Cenozoic History of Northeastern Montana and Northwestern North Dakota With Emphasis on the Pleistocene

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Cenozoic History of Northeastern Montana and Northwestern North Dakota With Emphasis on the Pleistocene

By ARTHUR DAVID HOWARD

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A study emphasizing the Pleistocene history of the north-central Great Plains, with descriptions of glacial and nonglacial deposits and of major drainage changes

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CENOZOIC HISTORY OF NORTHEASTERN MONTANA AND NORTHWESTERN NORTH DAKOTA WITH EMPHASIS ON THE PLEISTOCENE

By Arthur David Howard

ABSTRACT

The exposed bedrock formations of northeastern Montana and northwestern North Dakota range from Late Cretaceous to Oligocene in age. Surficial deposits include middle and upper Tertiary stream gravels and a variety of glacial and nonglacial Quaternary sediments.

The virtually flat-lying lower Tertiary formations rest unconformably on the beveled edges of the older rocks. Locally, however, these younger formations are deformed, indicating that Laramide deformation continued into early Tertiary time. The most extensive of the Tertiary deposits is the lignite-rich Fort Union formation of Paleocene age.

The area lies almost entirely within the Missouri Plateau. The plateau surface is a degradational feature, herein called the Missouri Plateau peneplain. Major valleys are trenched as much as 500 feet below the peneplain. The plateau terminates to the east at the Missouri escarpment overlooking the Central Lowland.

Mesas and buttes rise above the peneplain in both North Dakota and Montana. Those in North Dakota are relatively small; the largest, the Killdeer Mountains, occupy only a few square miles. The gravel-capped Flaxville Plateaus in Montana cover several hundred square miles and stand 300 to 600 feet above their surroundings. The most extensive group of plateaus is north of the Missouri River. A smaller group, south of the Missouri and west of the Yellowstone River, is surmounted to the west by a narrow gravel-capped ridge forming the present western divide of the Yellowstone drainage basin.

It is generally agreed that during the Paleocene the area of the northern Great Plains consisted of a low, marshy, and forested plain that sloped gently away from the Rocky Mountains. The Rocky Mountains were presumably low, permitting free access of moisture-bearing winds. The change toward aridity is attributed to progressive uplift of the mountains. The early Tertiary drainage is largely unknown.

The high narrow divide overlooking the Flaxville Plateaus west of the Yellowstone River is capped by Oligocene or Miocene gravels lithologically similar to the gravels at all lower elevations along the river but markedly different from those capping the plateaus north of the Missouri River. This gravel, the Rimroad gravel, is fluvial, and the river that deposited it was probably the Yellowstone, at a level 1,200 to 1,300 feet higher than it is now. The relief at this time, at the close of the Rimroad cycle, was probably low, although the gradients of the graded streams were steep enough to permit transportation of gravel. The steep gradients were probably due to the coarse and abundant load supplied from the newly elevated mountains and to the semiarid climate which resulted from the orogeny.

The Rimroad cycle was brought to a close by regional tilting or climatic change, or a combination of both, and the Flaxville erosion cycle was initiated. The master streams incised their courses in the Rimroad valley floors, leaving the latter temporarily perched as benches. In consequence of their gravel caps, these benches persisted long after the former divides were obliterated. The protective nature of the gravel is apparently due to its high porosity and permeability, which prevent development of surface drainage and retard erosion. The erosion cycle came to a close at a time when the gravel-floored Flaxville valley bottoms were about 500 feet below the Rimroad valley floors. The floor of the Yellowstone River valley at this time may have been 15 to 20 miles wide, and the floor of the Missouri River valley may have been wider. Thus, a complete reversal of topography took place during the Flaxville erosion cycle; the valley bottoms of Rimroad time became the divides of Flaxville time. The Flaxville gravel has been identified as Miocene or Pliocene in age on the basis of an abundant vertebrate fauna. It has been suggested, however, that the Flaxville surface may be even younger.

A second uplift, climatic change, or combination of these, introduced a new cycle of erosion. The master streams began to trench the Flaxville valley floors. Widespread removal of the Rimroad gravel exposed the divides to more rapid lowering than the valley bottoms. North of the Missouri River, the former divides were completely obliterated, and the Flaxville valley floors remain as high plateaus. South of the Missouri, the divide west of the Yellowstone River valley was not completely obliterated, so that the Flaxville valley floor along the Yellowstone is preserved as a high bench. Erosion continued for 400 to 600 feet below the Flaxville valleys. Over much of the area, particularly in the east, nearly all vestiges of the Flaxville level were removed. The vast, level to gently undulating surface which remained is the Missouri Plateau peneplain, a surface which was formed, at least in part, under semiarid conditions. In the western part of the area the peneplain is confined between the Flaxville Plateaus and appears as broad, open valleys. Where these valleys have later been trenched, the peneplain stands as a broad bench. Belts of gravel, the Cartwright gravel, mantle the bench along most of the larger valleys. Patches of gravel elsewhere over the peneplain suggest ancestral paths of the master streams.

W. C. Alden referred to the Missouri Plateau peneplain as the No. 2 bench. Alden uses the term "bench" for "nearly flat features which are clearly remnants of ancient river plains and which are not due simply to the stripping of softer shales from harder flat-lying sandstone." The term is applied "more or less indiscriminately to the tops of gravel-capped plateaus or mesas which, like the Cypress Hills, stand well above their immediate surroundings, and also, in places, to typical terraces, which are bordered on the one hand by lower lands and on the other by slopes rising to higher lands."

At the peneplain level, the Missouri, Yellowstone, and Little Missouri Rivers flowed northeastward across the map area

along nearly the same paths they followed during the preceding Flaxville cycle.

The Missouri Plateau peneplain is probably Kansan, possibly Illinoian, in age.

The ancestral northeasterly drainage of the Missouri Plateau was blocked by ice from the Keewatin center of dispersal. The present Missouri River was formed as an ice-marginal stream. It is not certain whether the diversion took place in Kansan or Illinoian time, although the earlier date seems more likely. If the Kansan date is correct, then the diversion took place earlier than it did in South Dakota, where an Illinoian date has been assigned.

During the succeeding Crane erosion cycle, the Missouri River eroded its ice-marginal course about 200 feet below the ancestral peneplain valleys and about 400 feet below the ancestral divides. The gravelly floors of the valleys of the Missouri and Yellowstone Rivers were opened to widths of 3 to 4 miles at this level. The gravel is the Crane Creek gravel of the present report.

If the Crane episode of erosion is assigned to the Yarmouth stage, then the succeeding cycle of erosion, also pre-Wisconsin, may conveniently be attributed to the Sangamon stage. If the Crane erosion is assigned to the Sangamon, however, the succeeding erosion cycle would be Wisconsin in age. Available evidence favors a Yarmouth age for the Crane cycle.

During the Crane and the succeeding deep-trench erosion cycles, the southerly direction of flow of the principal tributaries north of the Missouri was established. These streams now flow in a direction opposed to the regional slope.

Problematical till underlies early Wisconsin(?) till along Smoke Creek in Roosevelt County, Mont. If the deposit itself is not till, its glacial components clearly indicate derivation from till, a till presumably older than the overlying early Wisconsin(?) deposits. Geomorphic relations suggest, but do not conclusively demonstrate, that the till(?) was deposited after dissection of the ancestral valley of the Missouri, perhaps at the close of the Crane cycle. If the date of the diversion of the Missouri from the ancestral valley is not older than Kansan, then the Smoke Creek glaciation may be Illinoian.

Following attainment of grade at the Crane level, a new cycle of erosion was initiated, the deep-trench erosion cycle. The valleys of the master streams, and of their tributaries for varying distances upstream, were incised 140 feet or more below the Crane valley floors or more than 100 feet below present stream level. The Crane valley floors now appear as the gravel-capped Crane terrace. During the deep-trench stage, the Missouri River presumably still occupied several prominent abandoned valleys north of the present trench. One of these, in eastern Montana, is occupied by the towns of Culbertson and Bainville; a second is at Hofflund in Williams County, N. Dak.; and a third extends southeast of Sanish in Mountrail County.

The drainage pattern at the time of the advance of the early Wisconsin(?) ice was virtually as it is now, although the valleys were much deeper. The Little Missouri River, however, still followed a now-abandoned path north for 50 miles into the Missouri River, and the Missouri itself occupied the channel between Culbertson and Bainville and the channels at Hofflund and Sanish. The early Wisconsin(?) ice was the most widespread in the area and extended to the glacial limit. North of the Missouri only the highest parts of the Flaxville Plateaus projected above the ice. The ice completely buried the Missouri valley, shunting the river to the ice border 20 to 50 miles south. Here, the ice-marginal waters cut deep trenches across divides between lakes impounded in the north-draining valleys. On recession of the ice, the Little Missouri River did not return to its ancestral northern path; instead, it continued to flow in its ice-marginal path to the east. The Yellowstone, however, resumed its northerly path. During the general withdrawal of the ice north of the Missouri valley, the Missouri River was probably segmented by trailing ice lobes or by masses of residual ice. This may account for the difficulty in correlating remnants of valley fill within the trench.

The Missouri River was displaced from the Culbertson-Bainville and the Hofflund channels at this time, but the Sanish channel remained in use.

A succession of low till ridges extends 110 miles southwest of Coalridge, Mont. These ridges are interpreted as annual recessional features. The ridges average about 7 per mile, suggesting a minimum rate of retreat of the ice of about 1 mile in 7 years. This is about half the rate determined for the Mankato lobe in Iowa on the basis of somewhat similar features.

The surface of the early Wisconsin(?) till is smoothly graded. The smoothness is due in many places to the smoothness of the underlying topography, but glaciofluvial deposition, stream grading, rill wash, and creep have undoubtedly contributed to the leveling.

In the valley of Little Muddy Creek, N. Dak., middle Wisconsin(?) moraines are found below terraces of early Wisconsin(?) till. Erosion of at least 40 to 50 feet of till preceded deposition of the middle Wisconsin(?) drift. Outside the limited area of exposure of the middle Wisconsin(?) drift it has not been possible to distinguish evidence of erosion which took place at this time from evidence of that which took place subsequently.

A lobe of middle Wisconsin(?) ice extended into northwestern North Dakota and reached the Missouri River near Williston. Its drift occupies a lobate area of about 800 square miles south of the border of the Mankato drift. Pebble analyses of the till are distinct from those of the other tills and suggest a different avenue of approach of the ice. The drift has been differentiated on the basis of lithology, areal distribution, and stratigraphic and geomorphic relations.

Except for its thin deposits, the middle Wisconsin(?) ice exerted little influence on the topography. Some of the glacial channels north of Willston probably date from this glacial episode.

Terraces of middle Wisconsin(?) outwash are present in lower Little Muddy Creek valley. Williston itself is located in part on such a terrace. Middle Wisconsin(?) outwash may also contribute to some of the terraces in the Missouri valley.

Evidence of a middle(?)-late Wisconsin interval is restricted to the area of middle Wisconsin(?) drift. The middle Wisconsin(?) ice displaced Little Muddy Creek laterally in two places. The stream carved new channels in its diversion paths; these channels are now occupied by Mankato outwash. The Mankato outwash passes below the Recent alluvium in lower Little Muddy Creek valley and presumably underlies the alluvium of the Missouri valley. The Missouri River was apparently flowing at a level lower than it is now just prior to the Mankato glaciation. The middle Wisconsin(?) outwash terrace stands about 45 feet above low-water river level at Williston. Downcutting of more than this amount, therefore, is indicated for the middle(?)-late Wisconsin erosion interval.

The Mankato ice overtopped the frontal scarp of the Missouri Plateau and flooded south for 20 to 50 miles. The ice front was irregular, one lobe projecting more than 25 miles beyond the main ice mass. The glacier reached the Missouri River in the eastern part of the problem area and displaced it from the abandoned valley southeast of Sanish.

The Mankato ice distributed a thick deposit of till. On the plateau the till is piled in morainal hills and the drainage is completely disorganized in contrast to the drainage of the regions farther south. Part of the relief of the morainal topography is due to underlying irregular topography, but part is due to irregular deposition on relatively level surfaces.

Here and there topographic swales through the morainal belt suggest the presence of buried valleys. Many of the swales were used by glacial melt waters and are floored with outwash. Outwash was also fed into streams draining south into the Missouri River. Sand swept from the outwash deposits was locally gathered into dunes, and silt was distributed as loess.

North of the plateau border, the drift surface is remarkably smooth. This is probably due in part to the smoothness of the underlying topography, in part to uniform deposition of debris from the ice, and possibly in part to subsequent grading.

In late Mankato(?) time a thick alluvial fill was deposited in the valleys of the region. It is not clear whether erosion preceded this deposition, hence the deposition could have resulted from ice or debris obstructions in the Missouri valley or from a subsequent climatic change. The presence of comparable alluvial fills in other drainage systems, far from the present map area, favors the climatic interpretation. In either interpretation, the presence of remains of *Bison antiquus* (?) in the fill suggests a late Mankato age.

Erosion followed deposition of the older alluvium. The alluvium of the present flood plain occupies a valley trenched within these deposits. Except for the episodes of cut and fill, which may have been climatically induced, the topography of the region has undergone little modification in post-Mankato time.

INTRODUCTION

PURPOSE OF STUDY

The glacial stratigraphy of the Interior Plains from Iowa to the Rocky Mountains of Montana has received little attention until recent years. This is particularly true of the older deposits beyond the massive Wisconsin end moraines in North and South Dakota and Montana, first described by Chamberlin (1883). The little that was known of the older drifts was incorporated in a few short papers, including one by Alden (1924). Much information, however, was embodied in many coal bulletins issued by the U.S. Geological Survey over a span of more than 40 years. Alden's report (1932) on the physiography and glacial geology of eastern Montana and adjacent areas was the first regional interpretation of the Pleistocene stratigraphy and physiographic development of the plains east of the Rocky Mountains of Montana as far as western North Dakota. Despite the vast area covered, Alden's report is an important milestone in the interpretation of the Pleistocene history.

Within the last few years areal mapping and exploratory drilling by the U.S. Geological Survey and other government agencies have supplied much new information. The present study, although largely reconnaissance, includes the results of the detailed studies indicated in the index map of plate 1. The author served in an advisory capacity in several of these mapping projects. The present map area (fig. 1) extends about 150 miles in an east-west direction, about 130 miles north-south, and includes nearly 20,000 square miles.

FIELD METHODS

A total of about 14 months was spent in the field during the summers of 1946, 1947, 1948, 1949, and 1951. Wherever possible, north-south and east-west traverses were made at intervals of a few miles. Detailed traverses were also made along main streams and all railroad lines. Prominent exposures elsewhere were examined wherever possible.

Pebble samples were collected from tills to determine possible lithologic differences, and were collected from gravel deposits to help distinguish between glacial and nonglacial origins and help decipher former drainage changes.

The term "lithology," as applied in this report to gravel and till, refers to composition based on the numerical abundance of different rock types among the pebbles.

Plate 1 shows the distribution of the drift sheets, glacial channels, major ice-marginal lakes, principal terraces, and a wide variety of minor glacial features. The upstream termination of alluvium in most valleys was mapped from photo-index sheets and is approximate. Numerous exposures too small to be mapped but bearing on the Pleistocene history are separately considered in the text.

PEBBLE ANALYSES

Gravel deposits were sampled by scraping a vertical swath 3 to 5 feet long down the surface of the outcrop. In thick exposures, only the coarsest layers were sampled. Till was sampled by first scraping surface wash from an area estimated to contain between 100 and 200 pebbles of identifiable size and then either prying the pebbles from the exposure or collecting chips from them.

After identification, the pebbles were divided into 10 groups, and a histogram was prepared for each analysis. It should be emphasized that the analyses are based on the numbers of pebbles, not on the volumes represented. Volume determinations are impractical if not impossible.

The results of 77 pebble analyses of till are tabulated

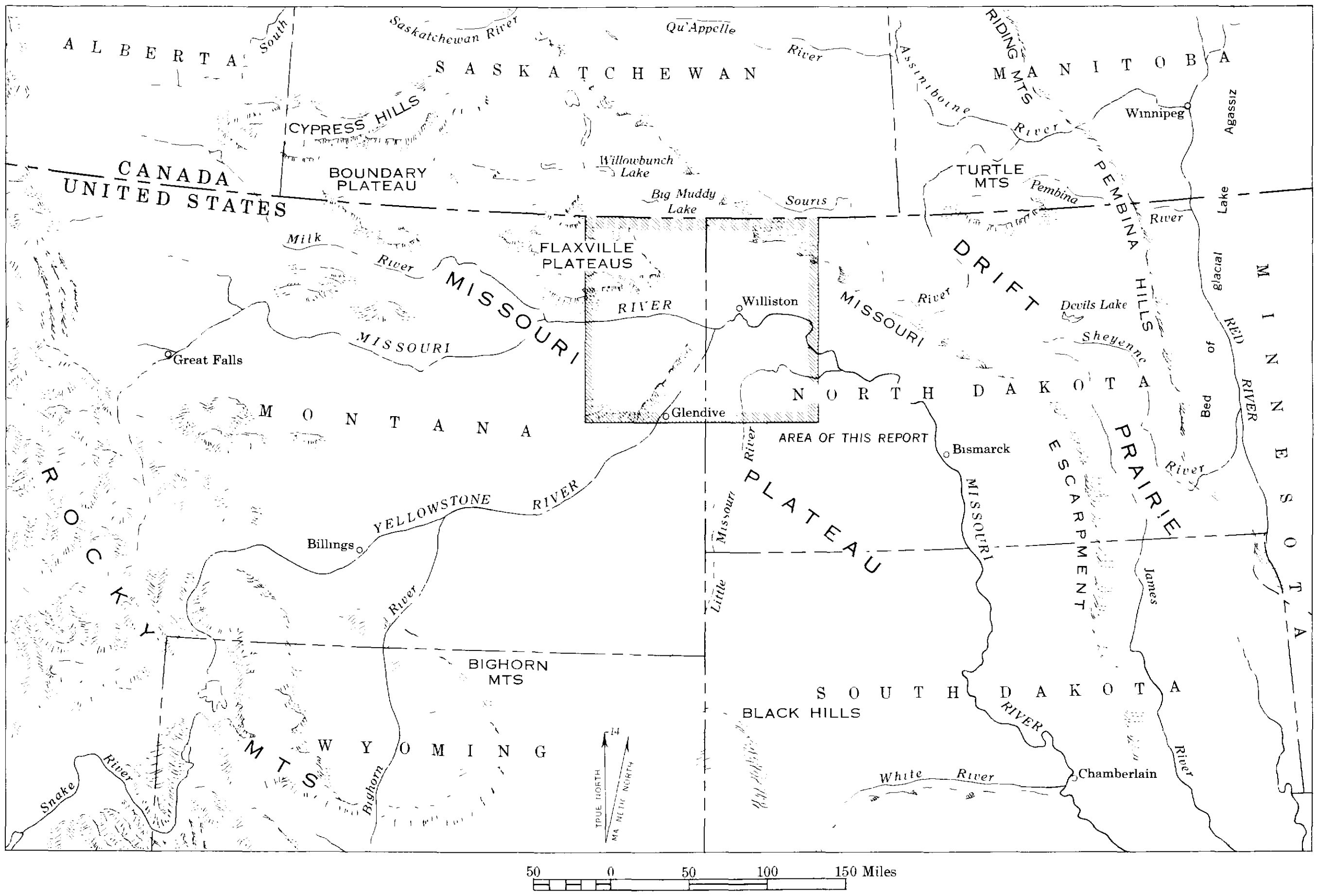

Figure 1.—Physiographic map of northern Great Plains. The area of this report is outlined in northeastern Montana and northwestern North Dakota.

in table 1. There are two analyses for each sample. The left-hand column is based on all 10 lithologic groupings; the right-hand column is based only on the obvious glacial erratics. Brown waterworn Flaxville-type pebbles, that is, pebbles from the Flaxville and similar-appearing gravels, were also excluded from the recast analyses. Although originally derived from the Rocky Mountains, they were local as far as the advancing glaciers were concerned. The Flaxville-type pebbles in till north of known outcrops of the gravels indicate that gravels are present farther north beneath the drift mantle.

