been inadequately differentiated

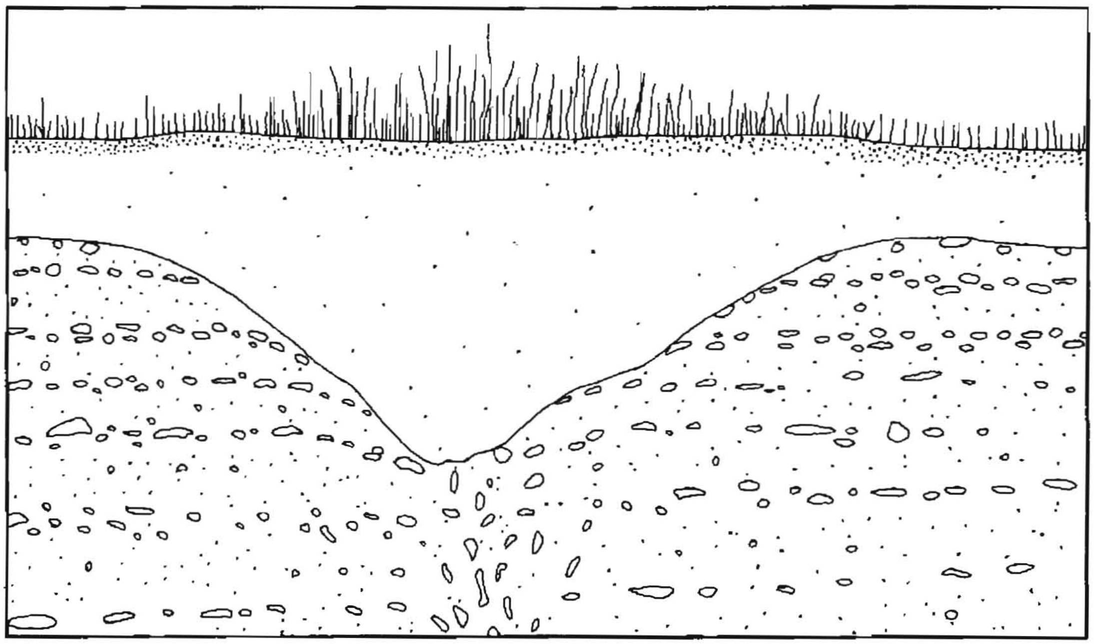

Figure 37. Cross section of a typical fossil ice wedge in southwestern North Dakota. Height of section is about 2 metres. The fluvial gravel slumped into the fissure left by the melting ice wedge, resulting in disrupted bedding with vertically oriented pebbles and cobbles. Wind-blown silt fills the resulting trough. During dry periods, vegetation is most lush where the moisture-retaining silt is thickest.

on previous maps and because they are indistinguishable on airphotos. The characteristic feature of map unit Qou is a low-relief surface with vague northwest-southeast lineations formed by wind scour.

OTHER MAPPED FEATURES

The following features shown on the Geologic Map of North Dakota cannot be conveniently grouped with any of the formations previously discussed.

Permafrost Polygons

Fossil ice-wedge polygons occur throughout southwestern North Dakota, typically in sand and gravel on remnants of pediments, alluvial fans, and river terraces (Clayton and Bailey, 1970). Cross sections of the wedges are commonly seen in gravel pits (fig. 37). The typical wedge is a zone that lacks bedding and has vertical foliation resulting from the growth of the ice wedge and from the overlying sand and gravel slumping into the crack as the ice melted out. This zone is generally less than 1 metre wide, and it extends to depths of at least 5 metres, but gravel pits are rarely deep enough to expose the bottoms of the deepest wedges.

The top of the wedge generally widens into a trough as the result of near-surface sand and gravel running into the underlying crack as the ice wedge melted out. The sides of the trough are near the angle of repose. The top of the trough is from 0.5 to 5 metres wide, averaging about 2 metres.

The sand and gravel containing the trough are generally overlain by about 1 metre of wind-blown sandy silt. The silt fills the trough, leaving a flat surface; the wedge and its trough generally have no topographic expression. The Manning Series is the surface soil mapped in most areas where fossil ice wedges occur (Larson and others, 1968).

