NORTH DAKOTA GEOLOGICAL SURVEY

**Wilson M. Laird, State Geologist**

MISCELLANEOUS SERIES 30

# GLACIAL GEOLOGY OF THE MISSOURI COTEAU AND ADJACENT AREAS

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**CHIEF EDITORS**

**GUIDEBOOK AND MISCELLANEOUS SHORT PAPERS PREPARED FOR THE 18TH ANNUAL FIELD CONFERENCE OF THE MIDWEST FRIENDS OF THE PLEISTOCENE IN SOUTH-CENTRAL NORTH DAKOTA, 19 TO 21 MAY 1967**

**GRAND FORKS, NORTH DAKOTA**

**1967**

## PREFACE

The Friends were last in North Dakota in 1958. At that time Colton and Lemke had finished the first comprehensive reconnaissance of the Pleistocene geology of the state (North Dakota Geological Survey Miscellaneous Publication 10 and U. S. Geological Survey Miscellaneous Geological Investigation Map I-331).

Since then, the Pleistocene geology of half of the 53 counties in the state have been mapped by the North Dakota Geological Survey, the U. S. Geological Survey, and the State Water Commission. County reports, with geologic maps at a scale of 1:126,720, are being jointly published as Bulletins of the North Dakota Geological Survey and County Ground Water Studies of the State Water Commission. Field work for the entire Missouri Coteau, except for about 10 townships, has now been completed. For this reason, and because many of the glacial-stagnation features found on the Coteau are not well known outside of this area, this year's Friends of the Pleistocene field trip will be a review of the late Pleistocene glacial stagnation on the Missouri Coteau in North Dakota.

Following the roadlog are a number of miscellaneous short papers that will give an indication of the present status of Pleistocene studies in North Dakota. The first half of these deal with the Missouri Coteau.

We wish to thank the many people who have contributed to the field trip and guidebook. Wayne A. Pettyjohn and John R. Reid helped edit much of the Guidebook. Walter L. Moore helped us with the roadlog. Those who helped organize the field trip include Walter L. Moore, John R. Reid, Wayne A. Pettyjohn, John A. Brophy, John P. Bluemle, Wilson M. Laird, E. A. Noble, Dan E. Hansen, H. A. Winters, Milton O. Lindvig, Samuel S. Harrison, Dennis N. Nielsen, Alan M. Cvancara, C. G. Carlson, LeRoy Staiger, M. J. J. Bik, and J. A. McAndrews. Support for the field trip has been contributed by the North Dakota Geological Survey, the Geology Department at the University of North Dakota, the U. S. Geological Survey, the North Dakota State Water Commission, and the North Dakota Geological Society. Cheryl Strehse typed the entire Guidebook.

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ROADLOG

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SATURDAY
JAMESTOWN TO BISMARCK BY WAY OF LEHR

Mile

-0.7 James River spillway.--Jamestown is in a large glacial meltwater spillway channel occupied by the grossly underfit James River. The spillway extends from north-central North Dakota, where it drained glacial Lake Souris, to southeastern North Dakota, where it emptied into glacial Lake Dakota (fig. R-1). The spillway is 1/3 to 2/3 mile wide and has a meander length of about 10 miles. The meltwater river that occupied this spillway, bank to bank, was about the size of the present lower Missouri River. The present James River has a meander length of about 1200 feet.

The James River spillway is a typical example of the kind of large glacial spillway that is characteristic of the northern plains, which were the site of numerous large proglacial lakes during Wisconsin time. Contrasting the spillways with normal meltwater channels, normal channels came directly from the ice margin, were much smaller, carried large amounts of outwash sand and gravel, generally had a braided pattern, and were generally aggrading. Spillways that drained glacial lakes were large, transported little bedload material because it had been trapped in the lakes, had a meandering pattern, and were degrading.

Two other very large spillways in this area are the Sheyenne River spillway (eastern North Dakota; drained Lake Souris at a later time than the James River spillway and emptied into Lake Agassiz; (fig. R-1); see p. 18-22, G. H. Dury, 1964, U. S. Geol. Survey Prof. Paper 452-A) and the Minnesota River spillway (drained Lake Agassiz and emptied into Mississippi River).

The glacial geology of Stutsman County (until mile 38.4) was described by H. A. Winters, 1963, in N. Dak. Geol. Survey Bull. 41.

-0.4 Climb out of James River spillway onto the flat to undulating ground moraine upland (fig. R-7a).

0.0 Junction of old U.S. 10 and N.D. 281. South on 281.

0.4 Overpass, Interstate 94. "Ground moraine."--We will drive over flat to undulating "ground moraine" until mile 7. It consists of a series of dozens of parallel "washboard moraines" 300 to 600 feet apart and only a few feet high. "Washboard Moraines" (Gwynn, 1942, Jour. Geol., v. 50 p. 200-208; and 1951, Geol. Soc. America Bull., v. 62, p. 233-250) are generally recognizable only from the air. They are a type of end moraine, though this area is so flat that it is usually called "ground moraine." Note the lobate pattern of the "washboard moraines" on figure R-7a.

0.6 Cross small valley, a tributary to the James River. It was cut into the "ground moraine" in post-glacial time, and probably in pre-hypsithermal time.

Drift Prairie.--We will be on the Drift Prairie (fig. R-1 and R-2) until mile 28.8. No stops. The Drift Prairie, like the till plains of the lower Midwest, is characterized by flat to undulating topography ($0^{\circ}$ to $4^{\circ}$ slopes). Dominant landforms are "ground moraine" or "washboard moraine," with associated meltwater channels, kames, and eskers. End moraines (other than "washboard moraines") consist of bands of low hummocks. The drift (latest Wisconsin in age) is generally several tens of feet thick. Drainage is integrated only along the edges of large meltwater channels. In between the channels are numerous shallow marshy depressions. Permanent lakes are shallow and uncommon (Fig. R-7a and R-8).

In general, the glacial landforms of the Drift Prairie were associated with a "normally" retreating glacial terminus. At stop 1 (mile 31) we will discuss the differences between the Drift Prairie and the Missouri Coteau, whose landforms were associated with large-scale glacial stagnation, rather than with a "normally" retreating terminus.

1.3 Small kame 1000 feet to the east (left), typical of the Drift Prairie (fig. R-7a).

3.4 Esker by farm on west (right) side of road; it is being mined for gravel (fig. R-7a).

4.9 Esker crossing road from northeast (left), parallel to railroad (fig. R-7a). Most eskers in North Dakota trend southwestward, parallel to the general direction of ice movement.

5.0 Midland Continental Railroad (which was to have extended from Canada to Mexico; it is 40 miles long).

7.2 Small meltwater channel originating at the end of an esker, 1 mile northeast of here (fig. R-7a).

7.9 Subtle Eldridge end moraine until mile 11.0: a band of gently-rolling topography (see fig. R-3 and R-7a).

11.0 Cross Beaver Creek meltwater channel. Obscure outcrops of black shale of the Cretaceous Pierre Formation are present along the sides of the channel. We will drive over "ground moraine" with obscure "washboard moraines" for the next 10 miles (fig. R-7a).

18.0 Junction N.D. 46. Turn west (right).

20.9 Drop down onto collapsed till (dead-ice moraine) for 1/2 mile (fig. R-7a).

21.6 Cross a 1-mile-wide valley-fill of gravelly sand deposited by small braided meltwater streams; the outwash is covered by Recent alluvium. It extends several miles north and south of here (fig. R-7a).

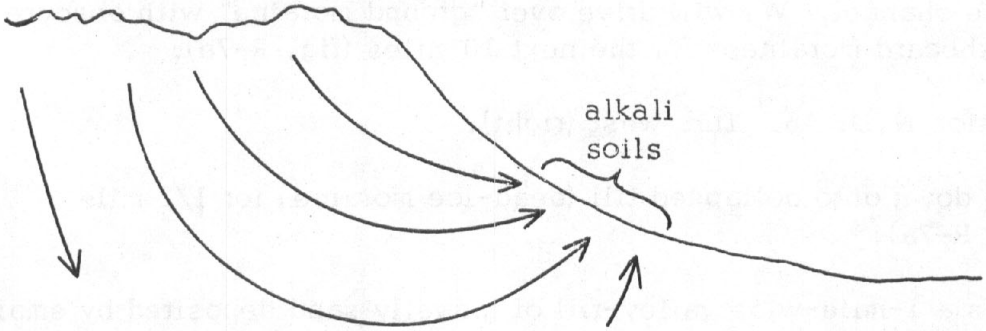
22.6 For the next 6 miles, we drive over ground moraine or low-relief dead-ice moraine, which in this area is an undulating moraine consisting of subglacial (lodgement) till overlain by a thin blanket of superglacial till.

Missouri Escarpment.--We will rise 300 feet in the next 8 miles. This rise is the Missouri Escarpment (or Coteau Escarpment or Missouri Coteau Escarpment), which is unspectacular at this point. Forty miles south of here the escarpment rises 300 feet in 1 mile (fig. R-9).

The escarpment marks the contact between the Missouri Coteau (eastern edge of the Missouri Plateau section of the Great Plains) and the Drift Prairie (western part of the Western Lake section of the Central Lowlands). The escarpment has been interpreted by some as either (1) a late Cenozoic fault scarp (especially in northwestern North Dakota) or (2) the erosional scarp between the (a) late Tertiary or early Pleistocene Missouri peneplain (also called the Alberta High Plains or the third prairie level) and (b) the early or middle Pleistocene Central Lowlands peneplain (also called the Saskatchewan Plains or the second prairie level).

At this point the escarpment might also be considered the proximal side of a great composite lateral-moraine complex formed along the west edge of the James glacial lobe. The drift of the Missouri Coteau is here several hundred feet thick and the buried bedrock escarpment is about 30 miles west of here, as can be seen in the generalized cross-section, figure R-10.

Alkali soils.--A 5-mile-wide band of alkali soils (Cresbard series; Solodized Solenetz or Glossic Natriboroll) generally occurs at the bottom of the Missouri Escarpment. These soils are the result of a regional groundwater flow system; groundwater recharges through the drift of the Coteau and discharges alkaline water at the bottom of the escarpment.




28.8 Enter Missouri Coteau; leave Drift Prairie. Enter area of rolling to hilly topography (largely collapsed superglacial drift); leave area of flat to undulating topography (largely subglacial till) (fig. R-7b).

31.0 Stop 1.--Observation stop. (We are on a small body of collapsed superglacial stream gravel.) Looking back toward the east you can see that we are near the top of the Missouri Escarpment; you can see for several dozens of miles east across the flat Drift Prairie.

For most of the rest of the day we will be on the Missouri Coteau. The Coteau is contrasted with the Drift Prairie in the following table.

| | Drift Prairie | Missouri Coteau |
|---|---|---|
| General topography | Undulating or flat | Rolling or hilly |
| Average maximum slope angles | $0^{o}$ to $4^{o}$ | $4^{o}$ to $20^{o}$ |
| Characteristic landforms | "Ground moraine," meltwater channels, kames, eskers, low hummocky end moraines, washboard moraines | Collapse topography (superglacial till, alluvium, and lake sediment), ice-walled lake plains, ridged end moraine loops, knobby end moraine |
| Drift age | Latest Wisconsin | Latest Wisconsin |
| Drift thickness average | Several tens of feet | A few hundred feet |
| Drainage | Partly integrated | Non-integrated |
| Lakes | Few, shallow | Numerous, less shallow |

As mentioned before, the landforms of the Drift Prairie were associated with a "normally" retreating glacier terminus, whereas those of the Missouri Coteau were associated with large-scale glacial stagnation.

See paper 30-A, following the roadlog, for a discussion of glacial stagnation features on the Missouri Coteau.

<u>Collapsed superglacial till</u>.--For the next 8 miles we will be driving on hilly collapsed superglacial till (fig. R-7b). About 3/4 of our route until stop 9 late this afternoon will be on collapsed till. Collapsed-superglacial-till topography may be called dead-ice moraine, stagnation moraine, disintegration moraine, or collapse moraine. It formed when a thick blanket of superglacial till collapsed or was let down as the underlying dead (stagnant) ice sheet melted. This superglacial "ablation" till is not noticeably coarser or more washed than basal (lodgement) till. It is about 1/3 sand, 1/3 silt, and 1/3 clay (largely montmorillonite) with only a few percent gravel-sized material.

The dominant depositional processes were mudflow (of "flow-till") and other kinds of mass movement down ice slopes; the superglacial topography was continuously changing as a result of differential melting caused by an irregular (and continuously changing) thickness of superglacial till. This blanket of till insulated the stagnant ice, allowing it to persist for a few thousand years.

Where the blanket of superglacial till was thin the ice melted more rapidly, resulting in a depression. Part of the adjacent superglacial till slumped or flowed into the depression, thickening the till in the depression and thinning the till near the depression. The ice around the depression then melted faster than that below the depression. Continuous alteration and inversion of topography on the stagnant ice surface resulted.

The conspicuous high hills in this area (shown surrounded by ice-contact faces in figure R-7b) are composed of till and are a part of the dead-ice moraine. The details of their formation is unknown, though they are probably the result of flowage of superglacial till into large depressions in the stagnant ice.

The sloughs of the Coteau are not true "kettles"; they did not form by the melting of single, buried blocks of ice. Rather, they resulted from the melting of a sheet of ice having an irregular thickness and an irregular cover of superglacial till.

The steep faces bordering many sloughs are seepage faces, rather than ice-contact faces or wave-cut faces. Groundwater discharge tends to sap away the material from the base of these slopes, causing them to remain steep. Seepage faces are generally recognizable on air photos and may therefore be used to delineate groundwater-discharge areas in morainic topography.

An analysis of the paleocology of a Coteau slough is given in paper 30-H, following the roadlog.

38.4 Town of Gackle--"Duck Capital of the Nation." (Gackle is, oddly enough, on a straight line between Jud and Zap.) The Missouri Coteau is the largest wild-duck producing area in the 48 states. The ducks nest in the thousands of "prairie potholes" or sloughs, which are so characteristic of collapse topography.

Junction N.D. 56; turn south (left). Geology of Logan and McIntosh Counties was described by Lee Clayton, 1962, in N. Dak. Geol. Survey Bull. 37.

38.6 Gackle Improvement Association.

39.5 <u>Collapsed superglacial alluvium</u>.--A half-mile-wide band of collapsed alluvium with hilly topography nearly identical to that of collapsed superglacial till. The alluvium is sand and gravel that was deposited by streams on top of the stagnant ice sheet. Melting caused the sand and gravel to collapse, producing faulting and other collapse structure (fig. R-7b).

Most of the alluvium on the Coteau is nonglacial. The stagnant ice was approximately 300 feet thick and took about 3000 years to melt (see paper 30-A), producing about 1 inch of meltwater per year. Precipitation was probably at least 25 inches a year because spruce and associated vegetation typical of a slightly moister and cooler climate than at present lived in the area (the present annual precipitation is about 17 inches). For these reasons, only a very small fraction of the water in the rivers on the ice was meltwater, especially after the superglacial drift thickened and its insulating effect increased. Therefore the collapsed superglacial alluvium is not called "outwash."

Note: "alluvium" is used here to mean any river-deposited sediment, including both glacial alluvium (outwash) and nonglacial alluvium.

40.1 Back onto collapsed till for 4 miles (fig. R-7b).

44.3 Stop 2.--Contact between typical collapsed superglacial till (most of north or right part of exposure) and typical collapsed superglacial alluvium (south or left end of exposure) (fig. R-11a and R-7b).

Note the bodies of contorted pond sediment within the till. These were deposited in the numerous depressions which are characteristic of the surface of a disintegrating stagnant sheet of ice. The contortions are the result of slumping or flowage that occurred when the superglacial drift collapsed or was let down as the underlying ice melted.

Alternating collapsed river sediment and collapsed till for next 5 miles. They can frequently be distinguished from the window by the presence of boulders on the till and the absence of boulders on the collapsed alluvium.

45.4 Junction N.D. 34; turn east (left).

48.5 Turn south (right).

49.4 Stop 3.--Typical polygonal disintegration trenches in collapsed river sediment (fig. R-11b and R-7b).

Disintegration trenches.--These features are characteristic of collapsed alluvium. Their orientation lacks any correlation with the present topography; they might go up a hillside, over the top, and down the other side, or they may cross through depressions. Their pattern may be polygonal (as here at stop 3), parallel (as at stop 12), or braided.

The origin of disintegration trenches is given in paper 30-A (fig. A-3). Briefly, they formed when a superglacial channel was filled with drift; the topography was inverted due to differential melting caused by differential insulating effect of the drift resulting in an ice-cored ridge; the ridge was buried under river sediment; and the buried ice core melted, forming a trench in the overlying sediment.

Disintegration trenches are generally visible on air photos of most collapsed stream sediment and are considered to be good evidence for collapsed superglacial stream sediment.

50.1 Stop 4-A.--A rather poorly-developed circular disintegration ridge or "doughnut" (fig. R-11b and R-7b).

Disintegration ridges.--These are characteristic features of both collapsed till and collapsed alluvium. Many disintegration ridges resemble eskers.

Disintegration ridges were formed when drift slid, flowed, was washed off, or was squeezed up from beneath stagnant ice into a cravasse, a channel, or a hole in the ice or along the side of a mass of ice (fig. A-1 in paper 30-A).

"Doughnuts."--"Doughnuts" are circular ridges a few hundred feet across and about 15 feet high. In collapsed-till and collapsed-lake-sediment areas they are usually interpreted as circular disintegration ridges formed by (fig. A-2 in paper 30-A) (a) mass movement of superglacial drift into a sinkhole in a stagnant ice sheet, (b) inversion of topography as a result of the insulating effect of the drift in the bottom of the sinkhole and mass movement of this drift off the sides of the ice cone thus formed, and (c) melting of the ice core (Gravenor, C. P., 1955, Am. Jour. Sci., v. 253, p. 475-481). Bik (paper 30-F following this roadlog), however, has suggested that many "doughnuts" may be collapsed pingos.

51.5 Turn west (right).

52.1 Stop 4-B.--Rather poorly-developed elongate disintegration ridges (fig. R-11b and R-7b).

52.5 Turn south (left).

58.1 Low-relief collapsed superglacial till for the next 6 miles (fig. R-7b).

63.3 Town of Fredonia.

64.1 High-relief collapsed till.

64.6 Junction N.D. 13; turn west (right). Alternating collapsed till and collapsed stream sediment.

71.6 Note flat skyline 1 1/2 miles south; this is a perched (but slightly collapsed) alluvial plain on the same level as the perched Lake Lehr plain, which we will see in a few minutes (fig. R-12).

75.0 Note flat skyline 1 1/2 miles southwest (left); this is the perched plain of ice-walled Lake Lehr (fig. R-12).

For the next mile we drive over an apron of collapsed stream sediment that surrounds the Lake Lehr plain (fig. R-12).

76.7 Climb up ice-contact face onto the Lake Lehr plain; curve southwest (fig. R-12).

77.1 Turn east (left).

77.4 Turn south (right). Continue across the flat elevated plain of glacial Lake Lehr. Note that it is concave, with a slightly raised rim (fig. R-13).

77.9 Turn east (left).

78.1 <u>Stop 5</u>.--Edge of Lake Lehr plain: observation and lithology stop. A hole will be augered into the lake sediment here.

<u>Lake Lehr--An ice-walled lake</u>.--Lake Lehr plain (fig. R-12 and R-13) is a good example of one of the most characteristic features of large-scale glacial stagnation: ice-walled-lake plains. These features at first seem "anomalous": they are flat glacial-lake plains underlain by horizontally-laminated silts and clays that are perched up above the surrounding moraine, like buttes or small mesas. The first geologist to work in this area, J. E. Todd (1896, U. S. Geological Survey Bull. 144, p. 19, 35, 42, 43) thought that they were remnants of "glacial terraces," but the drainage is non-integrated; there has not been enough postglacial erosion to form such features. However, they can be easily explained within a framework of large-scale glacial stagnation: they are ice-walled lake plains surrounded by outward-facing ice-contact faces (fig. R-13).

Sand and gravel commonly occurs on the margin of ice-walled lake plains (as at the northeast edge of Lake Lehr, by the graveyard) where meltwater streams flowed off the stagnant ice into the lake. Drillholes near Lehr showed that 60 to 80 feet of lake sediment underlies the north end of the lake plain. Bedrock is at a depth of 200 to 450 feet in this area.

Turn around; return west.

78.3 Turn south (left).

78.9 Drop down from perched Lake Lehr plain onto collapsed superglacial-lake sediment (right side of fig. R-13). We will continue on lake sediment for 1 1/2 miles and turn around and come back. Collapsed lake sediment has a topography very similar to that of collapsed till, though it tends to be slightly more subdued with gentler slopes. As we drive along, lake sediment can generally be distinguished from till by the lack of stones in the fields. Bedding is broadly folded to intricately contorted, in contrast to that of collapsed stream sediment, which tends to have numerous gravity faults.

80.3 Turn around, return to Bible Camp east of Lehr for lunch.

Lowenthall Cut.--A half mile east of here an excellent example of folded bedding resulting from collapse of the superglacial lake sediment was exposed in a new road cut in 1961 (destroyed now; fig. R-14).

Abundant in-place articulated fossil clam shells (*Lampsilis luteolus*) were collected from these beds. A large, lumpy, fossil pearl was found in one of the clams. Three species of large fossil clams have been found in collapsed superglacial-lake sediment or collapsed superglacial alluvium at a dozen or more localities in the Coteau. The presence of these clams suggests a rather mild superglacial environment. Because these clams are parasitic on fish in one of their larval stages, these superglacial lakes and streams must have been connected with the proglacial drainage system, allowing migration of fish from the Missouri River, up Beaver Creek, and up at least 20 miles onto the stagnant glacier. (See paper 30-E by Tuthill following roadlog.)

Mummy Cat Site.--In the unlikely chance that conditions permit, we will look at the fossil exposure in the ice-contact face south of Mummy Cat Slough, 2 miles southeast of stop 5 (fig. R-12); if not, a bulk sample of the fossiliferous sediment will be displayed at lunch time.

The Mummy Cat Site is one of the best ice-contact fossil sites on the Coteau. A generalized section is shown in figure R-15.

The upper unit (a) is a clayey glacial-lake silt; is broadly folded and tilted (as a result of collapse when the underlying stagnant ice melted); contains a dozen species of aquatic snails (belonging to genera *Amnicola*, *Valvata*, *Gyraulus*, *Helisoma*, *Lymnaea*, *Physa*, *Armiger*, and *Promenetus*) plus sphaerid clams, the mussel *Lampsilis* sp., ostracodes, and the algae *Chara* sp., bedding is much thicker and more indistinct than in the lower unit as a result of plant and animal activity on the lake floor (see paper 30-E).

The lower unit (b) is a silty glacial-lake clay; is highly contorted (resulting from flowage as topography on underlying stagnant ice shifted as the ice melted); is free of fossils; has much thinner and more distinct bedding than upper unit as a result of the lack of organic activity.

This cut, at the northeast edge of former Lake Lehr, illustrates a kind of succession that probably was common in superglacial and ice-walled lakes on the Coteau. At first the lake environment was unstable; shores frequently changed because the enclosing ice was melting rapidly, and large amounts of cold meltwater flowed into the lake. Later, however, the environment began to stabilize as the blanket of insulating superglacial drift on the surrounding ice got thicker, allowing the lake waters to become warm enough for abundant plants and animals. A similar succession of aquatic environments is occurring today on the stagnant part of the Martin River Glacier in Alaska (see references in paper 30-E). (See fig. A-7 of paper 30-A.)

(For discussion of characteristics of some ice-walled-lake sediments, see paper 30-C.)

82.7 Evangelical United Brethren Bible Camp. Lunch.

83.4 Turn north (left) through Lehr.

83.5 Descend ice-contact face at north edge of Lake Lehr plain onto the narrow apron of collapsed river sediment (fig. R-12).

84.1 Collapsed till for 4 1/2 miles (fig. R-12 and R-17).

88.7 Collapsed alluvium (fig. R-17). Notice flat-topped hills, which are uncollapsed remnants of original alluvial plain (where it was deposited on solid ground in holes in the stagnant ice sheet). Note the Streeter end moraine to the northeast. Much of this river sediment came from behind (east of) the moraine, through the gap (meltwater channel) in the middle of the moraine loop. The gap is 100 feet above and 5 miles northeast of the uncollapsed remnants of alluvial plain in this area.

91.6 Turn east (right).

91.9 Stop 6.--Observation stop. Collapsed-river-sediment topography with uncollapsed remnants (flat hilltops). Inconspicuous linear disintegration trenches here are easily visible on air photos (fig. R-16 and R-17).

93.6 Turn north (left).

93.9 Linear disintegration ridge or esker (fig. R-17).

94.6 Turn east (right). Alternating collapsed river sediment and collapsed till. Streeter end moraine is conspicuous 2 miles north (left) (fig. R-17).

96.6 Turn north (left).

98.2 Start to climb up the Streeter end moraine. Many small end moraine ridges superimposed on the larger ridge can be seen here (fig. R-17).

98.9 Crest of the Streeter end moraine.

100.2 Turn around, return to crest.

100.5 Stop 7.--Crest of Streeter end moraine (fig. R-17). Observation stop. This is one of the more striking end moraines in the state. It consists of a discontinuous series of individual "loops" formed by small ice lobes flowing out from the main ice mass. These lobes were 4 to 8 miles across. Each "loop" consists of a curved ridge, about 2 miles wide and 200 to 300 feet high. Superimposed on this ridge are smaller ridges 10 to 50 feet high and a few hundred feet apart.

The entire loop can be seen from here (fig. R-17). The north limb (4 miles north of here) joins the south limb of the next loop to the north, forming an interlobe area that is as much as 575 feet above adjacent depressions immediately behind the moraine.

The Streeter moraine is similar to the "push moraines" or "thrust moraines" described by Kupsch (1962, Jour. Geol., v. 70, p. 582-594) in Saskatchewan, except that those are composed of folded Cretaceous shale, rather than till and pre-existing outwash and lake sediment, as is the Streeter moraine.

Collapsed alluvium.--Extending 12 miles southwest from this loop to the east edge of the Burnstad end moraine is a 2 to 5 mile wide band of collapsed river sediment (which we were driving over from mile 89; fig. R-17). Much of the river sediment came through the channel breaching the middle of this loop (not easily seen from this point). Therefore, most of the stagnant Burnstad ice sheet had not yet completely melted by the time the Streeter ice stagnated. A prominent ice-contact face marks the front (west) side of many of the Streeter moraine loops; these loops were built up against stagnant Burnstad ice lying in front of the Active Streeter ice.

A similar relationship is illustrated in figure 58 of R. G. Ray's air photo manual (1960, U. S. Geol. Survey Prof. Paper 373), which shows the McIntosh County loop (not in South Dakota) of the Streeter moraine. Here, a gap, 1/2 mile wide and 150 feet deep, breaches the moraine. Leading away from this gap is a body of collapsed river sediment extending 17 miles to the west edge of the Coteau. A prominent ice-contact face can be seen along the west side of this loop.

Return 2.4 miles south and 2 miles west.

104.9 Turn north (right); still on collapsed river sediment (fig. R-17).

106.9 Turn west (left); a typical large saline lake--a local groundwater discharge area. Alternating collapsed river sediment and collapsed till for the next 4 miles (fig. R-17).

111.2 Climb up ice-contact face onto an elevated lake plain; depressions indicate the lake sediment was at least in part deposited on top of the ice (fig. R-17).

112.3 Stop 8.--Observation stop (fig. R-17). Top of the ice-contact face at the west edge of the lake plain where a "moat" surrounds the lake plain. "Moats" such as this are the result of a deficiency of superglacial drift where it had slumped into the lake, forming the slight rim characteristic of many ice-walled lake plains (fig. R-19 and R-17).

114.6 Continue on collapsed till for 2 1/2 miles. Large circular disintegration ridge on left (south side of road) (fig. R-17).

114.9 Collapsed alluvial sand and gravel (fig. R-17). This collapsed superglacial valley-train turns into a channel as it leaves the area of collapsed till and enters the Burnstad end moraine. Note the subparallel ridges, which are eskers and disintegration ridges, trending southwest toward the channel (fig. R-17).

115.9 Turn south (left).

116.9 Turn west (right).

117.7 At this point the collapsed valley-train turns into a channel cutting through the Burnstad end moraine (fig. R-17).

117.9 Burnstad end moraine.--The Burnstad end moraine (about 13,000 B.P.) has a much different morphology from that of the Streeter moraine. Instead of being a ridge or band of ridges it is a band of knobs closely resembling dead-ice moraine. However, it is not considered to be dead-ice moraine because: (a) it contains some live-ice features (a few transverse end moraine ridges) and lacks large-scale-stagnation features. For instance, as we just saw, the collapsed-alluvium bodies associated with dead-ice moraine are replaced by meltwater channels where the meltwater crossed the Burnstad moraine. (b) Statistical analysis of 920 lakes in a 90-square-mile area containing dead-ice moraine and end moraine (just north of here in T. 135 and 136 N., R. 71 and 72 W.) indicate that the lakes of the dead-ice moraine and the end moraine are significantly different: The azimuths of the long axes of the lakes in areas mapped as dead-ice moraine are random, whereas those in end moraine have a circular mean that is equal (within 2 degrees) to the azimuth of the long axes of the end moraine; lake density is significantly greater in the end moraine; and lakes are significantly more elongated and smaller in the areas mapped as end moraine. (c) The area mapped as Burnstad end moraine is known to mark the outer limit of the late Wisconsin drift sheet, as is shown in N. Dak. Geol. Survey Report of Investigation 44.

119.8 Another meltwater channel.

119.9 Turn north (right).

120.9 Turn west (left).

122.9 Turn south (left).

123.9 Turn west (right).

124.3 Stop 9.--Outer edge of Burnstad end moraine--marked by a ridge 40 feet high and 500 feet wide (fig. R-18).

This is the contact between the Missouri Coteau and the Coteau Slope. The following table contrasts these two regions:

| | Missouri Coteau | Coteau Slope |
|---|---|---|
| General Topography | Hilly or rolling hummocks and depressions | Undulating to rolling; valleys and divides; some badlands |
| Drainage | Almost completely non-integrated | Almost completely integrated |
| Dominant landforms | Dead-ice moraine, collapsed alluvium, ice-walled-lake plains, end moraines | Stream-eroded topography and ground moraine draped over stream-eroded topography |
| Drift Age | Latest Wisconsin Streeter and Burnstad drifts | Early Wisconsin or pre-Wisconsin Napoleon drift |
| Drift thickness | A few hundred feet; very rare bedrock outcrops | A few tens of feet; absent in much of area; many bedrock outcrops |
| Lakes | Abundant small ones in stagnant ice-block depressions | Rare; large ones in blocked valleys |
| Ducks | Numerous | Few |

In general, the Coteau Slope is characterized by much non-glacial topography, whereas the Missouri Coteau is characterized by glacial topography.

Continue westward onto the Coteau Slope.

124.9 Drop down into 1/4 mile-wide meltwater channel, which carried water from Burnstad moraine into Beaver Creek drainage to the southwest (fig. R-17).

125.5 Another meltwater channel, slightly higher elevation.

125.7 Napoleon "ground moraine." May be early Wisconsin in age (before 38,000 B.P.). It consists of a thin (a few tens of feet, at most) blanket of till draped over the pre-existing stream-eroded topography (fig. R-17 and R-18).

Napoleon stagnation.--The last glacier to cover this area also underwent large-scale stagnation, however, with results which are quite different from those of the late Wisconsin glacial stagnation on the Missouri Coteau. As is shown in the discussion on ice thickness in paper 30-A, the ice apparently was cleaner and carried thinner supergiacial till. As a result, probably only subdued dead-ice moraine was formed. Because the large-scale elements of the pre-existing drainage pattern had not been obliterated and because morainic depressions were very shallow, the drainage was soon completely re-integrated, and most of the subdued morainic topography was eroded away. This left a topography that today has few recognizable morainic features (such as till disintegration ridges) that are characteristic of collapsed, thick superglacial till. Non-integrated morainic topography is preserved on a few flat drainage divides, however.

Evidence suggesting large-scale stagnation in this part of the Coteau Slope includes meltwater channels crossing divides or running along drainage divides and a large body of collapsed outwash in the valley just west of Napoleon.

127.7 Stop 10.--The meltwater channel here is along the top of a drainage divide area. The channel begins in the slough 1/4 mile southwest of here; this slough was apparently a "plunge pool" where the meltwater river flowed off of the ice onto solid ground, which was then exposed through the ice only along the topographic highs (drainage divides) (fig. R-17 and R-18).

128.7 The railroad runs northwest to Napoleon down the preglacial valley. One mile south (left) of here a meltwater channel cuts across the drainage divide at the head of the valley, indicating that a lake was dammed in the valley by the wasting Napoleon ice.

128.9 South of here 1 1/4 miles is a classic exposure of pre-Wisconsin outwash gravel underlying Napoleon drift. It is much more weathered than the Napoleon drift. It is cemented (Napoleon outwash is not) and is reddish yellow (Napoleon drift is never so red). Most of the limestone pebbles have been leached out, leaving hollow molds (a rarity in Napoleon gravel). Similar weathered gravels are exposed at several other places in Logan County and at many places in the Coteau Slope in the northwestern part of the state, where they have been called the "Four Bears conglomerate" (from an exposure near the Four Bears Bridge near Newtown).

129.4 The small knobs on the divide areas are kames, which are abundant on the divide areas of the Napoleon drift.

130.8 Note small meltwater channel (fig. R-17), about 300 feet wide, crossing the divide between drainage to the south (into Beaver Creek) and drainage to north (into valley by Napoleon).

131.9 Junction N.D. 3; turn north (right).

133.9 In the gravel pit on right, organic material was found and radiocarbon dated at older than 38,000 B.P. (W-990). The gravel body was a delta along a vague strandline, near which we will be driving for the next mile. The gravel is probably Napoleon in age or younger (but definitely pre-Long Lake drift) (fig. R-4). Other evidence suggests that the Napoleon drift is fairly old: (1) it has a highly integrated drainage pattern. (2) In many areas the drift has been completely eroded away. (3) Another radiocarbon date (W-1433) from Napoleon outwash in Burleigh County is greater than 38,000 B.P. (4) A partial jaw of Equus hatcheri from Burleigh County may be pre-Wisconsin in age. However, the Napoleon drift appears to be only slightly more weathered than late Wisconsin drift in the area, and non-integrated morainic topography is preserved on a few flat drainage divides. Therefore, the Napoleon drift may be early Wisconsin (or perhaps pre-Wisconsin) in age.

136.4 Outwash plain composed of sand and gravel from the Burnstad moraine 4 miles to the east (fig. R-17).

136.9 Napoleon.

137.9 Junction N.D. 34; turn west (left).

139.7 The road is on connected, opposing cuspate bars (fig. R-17) of postglacial age for the next mile. Lakes are saline and have no outlet. The south one is commonly dry. Continue on Napoleon "ground moraine."

143.1 Butte capped by sandstone of the Fox Hills Formation (late Cretaceous marine) 1/2 mile north (left).

145.3 Emmons County line. The Pleistocene geology of Emmons County has not been studied in detail but thin Napoleon "ground moraine" and abundant outcrops of the Fox Hills Formation (fig. R-6) occur.

147.1 Outwash (fig. R-3).

151.7 Hill of Fox Hills sandstone; little drift.

152.5 Thin Napoleon "ground moraine" on the Fox Hills Formation for the next 6 miles.

158.5 Thin loess overlies the "ground moraine" for the next 11 miles. We are at the eastern edge of the area of the most widespread loess in North Dakota. It thickens to several feet or more towards the Missouri River. The thickest loess, in southwestern Emmons County and south into South Dakota, may be pre-Napoleon in age--possibly early Wisconsin.

163.3 Hazelton.

164.0 Junction U.S. 83; turn north (right).

169.2 Drop down into pre-glacial Cannonball Valley (fig. R-3). (See paper 30-J by Kelly and Buturla.) Fox Hills Formation on valley sides (no drift). Outwash and Recent alluvium fill the valley bottom.

172.9 Climb up out of preglacial Cannonball Valley. Fox Hills Formation.

173.4 Napoleon "ground moraine" for 3 miles.

174.1 Burleigh County; geology described by Jack Kume and D. E. Hansen, 1965, N. Dak. Geol. Survey Bull. 42.

175.3 Drop down into preglacial Heart Valley (fig. R-3).

176.8 The valley bottom is 5 miles wide and is underlain by fine outwash, alluvium and marsh and lake sediment. Long Lake, to the east (right), lies in the valley below the junction of the preglacial Heart and Cannonball Valleys. It has no outlet, is saline, and frequently dries up entirely.

182.1 Up onto a vague end moraine at the terminus of the Long Lake drift. Note the subdued morainic topography for the next 9 miles. The amount of postglacial erosion is apparently more similar to that of the Burnstad drift than to that of the Napoleon drift, and it therefore has been suggested that the Long Lake drift is late Wisconsin in age (fig. R-3 and R-4).

187.2 Junction U.S. 10; turn west (left).

188.7 Gravel pits in "kames."

191.0 Outer (west) edge of Long Lake end moraine (fig. R-3). Drive on proglacial outwash and lake plain for next 11 miles. The plain is underlain by fine-grained outwash, lake and marsh sediment, and alluvium. The depressions, such as the one at Menoken (mile 198.7) may be ice-block depressions (kettles).

193.0 McKenzie.

198.7 Menoken.

201.5 Apple Creek; occupies meltwater channel from Long Lake moraine to the east (fig. R-3).

202.0 Cannonball Formation (marine Paleocene) overlain by thin till is exposed at places between here and Bismarck (fig. R-6).

204.2 A radiocarbon date from the Napoleon drift in this roadcut is greater than 38,000 B.P.

209.0 Bismarck.

211.2 G.P. Hotel. End of first day's fieldtrip.

## SUNDAY
## BISMARCK TO JAMESTOWN

Miles

0.0 G.P. Hotel. First 24 miles are a repeat of the last part of yesterday's trip. Cannonball Formation overlain by thin till for 9 miles (fig. R-3 and R-6).

9.2 Cross Apple Creek; driving on outwash or lake sediment deposited in front of the Long Lake end moraine for next 10 miles (fig. R-3 and R-6).

20.2 Outer edge of the Long Lake end moraine (late Wisconsin) (fig. R-3).

31.0 "Ground moraine" behind Long Lake end moraine for next 6 miles.

34.8 Kidder County; geology described by J. A. Rau and others, 1962, in N. Dak. Geol. Survey Bull. 36.

37.4 Collapsed river sediment at the outer margin of the Burnstad drift; note gravel pits (fig. R-3).

39.8 Low-relief collapsed till of the Burnstad drift (fig. R-3).

41.6 Moat surrounding ice-walled-lake plain at Steele; plain is 2 miles wide (fig. R- 3).

43.4 Drop off the east edge of the lake plain into moat; plain has a slight rim ridge here. Continue on low-relief collapsed till for 4 miles.

47.0 Turn north (left).

47.8 Edge of the Kidder Sand plain (fig. R-3).

Kidder Sand Plain.--The sand plain, which occupies most of the 14 townships between the Streeter and Burnstad end moraines, is in large part collapsed river sediment that was deposited on top of stagnant Burnstad ice. The source of the sediment was largely behind (east of) the Streeter moraine. This relationship again shows that a nearly continuous mass of stagnant Burnstad ice, 25 miles wide, persisted until after the Streeter ice had stagnated (fig. R-2, R-3, and R-4).

Evidence that the river sediment of the Kidder Sand Plain has in most places been collapsed is its rolling topography and non-integrated drainage; it contains hundreds of salt lakes without outlets. Local relief is as much as 100 feet. Some uncollapsed areas occur as flat hill-tops.

Other evidence that the river sediment was let down from stagnant ice is the disintegration trenches, which are easily recognized on air photos and are an almost infallible indicator of collapsed river sediment.

The Kidder Sand Plain is the most outstanding area of collapsed superglacial river sediment in North Dakota.

We are now on the west edge of the sand plain, but here the sediment is largely uncollapsed except in isolated ice-block depressions.

50.7 Small abandoned river channel.

52.0 Larger abandoned river channel. Note Sibley Buttes ahead (fig. R-3).

54.0 Curve around southeast end of Sibley Buttes; fossil Cretaceous oysters exposed in roadcut.

Sibley Buttes.--These hills are a series of southeast-northwest trending hogbacks formed of sandstone of the Cretaceous Fox Hills Formation. The sandstone beds commonly dip northeast at about 60 degrees, though in some places they are nearly vertical. Upper parts of the north-facing slopes are dip-slopes, and individual hogbacks can be traced almost unbroken for nearly 4 miles (fig. R-20).

Surface tectonic structure of this intensity is unknown in this part of the northern plains. The structure of Sibley Buttes has been generally thought to be of glacial origin.

55.1 Turn east (right).

55.4 Stop 11.--Observation and lithology stop. View of the typical rolling topography of the collapsed river sediment on all sides of us. Sibley Buttes to the southwest. Sand in the roadcuts is typical river sediment of the Kidder Sand Plain. The sand is as much as 250 feet thick under some parts of the plain, though it is only a few tens of feet thick in this area (fig. R-20).

Note Streeter moraine on the horizon 7 miles to the east (fig. R-3).

Drive east over typical collapsed-river-sediment topography. Some thin wind-blown sand on the surface in places.

62.8 Turn south (right); Streeter end moraine on left (fig. R-3).

63.0 McPhails Butte Historic Site on left. Here, in the summer 1862, General Henry Hastings Sibley, with 2800 men, fought the Sioux in the battle of Big Mound. The Sioux living along the Minnesota River in southern Minnesota had killed several hundred whites. Sibley chased the Sioux, burdened with families and possessions, westward through North Dakota. Here, to revenge the Minnesota killings, Sibley captured and burned most of their food, robes, and household goods.

64.8 Turn east (left). The river sediment is coarser sand and gravel here next to the moraine. Note the gullies formed in postglacial time down the steep outer slope of the moraine (fig. R-3).

66.5 Note meltwater gap (600 feet wide) through crest of moraine to left.

66.8 Turn south (right). River sediment is flat and only slightly collapsed (pitted) in this area and characterized by disintegration trenches, which are conspicuous on the air photos of this area. Most are about 30 feet wide, several hundred feet long, straight or slightly curved, and trend north-northwest, perpendicular to the direction of ice movement. (See fig. A-6 of paper 30-A.)

69.8 Turn east (left).

70.0 Stop 12.--Disintegration trenches are well developed here. (Though they are not conspicuous from the ground, they are obvious on air photos.) Note that the trenches change into ridges where they cross the linear depression (fig. R-21).

To review the theory for the origin of disintegration trenches (given at stop 3 yesterday; fig. A-3 in paper 30-A): they formed during the melting of the ice in ice-cored eskers or ice-cored crevasse-fillings that had been buried under an alluvial plain. This theory can be tested by extending it to explain the ridges at this stop:

The broad linear depression is the result of the melting of ice in an ice-cored esker that had been buried under the alluvial plain, and the narrower trenches are the result of the melting of the ice in ice-cored crevasse fillings. The ridges were formed as follows:

(1) Crevasses crossed a superglacial meltwater channel and extended below the bottom of the channel, as shown in figure R-22a.

(2) The drift in the crevasses and superglacial meltwater channel insulated the underlying ice, causing differential melting and topographic inversion; this resulted in an ice-cored esker crossed by ice-cored crevasse fillings (fig. R-22b).

(3) The ice-cored esker and ice-cored crevasse fillings were buried under alluvial sand (fig. R-22c).

(4) The buried ice cores then melted, and formed disintegration trenches at the sites of the buried eskers and crevasse fillings, and formed disintegration ridges where they crossed each other (fig. R-22d). The ridges might be thought of as places where "less trench" formed.

Thus, these ridges can be explained by an elaboration of the same double inversion sequence used in the general theory for the origin of disintegration trenches (fig. A-3 in paper 30-A). No known alternate theory has explained the relationships seen here at stop 12.

This theory is complex; but the probability for such a sequence of events is fairly high. Ice-cored crevasse fillings and ice-cored eskers are fairly common in front of stagnant glaciers (see Sharp, 1949, Am. Jour. Sci. v. 247, p. 289-315). If these features were buried under river sediment, disintegration trenches would result when the ice cores melted; and any place where the esker ice-channels were crossed by crevasses which extended below the base of the channel-fill, ridges, like those shown in figure R-22d, would form.

(Another good example of disintegration trenches changing into ridges when crossing a depression is in the NW 1/4 NE 1/4 sec. 33, T. 142 N., R. 70 W., 3 miles southwest of Pettibone. Other less well-developed examples are common in most collapsed stream sediment in North Dakota.)

Turn around, return west.

70.2 Turn south (left).

70.4 Look east (left) up the broad disintegration trench; note the ridges crossing the bottom (fig. R-21).

71.2 Turn west (right).

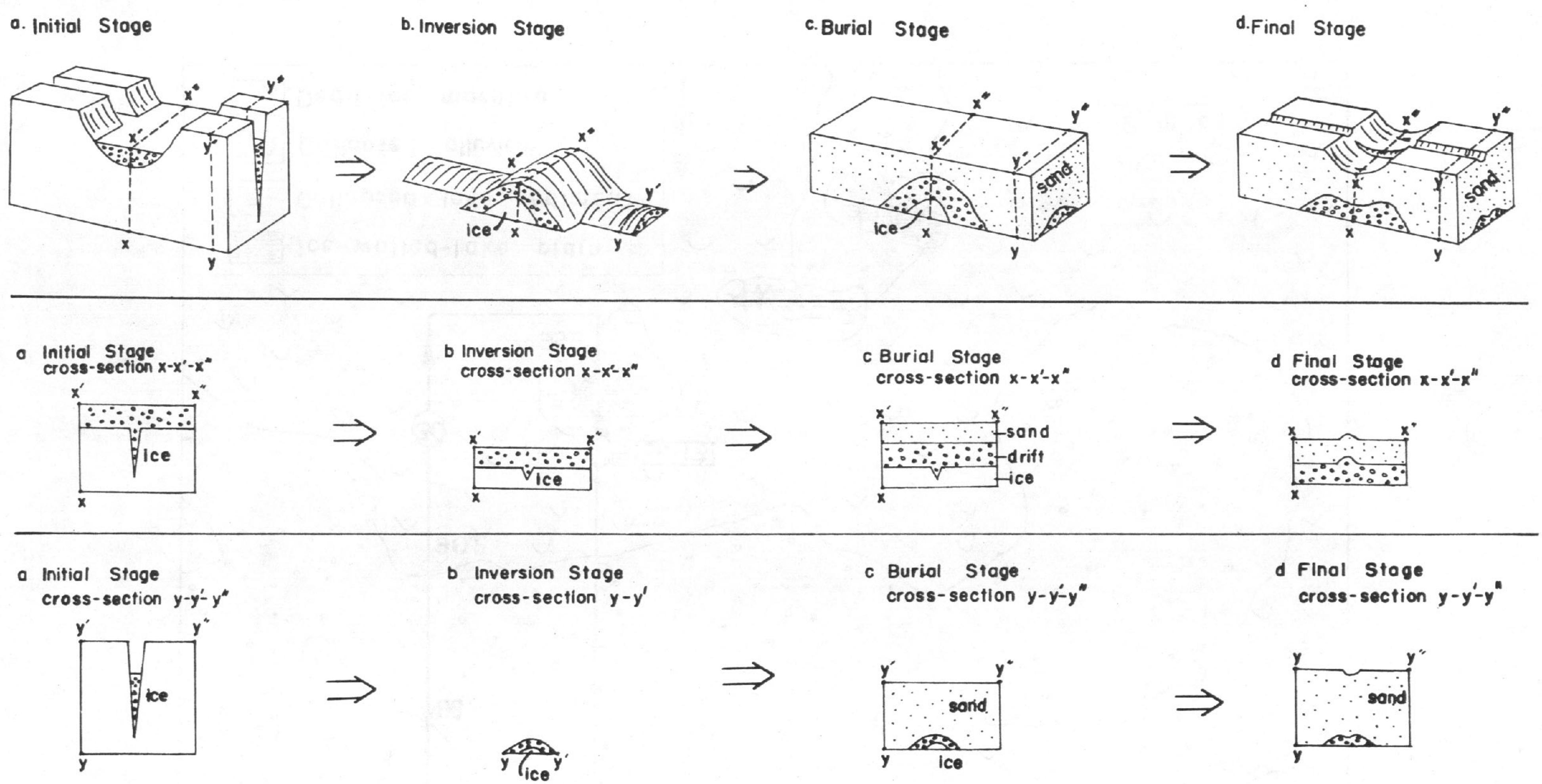



Figure R-22. Origin of disintegration trench that changes to ridge where it crosses a linear depression at stop 12.

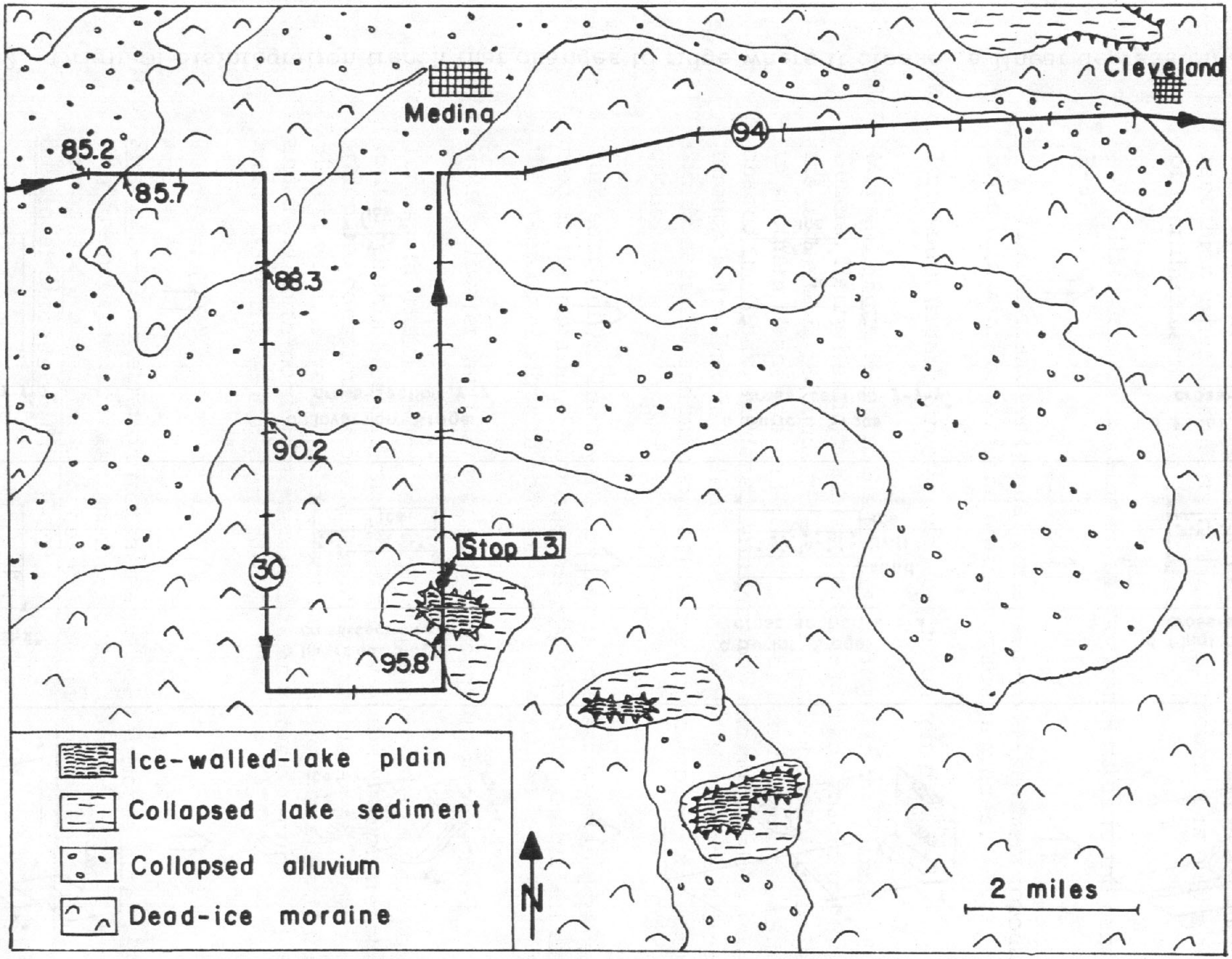



Figure R-23. Glacial geology along fieldtrip route in Medina area, with mileages and stops indicated; mile 84.2 to 110.3. Dashed line is wet-weather alternate route. Map location given in fig. R-2.

72.2 Turn south (left).

73.1 Turn east (left) onto Interstate 94. The Kidder Sand Plain here has a thin blanket of wind-blown sand on the surface (fig. R-3).

78.1 We are driving east of a body of collapsed river sediment that came out of an interlobe area of the Streeter moraine. The lwo loops of the Streeter moraine are not especially conspicuous here (fig. R-3).

82.2 Stutsman County line.

85.2 Angle right. Disintegration trenches across ridge crest on left.

85.7 Leave collapsed river sediment; enter area of dead-ice moraine (fig. R-23).

87.2 Junction N.D. 30; turn south (right).

88.3 Enter area of collapsed river sediment (fig. R-23).

90.2 Note flat-topped hill 2 miles to southeast (left, ahead). This is the perched, ice-walled-lake plain of the next stop. Dead-ice moraine for next 5.5 miles (fig. R-23).

93.2 Turn east (left).

95.2 Turn north (left).

95.8 Top of ice-contact face; ice-walled-lake plain, about 1/2 mile across.

96.4 Stop 13.--Roadcut exposures of ice-walled-lake sediment on the ice-contact face at the north edge of the lake plain. Note the collapse structure that was formed when the supporting ice wall melted away (fig. R-23). Continue on dead-ice moraine.

98.0 Collapse river sediment for 3 miles.

101.2 Junction Interstate 94; turn east (right). Continue across typical dead-ice moraine (fig. R-23 and R-3).

109.7 Cleveland Exit.

113.7 Windsor Exit.

114.2 Leave Missouri Coteau, (high relief collapsed superglacial till; dead-ice moraine) enter Drift Prairie (low-relief collapsed superglacial till or subglacial till; "ground moraine"). Begin to descend the Coteau Escarpment, which is here very vague. We will drop 200 feet in next 5 miles (fig. R-3).

119.6 Meltwater channel of Minneapolis Flats Creek (1/2 mile wide); base of Missouri Escarpment (fig. R-3).

120.2 Eldridge end moraine, a 1 mile wide band of hummocky topography (fig. R-3).

121.2 Flat to undulating "ground moraine" covered with washboard moraines (see discussion at mile 0.4, first day).

122.7 Eldridge Exit.

127.9 Jamestown Exit.

129.5 Descent into the James River spillway.

End of field trip.

30-A

# STAGNANT-GLACIER FEATURES OF THE MISSOURI COTEAU IN NORTH DAKOTA

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## CONTENTS

CONTENTS (cont.)



## ABSTRACT

About 13,000 B.P. the late Wisconsin glacier advanced to the present western edge of the Missouri Coteau, forming the Burnstad end moraine in south-central North Dakota. Stagnant ice accumulated behind the Burnstad moraine, and about 12,000 or 12,500 B.P. active ice formed the Streeter end moraine about 15 miles east of the Burnstad. The Streeter moraine was formed against stagnant Burnstad ice.

The active ice rapidly thinned during the general deglaciation near the close of the ice age. Where the thin ice flowed over the Missouri Escarpment onto the Coteau, it experienced compressive flow and intensive marginal thrusting, which dragged great quantities of subglacial drift up into the ice. As the ice melted, the till was concentrated on its surface, forming a nearly continuous blanket of superglacial till several tens of feet thick over most of the Coteau. This drift blanket insulated the ice, causing it to melt very slowly. A nearly continuous sheet of stagnant ice, initially a few hundred feet thick, covered almost the entire Coteau. Much of this ice persisted for at least 3000 years, until 9000 B.P.

The superglacial drift blanket was irregularly distributed, and, as a result, the ice melted irregularly, and the topography on the stagnant ice became hilly and pitted with irregular depressions. The superglacial till was water-saturated and was highly plastic or fluid; it slid or flowed down the ice slopes and accumulated in depressions. Mudflows were common. The thick drift in the depressions caused the underlying ice to melt more slowly, and the newly exposed ice melted more rapidly, resulting in continual inversions of the superglacial topography.

Locally, in lower areas, there had been extending flow, with no marginal thrusting, and no accumulation of superglacial till. The ice in these areas melted faster, forming large depressions that held lakes or superglacial valleys occupied by streams. The streams deposited much superglacial alluvium.

This environment was at first very active, with rapid melting of the ice, much mass movement of the superglacial drift, and continual alterations of the superglacial topography. As the ice continued to melt, the superglacial drift cover became thicker, causing the ice to melt more slowly. The environment gradually became more stable, and the water in the superglacial and ice-walled lakes and streams became more temperate; most of the water was runoff from local precipitation, rather than glacial meltwater. Fish were able to migrate up the superglacial streams from the Missouri tributaries. They carried the parasitic glochidia (larval stage) of at least three of the four species of mussels now found in the Missouri drainage. Birds probably brought aquatic snails and ostracods in mud on their feet into the lakes and streams in the stagnant glaciers. The algae Chara sp. (stonewort) and other aquatic plants also became established. Surrounding the lakes and streams, the superglacial drift was covered with spruce, tamarack, birch, poplar, aquatic mosses, and associated vegetation. The mean annual precipitation was probably several inches higher, and the mean annual temperature was a few degrees cooler than at the present time.

Eventually, all of the ice melted, and all of the superglacial drift was let down, forming the hilly collapse topography that is characteristic of the Coteau today. The plains of the ice-walled lakes were left perched above the surrounding collapsed drift.

The drainage was then non-integrated; there were few if any streams, and the lakes had no outlets. The aquatic environments then slowly began to deteriorate as salts accumulated. In contrast to the period 9,000 to 12,000 years ago, mussels do not now occur, and gill-breathing snails are now a minor element of the fauna.

The stagnant-ice environment on the Coteau was in many ways comparable to the stagnant, drift-covered parts of the Bering, Malaspina, and Martin River Glaciers of south-central Alaska today.

## INTRODUCTION

The Missouri Coteau (or Coteau du Missouri) is a geographic district extending about 800 miles northwestward from east-central South Dakota, through North Dakota, and into southwestern Saskatchewan. It is a band of morainic hills ("coteau" is French for "little hill") averaging about 30 miles wide. It occupies the northeast edge of the Missouri Plateau and is bounded on the northeast by the Missouri Escarpment, which is 200 to 400 feet high. On its west side, the Missouri Coteau is bounded by the Coteau Slope, an area characterized by integrated drainage and a scarcity of morainic topography (fig. R-1 of roadlog). (See paper 30-D for discussion of physiographic terminology.)

The outstanding characteristic of the Coteau is its hilly topography. It consists of closely-spaced hummocks or knobs alternating with marshy depressions called "sloughs" or "prairie potholes." There is an almost complete lack of streams or stream-cut valleys on the Coteau; the topography is almost entirely glacial in origin. The drift, which is a few hundred feet thick in many areas, generally obscures the preglacial topography.

Most of the landforms on the Missouri Coteau are the result of the collapse of drift that covered a nearly continuous sheet of stagnant glacial ice which melted between about 9,000 and 12,500 years ago. The characteristic landforms are dead-ice moraine, collapsed-stream-sediment topography, ice-walled-lake plains, collapsed-lake-sediment topography, and associated disintegration ridges and trenches. End moraines on the Coteau tend to be ridged thrust moraines similar to those described by Kupsch (1962).

## GEOLOGIC INVESTIGATIONS

The moraine of the Missouri Coteau in North Dakota was first recognized as a result of ablation by Townsend and Jenke (1951) and was called dead-ice moraine by Lemke and Colton (1958). The stagnant-glacier origin of the topography of the Missouri Coteau in Saskatchewan was recognized by Christiansen in 1956. A comprehensive summary of the features associated with late Wisconsin glacial stagnation in Saskatchewan and Alberta was presented by Gravenor and Kupsch (1959). The glacial geology of the entire Missouri Coteau in Saskatchewan has been mapped by Christiansen and associates of the Saskatchewan Research Council.

The Missouri Coteau in North Dakota is described in the following reports: Hard (1929), Rau and others (1962), Clayton (1962), Winters (1963), Kume and Hansen (1965), and Bluemle (1965). Much of the information in this paper is from these reports plus North Dakota Geological Survey bulletins in preparation by D. E. Hansen (Divide County), T. F. Freers (Williams and Burke Counties), Clayton (Mountrail County), and J. P. Bluemle and others (McLean and Wells Counties), and open-file reports by Sherrod and Gustavson (Sheridan County). A preliminary glacial-geology map of North Dakota showing stagnation features on the Coteau was prepared by Colton and others (1963).

## STAGNANT-ICE FEATURES

Dead-ice moraine.--The most characteristic landform on the Missouri Coteau is dead-ice moraine (also called collapsed-superglacial-till topography, stagnation moraine, disintegration moraine, hummocky disintegration moraine, hummocky moraine, collapse moraine, or ablation moraine), which covers about 90 percent of the Coteau.

The dead-ice moraine of the Coteau is composed of till consisting of about equal parts sand, silt, and clay, plus a small percentage of gravel; it is a slightly-gravelly loam (slightly-gravelly sandy mud). The clay-size fraction is dominantly montmorillonite.

The till is a few tens of feet to a few hundred feet thick. The "ablation till" of dead-ice moraine is not noticeably different in grain-size composition or compaction from associated subglacial or "lodgement till."

The topography of dead-ice moraine consists of closely-spaced hills and depressions averaging a few hundred feet in diameter. They are roughly equidimensional in plan and generally lack live-ice features. Differences in elevation between adjacent depressions and hills range from 20 to more than 200 feet. Slopes are steep, 5 degrees (9 percent) to 15 degrees (27 percent) or more. The depressions between the hills contain marshy sloughs averaging 300 feet in diameter; there are about 500,000 sloughs on the Coteau in North Dakota. The larger depressions contain lakes; there are about 50,000 lakes between 500 feet and 1 mile in diameter.

Origin.--Intensive marginal thrusting in the glacier brought large amounts of subglacial drift up into the glacier; the drift was concentrated on the surface of the ice as it ablated. This superglacial till insulated the ice and allowed a large, nearly continuous sheet of stagnant ice to persist for a few thousand years on the Coteau. The superglacial till was irregularly distributed, causing uneven melting of the stagnant ice and resulting in a very irregular ice surface. The superglacial till, having a high moisture content (from the melting ice) and a high montmorillonite content, was very responsive to slight changes in superglacial topography and was continually slumping or flowing to lower areas on the ice. Ice in these low areas, insulated with a thick layer of till, melted more slowly, while the higher, newly-exposed ice melted more rapidly. Thus, there were continuous inversions of the topography on the glacier. The final result was the hummocky topography so characteristic of the dead-ice moraine on the Coteau today.

Mass movement.--The dominant depositional process in dead-ice moraine apparently was flowage and sliding. Evidence for this comes in part from an analogy with the collapsed-superglacial-lake sediments that are associated with the dead-ice moraine; both have similar topography and are thought to be of similar origin. As will be shown in a later part of this report, the contortion and disruption of the bedding of collapsed lake sediment indicates that sliding and flowage are the dominant processes during collapse of superglacial-lake sediment.

Because the till of dead-ice moraine is generally free of bedding, evidence for flowage and sliding of superglacial till is less obvious than in collapsed stratified sediment. Only where the superglacial till or "flowtill" (Hartshorn, 1958) has been interbedded with stratified, washed drift is this origin obvious. An excellent example of flowtill interbedded with washed drift was exposed in several miles of new roadcuts in northwestern McIntosh County (Clayton, 1962, p. 39). Here mudflows that are a few feet thick were deposited in a superglacial lake. Most of the individual mudflow units have the appearance of normal homogeneous till, but some contain highly contorted bedding or possibly flow laminations. Between the individual flows are thin beds of laminated lake sediment or sand. Sole markings that are similar to those frequently found in turbidite sequences have been observed on the bottom of some of the mudflow beds. Much dead-ice moraine might also be called collapsed-superglacial-mudflow topography.

Slope angles.--Slope angles (excluding seepage slopes; see below) in dead-ice moraine are apparently closely related to the moisture content of the till at the time of deposition. During his studies in the Yorkton area, Saskatchewan, John Cherry (letter, February 1967) observed that in any given area of dead-ice moraine there exists, within narrow limits, a characteristic maximum slope angle. This slope angle approaches the steepest equilibrium slope that existed during the period of collapse.

In parts of the Coteau where maximum slope angles in dead-ice moraine are low (4 degrees or less), as in the southwestern half of Mountrail County, the superglacial drift was relatively thin. Local relief is low, drift exposed in roadcuts is no more than a few tens of feet thick, and the ice-walled-lake sediments indicate rapid deposition; the superglacial environment was unstable (see paper 30-B, following this paper). Where maximum slope angles in dead-ice moraine are steep, as in the hilliest part of the Coteau in northeastern Mountrail County, the superglacial drift was thick. Local relief is high, the uppermost drift sheet is at least several tens of feet thick, and the ice-walled-lake sediments indicate slow deposition; the superglacial environment was relatively stable.

This correlation between maximum present-day slope angles and presumed superglacial drift thicknesses suggests that moisture content at the time of deposition controlled the fluidity or plasticity of the superglacial drift. Superglacial drift thickness controlled the rate of ice melting, and this in turn controlled the moisture content of the drift. Higher moisture content resulted in a lower maximum angle of slope at which the drift could remain stable.

Seepage slopes.--Postglacial erosion has altered the original slope angles. Several feet of sediment has been washed into the sloughs from adjacent hillslopes. This, in general, has reduced slope angles throughout the Coteau. The steepest slopes (commonly 25 degrees), however, occur at the edges of the larger sloughs. These steep slopes are not ice-contact faces or wave-cut faces, as might be thought, but are groundwater-seepage faces. Intermittent or permanent discharge of groundwater tends to sap material from the base of these slopes, causing them to become much steeper than nearby groundwater-recharge slopes.