The Flaxville-type pebbles eliminated in the recast analyses may have included some glacial pebbles not actually belonging to this category. Pebbles of the Flaxville type are so distinctive, however, that the probability of serious error is unlikely. Some of the Flaxville-type pebbles themselves, however, as well as pebbles of the country rock, were almost certainly brought in from the north. These cannot be separated from their local counterparts and are of no value in distinguishing tills by lithology.

The recast analyses of table 1 have been used in preparing the histogram map (pl. 5). The recast analyses include only four lithologic groups: granitic rocks, other plutonic types, metamorphic foliated rocks, and limestone and dolomite. The second group rarely constitutes more than a small percentage of the glacial pebbles, hence it has been disregarded in the preparation of the histogram map.

ISOPLETH MAPS

The significant factual data of table 1 and plate 5 are plotted diagrammatically in isopleth maps, that is, maps in which the regional distribution of a variable is shown by lines of equal magnitude called isopleths (Krumbein and Pettijohn, 1938, p. 200–202). The possible application of such maps to till studies was suggested by Howard (1950), and the maps of the present area have been described in detail by Howard (1956). The information revealed in these maps will be summarized briefly in those parts of the text concerned with the differentiation of the drift sheets.

TABLE 1.—*Pebble analyses of tills*

[The recast analyses include only glacial types known to be foreign to the area: granite and gneissic granite, other plutonic rocks, schist and other foliated rocks, and limestone and dolomite. Recasts involving relatively few pebbles are less reliable than others. The letter "X" indicates that no local bedrock types were collected. Where the "Miscellaneous" entry is high, the bedrock pebbles were abundant]

	Pebble analysis (percent) of sample [1]—													
	1		2		3		4		5		6		7	
Number of pebbles in each sample (left-hand number refers to original analysis, right-hand number refers to recast analysis)	218	96	93	74	131	101	154	135	151	133	138	130	103	85
Granite and gneissic granite	10.5	1.7	6.4	8.1	13.7	17.8	12.3	14.1	4.0	4.5	13.8	14.6	3.9	4.7
Other plutonic rocks	3.2	3.6	1.1	1.4	0	0	0	0	2.6	3.0	0	0	2.9	3.5
Acid felsitic and porphyritic types (rhyolite and trachyte)	.5		0		0		0		0		0		1.0	
Basic felsitic and porphyritic types (andesite and basalt)	4.6		1.1		9.9		5.8		2.0		1.5		2.0	
Quartz and primary quartz aggregate	.5		0		0		1.9		1.3		.7		1.0	
Schist and other foliated rocks	4.6	5.1	5.4	6.8	4.6	5.9	8.4	9.6	6.0	6.8	4.3	4.6	3.9	4.7
Quartzite	1.8		11.8		2.3		0		.7		.7		3.9	
Limestone and dolomite	71.6	79.6	66.7	83.8	58.8	76.2	66.9	76.3	75.5	85.7	76.1	80.8	71.8	87.0
Chert	1.4		7.5		4.6		0		7.3		.7		1.9	
Miscellaneous [2]	1.4		X		6.1		4.5		.7		2.2		7.7	
Total	100.1	100.0	100.0	100.1	100.0	99.9	99.8	100.0	100.1	100.0	100.0	100.0	100.0	99.9

	8		9		10		11		12		13		14	
Number of pebbles in each sample (left-hand number refers to original analysis, right-hand number refers to recast analysis)	122	95	213	171	118	89	121	72	217	171	152	119	215	118
Granite and gneissic granite	10.6	13.7	13.1	16.4	10.2	13.5	12.4	20.8	21.2	26.9	16.4	21.0	9.8	17.8
Other plutonic rocks	.8	1.1	1.9	2.3	0	0	.8	1.4	3.2	4.1	2.0	2.5	.9	1.7
Acid felsitic and porphyritic types (rhyolite and trachyte)	0		0		0		0		0		0		.5	
Basic felsitic and porphyritic types (andesite and basalt)	3.3		3.8		.8		0		5.5		5.3		3.3	
Quartz and primary quartz aggregate	1.6		1.9		.8		1.7		.9		2.6		2.3	
Schist and other foliated rocks	.8	1.1	7.0	8.8	3.4	4.5	.8	1.4	7.8	9.9	9.9	12.6	4.2	7.6
Quartzite	3.3		2.3		17.8		23.1		4.6		2.6		20.5	
Limestone and dolomite	65.6	84.2	58.2	72.5	61.9	82.0	45.4	76.4	46.5	59.1	50.0	63.8	40.0	72.8
Chert	4.1		4.7		5.1		14.1		2.3		4.6		12.1	
Miscellaneous [2]	9.8		7.0		X		1.7		7.8		6.6		6.5	
Total	99.9	100.1	99.9	100.0	100.0	100.0	100.0	100.0	99.8	100.0	100.0	99.9	100.1	99.9

	15		16		17		18		19		20		21	
Number of pebbles in each sample (left-hand number refers to original analysis, right-hand number refers to recast analysis)	170	43	204	92	164	70	197	153	96	73	162	92	103	90
Granite and gneissic granite	3.5	15.4	6.9	15.2	6.7	15.7	15.2	19.6	11.5	15.1	6.8	12.0	10.7	12.2
Other plutonic rocks	.6	2.6	0	0	.6	1.4	2.0	2.6	1.0	1.4	0	0	3.9	4.4
Acid felsitic and porphyritic types (rhyolite and trachyte)	1.8		1.5		2.4		0		3.1		3.1		0	
Basic felsitic and porphyritic types (andesite and basalt)	.6		1.5		2.4		3.6		2.1		6.2		3.9	
Quartz and primary quartz aggregate	2.9		1.0		1.2		.5		1.0		0		1.9	
Schist and other foliated rocks	4.1	17.9	2.0	4.3	2.4	5.7	7.1	9.2	5.2	6.9	0	0	2.9	3.3
Quartzite	37.6		26.5		28.0		4.1		8.3		11.7		3.9	
Limestone and dolomite	14.7	64.2	36.3	80.4	32.9	77.2	53.3	68.7	58.4	76.7	50.0	88.0	70.0	80.0
Chert	32.4		20.1		17.1		5.6		5.2		17.3		2.9	
Miscellaneous [2]	1.8		4.1		6.1		8.6		4.2		4.9		X	
Total	100.0	100.1	99.9	99.9	99.8	100.0	100.0	100.1	100.0	100.1	100.0	100.0	100.1	99.9

	22		23		24		25		26		27		28	
Number of pebbles in each sample (left-hand number refers to original analysis, right-hand number refers to recast anslysis)	123	32	179	46	200	163	198	95	159	124	143	90	193	85
Granite and gneissic granite	7.3	28.1	7.3	28.3	17.0	20.8	7.1	14.7	15.7	20.2	11.2	17.8	10.4	23.5
Other plutonic rocks	.8	3.1	4.5	17.4	2.0	2.5	2.0	4.2	0	0	0	0	2.1	4.7
Acid felsitic and porphyritic types (rhyolite and trachyte)	12.2		7.3		.5		9.6		.6		5.6		4.1	
Basic felsitic and porphyritic types (andesite and basalt)	7.3		10.6		2.0		5.0		4.4		7.0		1.6	
Quartz and primary quartz aggregate	2.4		4.5		1.0		4.0		0		2.1		2.6	
Schist and other foliated rocks	2.4	9.4	0	0	8.5	10.4	2.0	4.2	10.7	13.7	1.4	2.2	3.1	7.1
Quartzite	18.7		12.3		4.5		10.0		6.3		3.5		13.5	
Limestone and dolomite	15.5	59.3	14.0	54.3	54.0	66.3	36.9	76.8	51.6	66.2	50.0	80.0	28.5	64.7
Chert	30.9		36.9		5.0		19.2		10.0		16.1		32.1	
Miscellaneous [2]	2.4		2.8		5.5		4.0		.6		2.8		2.1	
Total	99.9	99.9	100.2	100.0	100.0	100.0	99.8	99.9	99.9	100.1	99.7	100.0	100.1	100.0

See footnotes at end of table.

TABLE 1.—*Pebble analyses of tills*—Continued

	Pebble analysis (percent) of sample [1]—													
	29		30		31		32		33		34		35	
Number of pebbles in each sample (left-hand number refers to original analysis, right-hand number refers to recast analysis)	125	44	80	44	134	121	171	157	185	165	128	118	122	106
Granite and gneissic granite	10.4	29.5	8.8	15.9	17.9	19.8	14.0	15.3	15.1	17.0	15.6	16.9	14.8	17.0
Other plutonic rocks	0	0	0	0	1.5	1.7	1.2	1.3	.5	.6	3.1	3.4	4.1	4.7
Acid felsitic and porphyritic types (rhyolite and trachyte)	4.8	--------	5.0	--------	0	--------	0	--------	0	--------	0	--------	0	--------
Basic felsitic and porphyritic types (andesite and basalt)	8.8	--------	5.0	--------	1.5	--------	2.9	--------	.5	--------	.8	--------	0	--------
Quartz and primary quartz aggregate	2.4	--------	1.3	--------	.7	--------	0	--------	.5	--------	.8	--------	1.6	--------
Schist and other foliated rocks	.8	2.3	0	0	6.7	7.4	7.0	7.6	8.1	9.1	3.9	4.2	4.1	4.7
Quartzite	7.2	--------	7.5	--------	1.5	--------	2.3	--------	1.1	--------	.8	--------	4.1	--------
Limestone and dolomite	24.0	68.2	46.2	84.1	64.2	71.2	69.6	75.7	65.4	73.3	69.6	75.4	63.9	73.6
Chert	25.6	--------	21.2	--------	4.5	--------	2.9	--------	3.2	--------	2.3	--------	1.6	--------
Miscellaneous [2]	16.0	--------	5.0	--------	1.5	--------	X	--------	5.4	--------	3.1	--------	5.7	--------
Total	100.0	100.0	100.0	100.0	100.0	100.1	99.9	99.9	99.8	100.0	100.0	99.9	99.9	100.0

	36		37		38		39		40		41		42	
Number of pebbles in each sample (left-hand number refers to original analysis, right-hand number refers to recast analysis)	108	90	145	120	175	154	185	160	152	130	133	117	160	117
Granite and gneissic granite	16.7	20.0	10.3	12.5	18.9	21.4	20.5	23.7	11.8	13.8	16.5	18.8	15.0	20.5
Other plutonic rocks	1.9	2.2	5.5	6.7	.6	.6	.5	.6	.7	.8	0	0	.6	.9
Acid felsitic and porphyritic types (rhyolite and trachyte)	0	--------	0	--------	0	--------	0	--------	0	--------	0	--------	0	--------
Basic felsitic and porphyritic types (andesite and basalt)	.9	--------	4.1	--------	2.3	--------	.5	--------	3.3	--------	1.5	--------	1.3	--------
Quartz and primary quartz aggregate	0	--------	1.4	--------	1.1	--------	1.6	--------	2.0	--------	.8	--------	3.1	--------
Schist and other foliated rocks	1.9	2.2	3.4	4.2	16.6	18.8	25.9	30.0	7.2	8.5	10.5	12.0	6.2	8.5
Quartzite	3.7	--------	2.1	--------	1.7	--------	3.2	--------	2.6	--------	.8	--------	7.5	--------
Limestone and dolomite	63.0	75.6	63.4	76.6	52.0	59.2	39.5	45.6	65.8	76.9	60.9	69.2	51.2	70.1
Chert	0	--------	1.4	--------	5.1	--------	4.9	--------	2.6	--------	5.3	--------	13.1	--------
Miscellaneous [2]	12.0	--------	8.3	--------	1.7	--------	3.2	--------	3.9	--------	3.8	--------	1.9	--------
Total	100.1	100.0	99.9	100.0	100.0	100.0	99.8	99.9	99.9	100.0	100.1	100.0	99.9	100.0

	43		44		45		46		47		48		49	
Number of pebbles in each sample (left-hand number refers to original analysis, right-hand number refers to recast analysis)	195	173	228	86	129	124	112	108	219	203	209	181	143	114
Granite and gneissic granite	27.7	31.2	11.0	29.1	17.1	17.7	19.6	20.4	33.3	35.9	17.7	20.4	14.7	18.4
Other plutonic rocks	1.5	1.7	4.0	10.5	3.1	3.2	7.1	7.4	2.7	3.0	4.8	5.5	2.1	2.7
Acid felsitic and porphyritic types (rhyolite and trachyte)	0	--------	1.8	--------	0	--------	0	--------	0	--------	0	--------	.7	--------
Basic felsitic and porphyritic types (andesite and basalt)	0	--------	14.5	--------	1.6	--------	.9	--------	.5	--------	4.8	--------	6.3	--------
Quartz and primary quartz aggregate	1.5	--------	2.2	--------	0	--------	0	--------	4.6	--------	1.4	--------	3.5	--------
Schist and other foliated rocks	21.0	23.7	2.6	7.0	19.4	20.1	9.8	10.2	30.1	32.5	6.2	7.2	7.0	8.8
Quartzite	1.0	--------	13.2	--------	.8	--------	0	--------	.9	--------	1.9	--------	3.5	--------
Limestone and dolomite	38.5	43.4	20.2	53.5	56.6	58.9	59.8	62.0	26.5	28.6	57.9	66.8	55.9	70.1
Chert	2.1	--------	25.0	--------	.8	--------	.9	--------	1.4	--------	3.3	--------	1.4	--------
Miscellaneous [2]	6.7	--------	5.7	--------	.8	--------	1.8	--------	0	--------	1.9	--------	4.9	--------
Total	100.0	100.0	100.2	100.1	100.2	99.9	99.9	100.0	100.0	100.0	99.9	99.9	100.0	100.0

	50		51		52		53		54		55		56	
Number of pebbles in each sample (left-hand number refers to original analysis, right-hand number refers to recast analysis)	119	99	158	90	203	189	211	192	191	157	159	145	142	129
Granite and gneissic granite	29.4	35.3	12.0	21.1	25.1	27.0	17.1	18.7	12.0	14.7	22.6	24.8	33.8	37.2
Other plutonic rocks	0	0	2.5	4.4	1.0	1.1	.9	1.0	3.1	3.8	5.7	6.2	1.4	1.5
Acid felsitic and porphyritic types (rhyolite and trachyte)	0	--------	5.7	--------	0	--------	0	--------	0	--------	0	--------	0	--------
Basic felsitic and porphyritic types (andesite and basalt)	4.2	--------	9.5	--------	1.5	--------	0	--------	4.7	--------	5.0	--------	4.2	--------
Quartz and primary quartz aggregate	3.4	--------	0	--------	.5	--------	1.9	--------	2.1	--------	.6	--------	1.4	--------
Schist and other foliated rocks	19.3	23.1	5.1	8.9	24.1	25.9	16.5	18.2	3.1	3.8	20.8	22.9	23.2	25.6
Quartzite	2.5	--------	11.4	--------	0	--------	1.9	--------	1.0	--------	.6	--------	.7	--------
Limestone and dolomite	34.5	41.5	37.3	65.6	42.8	46.0	56.4	61.9	63.9	77.7	42.1	46.2	32.5	35.6
Chert	.8	--------	13.3	--------	3.9	--------	5.2	--------	.5	--------	.6	--------	0	--------
Miscellaneous [2]	5.9	--------	3.2	--------	1.0	--------	0	--------	9.4	--------	1.9	--------	2.8	--------
Total	100.0	99.9	100.0	100.0	99.9	100.0	99.9	99.8	99.8	100.0	99.9	100.1	100.0	99.9

See footnotes at end of table.

TABLE 1.—*Pebble analyses of tills*—Continued

	Pebble analysis (percent) of sample [1]—													
	57		58		59		60		61		62		63	
Number of pebbles in each sample (left-hand number refers to original analysis, right-hand number refers to recast analysis)	180	165	177	157	126	103	200	191	169	158	173	156	189	177
Granite and gneissic granite	30.6	34.5	29.4	33.1	16.7	20.4	29.5	30.9	30.8	33.0	16.8	18.6	10.0	10.7
Other plutonic rocks	6.7	7.3	1.1	1.3	2.4	2.9	2.0	2.1	.6	.6	1.7	1.9	.5	.6
Acid felsitic and porphyritic types (rhyolite and trachyte)	.6	------	0	------	0	------	0	------	0	------	0	------	0	------
Basic felsitic and porphyritic types (andesite and basalt)	3.9	------	2.3	------	7.9	------	1.0	------	.6	------	0	------	2.1	------
Quartz and primary quartz aggregate	2.2	------	2.3	------	.8	------	0	------	.6	------	.6	------	.5	------
Schist and other foliated rocks	20.0	21.8	26.0	29.3	4.8	5.8	23.5	24.6	20.7	22.1	11.0	12.2	10.0	10.7
Quartzite	1.1	------	1.1	------	3.2	------	0	------	1.2	------	1.7	------	.5	------
Limestone and dolomite	33.3	36.3	32.2	36.3	57.9	70.8	40.5	42.4	41.4	44.3	60.7	67.3	73.0	78.0
Chert	0	------	0	------	2.4	------	.5	------	1.2	------	4.0	------	.5	------
Miscellaneous [2]	1.7	------	5.6	------	4.0	------	3.0	------	3.0	------	3.5	------	2.6	------
Total	100.1	99.9	100.0	100.0	100.1	99.9	100.0	100.0	100.1	100.0	100.0	100.0	99.7	100.0

	64		65		66		67		68		69		70	
Number of pebbles in each sample (left-hand number refers to original analysis, right-hand number refers to recast analysis)	147	129	143	117	121	103	142	130	189	168	158	135	164	138
Granite and gneissic granite	17.7	20.2	15.4	18.8	20.7	24.3	18.3	20.0	14.8	16.6	11.4	13.3	8.5	10.1
Other plutonic rocks	2.7	3.1	0	0	2.5	2.9	4.9	5.4	.5	.6	.6	.7	.6	.7
Acid felsitic and porphyritic types (rhyolite and trachyte)	0	------	0	------	0	------	0	------	0	------	0	------	0	------
Basic felsitic and porphyritic types (andesite and basalt)	4.1	------	3.5	------	7.4	------	4.2	------	.5	------	1.3	------	5.5	------
Quartz and primary quartz aggregate	2.0	------	2.1	------	2.5	------	1.4	------	3.2	------	2.5	------	.6	------
Schist and other foliated rocks	10.9	12.4	11.2	13.7	19.0	22.3	4.2	4.6	10.0	11.3	4.4	5.2	6.7	8.0
Quartzite	1.4	------	1.4	------	.8	------	0	------	0	------	3.2	------	1.2	------
Limestone and dolomite	56.5	64.4	55.2	67.6	42.9	50.5	64.1	70.0	63.4	71.4	69.0	80.7	68.3	81.2
Chert	2.0	------	4.9	------	.8	------	.7	------	4.8	------	3.2	------	1.8	------
Miscellaneous [2]	2.7	------	6.3	------	3.3	------	2.1	------	2.6	------	4.4	------	6.6	------
Total	100.0	100.1	100.0	100.1	99.9	100.0	99.9	100.0	99.8	99.9	100.0	99.9	99.8	100.0

	71		72		73		74		75		76		77	
Number of pebbles in each sample (left-hand number refers to original analysis, right-hand number refers to recast analysis)	130	103	182	138	142	113	167	141	171	104	149	112	155	107
Granite and gneissic granite	10.0	12.6	10.4	13.8	12.7	15.9	14.4	17.0	7.0	11.5	18.1	24.1	10.3	14.9
Other plutonic rocks	0	0	3.3	4.3	1.4	1.8	2.4	2.8	2.3	3.8	2.7	3.6	1.3	1.8
Acid felsitic and porphyritic types (rhyolite and trachyte)	0	------	0	------	0	------	0	------	.6	------	0	------	0	------
Basic felsitic and porphyritic types (andesite and basalt)	3.1	------	5.5	------	1.4	------	3.0	------	1.8	------	5.3	------	8.4	------
Quartz and primary quartz aggregate	5.4	------	.5	------	2.1	------	1.8	------	1.2	------	2.0	------	2.6	------
Schist and other foliated rocks	7.7	9.7	3.3	4.3	4.9	6.2	11.4	13.5	2.9	4.8	4.7	6.3	1.9	2.8
Quartzite	2.3	------	1.1	------	2.8	------	4.8	------	1.8	------	1.3	------	4.5	------
Limestone and dolomite	61.5	77.7	58.8	77.6	60.6	76.2	56.3	66.7	48.5	79.8	49.7	66.0	55.5	80.4
Chert	2.3	------	.5	------	3.5	------	1.2	------	1.8	------	4.7	------	6.5	------
Miscellaneous [2]	7.7	------	16.4	------	10.5	------	4.8	------	32.2	------	11.4	------	9.0	------
Total	100.0	100.0	99.8	100.0	99.9	100.1	100.1	100.0	100.1	99.9	99.9	100.0	100.0	99.9

[1] Sample numbers refer to localities in plate 5.
[2] Largely local bedrock types: sandstone, siltstone, shale, clay, coal, clinker, petrified wood, and limonite. Clinker is a dense, reddish, locally scoriaceous rock resulting from the baking of sediments over a burning coal bed.