Polygons have been seen at about 200 sites on 1:20,000 or larger scale airphotos in southwestern North Dakota; three of these sites are shown in figure 38. The polygons generally have four or five sides and are from 10 to 100 metres in diameter. Because they generally have no topographic expression, they are visible only when the soil is more moist or the vegetation is thicker over the fossil ice wedges. The wind-blown silt is thicker in the troughs, resulting in more available moisture. For this reason, the polygons are visible only during dry periods, and probably only a small portion of the existing polygons have been observed and plotted on the Geologic Map of North Dakota.

Similar fossil ice-wedge polygons have been identified in Ontario by Morgan (1972) and in England by West (1969, p. 85). They are similar to ice-wedge polygons found in Alaska and other northern areas (see, for example, Pewe and others, 1969). These polygons are not drying cracks, as suggested by Denny and others (1968, p. 47) for similar features in Montana, because they are in sand and gravel rather than silt and clay. They are probably not "sand-wedge" polygons (Black, 1976) because the presence of the overlying troughs indicate collapse when the ice wedges melted.

Ice-wedge polygons form today in areas having continuous permafrost colder than -5°C (Black, 1976, p. 9), such as northern Alaska. The late Cenozoic paleontology and paleoecology of southwestern North Dakota is largely unstudied, and as a result little or no other evidence of a tundra climate has been documented. Possible solifluction landforms and deposits have been noted (Bell, 1972, considered them to be glacial features), but have never been studied in detail; they probably existed throughout the area but have since been largely eroded away. The polygons have been preserved because they are in sites that escaped erosion where the soil is so permeable that there has been little runoff.

The exact age of the polygons is unknown because the Pleistocene stratigraphy of southwestern North Dakota is largely unstudied. Shells of more than a dozen species of mollusks have been found in the sand and gravel cut by the ice wedges at three sites (Tuthill and others, 1964; Bailey, 1970). None is extinct, with the possible exception of Pupilla sinistra, which was thought to be restricted to "Yarmonthian interglacial beds." The polygons are absent on surfaces formed during latest Wisconsinan time throughout North Dakota. They occur on stream gravel as young as the Napoleon Glaciation, which may have occurred during Early Wisconsinan time (Clayton, 1969; Bickley, 1972). The