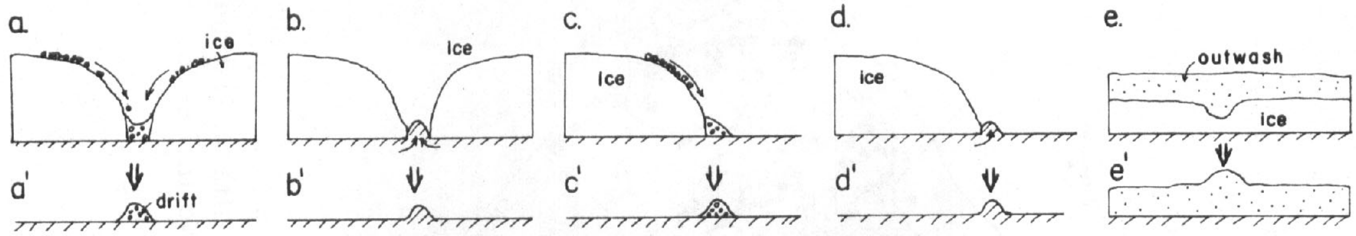


Figure A-1. Five methods of forming disintegration ridges; diagrammatic cross-sections.

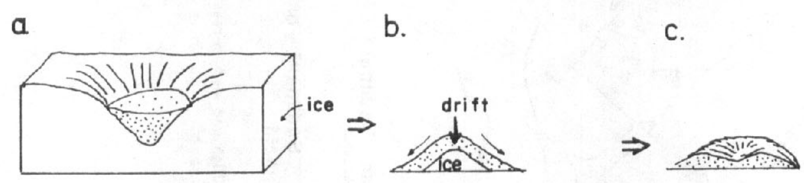


Figure A-2. Three stages in the formation of a circular disintegration ridge ("doughnut").

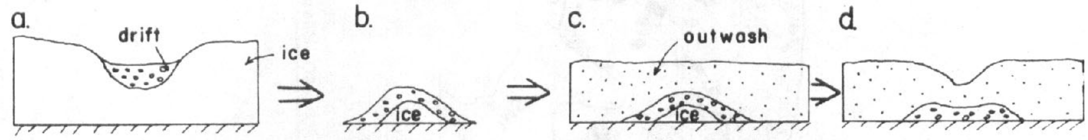


Figure A-3. Four stages in the formation of disintegration trenches.

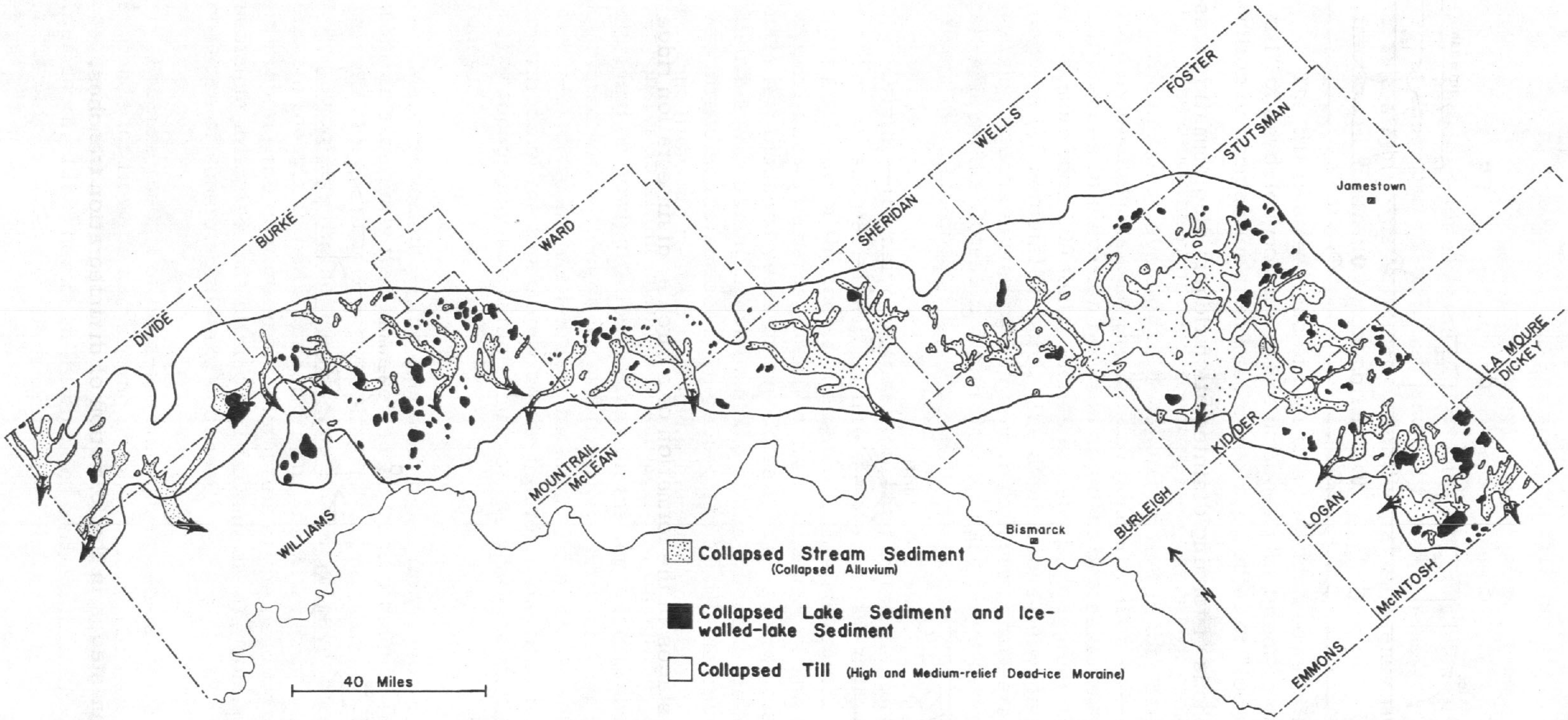


Figure A-4. Generalized distribution of high-relief and medium-relief dead-ice moraine, collapsed stream sediment or alluvium, ice-walled-lake sediment, and collapsed lake sediment on the Missouri Coteau in North Dakota. From the sources given in section on geologic investigations.

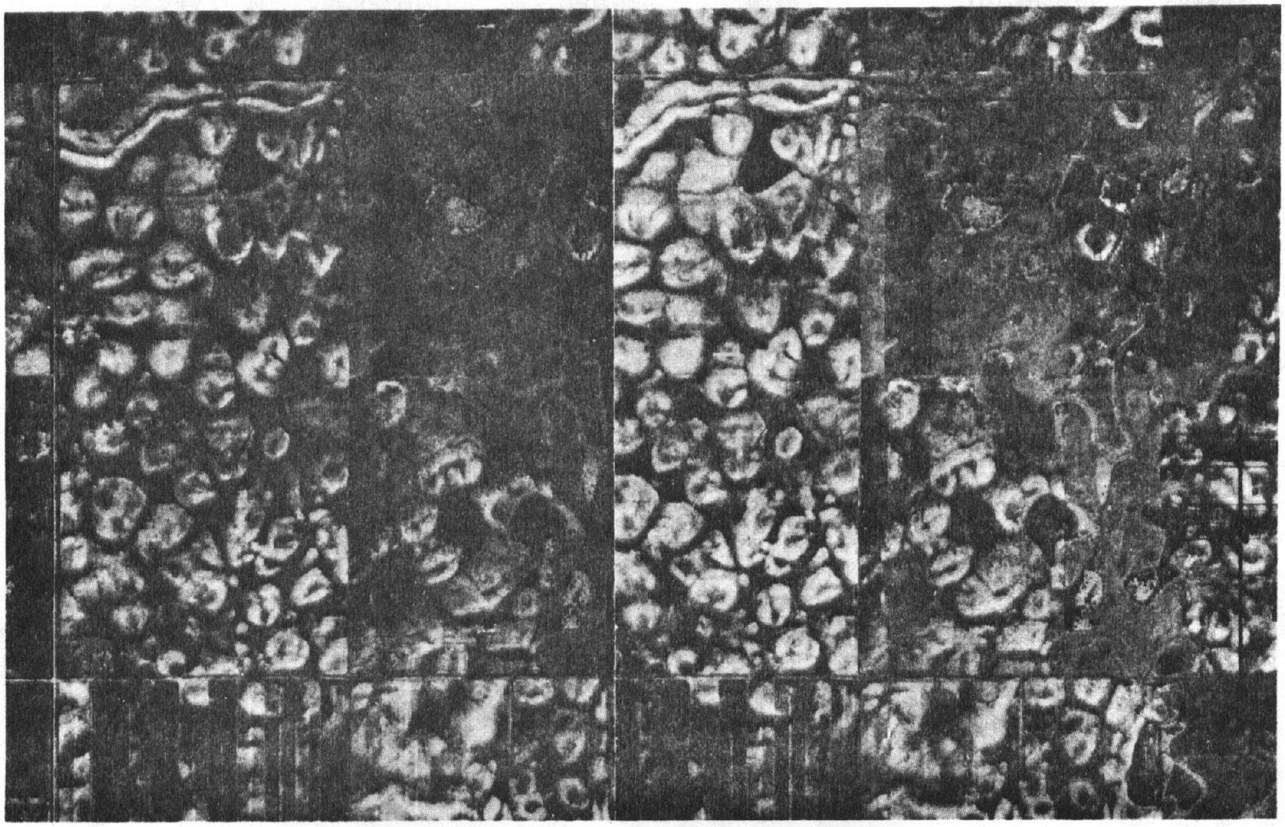

Figure A-5. Stereopair of circular disintegration ridges having high centers. Typical straight, double disintegration ridge in upper-left part of photo. In sec. 6, T. 161 N., R. 102 W., Divide County. (U.S. Dept. of Agriculture CC-3BB-83 and 84, 9-15-61.)

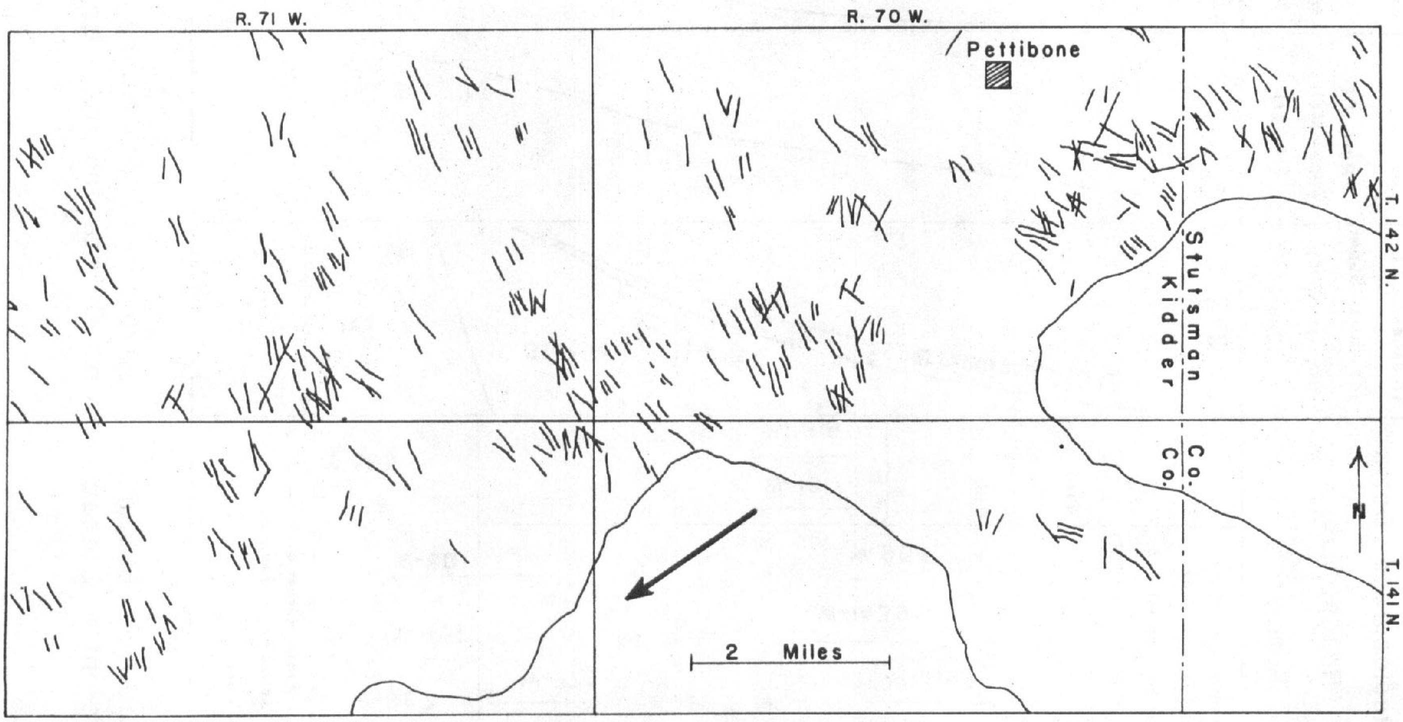


Figure A-6. Disintegration trenches in northeastern part of Kidder Sand Plain; traced from air photos; arrow shows direction of ice movement, as determined from orientation of end moraines.

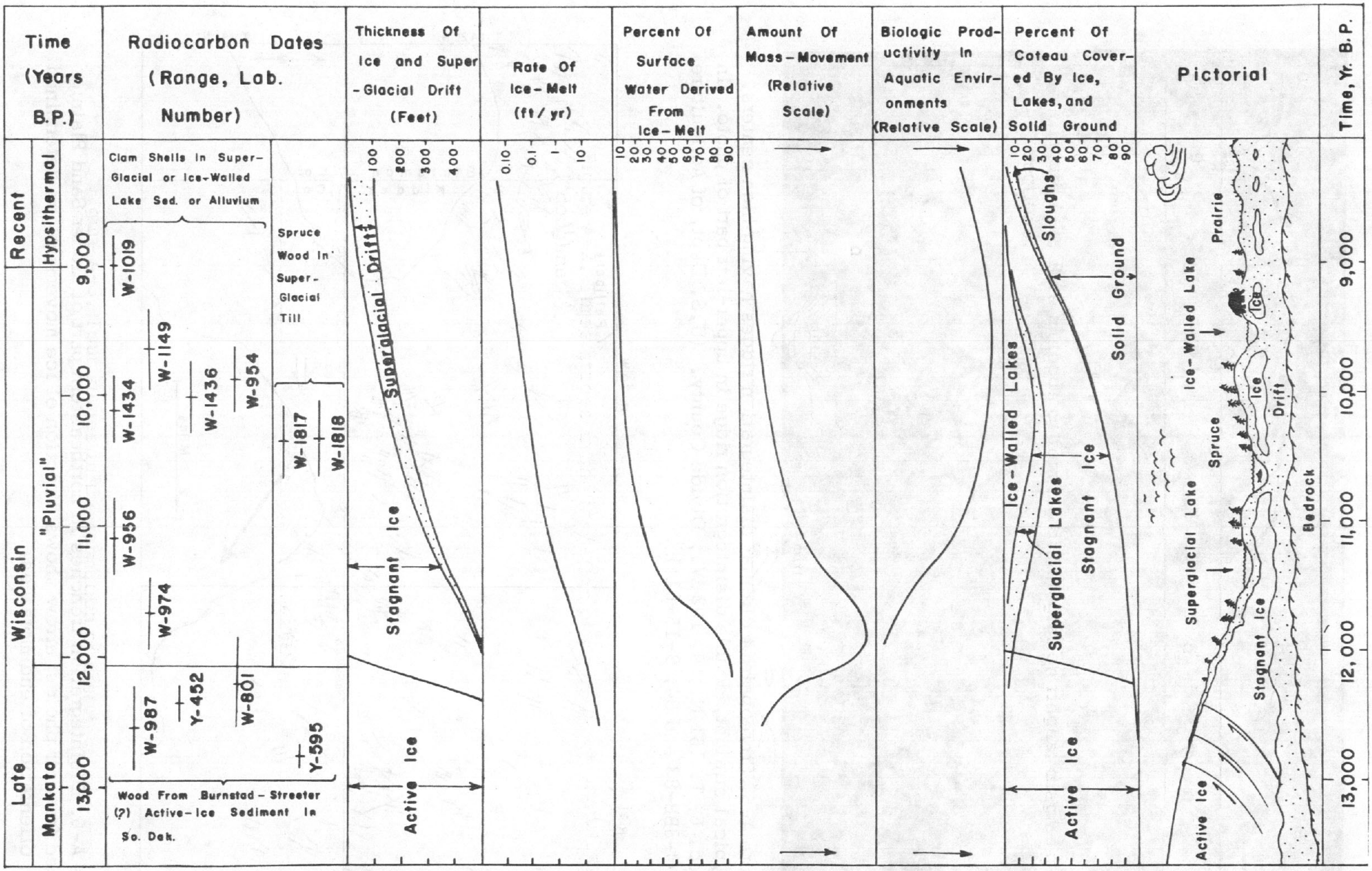


Figure A-7. Evolution of stagnant-ice environment on the Missouri Coteau. Numerical values are averages for the entire Coteau in North Dakota and are estimates only.

This may seem to conflict with the conclusions of the previous section on slope angles. The till of discharge areas is now saturated with water and should be stable at much lower slope angles than the adjacent slopes that were originally formed at presumably unsaturated conditions. Till slopes are now much more stable, however, because of compaction resulting from desiccation during the "hypsithermal" period when water-table levels were lower than at present.

Topographic density.--Local relief and slope angles of dead-ice moraine range greatly in magnitude. The size and spacing of topographic elements is relatively uniform, however. Individual elements (such as circular disintegration ridges) averages about 550 feet in diameter, and about 90 of these elements occur per square mile (see fig. 5-A).

The size and spacing of topographic elements of dead-ice moraine is inherited from the topography of the stagnant glacier. A depression in dead-ice moraine was a hill of stagnant ice, and a hill in dead-ice moraine was an ice depression during the last stage of melting of the stagnant ice. It should, therefore, be expected that the topography density on the stagnant glacier was the same as that of dead-ice moraine.

Beneath a depth of about 150 feet, ice of temperate glaciers tends to become plastic; ice depressions deeper than about 150 or 200 feet are unlikely. If maximum probable ice slopes were about 40 degrees, and maximum depression depths were 200 feet, the average ice depression would have been about 550 feet in diameter. As is shown in a later section of this report, these values are close to the equivalent values for depressions in modern stagnant glaciers.

Disintegration ridges.--Disintegration ridges (Gravenor and Kupsch, 1959, p. 52-54) are commonly found in dead-ice moraine, though they also occur in collapsed lake sediment and collapsed stream sediment. The ridges are commonly a few tens of feet high and a few hundred feet to a few miles long. The slope angles of their sides are the same as the angles that are characteristic of the surrounding dead-ice moraine.

Disintegration ridges originated when drift flowed or slid into a crevasse or channel in stagnant ice (fig. A-1a) or off the edge of a mass of ice (fig. A-1c). They may also have originated when drift was squeezed from beneath the ice (fig. A-1b and A-1d) or was washed into a crevasse or channel by superglacial streams (fig. A-1a, A-1c and A1-e). (See paper 30-I.)

Disintegration ridges tend to be straight, if they originated in crevasses, or braided or meandering, if they originated in stream channels in the stagnant ice--in which case they may be difficult to distinguish from eskers (fig. A-1a, A-1e). Circular disintegration ridges form at the edge of circular holes in the ice, and have been called "doughnuts."

"Doughnuts."--"Doughnuts" are circular ridges a few hundred feet across and, at most, a few tens of feet high. They are best developed in medium-and low-relief dead-ice moraine and collapsed lake sediment; they are uncommon in high-relief dead-ice moraine and collapsed stream sediment.

In glacial-stagnation areas, "doughnuts" are usually interpreted (Gravenor, 1955) as circular disintegration ridges formed by (a) sliding or flowing of superglacial till or lake sediment into a sinkhole in the stagnant ice, (b) inversion of topography as a result of the insulating effect of the drift in the bottom of the sinkhole, followed by mass movement of this drift away from the center and down the sides of the buried ice core, and (c) melting of the ice core (fig. A-2). Because the centers of adjacent circular disintegration ridges average 550 feet apart, the ice sink-holes in which they formed should also have averaged about 550 feet in diameter on the Coteau. As will be shown later, this is near the average size of sinkholes on modern stagnant glaciers.

"Doughnuts" are commonly doubly breached on opposite sides of the ring. Commonly several breached "doughnuts" are aligned in a row like a chain. These breaches are short disintegration trenches resulting from collapse of a buried ice-cored crevasse filling. Disintegration trenches occur at these positions because ice sinkholes are commonly initiated along a crevasse, as pointed out by Russel (1901, p. 116). A "chain" of breached "doughnuts" occurs where several ice sinkholes formed along a single crevasse.

Bik, however, believes that many of the "doughnuts" in western Canada are collapsed pingos; (see paper 30-F).

Collapsed stream sediment.--Collapsed superglacial-stream sediment on the Coteau is well-sorted and well-bedded gravelly sand or sandy gravel. It is primarily channel sediment; overbank sediment is uncommon. It differs from uncollapsed stream sediment in having faulted bedding and rolling or hilly topography similar to that of dead-ice moraine. Fossil aquatic snails and clams are common; (see paper 30-E).

As is shown later, the water that deposited this sand and gravel was mostly runoff from local precipitation. Only a small fraction of it was glacial meltwater. For this reason the sediment is referred to as "stream sediment" or "alluvium," rather than "outwash." ("Alluvium" is here considered to be a general term for any stream sediment, including both glacial stream sediment (outwash) and nonglacial stream sediment.)

Collapsed stream sediment contains abundant undrained depressions. Flat-topped hills, which are uncollapsed remnants of the former alluvial plain where it was deposited on solid ground rather than on the ice, are present in some areas of collapsed stream sediment. Straight disintegration ridges of the type "e" in figure A-1 are common, but circular disintegration ridges are rare. Disintegration trenches are also common; meandering or long, straight disintegration trenches are so characteristic of collapsed stream sediment that they may confidently be used to distinguish it from dead-ice moraine on air photos.

Collapsed stream sediment may occupy the highest parts of an area. Its distribution generally shows little relationship to any present-day valleys or low areas. Much of it is in the form of linear bodies a mile or two wide that extend several miles southwestward across the Coteau; these are "collapsed valley trains" deposited in ice valleys on the stagnant glacier (fig. A-4).

The sand and gravel deposits labeled "collapsed alluvium" in figure A-4 are known to be collapsed sediments of former superglacial alluvial plains and superglacial valley trains because of (a) the faulted bedding and other collapse structures, (b) the presence of hilly topography with abundant undrained depressions, (c) its elevated position above adjacent dead-ice moraine, and (d) the presence of numerous disintegration trenches.

The origin of superglacial stream sediment is related to "pluvial" runoff and hillslope stability. Paleobotanical evidence suggests that precipitation during this time (9,000 to 12,500 B.P.) was greater than at present. Non-glacial stream channels formed on the Coteau Slope and the Drift Prairie at the time, commonly have meander radii five times as great as the meander radii of the underfit streams flowing in them today. For these reasons, the period 9,000 to 12,500 B.P. is thought to have been a "pluvial" period (fig. A-7) (see Schumm, 1965, p. 786-788). Because of the melting of the underlying ice, the superglacial environment was unstable, and superglacial till frequently slumped into the "pluvial" rivers. Thus, the superglacial rivers on the Coteau produced much more alluvium than those flowing on the stable solid ground of the Drift Prairie to the east.

Disintegration trenches.--Disintegration trenches (Clayton, 1962, p. 42) characteristically are associated with collapsed stream sediments. They are shallow trenches, generally only a few feet deep, and are best observed on air photos. Most are straight, only a few tens of feet wide, and from several hundred feet to as much as a mile in length. Straight disintegration trenches generally show a strong preferred orientation perpendicular to the ice-movement direction (fig. A-6). Others have polygonal or meandering pattern. They may extend over hill tops or through depressions, where they may become ridges (see stop 12 in the roadlog).

Disintegration trenches probably originated as follows (fig. A-3): (a) a superglacial channel or crevasse was filled with drift; (b) an ice-cored esker or ice-cored crevasse-filling resulted when differential melting due to the insulating effects of the drift caused topographic inversion; (c) these ice-cored ridges were then buried under stream sediment; and (d) the buried ice cores then melted, leaving disintegration trenches.

Ice-walled-lake plains.--Ice-walled-lake plains are flat to undulating elevated plains that are perched above the surrounding dead-ice moraine like buttes or small mesas. They are underlain by flat-lying, finely-laminated, pebble-free silt and clay, which contrasts with the slightly pebbly till of the surrounding dead-ice moraine. Most contain fossil mollusks and ostracods. Ice-walled-lake plains are scattered throughout the Coteau (fig. A-4). These lake sediments were deposited in depressions in the stagnant ice sheet, as shown in figure R-13 and R-19 of the roadlog.

Ice-walled-lake plains are discussed in more detail in paper 30-B.

Collapsed lake sediment.--Collapsed lake sediment (see paper 30-B) is similar in origin to collapsed stream sediment; it was deposited in superglacial lakes and later collapsed and was let down as the underlying ice melted. Like collapsed stream sediment, its topography is similar to that of dead-ice moraine, but tends to be more subdued. Because of the higher clay content, it was stable at lower slope angles than till. Like ice-walled-lake sediment, it is finely-laminated silt and clay, which commonly contains fossil mollusks and ostracods. Unlike ice-walled-lake sediment, its bedding tends to be broadly folded or complexly contorted, in contrast to that of collapsed stream sediment which tends to be faulted (fig. B-2 of paper 30-B and fig. R-14 and R-15 of roadlog).

Only about half of the lake sediment mapped on the Coteau (fig. A-4) is collapsed lake sediment. In some parts of Saskatchewan (paper 30-B) collapsed lake sediment is much more widespread than in North Dakota.

## FURTHER INTERPRETATIONS

Extent of stagnant ice.--The entire Coteau was covered with a nearly continuous sheet of stagnant ice. There are two kinds of evidence for this:

(a) Radiocarbon dates indicate that lakes and streams on or bounded by stagnant ice existed in various parts of the Coteau for at least 3000 years. The lake and stream deposits now in elevated positions were then surrounded on all sides by stagnant ice. Few if any lakes or streams existed in intervening areas because lake or stream sediments are absent in these areas. The stagnant ice there must have been at higher elevations; these were ice-cored watershed areas. Because ice existed under both the low areas and the divides, the stagnant ice sheet must have been nearly continuous.

(b) Masses of collapsed stream sediment extend almost entirely across the Coteau in many places (fig. A-4). They are elevated, are hilly, and contain numerous undrained depressions, indicating that they were underlain by stagnant ice. Many deposits of collapsed stream sediments extend southwestward several miles to rivers of the Coteau Slope. Therefore, most of the rivers, the deposits of which now have non-integrated topography, must have flowed southwestward across the Coteau from an elevated surface underlain by a nearly continuous mass of stagnant ice.

During later stages of melting, the ice sheet became more discontinuous. The central parts of the upper-most beds of most ice-walled-lake deposits are uncollapsed because they were deposited after the basal lake deposits had been lowered to solid ground. Some areas of flat, uncollapsed "pluvial" stream sediment occur in parts of the Coteau; examples are the southwestern part of the Kidder Sand Plain and most of the 5-mile-wide sand plain in front of the Streeter moraine in north-central Logan County. Most of the Burnstad and Streeter end moraines in the southern part of the Coteau were deposited on solid ground.

Fossils.--Fossil aquatic snails, clams, ostracods, stoneworts, (*Chara sp.*), and other aquatic animals and plants occur in most deposits of ice-walled-lake sediment, collapsed superglacial-lake sediment, and collapsed superglacial-stream sediment on the Coteau (see paper 30-E).

More than two dozen species of mollusks have been identified; none are extinct. The faunal association is characteristic of that found in fresh-water lakes at this latitude in Minnesota today. Most present-day Coteau lakes are too saline to support such a diverse fauna.

Cvancara (University of North Dakota, conversation, February 1967) has observed that the three species of mussels found in superglacial washed drift on the Coteau are the three common species (only four are known) now found in the Missouri River drainage in North Dakota. None of the 10 species restricted to drainage east of the Coteau in North Dakota has been found on the Coteau. The distribution of collapsed stream sediment (fig. A-4) indicates that nearly all of the superglacial streams flowed westward into Missouri River tributaries; few if any flowed eastward down the Missouri Escarpment. Mussels are parasitic on fish during their glochidial stage and may be carried throughout a drainage basin by them. For this reason the mussels in the superglacial streams and lakes originated in the Missouri River and its tributaries rather than in rivers east of the Coteau.

The superglacial or ice-walled aquatic environments were "nonglacial" and supported life 9,000 to 12,000 years ago because they were well insulated from the stagnant ice by a thick blanket of drift. An open, boreal spruce forest grew adjacent to the water bodies. Wood, needles, cones, and pollen are commonly found in the base of slough deposits (see paper 30-H) or mixed with slumped superglacial drift. At least three spruce samples from the Coteau have been radiocarbon dated. Because the present vegetation is prairie grasses, it seems reasonable that the annual precipitation was perhaps 5 or 10 inches greater than the present 17 inches. Other evidence (see section on collapsed stream sediment) indicates that runoff was perhaps five times as great as at present. Only about an inch of meltwater was produced in a year, as is shown in the next section. Therefore, the meltwater contribution to the ice-walled and superglacial aquatic environments amounted to only a few percent of the total (fig. A-7). For this reason the temperature of the water in these environments was controlled almost entirely by the prevailing climate, not by the underlying ice.

Rate of melting.--The stagnant ice lasted for at least 3000 years. Clam shells from six different sites in superglacial or ice-walled stream or lake deposits on the Coteau in North Dakota have been radiocarbon dated by the U.S. Geological Survey, giving dates ranging from 11,650 to 9,000 B.P. (Clayton, 1966). Wood collected from collapsed superglacial till (paper 30-K) in Ward County has been dated at 10,350 B.P. (W-1817) and 10,330 B.P. (W-1818). Other dates indicate that the ice became stagnant about 12,000 or 12,500 B.P. (fig. A-7). Most of North Dakota northeast of the Coteau was free of ice after about 12,000 B.P. Thus, the glacial landforms and surface sediments of most of the Coteau are younger than glacial features a few hundred miles "up-glacier" to the northeast. Parizek (1964, p. 41) also concluded that radiocarbon dates from the Coteau in Saskatchewan indicate that stagnant ice lingered there for at least 2000 years.

Topographic considerations suggest that the stagnant ice on the Coteau averaged about 300 feet thick 12,000 years ago. For instance, the Streeter end moraine, which is 200 to 300 feet high, was, in most places, deposited against stagnant Burnstad ice. River sediment was deposited on the Burnstad ice by water flowing across the Streeter moraine, indicating that the stagnant Burnstad ice was 200 to 300 feet thick at that point.

Using an average of about 300 feet of ice, which required about 3000 years to melt, the ice-melt rate was roughly an inch per year. At first the rate was probably much greater when the superglacial drift was thin. Later, as more drift became concentrated on the ice surface, the rate was probably less than an inch per year (fig. A-7).

Evidence for superglacial drift.--Superglacial drift has been assumed to have existed on the Coteau. However, Stalker (1960a, 1960b, p. 34) and others have suggested that continental glaciers were relatively clean and that stagnation features are largely the result of the squeezing of subglacial drift up into holes and crevasses in the base of a stagnant ice sheet. However, Weertman (1966) has shown that continental glaciers should theoretically have englacial drift in a basal zone several hundred feet thick, and Behrendt (1963) has shown that this basal zone in glaciers in Antarctica contains several percent of englacial debris. Evidence for the existence of superglacial drift on the stagnant late Wisconsin ice sheet on the Coteau is as follows:

(a) Insulation of ice from aquatic environments.--The diverse association of aquatic animal and plant fossils indicates that the superglacial and ice-walled lakes and streams were more hospitable environments for gill-breathing mollusks than the present, rather saline, lakes and streams of this area. Such environments would have been possible only if the ice melted slowly; rapid melting would have produced large quantities of cold, silty water, making the lakes and streams improbable environments for plants and animals known to have been present. The climate at the time (12,000 to 9,000 B.P.) was apparently rather temperate. Spruce and tamarack needles, wood, cones, and pollen, grass, and other plant material occur in the base of sloughs in the Coteau (paper 30-H). One of these spruce samples from Kidder County was dated at 11,430 B.P. (Moir, 1958), and two spruce samples from Ward County (paper 30-K) were dated at 10,350 B.P. (W-1817) and 10,330 B.P. (W-1818). The combination of fairly temperate climate and slow rate of melting indicates that the ice was largely covered with a blanket of fairly thick superglacial drift.

(b) Radiocarbon dates.--The previously mentioned radiocarbon dates indicate that the ice required at least 3000 years, from 12,000 to 9,000 B.P., to melt. This would have been possible only if the ice were buried under superglacial drift.

(c) Collapse topography.--As mentioned previously, the contorted, folded, or faulted bedding of collapsed stream sediment and lake sediment indicates that these deposits had accumulated on stagnant ice. Their topography is nearly identical to that of dead-ice moraine, indicating it also formed in the same manner. This and other evidence presented in the section on dead-ice moraine indicates that it originated by the collapse of superglacial till when the stagnant ice melted.

<u>Cause of stagnation.</u>--The margin of any glacier may stagnate. The actual processes of stagnation on the Coteau was probably in no way different from the stagnation at the margin of most other glaciers. That is, the entire glacier on the Coteau need not have stagnated at one time in a zone many miles wide. Rather, there was probably normal stagnation along a narrow marginal zone. However, each successive zone of stagnant ice persisted because of the insulating superglacial drift, allowing a continuous mass many miles wide to accumulate. That is, there was probably progressive marginal stagnation rather than <u>en masse</u> stagnation.

<u>Superglacial drift thickness.</u>--Test drilling has shown that collapsed superglacial-stream sediment and collapsed superglacial-lake sediment is typically a few tens of feet thick. However, it is more difficult to determine the thickness of the superglacial till because the till stratigraphy is generally poorly known. Superglacial ("ablation") till and basal ("lodgement") till on the Coteau are nearly identical in grain-size distribution and compaction; they have not been differentiated in most places. Furthermore, the uppermost till sheet usually cannot be differentiated from older till sheets because they are nearly identical in mineralogy, and weathering zones have not been recognized in most places. There is, however, some evidence for the thickness of superglacial till.

Thin superglacial till.--In the Coteau Slope in Mountrail County (fig. R-1 of the roadlog and A-4), postglacial gullying has provided numerous complete drift sections, and thickness of the uppermost till sheet can be accurately determined. It averages 10 or 20 feet thick in regions where the dead-ice moraine has slope angles that are much lower (2 degrees to 4 degrees) than those on the Missouri Coteau. Local relief in these areas is 5 to 15 feet.

Medium-thick superglacial till.--In the southwestern part of the Coteau in Ward County, test drilling has shown (paper 30-K) that the uppermost drift sheet is about 40 feet thick in areas of medium-relief dead-ice moraine (4 degrees to 7 degrees slopes). Local relief in these areas is about 30 or 40 feet.

Thick superglacial till.--Spruce wood from till in a test hole at a depth of 135 feet in the hilliest part of the Coteau in Ward County has been dated at 10,350 B.P. (W-1817; paper 30-K). The ice was apparently active in this area about 12,500 B.P. (Clayton, 1966). This indicates that 135 feet of drift slid or flowed into that area 2000 years after the ice had stagnated. The youngest drift sheet in this area averages about 100 feet thick (paper 30-K). Maximum slope angles in this area are generally greater than 7 degrees, and local relief may be 100 feet or more.

Thus, the superglacial drift thickness was roughly equal to the present local relief, and slope angles are closely related to the former superglacial drift thickness. (This was previously concluded from other evidence in the section on slope angles.) Further evidence for the correlation between superglacial drift thickness and present local relief and maximum slope angles is presented in the section on uncoalesced ice-walled lakes in paper 30-B.

Origin of superglacial till.--The superglacial till originated by ablation-concentration of material that was thrust up along imbricate marginal shear planes when the ice was active. Continental glaciers generally have relatively clean surfaces except along a narrow marginal zone of thrusting only hundreds of feet wide, not 30 miles wide as on the Coteau in late Wisconsin time. In areas of extreme compressive flow near the margin of a glacier, however, thrusting is more active and extensive.

Compressive flow can be caused by several factors (Nye, 1952), but only one, a reversal in regional slope, appears to have been significant on the Coteau. The bedrock escarpment beneath the Missouri Coteau (fig. R-10 of roadlog) was a barrier to the glacier, causing intensive compressive flow at the glacier margin. As the ice thickened and moved across the escarpment, large amounts of shearing brought subglacial drift to the surface of the glacier.

Evidence that slope reversal and consequent compressive flow were the controlling factors producing superglacial drift is the close correlation between topography and the distribution of dead-ice moraine. Most dead-ice moraine in North Dakota and surrounding areas covers small plateaus or lies along the edge of larger plateaus. Typical examples are the Missouri Coteau (the northeast edge of Missouri Plateau), the Prairie Coteau (a plateau in northeastern South Dakota; fig. R-1 of roadlog), the Turtle Mountains (an "outlier" of the Missouri Plateau in North Dakota and Manitoba; fig. R-1 of roadlog); Moose Mountain, Last Mountain, Touchwood Hills, and Porcupine Mountain in Saskatchewan; Duck Mountain and Riding

Mountain in Manitoba; and the Leaf Hills (Alexandria-Detroit Lakes area) in Minnesota. All of these are low plateaus rising a few hundred feet above the surrounding plains.

Compressive vs. extending flow.--It was previously shown that most straight disintegration trenches were initiated at ice crevasses. Disintegration trenches are characteristically perpendicular to the former ice-flow direction, as is illustrated in figure A-6. Transverse crevasses are perpendicular to ice-flow direction. Thus most disintegration trenches are the result of transverse crevasses. As was shown by Nye (1952), transverse crevasses occur where there is extending flow in a glacier. Therefore, most disintegration trenches probably occur in former areas of extending flow: most areas of collapsed stream sediment in the Missouri Coteau (fig. A-4) have transverse disintegration trenches, and most superglacial stream sediment therefore occurred in areas of extending flow.

It was concluded, however, that compressive flow was dominant on the Coteau. Extending flow occurred locally where the ice spread out or flowed into lower areas that sloped westward toward the Coteau Slope. Areas of extending flow are characterized by transverse crevasses rather than thrust planes, causing the ice to have little superglacial till. In areas of extending flow, the surface of the ice was probably low as a result of the factors causing the extending flow; in addition, the lack of an insulating blanket of superglacial till caused the ice in these areas to melt faster, making the areas of extension flow much lower than the surrounding drift-covered areas of compressive flow. Superglacial rivers flowed into these low areas, depositing superglacial alluvium. Therefore, superglacial alluvium was deposited mainly in areas of extending flow; transverse crevasses occurred in most ice covered by stream sediment; and transverse disintegration trenches are characteristic of most collapsed stream sediment.

Ice-thickness.--Marginal thrusting is more intense if a glacial terminus is thin (gentle surface gradient). The ice on the Coteau was apparently relatively thin in late Wisconsin time. Evidence for this includes the radius of curvature of moraine loops and the effect of subglacial topography on ice movement. On the Missouri Coteau, end moraine loops have a radius of curvature ranging from 2 miles to as little as 1 mile; small variations in subglacial topography greatly influenced ice movement, indicating that the ice was very thin when last active. In contrast, on the Drift Prairie, where there is little or no dead-ice moraine, the radius of curvature of end moraine loops is 6 miles or more, indicating that the last active ice was much thicker than on the Coteau.

The ice was thin on the Coteau in late Wisconsin time because the climate was "nonglacial" and the regime of the glacier was negative. In contrast, the climate was "glacial," and the glacier had a positive regime during the time (early Wisconsin?) when the Napoleon drift was deposited west of the Coteau in Logan County (see discussion at mile 125.7 of roadlog). There, high-relief dead-ice moraine was never formed. The terminus of the glacier was thick (steep surface gradient) and ice-movement was little affected by the subglacial topography. Only minor marginal thrusting occurred, and the ice surface was relatively clean. For this reason, high-relief dead-ice moraine is lacking west of the Coteau.

## SUMMARY OF EVENTS ON THE COTEAU

Events on the Coteau from the time the glacier stagnated, about 12,000 or 12,500 B.P., until most of the ice melted, about 9,000 B.P., are summarized in figure A-7. This was a "pluvial" period. Precipitation and runoff were greater than at present, and temperatures were a few degrees lower. During the first part of this "pluvial" period, the superglacial environment was very active, with rapid ice melt and much slumping of superglacial drift. Gradually the environment became more stable, with a smaller amount of meltwater and less slumping. Rainfall-runoff streams, tributary to the Missouri River, flowed westward across the Coteau, depositing large amounts of superglacial sand and gravel. The rivers and ice-depression lakes contained abundant plant and animal life, and spruce grew on surrounding areas. By about 9,000 B.P. most of the ice had melted, and the drainage became completely non-integrated. As the "hypsithermal" period began, the climate became dryer, prairie vegetation replaced the spruce, and many species of aquatic animals dies out as the sloughs began to accumulate salts.

## MODERN ANALOGUE IN SOUTH-CENTRAL ALASKA

Most of the above interpretations had been made by 1962. Some of the interpretations, such as abundant superglacial mollusk populations and stagnant ice that took thousands of years to melt, were considered to be somewhat novel. It was therefore believed that we needed a better understanding of what an actual stagnant-glacier environment is like. For this reason the Department of Geology at the University of North Dakota organized in 1962 an expedition to the stagnant, drift-covered part of the Martin River Glacier in south-central Alaska.

The terminal zones of south-central Alaskan glaciers such as the Martin River, Bering, and Malaspina Glaciers are similar in many ways to the former stagnant glacier on the Missouri Coteau. South-central Alaska has a temperate climate with a dense spruce-hemlock forest growing on the termini of the glaciers. However, the mean annual precipitation, a large percentage of which is rain, is over 100 inches, or four times the probable late Wisconsin precipitation on the Coteau. The area is very mountainous, but the Martin River, Bering, and Malaspina are large expanded-foot piedmont glaciers in the lowlands in front of the mountains. The radius of curvature of the ice lobes is nearly the same as it was on the Coteau and the ice thickness and the surface slope of the ice are probably also similar.

Drift grain size.--The most significant contrast between the Coteau and this area in Alaska is the grain size of the drift. The superglacial drift on the Martin River Glacier is a coarse rubble with angular blocks averaging a few inches in diameter at the surface. The drift contains more than one-half sand and gravel. In contrast, the superglacial till on the Coteau consisted of about one-third montmorillonitic clay, one-third silt, and one-third sand, with only a few percent gravel. The drift on the Coteau was very plastic when wet and must have been susceptible to mass movement at much lower slope angles than the rubble on the Martin River Glacier.

Superglacial drift thickness.--The terminal, stagnant parts of the Martin River, Bering, and Malaspina Glaciers are almost completely covered with superglacial drift. Along a marginal zone (nearly a mile wide) of the Martin River Glacier, the superglacial drift averages between 10 and 20 feet in thickness. On the drift is a dense full-grown spruce and hemlock forest with a ground cover of thick moss on an incipient soil. Of several trees bored, the oldest has more than 100 annual rings. Those on the Malaspina are of similar age.

Rate of melting.--The forest required time to become established on the superglacial drift of the Martin River Glacier, and much buried stagnant ice still remains. Therefore, the stagnant ice at the terminus of these Alaskan glaciers should take at least a few hundred years to melt.

The superglacial drift there is much thinner than it was on the Coteau. Furthermore, the more abundant rainfall quickly seeps through the more permeable superglacial drift. The rainwater greatly increases the rate of melting of the underlying ice, causing the stagnant ice of these Alaskan glaciers to melt much faster than the Coteau ice. For this reason, the previously mentioned 3000 years required for the stagnant ice to melt on the Coteau is not unreasonable.

Aquatic environments.--Several superglacial or completely ice-walled lakes on the terminal stagnant ice of the Martin River Glacier are clear and relatively warm. Little meltwater gets into these lakes, and they are insulated from the surrounding stagnant ice by as little as 6 feet of superglacial drift. These lakes contain the normal aquatic life of the area, such as mollusks, fish, waterfowl, and aquatic plants. Beaver lodges occur at the edges of some of the lakes.

Other superglacial lakes in areas where the superglacial drift is only a few feet thick are cold and turbid and support little or no life. (See Tuthill, 1963.)

Superglacial topography.--The topography on the stagnant part of the Martin River Glacier is very rugged (fig. A-8). Local relief averages about 150 or 200 feet, although depressions as much as 350 feet deep were observed (Reid and Clayton, 1963). The surface is pock-marked with roughly circular sinkholes similar to those found in karst topography on limestone (Clayton, 1964; Tarr and Martin, 1914). Individual sinkholes are relatively uniform in size, averaging about 700 feet in diameter; their sides have uniform slopes of about 30 degrees. This is slightly larger than the 550 feet postulated for similar features on the Coteau earlier in this report.

Origin of superglacial till.--Much of the drift on the Martin River, Bering, and Malaspina Glaciers is in the form of medial moraines or coarse landslide debris from the valley walls in the upper reaches of the glaciers. However, superglacial drift of this origin forms only a very discontinuous cover, generally less than a foot thick over most of the active parts of these glaciers. In sharp contrast, their stagnant terminal zones, a few miles in width, are covered with a thick continuous blanket of finer superglacial drift at least several feet thick. This material was dragged up from the base of the glacier along thrust planes.

There are two striking evidences for this: (a) Dirty thrust planes dipping up glacier at about 30 degrees can be observed in cross-section in the walls of all of the ice sinkholes on the stagnant part of the Martin River Glacier (fig. A-8a). (b) Linear surface accumulations of drift (shear-moraine ridges) at the exposed edge of thrust planes are abundant at the up-glacial side of the terminal stagnant zones of these glaciers (fig. A-9). Both situations are shown diagrammatically in figure A-10.

Compressive flow.--Thrusting, which drags subglacial drift up into the ice, as a result of compressive flow, and open transverse crevasses are non-existent. The compressive flow in these glaciers is related to the subglacial topography. Seismic exploration has shown that the ground surface beneath the Malaspina glacier rises several hundred feet toward the terminus (Sharp, 1958, p. 624-625, 645), as is shown in figure A-10. This is analogous to the late Wisconsin glacier over-riding the Missouri Escarpment.

Thus, except for the smaller extent of the stagnant ice and the coarser, less plastic superglacial drift, the termini of the Martin River, Bering, and Malaspina Glaciers are very similar to the former stagnant late Wisconsin glacier that covered the Missouri Coteau.

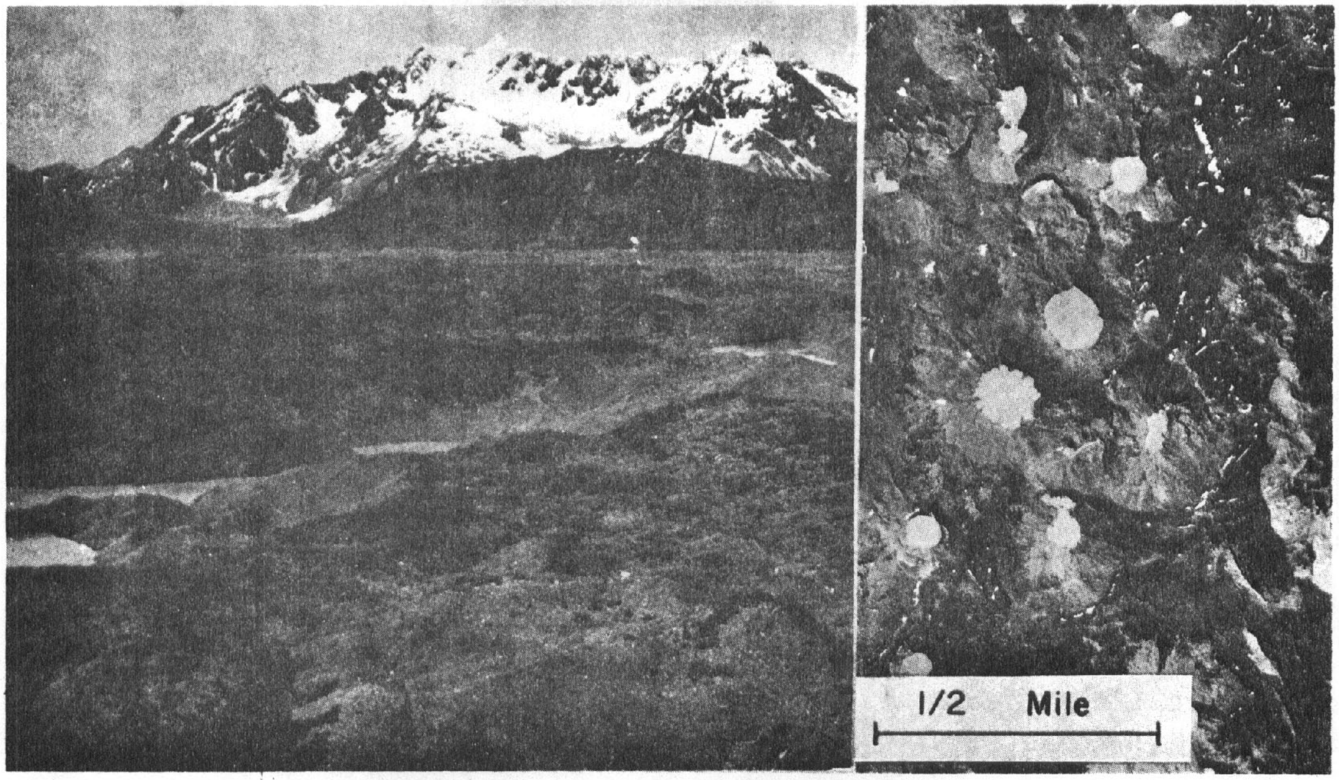


Figure A-8. Stagnant terminus of the Martin River Glacier. a: Entire view to base of mountains, 5 miles away, is stagnant ice covered with a few feet of drift; spruce and alder are growing on the more stable superglacial drift (photo by J. R. Reid). b: Air photo of same area (U.S. Forest Service EEV-15-18, 12-6-59).

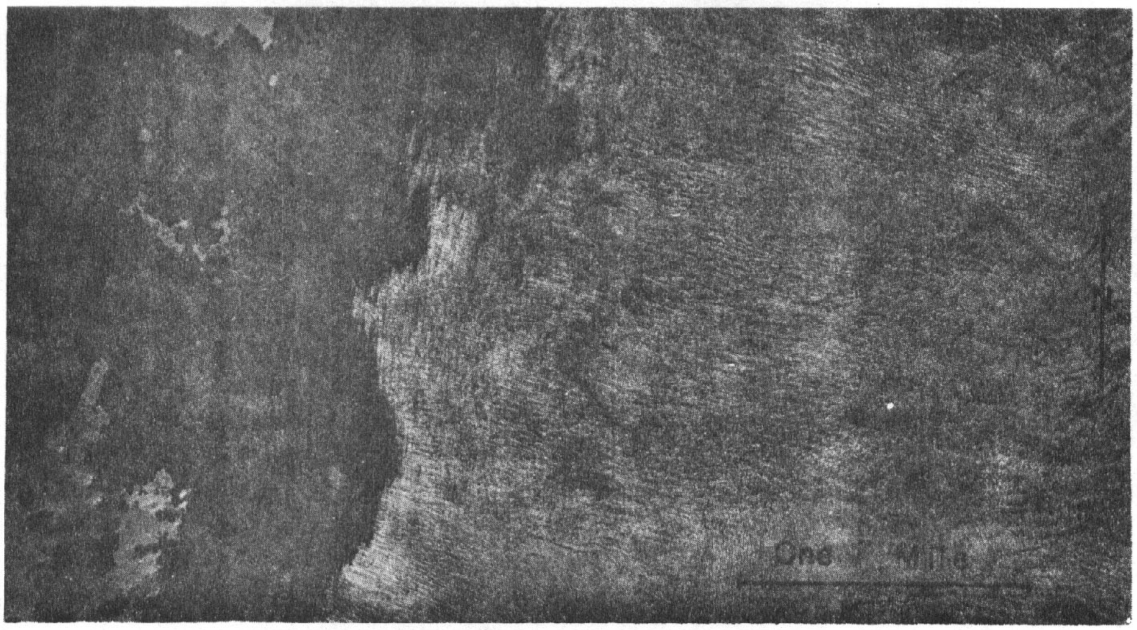


Figure A-9. Air photo of west edge of Bering Glacier. Clean, active ice on right; drift-covered stagnant ice with lakes in depressions on left; in between is an area of north-south shear-moraine ridges where drift has come to the surface along thrust planes. (U.S. Geological Survey 334vv 91RTS M 374 91 SRW 8 AUG 50 50M-6).

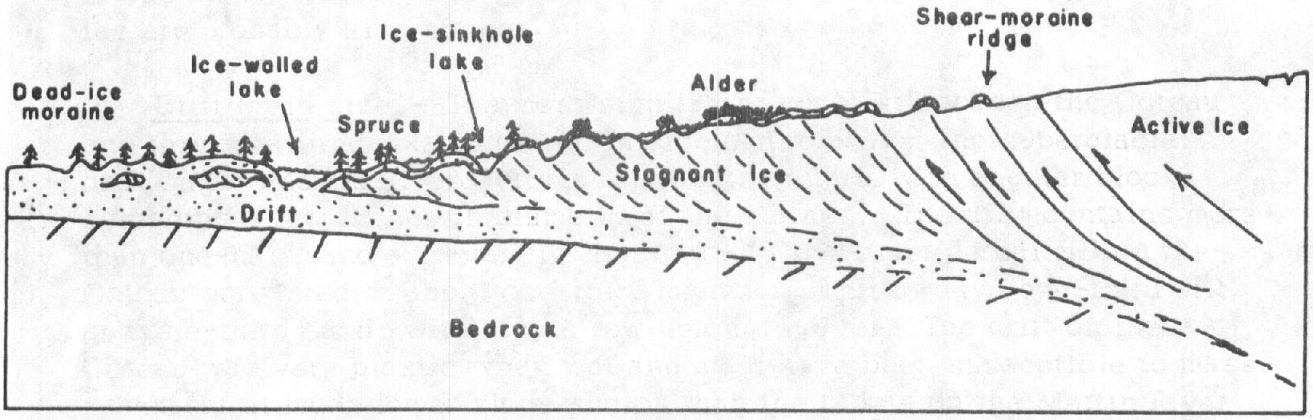


Figure A-10. Diagrammatic cross-section through terminus of Bering, Malaspina, or Martin River Glaciers.

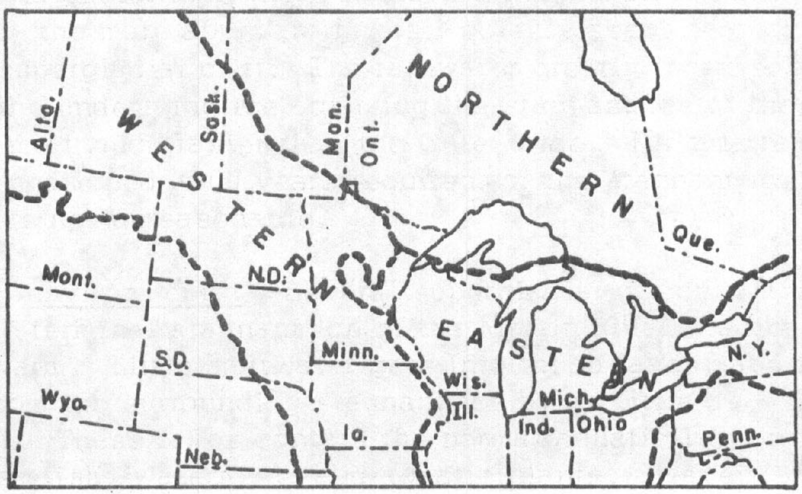


Figure A-11. Glacial-geology "provinces" of middle North America.

## PLEISTOCENE STAGNATION IN NORTH AMERICA

Stagnation features like those on the Missouri Coteau are abundant on most elevated glaciated parts of the Dakotas, northern Montana, western Minnesota, Alberta, southern Saskatchewan, and southwestern Manitoba.

Pleistocene glacial stagnation and collapsed superglacial river sediments were widespread in New England (Flint, 1932, and Hartshorn, 1952), Ontario (Deane, 1950), Michigan, Wisconsin, and eastern Minnesota (Flint and others, 1959). However, dead-ice moraine, disintegration ridges, collapsed lake sediment, and ice-walled-lake plains apparently are less common in these areas.

These apparent differences between "eastern glacial geology" and "western glacial geology" (fig. A-11) are in part the result of climatic differences, differences in ice-flow velocities, differences in postglacial erosion, and differences in loess cover. But perhaps there is a more basic cause for the differences between the eastern and western areas.

In the eastern area, drumlins tend to be of clastic spoon-shaped type, with a length-width ratio of about 3. In the western area, however, they tend to be of the long linear type, with a length-width ratio of perhaps 50 or more. Apparently, the western drift tended to be more easily molded into streamlined, subglacial forms. The cause of this may in part be related here to the differences in composition of the eastern and western tills. The eastern till was derived, in large part, from hard lower Paleozoic limestone and sandstone; the till tends to be sandy and not very plastic. In contrast, the western till was derived largely from soft Cretaceous and lower Tertiary montmorillonitic shale; the till is characteristically very clayey and highly plastic. Therefore, the western tills were more easily molded into highly streamlined drumlins. Once on top of the glacier, till in the western area tended to flow more easily when moist, forming the disintegration ridges that are so characteristic of the western area.

Perhaps, also, superglacial till was much more common in the western than in the eastern areas of continental glaciation. The glaciers in the west were more commonly moving up regional slopes, resulting in more widespread compressive flow and more intensive marginal thrusting.

## ACKNOWLEDGMENTS

I wish to thank those who have contributed to this paper. Included are Theodore F. Freers, Walter L. Moore, S. J. Tuthill, Wayne A. Pettyjohn, J. R. Reid, and W. O. Kupsch, who gave me numerous ideas and attempted to make the paper readable.

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30-B

# PLEISTOCENE SUPERGLACIAL AND ICE-WALLED LAKES OF WEST-CENTRAL NORTH AMERICA

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In areas where the Pleistocene glacial terminus retreated "normally" the characteristic type of glacial lake was the proglacial lake, an example of which was Lake Agassiz. However, in parts of the Upper Midwest and the Prairie Provinces (the "western" area of figure A-11 of the previous paper) where glacial stagnation occurred, the characteristic forms were superglacial or ice-walled lakes.

## SUPERGLACIAL LAKES

Distribution.--Sediments that were deposited in superglacial lakes cover large areas in Saskatchewan and Alberta (Bayrock and Hughes, 1962; Christiansen, 1965; Greer and Christiansen, 1963; Meneley, 1964; Parizek, 1964; Westgate, 1965). Individual superglacial lakes in North Dakota were smaller; one of the largest, near Lehr, is described at stop 5 in the roadlog.

Sediment.--The sediment of superglacial lakes is very fine sand, silt, and clay. It is commonly tens of feet thick. Laminations are common in some areas, but are lacking in others. The bedding is commonly folded or contorted as a result of collapsing due to the melting of underlying stagnant ice (fig. B-2).

Topography.--The topography of collapsed superglacial-lake sediment is nearly identical to that of collapsed superglacial till (dead-ice moraine), though the local relief and slope angles tend to be somewhat lower (fig. B-1). Where the superglacial-lake sediment was thin the relief is low.

Circular disintegration ridges, similar to those in medium-relief dead-ice moraine, are common. Straight disintegration ridges occur in lake sediment as much as 100 feet thick (Cherry, 1966).

Interpretation.--Field evidence that these sediments were deposited on top of stagnant glacial ice includes the following: (a) rolling undissected topography with abundant undrained depressions and other landforms that are similar to those of collapsed superglacial till (dead-ice moraine); (b) contorted or folded bedding (fig. B-2 and fig. R-14 of the roadlog); (c) an absence of active-ice features; and (d) a lake-sediment thickness great enough to preclude the possibility that the topography is a reflection of underlying dead-ice moraine.

Large superglacial lakes were possible on stagnant ice only where the ice was sufficiently laden with drift so that it could not float. Assuming the debris had a specific gravity of 2.7, a content of only 4.5 percent in or on the ice was necessary to prevent the ice from floating.

## ICE-WALLED LAKES

Ice-walled lakes are distinguished from superglacial lakes by having been bottomed on solid ground, not on ice. They were surrounded by stagnant ice. Two end members, with coalesced or uncoalsced basins, are distinguished. Individual uncoalesced basins were separated from each other by ice. If some of the separating ice melted, the basins coalesced.

### Coalesced basins

Ice-walled lakes with coalesced basins were rather uncommon. Sediment deposited in an outstanding example, in the Yorkton, Saskatchewan, area, is shown in figure B-3 and B-4. The sediment is massive or indistinctly-bedded silt and very fine to fine sand.

Evidence that the lake sediment of figure B-3 and B-4 was deposited around ice-blocks--that is, in coalesced basins--is as follows: (a) The high-relief topography consists largely of lake sediment resting on a relatively flat surface of till. (b) The areas that were presumably between the ice blocks are relatively flat plateaus underlain by uncollapsed lake sediment. (c) The plateaus rise no higher than 1675 and 1650 feet; they do not exceed the maximum water level of 1700 feet, determined by the elevation of the outlet of associated Lake Saltcoats (Christiansen, 1960, p. 16). (d) The areas which were presumably occupied by the ice blocks contain no hummocky collapse topography or superglacial lake sediment, indicating that the ice blocks extended above water level; the thin cover of lake sediment shown in figure B-4 in these areas is probably postglacial. (e) The plateaus grade into areas of hummocky collapse topography at several locations in the Yorkton area. Presumably the ice blocks extended below the lake in places, resulting in superglacial deposition, as shown at B in figure B-3.

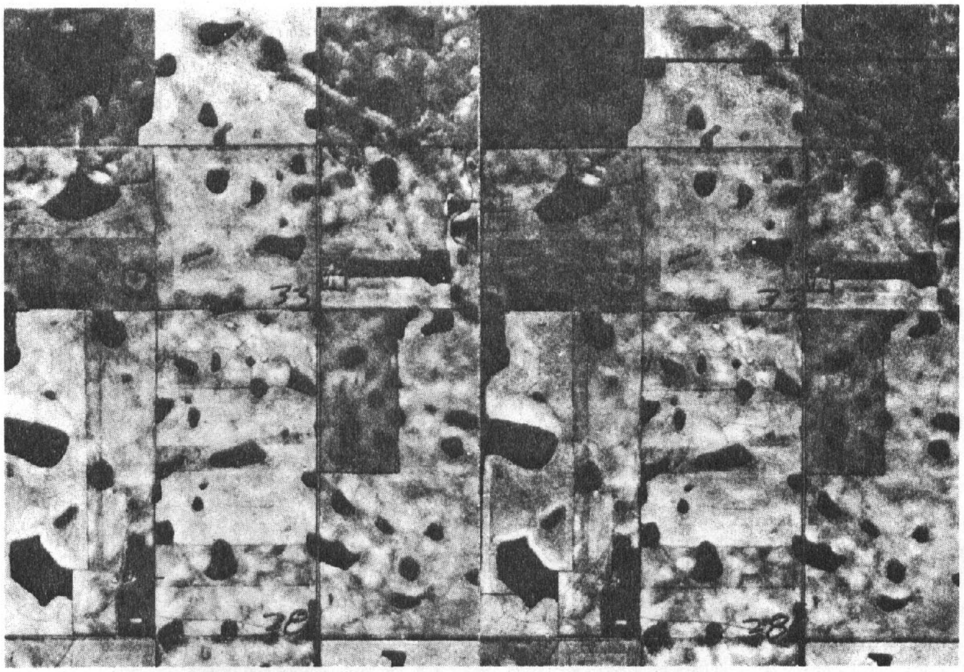

Figure B-1. Stereopair of high-relief collapsed superglacial lake sediment; 100 feet of lake sediment was penetrated in a near-by test hole. In T. 27, R. 17, W. 3, Kindersley area, Sask. (R.C.A.F. A15509-40 and 41.)

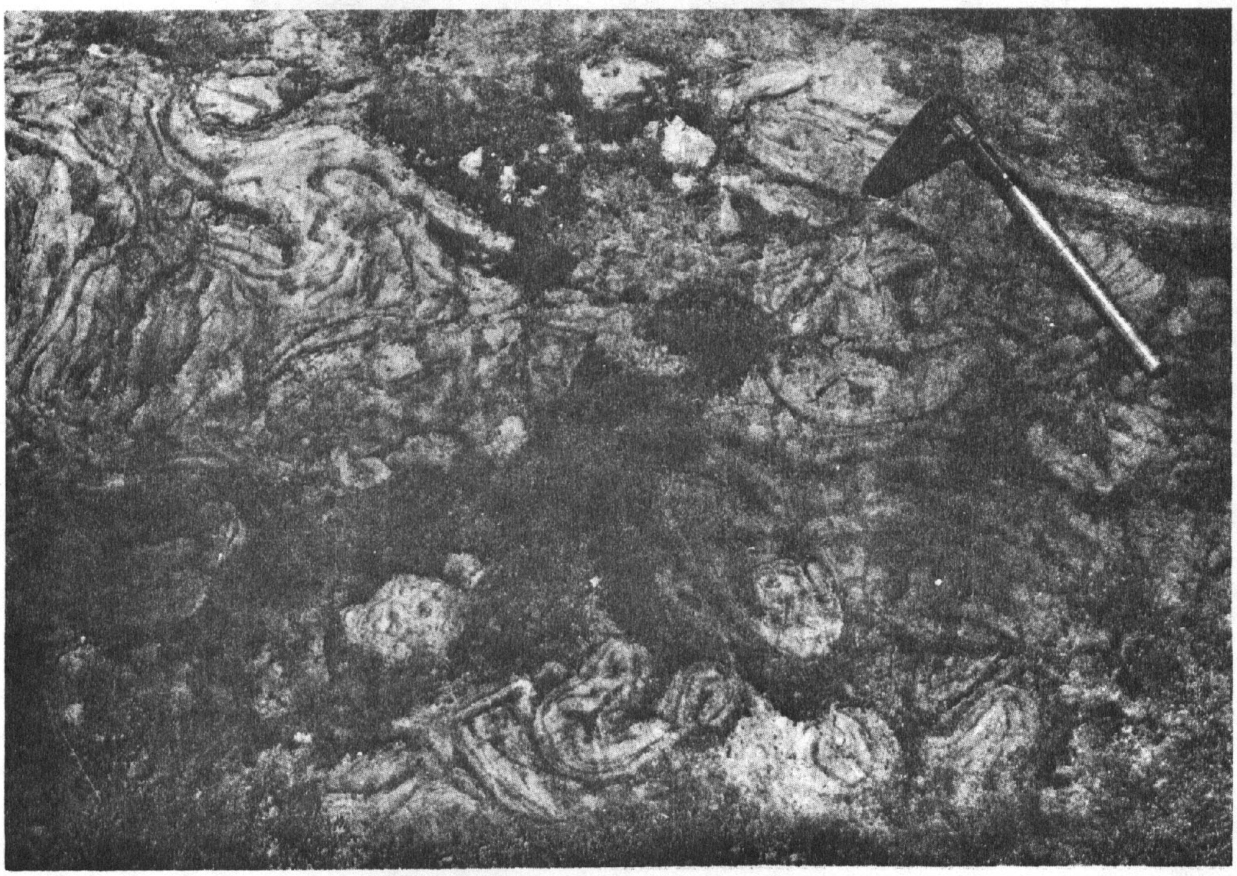

Figure B-2. Contorted fragments of bedded lake silt within a matrix of darker nonbedded lake clay; clay flowed from an unstable superglacial environment, carrying the silt with it. Roadcut 0.2 mile north of southeast corner of sec. 11, T. 157 N., R. 90 W., Mountrail County, N. Dak.