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PHYSIOGRAPHIC SETTING

REGIONAL SETTING

Three principal plains levels and two remnant upland levels dominate the landscape between the Red River, at the North Dakota-Minnesota boundary, and the Rocky Mountains in western Montana (fig. 1). The plains levels rise westward like the treads of a giant staircase, the treads becoming progressively wider in that direction. The two eastern steps are within the Central Lowland province. The lower is Red River valley, a level plain forming part of the floor of glacial Lake Agassiz; the higher is the drift prairie, a region of gently rolling topography interrupted by low morainal ridges. The vast western step, separated from the drift prairie by the Missouri escarpment, is the Missouri Plateau, in which the present area of study lies. The plateau surface is erosional and slopes eastward from the Rocky Mountains at an average rate of 10 feet per mile. Small isolated buttes and mesas dot the plateau surface. West of the Montana line are large remnants of two higher erosion surfaces. The lower of the two surfaces, the Flaxville Plain, is preserved in the Flaxville Plateaus of northern Montana; the higher, the Cypress Plain, is best preserved in the Cypress Hills of southeastern Alberta and southwestern Saskatchewan.

LOCAL SETTING

TOPOGRAPHY

The topography of northeastern Montana and northwestern North Dakota is shown on the physiographic map, plate 2. The entire area, except the low country in the northeast, is in the Missouri Plateau; the low country is part of the drift prairie. Total relief is about 1,600 feet. The high point, of more than 3,300 feet elevation, is in the extreme southwest on the high divide between the Yellowstone River and Redwater Creek; the low point, of about 1,720 feet elevation, is in the extreme east along the Missouri River.

DRIFT PRAIRIE

The drift prairie slopes gently northeastward from the Missouri escarpment to the valley of the Riviere des Lacs. Near the escarpment, the surface is moderately well drained by small channels trending northeast and by channels normal to these, probably formed by glacial waters marginal to the ice front as the latter receded to the northeast. Northeast of the Riviere des Lacs, much of the topography is morainal, although extremely subdued, and there is little integrated drainage.

The drift prairie abuts against the Missouri escarpment at an elevation of about 2,000 feet. Although the escarpment is several hundred feet high, its slope rarely exceeds a few degrees, and near Kenmare, N. Dak. (pl. 2) it is barely perceptible.

MISSOURI PLATEAU

West of the Missouri escarpment is the gently undulating surface of the Missouri Plateau, which is referred to in this report as the Missouri Plateau peneplain. The peneplain is most extensive in the North Dakota portion of the map area. Over at least one-third of the Montana portion, it is confined between the Flaxville Plateaus and stands as a terrace above the present valley bottoms.

The Missouri Plateau is dissected by the Missouri, Yellowstone, and Little Missouri Rivers and by their tributaries. The major streams have cut trenches from 1 to 4 miles wide and 300 to 600 feet deep. Episodes of cut and fill within the trenches have formed complex terraces. Rejuvenation of the tributary streams has formed badlands along the major rivers; those of the Little Missouri valley are the most spectacular (pl. 2 and fig. 2). The badland belts along the Missouri and Yellowstone Rivers are generally only 1 to 2 miles wide, and locally the plateau surface extends uninterruptedly to the very rim of the valleys.

FIGURE 2.—Badlands of the Little Missouri River, south of Watford City, McKenzie County, N. Dak.

Between the Little Missouri and Yellowstone Rivers in North Dakota is a maze of buttes, mesas, and narrow ridges, many capped by clinker 20 to 80 feet thick. Broad, deep, flat-floored troughs (pl. 2), former escape routes for proglacial lake waters, pass through the divides.

A conspicuous topographic feature of McKenzie County is the linear depression from half a mile to 2 miles wide that extends northeast about 50 miles from the Little Missouri valley, near the mouth of Hay Creek, to the Missouri valley. The depression is clearly the former pathway of the Little Missouri River into the Missouri River.

The eastern edge of the Missouri Plateau is thickly mantled with glacial drift. The pronounced morainal topography is partly due to the irregularity of the underlying topography.

The name "Coteau du Missouri" has been applied both to that portion of the Missouri Plateau between the Missouri escarpment and the Missouri River (Chamberlin, 1883, p. 394; Simpson, 1929, p. 11) and to the glacial hills which mantle this strip of plateau (Willard, 1902, p. 70; Willard and Erickson, 1904, p. 17, 20; and Fenneman, 1931, p. 74-75). Willard in 1902 and again with Erickson in 1904 presents the origin of the phrase. The following excerpt is from the 1904 publication:

> A great region lying east of the Missouri River was called by the early French explorers "Les Coteaux du Plateau," or the Hills of the Missouri Plateau. In the popular mind the hills or "coteaus" [the plural form is generally so anglicized] and the plateau are confused, the term "coteaus" being often applied to the great hilly upland which is the plateau, and a surface feature of which is the "coteaus" or hills. . . . The coteaus of the Missouri are morainic hills.

In this report, original usage will be followed; the morainal hills mantling the Missouri Plateau for as much as 50 miles back from the Missouri escarpment will be referred to by the term "coteaus" or the "coteaus of the Missouri Plateau". The belt of country characterized by the coteaus will be referred to by the terms "coteau belt" or "coteau morainal belt."

The coteau belt extends from the Missouri escarpment southwestward for 20 to 50 miles. Total relief from the floors of deep linear swales to the divides between them amounts to several hundred feet. Relief between individual depression floors and the tops of the surrounding hills may be 50 feet. Many depressions are occupied by saline lakes or salt flats.

The coteau belt is trenched by numerous southwest-sloping outwash channels. Some of the larger are shown in the north-central portion of plate 2. Elsewhere, as between Zahl and Crosby, N. Dak. (pl. 2), broad linear sags pass entirely through the morainal belt to the drift prairie beyond. These may mark the paths of ancestral northeast-draining valleys.

The coteau belt terminates abruptly in the northeast at the Missouri escarpment, but the southwest margin is highly irregular and not clearly defined. In places the morainal topography becomes more subdued toward the southwest and gives way to a level till plain. Elsewhere, outlying morainal areas are separated from the principal morainal belt by gently undulating or level till and outwash plains.

Beyond the confines of the coteau belt, morainal topography is rare and generally subdued. The only prominent moraine south of the Missouri River is in northeastern McKenzie County (pl. 2).

FLAXVILLE PLATEAUS AND FLAXVILLE PLAIN

The Flaxville Plateaus of northeastern Montana stand 300 to 600 feet above the Missouri Plateau peneplain (fig. 1 and fig. 3). The plateaus are

FIGURE 3.—Flaxville upland (background) north of the Missouri River, 16 to 18 miles west of Antelope and 8 miles south of Archer, Sheridan County, Mont. The Missouri Plateau peneplain is in the foreground.

capped by Miocene or Pliocene gravel, named the Flaxville gravel by Collier and Thom (1918) after the town of Flaxville atop one of the plateaus in Sheridan County, Mont. The gravel mantles an erosion surface beveling Mesozoic and Tertiary sedimentary rocks. The plateaus are presumably remnants of a former erosion surface, the Flaxville Plain. The term "plain," however, as in "Flaxville Plain" and "Cypress Plain," may be a misnomer. The level summits of the gravel-capped plateaus do not prove that the former landscapes were everywhere level. Evidence to be presented suggests that the Flaxville gravel, at least, was confined to broad valleys which, together with the intervening divides, contributed to a relief of hundreds of feet.

One large group of plateaus lies north of the Missouri River and west of Big Muddy Creek (pl. 1). A second group, south of the Missouri River, lies between the Yellowstone valley and the valley of Redwater Creek. Other scattered summits, some flat topped and bearing gravel, rise to the Flaxville level. Possible Flaxville gravel has been described (Howard, Gott, and Lindvall, 1946) in northwestern North Dakota, far east of the previous known limits. Few of the possible remnants could be examined in this reconnaissance.

PLATEAUS NORTH OF THE MISSOURI RIVER

The Flaxville Plateaus north of the Missouri River comprise the eastern end of a belt of plateaus about 175 miles long, east and west, and about 50 miles wide. The restored surface descends eastward from an elevation of 3,300 feet in Blaine County, far west of the map area, to 2,600 feet at its eastern terminus in Roosevelt County. The average slope, in the 180 to 190 miles, is between 3½ and 4 feet per mile.

In western Roosevelt County, and for some distance west of the map area, a large Flaxville upland forms a bench about 45 miles long and 12 to 15 miles wide along the Missouri valley. The surface of the bench descends eastward parallel to the direction of flow of the Missouri and also southward toward the present position of the river. In general appearance, it closely resembles the more restricted inner terraces of the valley and undoubtedly represents part of the floor of the Missouri valley when the river flowed some 700 to 1,000 feet above its present level. In this report, all high-level terraces along the Missouri and Yellowstone valleys within the present map area will be attributed to the high-level ancestors of these streams. It is recognized, however, that drainage changes upstream from the map area may have fashioned the modern courses of these two rivers from a significantly different ancestral pattern.

PLATEAUS SOUTH OF THE MISSOURI RIVER

The Flaxville uplands south of the Missouri River have the same relation to the Yellowstone River as those immediately north of the Missouri have to that river. They parallel the Yellowstone valley on the west for at least 60 miles, descending northeastward like the river and sloping toward it. The surface is surmounted to the west by the high divide between the Yellowstone and Redwater drainages (pl. 2 and fig. 4). This divide, which reaches an elevation of 3,300 feet in the map area, displays still older gravels. To the east, across the Yellowstone valley, are high isolated buttes equivalent in height to the Yellowstone-Redwater divide. These may be remnants of the ancient divide between the Yellowstone valley and the Little Missouri valley to the east. If so, the Yellowstone valley, in Flaxville time, was 30 to 40 miles wide, although the valley floor may have been only half as wide. The distribution of the Flaxville gravels in a relatively narrow belt along the Yellowstone valley, their surface slope to the north, and their lithologic similarity to the gravels of the lower terraces along the Yellowstone River, clearly indicate that the gravels mark a former high-level valley of the Yellowstone 700 or 800 feet above the present valley bottom.

The relations of the Flaxville Plateaus to other levels bordering the Yellowstone valley are shown by projected profiles in figure 4. It will be noted that much of the rugged country east of Yellowstone valley rises to the Flaxville level, although gravel has not been reported from this little-known area.

West of the Yellowstone-Redwater divide, a few linear, flat-topped ridges reach the Flaxville level. The capping gravel is mapped as Flaxville in plate 1. The remnants slope toward Redwater Creek. The marked asymmetry of the Yellowstone-Redwater divide may be due to more rapid erosion to the west because of the absence of "protective" gravels as extensive or thick as those within the valley of the Yellowstone itself.

CYPRESS HILLS AND CYPRESS PLAIN

The only possible remnants of the Cypress Plain in the present map area are on the Yellowstone-Redwater divide. High parts of the divide are gravel capped and stand above the Flaxville surface. The gravel, herein referred to as the Rimroad gravel after the village of that name in Dawson County, Mont., are apparently younger than the White River beds of Oligocene age and presumably older than the Flaxville gravel of Miocene or Pliocene age. They are herein interpreted as remnants of a high-level Oligocene or Miocene valley bottom.

The vertical intervals separating the Cypress Plain, the Flaxville Plain, the Missouri Plateau peneplain, and the present river flood plains diminish eastward and the surfaces converge in that direction. Correlation of these surfaces with surfaces in the Rocky Mountains has not yet been established to complete satisfaction.

GLACIAL MANTLE

During the Pleistocene the northern Great Plains were invaded by ice from the Keewatin center of dispersal. North of the glacial limit (pl. 1) the uplands are blanketed by drift that ranges in thickness from a few inches to 20 feet or more, whereas valleys are buried locally to depths of hundreds of feet. High parts of the Flaxville Plateaus projected above the ice and are drift free.

SAND DUNES

A final element of the landscape worthy of mention is the sand dunes. Colton (1954) has mapped a dune area of about 15 square miles immediately east of

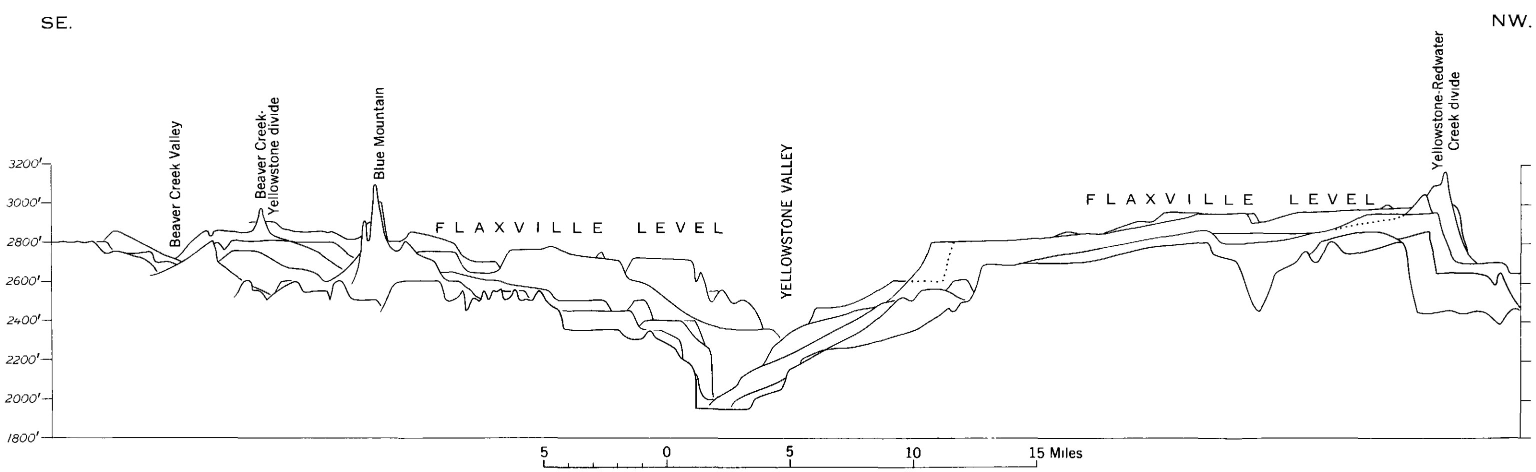


FIGURE 4.—Projected profiles of the Yellowstone River valley and borderlands. Width of block is 60 miles northwest-southeast; length of block along the Yellowstone valley is 40 miles (foreshortened). View is southwest from Savage, Mont. The surfaces, each of which consists of several closely spaced levels, descended toward the observer, hence the profiles are generally higher toward the background. The upland surface is the Flaxville; the prominent bench halfway to river level is the Cartwright bench, the correlative of the Missouri Plateau peneplain. Vertical exaggeration is approximately 40 times.

Poplar River valley near its junction with the Missouri River. I. J. Witkind (written communication, 1947) and G. B. Gott and others (written communication, 1949) mapped most of the dune areas east of Big Muddy valley. The largest of these, covering about 21 square miles, is southeast of Medicine Lake. A smaller area includes 2 principal segments and covers 14 to 15 square miles southeast of Big Muddy valley. Big Muddy valley has a thick fill of outwash below the clay-silt alluvium; the outwash presumably supplied the sand before burial under the alluvium. Poplar River valley may also contain buried outwash. Several ages of dunes are probably represented; in some areas the dunes are fairly well stabilized by vegetation, and elsewhere they are bare and apparently active.

Another dune area of a few square miles is northeast of Crosby, Divide County, N. Dak., near the Canadian border (R. C. Townsend, written communication, 1947).

DRAINAGE

The divide between drainage to the Gulf of Mexico and drainage to Hudson Bay lies in the coteau belt, generally quite close to the Missouri escarpment. The streams of the drift prairie (fig. 1) flow to Hudson Bay by way of the Souris, Assiniboine, and Nelson Rivers. The rest of the drainage is tributary to the Missouri River.

The major streams have cut deep trenches across the area. Where the valley sides are being actively undercut, precipitous bluffs rise abruptly to the upland level. Locally, landslides have produced belts of rugged, broken topography. Elsewhere, terrace remnants, as much as several square miles in area, separate flood plain and upland. Because of their importance in the present study, the terraces are considered separately.

The Missouri River flows eastward across the map area. A significant characteristic of its valley is the abrupt alternation of broad and narrow reaches (fig. 5), which is discussed later. No such marked variations occur along either the Yellowstone or Little Missouri valleys.

The Yellowstone River enters the Missouri River in about the center of the map area. Unlike the Missouri, it meanders only locally; it is coarsely braided.

The Little Missouri River flows north for 30 to 35 miles into the map area before turning abruptly eastward to join the Missouri. The abandoned valley marking the former path of the Little Missouri northeastward into the Missouri has already been mentioned.

The only other long tributary from the south is Redwater Creek, in the western part of the area. Redwater Creek meanders through a broad, flat-floored, terraced valley to join the Missouri opposite Poplar, Mont.

No large rivers enter the Missouri from the north. The four principal tributaries are, from west to east, the Poplar River, Big Muddy Creek, Little Muddy Creek, and the White Earth River. The White Earth valley has precipitous bluffs comparable in height with those along the larger valleys of the region. Steep bluffs are also present in Big Muddy valley, from the town of Reserve, Mont., to the Canadian border. The rest of Big Muddy valley, and practically all of Pop-

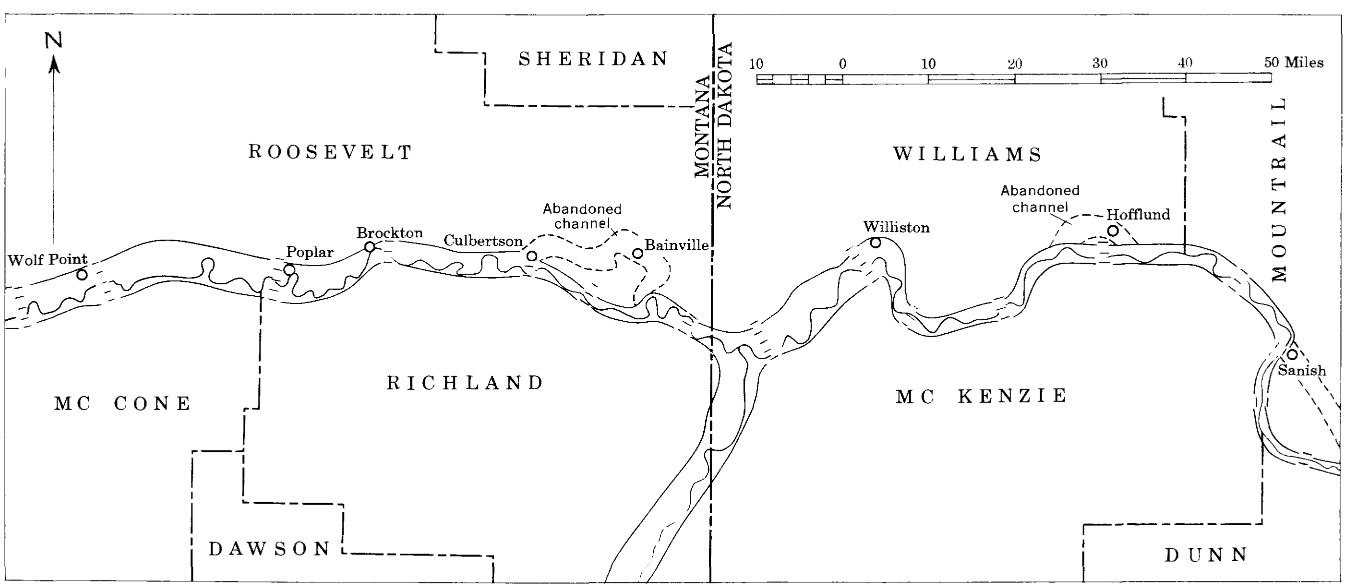


FIGURE 5.—Variations in width of the Missouri River valley below level of Missouri Plateau peneplain. The valley width has been exaggerated about one-third; tick marks show width across valley and are at intervals of 1 mile.

lar and Little Muddy valleys, are broad and gentle sided, although the flood plains in many places abut terrace scarps as much as 50 feet or more in height.