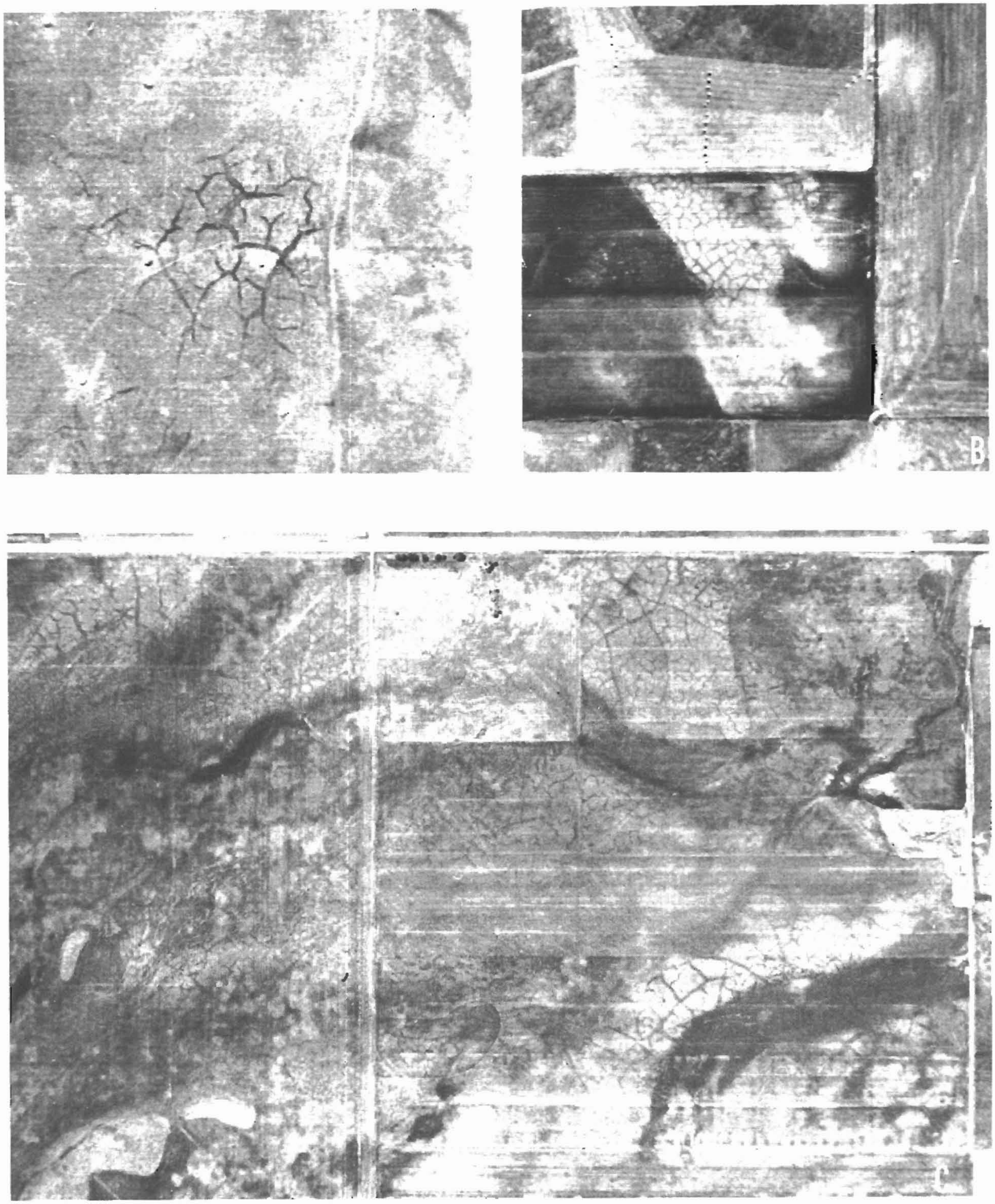

Figure 38. Fossil ice-wedge polygons in southwestern North Dakota. a: Center of section 32, T130N, R86W, Sioux County. U.S. Department of Agriculture airphoto BAA-4T-28. The area shown is 0.3 kilometres wide. b: Near the center of section 13, T137N, R94W, Stark County. U.S. Department of Agriculture airphoto AZV-2FF-114. Area shown is 0.3 kilometres wide. c: The southwestern quarter of section 15, T131N, R92W, Adams County. U.S. Department of Agriculture airphoto AZX-2T-19. Area shown is 1.4 kilometres wide. South is up.

wind-blown silt in the troughs overlying the fossil ice wedges may be part of the Mallard Island Member of the Oahe Formation, which was probably deposited during the last part of Late Wisconsinan time. Therefore, the youngest polygons may have formed during the first part of Late Wisconsinan time.

Two other kinds of polygons, which are unrelated to permafrost, occur in North Dakota. Soil polygons are forming today where surface soil falls down drying cracks; these polygons are typically about 1 metre in diameter in groundwater recharge areas (Clayton, 1962, p. 14, fig. 2) and about 3 metres in diameter in groundwater discharge areas, for example, on soil of the Fargo Series (Omodt and others, 1968, p. 26). Horberg (1951) interpreted markings on the plain of Lake Agassiz to be ice-wedge polygons, but they have since been interpreted to be ice-drag marks (Clayton and others, 1965); they formed when Lake Agassiz last drained, about 9000 B.P., when the climate was much too warm for permafrost. Some Lake Agassiz markings have been interpreted by Bluemle (1973, p. 31) to be ridges squeezed up between blocks of ice as the lake drained.

Blowouts

Large blowout depressions are indicated on the Geologic Map only in areas lacking wind-blown sand (they are also abundant in areas mapped as Qod and Qou). They occur in southwestern North Dakota on upland stability surfaces, far from major rivers.

The blowouts are commonly between a few tens of metres and several hundred metres across. Many are several metres deep, lack outlets, and contain ponds or marshes.

In some areas, the topography is almost entirely the result of eolian erosion (fig. 39). South of Rainy Buttes, in southeastern Slope County, interfluve areas a few kilometres wide have totally nonintegrated drainage; integrated drainage occurs for only a few hundred metres on either side of the major streams. The topography there could be mistaken for glacial topography if it were in an area that had been glaciated. However, it is 70 kilometres beyond the southwesternmost glacial erratics.

Most of the blowouts are elongated in a northwest-southeast direction, parallel to sandblast scour marks on tops of Paleocene chert beds capping some of the upland surfaces in the area. The sandblast marks indicate that the prevailing winds were from the northwest, as they are today.