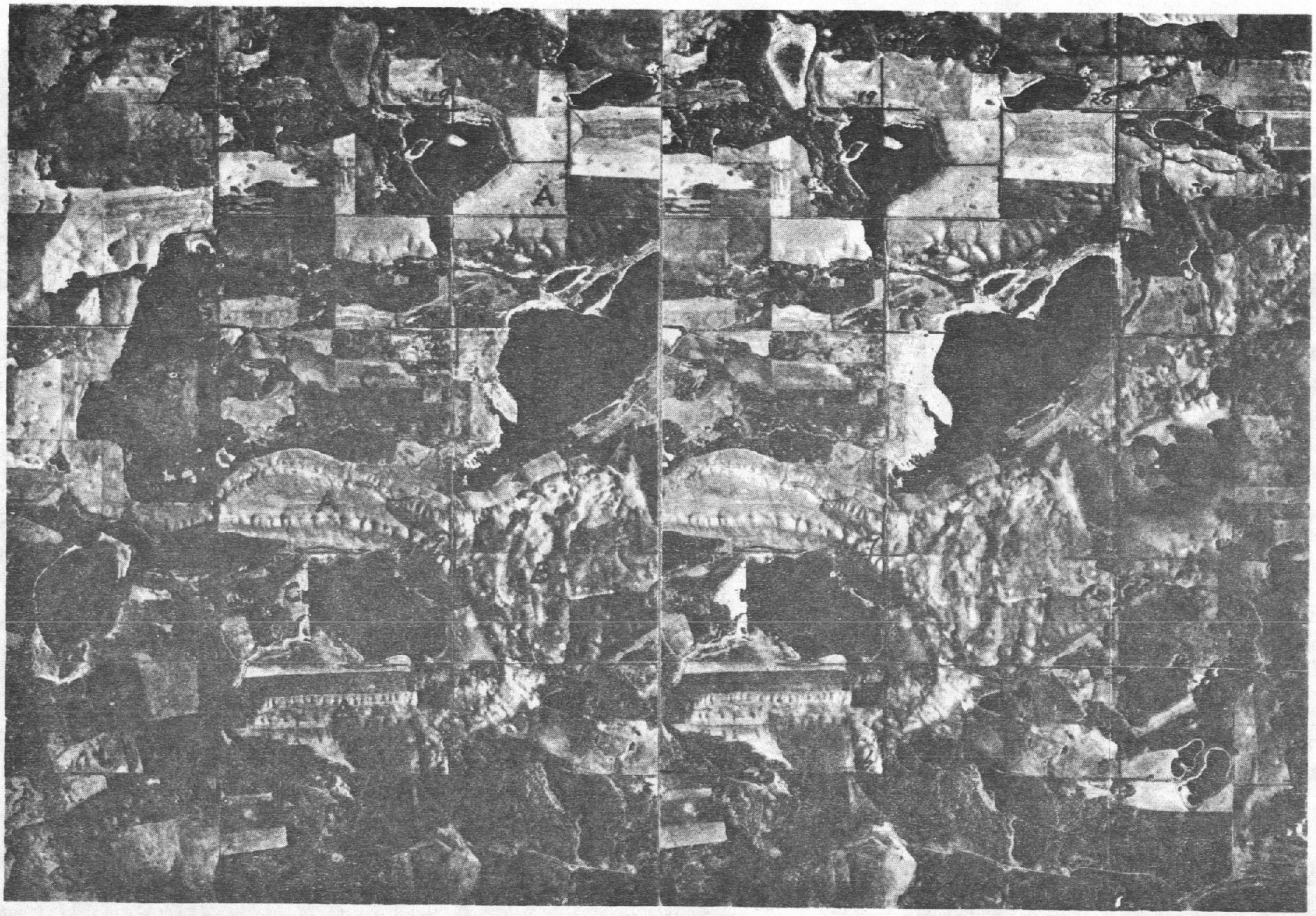


Figure B-3. Stereopair of ice-walled-lake plains formed in coalesced basins (A), collapsed superglacial lake sediment (B), and ice-block depressions (C); cross-section in figure B-4 is along the east edge of sections 6 and 7. In T. 33, R. 1, W. 2, Yorkton area, Sask. (R.C.A.F. A15960-36 and 37.)

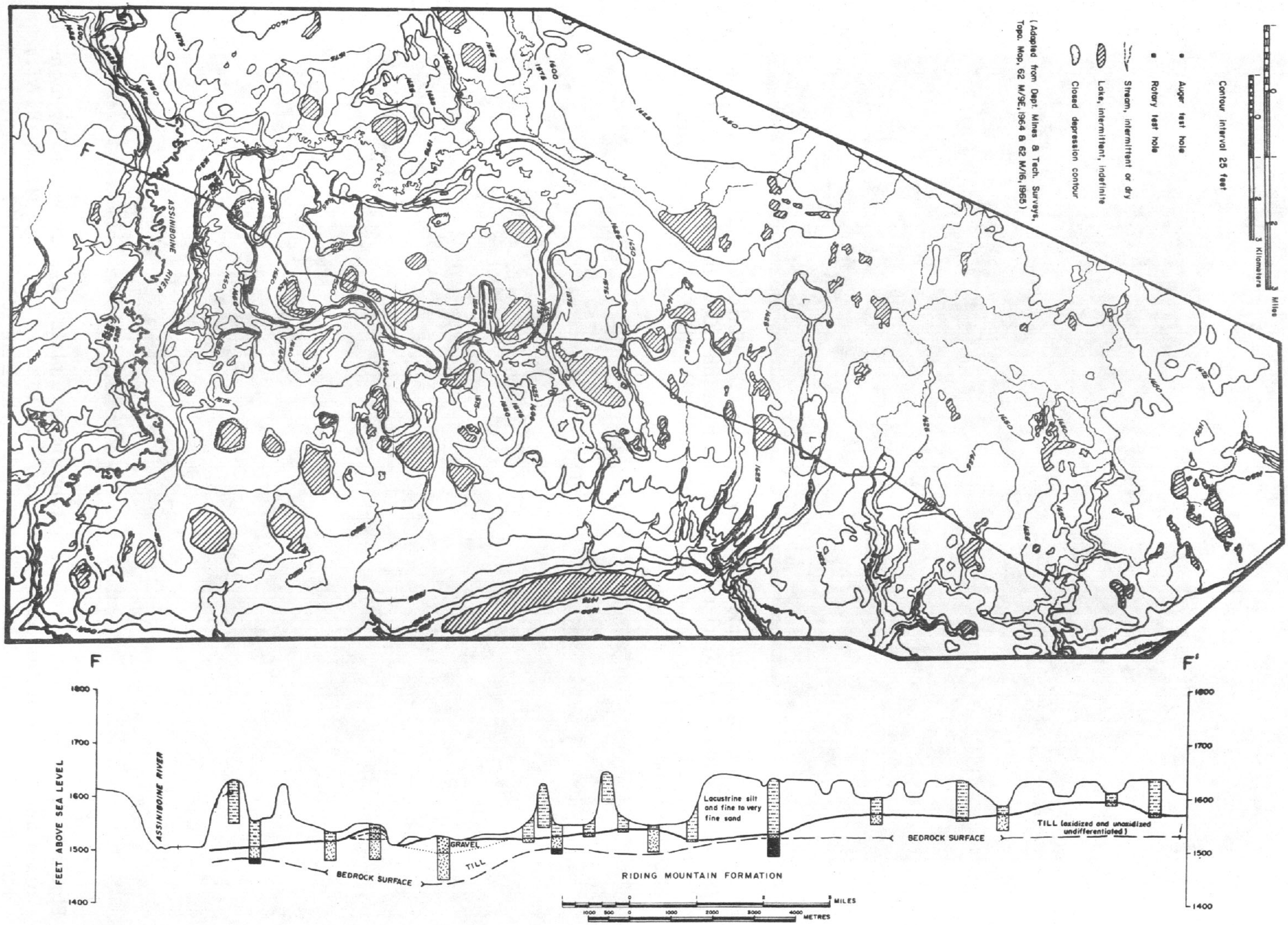


Figure B-4. Ice-walled-lake plains formed in coalesced basins. a: topographic map. b: cross-section. In T. 32, 33, and 34, R. 1, W. 2, Yorkton area, Sask. Fig. B-3 is in center of this area.

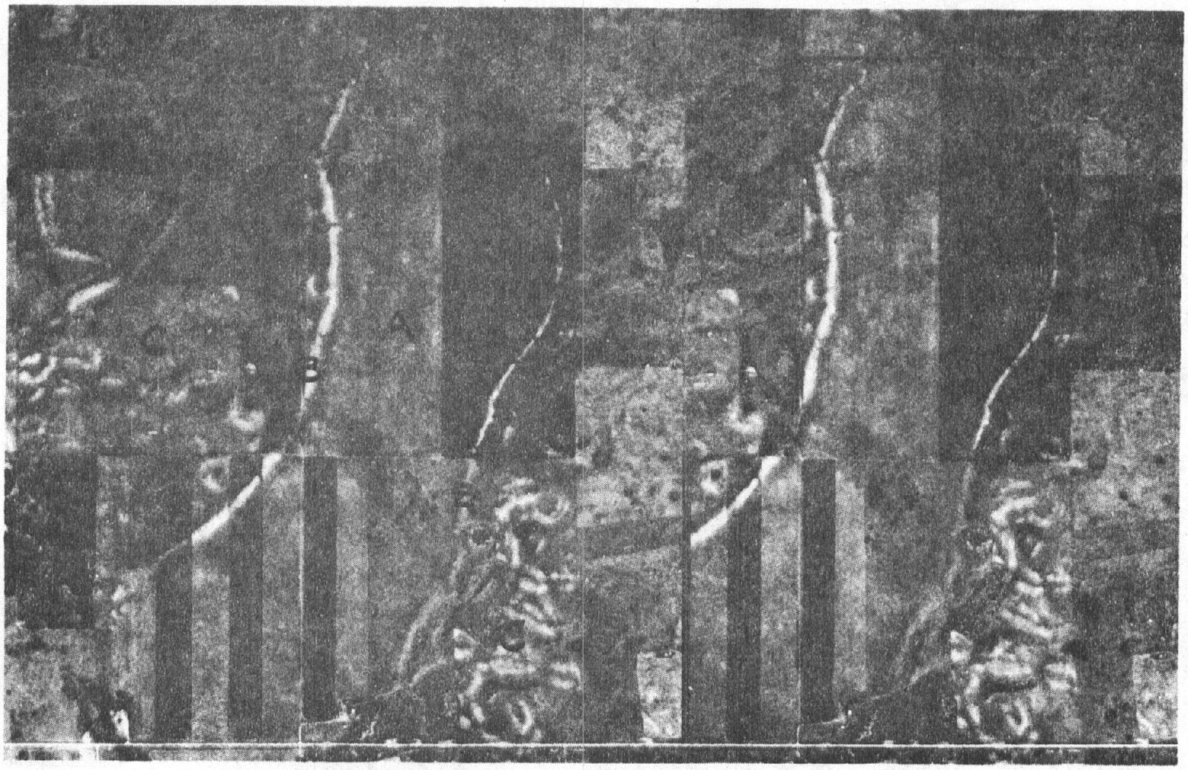

Figure B-5. Stereopair of unstable-environment lake plain (A) with till rim ridge (B); surrounded by medium-relief dead-ice moraine (C). In sec. 13 and 14, T. 154 N., R. 91 W., Mountrail County, N. Dak. (U.S. Dept. Agriculture BAL-4V-9 and 10.)

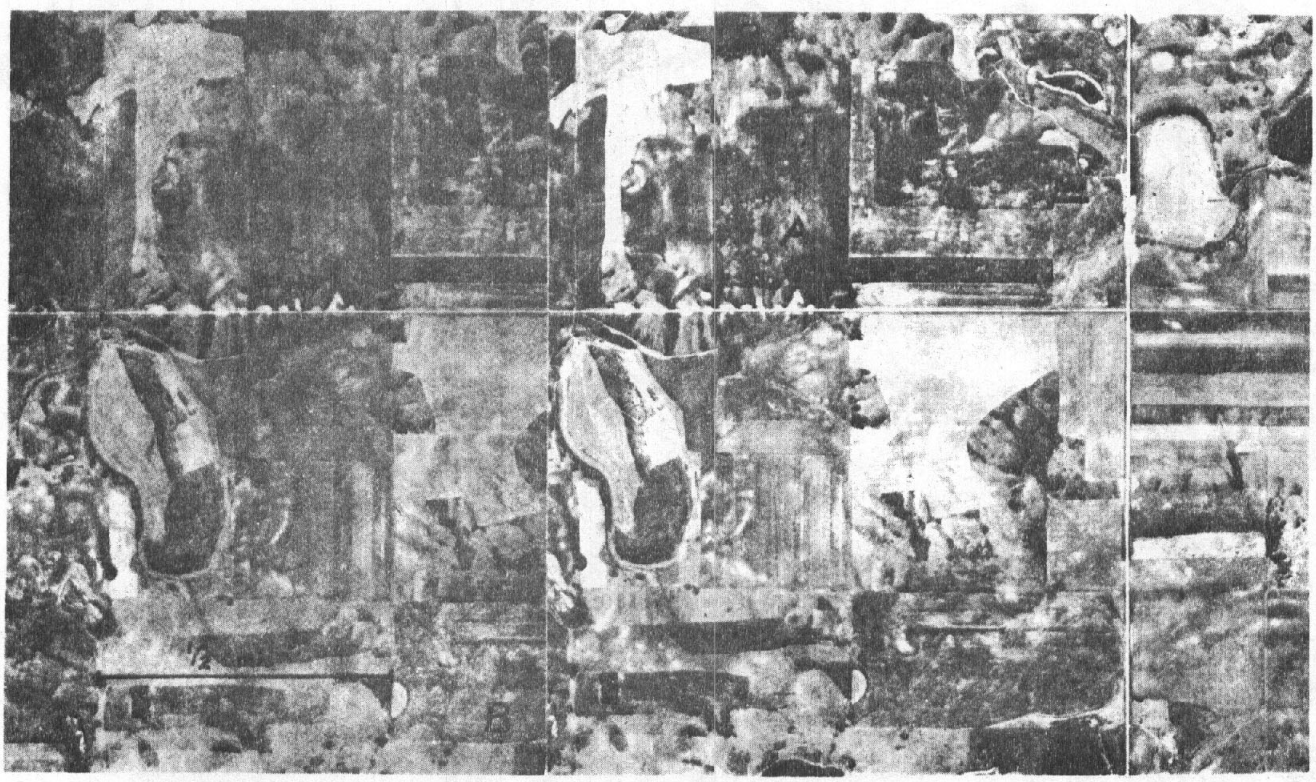

Figure B-6. Stereopair of stable-environment lake plain (A) surrounded by high-relief dead-ice moraine (B). In secs. 2, 3, 10 and 11, T. 157 N., R. 88 W., Mountrail County, N. Dak. (U.S. Dept. Agriculture BAL-6V-100 and 101.)

Uncoalesced basins

In contrast to the previous type, most lakes occupied single, separate ice basins. These lakes, which were commonly 1 or 2 miles wide, were widespread on stagnant ice in North Dakota, Saskatchewan, and Alberta in late Pleistocene time (Bayrock and Hughes, 1962; Bayrock and Jones, 1963; Christiansen, 1956 and 1961; Clayton, 1962; Ellwood, 1961; Gravenor and Ellwood, 1957; Gravenor and Kupsch, 1959; Parizek, 1961; and Winters, 1960, 1963). The landforms resulting from sediment deposition in these lakes have been called "ice-walled lake plains," "perched lacustrine plains," "perched lake plains," "elevated lake plains," "dead-ice plateaus," "moraine plateaus, " "moraine-lake plateaus," and "impounded glacial lake plains." Similar features, composed of outwash sand, have been described in eastern Minnesota (Farnham, 1956, p. 53-64; Schneider, 1961, p. 59-60). The characteristics and mode of origin of similar features in Denmark have been illustrated by Schou (1949).

Two end members of this lake type can be recognized: (1) lakes formed in a stable environment with thick superglacial till on the surrounding stagnant glacier and (2) lakes formed in an unstable environment with thin superglacial till on the surrounding ice. Detailed analysis of the sediments deposited in these lakes are given in paper 30-C.

<u>Unstable-environment lakes.</u>--Typical unstable-environment lakes occurred in T. 155 and 156 N., R. 92 and 93 W., near Ross in western Mountrail County, North Dakota (fig. B-5, B-7, and B-9).

The superglacial till in the "unstable" environment was thin (a few tens of feet or less), and ice melted rapidly. (Slopes in the resulting dead-ice moraine are gentle, 2 to 7 degrees, and the relief is low.) As a result, the topography on the glacier underwent continuous change, and mass movement of the superglacial till was very common. Much superglacial till slumped or flowed into the lakes, and cold silt-laden superglacial rivers carried much sand and gravel into the lake margins. Waves sorted and redistributed the mudflow and slump material, carrying the fines out into the lake, leaving the sand and gravel fraction on the beach along with the sand and gravel brought there by the rivers. Meltwater frequently moved through the lakes with a velocity sufficient to prevent the deposition of the finest suspended particles. Therefore, the lake sediments are typically much coarser than the stable-lake sediments. In the Ross area, the sediment is commonly clayey silt in the center of the lake plains, and gradually coarsens to sandy gravel at the margins.

Lake plains formed in the unstable environment, such as those near Ross, are concave upward. This is a result of the great amount of coarser material deposited near shore and the greater compaction of the finer mid-lake deposits.

Unstable-environment lake plains commonly have either till or gravel rims--till where the superglacial till slumped off the ice into the lake margin without being reworked by the waves, and gravel where it was reworked by the waves.

The plains of lakes in the unstable environment tend to be round in outline because of the greater shore activity than in the lakes in the stable superglacial environment; shore activity tended to remove any irregularities projecting into the lakes.

Stable-environment lakes.--Typical stable-environment lakes occurred in T. 157 N., R. 88 W. (fig. B-6, B-8, and B-9) and the southeastern part of T. 156 N., R. 88 W., near Tagus in eastern Mountrail County, North Dakota.

In the stable superglacial environment, the superglacial till was so thick (several tens of feet) that the ice melted very slowly. (Slopes in the resulting dead-ice moraine are presently steep, 7 to 20 degrees, and local relief is high; see section on slope angles in previous paper.) As a result, the topography of the stagnant glacier changed very slowly, and mass movement of the superglacial till was infrequent. Superglacial rivers were relatively tranquil and carried little silt; only the material that settled out very slowly reached the lakes. Therefore, the sediments that accumulated in these lakes are very fine grained (clay and silty clay), such as in the lake plains near Tagus.

Lake plains formed in the stable environment, such as those near Tagus, are commonly convex upward because there was relatively uniform deposition of clay over the entire lake bottom and the margins rested on ice; the margins collapsed when the ice melted.

Stable-environment lake plains do not have rims because there was little slumping of material off the stable ice walls.

Sediments deposited in the stable-environment lakes are commonly much thicker than those deposited in the unstable-environment lakes. Lake plains of the stable environment are perched above the surrounding dead-ice moraine, whereas the plains of the unstable environment are commonly in low areas, but are above the bottoms of depressions immediately adjacent to them. Apparently the lakes in a stable environment lasted for a much greater length of time, allowing more lake sediment to accumulate than in the short-lived lakes of the unstable environment.

Although an adequate sampling has not been made, aquatic fossils apparently are more abundant in lake sediments deposited in stable-environment lakes because the water was warmer, more silt-free, and there was more time for the establishment of plant and animal populations than in the unstable environment (see paper 30-E).

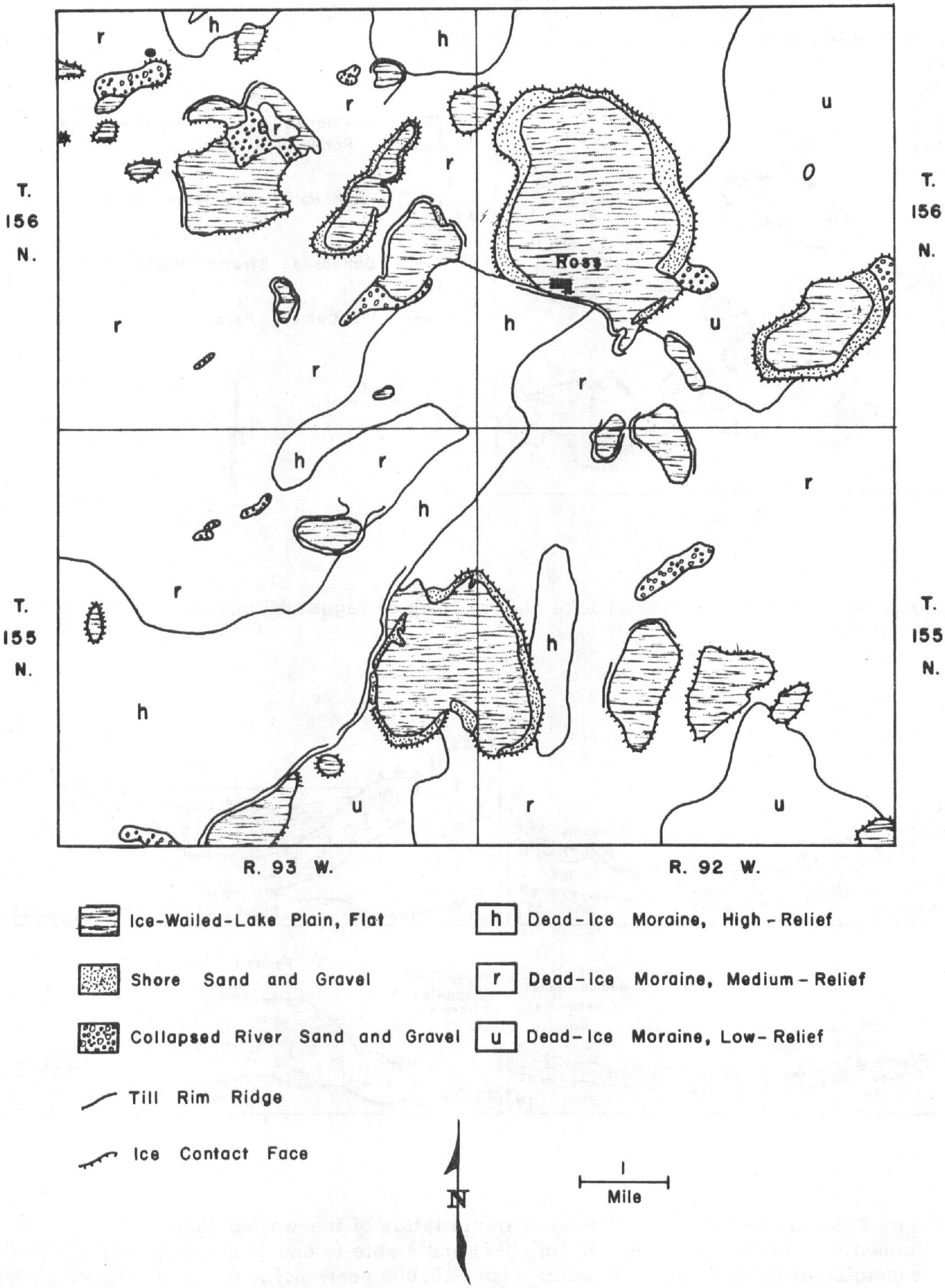


Figure B-7. Unstable-environment lake plains; around Ross, Mountrail County, N. Dak.

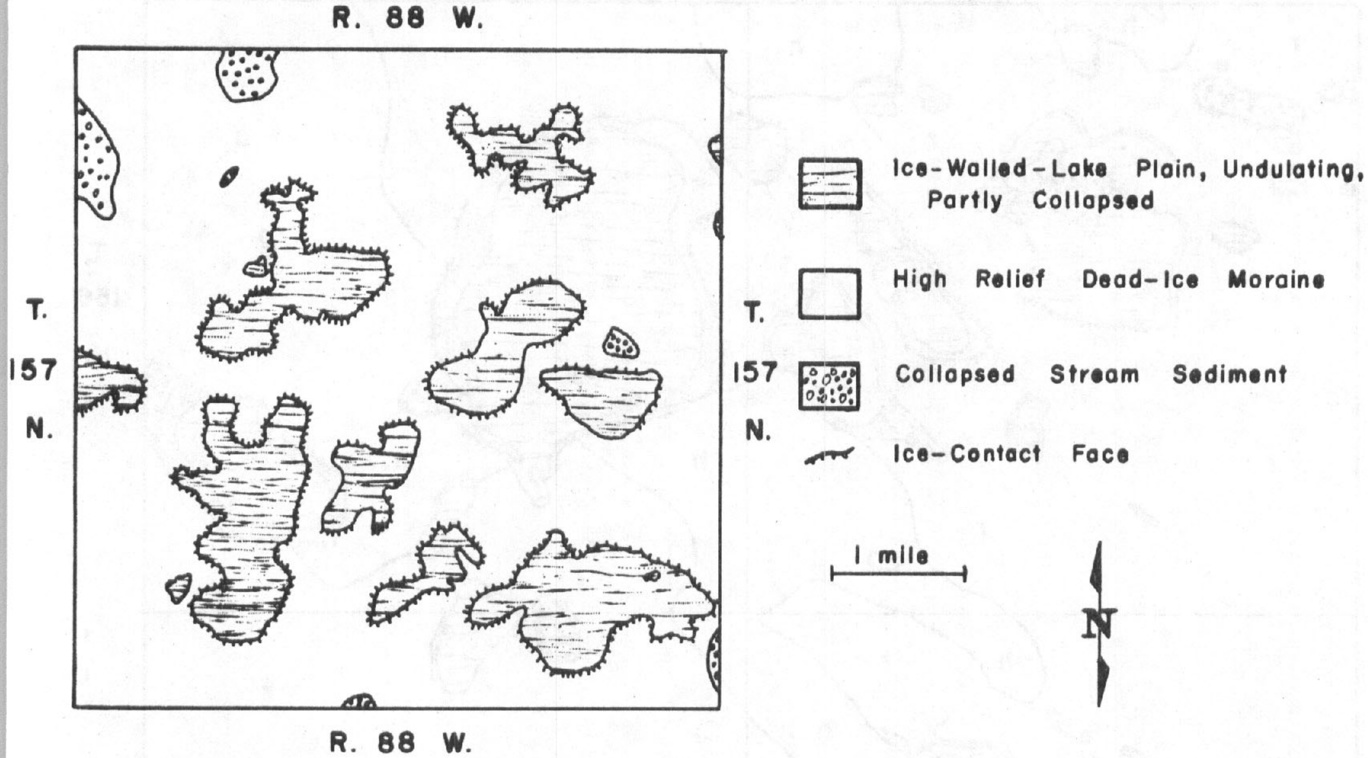


Figure B-8. Stable-environment lake plains; north of Tagus, Mountrail County, N. Dak.

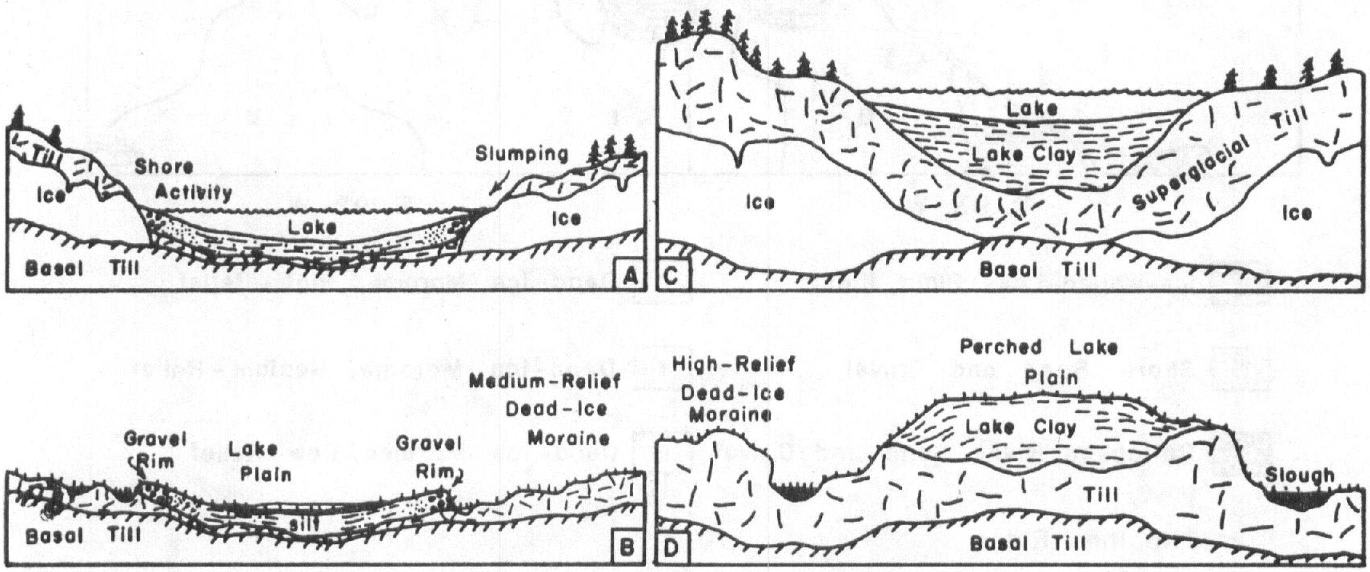


Figure B-9. Cross-sections showing characteristics of ice-walled-lake plains formed in unstable (a and b; see fig. B-7) and stable (c and d; see fig. B-8) superglacial environment. a and c: about 10,000 years B.P.; b and d: today.

## SUMMARY

Superglacial and ice-walled lakes were characteristic of late Wisconsin stagnant glaciers in central North America and are commonly found on present-day stagnant glaciers. The hilly collapsed sediments of superglacial lakes and perched plains of ice-walled lakes are today two of the most characteristic features of areas of former stagnant glaciers.

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30-C

# A PRELIMINARY REPORT ON SOME ICE-WALLED-LAKE DEPOSITS (PLEISTOCENE), MOUNTRAIL COUNTY, NORTH DAKOTA

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## INTRODUCTION

### Scope and objectives

A preliminary investigation of some Pleistocene lake-plain sediments from Mountrail County (figure B-1 in paper 30-B) was initiated in an attempt to interpret their relationship to late Wisconsinan ice-stagnation on the Missouri Coteau in North Dakota. It is believed that a knowledge of the physical and chemical nature of these deposits used in conjunction with paleontological and geomorphic evidence will aid in interpretation of both the lacustrine environment in which they accumulated and the adjacent stagnant-ice terrain. Although this study constitutes only a reconnaissance, it is felt that useful preliminary results have been obtained and that they point the way toward productive future work.

### Location and description

The general character and distribution of lake-plain deposits is given by Clayton and Cherry (paper 30-B). The size and shape of lake plains and sample locations discussed in this report are given in figure C-1. These localities are considered representative of the various ice-walled-lake plains present within the county. Lake plains A, B, and C are of the "low" ("unstable environment") type; and lake plain D is of the "elevated" ("stable environment") type as discussed by Clayton and Cherry (paper 30-B).

## ANALYTICAL METHODS AND PRESENTATION OF RESULTS

### Sampling

Samples were obtained from a single station in each of the lake plains using a 0.75 by 10-inch soil-sampler in conjunction with a 1.5-inch soil auger; samples from lake plain D were obtained by a 4-inch truck-mounted auger. A continuous core (in 10-inch sections) was recovered to depths commensurate with the capabilities of the soil-sampler, and spot samples were obtained from greater depths after deepening the hole with the soil auger. Lake plain A was penetrated to its base, and lake plain B nearly to its base. About 8 feet of core was recovered from lake plain C, which probably represents about half the thickness of the deposit at the sample locality. Although lake plain D was augered to a depth of 55 feet, its base was not reached. We estimate that the total thickness does not exceed 75 or 80 feet.

### Size analysis

Sieve and pipette analyses were made for subsamples from each station according to the general procedures of Krumbein and Pettijohn (1938). Measurements were made at single phi-unit intervals (Wentworth classes) and the results processed by a modified version of the University of Missouri Fortran program for evaluation of size data (Kane and Hubert, 1963). Due to the fine-grained character of many samples, measures of sorting, skewness, and other size parameters could not be obtained and their evaluation is deferred at this time. Replicate analyses indicate that the reproducibility of cumulative data is slightly better than $\pm$ 0.5 phi-unit at the 50th percentile.

The sand, silt, clay relationships shown on the three-point diagram of figure C-2 (Shepard, 1954) are dominantly clayey silt. Five samples from lake plain A and one from lake plain B fall within the sand-silt-clay class and a single sample from lake plain C is composed of silty sand. Inspection of size data shows that these coarser-grained sediments are all from the basal portions of their respective lake plains (figure C-3). Samples from lake plain D tend to contain greater amounts of silt. In general, there is little difference in the size components of samples from the four lake plains, but a definite trend toward increased particle size with depth is evident in each of them.

The coarse fractions (larger than 62 microns) of each sample was examined microscopically for authigenic constituents such as shell fragments, microfossils, pyrite, etc. Coarse biogenous material consists of plant fibers, shell fragments, broken chitinous insect carapaces, ostracods, radiolaria, and foraminiferids. The radiolaria bear great resemblance to specimens assigned to <u>Cyrtocalpis sp</u>. (Wilson, 1958) and the foraminiferids are the so-called "<u>Globigerina cretacea</u>"; both are derived from late Cretaceous Pierre and Niobrara Shales.

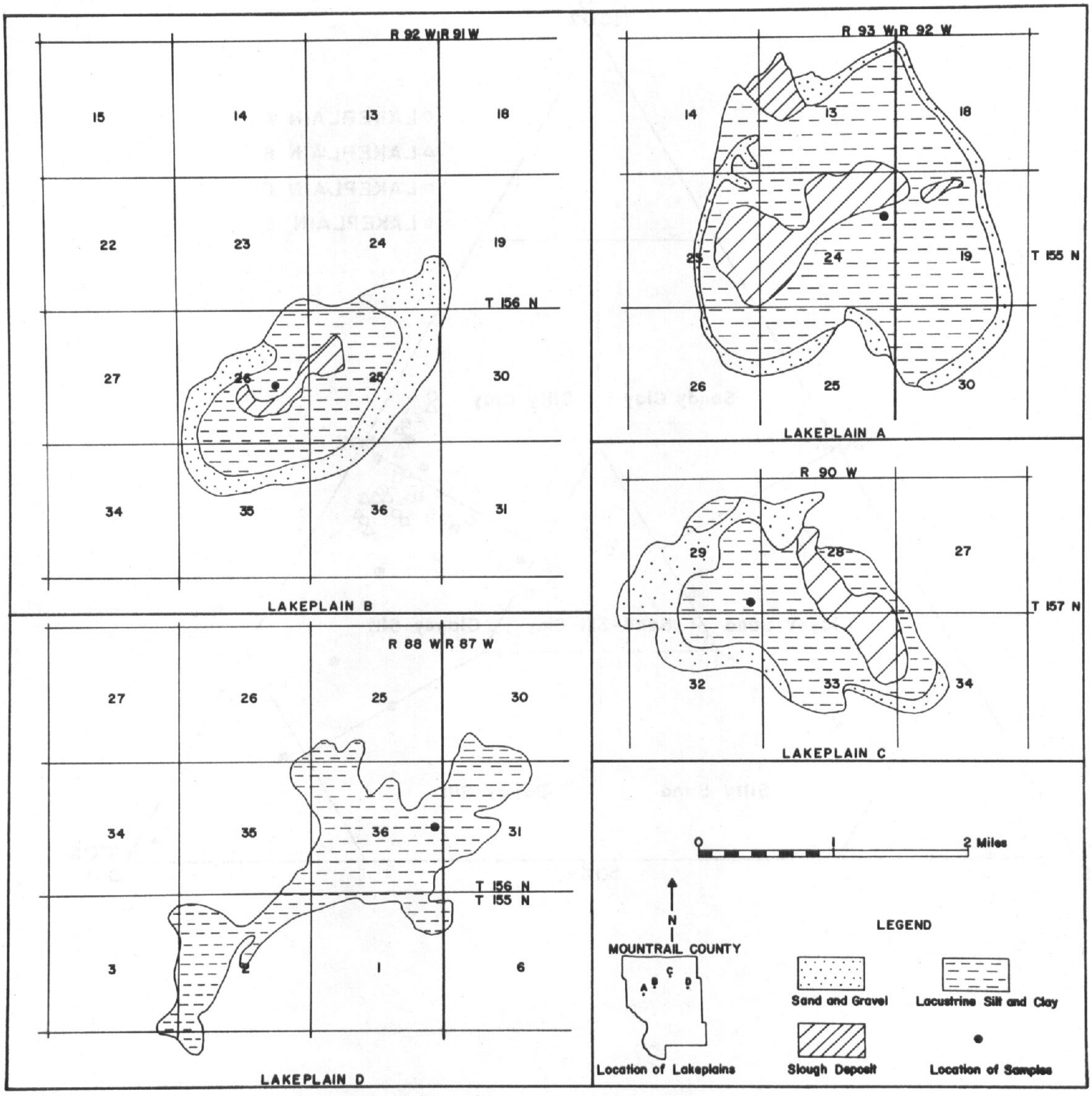


Figure C-1. Generalized geologic map of lake plains showing sample location.

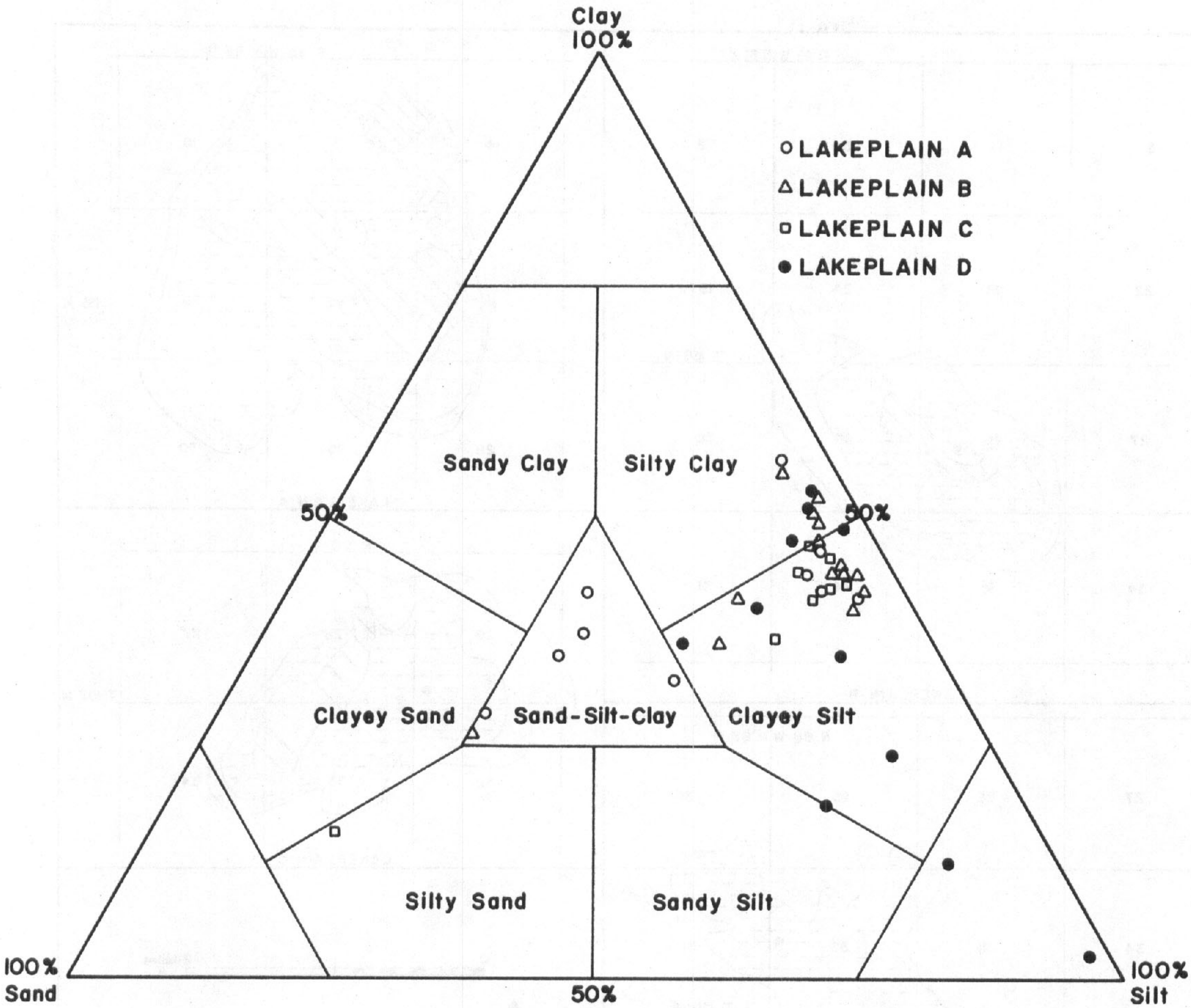


Figure C-2. Sand-silt-clay relationships in lake plain sediments.

Ostracods are absent in samples from lake plains A and D but occur in sufficient abundance in B and C to indicate that they could not have been reworked from older sediments. In the core from lake plain B, ostracods are common in the upper samples and diminish in relative abundance with depth, none being detected below a depth of about 10 feet. Likewise, samples from lake plain C above a depth of about 6.5 feet contain ostracods, while those of greater depth contain none.

### Carbonate analysis

The total reactive carbonate content of all samples was determined by the rapid titrametric method outlined by Herrin and others (1958), in which one gram of ground sample is reacted with 50 ml of standard $H_2SO_4$ for 10 minutes at 90°C and back-titrated with standard NaOH. All results were computed as calcium carbonate, although X-ray diffraction analysis indicates that a significant portion of some samples is dolomite. Results range between approximately 13 and 20 percent (by weight) calcium carbonate and average about 17 percent. Results are reproducible within $\pm$ 1 percent of reported values.

The carbonate component of sediments from the four lake plains is relatively constant. Lake plains A and D have the greatest average $CaCO_3$ content, each 17.8 percent, and C and B have 17.5 and 16.2 percent respectively. A plot of the median grain size of each sample against its carbonate content demonstrated that no correlation exists between sediment size and carbonate content. Results are plotted as a function of depth in figure C-3.

### Total-carbon and organic-carbon analyses

The total reactive carbon content was determined by the microdiffusion method of Maciolek (1962), in which the $CO_2$ evolved from dichromate oxidation of 100 to 200 milligrams of ground sample is absorbed in standard NaOH and back-titrated with standard HCL. Results are reproducible within $\pm$ 1 percent of reported values and are plotted in figure C-3.

Total-carbon values range from 1.84 to 3.04 percent. Samples from lake plains A and D yield the highest averages, 2.51 and 2.57 percent respectively, while B and C are only slightly lower with averages of 2.47 and 2.45 percent.

Organic-carbon content is estimated as the difference between total-carbon and carbonate-carbon. Lake plains A and B have the highest average organic-carbon content, 0.53 and 0.56 percent, and C and D are slightly lower, 0.37 and 0.42 percent respectively. A secondary trend of increased organic-carbon with increase in grain size is also evident.

### Total-nitrogen analysis

Total reactive nitrogen was determined by marco-Kjeldahl distillation (Jackson, 1958) of the solution remaining after total-carbon analysis (Maciolek, 1962).

Nitrogen values (figure C-3) are small and relatively constant, average values for lake plains A, B, C, and D are 0.029, 0.035, 0.041, and 0.033 percent respectively; values range between 0.02 and 0.07 percent. Results are reproducible within ± 3 percent of reported values and are plotted as a function of depth in figure C-3.

### Carbon-nitrogen ratios

The ratio of organic-carbon to total nitrogen was computed for all samples. Quotients range widely from 4.6 to 32.1. Lake plain A has the greatest single C/N value and also the largest average ratio, 18.0. Lake plains B and D have somewhat lower averages, 16.1 and 12.8 respectively, and lake plain C has the lowest average, 9.0. Results are plotted in figure C-3.

Although some reversals occur, a trend of increased C/N values with depth appears to be established, particularly in lake plains A and B. Lake plain C was not sampled to sufficient depth to validate any vertical trends. The increase in C/N values with depth is seen to be the result of increased organic-carbon values (rather than a decrease in nitrogen).

### Calcium-magnesium ratios

Calcium and calcium plus magnesium was measured by EDTA complexi-metric titration (Shapiro and Brannock, 1962) in samples from lake plain D. Ca/Mg ratios range from 2.23 to 3.21. There appears to be a significant difference between the values from the upper and the lower portions of the lakeplain. The average Ca/Mg ratio for the upper six samples is 2.87 and for the lower five samples is 2.37. Results are plotted as a function of depth in figure C-3.

### Clay mineralogy

Sedimented slides for X-ray diffraction study of the clay-sized fraction of all samples were prepared during pipette analysis. These were scanned through an angle of $30^{\circ}$ ($2\theta$) using standard Norelco equipment with a copper X-ray tube, nickel filter, and Bristol recorder at an angular speed of $1^{\circ}$ ($2\theta$) per minute. Successive analyses were made on air-dry, glycolated, acid-treated, and heat-treated slides in an effort to gain semi-quantitative data on clay-mineral composition. Details of procedure are given by Brown (1961).

Identification of clay minerals was based on the orderly repetition of the X-ray diffraction peaks from the basal (001) crystallographic lattice. The clay-mineral suite consists of mica species, referred to as illite, characterized by 5 and 10 A diffraction peaks; kaolinite; and montmorillonids. The presence of mixed-layer clays was not detected with certainty.

Estimates of the relative abundances of illite, kaolinite, and montmorillonite were obtained by comparing the intensities of major diffraction peaks. The ratio of 7.1A/10A peak areas was considered a measure of kaolinite/illite and the 17A/10A peak areas a measure of montmorillonite/illite. The value of the 7.1A peak area was divided by two (Weaver, 1958) and the 17A peak area by four (Bradley, 1953) to account for differences in crystallinity and geometry.

It was found that the high montmorillonite content, which has poor crystallinity and diffracting ability, masked the other clay components, making detailed comparison of peak areas somewhat impractical. Estimates, however, reveal that nearly all samples have 17A/10A values between 5 and 10 and 7.1A/10A values of 0.5 and 0.25. Thus, by rough approximation, clay fractions consist of 8 or 9 parts in ten montmorillonite, about one part in ten illite, and less than one part in ten kaolinite. A more precise statement of composition is not justified because the concentrations of illite and kaolinite are very near the limits of detectability. Because most samples contain minor amounts of sand and have nearly equal amounts of silt and clay (approximately 40 to 45 percent clay-sized sediment), the clay composition of an average bulk sample would be about 36 to 40 percent montmorillonite, 4.0 to 4.5 percent illite, and 2 percent kaolinite.

## DISCUSSION AND INTERPRETATION

### Source and composition of sediments

Mineralogical data are similar for all the lake plains studied, but, inasmuch as the lake-plain sediments constitute an homogenized sample of their sedimentary province, they reflect the composition of the till sheet upon which the lakes existed. Till composition may relate directly to the physical factors controlling its geomorphic character, which in turn influences the rate of deposition and the character of lacustrine deposits. It is possible that differences in till composition throughout the Missouri Coteau may be reflected in the lake-plain deposits.

The high montmorillonoid content (about 40 percent) of lake-plain samples from Mountrail County indicate that nearly half the deposits are derived from Pierre Shale (late Cretaceous) incorporated in the till. The sand and silt components (45 to 50 percent) are principally quartz and feldspar and are probably derived largely from local late Cretaceous and early Tertiary sediments (Fox Hills-Fort Union interval), and from Canadian Shield soils. A minor part of the sand fraction is very well rounded and may have come from basal Paleozoic sandstones in Manitoba (Deadwood-Winnipeg interval). The carbonate component is most likely heterogeneous, being derived in part from local early Tertiary sediments (the Tongue River Formation averages 25 to 30 percent $CaCO_3$; Royse, unpublished data) and from early Paleozoic carbonate rocks in Manitoba. All of these stratigraphic units outcrop (subcrop beneath drift) to the northeast of Mountrail County.

Early stages of lake development

Field and laboratory investigations indicate that, during initial stages of formation, these ice-walled lakes were either atrophic or only slightly oligotrophic. Early instability in the surrounding superglacial environment is indicated by the coarser texture, higher organic-carbon values, and greater C/N ratios at depth in the lake plains. The sand, silt, clay relationships of most samples (figure C-2) can be explained as the result of sediment deposition from fluid suspension under relatively quiescent conditions; such conditions commonly produce fine-grained deposits. It should be noted that, although deeper samples from most of the lake plains contain greater percentages of sand, the silt-clay ratio remains nearly constant. Similar textural relationships have been observed in turbidites on the continental slope off Washington State (Royse, 1964) and in sediments of proglacial (ice-walled) Miller Lake, Alaska (Callender, 1964), both of which are products of unstable depositional environments. Although more data are needed to substantiate the hypothesis, textural analysis suggests that suspension flows (or other high energy transport mechanisms) were active during early stages of lake development.

The increase in organic-carbon content with depth (figure C-3) appears anomalous. It would seem that the greatest concentration of authigenic organic material would result from accumulation during the slow deposition postulated for the latter stages of lake development. The anomaly is explained by the C/N values which also increase with depth. Animal matter contains more nitrogen (protein, fats, and lipids), relative to carbon, than does vegetation; thus sediments containing organic carbon derived from animals yield low C/N values. Previous studies (Trask, 1939; Bader, 1955; Emery, 1960; Scholl, 1963; and others) indicate that C/N values greater than about 14 to 16 are the result of significant contribution from terrestrial vegetation. Thus, the vertical distribution of C/N ratios for lake-plain sediments suggests that terrestrial vegetation contributed a significant portion of the organic fraction only during initial stages of lake development. It is suggested that transport of terrestrial organic debris into the lake basins by mass transport was prevalent during early stages of development when the adjacent ice and super-glacial debris was relatively unstable. Livingstone and others (1958) report that the organic content of sediment deposited during early stages of development of several glacial lakes in northern Alaska was greater than that of sediment deposited during subsequent stages. The rate of inorganic sedimentation was also greatest during early stages. They conclude that a large portion of the organic matter in the sediment is allogenic. The present source of sediment in the lakes is a thawed mantle, approximately a foot thick, which during the summer slides slowly over the frozen substratum into depressions. This mantle is composed primarily of clayey silt with some included peat and supports a tundra vegetation. It appears probable that the terrain surrounding these Alaskan lakes was even less stable during their early development, and that the contribution of terrestrial vegetation was greater then than it is at present.

The uniformity and lack of correlation of carbonate values with sediment size, organic-carbon or nitrogen contents, or depth below the surface suggests that the lacustrine system had no control on carbonate deposition and that it is largely allogenic. The probable source of such large amounts of carbonate is the lower Paleozoic limestone and dolostone which crops out to the northeast in Manitoba. Cobbles of this limestone are common in the till surrounding the lake plains. Because dolomite would not likely form in this environment, the presence of high percentages of dolomite, determined by X-ray analysis of bulk samples, tends to substantiate that carbonate is largely allogenic. Although statistical tests were not made, it is suspected that the difference in carbonate contents between the lake plains is not significant.

## Latter stages of lake development

The latter stages of lacustrine development are characterized by a general decrease in sediment size, organic-carbon content, and C/N ratios. These factors suggest that the rate of sedimentation had decreased significantly from that of the initial stages and that both the lakes and the surrounding terrain had stabilized considerably. The total biomass in the lakes was probably low throughout their history, but the decrease in C/N values in the upper sediments, combined with the presence of ostracods, indicates that organic productivity was greatest and contribution of terrestrial vegetation was minimal during their latter stages of development. Similar conclusions are reported by Livingstone and others (1958) who found that the rate of inorganic sedimentation in several Alaskan glacial lakes decreased with development and that the organic content of latter stage sediments contained a significantly greater authigenic component.

Ca/Mg values, although available only for lake plain D, show a relative increase (about 30 percent) of calcium with respect to magnesium in the upper half of the lake plain (figure C-3). This increase in Ca/Mg might represent a new mode of calcium addition to the lake sediments. Microscopic inspection revealed the presence of rounded grains of gypsum in the coarse fractions of these "upper" samples. Although it appears (intuitively) unlikely, the possibility that some of these lakes became saline during their latter stage cannot be entirely discounted. Additional study is needed to clarify calcium-magnesium relationships.

## Concluding remarks

The data presented here suggest a general sequence of development for the lake-plain deposits in Mountrail County. This sequence involved initial instability followed by increasing stability within and adjacent to the lacustrine environment. The lakes appear to have never evolved beyond the oligotrophic stage, the natural processes of normal lacustrine succession (Lindeman, 1942) apparently having been arrested by changes in physiographic factors.

Although the data of this report are inadequate to support a general hypothesis for the development of ice-walled-lake plain deposits throughout the Missouri Coteau, they afford limited insight into a few such deposits in a small locale. The general procedure utilized in this study appears to yield useful results and may afford an approach for additional study throughout the Missouri Coteau. Such study is encouraged to test the tentative conclusions offered here.

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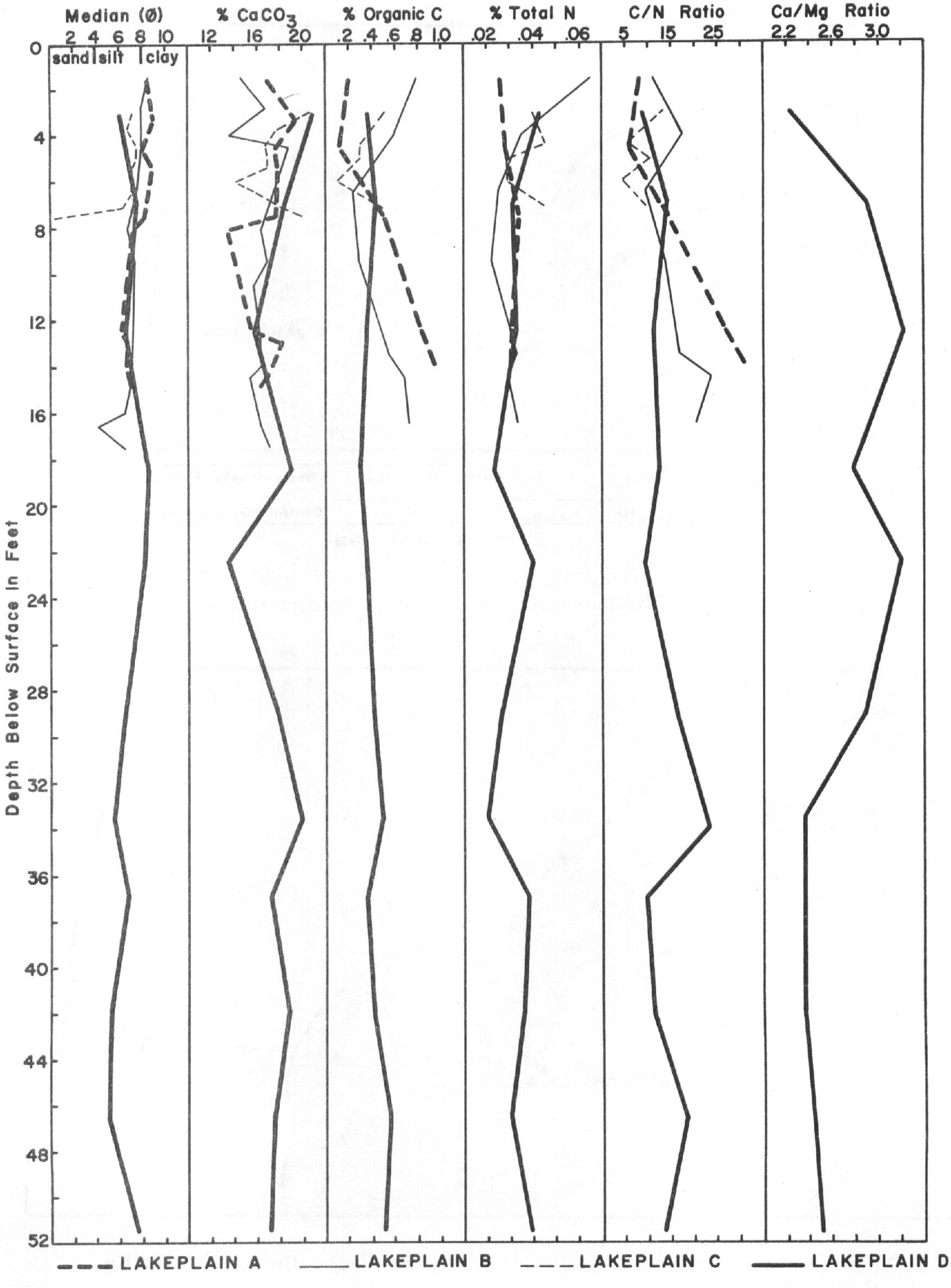


Figure C-3. Sedimentary parameters as a function of depth.

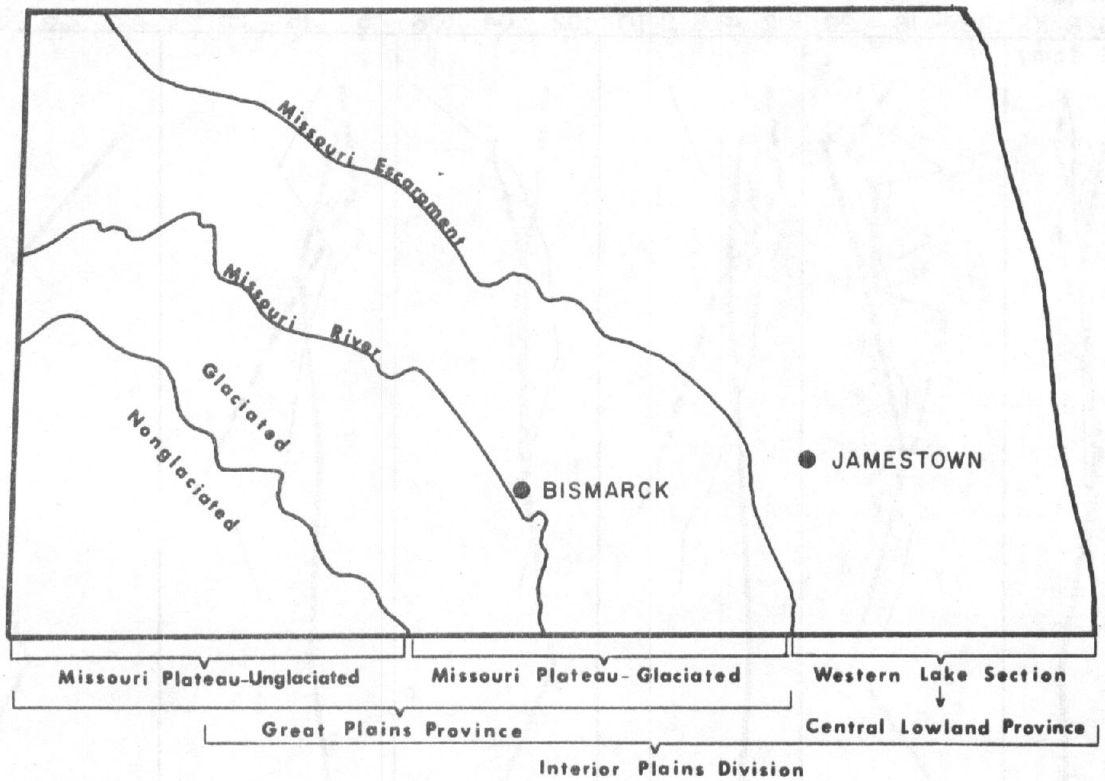


Figure D-1. Physical divisions of North Dakota according to Fenneman.

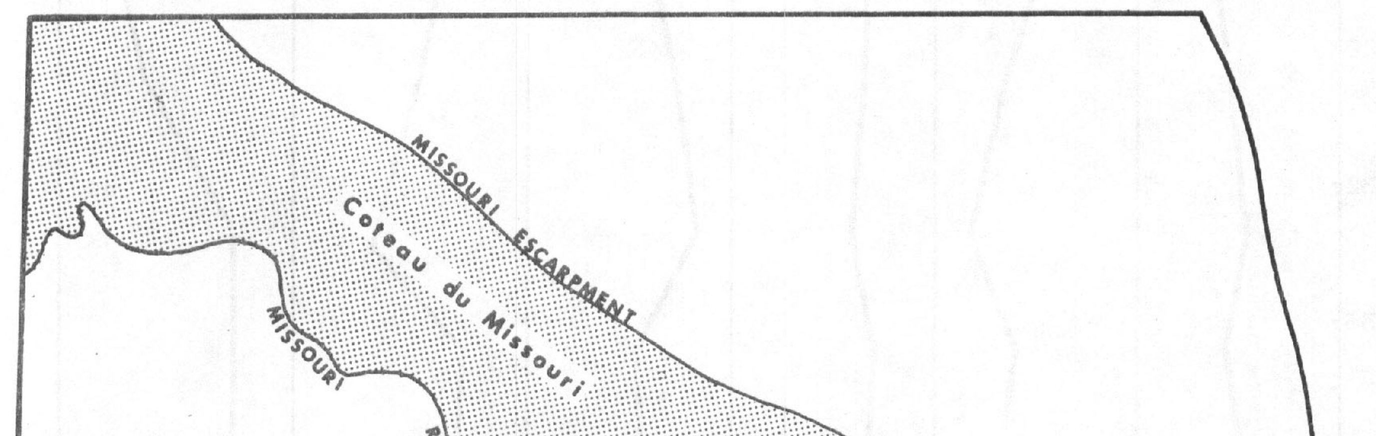


Figure D-2. The extent of the Coteau du Missouri according to Chamberlin (1883).

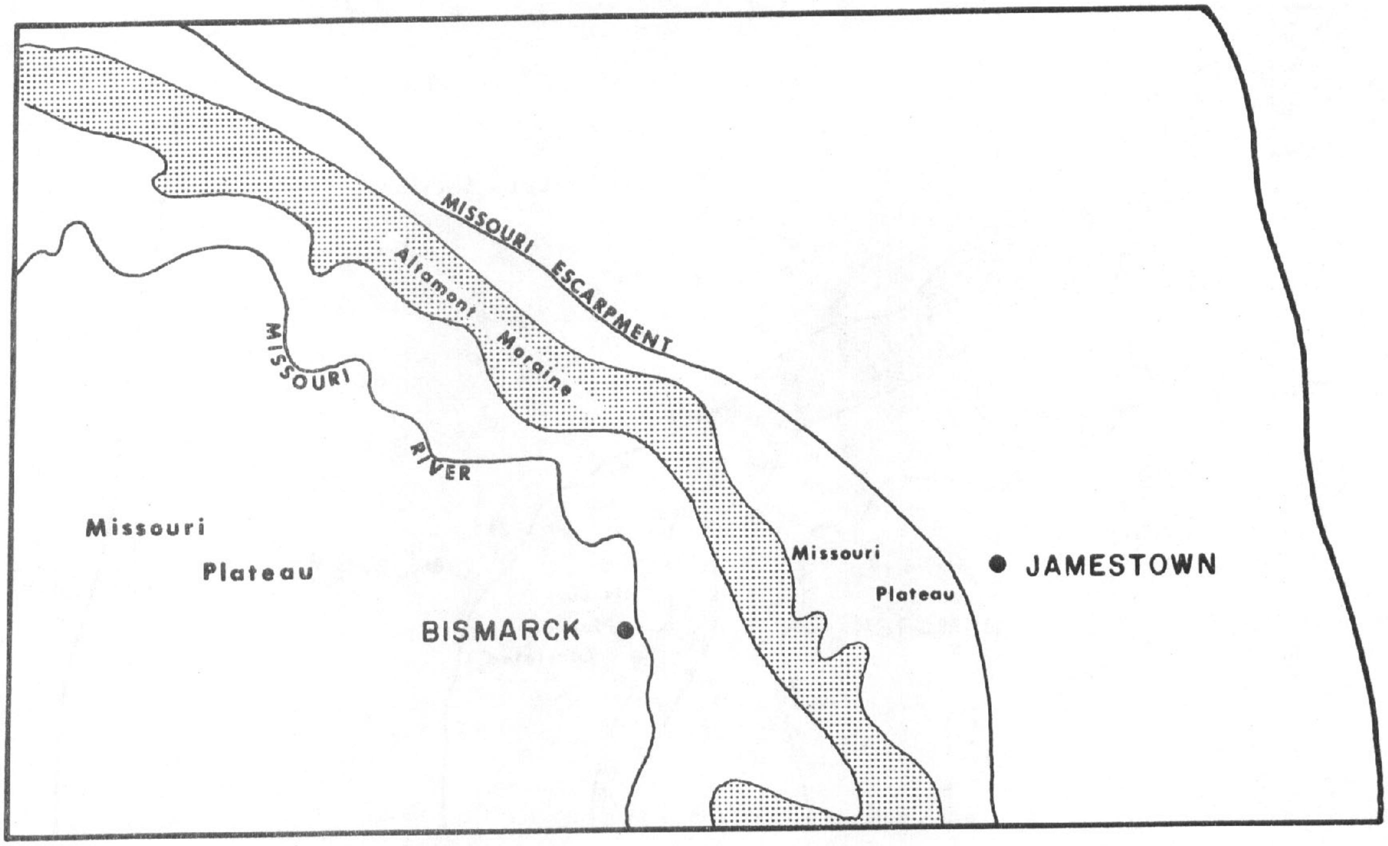


Figure D-3. The relationship of the Altamont moraine of the Missouri Plateau according to Leonard (1919).