Certain peculiarities in the courses of these streams are the direct results of glacial occupation of the region and will be discussed later along with the numerous channels and abandoned valleys of the region (fig. 6).

FIGURE 6.—Ice-marginal channel, about 3 miles north of Skaar, southwestern McKenzie County, N. Dak.

CLIMATE

TEMPERATURE

The climate of eastern Montana and western North Dakota is continental in character: long severe winters, short hot summers, low rainfall, and a dry atmosphere. Temperatures of 90° to 95°F are frequent during the summer. A high of 110°F has been recorded at Poplar, Mont., and a high of 124°F has been recorded at Medora, N. Dak., 10 miles south of the map area. A low of −63°F has been recorded at Poplar and one of −56°F in McKenzie County, N. Dak. The ranges of temperature thus reach 180°F. Extreme cold generally occurs in quiet weather. Periods of high temperature in summer are generally brief and not particularly oppressive, because of the low humidity.

The growing season, between killing frosts, averages about 115 days. Although the season is short, the duration of sunlight during the summer is long, as much as 16 hours per day; this and the generally clear conditions prevailing during the growing season permit cereals to mature before the first frost sets in.

Streams are ice locked for about 4 months of the year, and the ground is sufficiently frozen to prevent absorption of moisture for about 6 months. Fortunately, 75 percent of the precipitation occurs in summer when absorption is greatest.

PRECIPITATION

The region is semiarid; precipitation ranges from about 16 inches in the east to 12 inches in the west. Precipitation is associated with cyclonic storms moving eastward with the prevailing westerly winds. The summer rainfall is generally in showers and thunderstorms and is often spotty. Areas of crop failure due to insufficient rainfall are frequently surrounded by areas of prolific crops.

Snow may fall during all months except June, July, and August. The average depth of snow between December and March is between 6 and 12 inches and, except in drifts, rarely reaches 2 feet.

Sporadic hailstorms during the summer may destroy crops. In 1946, pear-shaped hailstones as much as 6 inches long fell in the Williston area, smashing windows, denting cars, and damaging roofs.

GROUND WATER

The water table is highest in early summer when rainfall is heaviest. Rainfall decreases in the late summer and early autumn, and evaporation and transpiration rapidly lower the water level. Many wells then go dry. The lesser evaporation of autumn may permit a secondary rise of the water table.

Snow is a relatively unimportant supplier of ground water; as long as the ground is frozen, precipitation cannot enter and the wells may again run dry. The ground is generally frozen even while the snow is melting. Wells tapping deep aquifers are generally unaffected by seasonal variations in precipitation.

WINDS

Winter winds are strong because of the steep pressure gradients between cold polar air masses from the north and warmer air masses from the south and because of the level terrain. The blizzards are really windstorms blowing small amounts of snow about. The winds are usually westerlies or northwesterlies. During the spring they do much damage to early seeded crops in the lighter soils. During the summer, hot winds from the southwest may cause heavy crop losses if the rainfall is below average.

VEGETATION

Eastern Montana and western North Dakota are virtually treeless. However, timber grows on the moist flood plains of the few large streams and may be found on protected slopes where the amount of evaporation is low. Cottonwood and other poplars grow profusely in the major valleys, and cedar and juniper grow in many ravines. Other principal varieties are boxelder, ash, elm, and oak. Many varieties of shrubs and vines grow in the timbered areas.

Except where cultivated, the region is covered with arid-land plants, such as grasses, sagebrush, and cactus. The prairie grasses form a thick matted sod, relatively impervious to water, and contribute to rapid runoff in heavy rains. The sod does, however, restrict evaporation of ground water.

Removal of the tough, impermeable sod during cultivation retards runoff, although soil erosion ensues if precautions are neglected. Rapid evaporation, in the absence of the protective sod, plus the demands of annual crops constitute a heavy drain on the ground water.

South-facing slopes are generally bare because of exposure to the sun and the warm southerly winds. Erosion is more active on such slopes than on the opposing grass-covered slopes. Thus south-facing slopes are generally more dissected than those that are north facing. To this extent the distribution of vegetation influences the topography.

For further details of the flora of the region, the reader is referred to Bergman (1912).

SOILS

A wide variety of soil series and types has been described by D E. Willard (1906), Rice (1910), Lapham (1910), Gieseker and others (1933), and Edwards and Ableiter (1942). In general, the rolling areas are occupied by sandy and stony soils, and the flat or gently undulating areas by silty soils. The many boulders and the roughly rolling topography of morainal areas make cultivation difficult. These areas, as well as areas of deep dissection, are largely used for grazing.

The thickness of the soil horizons differs considerably from place to place. An almost universal characteristic of the soil profile, however, is the abundance of lime in the B horizon. It occurs in streaks and spots and forms a light-gray band 1 to 2 feet or more thick. This is the so-called caliche horizon.

FLOOD PLAINS AND TERRACES

FLOOD PLAINS

The Missouri River flood plain stands terracelike 18 to 20 feet above the low-water river level (fig. 7). It is beyond reach of ordinary floods, but it is inundated by infrequent major floods, especially those caused by ice jams. Ice-jam levels of 24 and 28 feet above low water have been recorded along the Missouri River at Wolf Point and Culbertson, Mont., respectively. Flood levels of 27 and 26 feet have been recorded along the Yellowstone River at Glendive and Fairview, Mont., respectively. In March 1950, the Little Missouri River overtopped its flood-channel banks for a distance of 25 miles above its mouth. Intervals between major floods, however, are long. The Missouri River flood plain between Buford and Trenton in North Dakota is reported (Wilder, 1904, p. 18–19) to have been flooded only twice in one period of 28 years. It was flooded again on April 1, 1952. Comparable intervals between major floods prevail in the Yellowstone River valley.

Control now afforded by the Fort Peck Dam in the Missouri River valley west of the problem area has reduced the river fluctuations. Inundations of the 18- to 20-foot flat are not now likely. On the other hand, the controlled discharge of water from the reservoir has probably raised the midsummer low-water level. It was originally estimated (North Dakota State Planning Board, 1935, p. 66) that the August low-water level would be raised about 6 feet at Williston and about 5 feet at Bismarck.

The low-water channel of the Missouri River is generally only a few hundred feet across and a few

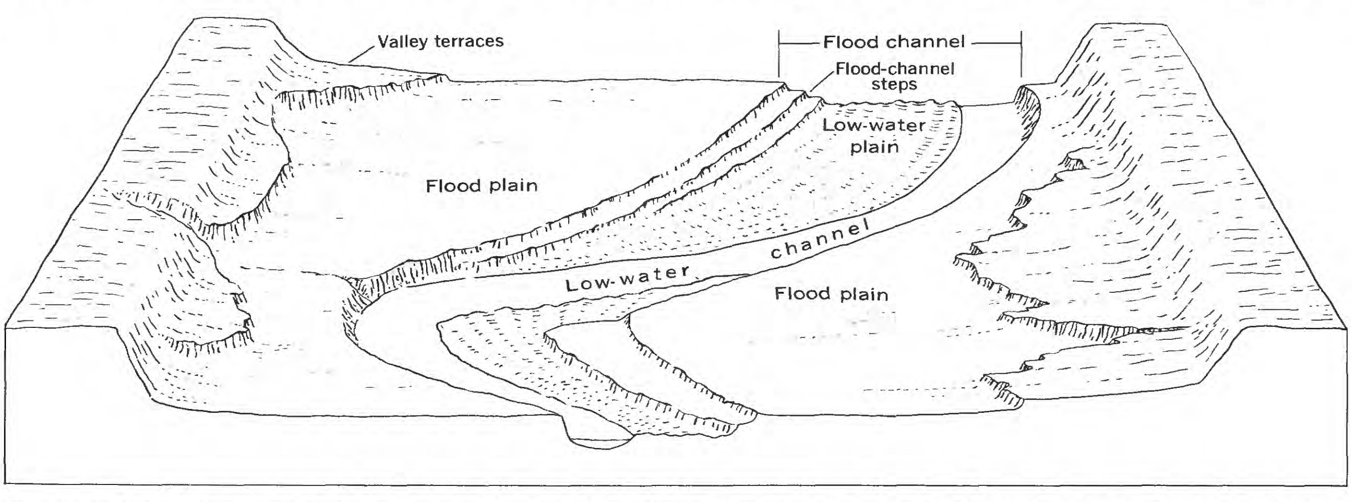


FIGURE 7.—Principal geomorphic elements of the Missouri River valley. The flood plain is as much as 4 miles wide; the flood channel ranges in width from 1,300 feet or less to 13,000 feet or more. Flood-channel steps reach widths of 7,000 feet.

feet deep. The low-water stream shifts continuously across an irregular surface corrugated with sandbars (fig. 7). This low-water plain, which is inundated at the first rise in water level, varies appreciably in width but probably averages three or four thousand feet. For convenience, it will be referred to as the low-water plain, in contrast with the flood plain above.

The flood plains of the larger rivers, although appearing remarkably level on casual inspection, have a varied relief (pl. 3). Natural levees along the flood channel of the Missouri River stand 5 feet or more above the general level, and some meanders abandoned within the last 60 years contain oxbow lakes whose bottoms lie below low water. Elsewhere over the flood plain the relief is due primarily to bars, swales, and sloughs. No other flat above the flood plain shows this pattern of levees, bars, swales, and cutoffs; the low-water plain, however, shows these to a lesser extent. It is relatively simple, therefore, to distinguish on aerial photographs higher valley terraces from the flood plain itself and the low-water plain. The vegetation is likewise distinctive: the native vegetation of the flood plain is deciduous forest, dominantly cottonwood trees as much as 2 feet thick and grassy or brush-covered open areas. The flood-channel steps below are covered by brush and willows; the higher valley terraces are sage or grass covered and practically treeless.

Levees of tributary streams, sand dunes, and wash from terraces and bluffs add to flood-plain relief. Because the flood plain is young, the wash slopes are smaller than they are on the terraces.

The varied topography of the flood plain gives it a relief of 30 to 40 feet, reckoned from low water. Too frequently, students of terraces are unaware of the magnitude of flood-plain relief and too hastily correlate isolated terrace remnants on the basis of elevation alone. If the Missouri River were now to entrench itself and the abandoned flood plain were to be thoroughly dissected, the scattered remnants would be distributed over a vertical interval of 30 to 40 feet. Obviously, correlation of remnants on the basis of elevation alone might lead to serious error.

FLOOD-CHANNEL STEPS

In many places terraces interrupt the slopes between the flood plain and the low-water plain (fig. 7). These terraces are annually(?) inundated and generally swampy. Whether they are entirely erosional or the products of cut and fill was not determined. To avoid confusion with terraces above the flood plain those within the flood channel will be referred to as flood-channel steps.

VALLEY TERRACES

One or more higher terraces—here called valley terraces (fig. 7)—are found within the larger valleys. One of these, the Crane terrace, a gravel-capped rock bench, is described in the section on the Crane Creek gravel. Remnants of fill terraces commonly bury the Crane terrace, particularly north of the glacial limit. The fill terraces consist of till, glaciofluvial sediments, alluvium older than that of the flood plain, slope wash, and combinations of these, all of which are described in later sections.

PRE-PLEISTOCENE STRATIGRAPHY AND STRUCTURE

REGIONAL GEOLOGIC SETTING

The regional geologic setting is simple. The Tertiary sedimentary rocks of the Great Plains rest on the truncated edges of deformed Paleozoic and Mesozoic rocks and dip gently eastward at a few feet per mile. Tertiary deformation is indicated by minor folds in the Fort Union formation of Paleocene age.

LOCAL GEOLOGIC SETTING

AREAL GEOLOGY AND STRATIGRAPHY

The pre-Pleistocene areal geology of the problem area is shown in plate 4. Quaternary alluvium in major valleys is also shown in this plate, and a more precise distribution of the Flaxville gravel, based on recent studies, is shown in the glacial map, plate 1.

The stratigraphy of the exposed formations, including the Quaternary deposits, is summarized in table 2. Upper Cretaceous and Tertiary rocks are the only pre-Pleistocene rocks exposed. North of the glacial limit, even these are effectively mantled by glacial drift and loess, and locally by eolian sand. Exposures are found only in valleys or in isolated buttes and mesas. Outside the glaciated area, exposures are more abundant, although loess is still widespread and level areas are blanketed by a deep soil.

UPPER CRETACEOUS

The oldest formation, the Bearpaw shale of Late Cretaceous age, is widely exposed in and adjacent to the Poplar anticline (pl. 4 and fig. 8). The formation thickens eastward and forms the upper part of the Pierre shale. The latter is exposed in the core of the Cedar Creek anticline (fig. 8).

The Fox Hills sandstone overlies the Bearpaw and Pierre shales and is disconformably overlain by the shales, sandstone, and conglomerates of the Hell Creek formation (Jensen, 1952, p. 40). The somber-appearing Hell Creek formation contains abundant dinosaur remains in sharp contrast with the overlying brighter colored, lignite-rich Fort Union formation of Paleocene age.

TABLE 2.—*Stratigraphy of exposed formations of eastern Montana and western North Dakota*

[The Montana section of the table is based on the studies of Jensen (1951, 1952), Collier and Knechtel (1939), Alden (1932), Hares (1928), Clapp, Bevan, and Lambert (1921), Calvert (1912), and Howard (present investigation). The North Dakota section is based on the studies of Benson (1949), Brown (1948), Benson and Laird (1947), Howard (1946), and Laird (1944)]

				Eastern Montana	Western North Dakota
Cenozoic	Quaternary	Recent		Alluvium, eolian sands and silts, slope wash, and products of mass movements	
Cenozoic	Quaternary	Pleistocene		Till, stratified drift, eolian sands and silts, slope wash, and products of mass movements	
Cenozoic	Tertiary	Pliocene or Miocene		Flaxville gravel: Thickness, 30 ft average. Fluvial sands and gravels. Pebbles waterworn and stained with iron oxide. Almost exclusively quartzite and chert north of Missouri River, but includes silicified igneous rocks in Yellowstone drainage basin. One probable occurrence reported in North Dakota.	
Cenozoic	Tertiary	Miocene or Oligocene		Rimroad gravel: Thickness, 30 ft average. Fluvial sands and gravels capping high divide between Yellowstone River and Redwater Creek. Similar to Flaxville gravel.	Not reported in North Dakota.
Cenozoic	Tertiary	Oligocene		White River formation: Thickness, 250 ft. Clays, shales, sands, limestone; fluvial and lacustrine.	
Cenozoic	Tertiary	Eocene		Not reported in easternmost Montana.	Golden Valley formation: Thickness, 100 ft. Micaceous sands and silts and clay lenses, underlain by gray carbonaceous shales and white and yellow-orange clays.
Cenozoic	Tertiary	Paleocene	Fort Union formation	Sentinel Butte member: Thickness ranges from 210 ft in Montana to 550 ft in North Dakota. Somber sandstones, shales, clays, and lignite coal; some bentonite. Lower part interfingers with upper part of Tongue River member.	
Cenozoic	Tertiary	Paleocene	Fort Union formation	Tongue River member: Thickness ranges from 700 ft in Montana to 300 ft in North Dakota. Light-gray, calcareous sand, silt, clay, and numerous lignite beds; weathers yellow to buff; loglike concretions as much as 30 ft thick.	
Cenozoic	Tertiary	Paleocene	Fort Union formation	Lebo shale member: Thickness, 400 ft. Dark shale and thin beds of white sandstone and sandy clay.	Ludlow (250 ft) and Cannonball (300 ft) members: The continental shales, sandstones, and lignite of the Ludlow grade eastward into the marine sands and clays of the Cannonball.
Cenozoic	Tertiary	Paleocene	Fort Union formation	Tullock member: Thickness 165 ft. Light and dark shales and sands, locally yellowish; coal seams in dark shales.	
Mesozoic	Cretaceous	Upper Cretaceous		Hell Creek formation: Thickness ranges from 575 ft to 100 ft, west to east. Largely shale in Montana but some sandstone in lower part and a persistent coal seam at top. Gray to brown bentonitic sandstone and shale in North Dakota.	
Mesozoic	Cretaceous	Upper Cretaceous	Montana group	Fox Hills sandstone: Colgate sandstone—thickness, 80 ft. Brown sandstone with concretions. Basal member—thickness 80–200 ft. Yellow clay, silt, sand.	Fox Hills sandstone: Thickness, 180–320 ft. Brown to gray sandstone.
Mesozoic	Cretaceous	Upper Cretaceous	Montana group	Bearpaw shale: Thickness, 1,200 ft. Dark-gray shale; layers of bentonite.	Pierre shale: Thickness, 2,390–930 ft, west to east. Dark-gray bentonitic shale with ironstone concretions and selenite. Present also in the Cedar Creek anticline in eastern Montana.

TERTIARY

FORT UNION FORMATION

The Fort Union formation of Paleocene age underlies the greater part of the area. It consists largely of interbedded mudstones, shales, siltstones, sandstones, and lignite coal, and it contains numerous fossil plants and fresh-water mollusks. Extremely dense siliceous siltstone caps a small upland in southern McKenzie County, N. Dak., about 12 miles northwest of Grassy Butte. The rock forms a layer as much as several feet thick. In thin section, it is seen to consist almost exclusively of angular grains of quartz of silt size firmly cemented by cryptocrystalline silica. Scattered grains of tourmaline, zircon, and other minerals attest to its detrital origin. The reason for the localization of the silicification is unknown. No float of this rock appears on the surrounding plateau surface, nor has rock of this type been found elsewhere.

GOLDEN VALLEY FORMATION

Locally unconformable on the Sentinel Butte member of the Fort Union formation are the clays, shales, silts, sands, and coal seams of the Golden Valley formation of Eocene age (Benson and Laird, 1947, and Benson, 1949). The formation outcrops in Killdeer Mountains and Blue Buttes, prominent landmarks in western North Dakota, and caps many other buttes and mesas. Benson considers the Golden Valley formation to be the most reliable horizon for structural mapping in southwestern North Dakota.

WHITE RIVER FORMATION

The White River formation of Oligocene age consists of over 200 feet of fluvial and lacustrine clays, shales, sands, and limestones. They locally rest unconformably on the Paleocene Fort Union, as in Sentinel Butte, N. Dak., and elsewhere they rest directly on the Golden Valley, as in the Killdeer Mountains.

RIMROAD GRAVEL

Remnants of a coarse sandy fluvial gravel—the Rimroad gravel—are distributed along the high divide between the Yellowstone River and parallel-flowing Redwater Creek to the west. Pebble size in the gravel layers ranges considerably, vertically and horizontally. The pebbles average an inch or less in size in the coarsest layers, but there are some as large as 6 inches. Quartzite, chert, and igneous rock are abundant, and scattered pebbles of quartz, agate, and silicified wood are present. Among the igneous rocks, porphyritic andesites are probably the most common, but all types from granite to gabbro are included. The plutonic types are rare, however. Porphyritic trachytes are conspicuous because of the large light-colored phenocrysts, many of which are more than a quarter of an inch in size. Local rock types, including clinker and black shale, are also present. No fossils have yet been recorded from these gravels.

Collier and Knechtel (1939, p. 13) reported a thickness of 20 to 100 feet of this gravel at Antelope Mountain, a few miles west of Rimroad in Dawson County, Mont. The writer believes that the latter figure may be excessive. Although none of the gravel remnants in or outside this area are known to overlie Oligocene beds, Alden (1932, p. 12) believes that Oligocene beds formed part of the landscape in which the gravels were deposited. He believes that the gravel is no older than Oligocene. Inasmuch as the gravels stand above the level of the Flaxville gravel of Miocene or Pliocene age, the Rimroad gravel may be Oligocene or Miocene in age, and possibly correlative with the capping gravels of the Cypress Hills in southern Saskatchewan, described by Mackin (1937, p. 871).