Little or no wind-blown sand has been observed in the blowout areas. It was probably carried away by the streams flowing through the blowout areas.

The age of the blowouts is unknown, but they occur primarily on stability surfaces formed before the Napoleon Glaciation (Clayton, 1969; Moran and others, 1976, p. 147). The blowouts occur in areas that are culti-

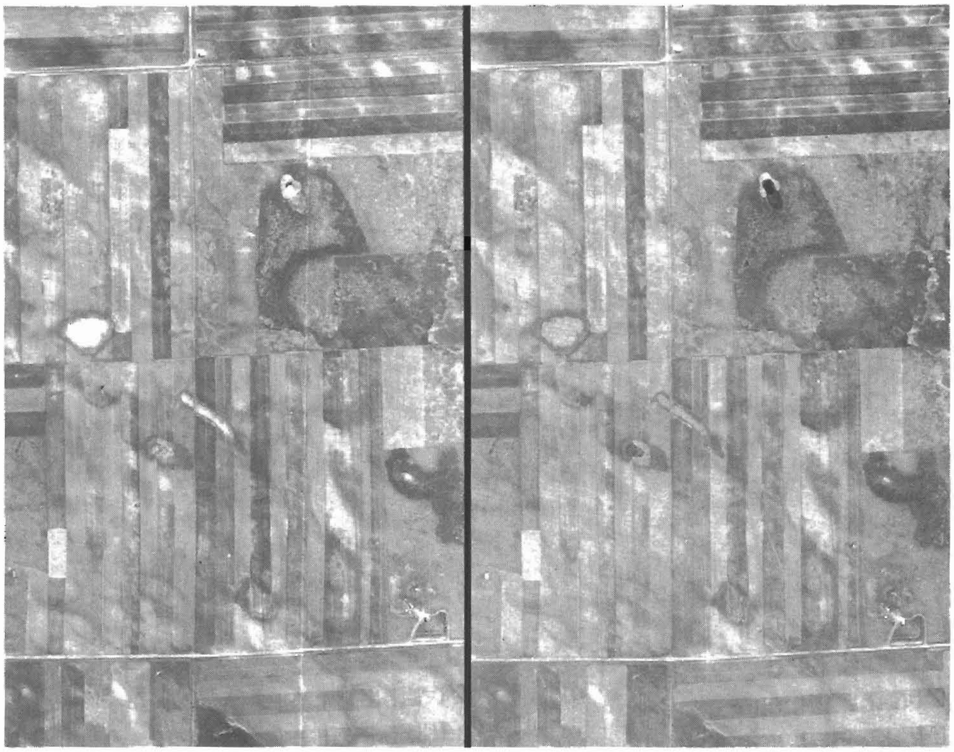

Figure 39. Blowouts in sections 34 and 35, T134N, R98W, in southeastern Slope County. U.S. Department of Agriculture airphotos AXE-1T-4 and -6. Area shown is 1.3 kilometres wide. South is up.

vated but undergo little eolian erosion, indicating that they formed during periods that were considerably drier than today.

Landslides

Only the larger areas of landslides, more than about 0.8 kilometre across, have been indicated on the Geologic Map of North Dakota. All are based on airphoto interpretation. Many of the older, now-stable landslides were probably overlooked because they no longer retain their characteristic topography.

Spring Pits

Spring pits have been indicated on the Geologic Map in eastern Grand Forks (fig. 40) and Walsh Counties. They are irregular depressions as much as 2 kilometres wide and 5 metres deep. Springs are abundant in the marshy bottoms of the depressions.

Downey (1973, p. 54-55) determined that the spring pits in Walsh County overlie several tens of metres of sand and gravel connecting them to the aquifer in the Cretaceous Dakota Group. The sand and gravel is probably part of a complex of clastic plugs