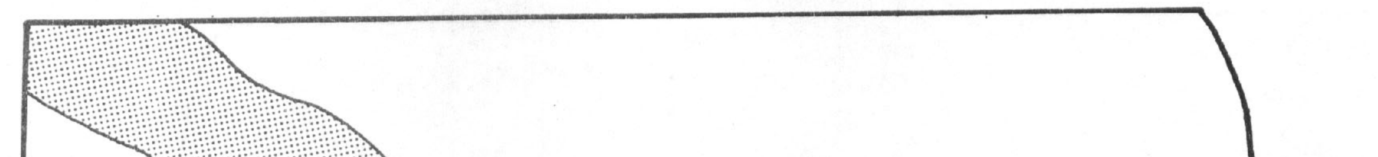


Figure D-4. The extent of the Missouri Coteau according to Lemke and Colton (1958).

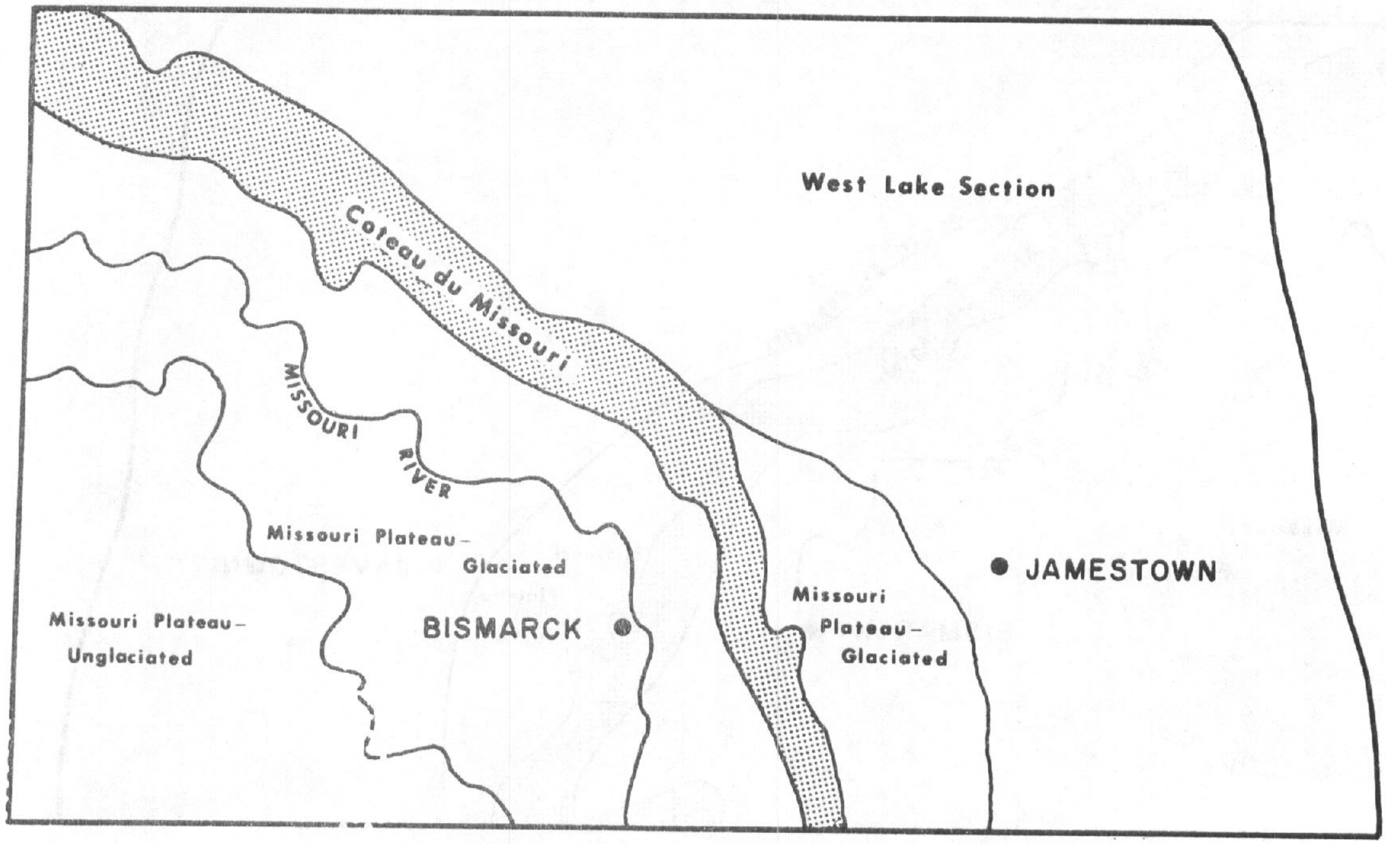


Figure D-5. The extent of the Coteau du Missouri after Lemke (1960).

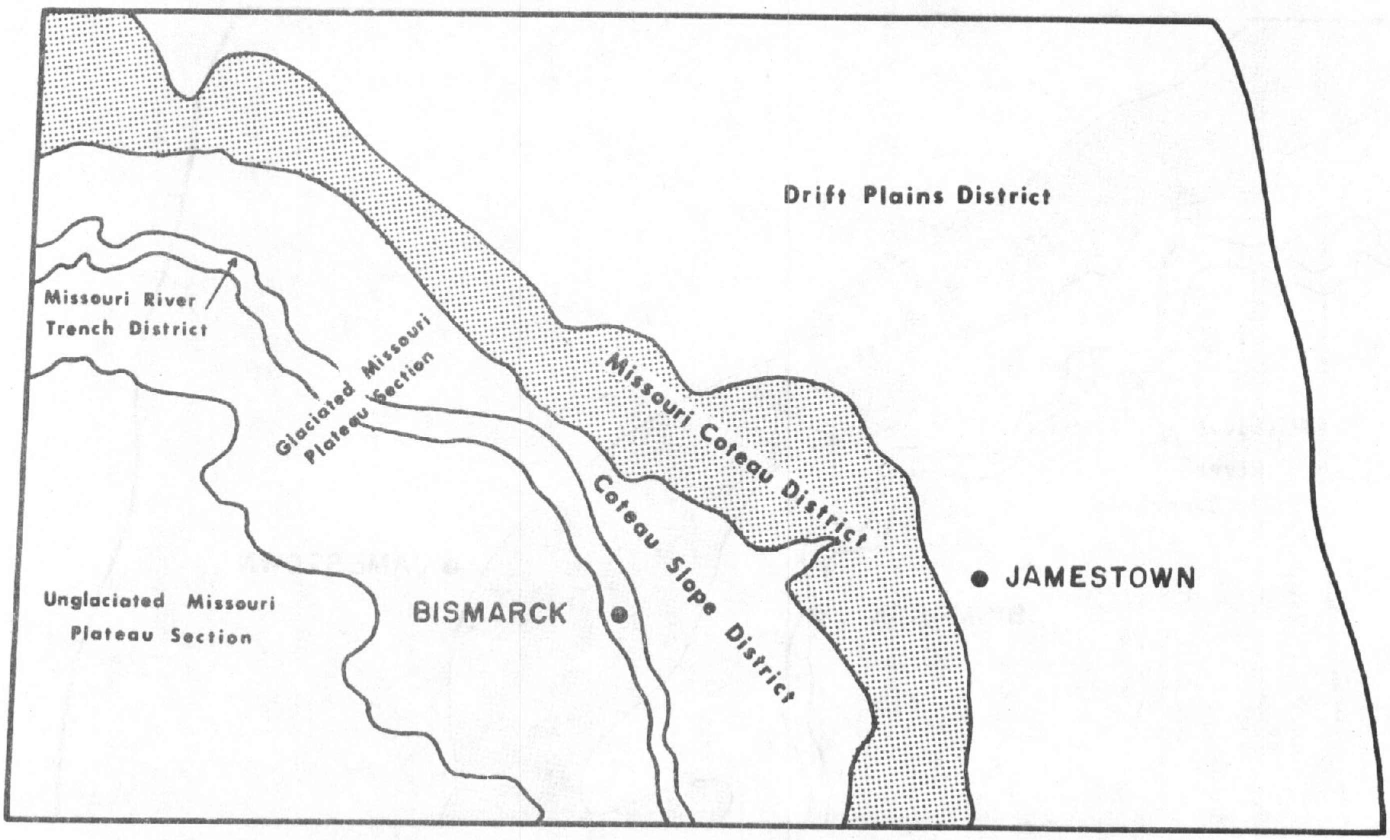


Figure D-6. The extent of the Coteau du Missouri after Kume and Hansen (1965), and Bluemle (1965).

30-D

# THE EXTENT OF THE COTEAU DU MISSOURI IN SOUTH-CENTRAL NORTH DAKOTA

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The areal extent and topography of the Coteau du Missouri in south-central North Dakota has been interpreted in several different ways. The purpose of this paper is to review some of these interpretations and to provide a clearer understanding of the area and associated terminology.

The term "Coteau du Missouri" originated with the early French explorers and traders and probably referred to the rugged or hilly topography, marked in part by the Missouri Escarpment (or Missouri Coteau Escarpment or Coteau Escarpment), that is located east of the Missouri River (fig. D-1). A precise delineation or definition of the area encompassed by the early usage of the term appears to be lacking and, in fact, may never have existed. Eventually the term "Missouri Coteau" was used as a synonym and is now preferred by many.

In 1883 Chamberlin specifically defined the area of the Coteau du Missouri as being bounded on the east by the foot of the Missouri Escarpment and on the west by the Missouri River. He also attributed the altitude of the area largely to the bedrock rather than to deposition of glacial drift (fig. D-2). Later, in 1904, Willard and Erickson discussed the term "coteau" as it applied to this area and stated, in contrast with Chamberlin, that it referred only to the morainic hills between the Missouri River and the Missouri Escarpment and did not pertain to the plateau or bedrock topography. In 1919 Leonard was even more specific in defining the Coteau du Missouri as a rough belt of morainic hills that he correlated with what has been called the Altamont Moraine (fig. D-3). Leonard shows the moraine on an accompanying map, and it is apparent that, according to him, and unlike the interpretations of Chamberlin and Willard and Erickson, the Missouri Escarpment does not form the eastern boundary of the area in south-central North Dakota. Leonard does, however, mention the escarpment and interprets it as a bedrock feature. Thus, at an early date conflicting interpretations were made of the areal extent and, to a lesser degree, of the nature of the landforms. The following discussion traces the problem of interpretation and terminology to the present.

In 1929 Hard described the Coteau du Missouri as largely morainal in character but did not define a western boundary. He notes that the Missouri Escarpment, which he interprets as marking the eastern boundary of the coteau, is not a bedrock feature in south-central North Dakota as had been suggested previously by several writers.

Fenneman mentioned the area several times (1914, 1916, 1928, 1931) and in his 1931 publication, under a major heading "Missouri Coteau," he discusses the area at some length but does not specifically define the term.[1] Fenneman does describe a strip of land 15 to 25 miles wide located east of the Missouri River in which the most conspicuous element of the landscape is a system of terminal moraines that are merely superimposed upon the underlying bedrock plateau. Fenneman applies the term "coteau" to this tract of moraines and says that it is not meant to refer to the plateau of stratified rocks upon which the moraines rest (agreeing with Willard and Erickson), though this is often spoken of as the "coteau district" or "coteau belt." Furthermore, he divides the area east of the Missouri River and west of the Interior Lowlands into (1) an east-facing slope of the Missouri Escarpment that consists mainly of ground moraine, (2) the "Coteau proper," and (3) the slope between the "Coteau proper" and the Missouri River. Thus Fenneman proposes the term "Coteau proper" and describes it as a morainal tract, but does not define its western boundary. These divisions and the terminology may well be satisfactory, but there is probably much less agreement regarding his additional interpretation that the lower altitude of the bedrock surface within the Interior Lowlands, immediately to the east of the Missouri Escarpment, is the result of a newer and lower peneplain than that which exists in the subsurface in the area of the coteau. According to Fenneman, the two peneplains are separated by a slope 5 to 20 miles in width which forms the Missouri Escarpment. The absence of a clearly defined bedrock escarpment coincident with the Missouri Escarpment has been noted, however, by Hard (1929), and to some degree, by Winters (1963) in south-central North Dakota. But a bedrock escarpment may coincide, at least in part, with the eastern edge of the plateau in north-central and northwest North Dakota, and differences in the altitude of the bedrock surfaces certainly do exist. Peneplanation, on the other hand, has not actually been proved, and the situation might be the result of other conditions.

[1]Some may argue that Fenneman does provide a definition of the coteau by inference because of the heading for the topic of discussion. (See p. 73, 1931.) Actually, a specific definition is lacking, and it may be that Fenneman avoided the problem because of previous conflicting usage.

Another part of Fenneman's discussion that may be somewhat misleading is his statement that moraines are the most conspicuous element of the topography. Fenneman is probably referring to the morainal systems that in the past have been called the Altamont and Gary Moraines. Actually these end moraines constitute only a part of the rugged glacial topography west of the Missouri Escarpment in south-central North Dakota. Much of the glacial topography actually consists of extensive areas of stagnation, or dead-ice moraine and associated features rather than simply end moraine. These areas of stagnation moraine might well have been included within Fenneman's "Coteau proper" even though they may be located east of the positions of the Altamont and Gary Moraines (according to early terminology) and west of the Missouri Escarpment.

In 1933 the U.S. Geographic Board defined the Coteau du Missouri as a narrow plateau beginning in the northwest corner of North Dakota between the Missouri River and the River de Lacs and Souris River and running southeast and south with its southern limit not well defined and its western escarpment forming the bluffs of the Missouri River. On the basis of areal extent the definition agrees with that of Chamberlin but differs from that of Willard and Erickson and, at least in part, Fenneman.

In 1940 Atwood described the Coteau du Missouri as a cuestaform with a thick mantle of glacial debris. A thick covering of drift is known to exist in the eastern part of the area in south-central North Dakota, but this is not necessarily the situation in the western part of the area or in the northwestern part of the state. Even more important, the Coteau du Missouri cannot everywhere be defined as a cuestaform because a clearly defined cuesta that coincides with the Missouri Escarpment is lacking in south-central North Dakota.

Paulson (1952) and Roth and Zimmerman (1955) have defined the Coteau du Missouri, as have Chamberlin and the U.S. Geographic Board, as being bounded on the west by the Missouri River and on the east by the Interior Lowlands. Neither study, however, explains the basis for this delineation although some references to the problem are listed in Paulson's work.

Hainer and Kresl each commented further on the problem in 1956. Hainer describes the coteau first as an escarpment that trends northwest and rises abruptly above the Drift Prairie (Interior Lowlands); he then defines the coteau as a belt of terminal moraines that rests upon the eastern edge of the Missouri Plateau. Hainer does not define the western edge of the area in his study. Kresl relates the Coteau du Missouri in central North Dakota to the Missouri Plateau, interprets the east-facing Missouri Escarpment as a bedrock feature, and states that the Altamont and Gary Moraines, located a considerable distance to the west of the escarpment, form the eastern boundary of the plateau.

In 1958 Lemke and Colton described the Coteau du Missouri as a "belt" of rugged glacial topography located in the northeastern part of the Missouri Plateau (fig. D-4). On accompanying maps the western margin of the "belt" appears to coincide in most places with what they have mapped as the post-Tazewell, pre-Two Creeks drift border, which is at least partially based on the boundary between integrated and non-integrated drainage. The eastern margin of the coteau coincides with the Missouri Escarpment on this map. Later, in 1960, another report by Lemke dealing with the Souris River area in north-central North Dakota was released (fig. D-5). An accompanying map also shows the coteau forming a "belt" east of the Missouri River, but the eastern boundary does not coincide, as it did in Lemke's previous joint publication, with the Missouri Escarpment in south-central North Dakota. Instead, it appears similar to the boundary proposed by Leonard in 1919. The western boundary is similar, although not identical, to that shown by Leonard in 1919 and by Lemke and Colton, 1958. It should be noted here that Lemke and Colton were among the first to recognize that many of the landforms within the area were complex stagnation, or dead-ice, features and that questions regarding the distribution and extent of these features were probably involved in the establishment of some of the boundaries.

In 1962 Rau and others described the areal extent of the coteau as being bounded on the west by the Missouri River and on the east by the Interior Lowlands; they attribute the definition to Fenneman. But a map in the publication shows the western boundary of the coteau far to the east of the Missouri River (see pages 3 and 41).

Clayton (1962) agrees in general with Lemke and Colton's location of the coteau boundary mentioned previously. Clayton proposes the term "Coteau slope" for that area between the "Coteau proper," as defined by Fenneman, and the Missouri River Trench (Lemke and Colton, 1958, fig. 1). He accepts the Missouri Escarpment as the eastern boundary of the coteau and agrees with Lemke and Colton that the area consists of both end-moraine topography and extensive tracts of stagnation, or dead-ice, moraine and associated landforms.

Winters (1963, 1965) mentioned that there was some controversy regarding the characteristics and boundaries of the area and in 1960 described some landforms resulting from glacial stagnation. He accepts the Missouri Escarpment as the eastern boundary of the area but does not indicate the western border of the coteau.

Kresl again commented on the Missouri Coteau in 1964 and stated that it consists of "a wide belt of stagnant ice moraine coincidental in areal extent and location with the Max Moraine of Townsend and Jenke (1951) and the Altamont Moraine of North Dakota . . ." (1964, p. 106). He reviews some of the terminology pertinent to establishing the western boundary of the coteau du Missouri and agrees with Lemke and Colton's and Clayton's placement. He also accepts the Missouri Escarpment as the eastern boundary of the coteau.

Kume and Hansen briefly discuss their interpretation of the physiographic terminology in 1965 and describe the Missouri Coteau as "the high morainic belt with non-integrated drainage" that is part of the Glaciated Missouri Plateau section of the Great Plains (Physiographic) Province (fig. D-6). This appears to be in agreement with Lemke and Colton's (1958) as well as Clayton's (1962) views regarding the western extent of the coteau, but there are some differences in the boundaries shown on the maps in these publications. Similar differences exist regarding the eastern border of the area shown on these maps, but they are minor and Kume and Hansen do not comment on this boundary in their text, undoubtedly because it was not directly pertinent to the area on which they were reporting.[2]

Additional publications might be discussed (for example: Flint, 1955; Townsend and Jenke, 1951; Howard, 1960; Thornbury, 1965), but it is obvious from the references cited that significant differences in interpretation exist in regard to the coteau in south-central North Dakota. Some factors that may have led to the many different interpretations are (1) the vagueness of the early French terminology, (2) lack of agreement on the level of generalization for physiographic description, and (3) insufficient detailed information on the various landforms and bedrock conditions in the area between the Missouri Escarpment and the Missouri River. Most of the writers cited agree that the Missouri Escarpment forms the eastern boundary of the area. Two exceptions are Leonard (1919) and Lemke (1960).[3] None disagree that the escarpment is a conspicuous part of the landscape in south-central North Dakota. The difficulty is that the escarpment does not everywhere mark the easternmost extent of the rugged morainal topography so characteristic of much of the coteau. If the Coteau du Missouri is defined, at least in part, as a high area of rugged topography consisting of end moraines and stagnation moraine along with their associated landforms, then a new subsection similar to Fenneman's "Coteau proper" and Clayton's "Coteau slope" might be appropriate for the areas east of such a topography,

[2]Bluemle (1965, p. 3) shows the same map used by Kume and Hansen and states that it is a modification of Lemke and Colton (1958) and Clayton (1962), but the basis for the modification and placement of the boundaries is not discussed.

[3]Another possible exception may be shown on the identical maps used by Kume and Hansen (1965, p. 6) and Bluemle (1965, p. 3), but neither publication contains an explanation for this boundary, probably because it is not directly pertinent to the studies.

but west of the foot of the escarpment.[4] If, however, a less specific definition is desirable, then the escarpment may well serve as the eastern border of the area.

The location of the western boundary is more involved. Many workers have indicated the Missouri River (or the valley, or "trench," of the river) as the western boundary. However, it has long been recognized that some of the topography immediately east of the river is more similar to the topography west of the river than it is to the extensive tracts of rugged glacial landforms located farther east which are so characteristic of the higher parts of the coteau. This situation presents two questions: should the landforms immediately east of the river be included within the Coteau du Missouri, and if they are not considered part of the coteau, where and on what basis should the western boundary of the coteau be placed? Lemke and Colton (1958) used the boundary between relatively well-integrated and non-integrated drainage to separate the Coteau du Missouri to the east from a less strongly glaciated and older topography to the west, which Clayton (1962) later named the Coteau slope.[5] This is certainly a refinement in the description of the topography, but two questions arise with this usage. First, is it best to define the Coteau du Missouri as a broad district east of the Missouri River and west of the Missouri Escarpment, as many previous workers have done, or is it desirable to redefine the area more precisely as detailed studies are completed? Both procedures have obvious advantages and disadvantages, not the least of which is that definitions applied in south-central North Dakota must also be meaningfully applicable to other areas of North Dakota as well as in South Dakota and Canada. Second, if the Coteau slope is differentiated from the Coteau du Missouri (the probable equivalent of Fenneman's "Coteau proper"), what are the best criteria for making such a distinction? As mentioned previously, Lemke and Colton (1958) use the degree of drainage development. Others might prefer the boundary to coincide with a rock-stratigraphic unit, but because the extent of till sheets and their associated outwash may be difficult, if not impossible, to determine, this procedure may not be practical. Clearly, the term "Coteau du Missouri" refers to the topography, and thus the area might best be defined as the basis of an assemblage of landforms rather than a drift sheet or the degree of drainage development. Landforms that may be considered typical of the coteau in south-central North Dakota as it is defined by Lemke and Colton include certain end moraines; hummocky stagnation, or dead-ice, moraine; features similar, if not identical, to moraine plateaus; ice-restricted (walled)

[4]If this procedure is followed, it might be appropriate to recognize another section that coincides with the east slope of the escarpment. This might well be referred to as the Escarpment slope.

[5]Lemke and Colton also indicated that this same boundary is (1) marked by the Burnstad Moraine in south-central North Dakota and (2) separates "Tazewell (?)" from "Post-Tazewell, pre-Two Creeks" drifts.

lacustrine plains; collapsed outwash topography; gravel trains, or ice-walled channels; and various types of disintegration ridges. The major difficulties are (1) that the extent of a topography consisting of such landforms is not precisely known everywhere, (2) that additional landforms might be included within this assemblage elsewhere and (3) that all those mentioned do not necessarily have to exist everywhere in the coteau. Thus, the coteau, as defined by Lemke and Colton (1958) and others, may be defined largely on the basis of the extent of depositional glacial landforms within a certain area of the Missouri Plateau that have resulted from extensive glacial stagnation.

In conclusion, because of the present state of the problem of definition and delineation, it may be awkward to use the term "Coteau du Missouri" to describe precisely either a specific area or a type of landscape in the upper Great Plains. However, the term is well established in the literature and is in current usage. Therefore, it may be unwise to abandon it completely, and when it is used, the problem of definition should be considered or the criteria for delineating the area should be explained. Possibly agreement among those who have studied the coteau, not only in North Dakota but in South Dakota and Canada as well, will be forthcoming as more is learned about the area. To be more precise, one may apply names other than the Coteau du Missouri to subsection, or districts, of Fenneman's glaciated Missouri Plateau section of the Great Plains Province (see, for example, Kume and Hansen, 1965, fig. 1, p. 6). Examples of such subsections may be the Coteau proper and the Coteau slope. If the boundaries between the subsections are not appropriate, changes can then be made without necessarily changing the meaning of the terminology or the borders of the area as a whole.

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30-E

# LATE PLEISTOCENE MOLLUSCA OF THE MISSOURI COTEAU DISTRICT, NORTH DAKOTA A NOTE and BIBLIOGRAPHY

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Fifteen genera (28 species and subspecies) of fossil, pre-Recent late Pleistocene Mollusca occur in ice-contact sediments of the Streeter and Burnstad drifts on the Missouri Coteau. Forty sites (see table E-1) have been studied in varying degrees of detail; 22 with a paleontologist collecting and describing the lithology in the field, and 18 with geologists collecting fossiliferous sediments and describing lithology in the field and submitting fossils for identification and paleoecologic interpretation by a paleontologist.

Published discussions of the fauna of the region and its paleoecology are listed in the bibliography. An article synthesizing studies of North Dakota Pleistocene Mollusca is being submitted to the Journal of Paleontology by this writer and publication is expected in 1967.

The faunules from the various fossil sites are dominated by branchiate (gill-breathing) mollusks, in contrast to the Recent molluscan fauna of the area, which is strongly dominated by pulmonates (lung-breathers). The ice-contact nature of the late-Wisconsin habitats is established not by the fauna, but by the surficial geology of the district. Fossils are found in flat-lying and collapsed lake sediments, which exist as topographically high, butte-like bodies, and in sinuous bodies of stream-laid sediments, which lie across hills and valleys in apparent violation of the physical principle that water flows downhill. The Coteau has experienced little post-glacial erosion, thus negating the explanation that these sediments are erosional remnants. The lakes were basined in stagnant glacier ice, and the streams flowed over topography which was ice-cored.

Mussel shells from the Huffnungsthall Site show that drift-insulated habitats had become established on the Burnstad ice as early as 11,650 ± 310 B.P. (W-974). The presence of mussel shells at the Gutschmidt site in

Streeter drift indicates that drift-insulated ice-basined lake environments endured in eastern Logan County through 9,000 ± 300 B.P. (W-1019).

The molluscan fauna is listed on table E-2. Its main significance is two-fold. First, it serves the geologist in irrefutably establishing that the sediments assigned to lake environments of deposition were indeed lacustrine. And second, it establishes the fact that at the close of late Wisconsin time the continental ice sheet stagnated in a climate and supported a biota which probably was not much different from that of north-central Minnesota today.

Analogues for the environments postulated for the Missouri Coteau district during late Pleistocene time (fig. E-1) have been found on the Martin River Glacier of south-central Alaska (Tuthill, 1963a, and 1966, Tuthill and Laird, 1964).

Changes in the population structure of the late-Pleistocene molluscan fauna since 9,000 ± 300 B.P. cannot be explained by the decrease in ground and surface water flow expectable at the time the buried ice eventually disappeared. The changes must have been caused by a shift in climate as previously postulated (Tuthill, 1961, 1963, and Tuthill, Clayton, and Holland, 1963, and Tuthill, Clayton, and Laird, 1964). As is shown by Clayton in paper 30-A, about 300 feet of stagnant ice existed in the Missouri Coteau district 12,000 B.P. Radiocarbon dates indicate that the topography remained ice-cored for at least 3000 years. The delivery of ground and surface water by the melting of the buried ice would have averaged only about 1 inch per year. When added to the present average annual precipitation of this region (17 inches) the precipitation would, given present evapotranspiration rates, still be too low to account for the permanent ponds necessary to explain the dominance of molluscan faunules by branchiates. Thus, climatic conditions must have been different, at least with respect to the evapotranspiration-precipitation ratio. The climatic shift necessary to explain the dominance of the modern fauna by pulmonates would require a westward displacement of the present evapotranspiration zero isopleth of only 150 miles.

Table E-1.--Location of studied Pleistocene molluscan fossil sites in the Missouri Coteau District, North Dakota.

| Site Name | ¼ | ¼ | Sec. | Tier N. | Range W. | County N. Dak. | C-14 Age Y.B.P. | Reference (see footnotes) | Type of Habitat (See Fig. E-1.) |
|---|---|---|---|---|---|---|---|---|---|
| 1. Cleveland | SE | SW | 17 | 139 | 67 | Stutsman | 11,070± 300,W.956 | 1 | 4 |
| 2. Schauer Farm | SE | SE | 29 | 137 | 69 | " | 9,870± 290,W.954 | 2,3 | 6,7 |
| 3. Biderman Farm | SW | SW | 9 | 136 | 69 | Logan | | 2,4,5 | 6,7 |
| 4. Schlenker Farm | NW | NW | 28 | 135 | 68 | " | | 2,4,5 | 6 |
| 5. Gutschmidt Farm | NW | NW | 20 | 135 | 67 | " | 9000± 300,W.1019 | 2,4,5 | 6 |
| 6. Brenneise Farm | NW | NW | 27 | 135 | 71 | " | | 2,5 | 6 |
| 7. Kroeber Farm | NW | NW | 35 | 135 | 73 | " | | 2 | 6 |
| 8. Huffnungsthall Farm | SW | NW | 20 | 132 | 68 | McIntosh | 11,650± 310,W.974 | 2,5 | 1,3,6 and/or 7 |
| 9. Lowenthall Farm | SW | SE | 16 | 132 | 69 | " | | 2,5 | 6 |
| 10. Mummy Cat Slough | NW | NW | 14 | 132 | 69 | " | | 2,5 | 6 |
| 11. Clear Lake | SE | SE | 21 | 132 | 70 | " | | 2,5 | 6 |
| 12. Rosenthall No. 1 | NW | NE | 36 | 131 | 68 | " | | 2 | 3 and/or 6 |
| 13. Rosenthall No. 2 | | N½ | 34 | 131 | 67 | " | | 2,5 | 6 |
| 14. Rosenthall No. 3 | NE | SE | 25 | 131 | 68 | " | | 2,5 | 6 |

Table E-1.--Continued.

| Site Name | ¼ | ¼ | Sec. | Tier N. | Range W. | County N. Dak. | C-14 Age Y.B.P. | Reference (see footnotes) | Type of Habitat (see Fig. E-1.) |
|---|---|---|---|---|---|---|---|---|---|
| 15. Rosen-thall No. 4 | NW | NE | 27 | 131 | 68 | McIntosh | | 2,5 | 6 |
| 16. Rosen-thall No. 5 | SE | NE | 6 | 131 | 68 | " | | 2,5 | 2 |
| 17. Antelope No. 1 | NW | NW | 30 | 131 | 67 | " | | 2,5 | 6 |
| 18. Antelope No. 2 | NE | NW | 16 | 131 | 67 | " | | 2,5 | 6 |
| 19. Nue Farm | SE | SE | 36 | 130 | 68 | " | 9620± 350,W.1149 | 2,5 | 4 |
| 20. Iowa No.1 | SE | SE | 34 | 130 | 68 | " | | 2,5 | 2,3 |
| 21. Iowa No.2 | SW | SW | 25 | 130 | 68 | " | | 2,5 | 6 |
| 22. Podell Farm | - | NW | 14 | 135 | 66 | LaMoure | | 2 | 6 |
| 23. Billing-meier Farm | - | NW | 12 | 144 | 75 | Burleigh | | 2,6 | 6 |
| 24. Salt Lake | - | NE | 12 | 144 | 76 | " | | 6 | 6 |
| 25. Florence Lake | NW | NW | 17 | 144 | 76 | Burleigh | | 2,6 | 2 |
| 26. Muller Ranch No. 1 | NE | NE | 14 | 144 | 76 | " | | 2,6 | 6 |
| 27. Muller Ranch No. 2 | NW | NW | 14 | 144 | 76 | " | | 2,6 | 2 and/ or 6 |
| 28. Detlef Ranch | NE | NW | 22 | 144 | 76 | " | | 2,6 | 6 |
| 29. Wheelock Ranch | NW | SW | 11 | 144 | 77 | " | | 2,6 | 6 |
| 30. Painted Woods | NW | NW | 12 | 144 | 79 | " | 10,100± 300,W.1434 | 2,6 | 6 |

Table E-1.--Continued.

| Site Name | ¼ | ¼ | Sec. | Tier N. | Range W. | County N. Dak. | C-14 Age Y.B.P. | Reference (see footnotes) | Type of Habitat (See Fig. E-1.) |
|---|---|---|---|---|---|---|---|---|---|
| 31. Pelican Lake | NW | SW | 13 | 144 | 77 | Burleigh | | 2,6 | 6 |
| 32. Boynton Ranch | SE | NE | 19 | 143 | 75 | " | 9,990± 300,W.1436 | 2,6 | 6 |
| 33. Toether Farm | NE | NE | 13 | 148 | 76 | Sheridan | | 2,7 | 5 and/ or 6 |
| 34. Neff Farm | SE | SE | 19 | 148 | 77 | " | | 2,7 | 6 |
| 35. Prophets Mt. | NE | NE | 29 | 147 | 78 | " | | 2,7 | 6 |
| 36. Schroeder Farm | NE | SE | 34 | 147 | 77 | " | | 2,7 | 2 |
| 37. Silber- mann Farm | NE | SE | 28 | 147 | 77 | " | | 2,7 | 2 |
| 38. McClusky | NE | NE | 31 | 147 | 77 | " | | 2,7 | 1,4,5, and/or 6 |
| 39. Stock Farm | SW | NW | 15 | 146 | 77 | " | | 2,7 | 6 |
| 40. Rognle Farm | SW | SW | 34 | 160 | 96 | Divide | | 2 | 6 |

1. Tuthill, 1961
2. Tuthill, 1963b
3. Winters, 1963
4. Clayton, 1961
5. Clayton, 1962
6. Tuthill, 1965b
7. Sherrod, 1963

Table E-2.--List of late Pleistocene Mollusca from the Missouri Coteau district, North Dakota.

| Taxon | Ca. 12,500 to 8,500 B.P. | Modern |
|---|---|---|
| **Pelecypoda** | | |
| Naiades | | |
| Anodonta grandis Say 1829 | X | X |
| Anodontoides ferrussacianus (Lea, 1834) | X | |
| Lampsilis luteolus Lamarck, 1819 (= L. siliquoidea Barnes, 1823 of some authors) | X | |
| Sphaeriidae | | |
| Sphaerium simile Say, 1817 | X | |
| Sphaerium sp. | X | X |
| Pisidium sp. | X | X |
| **Gastropoda** | | |
| Pulmonata | | |
| Aplexa hypnorum (Linné, 1858) | | X |
| Armiger crista (Linné, 1758) | X | X |
| Ferrissia sp. | X | |
| Gyraulus parvus (Say, 1817) | X | X |
| Gyraulus sp. | X | X |
| Helisoma anceps (Menke, 1830) | X | X |
| H. campanulatum (Say, 1821) | X | |
| H. trivolvis (Say, 1817) | X | X |
| Helisoma sp. (juveniles or fragments) | X | X |

| Taxon | Ca. 12,500 to 8,500 B.P. | Modern |
|---|---|---|
| *Lymnaea humilis* (Say, 1822) (*sensu* Hubendick, 1951) | X | X |
| *L. palustris* (Müller, 1774) | X | X |
| *L. stagnalis* (Linné, 1758) | X | X |
| *Lymnaea sp.* (juveniles or fragments) | X | X |
| *Physa spp.* | X | X |
| *Promenetus exacuous* (Say, 1821) | X | X |
| *Discus* ? | | X |
| *Euconulus fulvus* (Müller, 1774) | | X |
| *Gastrocopta armifera* (Say, 1821) | | X |
| *Gastrocopta sp.* (fragments) | X | |
| Pupillid shell fragments not *Gastrocopta* | | X |
| *Succinea avara* ? (Say, 1824) | X | |
| *Zonitoides arborea* (Say, 1816) | | X |
| *Branchiata* | | |
| *Amnicola limosa* (Say, 1817) (includes *A. leightoni* Baker, 1920) | X | X |
| *Amnicola sp.* (juveniles or fragments) | X | |
| *Valvata bicarinata* ? Lea, 1841 | X | |
| *V. lewisi* | X | |
| *V. tricarinata* (Say, 1817) | X | X |
| *Valvata sp.* (juveniles or fragments) | X | |

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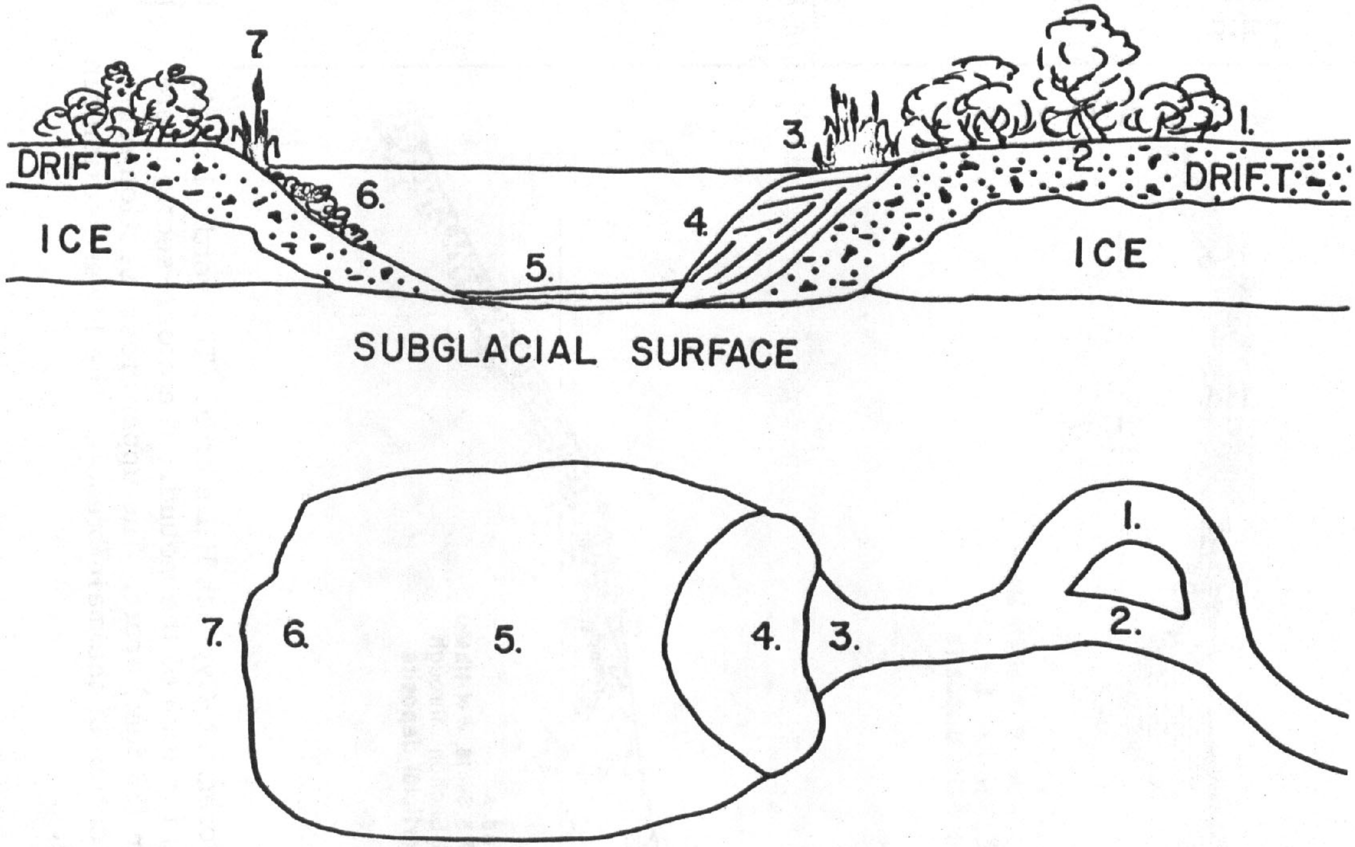


Figure E-1. Molluscan habitats postulated for late Wisconsin Missouri Coteau.

1. Backwaters in Stream Systems. Waters with seasonally high temperatures and erratic water levels. Ideal habitat for pulmonates. Seasonal mixing of shells of flowing-water and standing-water mollusks. Seasonal flushing of sediments and shells to depositional environments lower in drainage system.
2. Streams. Highly oxygenated waters. Suitable habitat for running-water forms. Mixing of shells likely in sediments brought from upstream. Post-mortem transportation of shells from this habitat likely. Water temperature cooler than in habitat 1 or 3.
3. Delta Top. Habitats ranging from exposed mud flats with low dissolved oxygen to running water. Water level highly variable, temperature likely high in standing water, cooler in running water but higher than in 2 above. Mixing of shells likely.
4. Delta Front. Cool to warm water with maximum delivery of organic detritus expectable. Firm substrate; maximum mixing of shells from quiet- and running-water environments upstream would be expected; optimum environment for mussels.
5. Deep Lake. If lakes, deep enough for stabilization of a thermocline existed, an oxygen-poor hypolimnial environment would have existed. No clear evidence of the existence of this type of lake environment has yet been found in the Missouri Coteau district, but the possibility of its existence cannot be discarded.
6. Shallow Lake. Well oxygenated, cool to warm waters. Optimum environment for maximum number of species of mollusks in stretches not frequently agitated by waves.
7. Lake Margins. Riparian flora, moist surface, and shade. Terrestrial snails and *Lymnaea humilis* should be expected here.

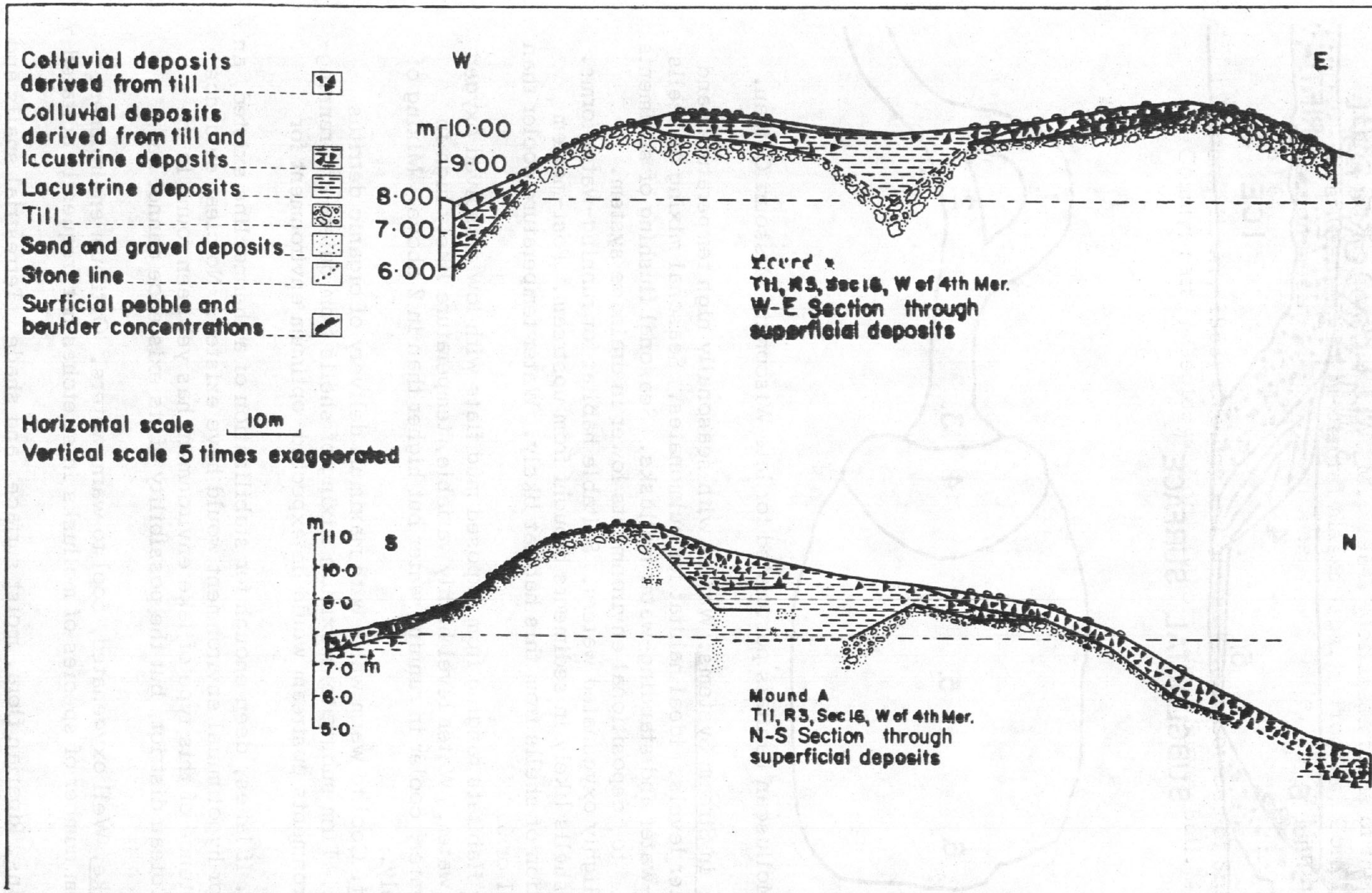


Figure F-1. Cross-sections through one of the mounds of the Foremost-Cypress Hills area. The vertical scale (in m) refers to an arbitrary reference level used in surveying the form of the mound. The cross-sections are based on hand augering data collected from holes spaced 10m (33 feet) apart. The upper cross-section runs across the center of the mound and perpendicularly to the direction of the main breach. The lower section runs along the main breach. The mound is portrayed in figure F-3.

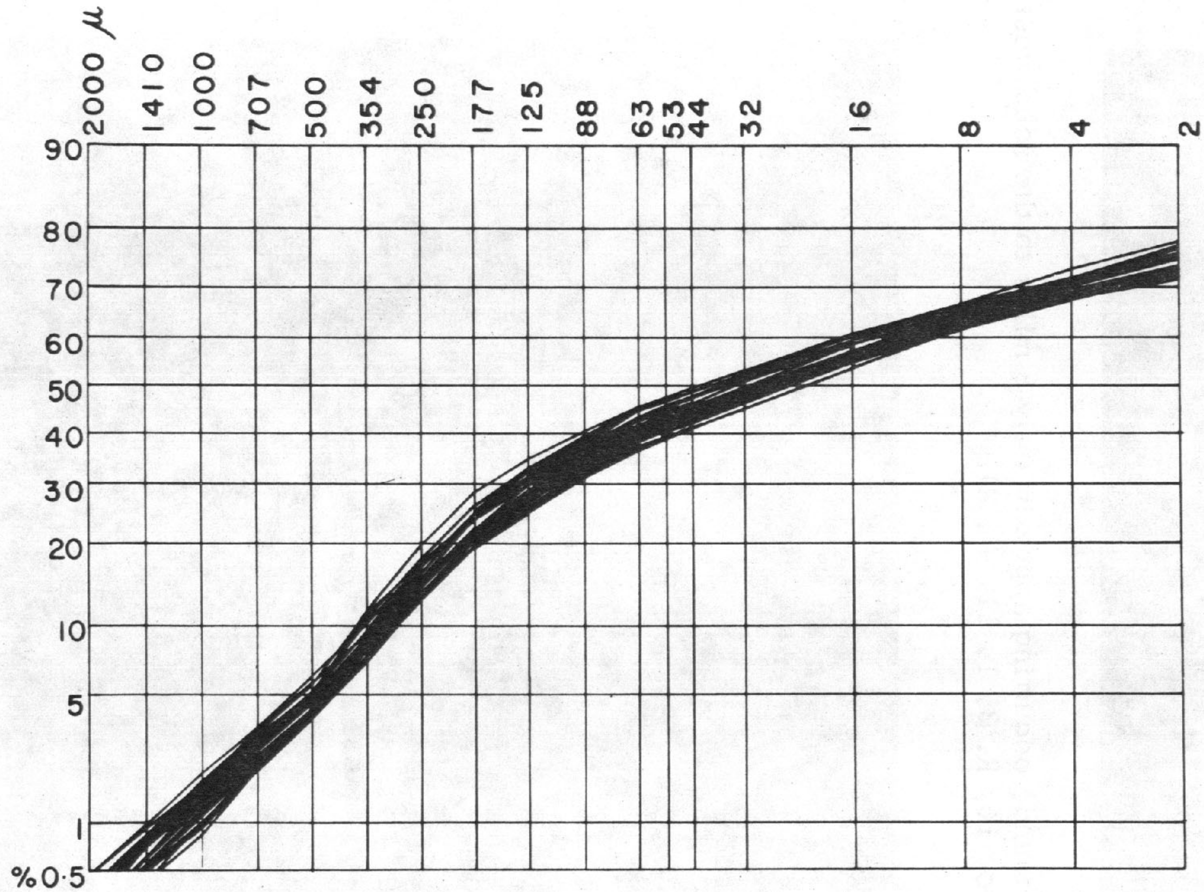


Figure F-2. Grainsize distribution of 38 samples of till taken from various locations and depths from one of the mounds and from the intermound areas adjoining it. Vertical lines represent the size-classes used in the analysis. The percentage scale (left) is drawn on a probability basis.

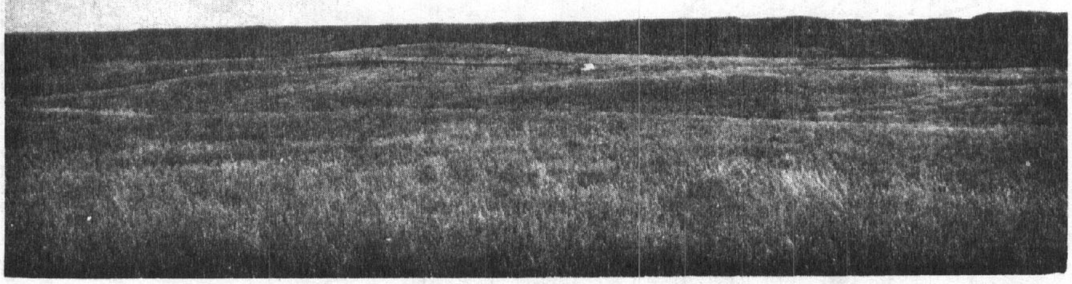

Figure F-3. Prairie mound, occurring approximately 8 miles to the southwest of Irvine, Alberta (sec. 16, R. 3, T. 11, W. 4).

Figure F-4. Prairie mound field, occurring North of Lake Pakowki, in the Foremost-Cypress Hills area of Alberta.

# ON THE PERIGLACIAL ORIGIN OF PRAIRIE MOUNDS

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## INTRODUCTION

Till and silt mounds (2, 3), prairie mounds (4), or plains plateaus (5) are a rather common landform of the glacial landscapes of southern and central Alberta and southern Saskatchewan (1).

On the basis of a study of till and silt mounds in the Sturgeon Lake area of west-central Alberta, Henderson (2) concluded that the features are of periglacial origin. He proposed a mechanism in which the centers of giant soil polygons would be raised by the growth of polygonal ice wedges, and an ice core would subsequently be segregated beneath the surface of the raised central part of the polygons. The former presence of an ice core below the centers of the mounds is assumed to explain the usual central depression of the mounds, which generally extends 2 to 3 feet below the rims of the mounds of the Sturgeon Lake area (2).

Henderson's and subsequent descriptions of the mounds are all similar in that the mounds are oval or round in plan, 250 to 600 feet in diameter, from 5 to 30 feet high, and contain commonly one, and less frequently several central depressions, whose floor level is never lower than the adjoining terrain (4, 5, 6, 7, 8). Henderson noted the frequency of occurrence of a breach in the rim of the mounds, which he concluded to be caused by the drainage of the central depression during the gradual melting of the ice core. Subsequent studies apparently overlooked the presence as well as the genetic significance of the breaches in the mound rims. Henderson considered a glacial origin of the till and silt mound to be unlikely, because identical forms could consist of till, lacustrine silt, or a combination of both.

Modern thinking on the formation of ice wedges (9) makes the mechanism of formation proposed by him less likely, since it is incompatible with the contraction theory of ice wedge formation.

Gravenor (4) proposed a supraglacial origin for features that are identical in size and shape with the till and silt mounds of the Sturgeon Lake area. He made particular reference to occurrences near Hemaruka in the Oyen area of southern Alberta. The mounds, renamed "prairie mounds," are explained from the trapping of ablation till in pits on the surface of stagnant ice bodies. According to Gravenor, mounds with central depressions result when debris accumulation in pits becomes sufficiently thick to inhibit the progress of melting beneath the deposit; relief inversion through continued melting of adjacent, exposed ice causes an ice core to be formed beneath the debris. Eventual melting of this core leaves a central depression within the mound. Gravenor and Kupsch (6), in a general review of ice-disintegration morphology in Western Canada, retained this hypothesis of origin. Stalker (5) however, explained the prairie mounds as ice-pressed drift features, caused by the intrusion of plastic basal till into cavities of a stagnating ice mass. According to him, prairie mounds are genetically related to moraine plateaus (10), and he introduced the term "plains plateau" for the prairie mound. Both Gravenor's supraglacial and Stalker's subglacial hypothesis of origin of the mounds fail to account for prairie mounds consisting substantially or wholly of lacustrine silts, as well as for the one or more breaches of the mound rim. Furthermore, the subglacial mode of origin appears less suited to explain the central depression.

Mathews (7) described the occurrence of many thousands of mounds in the Fort St. John area, British Columbia, which occur typically where late-glacial lacustrine sediments of Glacial Lake Peace form a continuous mantle in excess of 10 feet in thickness. According to him the great majority of the mounds could have developed only long after the ice had withdrawn from the sites where they occur. He proposed that the development of the mounds "possibly involved displacement of water-saturated soil, rather than water alone, during the development of permafrost, and that this soil moved at depth towards points of potential rupture, where permafrost was thinnest, as for example beneath the center of shallow ponds" (7, p. 18).

Observations of glacier-karst phenomena on the Martin River Glacier, Alaska, by Clayton (11), confirms that a pitted glacier surface, as inferred by Gravenor, may indeed be found on the stagnant zone of some present-day glaciers.

In summary, it seems that a non-glacial hypothesis of origin offers a better explanation for all prairie mounds, whether on glacial or on other deposits.

Studies of the regional distribution, size, shape and sediments of the prairie mounds of southern Alberta, within the larger context of a morphoclimatic and morphotectonic analysis of landscape development since deglaciation (1) have uncovered more evidence relating to the origin of the mounds. A brief discussion follows below.

## DISTRIBUTION

The distribution of prairie mounds was mapped on airphotos for approximately 60,000 square miles of terrain in southern and central Alberta and southern Saskatchewan. Hundreds of thousands of prairie mounds were found. The distribution patterns can be summarized as follows:

A. Fields containing almost exclusively prairie mounds that commonly adjoin each other closely. The fields form contour belts, and the distribution in the terrain is clearly controlled by elevation. The fields are associated with smooth topography of little internal relief and gentle angle of regional slope, such as ground moraine plains, lacustrine plains, broad ridges, and broad valley floors. Frequently, glacio-lacustrine deposits occur topographically below the highest contour belt of prairie mounds of an area, while, if more than one belt is found, lower belts are frequently composed of lacustrine deposits, if these deposits are thick. The highest prairie mound belt of an area is usually located on till.

B. Within or associated with hummocky moraine topography: 1. The first type includes those that cover the floor of depressions within hummocky moraine topography, sometimes forming small fields bounded by hummocky moraine. The regional distribution of this pattern does not appear to be dependant on elevation of the terrain. 2. The second type includes those that cover the depression floors as well as the larger protrusions of the hummocky moraine topography, and are associated with pitted winding ridges. Elevation here appears to control distribution, in the sense that at particular topographic levels, prairie mounds are intermixed with hummocky moraine, whereas at higher and lower levels they are not. Furthermore, the distribution pattern is laterally associated with the pattern described under A.

C. In the re-entrants of major topographic rises of the Canadian Prairies such as the Cypress Hills, the Sweet Grass Hills and the Milk River Ridge. Elevation does not appear to control this distribution pattern, but occurrences are limited to terrain sloping less than 5 degrees.

One prairie mound belt has been traced over 500 miles from the Cypress Hills, via Lethbridge, Gleichen, Drumheller, and Stettler to Edmonton. It is a continuous feature, consisting dominantly of patterns described under A (above) with patterns described under B (2) occurring less frequently. This prairie mound belt bends in an upstream direction, and thus forms salients into some of the major valleys of the Western Prairie area, such as those of the Bow and Red Deer systems.

## SHAPE AND PARENT FORM

Detailed analyses of prairie mound shapes were made in the Foremost-Cypress Hills area, southern Alberta. Breaching of the rim of the mounds is very common. But, while each prairie mound has at least one principal breach, each rim actually consists of a circular arrangement of small rises and intervening saddles. The majority of the central depressions of the mounds are not enclosed. Wherever full enclosure of the central depressions was noted, this is by means of a low swell across the principal breach of the mound.

Cross-sections of the landforms were drawn on the basis of hand augering and laboratory data collected from two mounds. Two cross-sections of one of these mounds are given in figure F-1. A photograph of this mound is presented in figure F-3. The various sediments other than till of figure F-1 are all non-glacial in origin. The sandy deposits contained in the till were tentatively explained as of cryogenetic origin, since they are very poorly sorted and could not be explained as stratified intratill deposits. Sorting by repetitive freezing and thawing of till presumably formed these deposits (1). With the exception of the till and the sandy deposits, the various sediments were found to be younger than the till and younger than the mound itself. Because the till of the mound rim contributed to the colluvial deposits found on the lower slopes of the mound the rim must have been higher in the past. The central depression is filled with as much as 7 feet of sediments that are younger than the mound itself (see below). The areas between the mounds are covered by 3 feet or more of superficial sediments that are also younger than the mound. Saddles between closely adjoining mounds received proportionately more superficial deposits.

At the time of formation the fields of prairie mounds thus had greater internal relief. The mound rims were higher (up to 4 feet) and of sharper form, and the central depression was deeper. The internal relief of the mound portrayed in figure F-3 was as much as 10 feet greater than found at present. Not only the breaches of the rim but also the minor saddles within the crestline of the rim were found to pre-date colluvial deposition. Major and minor re-entrants in the outer slope of the mounds are thus an essential part of the parent form of the present-day prairie mound. In summary, the original feature (or parent form) prior to deformation by various subaerial processes was crater-like in shape, with a notched ringwall, and a central depression that did extend near but not below the level of the adjoining intermound areas. Since the prairie mound neither originally was, nor presently is, a closed depression, but is characterized by the presence of breaches, the term "doughnut," inasmuch as it has been applied to these landforms, is a misnomer.

## AGE OF THE PARENT FORM

The superficial lacustrine deposits (fig. F-1), though of primarily lacustrine origin, were redeposited by wind action (1) and are of niveo-aeolian origin in the sense of Vink (12). These deposits form a lateral stratigraphic continuum with colluvial deposits of the lower rim slopes, with which they are intermixed. Freshwater shells occurring at the base of the niveo-aeolian deposits in another mound (not portrayed in figure F-1) have an age of 10,550 radiocarbon years. The parent form was thus in existence prior to the Allerød Period. The niveo-aeolian deposits were tentatively assigned to the Younger Dryas Period (in the sense of Hammen and Vogel, 13).

Inasmuch as the principal, and topographically highest belt of prairie mound fields of the Foremost-Cypress Hills area of southern Alberta occurs on the Etzikom Till Sheet (as defined by Westgate, 14) the prairie mounds of this belt can safely be considered to be younger than the maximum age of 25,000 years (14) found for this till sheet. The age-span cannot be defined more precisely on the basis of available data.

## THE DEPOSIT OF THE PARENT FORM

The grain-size distributions of 38 samples taken from various locations and depths below the rim, central depression and outer slopes of mound A as well as from below the adjoining intermound areas is portrayed in figure F-2. Very little variation of the grain-size distribution of the till is apparent; only one kind of till is obviously present. Comparison with grain-size distributions of many till samples collected throughout the Foremost-Cypress Hills area by Westgate (14), as well as the lack of variation in the granulometries of the 38 samples taken from and in the vicinity of mound A, indicate that this mound consists of basal till insofar as the parent form is concerned.

## ORIGIN OF THE FORM

Inasmuch as the prairie mounds occurring on till are identical in shape and form to those occurring on proglacial lacustrine deposits, the landform cannot have a subglacial or supraglacial genesis, but is either of sub-aquatic or subaerial origin. Furthermore, as the parent form in the case under discussion is made up of basal till, the genesis of the form is probably the result of deformation of an already existing deposit (see also Mathews, 7), rather than of concurrence of deposition and deformation.

The apparent similarity of prairie mounds with glacier karst phenomena and deposits (Gravenor, 4; Clayton, 11) only partly explains the prairie mound fields of southern Alberta, which contain almost exclusively prairie mounds and none of the other expected stagnation or collapse landforms and deposits.

Mathews' proposal of a process similar to pingo formation is more likely however, since:

1. The parent form, and to a lesser extent the present-day prairie mound, has all the form characteristics of the collapsed pingos described from western Europe (15, 16, 17, 18). But, the European collapsed pingos consist of a breached ringwall surrounding a depression that extends below the level of the surrounding terrain. The prairie mound of southern Alberta is a definite mound with a breached ringwall, which surrounds a depression in the mound itself. In these aspects the prairie mound is therefore similar to, but not identical with the collapsed pingos of western Europe.

2. Prairie mound fields in the Red Deer and Wabamun areas of Alberta and the Kindersley area of Saskatchewan contain a minority of forms that are identical with the European collapsed pingos. The same was observed in the Whitecourt area of Alberta (19). In the Edmonton area, Alberta, prairie mounds are generally less strongly "mounded" than found to the west of the Cypress Hills. Thus within this landform group as well as within the prairie mound belts as landform association, there is a variation of the form that ranges from replicas of the generally acknowledged collapsed pingo of western Europe to the "mounded" pingo of prairie mound.

An explanation of the origin of the prairie mound as resulting from a kind of pingo formation, however, raises the following questions:

1. Can till be displaced below the surface under an aggrading permafrost layer, as suggested by Mathews (7) for lacustrine deposits?
2. Can discreet ice cores be segregated in till?
3. What controls the distribution of prairie mound fields which appear to be of limited vertical and rather unlimited lateral extent?

All three questions are interrelated in that they bear on the genesis of the mounds.

1. The liquid limit of the till of the two mounds that were studied in detail is low (24 to 32 percent); the plasticity index is also low (5.5 to 15.5 percent). The clay fraction of the till (22 to 28 percent smaller than 2 microns) is rich in montmorillonite. Rather high mobility under saturated or supersaturated conditions may thus be expected.

The distribution of prairie mounds within the mound fields and their round or oval plan indicates that if they owe their positive relief partly to subsurface displacement of mobile till, this displacement was directed to points rather than to lines of weakness in the frozen layer below which displacement occurred. This would suggest that the mounds are located either over the initial depressions or over the initial high areas of the surfaces (see above) on which they were formed. Location over initially high areas seems unlikely, as permafrost establishment is thought to proceed more rapidly below these; moisture content of the upper horizons below such areas may be expected to be lower than in depressions. However, since the total mass of material in the parent form of the prairie mound (the mass of the rim less the volume of the depression) rises above the surface of the surrounding terrain, inversion of the initial relief then seems to have occurred. Perhaps not initially, but certainly in the later stages of the development of the mounds, the forces that formed the protrusion must have been derived from the process of permafrost aggradation itself. For the less likely case of mound formation over the existing topographic highs of the terrain, this also applies to the initiation of the process. The internal relief of the surfaces on which prairie mounds were formed (above) did not exceed 10 feet. Finally, though growth of a discreet ice core within an existing high area of a till plain is not unlikely (see below), such a process cannot explain all prairie mounds; in particular, the mounds on thick lacustrine silts described by Mathews (7), cannot be explained by growth of discreet ice cores below pre-existing convexities of lacustrine plains that are underlain by up to 100 feet of silts.

In the more probable case, mounds over previous depressions of the terrain, it seems that the greater thickness of frozen overburden below the topographic highs results in greater pressure being exerted on the unfrozen layer below it, relative to the depressions. However, the low relief of the undulating landforms that are laterally associated with the prairie mound belts with a distribution pattern described under A (see paragraph on distribution) would suggest that pressure differences are small, perhaps in the order of 0.5 kg/cm$^2$, since relief is usually only 5 to 10 feet. The continuous strength of frozen soil is from 5 to 15 times smaller than the momentary one (20). However, strength values of soils (20) with a grain-size distribution comparable to the till of the mounds are higher than 0.5 kg/cm$^2$. Unusually low values for the continuous shear strength of this till in a frozen state may be expected in view of its high content of montmorillonite. No test results are available to date, however.

Continued subsurface displacement would lead to relief inversion, and pressure differentials would be maintained only if permafrost aggradation would continue to occur more rapidly beneath the initial convexities of the relief.

Although initial subsurface displacement can be caused by the subsidence of the higher areas of the initial relief, beneath which permafrost is thicker, the continued growth of the prairie mounds can be attributed only to volume expansion resulting from downward progress of the freezing front.

2. Contrary to Mackay's (21) opinion that the phenomenon of pingo growth is limited to a narrow range of sufficiently permeable parent materials, active pingos have been reported as occurring on a wide range of parent lithologies (22, 23). Collapsed pingos also have been found on a variety of deposits, including till (15, 16, 17, 18).

The air intrusion values (24) (a measure of the frost-heave susceptibility of the till) of four samples taken from the mounds, are quite high (average 1.85 kg/cm$^2$). As pointed out above, the grain-size distribution of 38 samples taken from various locations and depths on and near one of the mounds indicates that the till of the mounds is similar in texture to the till beneath the intermound depressions.

In this section, it is assumed that the parent form of the mound fields prior to the formation of the prairie mounds is a till plain with relief of 2 m. Before freezing there is a free water table at 1 m below the surface of the high areas and at 0.2 m below the surface of the depressions. The freezing front or frostline advances twice as fast into the soil above the water table as below it.

The bulk density of frozen till is assumed to be 2 g/cc. On the basis of the average of the four air intrusion values for the till, 1.85 kg/cm$^2$, the pore pressures at the freezing front were calculated following Williams (24). When the frostline reaches 150 cm below the high areas, it is at a depth of 110 cm below the depressions. In the former case, the pore-water pressure before freezing would be 0.050 kg/cm$^2$ (atmospheric pressure at the surface is assumed to be 0.0 kg/cm$^2$). The pore pressure at the penetrating frost line would be -0.477 kg/cm$^2$. In the latter case the pore-water pressure before freezing would be 0.090 kg/cm$^2$, and pore pressure at the penetrating frost line would be -o.557 kg/cm$^2$. According to Williams (24) water would migrate to the frost line until pore-water pressure would equal the pore pressure at the penetrating frost line, and ice segregation would occur. It would in fact occur before the frost line reached the water-table; but this and other effects are not considered here for the sake of brevity.