FLAXVILLE GRAVEL

The Flaxville gravel is reported to be from a few feet to 100 feet thick (Collier and Thom, 1918, p. 181). The pebbles average about 1 inch in size but include cobbles as much as 12 inches in size. They are waterworn and coated brown by a patina of iron oxide. The gravels within the Missouri River drainage consist almost entirely of quartzite and chert but include scattered pebbles of quartz. The gravels along the Yellowstone River include numerous pebbles of aphanitic and porphyritic igneous rocks, largely andesitic. The ancestral course of the Yellowstone has been traced 10 miles north of the Missouri valley by means of the distinctive gravels.

Numerous vertebrate fossils were collected from the Flaxville gravel north of the Missouri River by Collier and Thom and were identified by J. W. Gidley of the U.S. National Museum as being Miocene to early Pliocene in age. No fossils have yet been found in the Flaxville gravel along the Yellowstone River. It will be recalled that the Rimroad gravel on the Yellowstone-Redwater divide was also barren.

West of the Yellowstone-Redwater divide, gravels cap ridges sloping toward Redwater Creek. These gravels, too, are below the level of the Rimroad gravel and are mapped as Flaxville.

The remainder of the deposits of the region are Pleistocene and are the subject of the greater part of this report.

STRUCTURE

The structure of the map area influences the topography and bears on the problem of the origin of the master drainage. Figure 8 shows the major structures in and adjacent to the study area: the Poplar anticline, the narrow Cedar Creek (Glendive-Baker) anticline, and the Nesson-Keene anticline.

The Fort Union formation is involved in the major folding and is locally faulted. The largest known fault is the Weldon fault (fig. 8), which has displacement of 100 to 160 feet (Collier and Knechtel, 1939, p. 17). Townsend (1950) has reported a zone of complicated minor folding and faulting near Lignite, Burke County, N. Dak.

The folding and the faulting of the Fort Union indicates that deformation continued until or near the close of the Paleocene epoch. Ballard (1942, p. 1559) has reported gentle folding in the White River formation in the southern part of the Williston basin and concludes that deformation continued into the Oligocene.

RELATION OF TOPOGRAPHY AND DRAINAGE TO STRUCTURE

Areas of steep dip, such as the west side of the Cedar Creek anticline, are characterized by a ridge-and-valley topography. The greater part of the region, however, is underlain by virtually horizontal strata and the topography consists of broad, level upland tracts dissected by a dendritic system of valleys.

The major streams of the region are independent of structure. The Missouri River crosses the Poplar

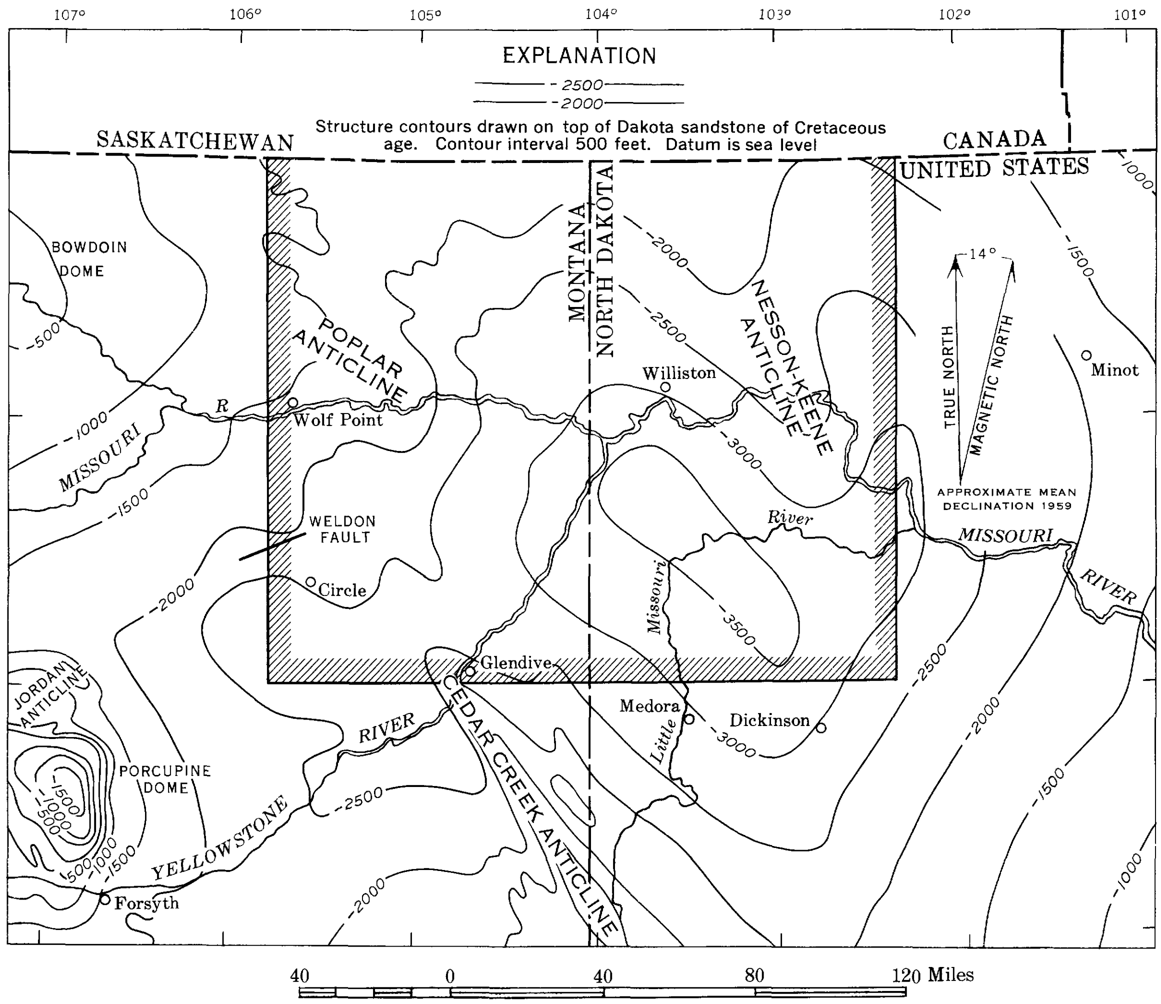


FIGURE 8.—Structure-contour map of eastern Montana and western North Dakota. The present map area is outlined. The structure-contour lines are from the map of the Missouri Valley Region, Part III, Fuel Resources, U.S. Geological Survey, Missouri Basin Studies No. 1, 1945. Only major folds are labeled. The Weldon fault has been described by Collier and Knechtel (1939).

and Nesson-Keene anticlines, and the Yellowstone River crosses the north end of the Cedar Creek anticline (fig. 8). The streams are either antecedent or were superposed from some higher surface, perhaps from the level of the Rimroad gravel of Oligocene or Miocene age.

In areas of mild deformation, streams appear to follow synclines. Thus, O'Fallon Creek, a tributary to the Yellowstone River about 15 miles south of the map area, follows a syncline parallel to the Cedar Creek anticline (Calvert, 1912, p. 201) and the White Earth River, in the eastern part of the map area, occupies a syncline adjacent to the Nesson-Keene anticline.

Some streams may be of compound origin, with post-Paleocene superposition onto structures in the Fort Union formation, followed by antecedent relations relative to later movements.

QUATERNARY STRATIGRAPHY

The surficial deposits of the area are largely Quaternary. Alden (1932) mapped three drifts in the northern Great Plains: one of late Wisconsin age, another of early Wisconsin age, and a third of Iowan or Illinoian age. At the time of Alden's report, many geologists regarded the Iowan as a separate age of the Pleistocene, older than the Wisconsin, younger than the Illinoian. Alden's map shows the drifts in offlap relations, the oldest being the most extensive. The results of the present study agree essentially with Alden's conclusion regarding the existence and distribution of the oldest drift. There is agreement, too, on the presence of the late Wisconsin drift, but considerable disagreement on its distribution. The greatest disagreement, however, concerns the intermediate drift. The middle Wisconsin(?) drift of this report has no resemblance to Alden's early Wisconsin drift either in physical character or distribution. Finally, the presence of still a fourth drift is suggested by a few scattered deposits in Roosevelt County, Mont.

In addition to glacial drifts, the Quaternary deposits include inwash,[1] outwash, lacustrine deposits, eolian sand and silt, and at least one interglacial gravel. North of the glacial limit, the surficial mantle consists largely of till and glaciofluvial deposits. Both north of the glacial limit and for short distances to the south, valley bottoms are occupied by outwash. Inasmuch as most valleys drain generally northward, outwash and inwash are probably intimately mixed. In addition, most valleys of the region are choked with silts of at least two different ages, and major valleys show at least one older interglacial (?) gravel. Both upland and lowland alike are blanketed by a veneer of loess generally less than 2 feet thick, but locally as much as 5 feet thick. Several large sand-dune areas are east of Big Muddy Creek valley in Montana and eolian sand mantles the surrounding countryside.

The sequence of Quaternary deposits is shown in table 3.

TABLE 3.—*Quaternary stratigraphy in northeastern Montana and northwestern North Dakota*

[Wavy lines indicate erosion]

Recent		Loess [1]	1-5 ft
		Flood-plain alluvium [2]	20+ ft
Pleistocene	Wisconsin	Older alluvium and slope wash / Loess and eolian sand / Mankato drift	Loess: 1-2 ft Sand: 25+ ft Older alluvium and slope wash: as much as 100 ft exposed. Mankato drift: as much as 300 ft (estimated)
		Middle Wisconsin(?) drift	25+ ft, max
		(Fossil soil on drift) Early Wisconsin(?) drift	As much as 80 ft exposed.
	Sangamon(?)	Deep-trench erosion cycle	
	Pre-Wisconsin: Illinoian(?)	Drift(?) at Smoke Creek	12-15 ft
	Pre-Wisconsin: Yarmouth(?)	Crane Creek gravel	10-15 ft, avg
		Crane erosion cycle	
	Pre-Wisconsin: Aftonian(?)	Cartwright gravel	30 ft, avg

[1] Possible pre-Mankato loesses are described in the text but are not included in the table.
[2] The flood-plain alluvium is mantled by, and probably interfingers with, slope wash along its margins.

The sequence must be considered tentative in view of the reconnaissance nature of the study. Stratigraphic relations of the deposits are shown schematically in figure 9. The Quaternary deposits will now be described in detail and the evidence for their age relationships considered.

[1] Inwash is sediment, largely of nonglacial origin, deposited around the margin of a glacier by streams draining toward the ice. Close to the ice, inwash is intimately mingled with outwash.

CARTWRIGHT GRAVEL

Here and there over the Missouri Plateau peneplain are belts and patches of gravel marking either higher levels of present streams or the paths of former streams. The gravel is similar to the Flaxville gravel. The largest area of exposure is along the Yellowstone River valley, where the gravel caps a high bench. The gravel is less extensively exposed north of the glacial limit because of the mantle of drift. However, patches are present along the Missouri River, Little Missouri River, and other streams, as well as in interfluve areas. The distribution shown in plate 1 should not be considered final, because of the reconnaissance nature of the study.

The gravel is well exposed along a road that descends into Yellowstone valley 5 to 6 miles north of the town of Cartwright in northwestern McKenzie County, N. Dak., not far from the Yellowstone-Missouri confluence. The gravel will be referred to as the Cartwright gravel after the exposures at this locality. The plateau is here mantled by stratified glacial sediments and till. Beneath the glacial sediments (early Wisconsin?) in the southwest corner of sec. 31, T. 152 N., R. 103 W., is a thick sequence of nonglacial sands and gravels. The pebbles of the gravels average about an inch in size, but include cobbles as large as 8 inches. In some localities, gravel layers are cemented by calcium carbonate and form ledges. The total thickness of the deposit at the Cartwright locality is between 35 and 40 feet, but the regional average is probably about 30 feet.

Like the lithology of the older gravels, the lithology of the Cartwright gravel varies depending on whether the gravel was imported by the Yellowstone or Missouri Rivers. The deposits of the former include numerous reddish and yellowish quartzites, a variety of acidic and basic porphyritic and nonporphyritic volcanic rocks, abundant cherts, small numbers of plutonic igneous rocks including granite, and scattered pebbles of agate, petrified wood, clinker, and local sedimentary types. Alden (1932, p. 57) reported a few pebbles of fossiliferous limestone of Paleozoic age. The lithology of a random sample is shown in figure 10. The Missouri River facies of the gravel, upstream from the mouth of the Yellowstone, consists almost exclusively of quartzite, chert, and quartz. Contamination of glacial till by these lithologically unlike facies led the writer into the early error of assuming two different glacial drifts where he now recognizes only one. Except for lithology, the two facies are almost identical. Most pebbles bear a brown patina and all but local types are well rounded.

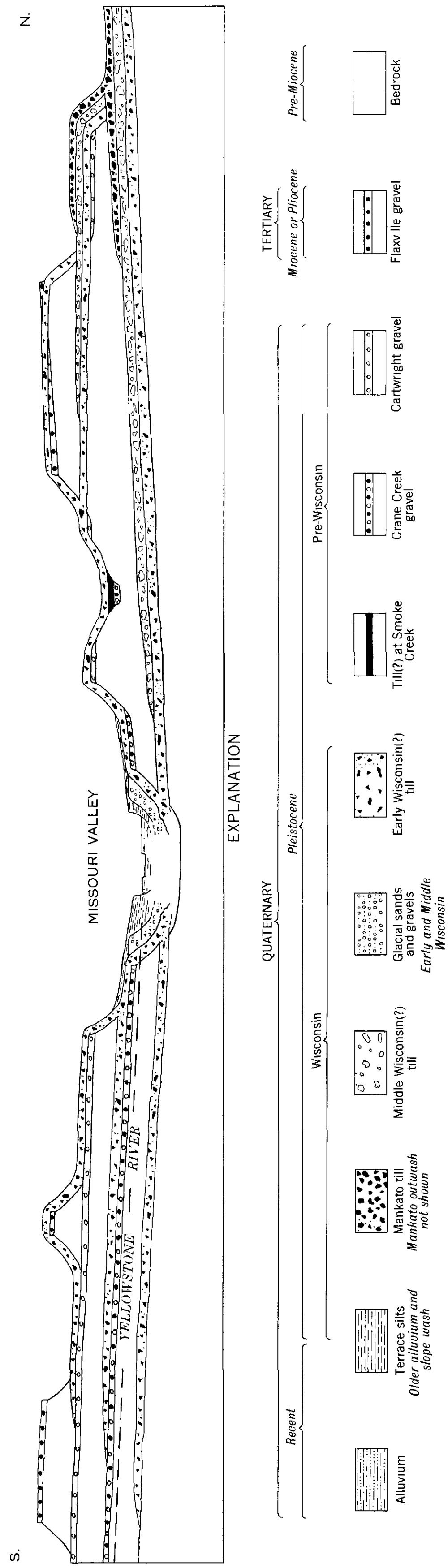


FIGURE 9.—Composite section of Quaternary deposits of northeastern Montana and northwestern North Dakota. The Flaxville gravel of Miocene or Pliocene age is added for completeness.

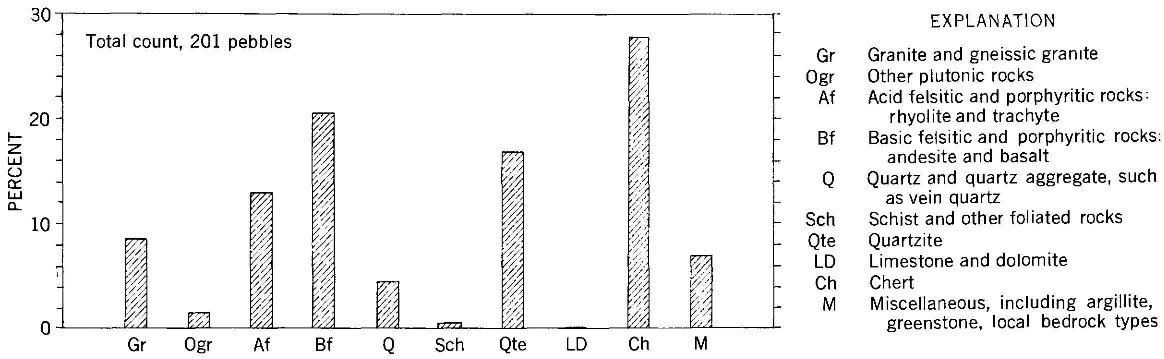


FIGURE 10.—Pebble analysis of Cartwright gravel from about 10 miles northwest of Intake, Mont.

No fossils were found by the writer in the Cartwright gravel. Fragments of bone and a single tooth were found by Collier and Thom (1918, p. 182) in possibly equivalent gravels west of the problem area near Malta, Mont. The tooth was identified by J. W. Gidley of the U.S. National Museum as that of a horse resembling the living species. Gidley did not believe it could be as old as the fauna of the Flaxville gravel (upper Miocene or lower Pliocene).

Alden (1932, p. 44) concluded that the No. 2 bench (Missouri Plateau peneplain) and the gravels which mantle it are Pleistocene in age. He writes:

> If the Flaxville Plain is not itself of Pleistocene age, it seems probable that Pleistocene time in this region began with the uplift which caused the streams to cut below the Flaxville Plain—that is, the Pleistocene began a long time prior to the completion of the second set of terraces or No. 2 bench [Missouri Plateau peneplain] by planation and spreading of the gravel thereon. From the evidence now in hand, it seems necessary to regard the second set of terraces as representing the first of the Pleistocene benches.

Although there is little evidence for precise dating in the northern Great Plains, a Pleistocene age for the Missouri Plateau peneplain seems justified by available evidence. For example, both the Missouri and Yellowstone rivers formerly flowed northeast in broad valleys on the peneplain. Only glacial obstruction seems capable of explaining the diversion of these rivers to the east and southeast. C. R. Warren (1952), as a result of his studies near Chamberlain, S. Dak., suggests that entrenchment of the White River below what is probably the Missouri Plateau peneplain of the present report took place in post-Kansan time. The suggestion is based on the presence of a high-level gravel containing vertebrate fossils identified as late Kansan or early Yarmouth by C. Bertrand Schultz of the Nebraska Museum. Inasmuch as Warren's area is downstream from the area considered in this report, entrenchment of the Missouri River could hardly have begun earlier upstream if the drainage were already integrated. It is possible, however, that the Missouri did not become integrated simultaneously throughout its length. As far as the available evidence in northeastern Montana and northwestern North Dakota is concerned, the Missouri Plateau peneplain could have endured through Yarmouth (Kansan-Illinoian) interglacial time, but indirect evidence favors termination in Kansan time.

The Cartwright gravel is probably equivalent, at least in part, to the Wiota gravel of Jensen (1952, p. 45–50) and the South Saskatchewan gravel of McConnell (1886, p. 70–71). Wiota gravels mapped by Colton (1950) in the Otter Creek quadrangle are shown in plate 1 of the present report as Cartwright (?) gravel.

Although gravel indicates vigorous streams, its presence on a late-mature or old surface is not necessarily contradictory, provided the surface was formed in an arid or semiarid climate. The graded streams of such surfaces may have gradients of hundreds of feet per mile close to the mountains and gradients of several feet per mile even at distances of hundreds of miles. The requisites for such steep gradients are large and coarse loads. Conditions during the latter part of the Tertiary were conducive to high-gradient graded streams. From the Oligocene on, the Rocky Mountains stood high. This is indicated not only by the coarseness of the debris fed to the streams, but also by the change from forest-swamp to grassland conditions on the plains. Such a change, reflecting diminished rainfall, is expectable in the rain shadow of newly risen mountains. The Missouri Plateau peneplain, therefore, was not a typical humid-cycle peneplain. The relatively steep gradients of the graded streams and the coarseness of the alluvium suggest that it was formed under semiarid conditions.

CRANE CREEK GRAVEL

The broadest terrace in the Yellowstone River valley, as much as 2 miles wide, is capped by gravels similar to the older and higher Flaxville and Cartwright gravels. The gravel is herein named the Crane Creek gravel, after the creek of that name located near the town of Crane, in southern Richland County, Mont.[2] Considerations of regional history suggest that the gravel is Yarmouth in age (p. 70).