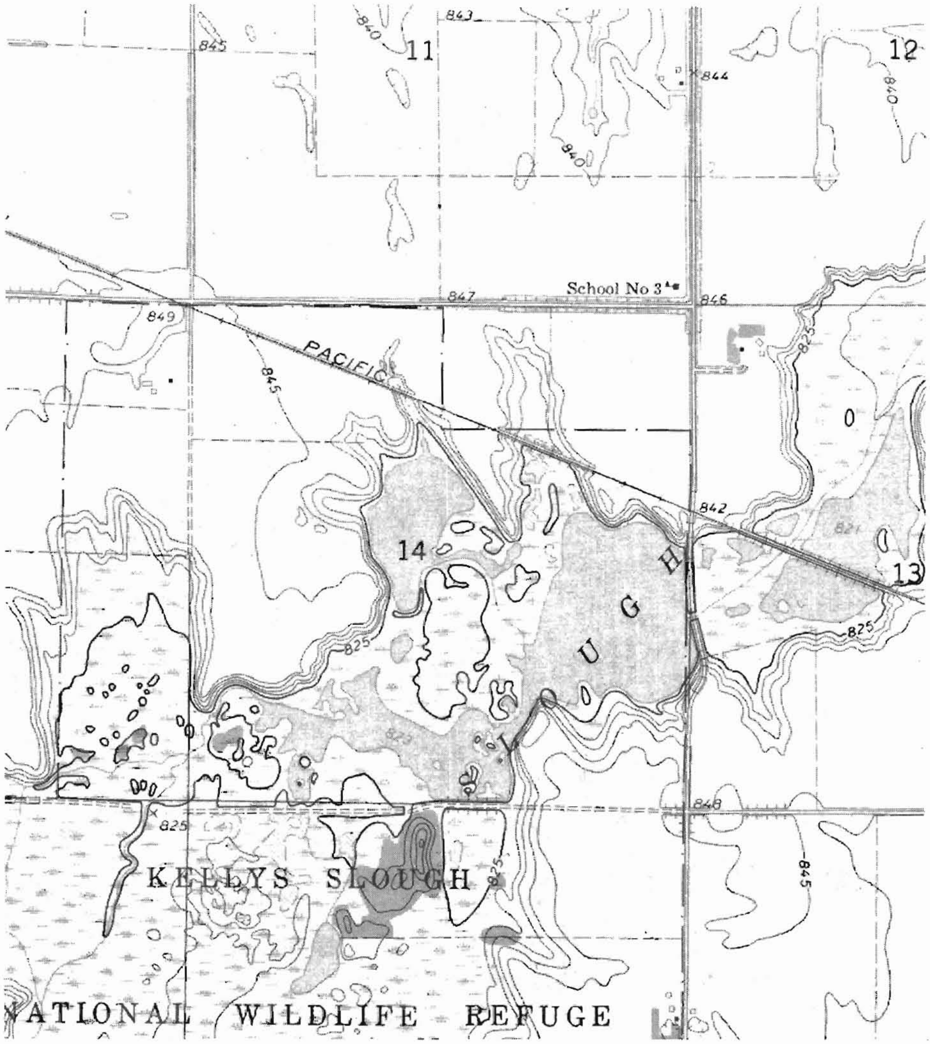

Figure 40. Depression (Kellys Slough) sapped by springs in the plain of Lake Agassiz in T152N, R52W, in eastern Grand Forks County. U.S. Geological Survey Emerado Quadrangle (5-foot contour interval). Area shown is 3.0 kilometres wide.

deposited by upward-moving groundwater. The clastic plugs probably formed near the beginning and end of each glacial advance across the area, when groundwater flow was most intense as the result of the large heads produced by the weight of the glaciers. The pits have been cut into the Sherack Formation. They therefore formed during or after the Emerson Phase of Lake Agassiz.

Badlands

Areas of badland topography (fig. 41) on the Geologic Map were determined by airphoto interpretation. Badlands have steep slopes with little vegetation because of rapid soil erosion. The largest area of badlands in the state occurs along the Little Missouri River; it is generally called "the Badlands" or "the Little Missouri

Figure 41. Badlands in the Sentinel Butte Formation in sections 28 and 33, T147N, R98W, in southern McKenzie County. U.S. Department of Agriculture airphotos AXD-3V-209 and -210. Area shown is 1.3 kilometres wide. South is up.

Badlands." Other small patches of badlands occur throughout the southwestern part of the state. Large areas of badlands occur on the Hell Creek, Ludlow, Slope, Bullion Creek, and Sentinel Butte Formations and on the White River Group.

Mines

Although some open-pit lignite mines are more than 0.8 kilometre across and therefore mappable, none has been indicated on the Geologic Map, because they have been rapidly increasing in size in recent years, and their indicated distribution would soon be out of date. Instead, an attempt has been made to indicate the geologic material that was present before the area was mined.

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