As the critical pressure difference is greater below the depressions, more ice would be segregated here than under the high areas. Furthermore, as there is a pressure gradient towards the depression it is likely that ice segregation in part occurs with moisture derived from the high areas. Finally, the maximum depth to which frost-heave may be expected to occur (24) is 6.77 m below the high areas and 7.57 m below the surface of the depressions. The retardation in the progress of the freezing front is thus also greatest below the depression. The pressures needed to maintain the subsurface displacement of plastic till thus are more easily generated beneath the higher areas of the initial relief.

In summary, it may be concluded that subsurface displacement of till towards weaker areas in the frozen overburden is not incompatible with a greater measure of ice segregation below these areas.

3. The limited vertical and relatively unlimited lateral extent of the prairie mound belts is best explained as being related to the shore zone of one or more proglacial lakes. The arguments presented for the periglacial and pingo-like origin of the mounds apply to all till and lacustrine plains where the parent material has a large frost-heave potential. However, although large tracts of land in southern and central Alberta could have been covered with these mounds, assuming the proposed mode of origin, they are restricted to rather sharply defined belts that bend upstream into major valleys and sometimes surround higher tracts of land. These belts are therefore interpreted to have formed along the shores of proglacial lakes. Moisture conditions more favorable to prairie mound formation, as outlined above, would more likely occur near such a shorezone. The other patterns of distribution of prairie mounds, in the re-entrants of hilly topography, and in the depressions within hummocky moraine topography, are also explained from the occurrence of locally more favorable moisture conditions than in adjacent higher areas.

The distribution of prairie mound fields was found to be inconsistent with the stagnant ice front theory of Gravenor and Stalker (1). Other factors that could explain both the horizontal and the vertical extent of prairie mound belts have not become apparent. Furthermore, the distribution of the highest prairie mound belts in the Lethbridge, Gleichen, Drumheller, Red Deer and Edmonton areas of Alberta accords closely to the distribution of lacustrine deposits (25). The same was found for the Kindersley area of Saskatchewan (26). Finally, as the highest prairie mound belt is located on till and lower belts are frequently located on lacustrine deposits, the same cause of the distribution of belts apparently operated on differing parent lithologies. This would be consistent with formation of the mounds under periglacial conditions in the shore zones of proglacial lakes that were lowered in stages.

## STAGNATION

Through publications by Gravenor (4, 6), Kupsch (6) and Stalker (5), the prairie mound has become associated with a complex landform association, the origin of which has been explained with large-scale stagnation of the margin of late-Wisconsin continental ice in Alberta and Saskatchewan. Inasmuch as the stagnating ice-front was identified solely on the occurrence of prairie mounds (as proposed by Westgate for the Foremost-Cypress Hills area, 14), some re-interpretation of the glacial geomorphology would follow from a change in explanation of the origin of the prairie mounds. Yet, for the phenomenon of large-scale stagnation itself there is better evidence, the interpretation of which is not affected by rejection of a glacio-genetic origin of the prairie mounds.

## CONCLUSION

Since both a supraglacial and subglacial origin of the prairie mounds cannot explain all characteristics of form, composition and distribution of these mounds, and because the earlier proposed mode of periglacial origin is inconsistent with the contraction theory of ice-wedge formation, the alternative origin, pingo formation combined with subsurface displacement of plastic material, offers a more likely explanation. This is further supported by the fact that the mounds strongly resemble the collapsed pingos described from western Europe. An origin of prairie mound fields in the shore-zone of proglacial lakes is consistent with the relationship of lacustrine deposits and prairie mound belts.

## ACKNOWLEDGMENTS

The air intrusion values of the till samples were determined by P. J. Williams of the Division of Building Research, National Research Council. Discussions on the origin of the Prairie mounds with Drs. Bayrock, Berg and Green of the Alberta Research Council, Drs. Stalker and St. Onge of the Geological Survey of Canada, Dr. Clayton of the University of North Dakota, and with Mr. Williams of the National Research Council are thankfully acknowledged.

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30-G

# GROUND-WATER MOVEMENT AS INDICATED BY PLANTS IN THE PRAIRIE POTHOLE REGION[1]

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## INTRODUCTION

Prairie potholes are water-holding depressions of glacial origin that occur in the drift-covered prairies of North America. Some of the best pothole terrain is on the Coteau du Missouri where glacial stagnation has created a hummocky, knob-and-kettle topography.

Prairie potholes have been called the backbone of duck production in North America. Although the prairie pothole region makes up only 10 percent of the total waterfowl breeding area of this continent, it produces 50 percent of the duck crop in an average year--more than that in bumper years (Smith and others, 1964). Potholes also furnish water and forage for livestock and, when drained, provide rich organic soils for the production of crops.

The results of ecologic studies of potholes have revealed that plants are good indicators of water permanence and relative salinity. Hydrologic studies indicate that inflow to the potholes is mainly from precipitation and overland flow. However, salinity and permanence are largely controlled by ground-water flow systems. Thus, plant indicators of salinity and permanence also indicate ground-water movement surrounding the potholes.

## WATER BUDGET

Consideration of a water budget aids in understanding the effects of ground-water flow on pothole hydrology. Inflow comes from precipitation on the pothole surface, overland flow, and ground-water inflow (seepage). Outflow results from evapotranspiration, overflow, and ground-water

[1]Publication authorized by the Director, U. S. Geological Survey.

outflow (seepage). All inflow contains dissolved solids, but in widely varying concentrations depending on the inflow process. Concentration of dissolved solids is extremely low in precipitation, moderately low in runoff, and relatively high in ground-water inflow. Dissolved solids can be removed from a pothole by overflow and ground-water outflow, but not by evapotranspiration. This results in a mechanism for increasing the salinity of pothole waters. Ground-water outflow is the main agent for removing dissolved solids because most potholes do not generally overflow.

## GROUND-WATER FLOW

Direction and rate of ground-water flow, largely controlled by topography and lithology of the drift, affect the salinity and permanence of pothole water.

The direction of ground-water flow with respect to a pothole--whether inflow, outflow, or some combination of both--is controlled primarily by topography. The phreatic surface, or water table, slopes away from topographically high potholes resulting in continuous outflow seepage, and relatively temporary and fresh pothole water. Conversely, the phreatic surface slopes into topographically low potholes resulting in continuous inflow seepage, and relatively permanent and saline pothole water. Most potholes, between these extremes of either total inflow or outflow seepage, have simultaneous inflow and outflow seepage, depending on the position of the adjacent phreatic surface. Inflow seepage occurs where the adjacent phreatic surface is higher than the water level in the pothole and outflow seepage occurs over the remainder of the pothole where the phreatic surface is lower. Imtermediate salinities ranging from slightly to moderately saline prevail in these potholes, depending on the dominance of inflow or outflow seepage.

Direction of ground-water flow is also affected by the lithology of the drift. Flow tends to be vertical in the poorly permeable till and horizontal in the permeable outwash.

The general movement of ground water is downward on the Coteau because it is a regional topographic high. Outflow seepage is common and relatively fresh pothole water predominates in the till areas on the Coteau. Inflow seepage and saline pothole water are common in the outwash areas which are topographically low on the Coteau.

The rate of ground-water flow with respect to potholes depends largely on the permeability of the underlying drift. Ground-water flow rates are extremely low in the poorly permeable till that underlies most of the pothole terrain on the Coteau. Because of this low rate, only a small amount of water seeps into or out of a pothole. Average seepage rates from potholes in till are about 0.003-foot per day, or about a foot per year. Owing to the slow ground-water movement in till, the salinity of pothole waters may be affected by ion-diffusion processes. If ground water and pothole water salinities are sufficiently different, it is possible that a diffusion gradient could be established whereby ions could diffuse counter to a weak hydraulic gradient. This process may explain the relative freshness of potholes in till that have continuous ground-water inflow.

Ground-water flow rates are fairly high in the permeable outwash deposits that underlie many of the discharge channels and other low areas on and adjacent to the Coteau. Springs and other points of ground-water effluence are abundant in the outwash areas, and the water in potholes and lakes is relatively saline. Ground-water flow rates are high enough that diffusion processes are ineffective, thus evaporites (principally sodium sulfate) accumulate in many of the depressions in the outwash areas.

## PLANT INDICATORS

The kinds and amounts of pothole vegetation are related to the permanence and relative salinity of pothole waters. Salinity and permanence of water in a pothole fluctuate rapidly in response to inflow and outflow. Vegetation changes rather slowly in response to fluctuations in salinity and permanence, so that it tends to integrate the short term changes and adjust to a seasonal or even longer term hydrology of a pothole.

According to R. E. Stewart and others (written communication) of the U. S. Bureau of Sports, Fisheries, and Wildlife, who have made detailed studies of pothole ecology for several years, emergent plants in potholes can be grouped into three distinctive vegetative zones which are called wet-meadow, shallow-marsh, and deep-marsh zones, based on increasing average water depth. The wet-meadow zones are dominated by fine-stemmed grasses or grasslike plants of relatively short stature and including species such as fowl bluegrass (Poa palustris), northern reedgrass (Calamagrostis inexpansa), wild barley (Hordeum jubatum), saltgrass (Distichlis stricta), and Baltic rush (Juncus balticus). The shallow-marsh zones are dominated by medium-stemmed grasses and grasslike plants that are intermediate in height compared to plants of the wet-meadow and deep-marsh zones and include tall mannagrass (Glyceria grandis), giant burreed (Sparganium eurycarpum), slough sedge (Carex atherodes), Whitetop (Scolochloa festucacea), sloughgrass (Beckmannia syzigachne), common-spikerush (Eleocharis palustris), and common three-square (Scirpus americanus). Plants in the deep-marsh zones, coarser and taller than characteristic plants of other zones, commonly include cattails (Typha sp.), river bulrush (Scirpus fluviatalis), hardstem bulrush (Scirpus acutus), and alkali bulrush (Scirpus paludosus).

According to Stewart, the occurrence of each of the emergent zones is more closely correlated with water permanence than average water depth. Wet-meadow zones are found in potholes, or portions thereof, that contain water of varying depths for only brief periods after spring snowmelt or immediately following heavy rainstorms. Shallow-marsh zones generally contain water through spring and early summer, but often are dry from midsummer through fall. In deep-marsh zones, water is ordinarily maintained throughout the spring and summer, frequently extending into fall and winter.

Steward has also found that the distribution of emergent cover within a pothole can be correlated fairly well with water depth. Closed stands of emergents that cover an entire pothole occur in comparatively shallow water. Emergent stands that cover a peripheral zone surrounding a central expanse of open water indicate deep water, while isolated stands interspersed with open water are characteristic of intermediate depths. This correlation of emergent cover with depth may be directly related to the length of time since the soil was last uncovered by water. It appears that even shallow water will drown most emergent vegetation if it persists for a sufficient period of time. Expanses of open water may occur in any portion of a pothole, regardless of depth, when subjected to certain land-use conditions such as heavy grazing.

Stewart has found that species composition of vegetation in the emergent zones and open-water zone is closely correlated with relative salinity or specific conductance (micromhos/cm). Certain plant species including giant burreed (Sparganium eurycarpum), broadleaf water-plantain (Alisma triviale), slender bulrush (Scirpus heterochaetus), and variable-leaf pondweed (Potamogeton gramineus) are intolerant of saline conditions and are restricted to the fresher potholes. A limited number of halophytic species including alkali grass (Puccinellia nuttalliana), samphire (Salicornia rubra), alkali bulrush (Scirpus paludosus), and widgeon-grass (Ruppia maritima), occur commonly in saline potholes. Many species are found in brackish potholes, and a few, such as hardstem bulrush (Scirpus actus) and sago pondweed (Potamogeton pectinatus), reach their best development under these intermediate conditions. The predominant plant associations within the deep-marsh zones may be correlated closely with relative salinity. The dominant species in most of these associations are slender and (or) river bulrush (fresh), cattail or mixtures of cattail and hardstem bulrush (slightly brackish), hardstem bulrush (moderately brackish), mixtures of hardstem bulrush and alkali bulrush (brackish) and alkali bulrush (saline).

## SUMMARY

In summary, plants indicate the salinity and permanence of water in potholes, variables that are controlled primarily by ground-water flow systems. Thus, plants indirectly indicate the effects of ground-water flow systems on the hydrology of potholes.

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30-H

# PALEOECOLOGY OF A PRAIRIE POTHOLE: A PRELIMINARY REPORT

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## INTRODUCTION

The abundant glacier-formed lakes and potholes of the Missouri Coteau contain sediments that record the biota and environment of late-Wisconsin and postglacial time. In central North Dakota the late-Wisconsin deposits contain abundant conifer pollen (Thompson, 1962), and Moir (1958) identified wood and cones of spruce (Picea) that were C-14 dated at 11,430 years ago. Tuthill, Clayton, and Laird (1964) summarized molluscan assemblages collected from ice-contact sediments of the Missouri Coteau. The several assemblages, including 13 genera and 21 species, had a range of C-14 dates of 9,000 to 11,650 years ago. These late-Wisconsin assemblages resembled the modern ones in a large permanent lake in northern Minnesota more than those present-day assemblages in potholes in relatively more arid North Dakota.

The Missouri Coteau of central North Dakota today lies in the prairie region. The prairie is separated from the boreal forest to the north by aspen parkland and to the east it is separated from the pine-hardwood forest by a belt of deciduous forest (fig. H-1). Pollen analyses of sediments on the northern plains indicate that the composition and position of these major vegetation units has changed drastically since the retreat of the last Wisconsin glacier. Late-Wisconsin spruce-dominated vegetation was widespread, according to pollen studies in Saskatchewan (Ritchie and de Vries, 1964), northwestern Minnesota (McAndrews, 1966) and South Dakota (Watts and Wright, 1966; Watts and Bright, manuscript). In horizons dated 11,000 to 10,000 years ago spruce pollen percentages decrease abruptly, and indicators of prairie (herb pollen) increase. Pollen diagrams from postglacial sites in the modern prairie, for instance Pickeral Lake, South Dakota, (Watts and Bright, manuscript) and northwestern Minnesota (McAndrews, 1966), indicate subsequently continuous prairie vegetation. Postglacial pollen diagrams from aspen parkland, (Glenboro site of Ritchie, 1966) and deciduous forest of northwestern Minnesota (McAndrews, 1966) indicate that these vegetation zones developed from prairie or the prairie-related oak savanna 4,000 to 3,000 years ago. Thus, during the period 10,000 to 4,000 years ago the prairie extended eastward into what are now forested areas of Minnesota (Wright, Winter and Patten, 1963; McAndrews, 1966) implying widespread aridity. Evidence for increasingly lower water levels in lakes beginning about 8,000 years ago indicates that this was a time of a marked climatic shift to aridity (McAndrews, 1966; Watts and Bright, manuscript). This aridity may be causually connected with mammal extinction because the youngest C-14 dates of large extinct Pleistocene mammals (Hester, 1960) cluster in this millennium. Paleontological studies of late-glacial and especially postglacial sediments in North Dakota are few, and the investigation of Woodworth Pond was designed to fill this gap.

Woodworth Pond (Stutsman County, North Dakota; SW 1/4, sec. 6, T. 142 N., R. 67 W.) is located in collapsed superglacial stream sediment on the Missouri Coteau at approximately 1,900 ft (580 m) altitude (Winters, 1963). This stream sediment is part of the Streeter drift, which has been assigned an age of about 9,000 to 12,500 years (Clayton, 1966). The pothole is on the Woodworth substation of the Northern Prairie Wildlife Research Center, and the substation director, Mr. Leo Kirsch, supplied the following information. The pothole occupies 25 acres (10 hectares), 16 acres (6.5 hectares) of which are open water where coontail (Ceratophyllum), mare's-tail (Hippuris), water-milfoil (Myriophyllum), pondweed (Potamogeton) and bladderwort (Utricularia) grow. Surrounding the open water are concentric zones of marsh and wet meadow that occupy increasingly shallower water. The marsh zone is dominated by bulrush (Scirpus), cat-tail (Typha) and sedges (Carex) and includes as subdominants spike-rush (Eleocharis), water parsnip (Sium), arrowhead (Sagittaria) and willow (Salix); the wet meadow zone is dominated by grasses and sedges. The pond is permanent, and its outlet flows at least part of the year; the water is considered "fresh."

This report is preliminary because we lack C-14 dates, chemical and physical analyses of the sediments, and detailed knowledge of pothole vegetation in relation to water chemistry.

McAndrews and Stewart did the field work and identified the pollen and plant macrofossils; Bright identified the mollusk shells. Good summaries of mollusk habits are in Baker (1928), and Hibbard and Taylor (1960).

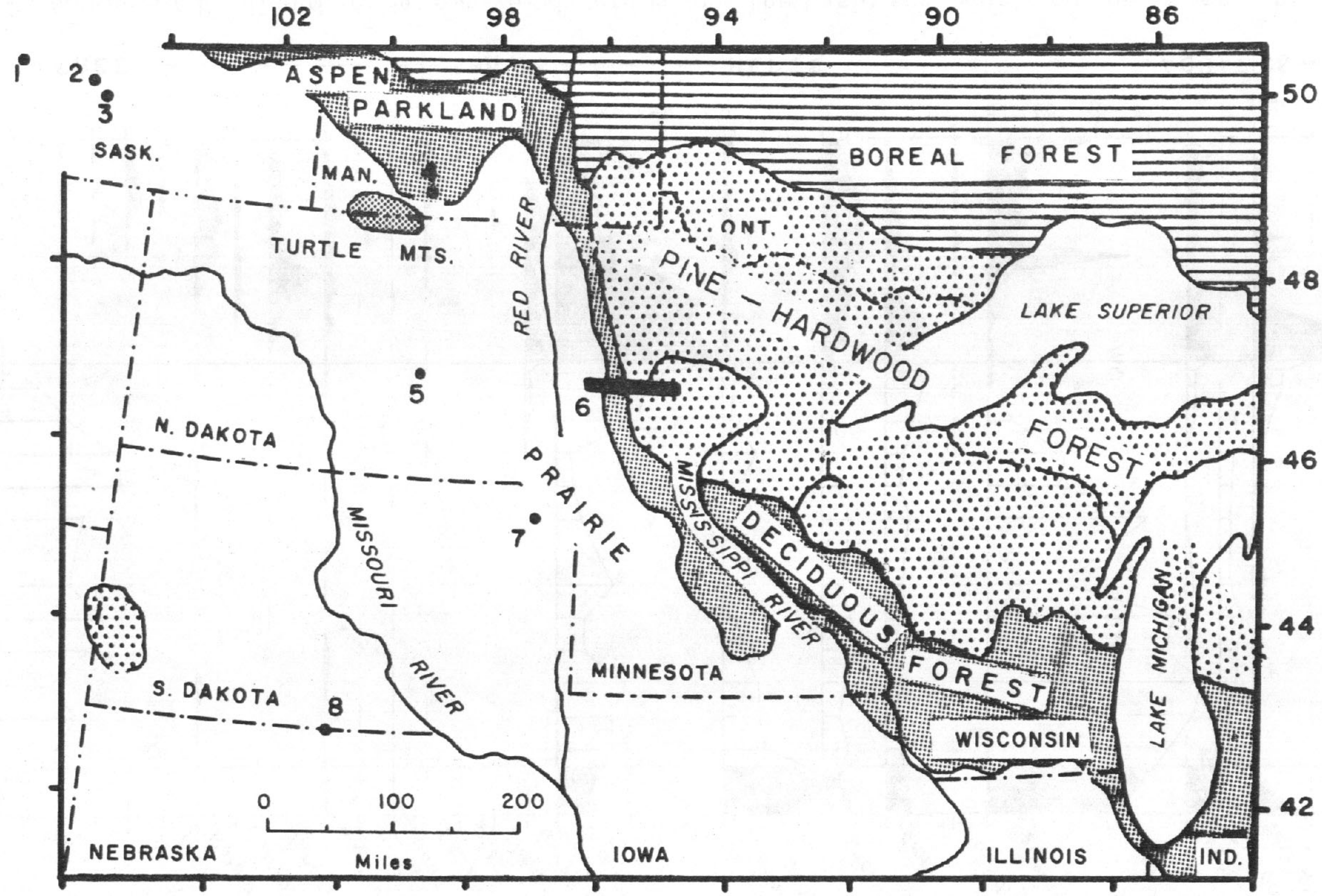


Figure H-1. Map showing major vegetation zones and sites referred to in the text; 1) Herbert (Kupsch, 1960), 2) Hafichuk (Ritchie and deVries, 1964), 3) Crestwynd (Ritchie, 1966), 4) Glenboro (Ritchie, 1966), 5) Woodworth Pond (herein), 6) Itasca transect (McAndrews, 1966), 7) Pickeral (W. A. Watts, unpublished), 8) Rosebud (Watts and Wright, 1966).

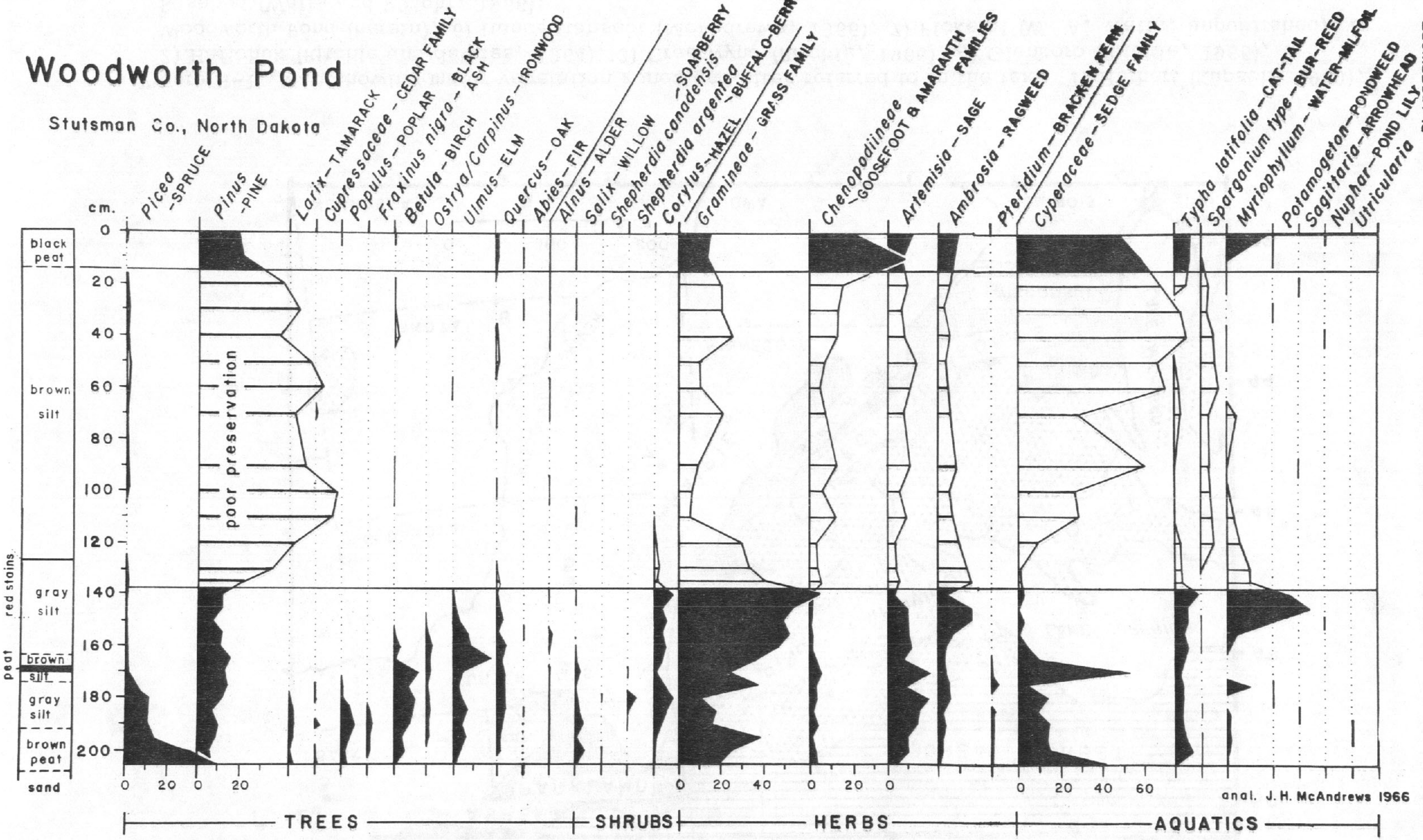


Figure H-2. Abbreviated pollen diagram from Woodworth Pond. The basic sum which includes trees, shrubs, and herbs, is about 150 in the poor-preservation horizons and about 300 elsewhere.

## METHODS

On March 13, 1965, a 220 cm core (diameter 3.8 cm) was collected through water 110 cm deep. The core was complete except for gaps at 70 to 90 cm and 130 to 135 cm. The stratigraphy was noted in the field, particularly the sediment color which changed upon exposure. Pollen analyses were made at 5 or 10 cm intervals. Five or 10 cm core segments were washed through number 30 mesh (0.5 mm) screen to obtain plant macrofossils and mollusk shells.

Figure H-2 shows the sediment stratigraphy together with the main curves of the pollen diagram, table H-1 shows the number of plant macrofossils per 100 cc of sediment, and table H-2 shows the number of mollusk shells per 100 cc of sediment. Pollen and seeds are generally abundant and well-preserved except in the upper brown silt; mollusks are abundant except in the basal peat.

## DISCUSSION

### 210 to 190 cm

Fifteen cm of moderately humified sedge peat overlies sand and gravel outwash. The lower 5 cm of the peat, relative to the upper 10 cm and overlying limnic sediment, contains high values of spruce and sedge pollen, and macrofossils of open-water aquatic plants are scarce. This indicates that the initial community at the site was a shallow-water or damp-soil sedge meadow that was destroyed when the water subsequently deepened. We suggest that deepening of the water or flooding of the soil was due to melting of an underlying iceblock that was buried in the stream sediment, and that melting was caused by a warming climate. The abrupt decrease in spruce pollen implies fewer spruce trees and probably reflects climatic change. It is doubtful if the high values of spruce represent local trees growing on the peat because no spruce needles occurred at this level.

The abrupt decrease of spruce is a stratigraphic marker commonly used to separate late-glacial and postglacial in the upper Midwest. At Pickeral Lake, South Dakota, (fig. H-1) the spruce fall is dated at 10,670 (Watts and Bright, manuscript). Northward at the Herbert site in Saskatchewan the boundary is dated at 10,050 years ago (Kupsch, 1960); at the Hafichuk site the late-glacial is bracketed by dates of 11,650 and 10,270 (Ritchie and deVries, 1964), and at the Crestwynd site the date of an early postglacial horizon is 9,390 (Ritchie, 1966). Thus the late-glacial, postglacial boundary is judged to be 10,700 to 10,000 years ago.

Above 200 cm the peat grades into calcareous gray silt containing fossils of open water aquatics, including naiad (Najas flexilis) and pond lily (Nuphar), species that do not occur today in prairie potholes. The absence of mollusks indicate that the water might have been acidic.

Table H-1.--Plant macrofossils per 100 cc sediment from Woodworth Pond.

| DEPTH (cm) | OPEN WATER | | | | | | | DEEP MARSH | | | SHALLOW | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | Najas flexilis | Nuphar sp. | Potamogeton spp. | Potamogeton pectinatus | Hippuris vulgaris | Myriophyllum sp. | Zanichellia palustris | Typha sp. | Scirpus fluviatilis | Scirpus validus-type | Carex atherodes | Carex spp. (trigonal) | Carex sp. (lenticular) | Eleocharis palustris-type | Alisma sp. |
| 0-10 | | | | | | 0.9 | 0.9 | 0.9 | | | | 7.1 | | 3.5 | 0.9 |
| 10-20 | | | | | | 6.2 | 6.2 | 5.3 | | 1.8 | 4.4 | 33.6 | | 2.6 | |
| 20-30 | | | 0.9 | | 10.6 | | 0.9 | 1.8 | | 21.2 | | 28.3 | | 42.4 | 0.9 |
| 30-40 | | | | | 8.8 | 2.7 | 4.4 | | | 15.9 | | 39.8 | | 69.0 | |
| 40-50 | | | 2.7 | | 12.4 | | | | | 2.7 | | 71.5 | | 28.2 | |
| 50-60 | | | | | .9 | | | | | | | 13.2 | 0.9 | 7.6 | |
| 60-70 | | | | | | | | | | | | | | | |
| 70-90 | NO DATA | | | | | | | | | | | | | | |
| 90-100 | | | | | | | | | | | | | | | |
| 100-110 | | | | | | | | | | 0.9 | | | | | |
| 110-120 | | | | | | | | | | | | | | | |
| 120-130 | | | | | | | | | | | | | | | |
| 130-135 | NO DATA | | | | | | | | | | | | | | |
| 135-140 | | | | | | | | | | | | 1.8 | | | |
| 140-145 | | | | 3.5 | | | | | | 3.5 | | | | | |
| 145-150 | | | | 7.1 | | | | | | 5.3 | | | | | |
| 150-155 | | | | 15.9 | | | | | | 7.1 | | | | | |
| 155-160 | | | | 3.5 | | | | | | | | | | | |
| 160-165 | | | | | | | | | | 3.5 | | | | | |
| 165-170 | | | | | | | | | | 106.1 | | | | | |
| 170-175 | | | | | | | | | | 258.2 | | | | | |
| 175-180 | | | | | | | | | | 1.8 | | | | | |
| 180-185 | | | 1.8 | | | 3.5 | | | 1.8 | 5.3 | | 10.6 | | | |
| 185-190 | | | | | | | | | | 1.8 | | | | | |
| 190-195 | 3.5 | | 1.8 | | | | | | | 3.5 | | 5.3 | | | |
| 195-200 | 1.8 | 1.8 | | | 1.8 | 1.8 | | | 1.8 | 5.3 | | 8.8 | | | |
| 200-205 | | | 10.6 | 3.5 | 1.8 | 1.8 | | | 1.8 | 10.6 | | 122.1 | 3.5 | | |
| 205-210 | | | | | 1.8 | | | | | | | 8.8 | 3.5 | | |

| MARSH | | | | | | WET MEADOW | | MUD FLAT | | | | | UPLAND | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| cf. *Sagittaria* sp. | *Sagittaria cuneata* | *Sium suave* | *Ranunculus* Batrachium-type | *Ranunculus sceleratus* | *Ranunculus* sp. | *Lycopus americanus* | *Mentha arvensis* | *Chenopodium* spp. | Circium cf. *arvense* | *Polygonum lapathifolium* | *Rumex* sp. | *Rumex maritimus* | *Physalis*-type | cf. *Betula* sp. | *Larix laricina* | *Larix laricina* needles | *Picea* sp. needles | Conifer wood | Unknown |
| — | 0.9 | 0.9 | 0.9 | 3.5 | 1.8 | — | — | 49.5 | 0.9 | — | — | 13.2 | — | — | — | — | — | — | 1.8 |
| 3.5 | — | 2.6 | 0.9 | 16.8 | 0.9 | — | — | 226.0 | — | — | — | 66.2 | — | — | — | — | — | — | 11.5 |
| 8.8 | 0.9 | 0.9 | — | 0.9 | — | — | 0.9 | 37.1 | — | — | — | 6.2 | — | — | — | — | — | — | 8.8 |
| 7.6 | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | 9.7 |
| 1.8 | — | — | 0.9 | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | 10.6 |
| 3.5 | — | — | — | — | — | — | — | 1.8 | — | — | 0.9 | — | — | — | — | — | — | — | 0.9 |
| — | — | — | — | — | — | — | — | 0.9 | — | — | — | — | — | — | — | — | — | — | 2.6 |
| — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — |
| — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — |
| — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — |
| — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — |
| — | — | — | — | — | — | — | — | — | — | — | 1.8 | — | — | — | — | — | — | — | 1.8 |
| — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | 1.8 |
| — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | 1.8 |
| — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — |
| — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — |
| — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — |
| — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — |
| — | — | — | — | — | — | — | — | — | — | — | — | — | 10.6 | — | — | — | — | — | — |
| — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — |
| — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | 1.8 | 1.8 | + | 1.8 |
| — | — | — | — | — | — | — | — | — | — | — | — | — | — | — | 5.3 | 10.6 | — | — | — |
| 1.8 | — | — | — | — | — | — | 1.8 | — | — | — | — | — | — | — | 1.8 | 3.5 | 3.5 | + | — |
| — | — | — | — | 1.8 | — | 1.8 | — | — | — | — | — | — | — | 1.8 | — | 1.8 | 5.3 | + | — |
| 3.5 | — | — | — | 8.8 | 3.5 | 1.8 | — | — | — | 1.8 | — | — | — | — | — | — | 10.6 | + | 1.8 |
| — | — | — | — | — | — | 1.8 | 7.1 | — | — | — | — | — | — | — | — | — | — | + | 10.6 |

Table H-2.--Number of mollusk shells per 100 cc of sediment.

| | AQUATIC | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|
| | Prosobranchs | | | | | | | | | |
| DEPTH (cm) | <u>Aplexa</u> <u>hypnorum</u> | <u>Armiger</u> <u>crista</u> | <u>Ferrissia</u> <u>parallela</u> | <u>Fossaria</u> cf. <u>F.</u> <u>obrussa</u> | <u>Gyraulus</u> <u>hirsutus/deflectus</u> | <u>Gyraulus</u> <u>parvus/circumstriatus</u> | <u>Helisoma</u> (immature) | Lymnaeidae (immature) | <u>Physa</u> (immature) | <u>Planorbula</u> <u>armigera</u> |
| 0-10 | .9 | 77.2 | — | 1.8 | 28.9 | 72.7 | X | 21.1 | 4.4 | — |
| 10-20 | | NO DATA | | | | | | NO DATA | | |
| 20-30 | .9 | 568.0 | — | 1.8 | 201.0 | 351.0 | X | 77.2 | 27.2 | 6.1 |
| 30-40 | 3.5 | 231.0 | — | 2.6 | 49.1 | 271.0 | X | 74.5 | 14.1 | 6.1 |
| 40-50 | 7.0 | 288.0 | — | 7.0 | 141.0 | 207.0 | X | 82.4 | 27.2 | 6.1 |
| 50-60 | 7.9 | 353.0 | — | 3.5 | 86.8 | 589.0 | 1.8 | 229.0 | 32.4 | 11.4 |
| 60-70 | 2.6 | 325.0 | — | 1.8 | 61.3 | 586.0 | 2.6 | 256.0 | 31.6 | 17.6 |
| 70-90 | | NO DATA | | | | | | NO DATA | | |
| 90-100 | 4.4 | 335.0 | — | 2.6 | 13.2 | 755.0 | 3.5 | 199.0 | 31.6 | 14.0 |
| 100-110 | | NO DATA | | | | | | NO DATA | | |
| 110-120 | — | 88.5 | — | 7.9 | 37.7 | 970.0 | .9 | 80.6 | 15.8 | 13.2 |
| 120-135 | | NO DATA | | | | | | NO DATA | | |
| 135-140 | — | 52.6 | — | 5.3 | 7.0 | 667.0 | 3.5 | 28.1 | 3.5 | — |
| 140-145 | — | 57.9 | — | — | 17.5 | 442.0 | 1.8 | 8.8 | — | — |
| 145-150 | — | 94.7 | — | 3.5 | 21.1 | 1019.0 | 15.8 | 21.1 | 14.0 | — |
| 150-155 | — | 93.0 | — | — | 19.3 | 2030.0 | 15.8 | 24.6 | 5.3 | — |
| 155-160 | — | 33.4 | — | — | 7.0 | 985.0 | 10.5 | 5.3 | 7.0 | — |
| 160-165 | — | — | — | — | — | 40.4 | — | — | — | — |
| 165-170 | — | — | — | — | 7.0 | 200.0 | 3.5 | — | 1.8 | — |
| 170-175 | — | 29.8 | 24.6 | — | 5.3 | 304.0 | 12.3 | — | 1.8 | — |
| 175-180 | — | — | — | — | — | 153.0 | 5.3 | — | — | — |
| 180-185 | — | — | — | — | — | 279.0 | — | — | 1.8 | — |
| 185-190 | — | — | — | — | — | 5.3 | — | — | — | — |
| 190-210 | NO MOLLUSK SHELLS | — | — | — | — | — | — | — | — | — |

| SNAILS | | | | | | AQUATIC CLAMS | | | LAND SNAILS | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Pulmonates | | | | | | | | | | | | | |
| *Promenetus exacuous* | *Promenetus umbilicatellus* | *Stagnicola caperata* | *Stagnicola exilis* | *Stagnicola reflexa* | *Valvata lewisi* | *Pisidium casertanum* | *Pisidium* cf. *P. compressum* | *Pisidium* undet. | *Gastrocopta holzingeri* | *Hawaiia minuscula* | Pupillidae indet. | cf. *Succinea* | *Vertigo ovata* |
| 18.4 | 1.8 | 6.2 | .9 | .9 | — | 5.3 | | | | | | 8.8 | |
| 50.8 | | 22.8 | .9 | 3.5 | — | 12.3 | .9 | 2.6 | | | | 14.9 | 2.6 |
| 21.9 | | .9 | 3.5 | 4.4 | — | 1.8 | | | | | | 14.9 | |
| 36.0 | | 11.4 | | 2.6 | — | 3.5 | | 7.0 | | .9 | 2.6 | 21.2 | |
| 45.6 | | 19.3 | 1.8 | 5.3 | — | 16.7 | | 2.6 | | | | 27.2 | .9 |
| 99.2 | | 7.0 | .9 | 22.8 | — | 18.4 | | 12.3 | .9 | | | 24.6 | 2.6 |
| 269.0 | | 8.8 | | 4.4 | — | 101.0 | | 9.7 | | | | 31.6 | 1.8 |
| 133.0 | | .9 | 7.0 | 11.4 | .9 | 43.0 | | 11.5 | | | | 20.2 | |
| 73.7 | | | 8.8 | | 1.8 | 8.8 | | | | | | | |
| 31.6 | | | 1.8 | | 5.3 | 17.6 | | | | | | | |
| 102.0 | | | 1.8 | 1.8 | 66.6 | 125.0 | | 14.0 | | | | | 1.8 |
| 100.0 | | | 1.8 | 1.8 | 170.0 | 193.0 | | 35.1 | | | | | |
| 5.3 | | | | | 54.4 | 140.0 | | 40.4 | | | | | |
| | | | | | 8.8 | 21.1 | | 3.5 | | | | | |
| | | | | | — | 164.0 | | 49.2 | | | | | |
| 10.5 | | | | | — | 35.1 | | 19.3 | | | | | |
| | | | | | — | 61.4 | | 8.8 | | | | | |
| 5.3 | | | | | | 56.2 | | 36.9 | | | | | |
| | | | | | | | | 1.8 | | | | | |
| NO MOLLUSK SHELLS | | | | | — | | | | | | | | |

X = fragments

190 to 175 cm

High pollen values of grass (Gramineae), chenopods (Chenopodiineae), sage (*Artemisia*) and ragweed (*Ambrosia*), indicate that the upland vegetation was largely prairie with stands of coniferous and deciduous trees. The presence of spruce, tamarack (*Larix*) and birch (*Betula*) is confirmed by macrofossils, whereas poplar (*Populus*), black ash (*Fraxinus nigra*), ironwood (*Ostrya-Carpinus*), elm (*Ulmus*), oak (*Quercus*) and hazel (*Corylus*) are recorded only by their pollen. Pine (*Pinus*), a heavy pollen producer, was probably absent as indicated by its low values; the nearest source of pine pollen at this time was probably the Sheyenne delta of Lake Agassiz in eastern North Dakota (McAndrews, unpublished).

The abundance of the mollusk *Gyraulus* in this interval indicates that aquatic vegetation at the coring site was rather thick, but with open spaces that were inhabited by fingernail clams. The water was quiet, less than 2 or 3 m deep, with pH greater than 7, and with moderate alkalinity.

175 to 165 cm

The brown silt is largely devoid of plant macrofossils, but the black, highly humified peat at 170 cm contains abundant bulrush seeds. The silt, according to McAndrews, probably represents temporary low-water levels in the pothole and a time when plant macrofossils were largely destroyed by oxidation. The peat layer represents a brief period of marsh vegetation. The mollusk suite of this horizon is essentially the same as in the underlying one, except for a few more species. Aquatic vegetation at the coring site was probably rather thick, but with numerous openings for *Pisidium* to inhabit. The mollusks indicate that the water was quiet, moderately hard, of moderate alkalinity, and with a pH greater than 7. The presence of *Ferrissia paralella* in the 170 to 175 cm interval is perhaps indicative of permanent water (Tuthill and others, 1964, p. 359). The marked increase of *Pisidium casertanum*, a highly adaptable and drought tolerant sphaeriid, in the 165 to 170 cm interval is consistent with but not proof of the interpretation of a temporary drought indicated by accumulation of peat.

The pollen spectra indicate that spruce, tamarack, poplar and black ash disappear at this time, but that birch, elm, ironwood, oak and hazel persisted into the time represented by the overlying gray silt. These deciduous forest trees were widespread in early postglacial prairie, as shown in pollen diagrams from the Itasca transect and from Pickeral Lake, but they, except for oak, are quite absent at the Rosebud site in the Nebraska sandhills (Watts and Wright, 1966).

165 to 125 cm

This gray silt horizon has red iron-oxide stains in the upper 20 cm. The mollusks are more helpful in reconstructing the limnic conditions than are the relatively few plant macrofossils. In the interval 165 to 140 cm Valvata lewisi indicates clear, permanent water. The pond at the coring site was probably less than 2 or 3 m deep, the water being of moderate alkalinity, moderately hard, and with a pH greater than 7. The mollusks indicate dense aquatic, vegetation with open areas. However, the virtual lack of seeds of aquatic plants indicates either that the plants were not dense or that the pond was not permanent. The pond was occasionally dried out, and seeds in the surface sediment were destroyed by oxidation. The pollen diagram indicates water-milfoil was abundant, but no seeds are present.

In the upper red-stained gray silt the pollen diagram shows essentially the disappearance of deciduous trees. Here the abrupt increase in pine pollen values is related to increasingly poor pollen preservation. The seeming abundance of the relatively easily fossilized and easily identifiable pine pollen is an artifact and does not indicate the presence of local pine trees. That the pine increase can be explained by the reworking of older sediment is doubtful for the older horizons do not contain abundant pine pollen.

125 to 15 cm

The period represented by brown silt is difficult to interpret because the pollen is poorly preserved, and the sediment below 60 cm is almost devoid of plant macrofossils. In the interval 140 to 60 cm the marked increase of the three species of Stagnicola, the appearance of Aplexa and Planorbula, and the increase of Gyraulus hirsutus/deflectus indicate unstable water levels. The presence of Fossaria cf. F. obrussa, a common littoral or marsh snail, along with cf. Succinea, which is commonly found in moist to dry littoral situations, and other land snails indicates that the shore was closer to the coring site than previously. The suite of aquatic snails is one that now characterizes the temporary pools, ponds, and roadside ditches. Therefore, the water was no doubt temporary, usually drying up each fall, and rarely more than 1 m deep at the coring site, harder than previously, with moderate to high alkalinity, and with pH about 8. The appearance of Aplexa and Planorbula in the upper half of the unit point to more stagnant conditions than previously. Marshes were probably more common than before.

The suite of mollusk shells recovered from above 60 cm is indicative of a water body similar to the one of the previous period, but perhaps with fewer periods of desiccation. Above 60 cm seeds become increasingly more abundant and well-preserved further supporting the view of more permanent water.

15 to 0 cm

The upper 15 cm of black detrital peat contains well-preserved pollen and a wide variety of seeds of aquatic plants. The pine pollen values are much lower in contrast to the values in the earlier horizons where pollen preservation was poor. The better pollen preservation may be related to recently changed hydrologic conditions. The pond outlet is through a culvert in a railroad embankment built in the early 1900's; this may have raised the outlet level. Gravel mining around the pond has resulted in the removal of soils and vegetation. These alterations could have maintained the pong permanently and not allowed drying and subsequent oxidation of the fossil pollen.

## CONCLUSIONS

We heed Ritchie (1966) and do not attach great climatic significance to the successions of upland vegetation which may respond primarily to soil development and the availability of plant species. However, we believe that geochemical and paleontological studies of prairie-pothole sediments that assess water depth and water quality may ultimately provide a sound basis for the reconstruction of late-glacial and postglacial climatic history.

Because C-14 dates are lacking, we can only estimate the antiquity of the various periods represented by fossils and sediments. Our conclusions are summarized in table H-3.

## ACKNOWLEDGMENTS

We wish to thank Ivan Hernandex for aid in field work. Robert E. Stewart, Sr., critically read the manuscript and kindly supplied unpublished information on the ecology of aquatic plants. The use of the facilities of Jamestown College, and the Itasca Forestry and Biology Station and the Limnological Research Center of the University of Minnesota, is greatly appreciated.

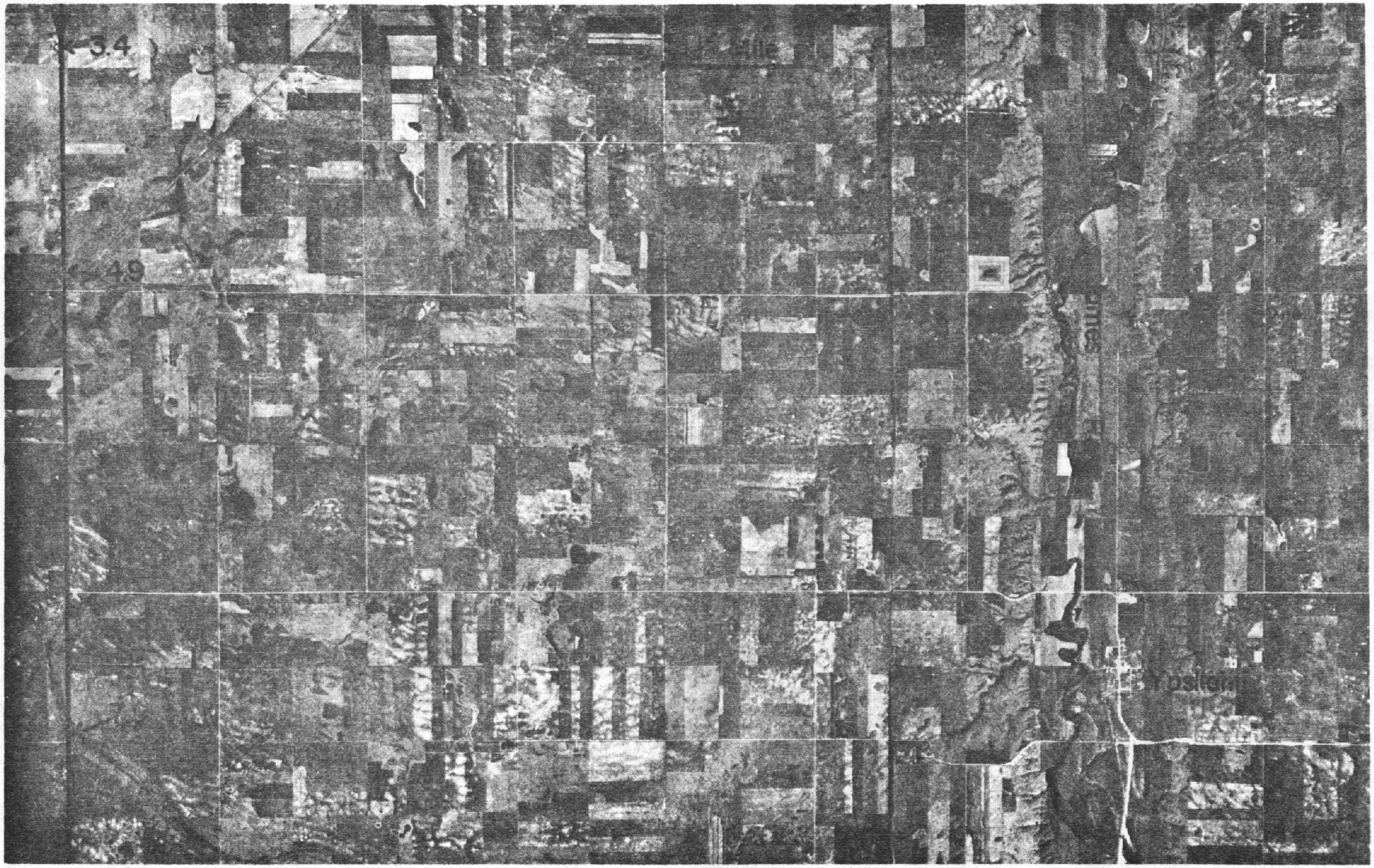


Figure R-8. Air photo of typical section of the Drift Prairie showing "washboard moraines" and James River spillway, with fieldtrip route and mileages indicated. Area is east-central part of fig. R-7a. (Army Map Service BE M1 118 and 119, 21 July 52.)

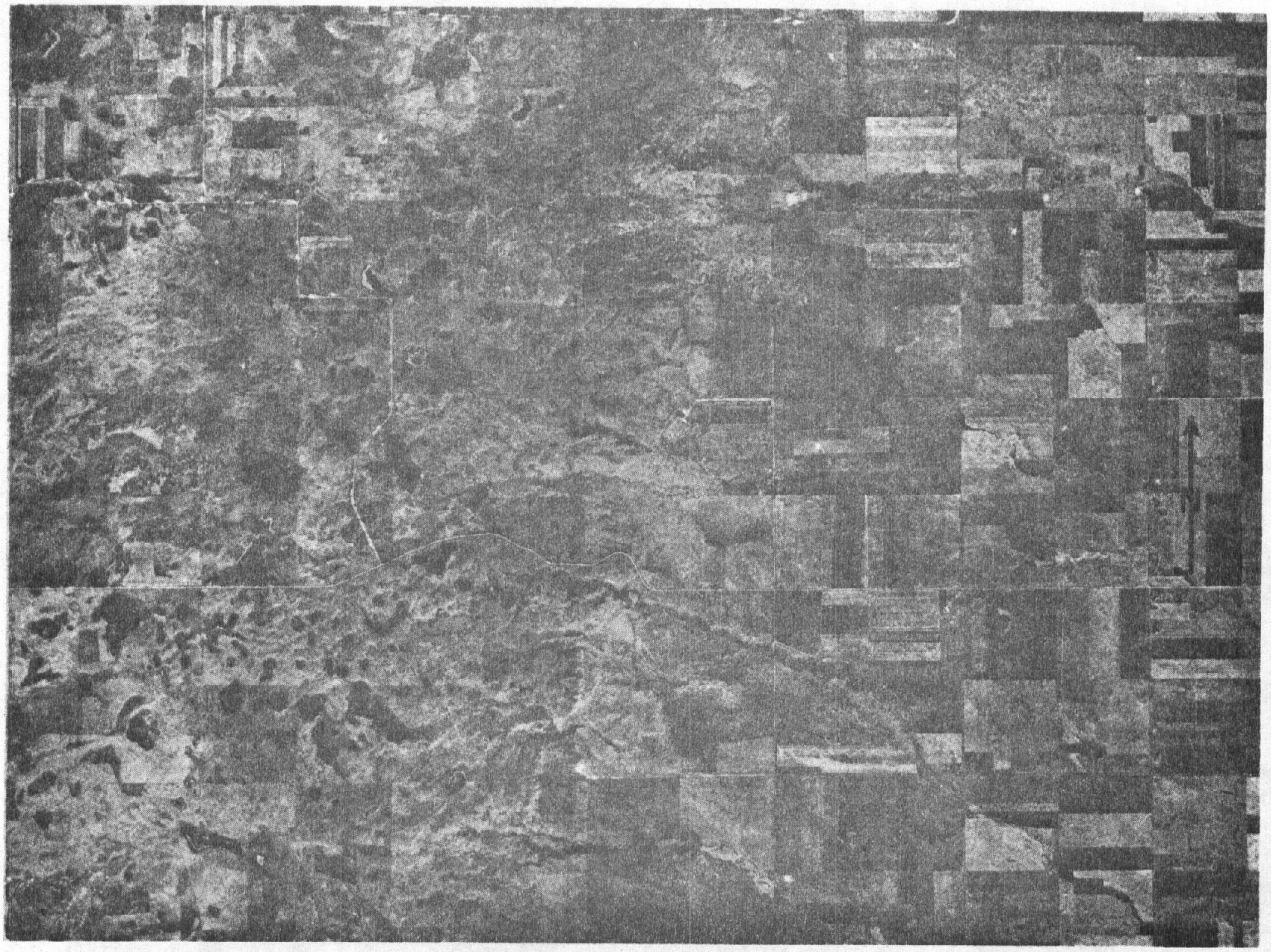

Figure R-9. Air photo of Missouri Escarpment, Dickey County, 40 miles south of stop 1. Gullies on escarpment, dead-ice moraine on Missouri Coteau (left), and ground moraine on Drift Prairie (right). (Army Map Service BE M19 2193 21 Aug. 52.)

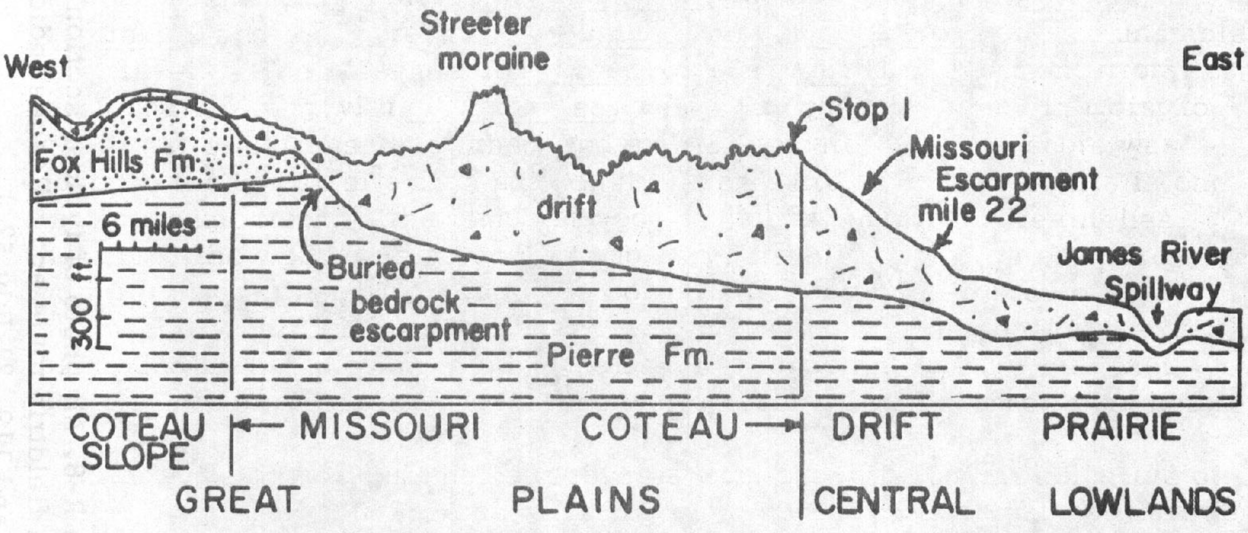


Figure R-10. Generalized east-west cross-section through Missouri Coteau near stop 1. Vertical exaggeration 100 times.

Table H-3. Summary of the history of Woodworth Pond. Read upward from base.

| Depth (cm) | Sediment type | Estimated age (years ago) | Conditions in the pothole | Upland vegetation | Climate |
|---|---|---|---|---|---|
| 0-15 | Black, poorly humified, limnic peat. | 0-50 | Permanent pond with open-water, and with marsh and meadow vegetation around margin. | Prairie disturbed by gravel mining and agriculture. | As now. |
| 15-60 | Brown silt, with plant detritus increasing upward | 50-3,000 or 4,000 | As below but with decreasing frequency of desiccation. Open water with marginal marsh vegetation. | Prairie. | Decreasing aridity. |
| 60-125 | Brown silt. | 3,000 or 4,000-8,000 | Temporary water usually $< 1$ m deep, moderate to high alkalinity, moderately hard, and pH ca.8. Poor conditions for plant fossil preservation. | Prairie; disappearance of Corylus. | Arid. |
| 125-165 | Gray silt, with red stains in upper 20 cm. | 8,000-9,000 | Initially permanent water probably 2 or 3 m deep, moderate alkalinity, moderately hard and with a pH $>7$. Interval 145-125 indicates a transitional period to conditions represented by overlying horizon. | Prairie; disappearance of trees. | Increasing aridity. |
| 165-175 | Brown silt, with 2 cm black highly humified peat at 170 cm. | 9,000 | Shallower. (See discussion). Water intermittant, moderately hard, moderate alkalinity, and with a pH $>7$. Marsh dominated by Scirpus but with some open water. | Prairie; disappearance of Picea, Larix, Populus and Fraxinus. | Temporary drought. |
| 175-205 | Terrestrial peat grading upward into lacustrine, calcareous gray silt. | 9,000-10,500 | Formation of pothole by melting of buried ice. Water 2 or 3 m deep with moderate alkalinity, initially perhaps acid but later with pH $>7$. Open water & marsh vegetation. | Prairie with stands of Picea, Larix, Populus Betula, Ostrya, Ulmus. n. Fraxinus, Quercus and Corylus. | Warmer. |
| 205-210 | Brown, moderately humified, terrestrial peat. | 10,500-11,000 | Sedge meadow on outwash overlying buried ice. | Open, boreal-type Picea forest. | Cool. |

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30-I

# DUMP RIDGES AND COLLAPSED SUB-ICE CHANNELS IN WARD COUNTY, NORTH DAKOTA[1]

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A dump ridge is one of several types of disintegration ridges that are relatively common on the Coteau du Missouri in Ward County. It generally consists of boulders and coarse gravel that have an esker-like form (fig. I-1). Dump ridges were deposited at the edges of stagnant ice fields and, like end moraines, mark the positions of the ice margins. The ridges were formed by dumping of outwash at the margin of stagnant ice. The coarse material formed a ridge whereas finer material was carried downslope forming adjacent outwash deposits. A typical dump ridge is 20 to 45 feet wide at the base, 5 to 20 feet in height, and as much as a mile in length.

In at least two places in Ward County, a series of parallel closely spaced dump ridges indicates successive retreats of the margin of the stagnant ice field (fig. I-1). They represent miniature recessional moraines.

Dump ridges differ from eskers which are more sinuous, consist of finer grained material, and trend roughly perpendicular to the ice front. Dump ridges differ from cravasse fillings which are relatively straight and are active ice features.

Commonly associated with dump ridges are collapsed sub-ice channels which trend nearly perpendicular to the former ice margin (fig. I-1). Sub-ice channels are former water courses that flowed beneath or within the glacial ice. Apparently there was little deposition of glaciofluvial material in the channels. As the ice fields melted, superglacial drift filled the channels, or the channels were partly buried by outwash.

[1]Publication authorized by the Director, U. S. Geological Survey.

The topographic sag of collapsed sub-ice channels can be traced for short distances from the till-outwash boundary into the hummocky dead-ice moraine. They may be traced for as much as 2 miles from the till-outwash boundary into adjacent collapsed outwash plains. They range considerably in size, but are always more prominent in outwash deposits than in dead-ice moraine. They generally contain chains of small lakes, which in the lower parts of the channels are nearly always saline.

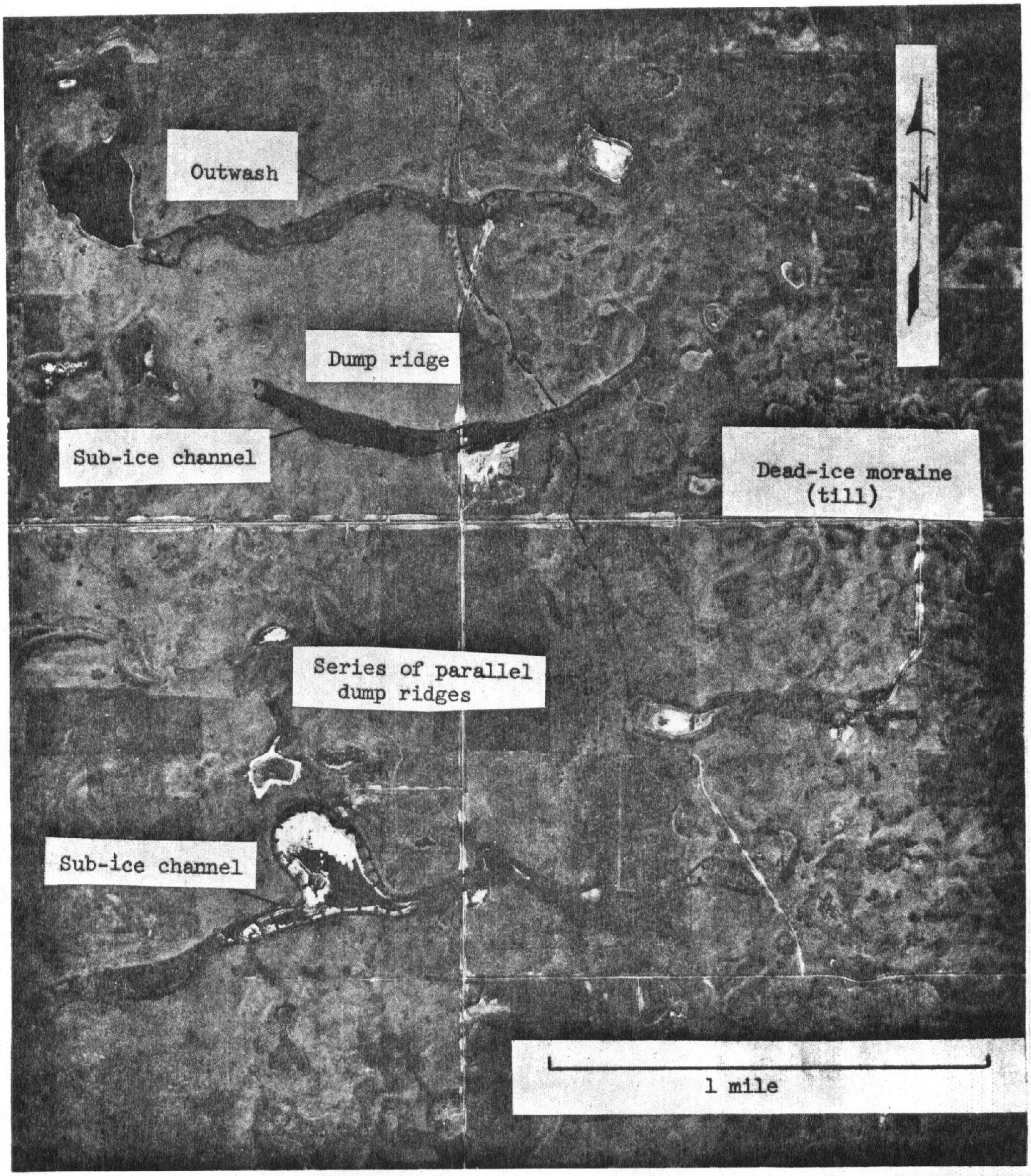


Figure I-1. Dump ridges and sub-ice channels on the Coteau du Missouri (secs. 17, 18, 19, and 20, T. 152 N., R. 84 W.) in Ward County, North Dakota.

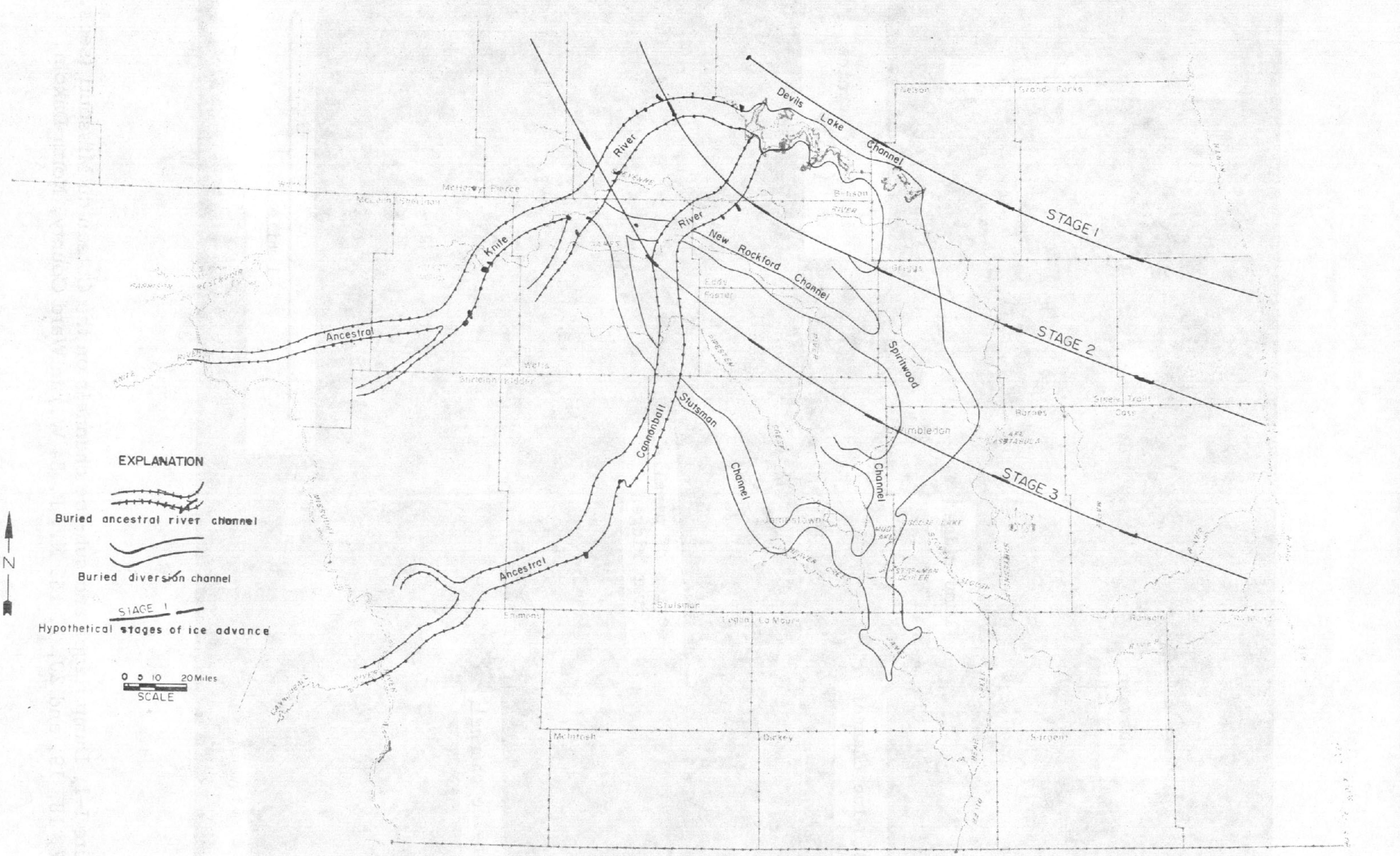


Figure J-1. Generalized map showing elements of preglacial and inferred proglacial drainage courses of North Dakota.