The gravel is exposed in many places along the terrace scarp (fig. 12*A*), in the banks of side streams trenching the terrace, in road cuts along the main highway between Fairview and Glendive, Mont., and in numerous pits.

Exposures of the gravel have been observed along several of the larger tributaries of the Yellowstone, as in Deer Creek valley south of Glendive, but no attempt was made to trace the gravels throughout the tributary system. It seems reasonable to suppose that the Crane Creek gravel in the tributary valleys was secondarily derived from the higher and older Rimroad, Flaxville, and Cartwright gravels. These secondarily derived gravels contributed to the deposits of the main valley.

The Crane Creek gravel is exposed at three places within the present trench of the Missouri River. This suggests that the Crane cycle of erosion followed the glacial blocking and diversion of the upland drainage. It should be pointed out, however, that only one of these three exposures is in a part of the trench which was unquestionably formed after diversion of the Missouri. This is the exposure along the Missouri south of Bainville, Mont., between the ancestral north-trending courses of the Missouri and Yellowstone Rivers (pl. 1). The second exposure, 5 to 6 miles southwest of Williston, is in a stretch of the Missouri valley which locally coincides with the preglacial course of the Yellowstone River. As far as this part of the valley is concerned, the trench within which the gravel was deposited could have been eroded along the ancestral course of the north-flowing Yellowstone before glacial diversion, or it could have been eroded afterwards along the diversion course. The third exposure is south of Hofflund, N. Dak., where the trench of the Missouri River may coincide with the ancestral course of Little Missouri River valley. Evidence will be presented later, however, indicating that, at both the second and third localities, the streams on the north side of the Missouri valley were already flowing south into the newly acquired course of the Missouri at the time of deposition of the Crane Creek gravel. This supports the conclusion reached for the first locality, namely that the Crane Creek gravel was deposited by the Missouri in its diversion trench rather than by the master streams still following their ancestral courses to the north.

The Crane Creek gravel is superficially similar to the other nonglacial or largely nonglacial gravels of the region. The bulk of the pebbles are brown and waterworn and are of resistant types. The texture varies from layer to layer, but the average pebble size probably exceeds 1 inch and there are scattered pebbles as large as 6 inches. The gravel locally includes fragments of bedrock over a foot long, as well as layers of sand. A typical exposure is in a gravel pit along the main highway about 3 miles south of Sidney, Mont., but the proportions of sand and gravel, as well as the texture of the gravel layers, differ from one exposure to the next. No fossils have yet been found in these gravels.

The Crane Creek gravel does not contain glacial pebbles south of the glacial limit, nor are glacial pebbles abundant to the north. Figure 11 is a graphic summary of the lithology of two samples of the gravel, one from near Intake, Mont., close to the glacial limit, and another from a gravel pit 3 miles south of Sidney, within the glacial limit. Except for

[2] The best exposures of the gravel are nearer the town of Sidney, Mont., 10 miles north of Crane, but the name "Sidney gravel" has already been applied to a deposit in Nebraska.

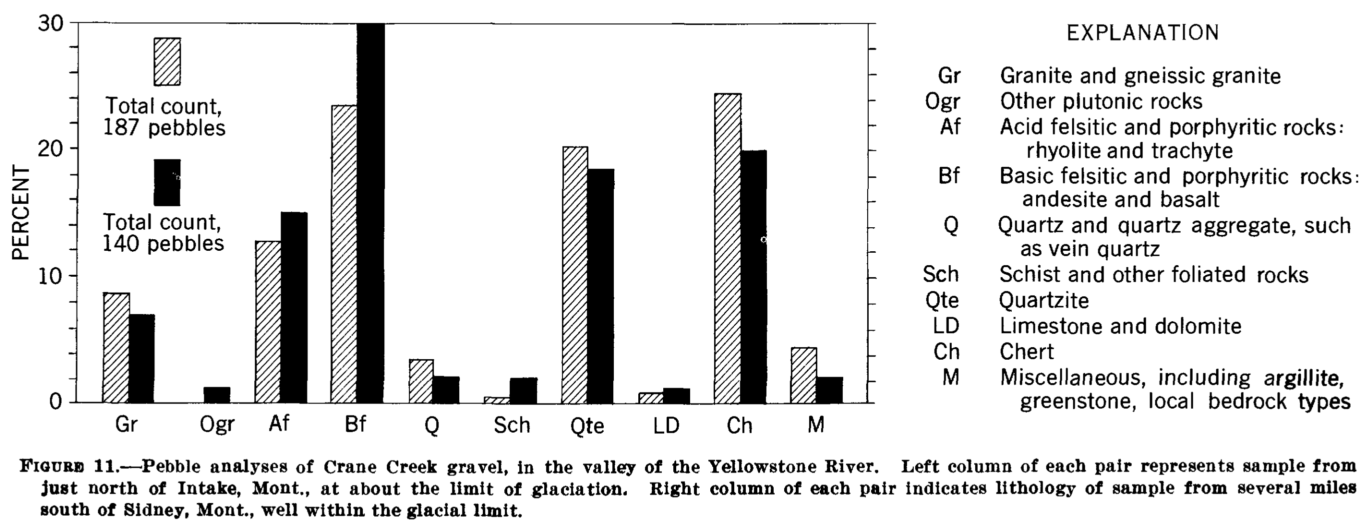


FIGURE 11.—Pebble analyses of Crane Creek gravel, in the valley of the Yellowstone River. Left column of each pair represents sample from just north of Intake, Mont., at about the limit of glaciation. Right column of each pair indicates lithology of sample from several miles south of Sidney, Mont., well within the glacial limit.

a few pebbles of dolomite such as occur in the glacial drift and a few pebbles of granite and metamorphic foliated rocks, the histograms differ little from each other or from the histograms of higher and older gravels beyond the reach of the ice or glacial melt waters.

Like the lithology of the higher and older gravels of the region, the lithology of the Crane Creek gravel depends on whether it was brought into the area by the Yellowstone or the Missouri Rivers. The gravels of the latter consist almost entirely of quartzite, chert, and quartz (exclusive of the small percentage of glacial pebbles), whereas the deposits of the former include a wide variety of igneous rocks. Among these igneous rocks are many porphyritic and nonporphyritic volcanic types, especially andesite, and scattered pebbles of acidic and basic plutonic types including granite. Very rarely is a limestone or metamorphic foliated rock found.

The height of the frontal scarp of the Crane terrace generally ranges from about 20 to 50 feet above the flood plain, depending on the amount of lateral undercutting of the sloping surface. The back of the terrace, at the bluffs, is locally more than 80 feet above the flood plain, so that, theoretically, the frontal scarp might attain that height. Where the scarp rises directly from the low-water plain, a height of 100 feet is possible.

In the lower, outer portions of the bench, the gravels lie below a thin mantle of silt and rest on a level bedrock surface. The gravels are uniformly thin, the average thickness probably being less than 15 feet. At least in its outer portion, the Crane terrace is a gravel-veneered strath. The mantle of debris which accounts for the rising surface between the flat distal portions and the bluffs consists in large part of slope wash. In the Yellowstone River valley the overburden may include lacustrine deposits of a former glacial lake (Lake Glendive).

The elevation of the Crane Creek gravel ranges from about 35 to 70 feet above the low-water plain (fig. 12*A*). It seems likely that the Yellowstone River, during much of the development of the terrace, was regrading downward. At the Lewis and Clark Bridge, in the Missouri River valley, a few miles southwest of Williston, a step in the gravels suggests that, locally at least, the regrading stream left shallow terraces in its wake.

Where lowest and flattest, the gravel surface is 15 to 20 feet above the flood plain or 35 to 40 feet above low-water level. This is presumably the level at which the Yellowstone River finally reached grade in the Crane cycle. An appreciable halt at the level of grade is indicated by the breadth of the distal flat portions of the terrace and by the local impingement of the low-level gravel sheet against the valley-side bluffs. About 4 miles south of Intake, Mont., the road and railroad follow a narrow strip of flood plain between the low-water plain of the Yellowstone River and a high scarp. The scarp rises more than 100 feet above the low-water plain. The upper two-thirds consists of laminated silts with a few scattered pebbles and, near the base, several thin seams of placer coal. Several feet of gravel is exposed below the silts, the base of the gravel being concealed under wash. The gravel is the Crane Creek gravel, which occurs at the same elevation north and south of this large crescentic exposure. Gullies extending back into this narrow terrace reveal bedrock immediately back of the gravels and silts. The Yellowstone River apparently impinged against the valley-side bluffs at this point while flowing at grade in the Crane cycle.

For several miles north of the glacial limit in the Yellowstone valley, only scattered boulders overlie the Crane Creek gravel beneath a mantle of silt. Northward, however, till is found more and more frequently resting on the gravels. Adjacent to steep slopes, the till, in turn, is mantled by slope wash.

In contrast with the Yellowstone valley, the entire course of the Missouri valley in the problem area is north of the glacial limit. Almost without exception, therefore, the Crane Creek gravel is covered by the early Wisconsin(?) till, as in the exposures between Wolf Point and Chelsea Creek in Montana (fig. 12*B*), in the railroad cut immediately under and to the east of the northern abutment of Lewis and Clark Bridge southwest of Williston, and at the mouth of Tobacco Garden Creek in northern McKenzie County, North Dak. Locally, the rock bench is preserved, but the gravel is missing.

The gravel plain had been trenched to form the Crane terrace prior to advance of the early Wisconsin(?) ice. This is indicated by the fact that the till not only mantles the terrace but fills deep valleys cut into it and extends well below the present floor of the Missouri valley. The Crane terrace, prior to advance of the early Wisconsin(?) ice, was much higher than it is now.

POSSIBLE PRE-WISCONSIN DRIFT

TILL(?) AT SMOKE CREEK

The only exposure of a possible till older than the early Wisconsin(?) drift is in Roosevelt County, Mont., on the east side of Smoke Creek valley about 2 miles above its mouth (fig. 14). The description of the deposit which follows is a composite of observations by Witkind (1959, p. 18, 19, 27) and the writer. The essentials are summarized in figure 13.

A

B

FIGURE 12.—*A*, Edge of Crane terrace opposite Glendive, Mont. The Crane Creek gravel (*b*) forms the slope between the silt bluffs (*a*) above and the bedrock (*c*) below and is about 70 feet above the low-water plain of the Yellowstone River. At least two genetic types of silt may be represented, eolian and alluvial. *B*, Early Wisconsin (?) till (*Qewd*) resting on Crane Creek gravel (*Qcg*) at Wolf Point, Mont.

At this locality the Fort Union formation is overlain by 4 to 7 feet of brown waterworn gravel which resembles the Cartwright and Flaxville gravels. The scattered exposures of the gravel are grouped as one in plate 1. The top of the gravel is about 60 feet above the creek at an absolute elevation of about 2,000 feet. The pebbles average less than 1 inch in size but include some as much as 4 inches in size. Imbrication suggests deposition by a stream flowing southeast in the direction of the present valley. The gravel and the underlying bedrock surface constitute an ancient strath.

Witkind (1959, p. 15) has suggested that the gravel may be the equivalent of the South Saskatchewan gravel of McConnell (1886, p. 70–71). However, Witkind includes all Flaxville-type gravels below the level of the gravel-capped Flaxville Plateaus in this category. The Flaxville-type gravels of the present

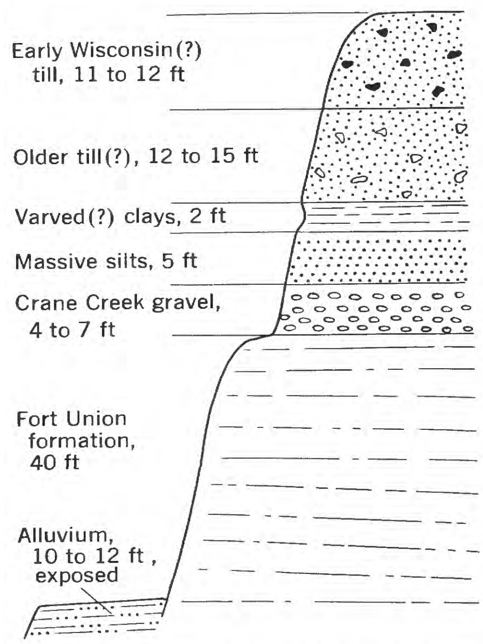


FIGURE 13.—Section including till (?) of pre-Wisconsin age in east wall of Smoke Creek valley about 2 miles above its mouth, sec. 20, T. 30 N., R. 54 E., Roosevelt County, Mont.

FIGURE 14.—Early Wisconsin (?) till resting on older till (?) about 2 miles above the mouth of Smoke Creek, Roosevelt County, Mont. Broken line indicates contact.

report include the Cartwright gravel, which is exposed in belts and patches on the Missouri Plateau peneplain, and the Crane Creek gravel, which forms a bench, frequently covered by glacial debris, in the larger valleys. Although the gravel below the till(?) at Smoke Creek crops out along the line of the ancestral valley of the Missouri River on the Missouri Plateau peneplain, the writer believes it is the Crane Creek gravel rather than the older Cartwright gravel. The reasons are: (1) the gravel contains a greater number of glacial pebbles than the Cartwright gravel in this area, and (2) the decrease in elevation of the gravel patches down Smoke Creek valley, plus the southeasterly imbrication, suggest deposition by a stream crossing the floor of the ancestral Missouri valley. Witkind (1959, p. 13, 15) also believes that the gravel below the lower till(?) is younger than similarly appearing, but probably older, Tertiary or early Pleistocene gravel. He describes gravel on a topographic high south of Medicine Lake which is superficially similar to the gravel along Smoke Creek but which lacks glacial pebbles. It is probably Cartwright gravel.

The gravels at the till locality along Smoke Creek are overlain by 5 feet of massive yellowish silts containing scattered tiny unidentifiable pebbles and small, flat lime nodules. Immediately overlying the massive silt is about 2 feet of varved clay. The origin of the underlying massive silts is not clear, but there seems little reason to doubt the lacustrine origin for the varved clays above.

The till(?) at Smoke Creek rests with sharp contact on the lacustrine clays. The till(?) differs in appearance from all other tills in the area. It is a compact, massive gray-to-tan mixture of silt and clay with widely scattered pebbles and a few striated cobbles. No boulders were observed. Most of the pebbles are stained with manganese oxide and many of the fractures are stained with either iron oxide or manganese oxide. Irregular, but sharply defined, streaks of iron oxide are unrelated to visible fractures. The deposit is calcareous throughout. The pebbles include limestone, dolomite, granite, and other crystalline rocks. The lithology of the deposit is recorded as sample 11, table 1. A recast of the original analysis, omitting local Flaxville-type pebbles, is also included. The recast shows 76.4 percent limestone and dolomite, 20.8 percent granitic rocks, 1.4 percent plutonic igneous rocks other than granite, and 1.4 percent metamorphic foliated rocks.

Resting with sharp contact on the till(?) is 11 to 12 feet of typical dark brown compact clay-rich early Wisconsin (?) till with abundant pebbles and cobbles and scattered boulders. The till is oxidized throughout, but unleached. The lithology is recorded in pebble analysis 10, table 1. The recast shows 82 percent limestone and dolomite, 13.5 percent granitic rocks, and 4.5 percent metamorphic foliated rocks. It is interesting to note that whereas the original samples of the Smoke Creek and early Wisconsin(?) tills differ in the amount of Flaxville-type pebbles (39 and 24 percent, respectively), which in turn affects the percentages of the other pebble types, the differences among the other pebble types are largely eliminated in the recasting.

Both Witkind and the writer have been hesitant about accepting as till the deposit below the early Wisconsin(?) till. The deposit differs from other tills in the region. Its texture is uniformly fine, there are relatively few pebbles and cobbles, no boulders, and the color is extremely light. The absence of boulders is least disturbing, inasmuch as there are large exposures of other tills in which boulders are rare. The scarcity of pebbles and cobbles is unusual, however. The available evidence suggests a pre-Wisconsin age for the problematical till. The deposit is sufficiently important from the standpoint of glacial history to justify detailed consideration of its possible origin and age.

The possibility was first considered that the problematical till is the basal portion of the early Winconsin(?) till enriched by silt from the Fort Union or pre-glacial deposits However, the early Wisconsin(?) till rests on silts of the Fort Union formation at a great number of places in the problem area and on glacial silts at several places and nowhere does it have a comparable basal zone. The hypothesis fails, too, to account for the sharp break above the basal zone and the restriction of manganese to this zone. Furthermore, at this locality the silty till(?) is underlain by clays, not silts.

A second possibility is that the basal zone owes its unusual appearance to deposition in a lake along the front of the advancing early Wisconsin(?) ice which subsequently deposited the clay-rich till above. This fails to explain, however, the scarcity of pebbles and cobbles in the material washed from the advancing ice and the absence of stratification.

Other possible explanations involve a nontill origin. For example, the deposit may represent wash from the slope at the back of the buried gravel terrace. The silt may thus have been derived from the Fort Union formation, and the glacial pebbles may have been contributed from a thin veneer of till or glaciofluvial sediments at the head of the slope The scarcity of the coarse fraction might then be a function of the distance to the till source or the thinness of the glaciofluvial deposits. According to this explanation the deposit is not till, but requires that drift have been present in the vicinity to furnish the included glacial pebbles. If the source material were glaciofluvial, its presence at the edge of the upland overlooking Smoke Creek valley and its absence within the valley itself below the till-like deposit makes it seem unlikely that it could be outwash from the advancing early Wisconsin(?) ice. There is no reason to believe that such material has been flushed out of the valley prior to accumulation of the till-like deposit. It seems reasonable to assume, therefore, that if the till-like deposit were secondarily derived from glaciofluvial sediments, these sediments were pre-early Wisconsin(?). If the source material was till, the till could have been deposited during an earlier advance of the early Wisconsin(?) ice.

The preceding discussion suggests that even though the problematical deposit may have been secondarily derived, a drift source is indicated. A glaciofluvial source favors a pre-early Wisconsin(?) age for the deposit; a till source leaves the question of age open. The problem of whether Witkind's Smoke Creek till(?) (Witkind, 1959, p. 18) was formed within early Wisconsin(?) time or earlier, hinges largely on the interpretation of the break between this deposit and the overlying early Wisconsin(?) till. Unfortunately, there is little additional to be said about this break. There is no recognizable soil on the till(?) at Smoke Creek, although such could have been removed prior to advent of the early Wisconsin(?) ice or by that ice itself. The restriction of manganese to the till(?) at Smoke Creek does not help to define the duration of the time break involved in the unconformity.

The writer believes that considerations of regional history indirectly support the validity of a pre-Wisconsin Smoke Creek glaciation. The cycle of erosion during which the Missouri Plateau peneplain was developed was brought to a close by the early ice advance which diverted the Missouri River to its present course. The following Crane cycle of erosion, during which the broad Crane terrace was developed, was also ultimately brought to a close. The most logical cause is another glacial advance, of which the till(?) at Smoke Creek may be evidence. The writer has no other explanation for the termination of the Crane cycle. It is interesting to note that the deep-trench erosion cycle which followed the Crane cycle was also brought to a close by a glacial advance, the early Wisconsin(?). These episodes are described fully under the heading "Geologic History."

DEPOSITS NEAR BAINVILLE

Several deposits which may be contemporaneous with the till(?) at Smoke Creek are exposed along an unimproved road leading south to the Missouri River from a point about 4 miles west of Bainville, Roosevelt County, Mont. (pl. 1). The southernmost road cut is about half a mile from the Missouri It reveals about 20 feet of gravel and some interbedded silt and sand. The finer deposits are unoxidized, but the degree of oxidation increases with grain size. Because the coarser beds predominate, the whole exposure appears a vivid reddish brown. The lower part of the section is a coarse boulder concentrate consisting of large numbers of rock types similar to the Fort Union formation, but including dolomite, granite, and other glacial pebbles. Some of the boulders are several feet in size. Most of the local types and the matrix are thoroughly impregnated with iron oxide and local rock types are thoroughly rotted. The resistant Flaxville-type pebbles are least affected and the glacial types are fresh or moderately altered depending on their lithology. The strata, which dip to the north, rest on the Fort Union. The material is clearly glaciofluvial, but the degree of oxidation is unlike anything the writer has seen elsewhere in the region.