# PLEISTOCENE DIVERSION OF STREAMS IN CENTRAL NORTH DAKOTA[1]

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## INTRODUCTION

The preglacial drainage of North Dakota bore little resemblance to the drainage that exists today. Prior to glaciation, streams traversed the State following patterns that were established in late Tertiary time. These streams flowed northeastward across the State and emptied into Hudson Bay (Alden, 1932, pl. 1; Benson, 1952, p. 165-175; Flint, 1955, pl. 7).

The preglacial topography in western North Dakota contrasted sharply with the topography that prevailed in the eastern part of the State. Rolling hills, buttes, and badland topography had been carved on the Tertiary deposits in the west, whereas a more subdued topography had been carved on the Cretaceous shale to the east. The bedrock surface sloped toward the northeast at about 15 feet per mile.

According to Witkind (1959, p. 55), slight uplift of eastern Montana and western North Dakota at the end of the Pliocene resulted in rejuvenation of the streams traversing the area. These streams include the Yellowstone, Little Missouri, Knife, and Cannonball, all of which now drain into the Missouri River. Prior to glaciation, however, the Missouri River emptied into Hudson Bay and the smaller Missouri tributaries extended across much of the State (Lemke and Colton, 1958, fig. 2). All or part of the channels of the Yellowstone, Little Missouri, Knife, and Cannonball have since been obliterated by glaciation north or east of the present course of the Missouri River. The ancestral courses of the Yellowstone and Little Missouri Rivers have been described by several writers (Bauer, 1915;

[1]Publication authorized by the Director, U. S. Geological Survey.

Alden, 1932, p. 129; Howard, 1960, p. 66-70), and T. F. Freers, (written communication). In general, these two streams were restricted to western North Dakota and southeastern Saskatchewan.

## KNIFE RIVER

The preglacial extent of the Knife River is not well known. The river and its tributaries headed in southwestern North Dakota and extended northeastward into Sheridan County (fig. J-1) where the channel has been obscured by ground moraine. The valley of the ancestral Knife River traverses McLean and Sheridan Counties and northwestern Wells County. Test drilling by the Bureau of Reclamation in an area north of Lincoln Valley indicates the trend of the ancestral Knife to have been northeast. The channel fill is more than a mile wide and contains fine-grained sand and silt as much as 240 feet thick (P. D. Wold, oral communication, 1966).

In Wells County the channel, which is about 2 miles wide, was eroded more than 300 feet into the bedrock. It probably extends north and east across Benson County.

## CANNONBALL RIVER

Kume and Hansen (1965, p. 64, 66) reported that the ancestral Cannonball extended eastward into Emmons and Burleigh Counties from the present confluence of the Cannonball River with the Missouri River (fig. J-1). The valley was eroded more than 250 feet deep and 2 1/2 to 3 miles wide. In Kidder County, the ancestral Cannonball valley was eroded as much as 300 feet deep and developed parallel terraces on each side of the channel (Rau and others, 1962, p. 37). The gradient of the ancestral Cannonball in Kidder County was about 2 feet per mile. In Wells County the channel is 2 1/2 to 3 miles wide and 300 feet deep.

The ancestral Cannonball extended north and joined the ancestral Knife River west of Devils Lake in Benson County. The lack of testhole data precludes tracing the channel farther. The trunk stream probably extended north and eventually flowed into Hudson Bay.

The valleys and drainage systems of both the ancestral Knife and Cannonball Rivers are similar. Both trenches have northeastward gradients, and the numerous tributaries joining the main channels display typical dendritic patterns. In Kidder, Stutsman, and Wells Counties, the deepest part of the Cannonball channel contains fine-grained clastics that are characteristic of alluvium. At numerous locations, glacial erosion has scoured the alluvium and the channels are filled with glacial till.

## GLACIAL HISTORY

During an early glacial advance, the combined Knife and Cannonball Rivers were blocked and diverted through the Devils Lake diversion channel (fig. J-1, Stage I). This diversion channel, which underlies the Devils Lake chain of lakes, has an average width exceeding 2 miles and a depth in excess of 250 feet (Paulson and Akin, 1964, fig. 9). The diversion channel system is known to extend east-southeastward into the vicinity of Stump Lake (fig. J-1). Little subsurface data are available in southwest Nelson County, and the size and location of the system is problematical in that area.

A large north-south oriented valley cut into the bedrock was delineated in eastern Stutsman and western Barnes Counties. Huxel (1961, p. 179) called this the Spiritwood valley. The valley was eroded more than 250 feet into the Pierre Shale in Barnes County. As defined by the 1,300-foot contour, it has an average width of more than 6 miles (Kelly, 1964, p. 161). Test-hole data indicate that the gradient is approximately 3 feet per mile toward the south. The bottom of the valley contains a thick accumulation of outwash, which locally exceeds 150 feet in thickness. Several tributaries join the Spiritwood valley from the west and northwest (Kelly, 1964, fig. 1).

Initially the Spiritwood channel was believed to be a continuation of the Cheyenne River, which reportedly flowed northward from South Dakota (Huxel, 1961, p. 180). However, detailed test drilling has produced sufficient evidence to indicate that the Spiritwood valley drained southward. The width of the valley is compatible with those of the ancestral Knife and Cannonball, assuming that additional water would have been contributed by the glacier that diverted the streams. The angles at which tributaries join the Spiritwood, as well as the gradient, indicate that the parent stream flowed southward. Only glacial outwash was penetrated by test drilling in the Spiritwood channel. The Knife and Cannonball Rivers probably had a very low competence and were capable of transporting only fine sand, silt, and clay such as that found on bedrock in the bottom of the channels in Wells and Sheridan Counties. However, at the point where diversion occurred, additional water and large quantities of outwash would have been picked up. This would explain the thick deposits of outwash in the Spiritwood channel. This is also supported by the fact that the outwash deposits in the Spiritwood valley thicken northward (Kelly, 1964, fig. 2).

As the glacial front advanced, the Devils Lake diversion channel was overridden and the Knife and Cannonball Rivers were diverted southeastward through a deep, narrow trench that crosses Wells, Eddy, and Foster Counties and joins the Spiritwood valley in southwestern Griggs County (fig. J-1, Stage II). This diversion trench was called the New Rockford valley by Trapp (personal communication). The New Rockford valley was eroded more than 200 feet below the adjacent bedrock surface and the trench has an average width in excess of 4 miles (Bluemle, 1965, pl. 2). As the diverted Knife and Cannonball Rivers were cutting the New Rockford diversion channel, water from the glacier continued to erode and widen the Spiritwood. Consequently, the Spiritwood trench is nearly 12 miles wide in Griggs and northern Barnes Counties.

Further advance of the glacial front (fig. J-1, Stage III) resulted in diversion of the Knife and Cannonball Rivers across Wells and Stutsman Counties to a more southerly part of the valley. The bedrock valley that traverses Stutsman and Wells Counties is herein called the Stutsman diversion channel. This diversion channel is similar in magnitude to the ancestral Cannonball River system in Kidder County (Winters, 1963, pl. 2). During diversion Stage III, the Knife River probably joined the ancestral Souris River drainage, as suggested by Lemke and Colton (1958, p. 43), and flowed northward.

Sufficient subsurface data are not available to determine the southern extension of the Spiritwood system in LaMoure and Dickey Counties. The Spiritwood probably emptied into either the ancestral Cheyenne River drainage system in South Dakota or the ancestral Red River drainage system of North Dakota.

Additional test drilling is necessary to further delimit the diversion channels in central North Dakota. Also, it would be desirable to accurately determine the geomorphic characteristics of these proglacial channels in order to better understand the early glacial history of the area.

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30-K

# MULTIPLE DRIFT SHEETS IN SOUTHWESTERN WARD COUNTY, NORTH DAKOTA[1]

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## INTRODUCTION

During the course of reconnaissance geologic mapping and test drilling in southwestern Ward County in 1965, the presence of at least two glacial drift sheets became evident. Exposed and buried oxidized zones, consisting of accumulations of iron oxide in the ferric state, permit correlation of the deposits. It is the purpose of this report to briefly describe the drift sheets and speculate on their age.

The information given in this report was obtained during a study of the geology and ground-water resources of Renville and Ward Counties, a cooperative program between the U. S. Geological Survey, the North Dakota State Water Commission, the North Dakota Geological Survey, and the county commissioners. The classification and nomenclature of the rock units discussed in this report conform to the usage of the North Dakota Geological Survey.

### Location and extent of area

Ward County, in north-central North Dakota, is included in both the Central Lowland and Great Plains physiographic provinces (fig. K-1). The southwestern part of the county lies wholly in the rolling, high to low relief dead-ice or stagnation morainal topography characteristic of the Coteau du Missouri (Great Plains). The area covered by this report includes about 1,000 square miles in the south and southwestern parts of the county.

[1]Publication authorized by the Director, U. S. Geological Survey.

## Previous work

The area was mapped previously by Andrews (1939), but the lack of aerial photographic coverage seriously hampered his work. The area also was mapped by Colton and others (1963). Their work was based entirely on the interpretation of high altitude aerial photographs.

# GEOLOGY OF THE AREA

The Coteau du Missouri is a large plateau-like highland that trends northwest-southeast through the state. The east-facing escarpment is abrupt in most areas and local relief, in places, exceeds 300 feet. The Coteau du Missouri is underlain by a dissected bedrock "high," and is mantled with drift that may exceed 300 feet in thickness. Locally, bedrock of the Fort Union Group (Dorf, 1940) of Paleocene age crops out.

Records of wells and test holes in Ward County suggest that the drift that forms the present escarpment of the Coteau may be 200 to 300 feet thick. This belt of unusually thick drift appears to range between 1 and 3 miles in width. It probably resulted from a piling up of drift as the glacial ice overrode the bedrock escarpment.

Much of the Coteau area has been referred to as the "Altamont moraine," "terminal moraine," or "Max moraine." However, the glacial deposits that cover it do not represent a single end moraine but probably several end or recessional moraines, some ground moraine, and a considerable amount of dead-ice moraine. There appear to be substantial age differences between some of the glacial deposits.

The southwestern part of the mapped area is atypical of much of the Coteau because here the glacial cover consists predominantly of linear end moraines and low relief ground moraine. Adjacent areas to the north and east are represented by typical dead-ice moraine characterized by an abundance of ice-disintegration features. The form and origin of ice-disintegration features are described by Gravenor and Kupsch (1959).

## Fort Union Group

The Fort Union Group is oxidized as much as 10 feet in outcrops in southwestern Ward County and as much as 40 feet beneath the drift, probably as a result of a pre-Wisconsin warm period.

## Glacial deposits

Older drift.--The oldest drift exposed in the county is a thin sheet of ground moraine on the distal (southeast) side of the Blue Mountain end moraine (fig. K-1). The entire thickness of older drift appears to be oxidized,

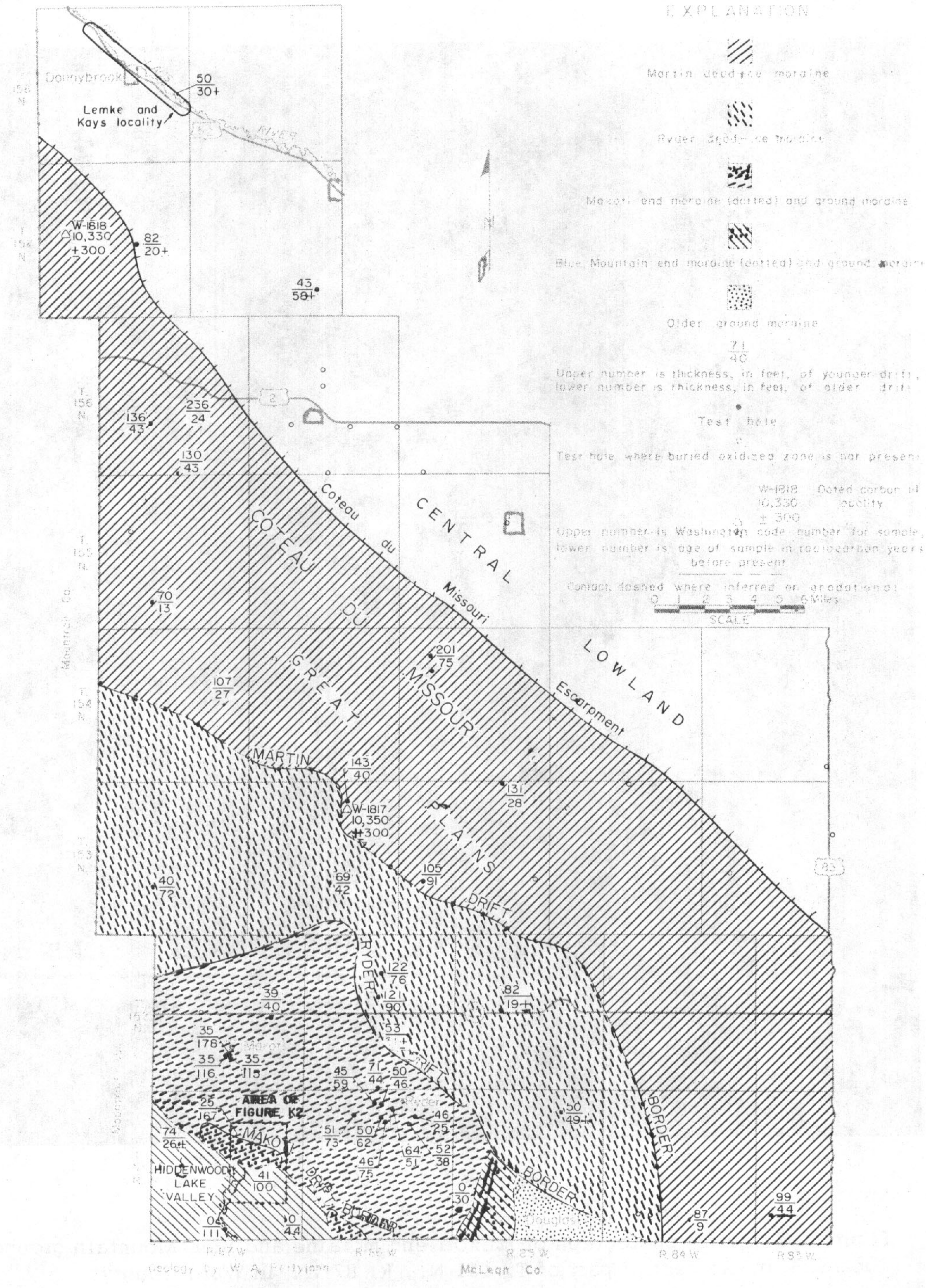


Figure K-1. Generalized landform map showing location of glacial drift boundaries and test holes in southwestern Ward County.

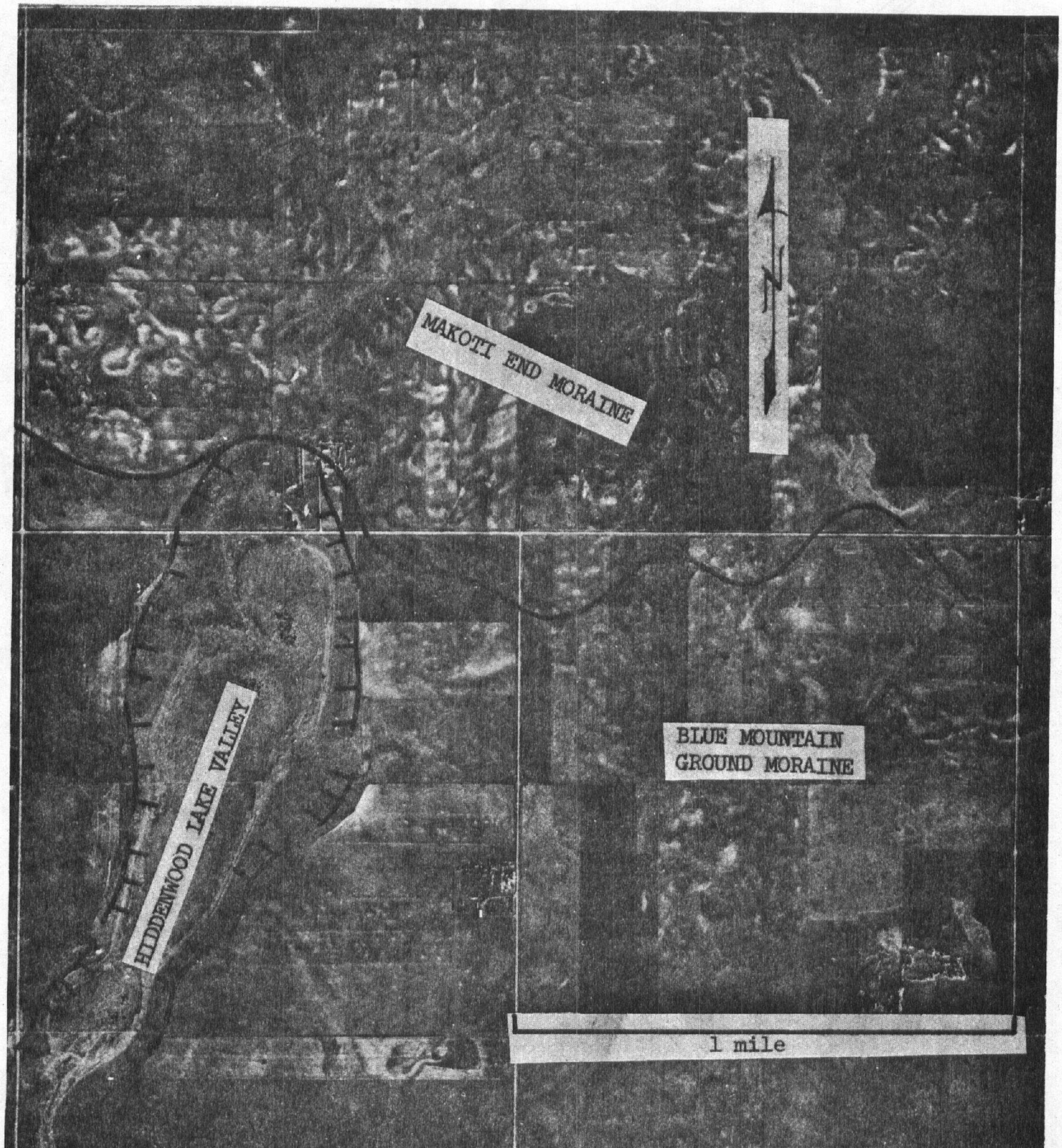


Figure K-2. Aerial photograph of Makoti end moraine and Blue Mountain ground moraine in the central part of T. 151 N., R. 87 W. in Ward County.

and the drift consists of very silty and very sandy clay till with an abundance of lignite chips. This deposit has not been dated and little is known of its relationship to other deposits farther south. It might represent a pre-Wisconsin drift, but data are too scarce for speculation at this time.

Blue Mountain drift sheet.--Adjacent to the older thin oxidized till is the Blue Mountain end moraine (fig. K-1). This drift sheet consists predominantly of gravelly and very sandy clay till. Lignite chips are abundant. The Blue Mountain end moraine has not been dated and it may be too old to be dated by radiocarbon methods.

The blue Mountain end moraine in the mapped area is an arcuate ridge generally ranging from 1 to 2 miles in width. It rises as much as 60 feet above the surrounding low-relief ground moraine. The ridge is covered by younger glacial deposits north of the border of the Ryder dead-ice moraine (fig. K-1).

The Blue Mountain ground moraine, where exposed, is characterized by low relief and poorly integrated drainage. In the extreme southwestern part of the county where these deposits are exposed at the surface, the entire thickness (about 44 feet) has been oxidized to various shades of red, yellow, and brown. North and east, these deposits are covered by younger drift. The Blue Mountain ground moraine appears to have a maximum thickness of about 178 feet where it fills the Hiddenwood Lake valley near Makoti (fig. K-1). Throughout most of the Coteau area in Ward County, however, these deposits average about 70 feet in thickness. On the basis of test-hole data, the depth of oxidation in the buried Blue Mountain moraine in these areas ranges from 13 to 36 feet; it averages 26 feet.

The Blue Mountain drift sheet extends largely beneath younger drift, over a wide area. The buried oxidized zone has been traced in test holes for nearly 40 miles north and more than 30 miles east of the Blue Mountain end moraine in Ward County. It is most commonly found in test holes drilled on the Coteau and, at present, has been seen only in four test holes and one exposure in the ground moraine plain (Central Lowland) north of the Coteau. Presumably, the oxidized material in the Central Lowland was, in most places, incorporated into glacial ice during subsequent advances.

Lemke and Kaye (1958) described an exposure of two drifts near Donnybrook (fig. K-1). In this area, the older drift consists of till and associated glaciofluvial deposits that are at least 30 feet thick. Although the drifts are lithologically similar, the older material shows a greater degree of oxidation. Separating the drifts is a boulder pavement, generally less than 3 feet thick, that grades laterally into a lime-rich zone. The younger drift is at least 50 feet thick. Lemke and Kaye suggested that the two drifts belong to different stades of the Wisconsin glaciation and that the older could be as old as Iowan.

In a test hole (157-87-22aaa) about 5 1/2 miles south of the Donnybrook exposure, a similar stratigraphic sequence was noticed. At this site, 82 feet of till overlies at least 20 feet of oxidized silt and sand.

On the basis of similar relationships, indicated by the aforementioned test hole and other test holes drilled between Donnybrook and the exposed Blue Mountain ground moraine as well as throughout much of the Coteau area, one can conclude that the older drift near Donnybrook is an equivalent of the Blue Mountain end moraine.

It is interesting to note that ground water from the Blue Mountain and older deposits has undesirable qualities for most purposes. Generally the water has a high specific conductance, and the sulfate concentration exceeds 1,000 parts per million. Presumably, the poor quality water is a product of recharge from the Fort Union Group, base exchange, and leaching of the abundant lignite fragments in the drift.

### Younger glacial deposits

Makoti drift sheet.--The Makoti end moraine is well developed 4 1/2 miles south of the village of Makoti where it blocks the Hiddenwood Lake valley (fig. K-2). The distal margin of the Makoti drift sheet trends northwest-southeast across a 10-mile stretch in Ward County. The moraine continues southeastward into McLean County, but becomes indistinct near the Blue Mountain end moraine. It has not been traced northwestward into Mountrail County; it is indistinct on aerial photographs.

The Makoti drift sheet consists mainly of gravelly and sandy clay till. It is very similar to the Blue Mountain drift sheet, but the Makoti drift sheet generally contains more boulders and pebbles and less sand and lignite chips.

In Ward County, the Makoti end moraine consists of discontinuous, generally elongate, isolated ridgelike masses. Most of the gaps between the ridges appear to represent areas in which no end moraine was ever built, because of the small load of debris in the ice or because the terminal part of the glacier had become stagnant at the time.

The larger ridgelike masses of the Makoti end moraine average about a mile in width. The maximum local topographic relief is about 40 feet. The distal margin of the Makoti end moraine generally has the steepest slope, which contrasts with the very low relief of the older ground moraine to the south. The proximal slope of the end moraine is gentle in most places.

The Makoti ground moraine is characterized by low relief, abundance of small undrained depressions, and unintegrated drainage. The ground moraine deposits range from 25 to 71 feet in thickness. The depth of oxidation ranges from 8 to 25 feet, but the average depth is about 20 feet.

The Makoti ground moraine appears to grade imperceptibly into the moderate-relief dead-ice moraine forming the border of the Ryder moraine. In the Ryder area, the Makoti ground moraine is covered by perched and collapsed outwash deposits that formed at or near the border of the Ryder ice margin.

<u>Ryder and Martin drift sheets</u>.--The Ryder dead-ice moraine is separated from the Martin dead-ice moraine by trenches partly filled with collapsed outwash, by relatively small uncollapsed outwash aprons, or by perched outwash.

North of the Martin drift sheet border, the Coteau is characterized by high-relief dead-ice moraine. In these areas the drift overlying the oxidized zone on the Blue Mountain drift sheet is generally thicker than in areas farther southwest. In several test holes the younger drift exceeds 100 feet in thickness (fig. K-1).

The terminus of the Makoti ice left a discontinuous but distinct end moraine in Ward County. It is not possible to distinguish similar end moraines for the Ryder and Martin dead-ice deposits, as shown by Colton and others (1963), Lemke and others (1965), or Clayton (1966). The margins of these drift sheets are distinct in many places where the age relations of their outwash deposits are discernable. However, in no place are their ice margins marked by linear ridges. In addition, nearly all of the outwash from the Ryder and Martin ice is collapsed, suggesting that stagnant ice was still present in front of the main ice fields as they, themselves, were melting. This suggests that the Makoti, Ryder, and Martin moraines are nearly the same age.

The difference in relief between the Makoti and Ryder drift sheets and the Ryder and Martin drift sheets probably reflects the quantity of superglacial drift on the ice. One would expect considerable superglacial drift where marginal imbricate ice thrusting took place--in this case, as it overrode the Coteau escarpment. The minor age difference between the outwash of the Ryder and Martin deposits may be more apparent than real, and may reflect only the difference in thickness of the insulating layer of superglacial drift.

If the Makoti, Ryder, and Martin drift sheets are nearly synchronous as is herein suggested, a problem becomes evident. Nearly all the reports that have been published in the last 5 years dealing with the Coteau du Missouri in North Dakota (Colton and others, 1963; Lemke and others, 1965; and Clayton, 1966) have described or referred to the Martin end moraine (Martin drift sheet border, in part, as described herein). In actuality, however, a Martin end moraine, in the strictest sense, does not exist in Ward County and the border shown on the published maps should be several miles farther south so that it coincides with the Makoti end moraine.

AGE DETERMINATION OF GLACIAL DEPOSITS

Two wood samples from the Coteau have been dated by radiocarbon methods. A white spruce log, nearly 2 feet long, was collected from a depth of 12 feet from dead-ice moraine 8 miles west of Berthold (157-87-18db). This sample (W-1818) was dated 10,330±300 B.P. Small chips of wood, white spruce cones, and an abundance of grass were collected from a depth of 137 feet from a test hole about 10 miles northeast of Makoti (153-86-3daa). This material (W-1817) was dated as 10,350± 300 B.P. The samples were collected from deeply collapsed dead-ice moraine, and both are from the youngest drift sheet.

These radiocarbon dates are significant because they indicate the approximate time that the drift was deposited in its present position on the Coteau in Ward County and perhaps in adjacent areas. The dates do not necessarily indicate the age of ice advance, which may have been at least 2000 years earlier. The latest (Makoti-Ryder-Martin) advance occurred before the Two Creeks Interstade, and Valders ice did not advance into this part of North Dakota.

The oxidized zone that has formed on the surface of the Makoti, Ryder, and Martin drift sheets in the Coteau area averages about 20 feet in thickness. One can speculate that it required about 10,000 years for the oxidation to penetrate an average depth of 20 feet or that oxidation proceeded at a vertical rate of about 2 feet per 1,000 years. Assuming similar climatic conditions during the pre-Makoti oxidation period, it would have required 13,000 or more years for oxidation to penetrate the average depth of 26 feet that is present on the buried Blue Mountain deposits. Consequently, one could assume that it required at least 23,000 years to form the two oxidized zones. If the data were available to determine the length of time required for the Makoti, Ryder, and Martin ice to advance, cover the buried oxidized zone, and retreat, it might be possible to determine the relative age of the Blue Mountain drift sheet. Highly speculative interpretations would indicate that the Blue Mountain drift sheet was deposited in excess of 25,000 years B.P. On the other hand, there is serious doubt as to the reliability of such interpretations because the oxidation rate probably varies with climate as well as rock texture and composition.

The depth of oxidation, no doubt, reflects the level of the water table since oxidation takes place in the zone of aeration. A warm dry period results in a general lowering of the water table. The depth of oxidation in the Blue Mountain drift sheet suggests a low water table (in excess of 26 feet below land surface) and a dry climate. The decline of the water table throughout a wide area and the resulting oxidation of the soil probably required several thousand years. Consequently, using this evidence, one could speculate that the Blue Mountain drift sheet is early Wisconsin (Altonian) and that the period of oxidation represents the Farmdale Stade.

## SUMMARY

In summary, it would appear that the Coteau du Missouri in Ward County consists of a bedrock "high" overlain by at least two end moraines, ground moraine, and an abundance of dead-ice moraine. At least two drift sheets are present and they differ considerably in age. The younger Makoti, Ryder, and Martin deposits have been dated by radiocarbon methods as $10,300 \pm 300$ and $10,350 \pm 300$ B.P. The upper surface of the older Blue Mountain drift sheet has been oxidized and this zone can be traced in the subsurface over wide areas. Inconclusive data suggest that the Blue Mountain drift sheet is early Wisconsin and that the buried oxidized zone represents the Farmdale Stade.

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## SUMMARY

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30-L

# PREGLACIAL TOPOGRAPHY AND DRAINAGE IN PART OF NORTHEAST NORTH DAKOTA

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During the 1964 field season 165 excavations under construction for Minuteman Missile installations in northeast North Dakota were examined by the North Dakota Geological Survey. In many of these, the shale of the Pierre and Niobrara Formations was exposed at shallow depths beneath the drift; in others, drill logs made it possible to determine the depth to shale. Over much of the area studied, the data obtained on bedrock elevations are the first available except for local water studies. The bedrock topographic map, figure L-1, is based on this information. The missile site information used on figure L-1 is supplemented in Eddy, Foster, Grand Forks, Wells, Ramsey, Benson and Traill Counties by test hole data obtained from the North Dakota State Water Commission.

The north-south trending Pembina Escarpment shown on figure L-1 separates a broad, relatively flat valley, the center of which appears to be to the east in Minnesota, from a rolling dissected upland. Elevations on the bedrock are generally about 600 to 800 feet higher on the west side of the escarpment. Although there is a surface escarpment, it is less pronounced because the drift east of the escarpment is commonly several hundred feet thicker than that to the west.

A 15-mile-wide upland just west of the Pembina Escarpment extends southward from the Manitoba border to Steele County. Although short tributary valleys head westward into the upland, there appear to be no deep valleys cut entirely through it; it must have been an effective drainage divide. In contrast to the steep eastern face, the western upland slope is gentle. On the north it appears to be cut by longer, shallower west-trending valleys, such as those in Cavalier and Ramsey Counties.

In the southwest part of the area, the deep Spiritwood Valley (Kelly, 30-J, this publication) has been diverted along the west side of the bedrock high. This valley is probably the result of glacial diversion of the drainage and does not reflect preglacial conditions. Rather, it seems likely that the regional preglacial drainage was northward to Hudson Bay and that valleys such as the Knife and Cannonball probably trended north through Towner County. Except for water well logs, data on bedrock elevations are lacking in Towner County. Paulson and Akin (1964, fig. 9) show a deep valley in Ramsey and Benson Counties that may be the preglacial Cannonball valley. It is conceivable that this deep valley may continue northwestward into Ramsey and Towner Counties and on into Manitoba, since the regional slope on the bedrock is northward in that area. It is likely that early glacial ice diverted much of the northern drainage southward and the Spiritwood valley was an outlet for this water. The same is true of the Heimdal trench in Wells, Eddy and Foster Counties; it formed at the margin of the ice, carrying a mixture of meltwater and water from the Knife and Cannonball Rivers. The Sheyenne trench in Nelson and Griggs Counties was probably not ice marginal, however; it simply formed on the surface glacial drift and cut down into the shale. It is unrelated to the preglacial drainage. Certainly, before the preglacial Knife and Cannonball River routes can be accurately determined, more data are needed from Towner, Ramsey and Benson Counties.

Permission to publish the information from the missile site excavations was granted by Colonel H. D. Weston, Corps of Engineers Area Engineer at the Grand Forks Air Force Base.

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Paulson, Q. F., and Akin, P. D., 1964, Ground-water resources of the Devils Lake area, Benson, Ramsey, and Eddy Counties, North Dakota: North Dakota State Water Conservation Commission Ground-Water Studies no. 56, 210 p.

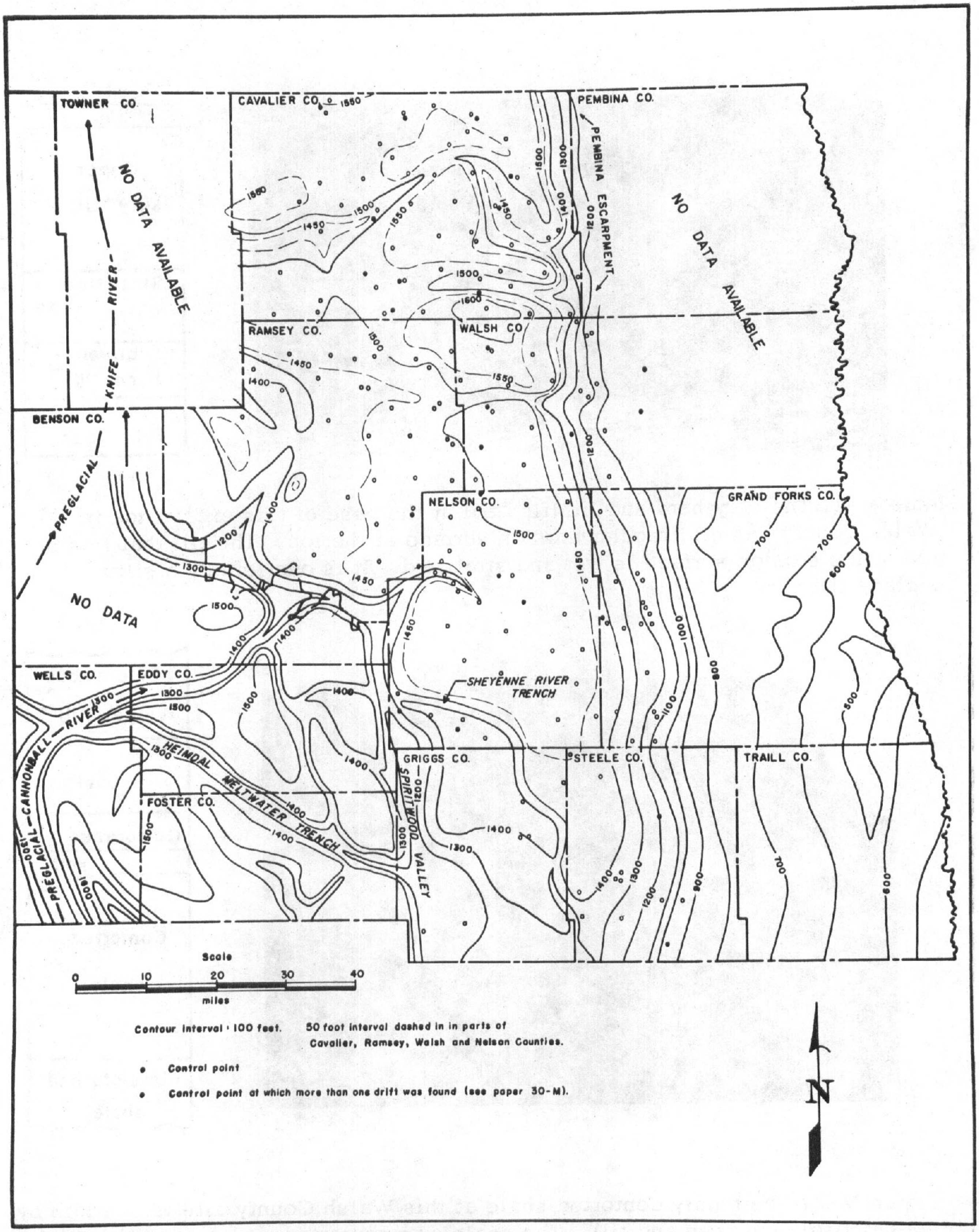


Figure L-1. Topographic map of the bedrock surface in northeast North Dakota.

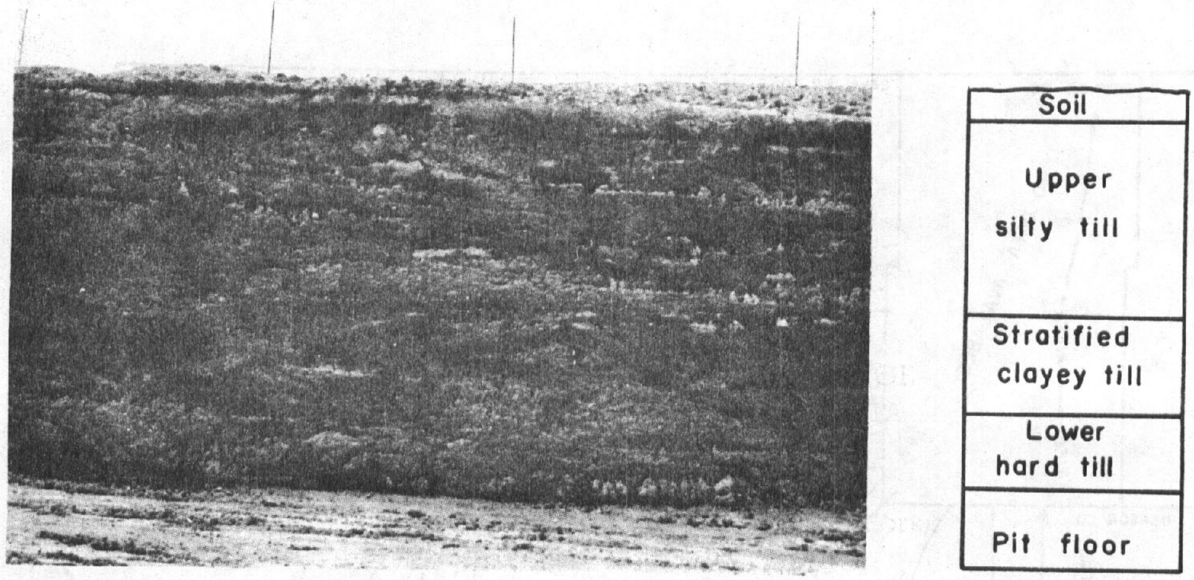


Figure M-1. The very hard and lithified till at the base of this excavation in Walsh County has an irregular erosion surface at the top. The till that lies on top of the erosion surface is wet and stratified. It is overlain by a silty surface till.

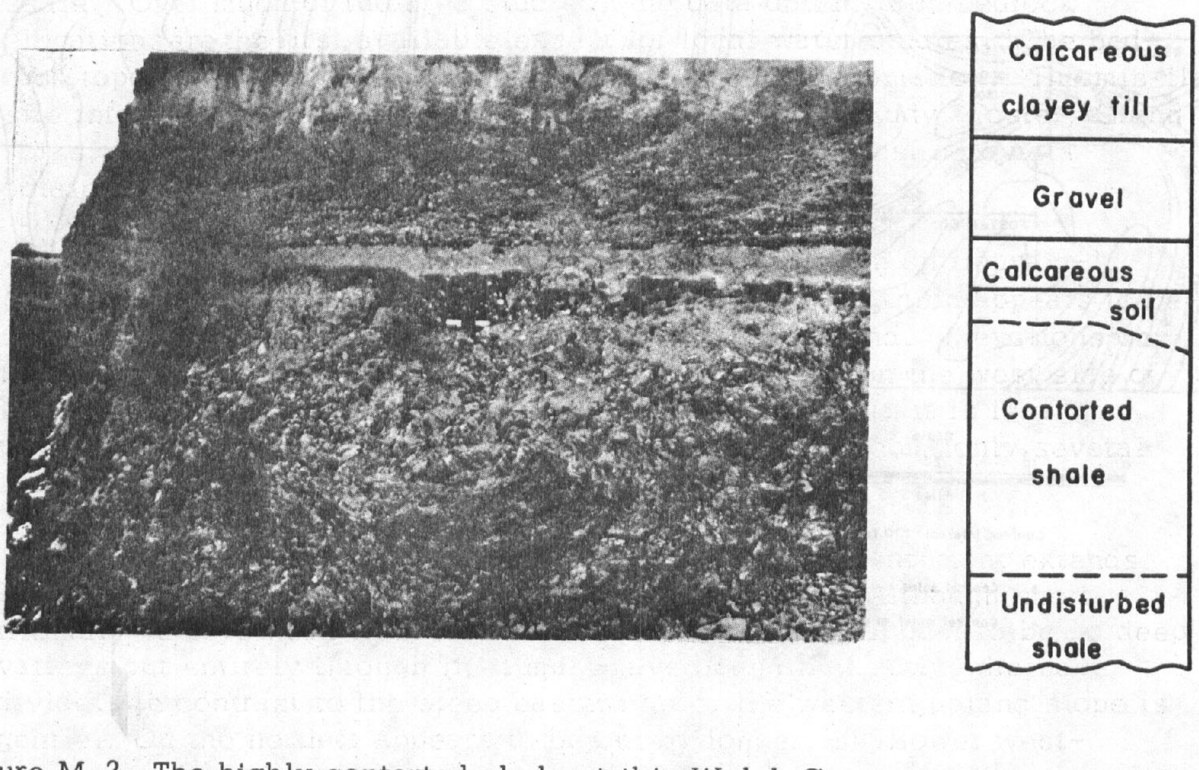


Figure M-2. The highly contorted shale at this Walsh County site is overlain by a paleosol, outwash and till. The shale was contorted by ice before the soil formed on it because the soil itself is undistrubed.

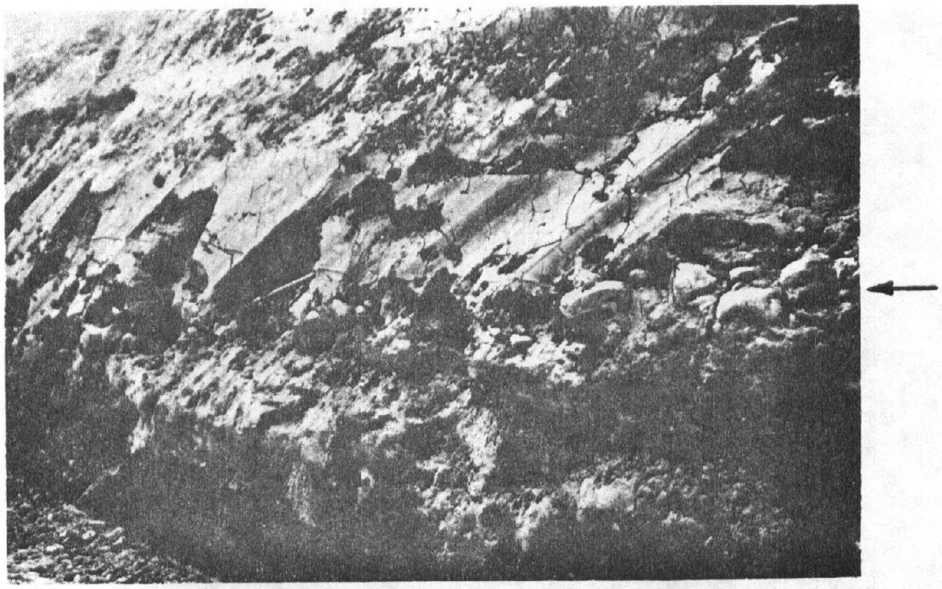

Figure M-3. Boulder pavement separating two tills in Walsh County (arrow shows boulder zone). It separates hard, sandy till at the base from an equally hard clayey till. A second boulder pavement occurs about 10 feet above the one shown here; it is overlain by loose, highly weathered till.

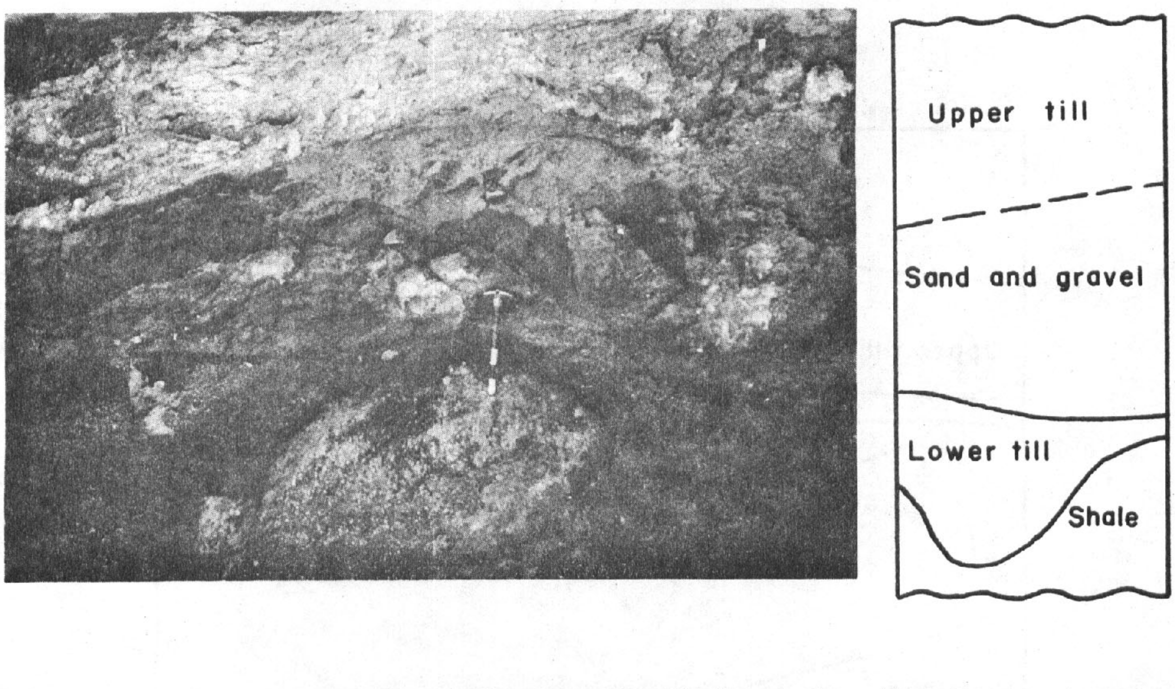


Figure M-4. Here, an erosion surface on the shale in Cavalier County is overlain by extremely hard and dense, highly jointed till that occurs only in lows on the shale surface. A 3-foot-thick horizon of sand and gravel that overlies the lower till is overlain by a loose, silty till.

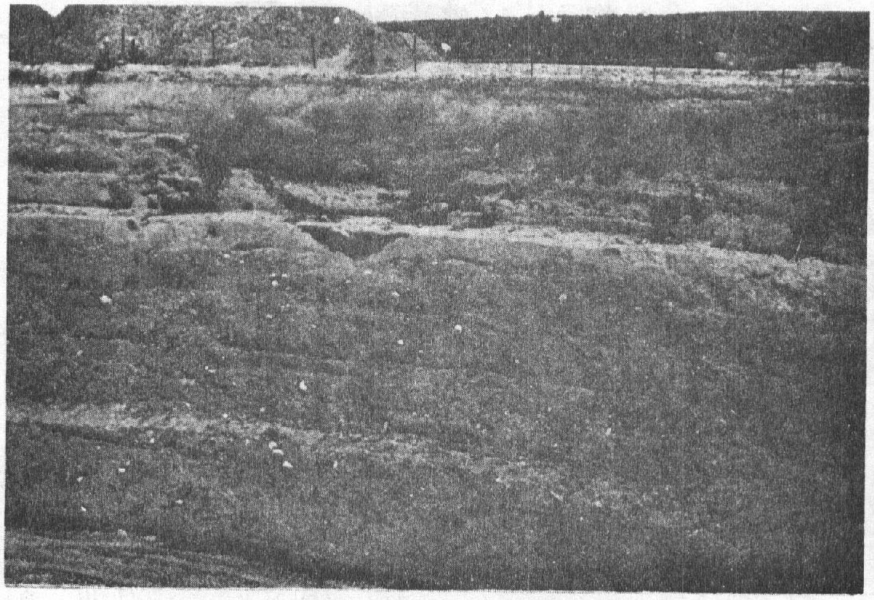

Figure M-5. A 50-foot-deep cut in Cavalier County exposes two tills. The lower till here is very hard, dense, bouldery and highly jointed, and has considerable limonite deposits. On top of the lower till is a limy zone that may be the base of an old soil horizon. It is overlain by a very silty, loose till. A lens of lake silt occurs in the upper till.

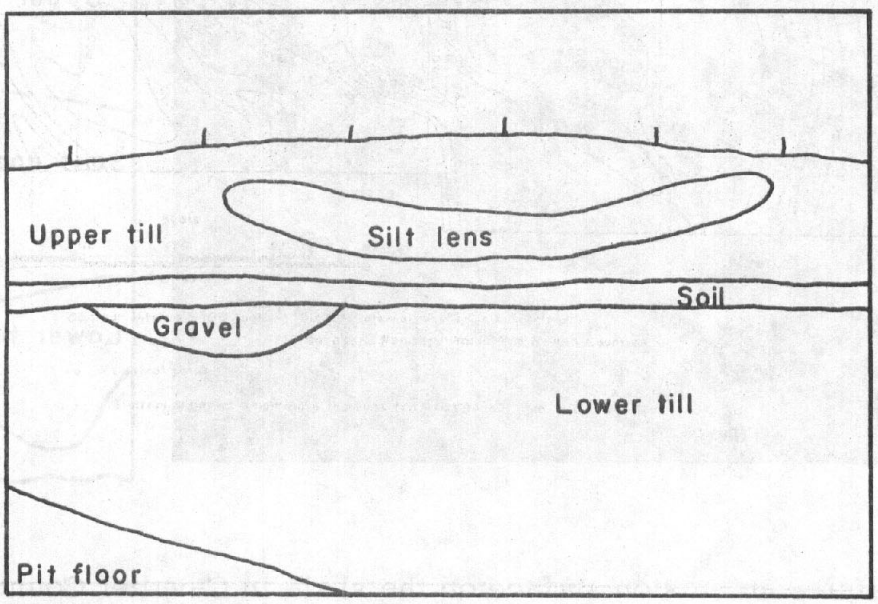


Figure M-5a. Schematic diagram of figure 5 showing stratigraphic relationships.

30-M

# MULTIPLE DRIFTS IN NORTHEAST NORTH DAKOTA

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North Dakota Geological Survey, Grand Forks

During the summer of 1964, evidence was obtained that northeast North Dakota is mantled by glacial deposits of significantly different ages. The North Dakota Geological Survey examined 165 excavations under construction for Minuteman Missile installations and found 25 exposures of two or more tills separated by outwash, boulder pavements, erosion surfaces, and buried soil profiles. Other exposures of two or more tills from various parts of North Dakota have previously been identified and described. They include exposures in Kidder County (Rau, and others, 1962, p. 22), Ward County (Lemke and Kaye, 1958, p. 95), Logan County (Clayton, 1962, p. 55), Barnes County (Kelly and Block, report in preparation), Mountrail County (Clayton, personal communication), and Ward County (Pettyjohn, personal communication). These are isolated occurrences, however, and they cannot be used to construct a regional picture of early glaciation. Tills separated by outwash and other materials are commonly penetrated in test drilling, but few fresh, open cuts showing three-dimensional relationships have been available until now.

Multiple till exposures were studied in an area that included parts of Cavalier, Walsh, Ramsey, Nelson, Griggs, Steele and Barnes Counties (fig. L-1 of previous paper). In general, the area where multiple tills were found is a gently rolling glacial till plain, mainly ground moraine, with some end moraine in the south. Outwash plains are less common in the area than they are farther west and south. The Agassiz Lake plain borders the eastern part of the area; a few of the sites are located on it. The surface deposits and topography are due to late Wisconsinan glacial deposition, although the Pembina Escarpment at the west edge of the Agassiz Lake plain is bedrock-controlled. The bedrock directly beneath the glacial drift in the area studied is shale of the Cretaceous Pierre Formation west of the Agassiz Lake plain and shale of the underlying Cretaceous Niobrara Formation beneath the lake plain. The glacial drift is underlain by 500 to 4000 feet of Mesozoic and Paleozoic sediments that thicken and dip gently westward into the Williston Basin.

The general configuration of the bedrock surface beneath the glacial drift in the study area is shown on figure L-1. The most prominent feature is the east-facing Pembina Escarpment in Pembina, Cavalier and Walsh Counties. Just west of the escarpment, the bedrock surface is high but it slopes again to the west resulting in a broad ridge in eastern Cavalier, western Walsh and Nelson Counties. This slope is probably due, in part at least, to the westerly dip of the shale bedrock; it is a modified dip-slope.

The area where multiple till exposures were most commonly found corresponds, in a general way, with the above-mentioned bedrock ridge. This is an area of thin glacial drift, commonly less than 30 feet thick in the north. In the south, and beneath the Agassiz Lake plain, the drift is as much as 400 feet thick. The large area of ground moraine underlain by thin drift has previously been considered to have relatively simply glacial geology but many of the exposures studied had very complex stratigraphic sequences.

Some of the general characteristics of the multiple till exposures will be discussed. A description of each site would be too cumbersome here; they are available, along with photographs, as an open-file report of the North Dakota Geological Survey.

Hardness and compaction of the tills.--At most of the sites the lower tills are considerably harder than the upper tills and at some of the sites the lower tills are remarkably hard and compact. At a few of the sites the lower tills are tillites as highly indurated as limestone. Such density and hardness may be a function of time; lithification processes are more advanced than in the younger tills. The compact tills may have been compacted by overriding Wisconsinan ice following settling, drying contraction, and weathering during the interglacial stages.

Jointing in the till.--The lower tills at most of the sites are much more strongly jointed than are the upper tills. At many of the sites, the joints form definite patterns with the vertical joints oriented in rectilinear patterns. At those sites where the upper tills have any jointing at all, it is randomly oriented and not at all extensive. Iron and manganese oxides are much more conspicuous in the lower tills than in the upper ones. Quite commonly, limonite and pyrolusite staining is concentrated on and near joint faces. At a few of the sites gravels in the lower tills are cemented with iron oxide and are highly ferruginous. At some of the sites pebbles and cobbles have weathered to "ghosts."

Buried erosion surfaces and paleosols.--The presence of well-developed buried erosion surfaces, buried oxidized zones, and buried soil profiles (figs. M-1 and M-2) on top of the lower tills at several places is further evidence of their relatively ancient age. Flint (1955, p. 31) states that reddish paleosols in South Dakota are probably of Sangamon age, and the underlying till is, therefore, Illinoian, Kansan or Nebraskan. Many of the paleosols observed during this study have been partially eroded and decapitated so that only part of the original soil profile remains (fig. M-5). The soils themselves were not red, but limonite accumulations commonly give reddish hues to the lower tills.

Boulders in the tills.--Buried boulder pavements are common both within single tills and between tills (fig. M-3). At many of the sites, the tills above and below the boulder pavements are not appreciably different. In several places, however, the boulder pavements are closely associated with an outwash zone, erosion surface or a buried soil profile; they may mark significant breaks in deposition. Many of the boulder pavements observed are probably lag deposits, left when meltwater from nearby ice washed the fine materials away, leaving the less easily transported boulders and cobbles behind.

At a few of the sites the lower tills have many more cobbles than the upper tills (fig. M-5). No particular reason for a systematic difference in the relative abundance of cobbles in the various tills is apparent. The number of cobbles must depend on what the ice was moving over; if it moved over a cobble pavement for example, more cobbles might be expected in the upper till. No obvious differences in lithologies of the cobbles were found.

Thickness of the glacial drift.--Most of the multiple till sites, particularly those in the northern part of the study area, are in thin drift where bedrock was either exposed or was only a few feet below the bottom of the excavation. Exceptions in the northern part of the state were mainly on or near the Agassiz Lake plain. In the southern part of the area, drift is thicker and multiple till exposures were less common. Multiple tills are not necessarily absent there. Because the excavations were only 30 to 50 feet deep, it is likely the lower tills were not reached in the south, where the drift is as much as 300 or 400 feet thick. In the north, where the drift is commonly less than 30 feet thick, the excavations cut through the entire section of glacial deposits, thereby increasing the probability of finding thin deposits of till that were present in lows on the bedrock surface (fig. M-4).

Age of the lower tills.--The age of the lower tills is unknown. Most of them fit many of the criteria used to identify pre-Wisconsinan till, and some of them are almost surely pre-Wisconsinan. In northeast North Dakota the stratigraphic and physical characteristics of the various lower tills are not outstanding enough to permit their correlation either with each other or with any particular stage or substage. A piece of wood taken from the third drift from the surface at a depth of 44 feet in Ramsey County was radiocarbon-dated at greater than 28,000 years (W-1528).

Acknowledgments.--Permission to publish the information from the missile site excavations was granted by Colonel H. D. Weston, Corps of Engineers Area Engineer at the Grand Forks Air Force Base.

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30-N

# A PRELIMINARY REPORT ON THE POSTGLACIAL SEDIMENTOLOGY OF DEVILS LAKE, NORTH DAKOTA

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## INTRODUCTION

Devils Lake, a closed lake, is located in northeastern North Dakota (fig. N-1) and is part of a chain of waterways situated in the Devils Lake drainage basin which has an approximate area of 3,900 square miles. During the late 1800's the lake had an approximate area of 80 square miles, but has steadily decreased and presently occupies an area of 18 square miles.

The mean annual air temperature at Devils Lake is 4$^{o}$C with a range of -40$^{o}$C to 36$^{o}$C. The mean annual precipitation is 44.5 cm (17.5 inches), and the prevailing wind is from the northwest with an average velocity of 10 mph (Swenson and Colby, 1955).

The surficial deposits in the Devils Lake area are primarily glacial and lacustrine in origin. The glacial drift varies in thickness from 3 to 120 m (10 to 400 feet) and was derived primarily from lower Paleozoic limestone and Cretaceous shale. Glacial Lake Minnewaukan, predecessor of the present Devils Lake-Stump Lake complex, originated as a moraine-dammed proglacial lake behind the North Viking morainal complex. The lake declined rapidly as the volume of meltwater subsided and eventually split into several isolated lakes and bays of which Devils Lake is the largest (fig. N-1).

The purpose of this study is to interpret the postglacial sedimentary history of Devils Lake through physical and chemical analyses of its sediments, especially those from the present Main Bay (fig. N-1). This paper is only a preliminary report on the results of this study and will not discuss in detail many interesting and problematical aspects of this research. A general outline of the sedimentary history of Devils Lake based upon logical conclusions from the analytical data presented will be given.

This work was supported in part by the North Dakota Water Resources Research Institute with funds provided by the U. S. Department of Interior, Office of Water Resources Research under P. L. 88-397.

## METHODS

Two 7 m and one 8 m cores (see fig. N-1 for location of cores) were taken with a piston coring device designed after that of Colinvaux (1964). The cores were kept in sealed plastic core liners and transported back to the laboratory for sampling. The sediment was extruded with the aid of a hand-operated piston and the extruded cores were thoroughly scraped to remove all contaminating material and sampled according to lithology and color. One portion of each sample was weighed, dried for 48 hours at 90°C, and weighed again to determine water content. The resultant dried sediment was ground for chemical analysis. The other portion was placed in an air-tight jar. Part of this remaining sample was used later for grain-size analysis performed by the standard sieving and pipette method (Krumbein and Pettijohn, 1938). Chemical analysis of the dried, ground sediment samples consisted of: total carbon by quantitative dichromate oxidation and measurement of the $CO_2$ evolved (Maciolek, 1962); carbonate carbon by reaction with hydrochloric acid and measurement of the $CO_2$ evolved (Maciolek, 1962); organic carbon as the difference between total carbon and carbonate carbon; and total nitrogen by macro-Kjeldahl distillation of the solution which remained from the total-carbon analysis (Maciolek, 1962). Calcium, magnesium, and iron were determined compleximetrically (EDTA titration of Ca, Ca + Mg) and colorimetrically (Fe) on filtered solutions from the carbonate analysis (Shapiro and Brannock, 1962).

## PRESENT ENVIRONMENT OF DEVILS LAKE

Devils Lake has a simple morphometry. The lake basin is shaped like a pie plate with a present mean depth (1965) of 2.6 m (8.5 feet) and a maximum depth of 3.4 m (11 feet). Strong winds frequently mix the lake waters during the ice-free months and any thermal stratification is temporary and easily destroyed by the slightest wind. Summer water temperatures reach 28°C, while winter temperatures under the ice approach 0°C. Transparency is low and an average Secchi disc reading during the ice-free months is 0.5 m. The lake is sealed off by ice during the winter and early spring months and reducing conditions develop.

During the summer of 1965 the salinity of Devils Lake was 12.3 parts per thousand, but beneath the ice in March of 1966 it reached nearly 16 parts per thousand due to concentration by freezing. Devils Lake water differs from normal sea water in that it has more sulfate and much less chloride. Sulfate comprises nearly 50 percent of the ions in Devils Lake, whereas chloride makes up nearly 60 percent of the ions in the ocean and

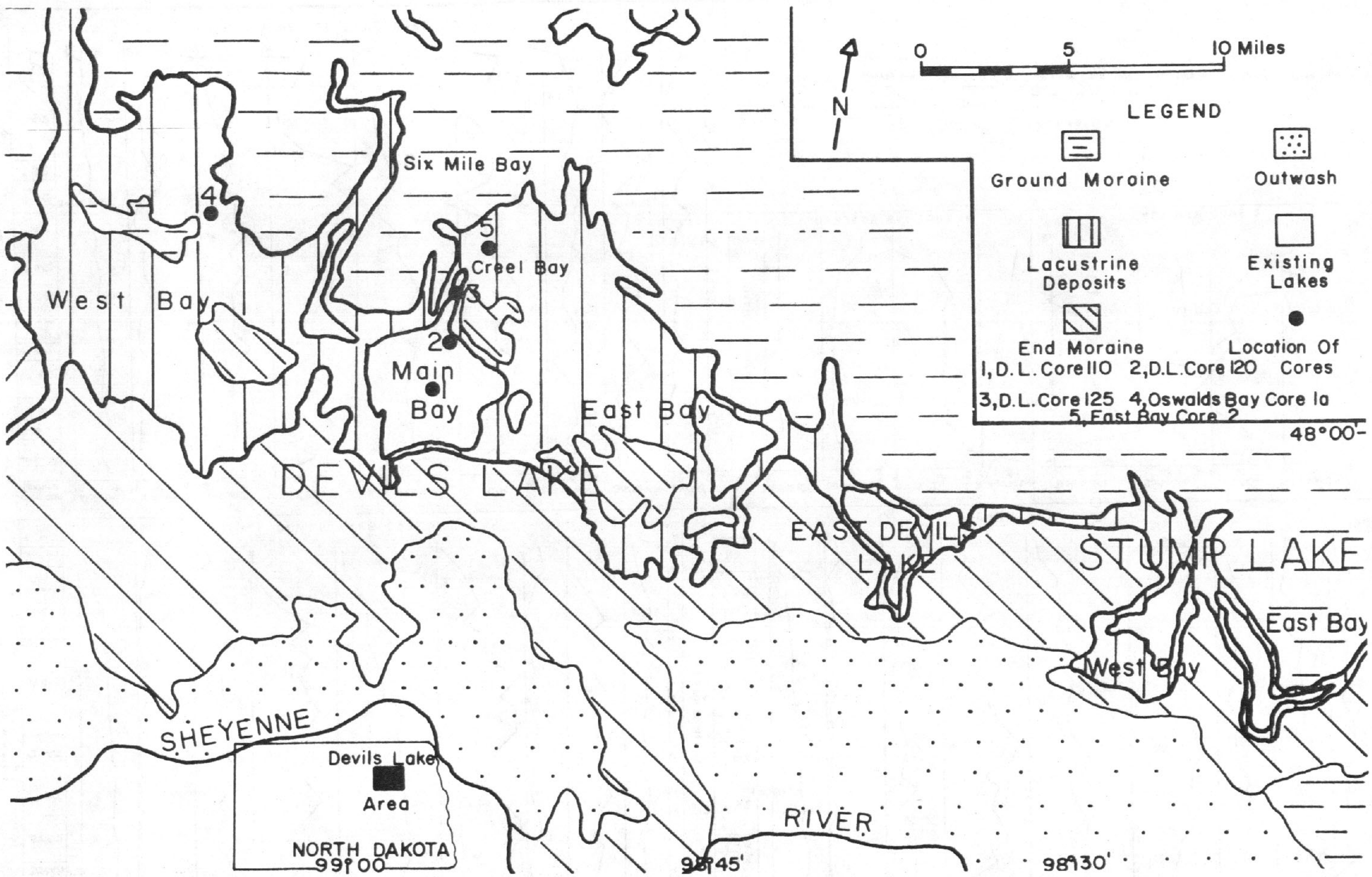


Figure N-1. Generalized geologic map of the Devils Lake area showing location of cores.

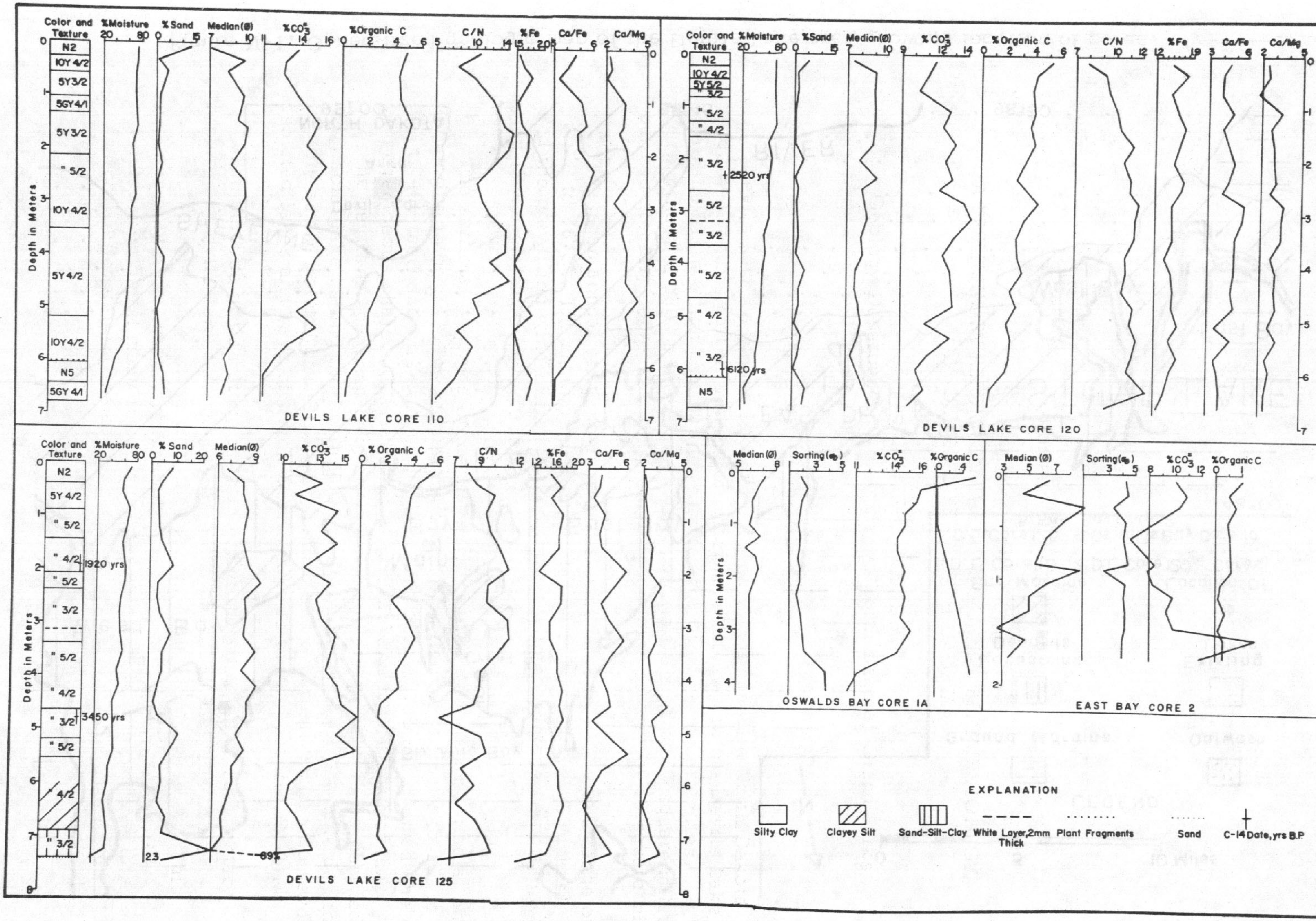


Figure N-2. Analytical data for Devils Lake cores.

sulfate less than 1 percent. The very high organic nitrogen and phosphorous contents of Devils Lake are a result of the influx of sewage and agricultural pollutants.

In spite of apparently high concentrations of available nutrients, the biota of the lake is not especially rich in terms of biomass and is distinctly poor in terms of biological diversity. However, the lake can be definitely classified as an eutrophic lake (Armstrong, and others, 1967). The vertebrates are represented by one species of salamander. The only rooted plant in Devils Lake is Ruppia maritima. Discontinuous beds of filamentous algae occur in the littoral zone and constitute the major part of the attached aquatic vegetation.

There are three main types of surface sediments in Devils Lake: (1) a black hydrogen sulfide-rich silty clay that occurs throughout 80 percent of the bottom area, (2) a light grayish brown near-shore sand, and (3) a calcareous mud which is found in the shallower areas heavily populated by attached algae. Cobbles and boulders are common in these shallow areas. The average pH of the black silty clay is 8.1, and the Eh ranges from -50 to -150 mv.

The sediment contains 8 to 14 percent (by weight) carbonate, 5 to 7 percent calcium, and 1 to 4 percent magnesium. The Ca/Mg ratio varies little (1.44 to 1.78) throughout most of the lake except in the littoral areas occupied by attached algae, where it is significantly higher (2.20 to 3.11), reflecting carbonate precipitation associated primarily with physiological processes of the algae. Equilibria calculations based on chemical analyses of the lake water during 1965 (taking into account the appropriate activity coefficients and the effects of inorganic complexing) indicate that the lake was supersaturated on an average of 250 percent with respect to calcite and 150 percent with respect to dolomite during the summer and fall of 1965. Preliminary mineralogical analyses indicate that the surface sediments contain approximately equal amounts (9.5 percent) of calcite and dolomite. The large difference between the Ca/Mg ratio of the glacial till surrounding the lake (5.0) and the surface sediments (1.5) suggests that a substantial portion of the carbonate minerals was not transported mechanically into the lake but precipitated from the water as a result of physico-chemical and biochemical processes.

The organic carbon content of the silty clay ranges from 5.0 to 5.6 percent and the C/N ratio varies from 7.2 to 8.4, ratios which reflect the planktonic source of organic matter in the sediments.

## ANALYSIS OF CORES-RESULTS

### Water content and texture

The variation of water content (expressed as percentage wet weight of sediment) with depth in the three cores from Main Bay of Devils Lake is plotted in figure N-2. Cores 110 and 120 show a gradual decrease from approximately 70 percent in the upper meter of sediment to 42 percent at a depth of 6 m. The water content decreases more rapidly to 26 percent in the lower 65 cm of the two cores. The water content of core 125 from Creek Bay decreases gradually from 70 percent at the top to 38 percent at the bottom. The very low water content of the last sample is a result of reduced pore space due to its coarse texture (sand). There appears to be no general correlation (with a few exceptions) between water content and median grain size, suggesting that the water content is controlled by other processes such as compaction due to the weight of overburden or dessication.

The texture of cores 110 and 120 is very uniform (silty clay) and the median grain size varies little throughout their length. The coarser sediment in the top of both cores appears to be related to postdepositional mineralogical changes and does not reflect an increase in detrital sediment size. The average median diameter of the sediments in core 110 (from the middle of the lake) is approximately 9 phi units, while that of core 120 (from the edge of the lake) is approximately 8 phi units, the difference reflecting proximity to the shoreline. However, a greater difference was expected. The fine uniform texture of both cores is probably due to a relatively constant slow rate of inorganic sedimentation into the lake during the last 7000 years (see section on rate of sedimentation).

The texture and median grain size of the sediments from core 125 varies considerably more than in the other cores from Main Bay. The sediment contains noticeably more sand, reflecting the influence of one of the few streams entering Devils Lake. The upper 5 m of sediment is generally finer (median diameter of 8 phi units) than the lower 2 m which has an approximate median diameter of 7 phi units. The sediment in all three cores from Devils Lake is very poorly sorted with the exception of the sample from the base of core 125 which is a well-sorted fine sand.

The median grain size of sediments from East Bay core 2 varies from 3 to 9 phi units with a trend toward coarser sediment with depth. The sorting, which is extremely poor for all but two samples, is fairly constant throughout the core with a standard deviation of approximately 4 phi units. The median grain size of sediments from Oswalds Bay core 1A varies from 5.5 to 7.2 phi units, and there is no general trend with depth. The sorting (poor) is also fairly constant at a standard deviation of approximately 2 phi units except in two samples from the base of the core where it becomes extremely poor.

### Carbonate Content

The variation in carbonate content, expressed as percent analytical $CO_3^{=}$, of each core is shown in figure N-2. The carbonate content of the three cores from the present lake varies from 9 to 16 percent. There appears to be good positive correlation between the curves for cores 110 and 120, and they exhibit significant minima at 60 cm and 600 to 665 cm. Core 120 generally has 2 percent less carbonate than core 110 even though it is much closer to the edge of the lake where one would expect to find more, instead of less, carbonate. This difference might be explained by its slightly coarser texture. The carbonate content curve for core 125 generally does not correlate well with the others. It varies considerably in the upper 2 m of sediment and there is only one significant carbonate minima at 600 to 660 cm.

The carbonate content of the two cores from the dry part of Devils Lake varies from 7 to 20 percent (fig. N-2) with the Oswalds Bay core containing several percent more than that from East Bay. In the East Bay core there is a pronounced minima at 50 to 70 cm with maxima at the top and bottom (160 cm), while in the Oswalds Bay core there is a maxima at the top (10 cm) followed by relatively constant values (75 to 300 cm), and a minima at the bottom (415 cm).

### Calcium and magnesium

The calcium and magnesium content of cores 110, 120, and 125 was determined, but the results are not plotted in figure N-2. The average calcium content (and the range of values) for the three cores respectively is: 7.87 percent (5.45 to 9.57 percent), 6.95 percent (4.90 to 8.45 percent), and 7.07 percent (5.20 to 9.10 percent). The average and range of values for the magnesium content of the cores is: 2.47 percent (2.00 to 3.25 percent), 2.46 percent (2.10 to 3.30 percent), and 2.54 percent (2.05 to 3.15 percent). Calcium-magnesium ratios were calculated for all samples and are plotted in figure N-2. It is apparent from the above data that calcium varies more widely than magnesium and that significant fluctuations in the calcium content generally result in significant fluctuations of the calcium-magnesium ratio. All three curves exhibit several maxima and minima with a general trend toward higher ratios in the middle of the sediment column.

### Organic carbon and nitrogen

The organic carbon content of the three cores from the present lake varies from 0.5 percent to 5.6 percent. All three cores have a similar average organic carbon content of approximately 3 percent. The general trend is a decrease in organic carbon with depth (fig. N-2), which is more rapid in the upper and lower meter of sediment and more gradual in the middle part of the sediment column.

The organic carbon content of the two cores from the dry part of Devils Lake is very low (0 to 1 percent). There is a general trend toward an increase in organic carbon with depth in the Oswalds Bay core, while the reverse is true for the East Bay core.

Although the total nitrogen content of the samples from cores 110, 120, and 125 was determined, the results are not plotted in figure N-2. The organic carbon-nitrogen ratio is plotted instead. The curves for cores 110 and 125 show considerable sample to sample variation with ratios ranging from 5 to 15 for core 110 and from 6 to 11 for core 125, while that for core 120 shows much less inter-sample variation even though the ratios range from 8 to 12. The general trend of each curve is different and somewhat obscured by sample to sample variation.

Iron

The iron content of samples from all three Devils Lake cores is plotted in figure N-2. It is readily apparent that the iron content varies little with values ranging from 1.3 to 1.9 percent. There is no apparent general trend with depth in the cores.

## DISCUSSION OF RESULTS

Texture

The predominantly silty clay texture of all three cores from the present lake is very uniform (fig. N-2) except at the base of core 125. The sand-silt-clay relationships of most samples from these cores can be explained as the result of sediment deposition from fluid suspension under relatively quiescent conditions; such conditions commonly produce fine-grained sediments. The very poor sorting of these sediments, although partially the result of their very fine texture, is characteristic of sediments deposited in low energy environments (Emery, 1960). The coarser texture, sand-silt-clay and sand, of the lowest two samples from core 125 indicate a change from a low energy basin environment to a high energy littoral environment. The sand at the base of the core is moderately well sorted (standard deviation of 0.625 phi unit) and has a particle-size distribution very similar to that of near shore sands present in the lake today.

The texture of sediments from the dry part of Devils Lake is significantly coarser and, in one core (East Bay core 2), less uniform than the sediments from the present lake. This is attributed to deposition in a mixed-energy environment. The texture of the sediments from Oswalds Bay core 1A is predominantly silt with lesser amounts of clay and minor amounts of sand. Exceptions to this are the lower two samples in which the sand fraction becomes significant (26 percent). The sorting of these two samples is very poor in comparison to the remaining samples which exhibit poor sorting (fig. N-2). The significantly larger median diameter and the leptokurtic and fine-skewed nature of nearly all the samples from the Oswalds Bay core suggest deposition in a medium-energy environment where the silt contributed by the stream(s) was not mixed with large quantities of clay. This mixing would impart a strong bimodal character to the sediment. Such a bimodal distribution is characteristic of many marine neritic sediments (Folk and Ward, 1957). The extremely poor sorting and near-symmetrical unimodal nature of the lower two samples from the Oswalds Bay core suggests deposition from a high-energy fluid medium which has a low sorting efficiency, perhaps a proglacial or meltwater environment.

The sediments in East Bay core 2 are predominantly of the sand-silt-clay textural type. Two samples can be classified as silty clay and one as silty sand. The sorting is extremely poor for all samples except the two classified as silty clay, in which it is poor to very poor. The extremely poorly sorted, often nearly symmetrical, strongly bimodal character of most samples suggests deposition in an environment in which the sorting efficiency was low and the sediment source was polymodal in nature. Such a situation might exist where the sediment source is glacial till or other extremely poorly sorted material which is transported to the deeper parts of the lake and deposited with the normal basin sediment. The polymodal nature of most samples suggests this is the case rather than deposition of well sorted littoral sediment transported _en masse_ by storms to greater depths, mixed with clay, and deposited in a medium of low sorting efficiency. The two finer-textured samples appear to be normal deep water deposits. Better sorting relative to the other samples is probably a result of their very fine texture.

## Carbonate

The carbonate curves for cores 110, 120, and 125 exhibit several major fluctuations. However, they do not represent any significant changes in the environment of carbonate deposition. The environment was probably very similar to the present one in which nearly half of the carbonate is biochemically precipitated and the remainder is probably detrital in origin. The significant decrease in carbonate at the base of cores 110 and 120 may be the result of an increase in inorganic sedimentation and a concomitant decrease in biogenic carbonate deposition.

The carbonate content of the Oswalds Bay core varies considerably. Although most of the carbonate is detrital in origin, some is in the form of calcareous shell material, particularly in the top two samples. The same is true for the East Bay core except that the carbonate maximum at 1.5 m is the result of an increase in detrital mineral content and not shell material.

Ca/Mg and Ca/Fe ratios

The Ca/Mg ratio curves closely follow those of carbonate, suggesting some direct relationship. There is a very high degree of correlation (correlation coefficient = +.91) between calcium and carbonate and a very low degree between calcium and magnesium (correlation coefficient = +.05). Most of the carbonate is in the form of mineral calcium carbonate. This is also true for the glacial deposits around Devils Lake.

The Ca/Mg and Ca/Fe curves also correlate very well. Because the iron content of the cores varies little (fig. N-2), fluctuations in the ratio are due largely to changes in the calcium content. The iron content of the till surrounding the lake is essentially the same as that of the cores. This, combined with the constant median grain size of the sediment in the cores, suggests that the rate of sedimentation has been relatively constant for the last 6000 years (see section on rate of sedimentation). This being the case, then fluctuations in the calcium and carbonate content of the sediment reflect changes in authigenic carbonate deposition; that is, a significant amount of the carbonate is authigenic in origin.

Organic carbon

The organic carbon content of all cores from the present lake gradually decreases from top to bottom. There is no strong positive correlation (correlation coefficient = +.36) between organic carbon and carbonate, suggesting that either the same processes did not affect both parameters in the same way, or one or both have undergone diagenetic change. If a significant portion of the carbonate is a result of biogenic precipitation, then it seems reasonable to assume that fluctuations in this parameter will follow similar fluctuations in organic matter; that is, increased primary productivity. This might increase the rate of biogenic carbonate deposition and should cause an increase in the depositional rate of organic matter. The order of magnitude of change, however, may not necessarily be the same for both parameters.

The organic carbon content of sediments from the Oswalds Bay core is very low and increased steadily with depth. This increase may be due to greater contribution of allogenic organic material during earlier stages of lacustrine development. The organic carbon content of sediments from East Bay, although low, is not as low as that from Oswalds Bay. There is a gradual decrease in organic carbon with depth in the East Bay core as in the cores from Main Bay. This decrease in organic carbon with depth appears to be the result of normal lacustrine succession (Lindeman, 1942; Hutchinson and Wollack, 1940).

Carbon-nitrogen ratios

The C/N ratios of the cores from Main Bay exhibit significant inter-sample variation which reflects changes in the source of organic material or diagenetic effects. Carbon-nitrogen ratios of 8 to 14 in sediments indicate that a major part of the organic matter is authigenic (Emery, 1960; Scholl, 1963) while higher ratios indicate that the allogenic component (terrestrial vegetation) is becoming increasingly significant. The C/N ratios for the Devils Lake cores never exceed 14, but it is possible that sediments with this high a ratio may contain significant quantities of allogenic organic material. The lowest ratios (5 to 7) are also noteworthy. This reflects either organic matter whose originally low C/N ratio has been essentially unaltered, or diagenetic alteration or organic carbon and nitrogen. The C/N ratio of surface sediments (silty clay) in Devils Lake averages approximately 8. It is unlikely that the C/N ratio of newly deposited sediment has been significantly lower at any time during eutrophication of the lacustrine environment. It appears, therefore, that the low C/N ratios of several samples in the cores resulted from a loss of organic carbon relative to total nitrogen. Loss of nitrogen from the sediment could explain the highest C/N ratios observed (Emery, 1960). The low C/N ratios at the base of all three cores are probably the result of more severe alteration (oxidation) of organic carbon in a more oxidizing lacustrine environment.

Rate of sedimentation

Radiocarbon dates were obtained on four samples from the Devils Lake cores; two from core 120 and two from core 125. The radiocarbon age determinations were used to compute rates of sedimentation expressed as: (1) thickness of the annual layer and (2) weight of sediment deposited per unit area per year.

The depositional rate (expressed as thickness of sediment deposited per year) for core 120 is 1.04 mm/year for 0 to 3 m, and 1.13 mm/year for 3 to 7 m; the depositional rate for core 125 is 1.34 mm/year for 0 to 3 m, and 1.68 mm/year for 3 to 7 m. It is apparent that the rate of sedimentation is relatively constant for core 120 and that it is significantly greater for core 125. In addition, the rate of sedimentation increases by 25 percent throughout the length of core 125.

If the rate of sedimentation is calculated on a weight per unit area per year basis, thereby avoiding the uncertainties of compaction inherent in the use of thickness of annual layer as an expression of rate of deposition, then the rate of core 120 becomes 81 $mg/cm^2$/year for 0 to 3 m and 122 $mg/cm^2$/year for 3 to 7 m; and the rate for core 125 becomes 119 $mg/cm^2$/year for 0 to 3 m and 277 $mg/cm^2$/year for 3 to 7 m. It is readily apparent that the rate of sedimentation is significantly greater for the lower 4 m of sediment in both cores.

## PRELIMINARY INTERPRETATION OF SEDIMENTARY HISTORY OF DEVILS LAKE

The analytical data presented here suggest that the postglacial sedimentary history of Devils Lake has been relatively simple. Only two types of environments can be inferred from the sediment data: (1) a deeper, oligotrophic lake existing from at least 7000 years to 4000 years B.P. and (2) a shallower, eutrophic lake which attained eutrophy around 3500 years B. P. Although the lake level has fluctuated at least 35 feet in historical times (Swenson and Colby, 1955), similar fluctuations during the last 7000 years are not readily discernible from sedimentary parameters. Either these parameters are not sensitive enough to water-level fluctuations of this magnitude, or the results of such fluctuations have been modified by other processes. The question arises, how long has Devils Lake been a closed lake? If it has been a closed lake for any length of time, should we not see the result of this in the sediments? With the exception of a 2 mm thick white layer (see fig. N-2) which occurs at a depth of approximately 320 cm in cores 120 and 125, there are no visible evaporite and related deposits in the cores from Devils Lake. It is entirely possible that no such deposits ever formed in Devils Lake because the lake never became very saline in the last 6000 years, or that saline minerals were subsequently dissolved. Substantial ground water inflow into the lake could maintain higher lake levels and dissolve readily-soluble minerals as it moved through the sediment. A ground water gradient, whereby water moves from Devils Lake to East Devils Lake to East Stump Lake, could explain the much greater salinity of East Stump Lake (235 parts per thousand in 1965) which is situated at the eastern end of the Devils Lake drainage basin.

The low carbonate content and the very low organic carbon content of the lower 75 cm of sediment in cores 110 and 120 suggest that this sediment was deposited in an oligotrophic lake. The low C/N ratios of these lower samples indicate that the entire lacustrine environment, both water and sediment, was oxidizing and that abundant available oxygen caused partial oxidation of the organic matter deposited in the lake. The significantly lower water content of this sediment relative to the overlying material precludes water loss only by the process of compaction, and suggests drying. The radiocarbon date of 6120±120 years B. P. for the top of this stiff gray, organic poor silty clay falls within the age range for the hypsithermal interval (5500 to 7500 years B.P.) as dated in several Minnesota lakes and bogs (Wright, and others, 1963). Pronounced evaporation during this dry period probably caused the lake to dry up and dessicated the recently deposited sediment. Calculations using the rate of sedimentation computed for the 3 to 6 m interval and the analytical data suggest that the rate of inorganic sedimentation was not significantly greater during this time than that represented by the lower 4 m of sediment in core 120. Therefore, it appears that the low organic productivity in the lake rather than dilution of organic matter by increased influx of detrital sediment resulted in the low organic content of the sediment.

Sometime around 6000 years B.P. the dry climate may have begun to ameliorate in the Devils Lake area as it did in Minnesota, thereby raising the lake level (Wright, and others, 1963). However, the lake could not have been too deep when the well-sorted littoral sand at the base of core 125 from Creel Bay was deposited because no other sand layers are present in the core, suggesting that this part of the lake was always too deep for development of near shore sands. The age of this sand is approximately 5000 years B.P.

The lake level probably rose gradually from this point. Continued weathering in a more temperate climate added more and more nutrients to the lake resulting in greater primary productivity. The increased depositional rate speeded the processes of filling in the deeper parts of the lake. The gradual increase in organic productivity and the filling in of the lake basin culminated in the development of an oxygen-deficient hypolimnion where the production of organic matter exceeded that which could be oxidized by respiration and bacterial decomposition. It is at this point in lacustrine succession that a lake changes from oligotrophy through mesotrophy to eutrophy (Lindeman, 1942). The gradual increase in organic matter in the lower 3 m of sediment in cores from Devils Lake indicate that the rate of detrital sedimentation was gradually decreasing and the organic productivity gradually increasing in the advent of cooler wetter climatic conditions toward the end of the hypsithermal. The sediment data suggest that biogenic carbonate deposition became important around 5000 years B.P.

Devils Lake probably attained eutrophy sometime around 3500 years B. P. and appears to have remained eutrophic until the present. The rate of detrital sedimentation was significantly decreased and the contribution of organic matter significantly increased by this time. The lake is still accumulating organic matter, partly from an influx of fertilizer and sewage during the last 50 years, and may not yet have reached ideal eutrophic stage-equilibrium (Lindeman, 1942).

No radiocarbon dates are available for the cores from Oswalds Bay and East Bay, and thus absolute time correlation with the cores from Devils Lake is impossible. However, the sediment data suggest a sequence of events from proglacial sedimentation in the earliest stages of lacustrine development to nonglacial sedimentation during the later stages of development which are represented by the cores from Main Bay.

The dry period (650 to 400 years B.P.) postulated by Aronow (1957) for the Devils Lake region is not reflected in the cores from Devils Lake. There is a significant decrease in the carbonate content of the cores at a depth corresponding to this time, but a pronounced lowering of lake level should cause increased carbonate deposition. The C/N ratio of the sediment from core 110 for this interval (20 to 50 cm) is also significantly lower, a fact which suggests loss of organic carbon possibly by oxidation during this dry period. However, the C/N ratio of sediment from the other two cores (fig. N-2) does not decrease significantly. In any case, the climatic change postulated by Aronow (1957) was not severe enough to dry up the main part of Devils Lake and cause dessication of the sediment. The water content data (fig. N-2) indicate no drying surface in the upper meter of Devils Lake sediment.

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30-O

# LATE-RECENT ALLUVIUM IN WESTERN NORTH DAKOTA

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## INTRODUCTION

The purpose of this report is to present a preliminary account of a part of the field investigations on badlands erosion conducted during the summer of 1966 as a part of the University of North Dakota Department of Geology's Seepage Erosion Effects Project. This work was supported in part by the North Dakota Water Resources Research Institute with funds provided by the U. S. Department of Interior, Office of Water Resources Research under P. L. 88-379, and was under direction of Lee Clayton of the Department of Geology.

The area of investigation included several small tributaries of the Little Missouri and Missouri Rivers. The drainage basins studied were selected as typical of the majority of the small basins in the area that are characterized by vertical-walled gullies, or arroyos. The alluvial stratigraphy is easily observed in the walls of these gullies.

The basins studied ranged in extent from a few acres to a few square miles. The valleys were cut into clays, silts, and sands of the Paleocene Fort Union Group. The alluvial deposits are in the form of a valley fill that ranges in thickness from a few feet to a few tens of feet. The surface of the till has an average slope of 1 degree.

## TERRACES AND ALLUVIUM ALONG THE LITTLE MISSOURI RIVER

Several terraces occur along the Little Missouri River, but little has been written about them. Laird (1950) described four terrace levels present along the Little Missouri River in the south unit of Roosevelt Park. The upper terrace (No. 4) was correlated with the advance of the Pleiscotene ice sheet which diverted the river in its lower reaches. The ages of the lower terraces are unknown.

Stratigraphic units within the level of the lowest valley fill were first observed in 1966 in a gully wall of Jones Creek in the south unit of Roosevelt Park (sec. 6, T. 140 N., R. 101 W.). Five alluvial units were identified; two are paleosols. These units were traced for a distance of some 100 miles north of the park, and are therefore assumed to occur throughout western North Dakota.

## DESCRIPTION OF THE ALLUVIAL SEQUENCE

The stratigraphic units were differentiated on the basis of position, color, grain size, and the presence or absence of bones. The paleosols were identified on the basis of color, carbonate content, organic material, clay content, and areal distribution.

The colors cited are those of dry samples obtained from the Munsell Soil Color Charts. The carbonate content of the sediment was determined in the laboratory using a volumetric analysis. The oxidizable material present in the sediment was determined using chromic acid as an oxidizer and back-titrating with a ferrous ammonium sulfate solution. The clay content includes all particles smaller than 0.004 mm and was determined by pipette analysis.

Unit A, the oldest unit (fig. O-1), contains a clayey-sand bed underlain by gravel, and constitutes the lowermost unit exposed. The unit is of variable thickness, but does not exceed 5 feet.

Unit B is a light gray to yellowish brown paleosol ranging in thickness from 1 to 2.5 feet, but has been highly eroded and is absent in most areas. Many trees buried by the overlying unit were observed to be rooted in this zone. Laboratory analyses (fig. O-2) indicate a greater clay content (up to 50 percent) and a lower carbonate content than adjacent older and younger units.

Unit C is from 8 to 15 feet thick and is light gray to very pale brown. Grain-size composition averages 20 percent sand, 50 percent silt, and 30 percent clay. The unit contains a greater amount of carbonate and organic material than Unit E. An abundance of bones have been found, the majority of which belong to the modern species of bison. One big-horn sheep skull and some gastropods (as yet unidentified) have also been found. This unit is present in most tributary valleys in western North Dakota and accounts for the bulk of the exposed valley fill.

Unit D is a brown to dark brown paleosol ranging in thickness from 0.5 to 2 feet, but is absent in places. Laboratory analyses indicate a higher clay content (up to 75 percent) than adjacent older and younger units and an increase in organic material. The carbonate content is 65 percent lower than the underlying unit, presumably indicating some leaching. The zone has a granular structure, although no soil horizons were differentiated. The upper 0.5 foot contains an abundance of plant remains; some living cottonwood trees rooted in this soil extend up through Unit E.

Unit E, the upper deposit, ranges in thickness from 1 to 10 feet, but is absent in places. The sediment averages 25 percent sand, 60 percent silt, and 15 percent clay and is pale yellow in color. This unit is lighter in color and contains less clay than Unit C. The unit is devoid of bison bones, but does contain small terrestrial and aquatic gastropods ( as yet unidentified). Only a faint soil zone is present on this surface; preliminary evaluation of laboratory data, however, indicates that there has been some leaching of carbonates and an accumulation of organic material.

## AGE OF UNITS

A buried tree rooted in Unit B was radiocarbon dated at less than 185 years B.P. (I-2325), indicating that units C through E, 15 to 20 feet thick, have been deposited since A.D. 1765.

The age of Unit A is unknown because no evidence has been found to indicate the time of deposition. The soil forming processes acting during the formation of Unit B ceased prior to about 1765. The character of Unit B indicates that following its formation the area underwent a period of degradation. Several trees rooted in Unit B and buried by Unit C were observed. The largest trees were about 2 feet in diameter and were probably well established before erosion began. The smallest trees observed were less than a foot in diameter and were estimated to be about 20 years old before burial. Unit B was present where these trees had established a root system capable of holding the soil together and thereby retarding erosion. In other exposures of Unit B, the local relief was probably not great enough to promote erosion. The age of the youngest trees present might therefore be used to approximate the length of time associated with the period degradation. Trees not established before degradation began would not likely have survived the erosion. Because the youngest buried trees observed were estimated to be 20 years old, the period of degradation preceeding the deposition of C was probably no longer than 20 years.

Following the erosion of Unit B, alluviation began, which resulted in the formation of Unit C. The beginning of this period was after 1775. Unit E, which directly overlies C in places, was deposited sometime after 1880. This date can be established because Unit E lacks bison bones, indicating that it formed after the 1880's when bison became extinct in this area (Robinson, 1966). The total amount of time involved in the deposition of Unit C can be further shortened by considering the time required for the development of the soil on its upper surface. By visible and chemical comparison of this soil with the faintly developed soil on the present surface, it is felt that at least three times the amount of time required for the present soil was required for the development of Unit D, assuming the same climate existed. Because the present surface has been exposed for approximately 30 years (see following paragraph), the period of time required for the formation of Unit D would be on the order of 100 years. When this figure is subtracted from the beginning of deposition of Unit E, which is about 1920, a date of about 1820 is obtained for the end of deposition of Unit C.

As was mentioned previously, the absence of bison bones in Unit E indicates that the earliest date for its deposition is sometime after the 1880's. However, the presence of living cottonwood trees rooted in Unit D, estimated to be approximately 40 to 50 years old indicate that the beginning of deposition of that unit was about 1920. The presence of 20 to 30 year old trees on the present surface indicates that deposition ended about 1935.

## DEPOSITIONAL RATES

Units C through E, which are 15 to 20 feet thick, probably formed between A.D. 1775 and 1935, a total of 160 years. However, the deposition of this sediment was interrupted for a period of approximately 100 years, during which time a soil formed and underwent a slight period of erosion. This leaves a period of approximately 60 years for the actual deposition of these sediments. The depositional rates indicated are on the order of 0.30 to 0.40 feet per year. These figures are of the same order of magnitude as the 0.34 feet per year calculated for the deposition of recent alluvium in a somewhat similar environment in southwestern Nebraska (Brice, 1966).

It is not likely that this deposition was constant for each year. Discoloration and fine banding in Unit C probably indicate times during which the deposition was much less for some years. Conversely, the deposition probably exceeded 0.4 feet per year at times. It is felt, however, that the range from 0.30 to 0.40 feet per year is representative of the average sedimentation rate occurring during this period.

## CORRELATION

In studies of the postglacial chronology of alluvial valleys in Wyoming, Leopold and Miller (1954) identified several formations ranging from Mankato to recent in age. These formations were correlated with similar units throughout the western United States. The most recent period of deposition began about A.D. 1200 and ended approximately 1880, resulting in the Lightning Formation. The recent valley fill of western North Dakota, possibly excluding Unit E, can be tentatively correlated with the upper part of the Lightning Formation.

In a study of erosion and deposition in Nebraska, Brice (1966) describes a sequence of recent alluvium which is divided into a banded upper part and a less banded lower part. The banding results from concentrations of clay and organic matter. A carbon-14 date from charcoal in the upper alluvium yielded a date of 420±160 years B.P., suggesting a correlation with the Lightning Formation and probably also with the recent alluvium of western North Dakota.

Any attempt at exact correlation between recent alluvial deposits should be done carefully, because the factors which cause aggradation in the individual valleys may have been important at different times. In a semiarid environment, only a slight change in the factors controlling either erosion or deposition may result in a local change in the rate of that process. A correlation of this type can at best group only similar occurrences which may have been independent of one another.

## GEOMORPHIC AND CLIMATIC IMPLICATIONS

The sequence of alluvial fills are related to either climatic fluctuations, animal activity, or a combination of both. The deposition of sediment prior to 1880 was most likely the result of a climatic shift because the influx of settlers and large scale grazing did not occur until after 1880.

Prior to the deposition of Units A through E in western North Dakota, the area underwent a cycle of stream incision, during which time most pre-existing valley fill was stripped from the valleys. The remnant of this incision is in the form of a basal lag gravel (lower Unit A) which overlies the Fort Union Group in all areas where present erosion has exposed it. This incision was probably caused by a greater than average amount of precipitation which may have caused the streams of the area to become unstable and to begin downcutting. This entrenchment may in part be associated with 39 years (1663 to 1702) of excess precipitation in North Dakota (Will, 1946). This correlation is strictly speculative as no age has yet been determined for Unit A.

Following this incision, a period of quiescence was reached, which was probably associated with a return to more normal precipitation. During this time the streams resumed a state of equilibrium with the environment. The quiescence resulted in the development of a soil (Unit B).

A climatic shift toward lesser precipitation, probably about 1775 caused a reduction in the slope cover with an increase in the sediment yield to the basins and a decrease in the stream flow to carry it away. The result was the deposition of 10 to 12 feet of sediment, here designated as Unit C.

A return to normal precipitation about 1820 caused an increase in the slope cover thus decreasing the sediment yield to the basins. This decrease in slope erosion brought an end to the period of alluviation and was followed by the formation of a soil (Unit D).

Gullying probably began in the lower reaches of some of the larger tributaries as early as the late 1880's. Unpublished field notes made in 1905 by A. G. Leonard, State Geologist, indicate that some gullying had begun at that time. However, the number of gullies described does not seem as great as is evident today.

Renewed valley filling in this century was again associated with a decrease in precipitation. Precipitation records for North Dakota have been evaluated by Bavendick (1952) who showed that beginning with the year 1920 and culminating in 1936, when Unit E was being deposited, western North Dakota underwent an almost continuous series of droughts. Records show that the yearly precipitation fell below the average of slightly more than 15 inches to as low as 5 or 6 inches in 1934. This period was accompanied by strong winds which caused much blowing and drifting of sediment. Bavendick (p. 17) cites measurements indicating that from 2 to 7 inches of soil was removed from many fields in 1934.

Size analyses of the sediment from Unit E indicate good sorting with the mean grain size in the 0.0625 to 0.031 mm range, which is characteristic of loess. In the exposures examined, any loess present had probably been reworked by water before final deposition.

The rapid sedimentation which occurred in western North Dakota is thought to be almost entirely controlled by climate. Because modern records show precipitation deficiencies of enough magnitude to eliminate most of the established vegetation, it is not inconceivable that similar fluctuations have occurred in the near past.

## SUMMARY

The late-recent alluvium of western North Dakota can be divided into five individual units, two of which are paleosols. By carbon-14 dating and the estimation of tree age it was established that the upper 15 to 20 feet of this sediment was probably deposited between 1775 and 1936. This 160-year period can be further shortened by approximately 100 years which were required for the formation and subsequent partial erosion of a soil developed between these units. An average rate of sedimentation on the order of 0.30 to 0.40 feet per year is believed to be characteristic of these units.

The established age of this alluvium makes possible a tentative correlation with similar alluvial units in Wyoming and Nebraska, which have been studied and dated by investigators in these areas.

The periods of valley filling in western North Dakota are associated with periods of sub-normal precipitation which decreased the slope cover, thereby increasing the sediment yield to the basins. Modern precipitation records indicate that such a period occurred from 1920 to 1936 during which time the upper unit was deposited.

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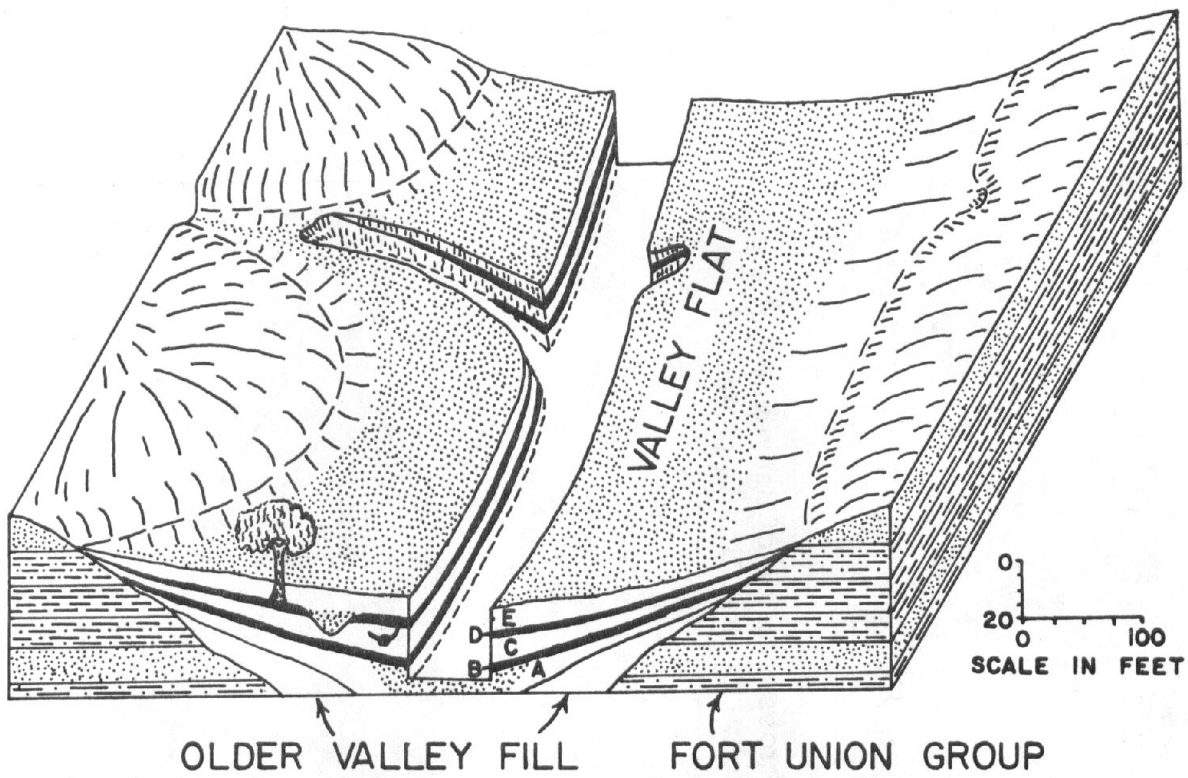



Figure O-1. Composite sketch showing the occurrence and relative positions of the alluvial units. Lettered units are described in text.

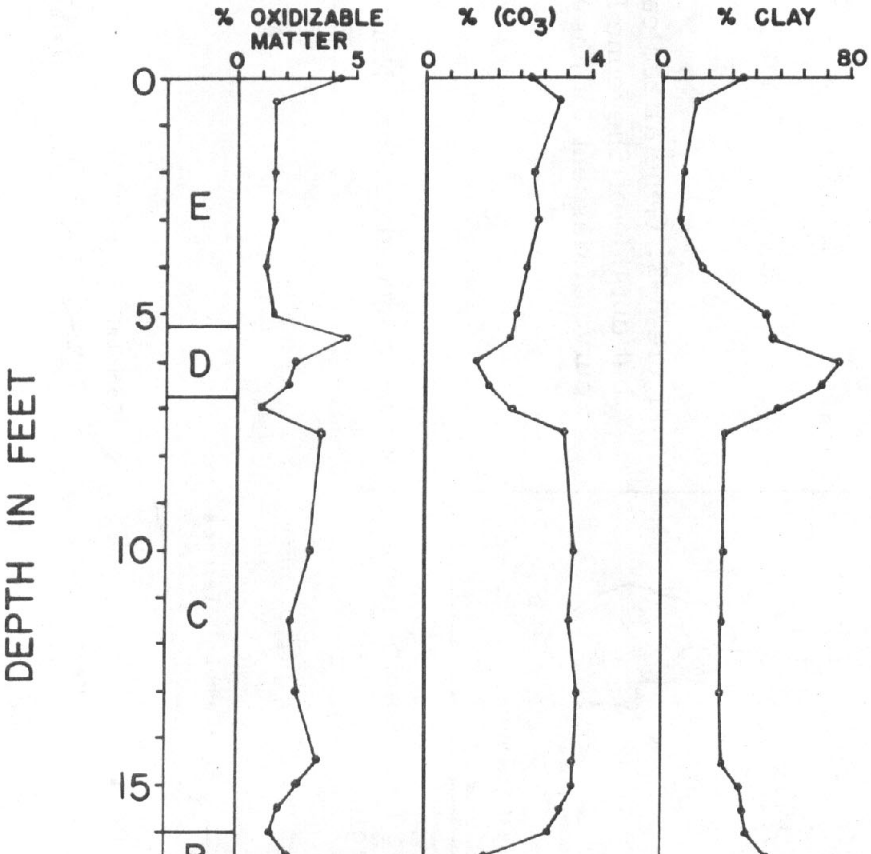
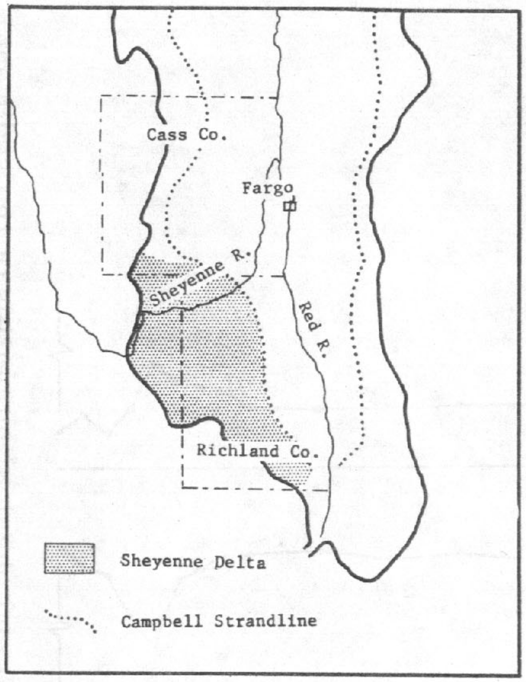



Figure O-2. Changes in the chemical and physical properties of the sediment with depth.



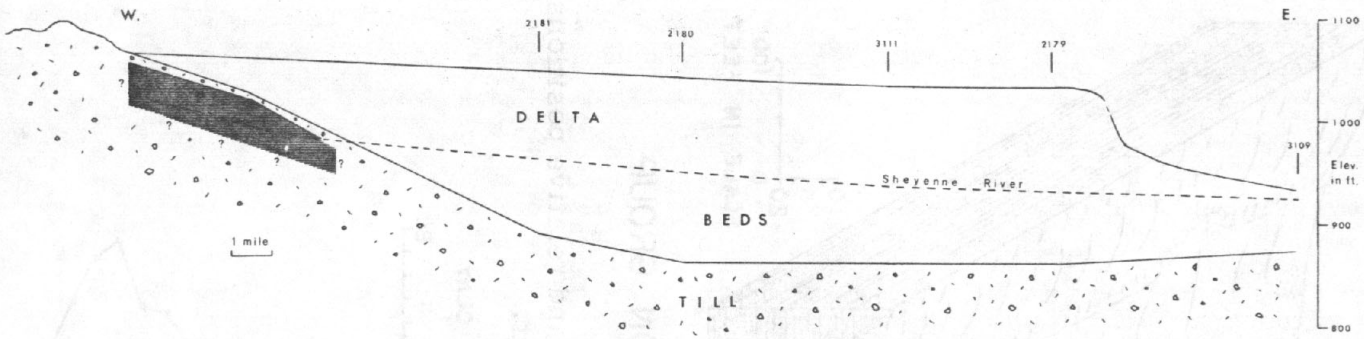




Figure P-2. Generalized east-west cross-section of Sheyenne Delta parallel to but outside of Sheyenne River trench. Dashed line is projected water-surface curve of descent for Sheyenne River.

Figure P-1. Map of south end of Lake Agassiz basin, North Dakota and Minnesota. Dotted lines mark basin margin.

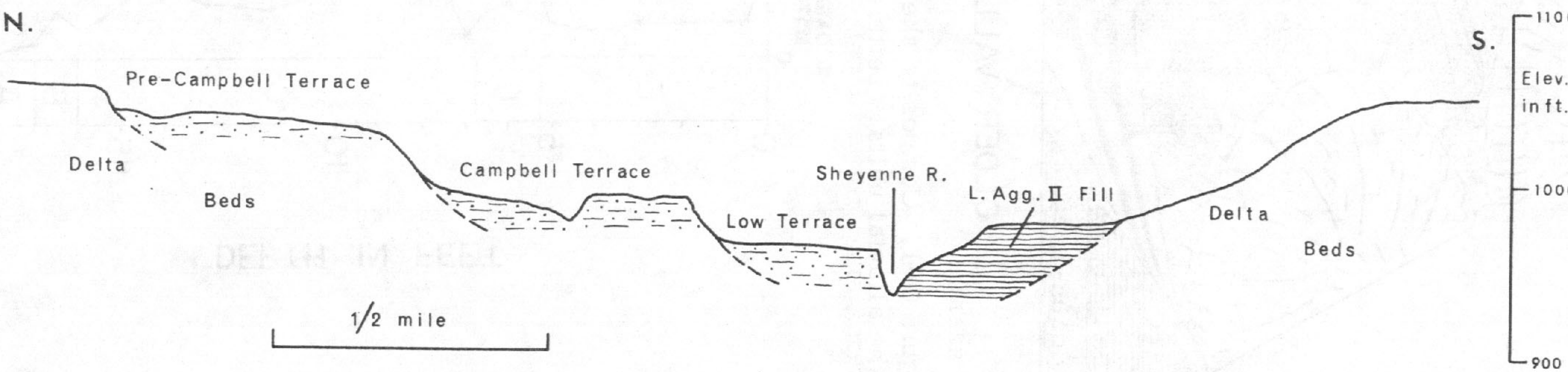



Figure P-3. North-south cross-section across Sheyenne River trench about 5 miles west of outer edge of delta.

30-P

# SOME ASPECTS OF THE GEOLOGICAL DEPOSITS OF THE SOUTH END OF THE LAKE AGASSIZ BASIN

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## LAKE DEPOSITS OF THE FARGO AREA

Through a combination of surface exposures and subsurface data, the stratigraphy of Lake Agassiz deposits in the vicinity of Fargo, North Dakota (fig. P-1) has become reasonably well known. The nature and thickness of these deposits is shown in the following composite section (1)* (from the surface downward):

1. Silt and clay, thin-bedded to laminated except in upper 10 feet where bedding is not evident (probably due to weathering), some minor, gently inclined cross-lamination, oxidized in upper 26 feet, unoxidized in lower 2 feet, no fossils observed------------------------------------ 28 ft.

2. Silt, laminated, unoxidized, containing abundant organic remains, predominantly wood, but including seeds, pollen, insect parts, and mollusk shells (2)*-------------6 in. to 2 ft.

3. Clay and silt, thin bedded, locally faintly laminated, unoxidized, no fossils observed------------------------- 3 ft.

4. Clay, massive, bedding obscure or lacking, unoxidized, highly plastic, no fossils observed, rests on unweathered drift, generally till---------------------------------- 70 ft.

Unit 4 is interpreted as a deep water Lake Agassiz I deposit, units 3 and 2 as shallow lacustrine and paludal accumulations of the Agassiz I-II interval and Unit 1 as the deposits of Lake Agassiz II. The wood of Unit 2,

* Notes at end of paper.

radiocarbon dated at about 9900 B.P. (W-388 and W-993), is believed to represent trees killed by the rising waters of Lake Agassiz II, thus providing a good date for the arrival of the transgressing lake.

A subaerial Agassiz I-II interval in the southern basin is indicated also by a desiccation zone in the upper part of Unit 4, a local disconformity between Unit 4 and overlying beds, and linear sand bodies (interpreted as fluvial deposits) intercalated between units 1 and 4. How far this rather simple stratigraphy extends beyond the immediate environs of Fargo is not precisely known. There are some indications, however, that the two major units (1 and 4) can be recognized within an area of at least 100 square miles around Fargo and possibly much farther out (3).

So far, no evidence of an earlier low-water phase, as recently proposed (4), has been found in the southern part of the basin.

## THE SHEYENNE DELTA

Origin.--Upham (5) and Leverett (6) disagreed on the source of sediments of the Sheyenne Delta which occupies about 800 square miles at the southwest corner of the basin (fig. P-1). Upham believed the sediments were supplied by the Sheyenne River to Lake Agassiz in true deltaic fashion.

Leverett felt that the great size of the Sheyenne Delta (and other Lake Agassiz deltas) suggested that the material was not entirely supplied by inflowing streams. "On the contrary," he says (7), "it appears probable that the greater part was contributed directly by the melting ice sheet as its border was receding across them. The Sheyenne area of sand has an abrupt northeast border, like an ice-contact face, rising 30 to 50 feet above the clayey plain to the northeast. The sand on this border is but a few feet thick and covers a highly calcareous, somewhat pebbly clay, which seems best interpreted as a glacial deposit in ponded water."

Upham's earlier view now seems strongly supported by the following recent findings:

1. Exposures, shallow borings and deep test holes along the steep northeast delta front show only thin-bedded to laminated fine sands, silts and clays beneath an eolian sand veneer (8). These deposits, which seem best interpreted as deltaic in origin, continue downward about 50 feet below the base of the delta front. Both they and the lake beds to the northeast lie upon a glacial drift surface which has about the same elevation under both (fig. P-2).

2. New topographic maps (contour interval 5 ft.) show that the delta surface declines gently from the Herman level on the west to the Campbell level on the east. This would be the expected gradient for a delta built by

the Sheyenne River entering Lake Agassiz from the west, but would be hard to explain as a depositional surface built by a northeastward-retreating ice front.

3. Map and field studies of the steep northeast-facing edge of the delta show that the morphology and elevation of the steepest break are consistent with origin as a Campbell wave-cut feature.

4. Though a detailed discussion would not be appropriate here, it now appears that Leverett's evidence for a lake-laid end moraine looping southward across the basin as a correlative continuation of his delta-edge ice-contact face (9) was tenuous at best, and that no truly reliable criteria exist along that line for a stationary ice-front.

Stratigraphy.--The stratigraphic sequence described below is the outgrowth of studies of surface exposures along the trench which the Sheyenne has excavated across the northern part of the delta, plus somewhat limited data from various borings and drill holes. Because of great distances between good exposures and a general lack of reliable criteria for correlation, the stratigraphy should be considered tentative.

The oldest unit so far recognized is a series of non-fossiliferous, thin-bedded to laminated fine sands, silts and clays known only from outcrops in the Sheyenne trench near the landward (western) edge of the delta. The texture and bedding of this unit and its position beneath a thin till suggest that it may be a pro-glacial lacustrine deposit (fig. P-2). If so, it may correlate with sub-till lacustrine deposits recognized elsewhere in the southern part of the Agassiz Basin (10). Its extent, thickness, and absolute age are not yet known.

Immediately above these beds in most outcrops is a thin till (fig. P-2), and in several places a north-south striated cobble pavement lies at the contact between the two. The till can be traced westward to the ground moraine behind the delta where it is the surface deposit, locally separated from a lower till by a similarly north-south striated cobble pavement. Eastward the thin till disappears beneath the floor of the Sheyenne trench, but it probably correlates with at least the upper portion of the till known from drilling farther east.

Next youngest are the deltaic deposits, or perhaps more accurately, a deltaic-lacustrine complex, which forms an eastward thickening wedge of sediments lying on the till (fig. P-2). Typically, these sediments are non-fossiliferous, well-bedded (often thin-bedded or laminated), well-sorted, sands, silts, and clays. At the head of the delta, where the glacial Sheyenne entered, these beds coarsen to gravel and coarse sand.

Outside of the Sheyenne trench the delta beds, or eolian sands derived from them, form the surface. Discounting the eolian increment, this surface, traced northeastward from the delta head, slopes gently downward from the elevation of the Herman Beach (about 1070 feet) to its lakeward terminus in an escarpment at about the Campbell Beach level (about 990 feet). The delta beds thus were deposited during the early high stages of the lake and can therefore be correlated with the lower part of the massive clay unit (Unit 4) of the lake deposits near Fargo.

Eolian sands derived from the sandy delta beds form a cover over much of the delta plain and parts of the Sheyenne trench as well. As a stratigraphic unit the eolian beds are strongly time-transgressive, for while the most intense eolian action was probably early in the subaerial history of the delta (before vegetation was established), new eolian deposits are still being derived from locally exposed sandy delta beds. Morphologically, the sand cover varies from a thin, low-relief veneer to highly irregular hills rising as much as 60 feet above the delta surface. Locally, a paleosol is found in the top of the delta beds beneath the eolian sand, and multiple paleoregosols occur within the eolian sequence itself.

Within the Sheyenne trench and restricted to it, there are several additional stratigraphic units all of which are younger than the deltaic sequence (fig. P-3). A series of fluvial deposits underlies a series of terraces standing at various levels above the present river. These terraces probably represent the attempts of the Sheyenne river to regrade the delta to the various base levels imposed by the fluctuation of Lake Agassiz, and the best developed of them is graded to the Campbell level. The fluvial deposits which veneer these terraces have a molluscan fauna and vary in texture from gravel to clay.

At two places in the trench, thin-bedded to laminated fine sand, silt and clay units were found which resemble the deltaic-lacustrine beds except that they contain a sparse fauna of small mollusks. At the top of one of these sequences there is a marly peat (11) wood from which has been dated at 9130 ± 150 years B.P. (I-1982), a date which is close to the time that Lake Agassiz II is presumed to have dropped from the Campbell level. Both of the known occurrences of these beds lie below the Campbell level in the trench. Considering their position, their nature and the C-14 date, these beds are interpreted as a deltaic-lacustrine fill laid down in the "estuary" created by invasion of Lake Agassiz II water into the deep trench cut by the Sheyenne River during the Agassiz I-II low-water interval. If this interpretation is correct, they would correlate with the lower part of the silt and clay unit (Unit I) in the lake sequence near Fargo.

The youngest beds in the trench underlie a low terrace the top of which is about 18-20 feet above normal stream levels. This terrace apparently was a flood plain until a few hundred years ago. It is seldom reached by modern floods, which in general are confined to both the main channel and a series of deep meander-scar channels which cut the low terrace. As shown by C-14 dating, the upper 20 feet of beds under this terrace was deposited since about 2500 years ago (W-1185, 2540 ± 300 B.P., wood from about 20 feet below top of terrace), and deposition of sediments near the top of the terrace has occurred as recently as about 235 years ago (I-2093, 235 ± 90 B.P., charcoal from 1 foot below top of terrace). Lithologically this latest increment of deposition is largely fine sand, silt and clay.

<u>History.</u>--The following sequence of geologic events is proposed for the delta area:

1. A proglacial lake was ponded between a southward-advancing late-Wisconsin ice front and high ground to the south, with the ice eventually overriding the lake deposits.

2. Ice retreat reopened the southern Lake Agassiz basin to lake-water accumulation, and the Sheyenne River deposited a delta during the Herman, Norcross and Tintah stages.

3. During Campbell I stability, the Sheyenne River trenched the delta until it was graded to the Campbell level. Concurrently, the northeast edge of the delta was steepened by wave attack.

4. During the Lake Agassiz I-II sub-aerial phase, the Sheyenne River adjusted to its lower base level by further trenching, probably to below present trench-floor levels.

5. The rising waters of Lake Agassiz II invaded the trench and a lacustrine deltaic complex was deposited. Local peat accumulation marked the end of this phase as Lake Agassiz fell from the Campbell II stand about 9100 years ago.

6. The Sheyenne River again cut downward in its trench adjusting to a falling base level and removing much of the Lake Agassiz II fill.

7. At least one aggradational episode has resulted from post-Agassiz readjustment of the Sheyenne, with fluvial filling of a somewhat deeper trench up to the level of the present low terrace which bounds the stream.

8. At present the Sheyenne appears to be slowly dissecting the low terrace.

9. Eolian activity, which probably began as soon as sandy delta beds stood above water, continues at present, primarily as blow-out activity in delta beds and previously deposited eolian sands.

Study is continuing on the nature of the various sedimentary units and their organic contents.

## NOTES AND REFERENCES

(1) The detailed description of Units 1, 2 and 3 and the upper 10 feet of Unit 4 was made from a Corps of Engineers high-water diversion cut on the Red River about 1 mile north of Fargo. The remainder of Unit 4 is known from large diameter auger holes dug to the top of the fill for emplacement of cast-in-place concrete pilings for high-rise buildings in Fargo.

(2) C. O. Rosendahl identified 39 plant species from an older exposure of this unit (Ecology, v. 29, pp. 284-315, 1948). A more recent paleobotanical study of this unit by J. H. McAndrews is in press in the proceedings of a conference on environmental studies held at the University of Manitoba, November, 1966. The insect remains and mollusk shells recognized in this unit have not yet been identified.

(3) See for example Dennis, Akin, and Jones, Ground Water in the Kindred Area, Cass and Richland Counties, North Dakota, North Dakota Ground-Water Studies 14, 1950, and Paulson, Q. F., Ground Water in the Fairmount Area, Richland County, North Dakota and Adjacent Areas in Minnesota, North Dakota Ground-Water Studies 22, 1953.

(4) Clayton, Lee, Notes on Pleistocene Stratigraphy of North Dakota, North Dakota Geol. Survey Rept. Inv. 44, 1966.

(5) Upham, Warren, The Glacial Lake Agassiz, U. S. Geol. Survey Mon. 25, 1898.

(6) Leverett, Frank, Quaternary Geology of Minnesota and Parts of Adjacent States, U. S. Geol. Survey Prof. Paper 161, 1932.

(7) Leverett, Frank, op. cit., pp. 126, 127.

(8) In my studies of many surface exposures and shallow auger holes, none of Leverett's "pebbly clay" was found, though the deltaic beds frequently have abundant pebble-size calcareous concretions. For deep hole data, see Baker, C. H., Geology and Ground Water Resources of Richland County, Part 2, North Dakota Geological Survey Bull. 46, 1966.

(9) Leverett considered this moraine a lake-laid extension of his Erskine moraine. On Glacial Map of the United States East of the Rocky Mountains, it is labeled "Wahpeton."

(10) Sub-drift silt and clay units interpreted as possible older lake beds have been described from northern Cass County (Brookhart, J. W., and Powell, J. E., Reconnaissance of Geology and Ground Water of Selected Areas in North Dakota, North Dakota Ground-Water Studies 28, 1961), central Cass County (Dennis, P. E., Akin, P. D., and

Worts, G. F., Geology and Ground-Water Resources of Parts of Cass and Clay Counties, North Dakota and Minnesota, North Dakota Ground-Water Studies 11, 1949), and in southern Richland County (see Paulson, Q. F., in note 3 above).

(11) For a paleobotanical analysis of this peat from the Mirror Pool Site, see McAndrews, reference cited in (2) above (in press).

(12) The support of the National Science Foundation (Grant GP-1347) is gratefully acknowledged.

30-Q

# A NOTE ON PREGLACIAL DRAINAGE IN THE NORTHERN GREAT PLAINS

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This short note is written to call attention to an idea regarding preglacial and glacial drainage which is not necessarily entirely new but may be worthy of further work. One of the main purposes of this note is to attempt an explanation of the width of the Minnesota River trench (generally 1 to 2 miles in width) and the presence of cross valleys in North Dakota which are oriented at high angles across known preglacial streams. At the present time the North Dakota Geological Survey, the Department of Geology of the University of North Dakota, the State Water Commission, and the U. S. Geological Survey are actively gathering a considerable amount of data which will be useful in elucidating this problem further.

A glance at the diagrammatic map (fig. Q-1) will show that in preglacial time rivers flowed northeastward across the northern plains to Hudson Bay. The portion of the drainage system which is outside the outermost glacial boundary is still in existence. It is also obvious that these northeastward-flowing rivers must have been blocked and had to seek other outlets when the ice advanced from the north. A few of the more prominent Late Pleistocene ice margin positions indicating still stands of the ice front are indicated on figure Q-1.

This blocking of the northeastward flowing streams forced the normal runoff water plus glacial meltwater to seek other outlets, which were largely (although not necessarily entirely) in a southeastward direction by way of the Minnesota River and the James River. The Minnesota River is the most logical first-used southeastern outlet as a result of low escarpments now partially obliterated by drift cover. Such escarpments are the Alta Ridge just east of Valley City, now covered by the Luverne end moraine (approximate equivalent of the Big Stone moraine), and the Coteau des Prairies. As the ice moved over these escarpments the water was forced into the James River valley. Finally when the ice advanced over the Missouri Escarpment, the water was forced up on the Missouri Plateau. There it probably formed the ice margin channels west of the present Missouri River (the Heart River-Cannonball River complex of channels and others). Later the present Missouri River was formed—perhaps even as the result of a much later advance.

There are several lines of evidence which would substantiate this line of reasoning. Consider the width of the Minnesota River trench, especially where it cuts through very resistant Precambrain rocks. Surely it took a stream of considerable magnitude some time to do this. True, much of this cutting has been attributed to the water released by the southward draining Lake Agassiz, but I suggest that probably much of the cutting may have been done by ice-diverted streams when Lake Agassiz did not exist; the Lake Agassiz runoff merely modified an already sizeable trench.

That there was an abundance of water to cut these valleys needs little elaboration when the amount of territory being drained by the Yellowstone, and Missouri and Saskatchewan Rivers is considered. In the general drainage area under consideration there are about 500,000 square miles. If Schumm (1965, p. 787) is correct, the runoff in the northern plains during "pluvial times" was 5 times present values. Presently the runoff in most of this area averages about 1 inch and about 10 inches in the high mountains (De Wiest, fig. 2-35). Using this data, 5 inches of runoff could, therefore, have been expected at times. This would result in about 100,000 second feet over the entire area. To this must be added additional runoff from the mountains not included in this figure, plus the meltwater from the glacial ice. This would at least double the runoff and thus the figure would be approximately 200,000 second feet. This is the equivalent of the main stem of the Mississippi River. Such a discharge would require a channel only 2000 feet wide (Leopold, Wolman, and Miller, 1964, fig. 7-21).

The channels such as the Minnesota River and the James River, as well as the cross valleys, have much wider channels than these. This indicates that the above rough computations are on the conservative side, and that more discharge water was available than indicated above to cut the trenches under consideration.

Another line of evidence is the presence of buried southeast-trending cross valleys in North Dakota; one is shown diagrammatically on figure Q-1. These cross valleys cut the preglacial valleys at nearly right angles and are commonly filled with glacial outwash, which makes them excellent potential aquifers. See papers 30-J and 30-L in this guidebook. These cross valleys are most logically explained as ice-marginal channels which existed mainly during the advancing phase of the ice sheet.

Thus I visualize the sequence as follows:

1. Northeast-flowing streams in preglacial time.

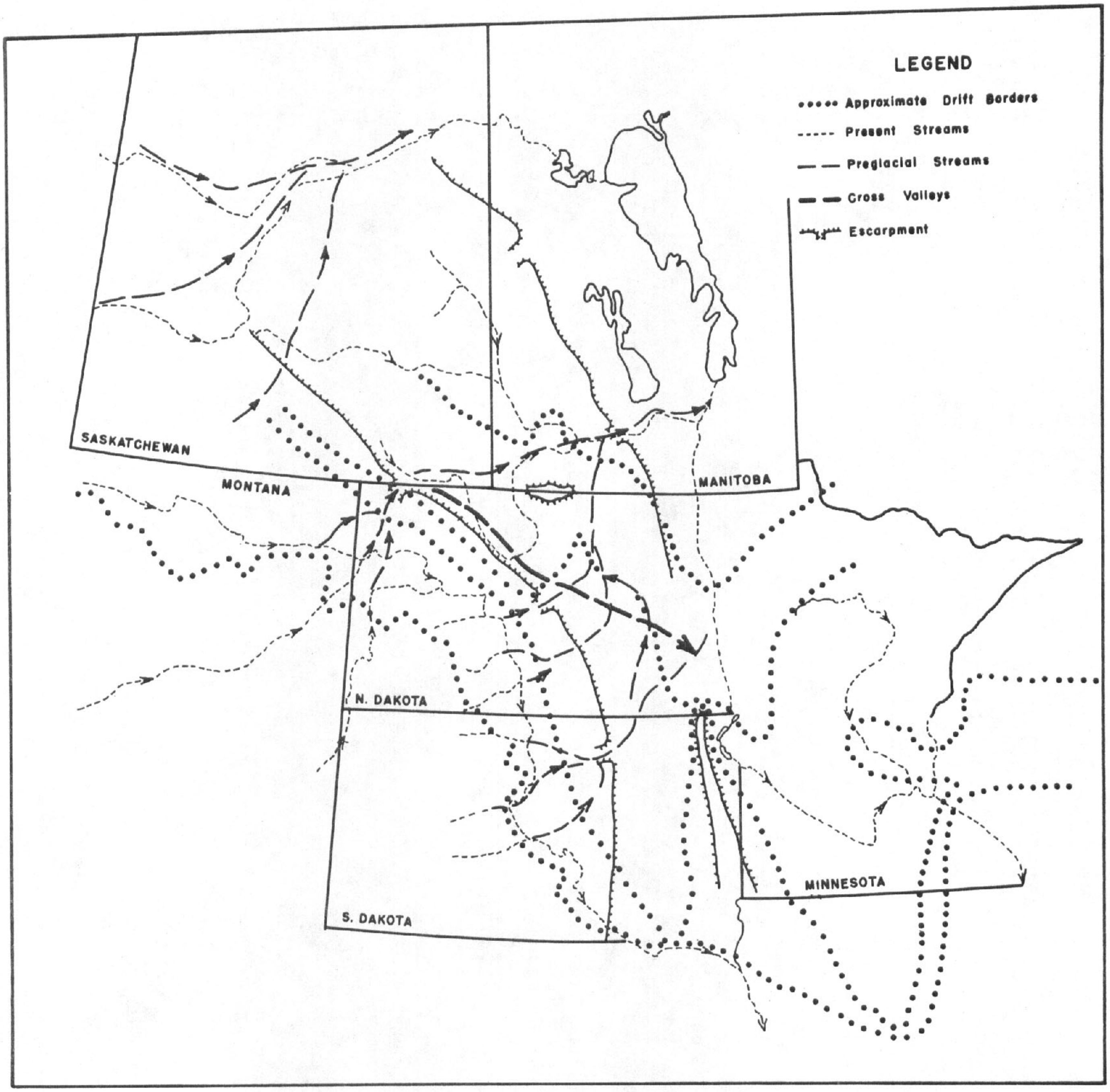


Figure Q-1. Map of northern Great Plains showing Pleistocene drainage changes.