About 150 feet north of this exposure is another one about 8 feet high. The material resembles that of the preceding section except that the boulder concentrate is missing and the oxidation is more spotty. Even some of the finer layers are partially oxidized.

One and a half miles farther north, a thoroughly oxidized silty till(?) is exposed in a road cut about 25 feet long and 5 feet high. Here too, the local rock types are thoroughly rotted, but the glacial boulders, including metamorphic foliated rocks, are relatively unaltered. This deposit, too, is on the south side of the divide separating the present valley of the Missouri from the abandoned Culbertson-Bainville channel to the north.

Half a mile or so north of the divide there is a fourth exposure resembling the last described.

Those four exposures have been mentioned specifically because of the unusual degree of oxidation and the high degree of alteration of the local pebbles and boulders from the Fort Union. Glacial deposits occur elsewhere in the region in comparable topographic environments, but they do not display these characteristics. Either the deposits are older than those elsewhere or oxidizing conditions are, or have been, unusual.

It is interesting to note that Bauer and Herald (1922, p. 114) have described drift cemented by gypsum and iron oxide at two localities in the Fort Berthold Indian Reservation in the eastern part of the present map area. The drift was assigned a probably Kansan age in contrast to the probably Wisconsin unconsolidated drift of the remainder of their area. The latter drift is tentatively dated as early Wisconsin(?) in the present report. If the heavily oxidized drifts reported by Bauer and Herald and by the writer are actually pre-Wisconsin, their low elevation in the present problem area indicates deposition by ice later than that which diverted the Missouri River at the level of the Missouri Plateau peneplain. The various facets of this problem are discussed more fully in a later section of this report.

EARLY WISCONSIN(?) DRIFT

The early Wisconsin(?) drift includes both till and glaciofluvial deposits. The glaciofluvial clays, silts, sands, and gravels resemble those of the remainder of the Pleistocene section and require no special treatment. However, individual exposures which shed light on the glacial history are described in detail under the heading "Geologic History."

DISTRIBUTION

The early Wisconsin(?) drift is the most widespread in the region (pl. 1). It extends south to the glacial limit and, except possibly in the very easternmost part of the area, is the only drift south of the Missouri River. The southernmost occurrences are in the Yellowstone valley, near Intake, Mont. The drift border was mapped at the southernmost occurrences of glacial boulders except where ice-rafting on proglacial lakes could have taken place. The location of marginal channels was used as supplementary evidence. The soil map of McKenzie County, N. Dak. (Edwards and Ableiter, 1942) was helpful in tracing the drift border in that county; the limit of the drift coincides closely with the limit of the Williams stony loam. The glacial boundary as shown in the southeastern part of plate 1, south of the Little Missouri River, is after Benson (1951).

The drift of northeastern McKenzie County, N. Dak., is mapped as early Wisconsin(?) on the basis of the author's interpretation of the glacial history rather than on direct evidence. West of Keene and Charlson is a northeast-trending morainal ridge, herein referred to as the Keene moraine. Except for a few closed depressions as much as 30 feet deep, the drainage is integrated. The westward-draining valleys of Clear and Sand Creeks cut entirely through the morainal belt and reveal good exposures of the underlying deposits (fig. 15). The till has a high content of limestone and dolomite and on that basis could be either early Wisconsin(?) or Mankato in age. Stratigraphic relations revealed along Sand Creek, however, suggest that the till is early Wisconsin(?).

The most informative section along Sand Creek is in the NE¼ sec. 30, T. 153 N., R. 95 W. Till is exposed at point *A* (fig. 15) about 80 feet above creek level. The upper 8 feet at *A* is massive and thoroughly oxidized. Below this is 8 to 10 feet of stratified till in which some layers are oxidized and others are not. Below the stratified zone, the till is massive and dark gray. Above the till at *A* is about 2 feet of cobbles, followed in turn by horizontal finely bedded gray and buff silts and thin seams of very fine glacial gravel. At *B*, a clearly defined shear zone cuts across the face of the exposure with a westerly dip of 45°. The joint

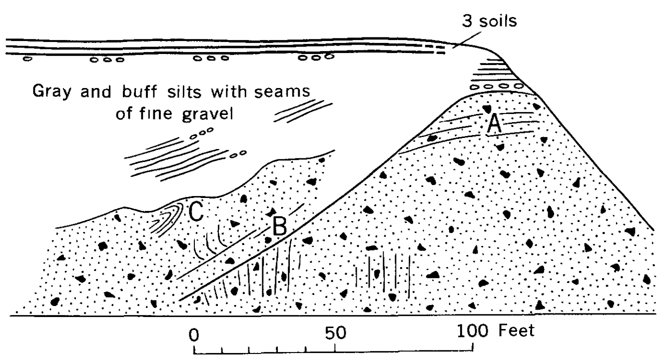


FIGURE 15.—Section along Sand Creek, McKenzie County, N. Dak., about 3 miles west of Charlson. Eighty feet of till is exposed in the 100-foot-high bluff. Letters indicate parts of the exposure described in the text.

columns of till are deformed at the shear zone, presumably as a result of drag during displacement, the upper part having moved down to the west. At *C*, stratified till is contorted. The surface of the till above the shear zone is irregular and generally rises less than half way up the 100-foot bluff. Above the till west of *A*, the gray and buff silts dip in a westerly direction at angles of as much as 35°. The silts and fine gravels are at least 60 feet thick as measured normal to the dip. Above the dipping silts, and presumably resting on their truncated edges, is a discontinuous pebble band. Below the pebble band a lime-rich zone extends about 6 inches into the underlying silts. Above the pebble band there is about 3 feet of uniformly fine silt, including 3 soil profiles. The 2 lower profiles consist of 6-inch black A horizons, underlain by 6- to 8-inch brown B horizons. The uppermost profile is relatively thin compared with those below, consisting of 6 to 8 inches of brown soil alone, with only locally a dark capping.

Till at stream level at a number of places along Sand Creek clearly indicates that the valley was established in its present course prior to deposition of the till. The valley, furthermore, was apparently deeper than it is now. The till did not fill the valley completely, because the stream was able later to resume its course. A sag in the morainal crest at each of the crosscutting valleys supports this view.

The thick section of silts above the till suggests quiet-water deposition. Included in the section are a few thin seams of gravel smaller than pea size. The scarcity of this gravel, and the absence of gravel of coarser texture, is inconsistent with a fluvial origin, because a wide range in pebble size was available from the till mantling the bottom and sides of the valley. The silts are interpreted as being lacustrine.

If the lacustrine interpretation is correct, then only ice could have served as the dam. Deposition of till may have created local barriers in the valley, but there is no evidence of a morainal barrier high enough to account for the high-level silts. Ice on the east side of the moraine could not have formed the dam, because the glacial waters would have flowed uninterruptedly down Sand Creek valley. The facts are readily explained if the ice stood to the west, impounding a lake in the headwaters of Sand Creek.[3]

The cobbly layer immediately below the silts at *A* (fig. 15) may represent material sloughed off the ice front. The steep dip of the silts may be due in part to slumping resulting from withdrawal of the ice on which the silts rested, and in part to slumping of the underlying valley-side till as the receding ice withdrew its support. That the till was wet during slumping is indicated by the bent joint columns and the convolutions at *C* in figure 15. It is unlikely that the slumping took place as a result of undermining after Sand Creek had reexcavated its valley, because the summit surface beveling the dipping beds could hardly have been developed at the top of the bluffs under circumstances similar to those that prevail today. The origin of this surface is a matter of speculation and need not concern us here. It does not affect the principal conclusion that the ice which deposited the Keene moraine came from the west. Inasmuch as there is no evidence that the Mankato ice crossed the Missouri River in the region west of the Keene moraine, it must be concluded that it was the early Wisconsin(?) ice that deposited the moraine. The till of the Keene moraine contains very few boulders over 1 foot long, yet the ground above is cluttered with boulders as much as 8 feet long. The presence of closed depressions nearby, plus the virtually boulder-free character of the till, makes it seem unlikely that the concentration of boulders on the surface is due to downwasting of the moraine. Conceivably, the boulders may constitute the sole record of a thin ice sheet which spread over the area at a later date.

Early Wisconsin(?) ice overtopped Blue Buttes (pl. 1), which stand about 300 feet above their surroundings in northeastern McKenzie County. The crests of the buttes are littered with glacial boulders. Had it been the Mankato ice which overtopped the buttes, the ice sheet would have presumably extended far to the southwest. There is no evidence that the Mankato ice ever had so great an extent.

Higher parts of the Flaxville Plateaus north of the Missouri River are largely driftless and apparently projected above the early Wisconsin(?) ice (Colton and Howard, 1951). The drift borders shown in plate 1 were mapped by the present writer. The reasons for drawing them as shown varied somewhat with the different plateau remnants and will be briefly discussed.

The large plateau remnant north of Wolf Point, Mont., may be referred to as the Wolf Point plateau. No till has been observed within the driftless area shown in plate 1. The southern limit of the driftless area has been drawn along a discontinuous channel which closely parallels the drift boundary. Glacial boulders are scattered north of the channel, but they were probably ice-rafted. If the mapping is correct, the ice stood at an elevation of approximately 2,700 feet on the plateau surface.

The large upland on which Madoc and Flaxville, Mont., are located will be referred to as the Madoc plateau, inasmuch as the name "Flaxville" is applied to the entire group of plateaus. The Madoc plateau is narrow and highly irregular; only small shallow lakes could have been impounded on its summit. Because of this, the distribution of boulders can be used as part of the evidence for outlining the drift border. Glacial boulders are increasingly numerous east of Flaxville, and the first till exposure is 1½ miles east of the town below the plateau crest. To the west, boulders are scarce; the westernmost occurrences are 2½ miles from town. Beyond here no boulders were observed to and beyond the west edge of the plateau. Boulders appear along the road south of Flaxville and 10 to 15 feet of till rests on varved clays in Eagle Creek valley, 3½ miles to the south. Apparently, a small lake was impounded in Eagle Creek valley and overflowed to the west by way of a shallow upland channel. The large north-south channel west of Flax-

[3] Leonard (1916b, p. 528), suggests that Dimick Lake (pl. 1) was formed by deposition of till across the path of a northwest-flowing stream. He implies, but does not specifically state, that the ice withdrew to the west.

ville was apparently occupied by melt waters flowing south from the ice front at the north edge of the plateau. Except for the southern boundary, the borders of the Madoc driftless area are probably fairly accurately defined. The ice, therefore, did not cover areas above about 2,785 feet in elevation.

The driftless area of the Peerless plateau was mapped almost exclusively on the basis of the distribution of glacial boulders, although the topography was considered in interpolating between traverse lines. The border of the driftless area is less accurate than those previously described. Elevations were estimated by barometer without benefit of frequent checks at control points. No boulders were found on this southeastward-sloping upland at elevations over about 2,780 feet. Peerless, Mont., itself is at 2,835 feet, and the northern part of the upland may exceed 3,000 feet. Thus, the entire northwestern part of the upland, above an elevation of about 2,780 feet, appears to have projected above the ice.

To the south of the Peerless plateau between the West Fork of the Poplar River and Cottonwood Creek, is an upland that reaches elevations of over 2,900 feet. Traverses revealed no glacial boulders, nor were boulders found in the crossing of the valley which splits the upland.

In the extreme northwest corner of the map area is the upland on which Carbert, Mont., is located. This upland connects with the Peerless upland just outside the map area. A barometric elevation of 2,970 feet was recorded a few miles west of Carbert, which was as far as the traverse went. No glacier boulders were found. Boulders, however, are common below the upland at barometric elevations up to 2,870 feet.

Some plateau remnants were not visited, hence it its not known whether they were covered by the ice.

Thus, at the height of the early Wisconsin(?) glaciation, the highest parts of the Flaxville Plateaus projected as islands above the ice. These now constitute driftless areas in an otherwise intensively glaciated region.

PHYSICAL CHARACTERISTICS OF TILL

Texture and structure.—The till is nearly everywhere a massive clay-rich deposit with scattered pebbles and cobbles but relatively few boulders (fig. 16). The till is well exposed in Jones cut, a railroad cut about 5 miles southwest of Williston, N. Dak. The section also includes Pleistocene deposits of several other ages. These are described in appropriate sections of the text. The complete stratigraphy of the cut is described on pages 98 to 101.

FIGURE 16.—Early Wisconsin(?) till along Poplar River, Mont., about 14 miles north of U.S. Highway 2. The section has slumped several feet, but without noticeable internal disturbance. One of the blocks of loose sediment within the till is outlined with a dotted line and is indicated by an arrow. Note the scarcity of large boulders.

The dry surface of the till is marked by irregular, hackly polygons from a fraction of an inch to several inches across. Examination of scores of exposures indicates that the higher the clay content, the smaller the polygons.

The early Wisconsin(?) till is not everywhere clay rich. Near Brockton, Mont., it is silty, the boulder content is higher, and weathered surfaces display a faint stratification (fig. 17). The stratification is not visible in the fresh till below the weathered surface. Hand-lens examination reveals no obvious textural

FIGURE 17.—Early Wisconsin(?) till along U.S. Highway 2 about 6 miles east of Brockton, Mont. The exposures in this area are more bouldery than those elsewhere. Note the faint stratification.

differences, but these might be revealed in petrographic or mechanical studies.

Locally the till contains thin seams of dark-gray, yellow, or pink silty clay. The clay follows shears and probably represents gouge.

In several exposures the till is crumpled as well as sheared (fig. 24). The deformation is clearly due to slumping, although in some places it may be the result of ice shove. The deformation of columnar structure and the development of small folds indicate that the till was wet and plastic when deformed.

Color.—Leonard (1916b, p. 529) described the color of the unaltered till as dark gray in contrast with the bluish-gray color of the late Wisconsin till. Although this color contrast may apply regionally, numerous local variations render it of doubtful value in differentiating these tills. Unaltered middle Wisconsin(?) till has not been observed. Oxidized portions of the early Wisconsin(?) till are generally darker than both the middle Wisconsin(?) and Mankato tills, but again, there are many exceptions.

Oxidation.—Oxidation has proceeded to depths of as much as 50 feet, but the average depth is probably 30 or 35 feet. A number of deeply oxidized exposures accessible by road are listed below: [4]

1. Missouri River valley between the White Earth River and Little Knife Creek, Mountrail County, N. Dak. Numerous exposures in the terrace scarp along the valley road show oxidation locally to depths of 50 feet, but spots of unaltered till are frequently present at the base.

2. Poplar River, Mont., about 14 miles north of U.S. Highway 2. An exposure in the riverbank reveals oxidation to depths of 45 to 50 feet (fig. 16).

3. Clear Creek, McKenzie County, N. Dak., about 5 miles west of Keene. About 50 feet of till is exposed in a bluff on the north side of the valley. The upper 30 feet is thoroughly oxidized, the next 15 feet is transitional, the basal 5 feet is largely unaltered but contains a few streaks of oxidized material.

4. Jones cut, a railroad cut about 5 miles southwest of Williston and about half a mile west of the Lewis and Clark Bridge. The entire 23-foot exposure is oxidized.

5. A road cut along U.S. Highway 2 just east of Boxelder Creek about 3 miles east of Brockton, Mont. Except for the lower few feet of the 35-foot section, the till is thoroughly oxidized; the basal few feet is partially oxidized.

[4] At Froid, Mont., an 18-inch-diameter auger hole penetrated 70 feet of till before being stopped by a glacial boulder. The upper 47 feet is thoroughly oxidized.

Below the zone of complete oxidation is a mottled gray and brown zone of partial oxidation. The oxidation is largely confined to the vicinity of joint cracks but frequently forms irregular patches.

Leaching.—The early Wisconsin(?) till is generally unleached. Only one exposure, in an excavation for a house in Froid, Mont., showed superficial leaching. Here, 4 to 6 inches of loess overlies a soil developed in the till. Both the loess and the dark soil A horizon of the till are leached.

Caliche crusts on pebbles.—Many of the pebbles in this and later tills bear a crust of calcium carbonate on their undersides. Many measurements were made to test the possibility that the thickness of the crusts is related to the age of the till. The investigation proved fruitless; there were frequently greater variations within a single exposure than in separate exposures of probably different tills (Howard, 1946, 1947).

Gypsum content.—Gypsum, generally in the form of selenite, is common in many places. The selenite occurs as individual blades, or as ball-like or more irregular agglomerations of blades. An easily accessible exposure where selenite may be observed is about 1 mile east of Williston, where the road to Spring Brook, N. Dak., leaves Little Muddy valley. Here, in a roadside ditch (fig. 19), the early Wisconsin(?) till is exposed beneath a later drift and contains clusters of selenite as much as one-half inch in size. It is interesting to note that the overlying middle Wisconsin(?) till contains no visible selenite.

Thickness.—The thickness of the early Wisconsin(?) till is highly variable over short distances, depending on the underlying topography. Exposures 30 feet thick are common, and several as much as 50 feet thick are known. At Froid, Mont., the auger hole referred to earlier revealed a minimum thickness of 70 feet. The thickest exposed section, 80 feet, is along Sand Creek in McKenzie County, N. Dak.

Lithology.—The lithology of the till, based only on the content of probable glacial erratics, is shown in figure 18. The approximate percentages of the lithologic types are as follows: limestone and dolomite, 71 percent; granitic rocks, 20 percent; metamorphic foliated rocks, 6 percent; and plutonic rocks other than granite, 2 percent. Figure 18 is based on the recast analyses of table 1. Comparison of the original and recast analyses of any one of the 77 till samples analyzed may be made by finding the sample number in plate 5 and referring to table 1.

In five places the early Wisconsin(?) till underlies the middle Wisconsin(?) till. The lithology of the

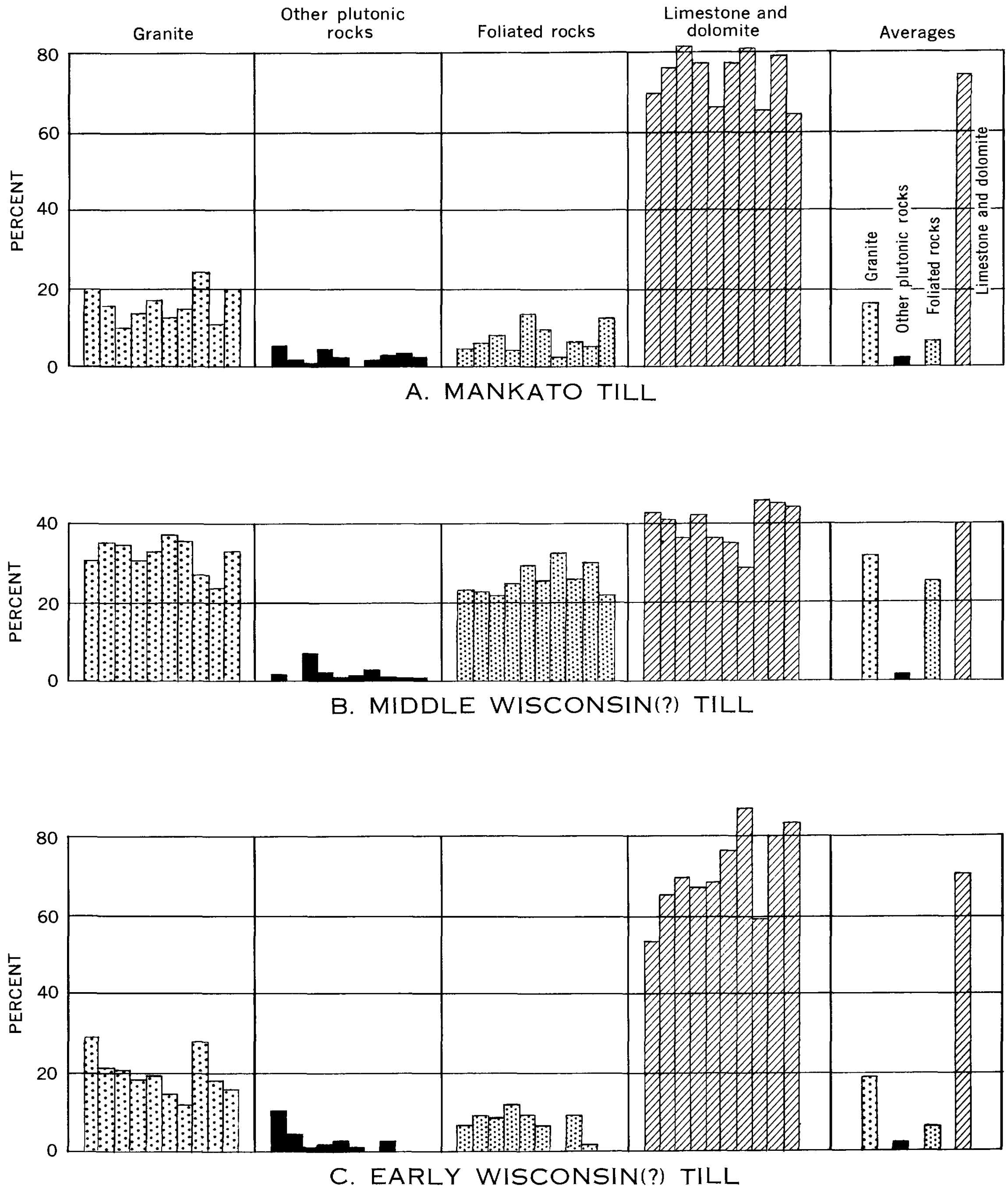


FIGURE 18.—Comparison of glacial-pebble content of Mankato, middle Wisconsin(?) and early Wisconsin(?) tills as revealed in 10 samples of each. The samples of the 2 younger tills were chosen at random. The first 4 samples of the oldest till are stratigraphically below the middle Wisconsin(?) till. The remaining 6 are far removed geographically from the Mankato drift.