2. Interruption of the northeastward-flowing streams by the advancing ice.

3. Diversion of the streams plus meltwater in a southeast direction largely by way of the Minnesota River.

4. Continued advance of the ice covering and filling cross channels with drift and diversion into the James River.

5. Formation of the ice-margin channels west of the Missouri River.

6. Formation of the Missouri River in its approximate present course.

7. Recession of the ice sheet to a series of positions east and north of the Missouri Coteau with the possible new establishment of southeasterly drainage by meltwater from the receding ice sheet.

This short note has simply indicated the existence of a fascinating problem. Some of the information needed to resolve it include more subsurface data to allow the construction of a more detailed bedrock contour map. Particularly is this needed in southeastern North Dakota, northeastern South Dakota and southwestern Minnesota. More detailed information is needed on the configuration of the Minnesota River trench, especially in southwestern Minnesota.

Because of their significance as aquifers, more accurate delineation of the cross channels in North Dakota is needed. They will tell us much about the history of the drainage, especially in pre-ice advance time and during the ice advance itself.

Detailed examination of the sediments found in these buried cross channels should be made to check source areas of sediments. If the Missouri River and its tributaries did use these channels, "western"-type gravels should be found in some abundance.

Very little study has been made of the distributary channels west of the present Missouri River. The sediments of these channels may tell much of the source of the waters that made them. Possibly datable materials may be found in them that will give more accurate time sequences than we now have. Paleontologic studies of the mollusks known to be present in these deposits may be of assistance in determining time and possibly climatic conditions of the area when the streams occupied these channels.

I wish to acknowledge the help of Lee Clayton, Theodore Freers, and John Bluemle in the preparation of this note.

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Leopold, L. B., Wolman, M. G., and Miller, J. P., 1964, Fluvial processes in geomorphology: San Francisco, W. H. Freeman and Company, 521 p.

Schumm, S. A., 1965, Quaternary paleohydrology in Wright, H. E., and Frey, D. G., The Quaternary of the United States: Princeton University Press, p. 783-794.

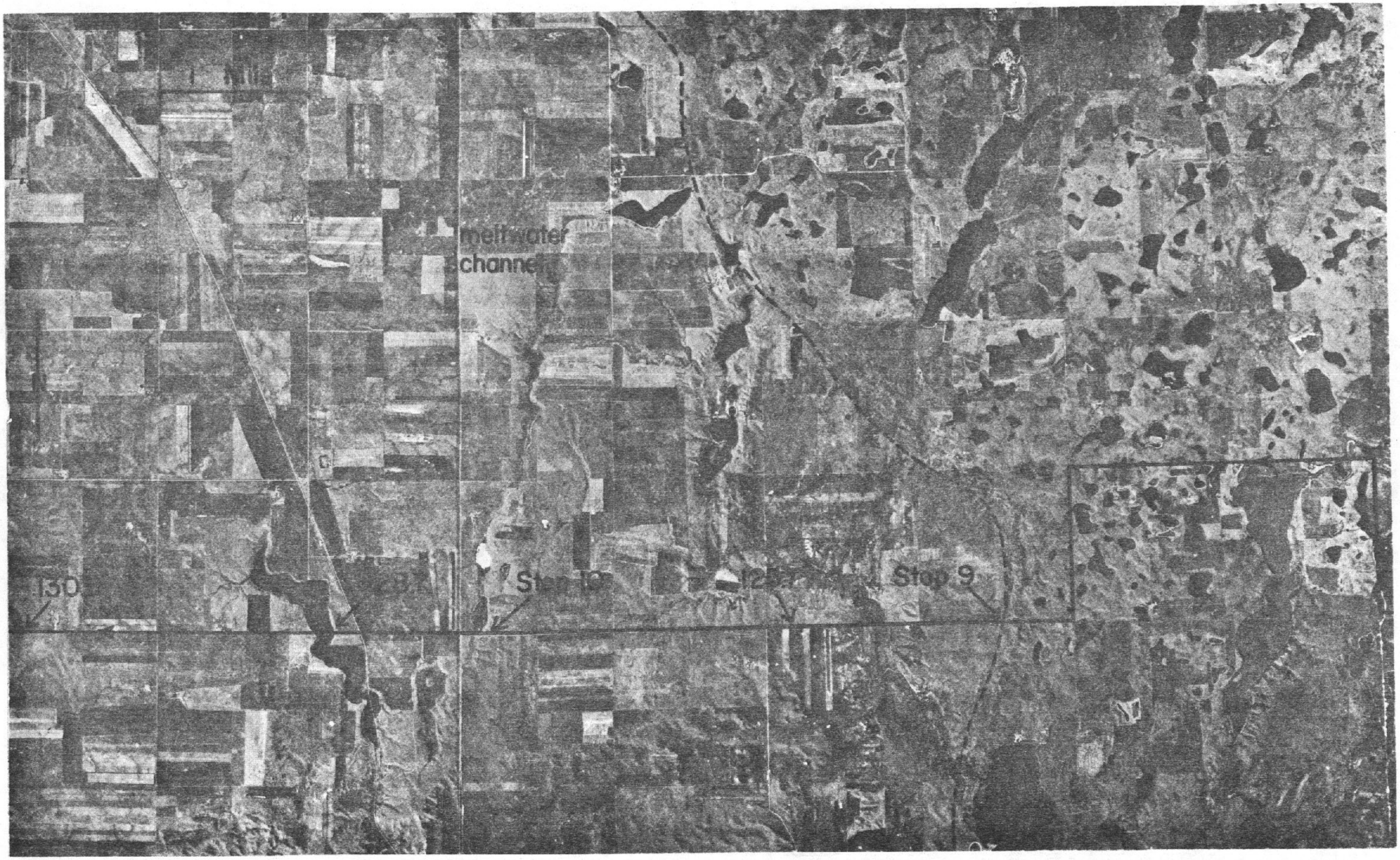



Figure R-18. Air photo of contact between Missouri Coteau (nonintegrated drainage) and the Coteau Slope (integrated drainage) with fieldtrip route, mileages, and stops indicated. Area is northwest-central part of fig. R-17. (Army Map Service BE M 5 520 and 521 28 July 52.)

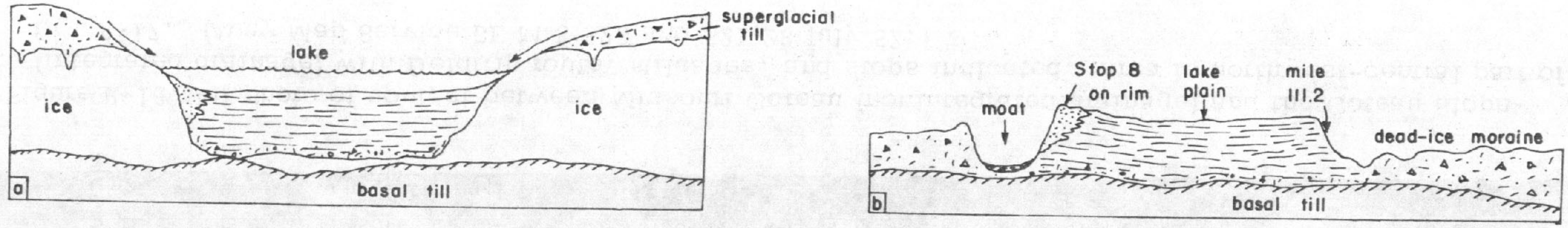



Figure R-19. East-west cross-sections showing formation of moat around ice-walled-lake plain at stop 8. a: Latest Wisconsin time. b: Today.

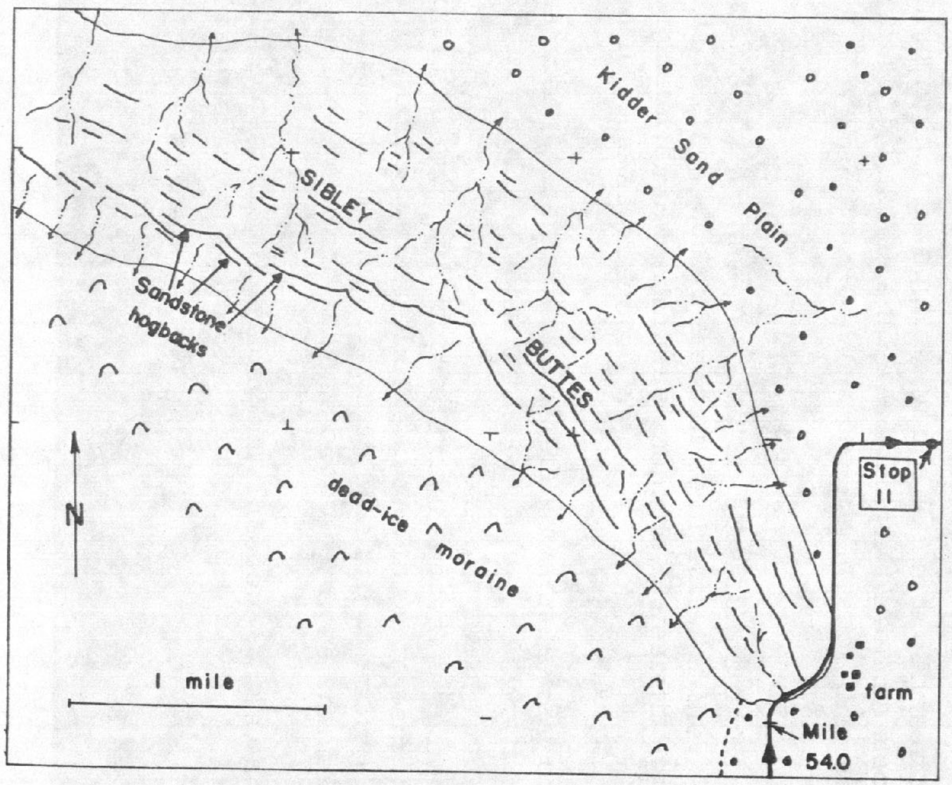



Figure R-20. Sibley Buttes and the sandstone hogbacks. Map location given in fig. R-2.

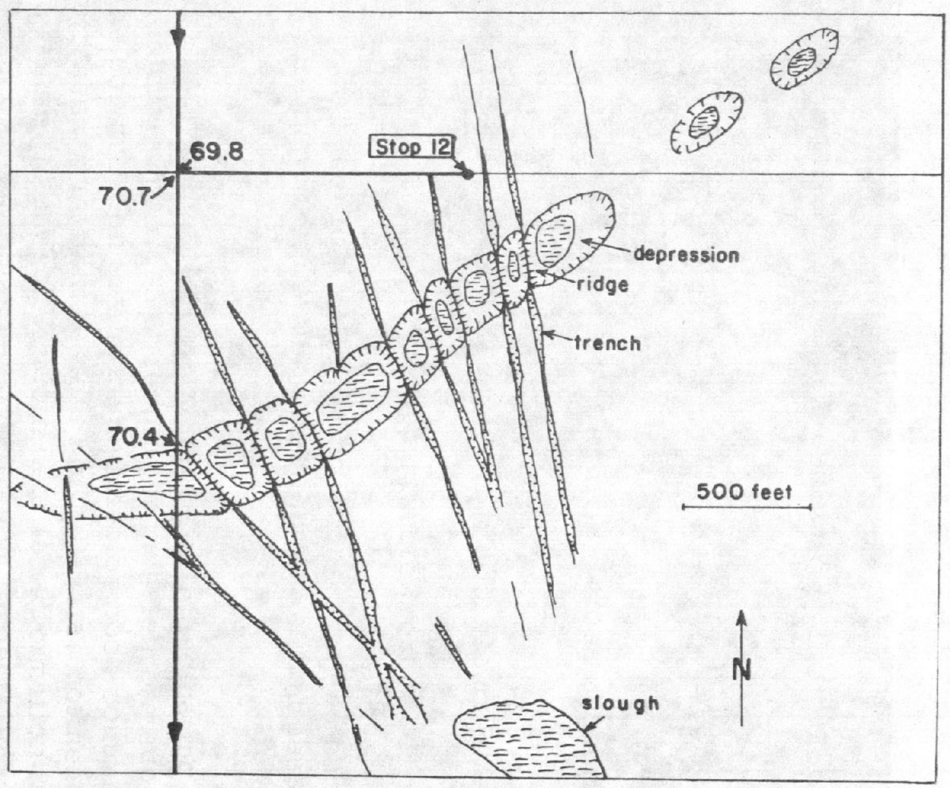



Figure R-21. Disintegration trenches changing to ridges where they cross a linear depression at stop 12.

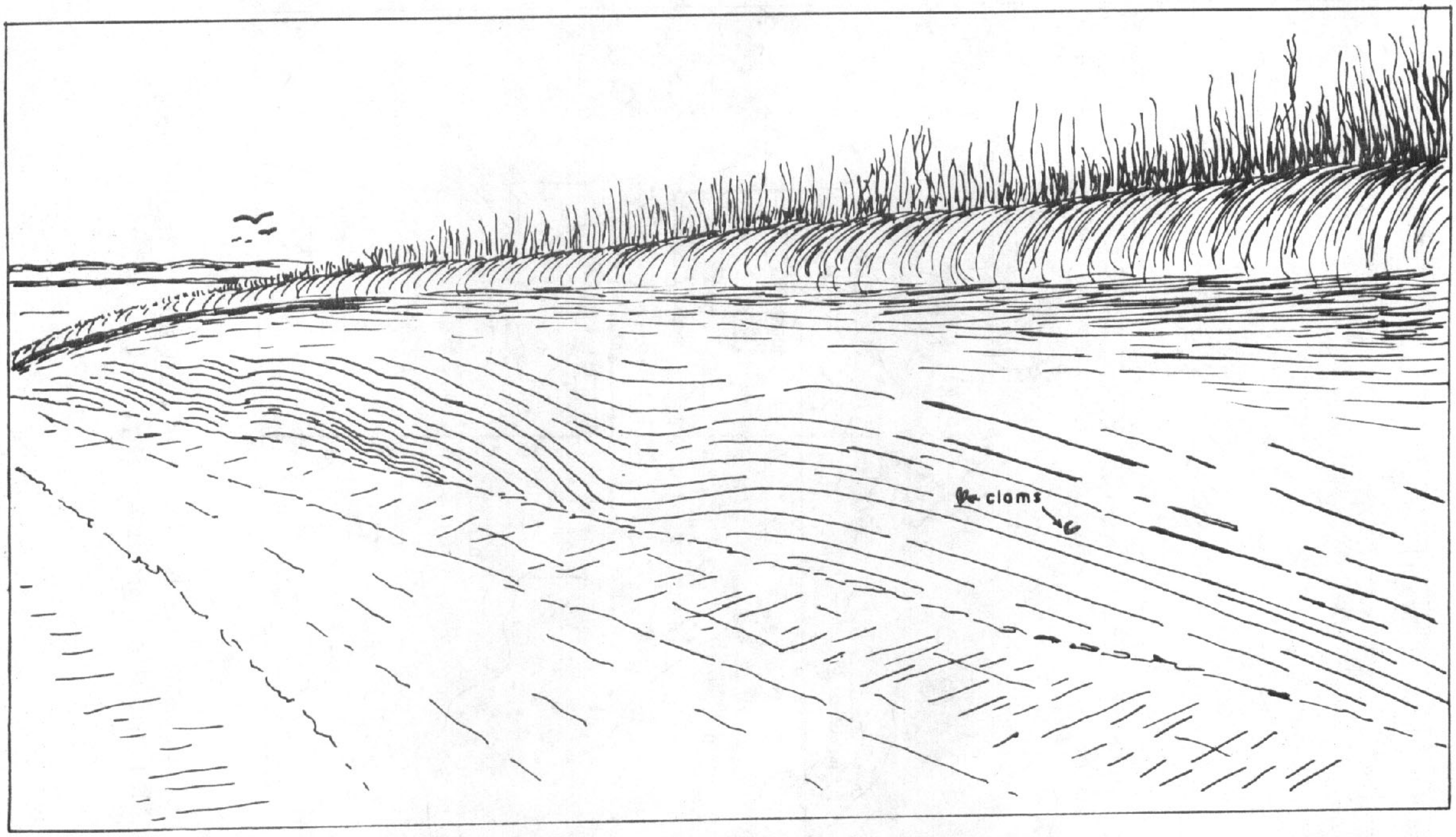



Figure R-14. Lowenthall Site: folded collapsed lake sediment exposed in vertical 4-foot roadcut in 1961 at southeast corner of SW 1/4 sec. 16, T. 132 N., R. 69 W., 2 1/2 miles southeast of Lehr. From a photo.

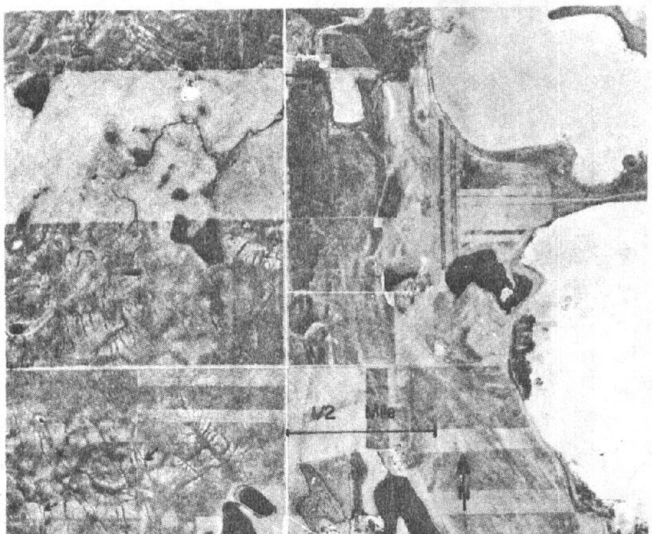



Figure R-16. Air photo of typical collapsed superglacial alluvium with disintegration trenches (arrows) around stop 6 on fig. R-17. (U.S. Dept. of Agriculture BAD-2K-19 7-3-52.)

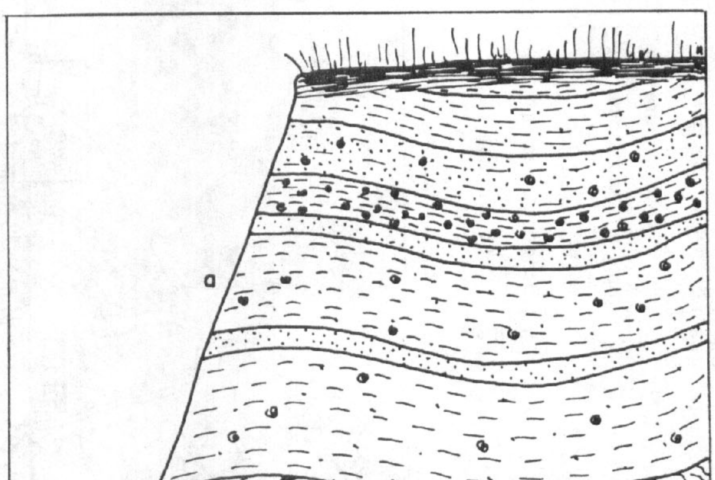



Figure R-15. Mummy Cat Site: slightly-folded collapsed lake sediment over contorted collapsed lake sediment exposed in 8-foot vertical roadcut in NW 1/4 NW 1/4 sec. 14, T. 132 N., R. 69 W. (fig. R-12).

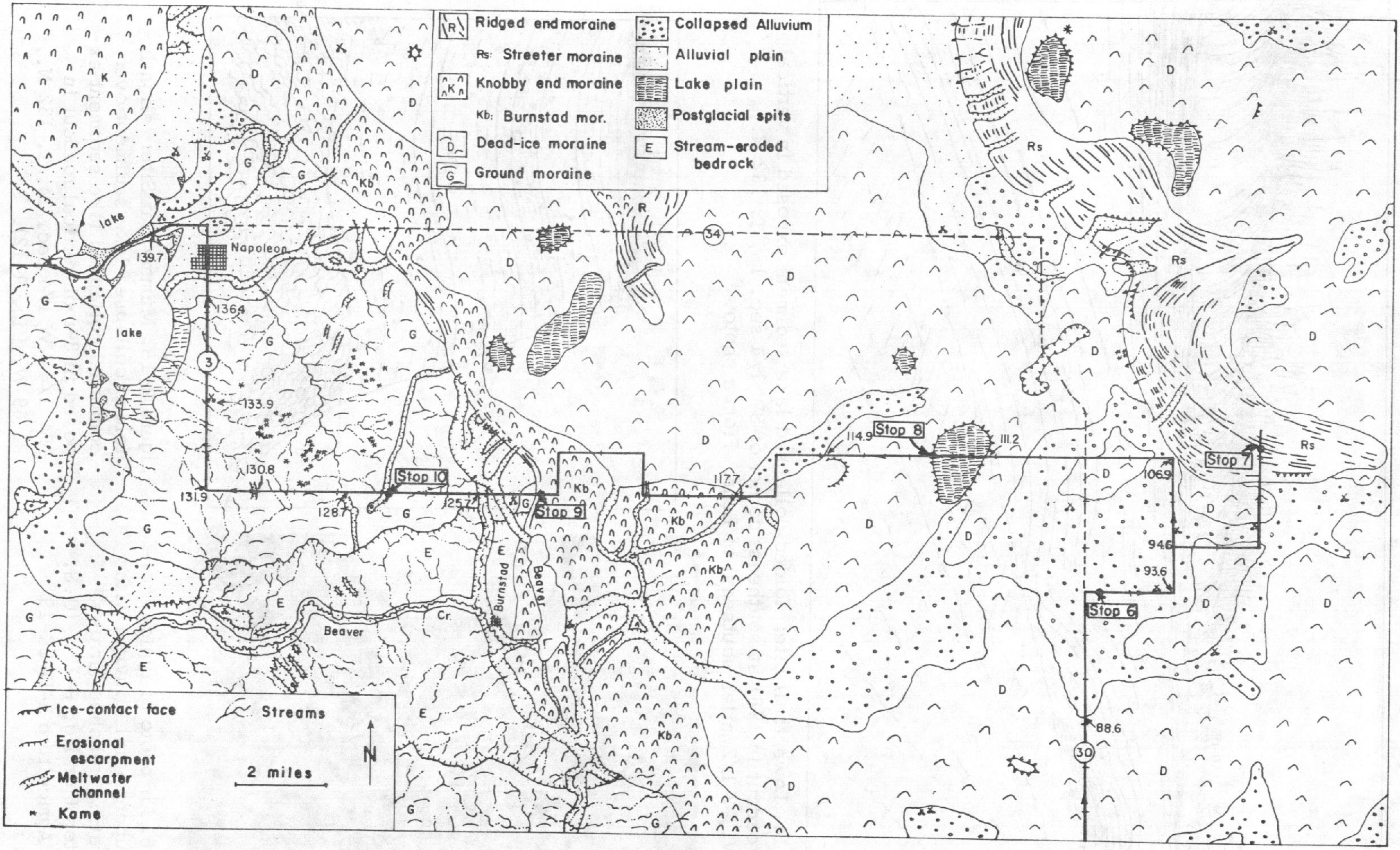



Figure R-17. Glacial geology along fieldtrip route in the area east of Napoleon, with mileages and stops indicated; mile 86.0 to 142.8. Dashed line is wet-weather alternate route. Map location given in fig. R-2.

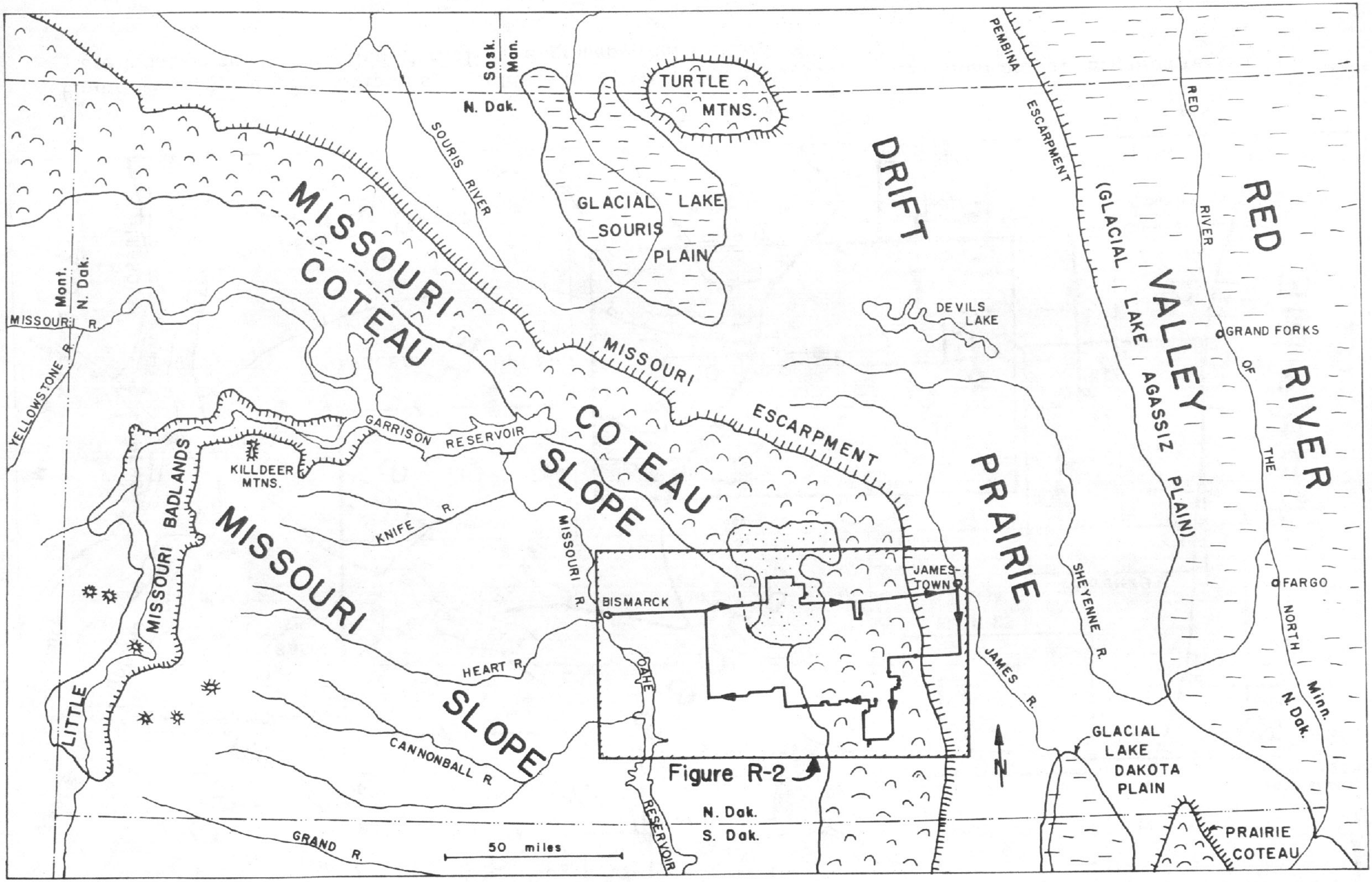



Figure R-1. Topographic divisions of North Dakota. Hachured rectangle is fieldtrip area, fig. R-2.

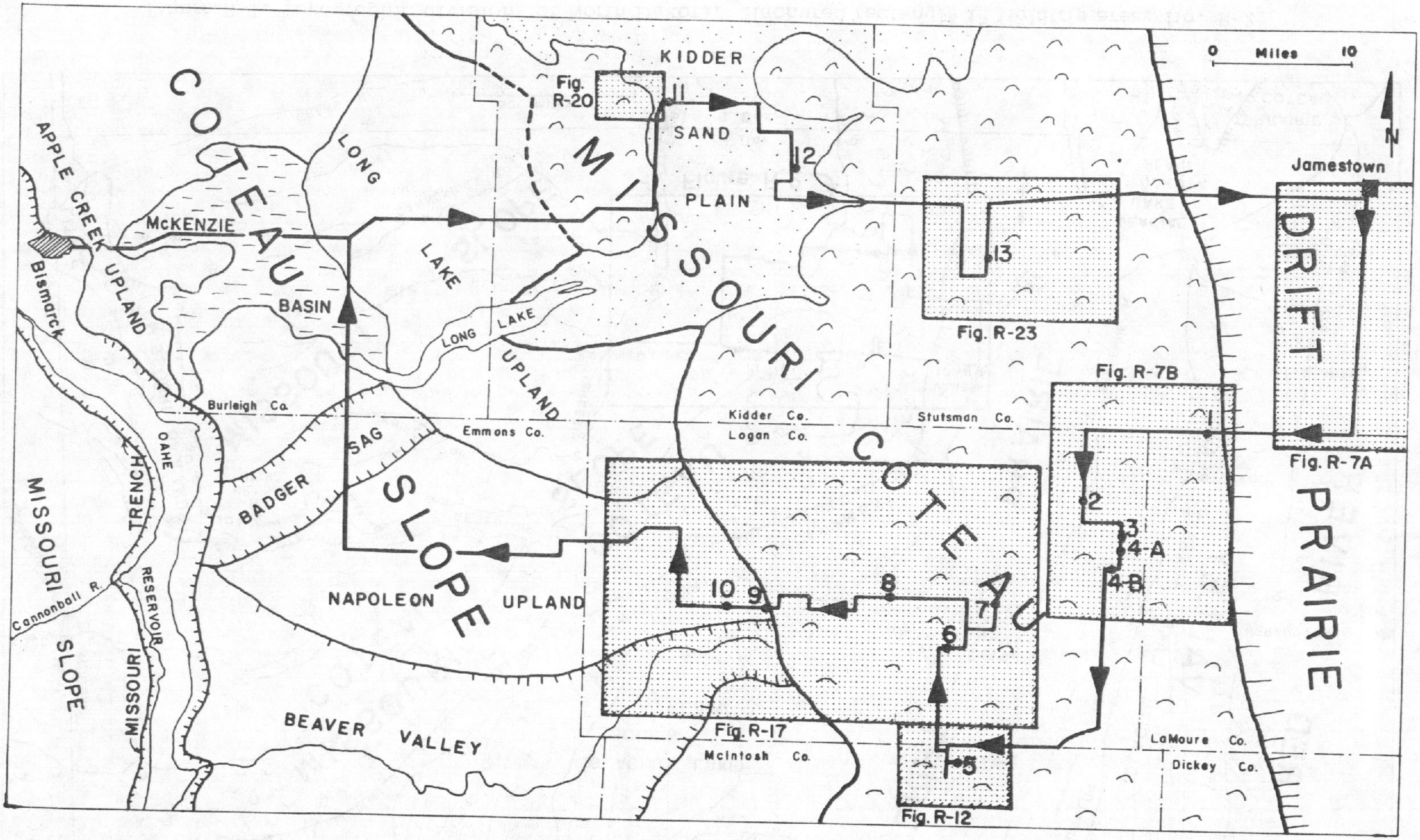



Figure R-2. Topographic divisions of fieldtrip area, with fieldtrip route and numbered stops. Stippled rectangles are locations of the detailed glacial-geology maps that follow.

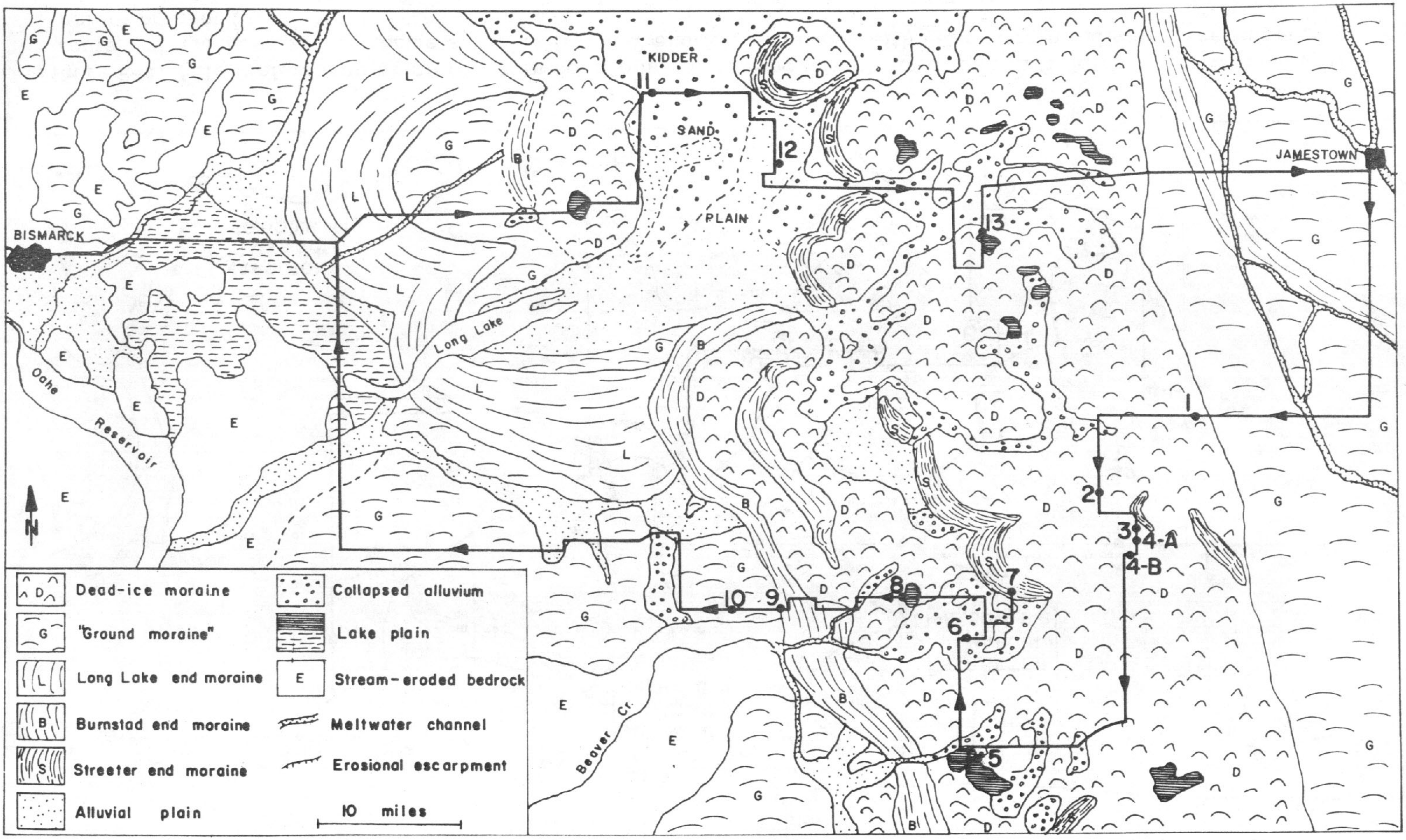



Figure R-3. Generalized glacial geology of fieldtrip area, with fieldtrip route and numbered stops. Same scale as fig. R-2.

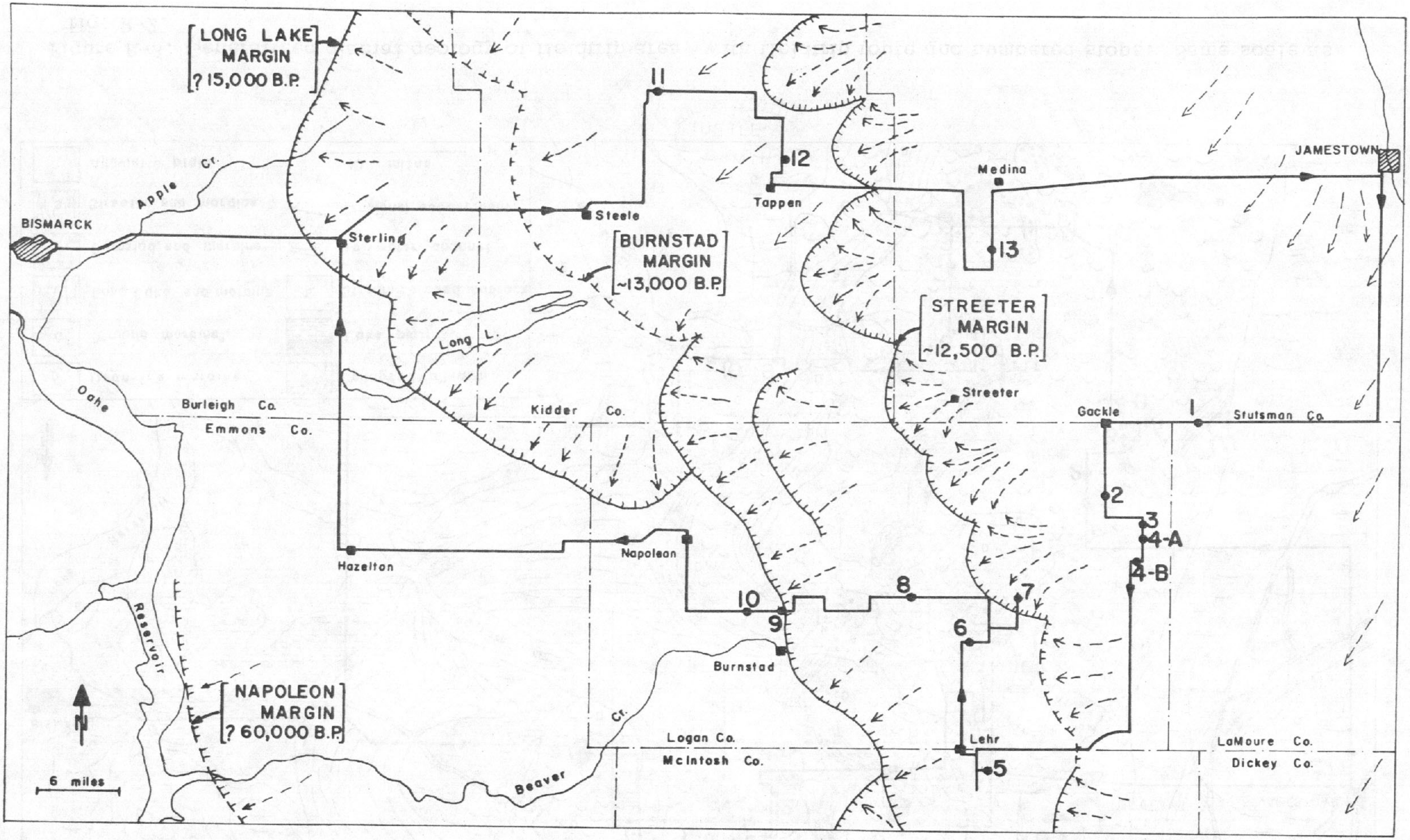



Figure R-4. Late Pleistocene glacier-margin positions and ice-movement directions in fieldtrip area, with fieldtrip route and numbered stops. Same scale as fig. R-2 and R-3. Number with each margin indicates age in years before present (B.P.).

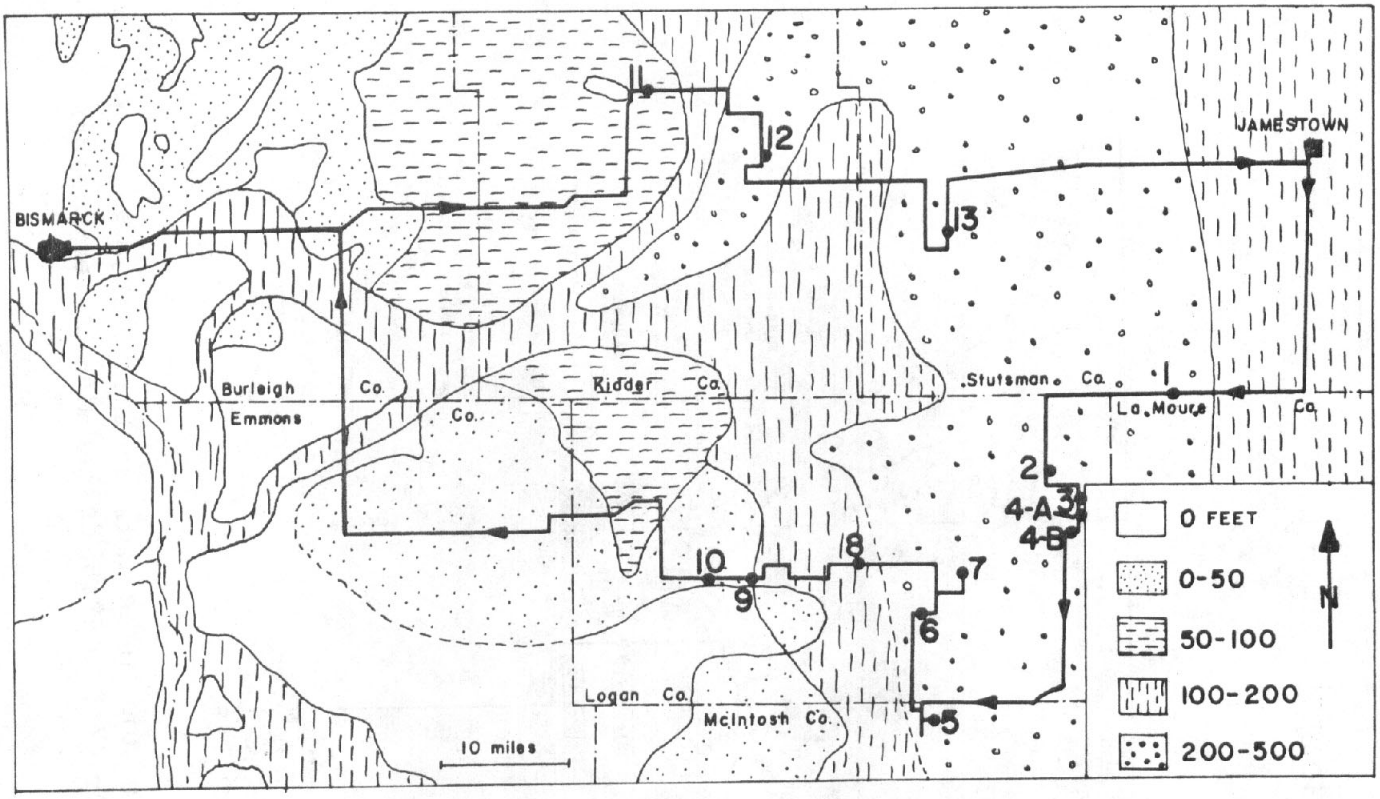



Figure R-5. Approximate drift thickness in fieldtrip area, with fieldtrip route and numbered stops. Same area as shown in fig. R-2 through R-4.

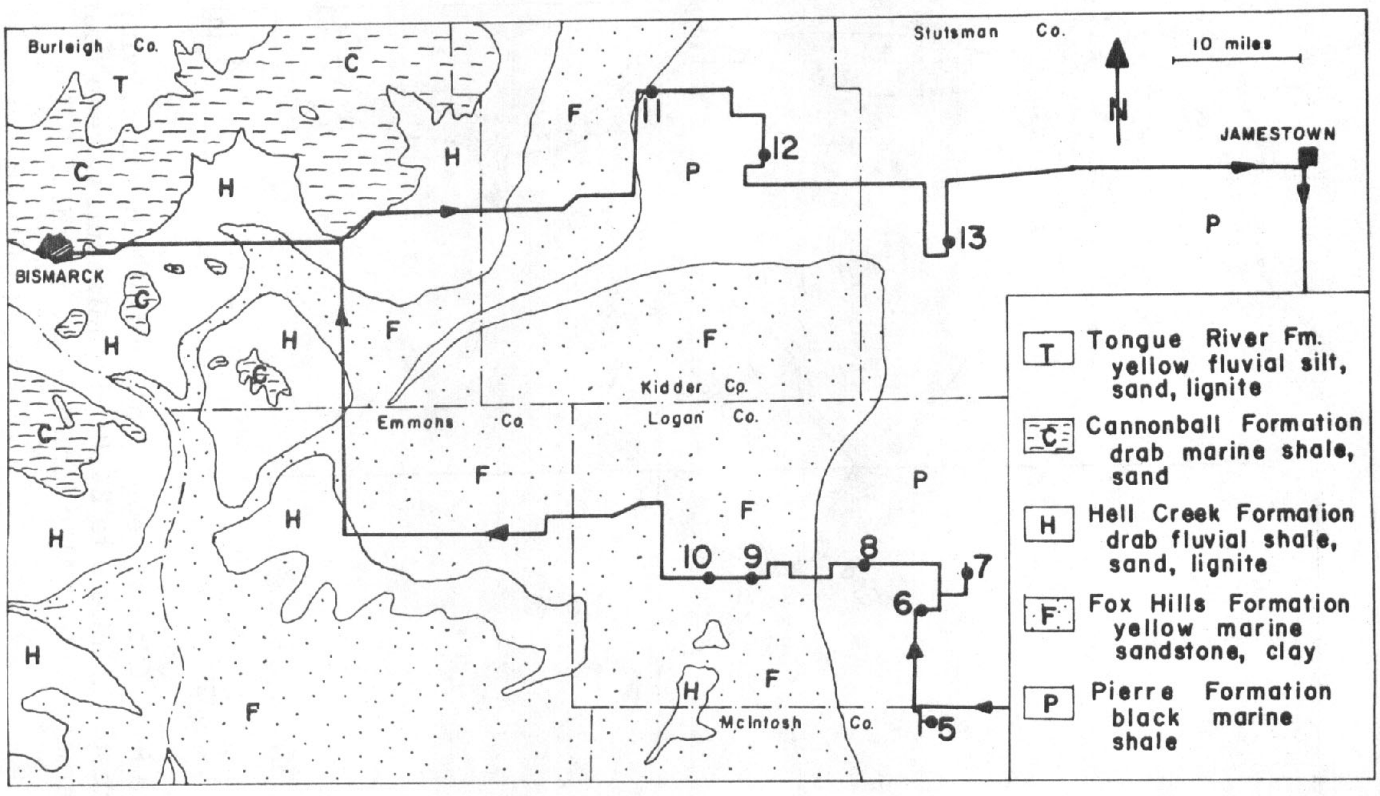



Figure R-6. Paleocene (Tongue River and Cannonball) and upper Cretaceous (Hell Creek, Fox Hills, and Pierre) formations in the fieldtrip area, with fieldtrip route and numbered stops. Same area as shown in fig. R-2 through R-5.

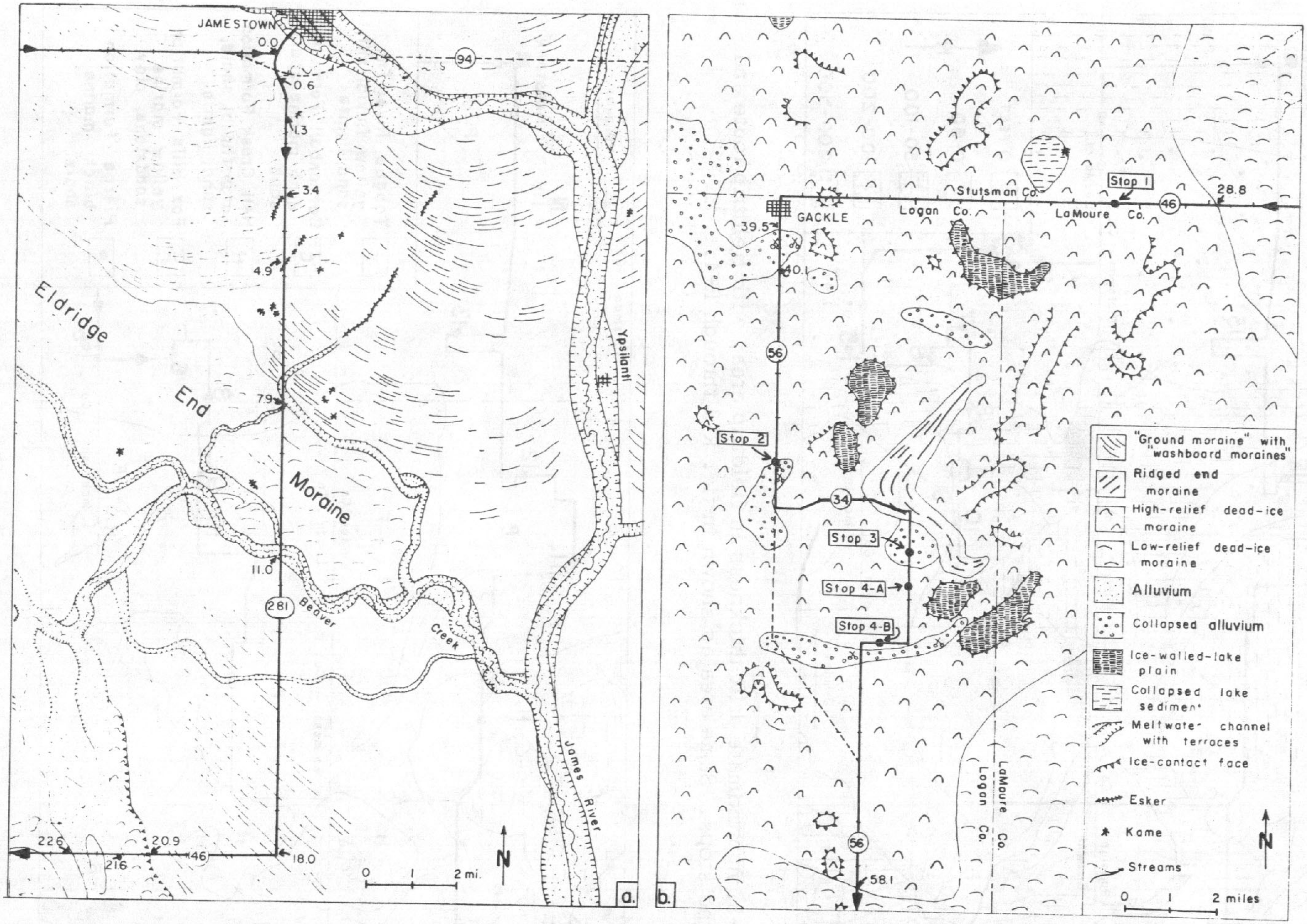



Figure R-7. Glacial geology along fieldtrip route, with mileages and stops indicated. a: Drift Prairie south of Jamestown; mile 0 to 24.0. b: Missouri Coteau near Gackle; mile 27.0 to 58.5; dashed line is wet-weather alternate route. Map location given in fig. R-2.

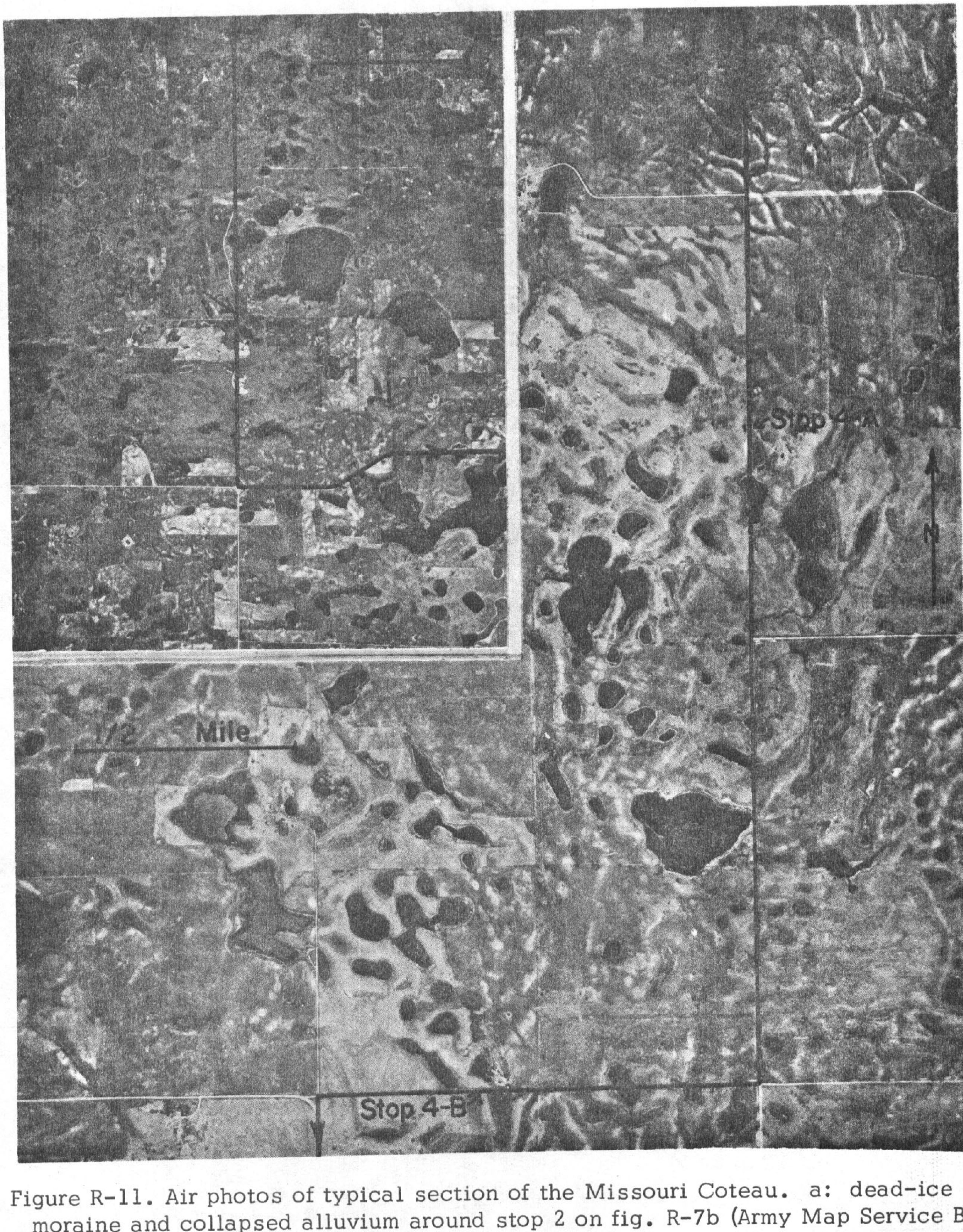



Figure R-11. Air photos of typical section of the Missouri Coteau. a: dead-ice moraine and collapsed alluvium around stop 2 on fig. R-7b (Army Map Service BE M2 253 21 July 52). b: dead-ice moraine and collapsed alluvium with disintegration trenches and ridges around stop 3, 4A, and 4B on fig. R-7b (U.S. Dept. of Agriculture BAD-1K-50 6-17-52).

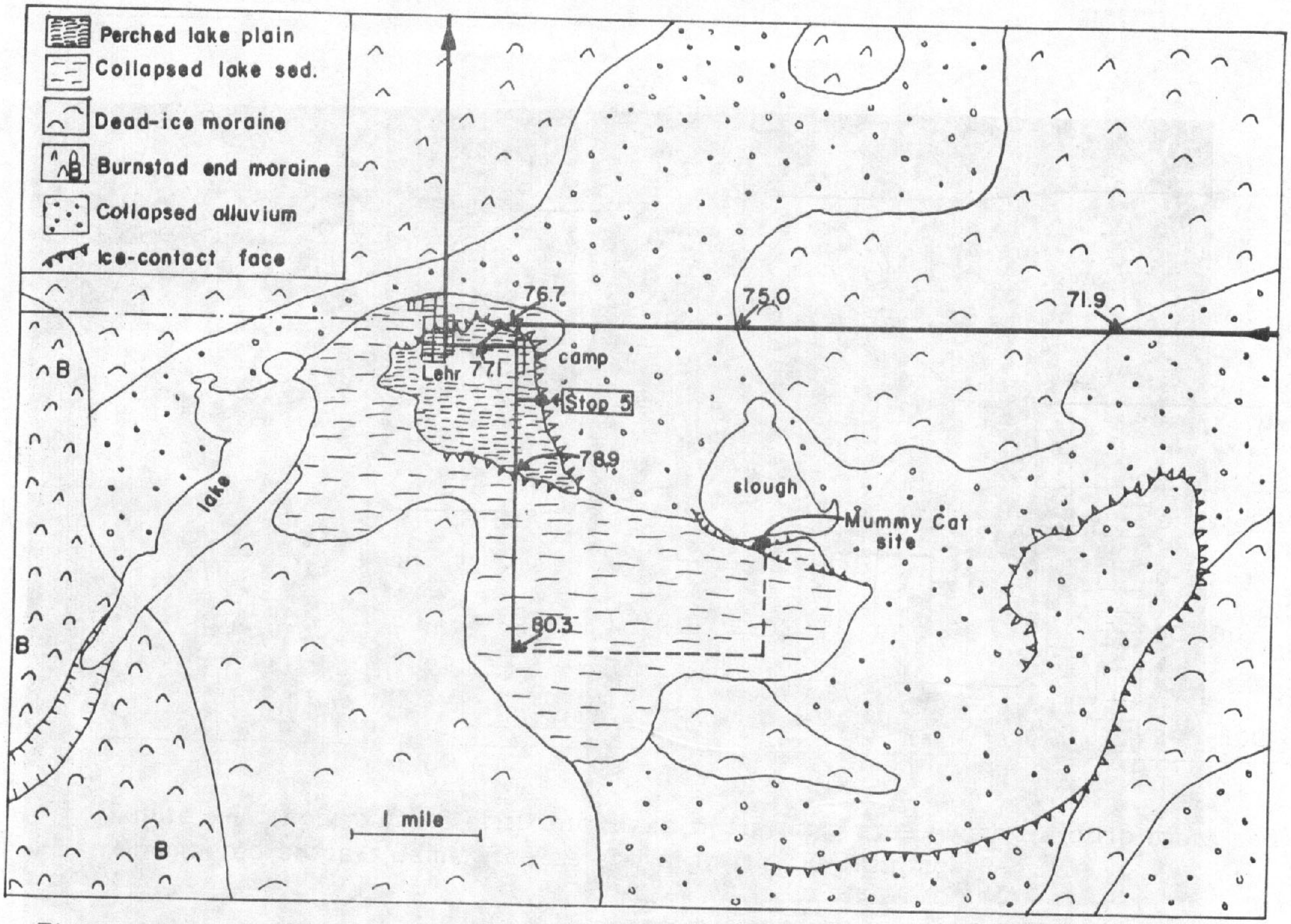



Figure R-12. Glacial geology along fieldtrip route in Lehr area, with mileages and stops indicated; mile 70.7 to 86.0. Dashed line is dry-weather alternate route. Map location given in fig. R-2.

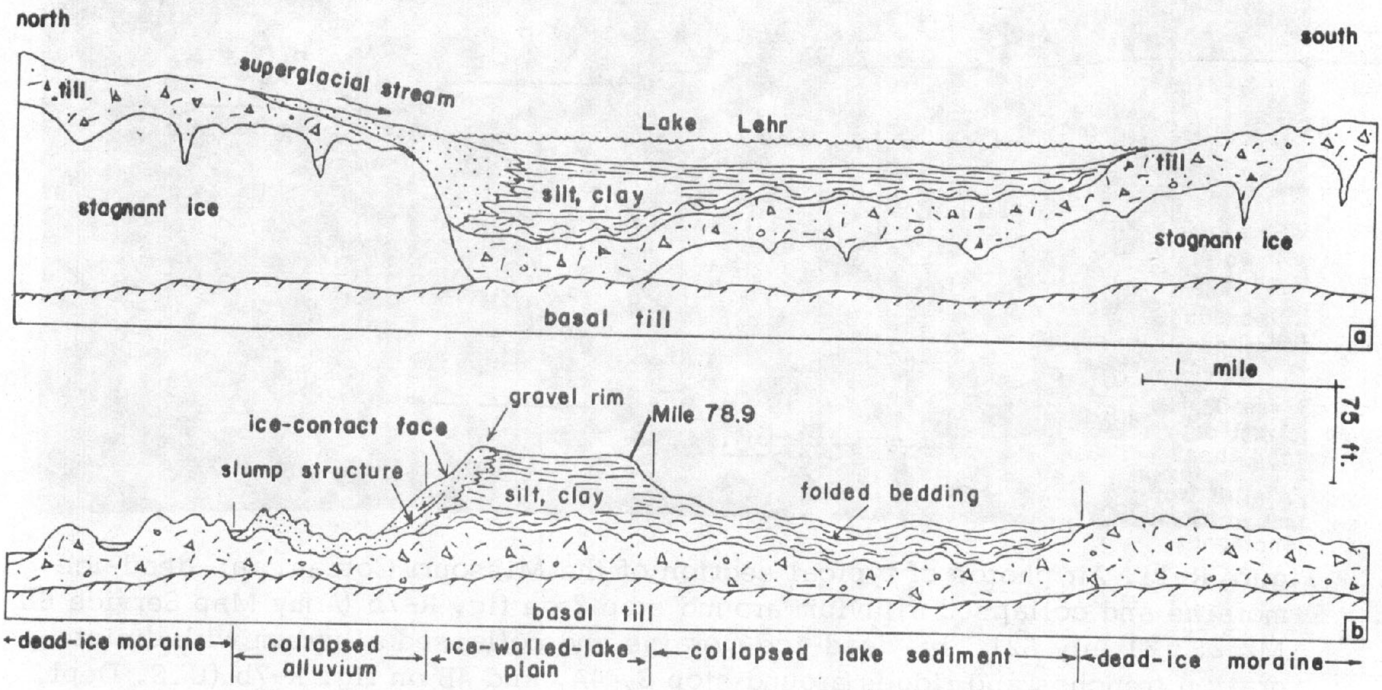
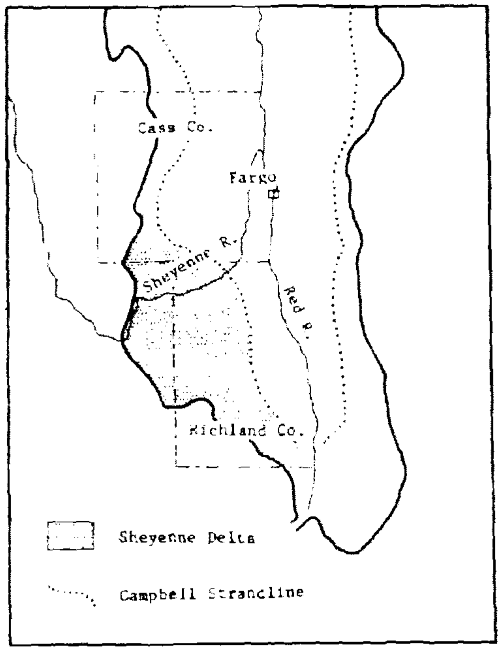



Figure R-13. North-south cross-sections showing formation of ice-walled-lake plain and associated features at Lehr. a: Latest Wisconsin time. b: Today.



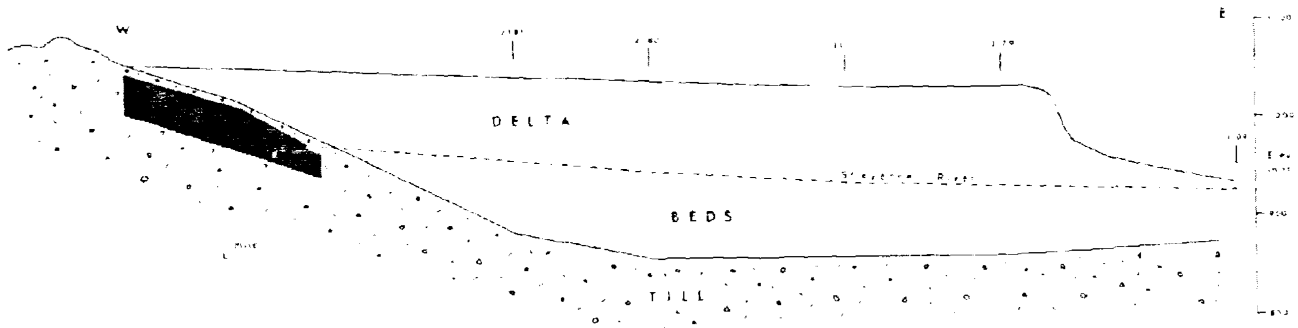




Figure P-2. Generalized east-west cross-section of Sheyenne Delta parallel to but outside of Sheyenne River trench. Dashed line is projected water-surface curve of descent for Sheyenne River.

Figure P-1. Map of south end of Lake Agassiz basin, North Dakota and Minnesota. Dotted lines mark basin margin.

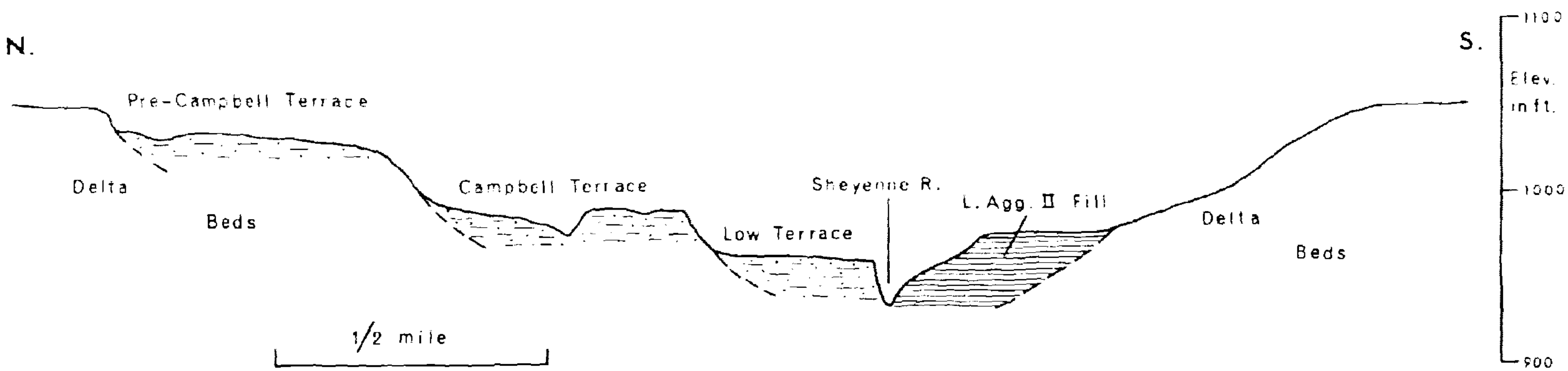



Figure P-3. North-south cross-section across Sheyenne River trench about 5 miles west of outer edge of delta.