TABLE 4.—*Glacial-pebble content of early Wisconsin(?) till*

[Based on five exposures stratigraphically below the middle Wisconsin(?) till]

Sample (table 1 and pl. 5)		Granitic rocks	Other plutonic rocks	Metamorphic foliated rocks	Limestone and dolomite
Number	Number of pebbles [1]				
42	117	20. 5	0. 9	8. 5	70. 1
44	86	29. 1	10. 5	7. 0	53. 5
51	90	21. 1	4. 4	8. 9	65. 6
53	192	18. 7	1. 0	18. 2	61. 9
62	156	18. 6	1. 9	12. 2	67. 3
Average		21. 6	3. 7	11. 0	63. 7

[1] After elimination of local types.

samples is given in table 4. The averages differ somewhat from those given above primarily because of individual variations in the few samples represented.

Analysis of the original complete samples reveals an expectably close correlation between the content of Flaxville-type pebbles and the principal areas of exposure of these gravels. The high concentration of Flaxville-type pebbles in the northwest is obviously related to the vast expanses of gravel capping the Flaxville Plateaus (pl. 1). The high concentration along the Yellowstone River is due primarily to the extensive exposures of Cartwright and Crane Creek gravels capping broad terraces along the valley. The third area of high concentration, southeast of the Yellowstone-Missouri confluence, is not related to large gravel exposures. The "high" straddles the ancestral path of the Little Missouri River, along the course of which only a few areas of Flaxville-type pebbles have been mapped. The presence of the "high" suggests that additional gravels may lie buried under the till or may have been overlooked in the present reconnaissance.

Over much of the north-central and northeastern part of the map area, the early Wisconsin(?) till contains less than 10 percent of Flaxville-type pebbles. Within the Little Muddy Creek drainage basin, however, where the early Wisconsin(?) till is mantled by a younger drift, it contains as much as 57 percent of Flaxville-type pebbles. The high concentration coincides with exposures of Flaxville-type pebbles marking the ancestral path of the Yellowstone River northward across this area.

Pebble orientation.—The pebbles in some exposures of the early Wisconsin(?) till show a preferred orientation. Platy pebbles lie flat and, in at least one exposure, the long axes seem to be parallel. Pebble orientation studies may provide a more detailed picture of ice movement in the problem area than is now available. With one exception, however, all known exposures in which pebbles are oriented are so located that the orientation was probably due to local topographic influences. The one exception is at the edge of the upland on the west side of the White Earth River valley. This, and similar upland exposures, may justify further study.

Pebble bands.—Pebble bands appear in a number of exposures of the early Wisconsin(?) till. Their significance is not always clear. Some appear to be lag gravels between tills of two different ages; others may represent "films" of superglacial debris deposited during slight withdrawals of the ice; and still others may represent clusters of englacial pebbles spread out in "plastering" deposition of lodgement till. Two occurrences are described below.

The first exposure is along U.S. Highway 2, about 1½ miles west of Culbertson, Mont., where the road ascends the divide between Culbertson and Big Muddy Creek valley. The early Wisconsin(?) till here stands in vertical cuts, one of which, on the east side of the road, is about 20 feet high. A pebble band is exposed near the top of this cut. The band is about 2 feet below the surface at the right side of the exposure, and here the pebble concentration is greatest. Even here, however, the band is not a continuous layer; the pebbles and cobbles, and one boulder over a foot long, are simply closer together than they are elsewhere. Tabular pebbles generally lie flat. The pebble band is increasingly discontinuous to the left where it lies at a uniform depth of about 4 feet below the surface. The till itself seems to be lighter in color along the line of pebbles. The pebble band could not be examined in detail because of the vertical slope. The material above the band on the right side of the exposure is probably a mixture of wash, loess, and lag gravels. To the left, however, the pebble band lies within the till. It is unknown whether any of the pebbles are wind abraded. "Plastering" deposition of the pebble band seems to be the most reasonable explanation here.

The second exposure is along U.S. Highway 2 at the edge of the upland on the west side of the White Earth River valley, North Dakota. Here, at the head of a long landslide slope, is a great crescentic scarp of till about 600 feet long. Except at the west end of the exposure where the Fort Union is exposed below the till, the base of the till scarp is mantled by slope wash and till talus. In the central third of the exposure, there is a discontinuous boulder band which is about 10 feet below the surface at its east end and about 20 feet below at the west end. Most of the boulders are granite, some of which are 3 feet across. In the very center of the scarp, the boulder band is a gravel concentrate 1 to 1½ feet thick. This is a far greater concentation of coarse materials than exists on the surface of the exposure today. The till below the boulder band is compact and clayey. It has relatively few pebbles, but a number of these show a preferred orientation as noted earlier. The upper foot or so of the clayey till, which is identical in appearance to the early Wisconsin(?) till elsewhere, contains seams of pinkish clay with scattered tiny pebbles. These seams are locally convoluted. The till immediately below the boulder band is unleached. As at Culbertson, Mont., the boulder band is largely out of reach, so that the presence or absence of wind abrasion could not be determined. The till above the boulder band is somewhat sandier than that below. In one place it includes a 3-inch layer of buff sand. The sandy phase extends from 2 to possibly 6 feet above the boulder band. Above this, the till is again more clayey. Metamorphic foliated rocks are uncommon above and below the boulder band. The upper till is unlike the middle Wisconsin(?) till in the relative scarcity of boulders and in the low metamorphic-rock content (estimated). In the writer's opinion the boulder band is a lag gravel separating early Wisconsin(?) till below from Mankato above. In the discussion of the Mankato drift, a clearer example of superposition of similar drifts at a locality about 6 miles away will be described.

UNDERLYING TOPOGRAPHY AND GEOLOGY

The topographic surface on which the early Wisconsin(?) till rests is level to highly irregular. The irregular floor is found marginal to valleys trenched below the Missouri Plateau peneplain or peripheral to plateaus which rise above it. The level floor represents valley terraces or flat upland areas.

The till is draped over more than 1,200 feet of relief, hence it straddles all stratigraphic horizons; it rests on Cretaceous and Tertiary bedrock; on Flaxville, Cartwright, and Crane Creek gravels; and on glaciofluvial and lacustrine sediments.

In a number of places, the till rests with sharp and level contact on coal beds. Lignite coal weathers and erodes easily; nowhere does it cap flat areas in the modern landscape. The beds are porous, however, and an ice cement during glacial climates may have imparted greater erosional resistance to the lignite layers than to the overlying deposits. Thus, periglacial streams or the ice itself may have experienced less difficulty in removing the overburden than the coal. The incorporation of fragments of lignite in the till may find similar explanation.

Much of the till rests with sharp and level contact on the Flaxville, Cartwright, and Crane Creek gravels (fig. 12*B*). There is no reason to believe, however, that these were buried under other deposits and that selective erosion was necessary to resurrect them.

In many localities, the early Wisconsin(?) till rests directly on proglacial sediments. The sediments range from coarse gravel to delicately laminated, locally varved silts. Proglacial sediments are exposed beneath the till of the Poplar River bluffs (fig. 16), beneath the till west of Jones cut (fig. 22), and below the southernmost exposure of till known to the writer (Yellowstone River valley, 2½ to 3 miles northwest of Intake, Mont., sec. 14, T. 18 N., R. 56 E).

The proglacial deposits in some places are crumpled and sheared. Locally this appears to be due to deformation by the ice, because till is kneaded into the sediments. Such kneading and deformation may be observed in a road cut 6 miles north and 1 mile west of Williston, in the northwest corner of sec. 23, T. 155 N., R. 101 W. An even more complicated situation is revealed along the main highway in a cut on the west side of Tule Creek, in Roosevelt County, Mont. Here the till is draped over a confused agglomeration of intermixed nonglacial and glacial materials. Periglacial frost activity may be indicated.

Locally, the early Wisconsin(?) till contains blocks of sediment so incoherent that they can be scooped out with the hand. Some of these blocks are large. In the bluff along Poplar River (fig. 16), one block of loose sand and gravel measures 10 by 4 feet. It is difficult to imagine how such blocks of loose sediment could have been incorporated in the till unless they were frozen solid.

SURFACE TOPOGRAPHY

Morainal topography is rare in areas of early Wisconsin(?) drift. Several of the larger moraines are shown in plates 1 and 2. Outside the morainal areas, the drainage is completely integrated and the surface of the till forms smooth slopes descending from the divides toward the valley bottoms. The till surface generally terminates as a terrace above present flood plains, or above intermediate terraces. Kames and eskers are not unkown in the area of the early Wisconsin(?) drift. An esker and an associated pitted valley train are shown in plate 1, in T. 23 N., R. 56 E., Richland County, Mont. Others are mapped in northeastern McCone County across the Missouri from Poplar, Mont. The kames shown in plate 1 probably represent only a part of those actually present. Melt-water channels of all dimensions occur throughout the area. Many of these were carved along the ice margin; others were eroded by proglacial waters flowing away from the ice front. The topography of the early Wisconsin(?) drift will be considered in more detail in the section dealing with the early Wisconsin(?) episode of glaciation.

DIFFERENTIATION

Except where it overlies the till(?) at Smoke Creek, the early Wisconsin(?) drift rests on nonglacial deposits from which it is easily distinguished. Most of the evidence for differentiating this drift from younger drifts is considered in the sections dealing with the middle Wisconsin(?) and Mankato drifts.

Prior to the work of Alden (1932) and Alpha,[5] only the early Wisconsin(?) drift of this report and the morainal drift of the coteau belt were recognized. Early workers placed a great deal of emphasis on topography in differentiating the drifts. Leonard (1916b), for example, emphasized the contrast between the level, well-drained topography of the older drift and the prominent morainal topography of the younger. Although surface appearances alone are probably never valid evidence for differentiating drift sheets, they may constitute valid evidence when supported by considerations of bedrock topography and glacial history. It is believed that such considerations in the present problem area constitute a strong argument by themselves for separating the drift herein under consideration—early Wisconsin(?)—from the drift which forms the moraines of the coteau belt. These considerations will be presented, largely in deference to the emphasis that contrasts in drift topography have received in earlier reports.

The presence or absence of morainal topography may depend on factors other than duration of exposure. Some topographic contrasts are original, owing either to variations in the underlying topography, to variations in the rate of recession of the ice, or to variable distribution of debris within the ice. The absence of pre-Illinoian moraines in the Midwest indicates that long exposure eventually leads to obliteration of morainal topography and suggests that drifts displaying morainal topography are probably Illinoian or later. This inference is based on the assumption that the kinds and rates of weathering and erosion during the glacial and interglacial stages of the later Pleistocene did not differ significantly from those during comparable stages of the early Pleistocene.

However, the absence of morainal topography does not stamp a till as old. Young drifts often form level till plains. The till plain may result from uniform deposition on a level surface with obliteration of minor irregularities by outwash. Thus, some areas of young drift, even late Wisconsin, are remarkably level and are topographically indistinguishable from areas of Kansan drift. Here, however, the absence of morainal topography is an original characteristic, just as it probably was in some areas of Kansan drift. It helps little, therefore, in considering the Pleistocene drifts of a small area, to know that, from the regional point of view, post-Kansan drifts frequently display morainal topography whereas earlier drifts do not. Of more importance is the fact that the absence of morainal topography does not prove that the drift of one area is necessarily older than the morainal drift of a neighboring area.

Unfortunately, in evaluating topographic contrasts in an attempt to differentiate drift sheets, it is generally impossible to trace the contrasts to one cause alone. For example, the writer is satisfied that the morainal topography of some parts of the coteau belt is not due to irregular bedrock topography, but to irregular deposition of till on locally level surfaces. The morainal deposits on either side of the White Earth River valley near the eastern boundary of the problem area, rest, at least in part, on a fairly level surface. This is indicated by the uniform level of bedrock exposed in parts of the White Earth valley at the edge of the upland and by records of a number of water wells. Elsewhere, additional evidence is provided by road and railroad cuts. Thus, whereas the greater part of the buried bedrock surface in the coteau belt has considerable relief, level areas remain, particularly at increasing distances from the Missouri escarpment. The presence of moraine on a level bedrock surface does not prove, however, that the morainal deposits are younger than those which form till plains on an equally level bedrock surface beyond. The topographic contrast may be due to uninterrupted advance and recession of the ice across a level area, resulting in a till plain, followed by a halt during which morainal deposits accumulated.

Such "attenuated" drift borders, without a terminal moraine, may lie many miles beyond the outermost morainal ridge formed during a halt in recession of the ice. Under these circumstances, a single drift may present an outer belt quite as level as the plain on which it was deposited and an inner belt which is decidedly morainal in aspect. The ice in the present problem area, however, did not withdraw rapidly from the glacial limit. This is indicated by the great size of the marginal channels at the drift border and by considerations of the history of the ancestral Little

[5] Alpha, A. G., 1935, Geology and ground-water resources of Burke, Divide, Mountrail, and Williams Counties in North Dakota: North Dakota Univ., unpub. thesis, 63 p.

Missouri River, which are discussed in detail under the heading "Geologic History."

In spite of the evidence indicating that the ice front remained for a long time at the glacial limit, there are no moraines to record this long halt. A reasonable explanation of the anomaly is that the deposits at the drift border once did have morainal topography, but that this topography was subsequently obliterated by the processes of gradation. If the above reasoning is valid, then the absence of morainal topography in the outermost drift is valid evidence that it is older than the ruggedly morainal drift of the coteau belt. The topographic contrast of the older drift—early Wisconsin(?)—with the middle Wisconsin(?) drift, however, is not nearly so great, and topography sheds little light on the problem of their relative ages.

AGE OF DRIFT

The exact age of the drift is unkown. Available evidence suggests that it is early Wisconsin(?), that is, Iowan or Tazewellian.

The drift was deposited after glacial diversion of northeast-flowing streams on the Missouri Plateau peneplain and after several hundred feet of subsequent downcutting during the Crane and deep-trench erosion cycles. This is indicated by exposures at river level in the diversion trench of the Missouri River and in the trenches of other streams. If the date of the trench cutting could be established, a maximum age could be assigned to the drift. The date of the trench cutting in turn depends on the age assigned to the Missouri Plateau peneplain. Evidence for a Wisconsin age of the drift will be considered first; the position of the drift within the Wisconsin will be considered subsequently.

WISCONSIN AGE

The Wisconsin age of the drift is based largely on considerations of geologic history. Warren (1952) finds that the Missouri Plateau peneplain in South Dakota is at least as young as the Kansan. According to him, the east-flowing White River eroded its valley 400 feet below the peneplain before the glacial diversion that created the south-flowing Missouri River. At the time of the diversion, the White River stood 100 feet or so above the present Missouri River. Warren believes that the diversion took place in Illinoian time and that the erosion of the deep Missouri trench took place subsequently.

Evidence in northwestern North Dakota and northeastern Montana also indicates two postpeneplanation cycles of erosion: the Crane cycle, which may correspond to the interval during which the White River eroded to within about 100 feet of the present level of the Missouri River, and the deep-trench cycle corresponding to the deep erosion of the diversion course of the Missouri in South Dakota. However, whereas the evidence in South Dakota suggests that the diversion took place during Illinoian time, at the close of the first postpeneplanation cycle of erosion, the evidence in the present area of study suggests that the diversion took place in Kansan time, prior to this cycle of erosion. These conclusions are not necessarily contradictory. The diversion course of the Missouri need not have been acquired simultaneously throughout its length.

The evidence for the Kansan date is as follows: (1) distinctive high-level gravels in Little Muddy Creek valley indicate that the Yellowstone River flowed north across the site of the present diversion trench at the peneplain level (pl. 8), but there is no evidence that the river continued along this course at a lower level; (2) Crane Creek gravel, marking the first postpeneplanation erosion cycle, is exposed within the diversion trench of the Missouri; and (3) evidence, to be presented in a later section, indicates that the tributaries which now enter the Missouri River from the north had already been established in their southerly courses, opposed to the regional slope, at the time of deposition of the Crane Creek gravel. Thus the Missouri was already established in its diversion course during the Crane cycle of erosion, the first postpeneplanation episode of erosion.

If the peneplain is as young as the Kansan, then the several hundred feet of downcutting to the level of the Crane Creek gravel could have taken place in Yarmouth time. This downcutting, therefore, may correlate with that along the former course of the White River which Warren (1952) believes to have been terminated by the advance of the Illinoian ice. The downcutting in the North Dakota-Montana area was probably similarly interrupted by an ice advance. The difference is that in South Dakota this new ice advance was responsible for the drainage diversions which created the Missouri River, whereas farther north this ice found the Missouri already incised in a diversion path determined by an earlier ice. The drift(?) at Smoke Creek, referred to in an earlier section, and believed to be pre-Wisconsin in age, may have been deposited by the ice which interrupted the Crane erosion cycle. If the Crane erosion cycle is the equivalent of that during which the White River eroded its valley 400 feet, then the ice which interrupted the Crane cycle in the north may be the same age (Illinoian) as that which interrupted the equivalent cycle of erosion along the White River. If the above reasoning is valid, then the episode of erosion which followed the Crane cycle, and during which the Missouri River valley was trenched to a depth of as much as 200 feet below present river level, is Sangamon in age. It is far more reasonable to attribute this great depth of erosion to Sangamon time than to the minor interglacial periods of the Wisconsin. The three drifts which postdate this erosion interval, including the one under discussion in this section, are thus interpreted as being Wisconsin.

The only direct evidence of a Wisconsin age is afforded by a meager fresh-water fauna from proglacial sediments below the drift just west of Jones cut, the railroad cut west of Lewis and Clark Bridge near Williston. The fauna was examined by Dr. A. Bryon Leonard of the University of Kansas, who found the assemblage too small for precise dating. He suggested, however, that it might be post-Iowan or even post-Tazewell-Cary in age. For reasons to be presented shortly, a post-Tazewell-Cary age is believed to be too recent.

W. E. Benson (written communication, 1952), working in west-central North Dakota, concluded that the drift in question is Iowan in age, whereas Jensen (1951, p. 25), working west of the present problem area in Montana, believes that the possible equivalent of the drift is Cary in age. Benson believes that he has successfully traced the Iowan drift north from the South Dakota border, to which it had been projected from the south by Flint (1955). Jensen, on the other hand, bases his conclusion on the discovery of a mammoth tooth (*Mammuthus primigenius*) in gravels underlying the drift in the Frazer quadrangle, about 12 miles west of the present map area. Assuming that the drift referred to by Jensen is the same as the one herein considered, the Cary age is doubtful because it would require that both of the later drifts, as well as the intervening cycle of erosion, be Mankato in age.

Jensen (1952, p. 55–56, 84–85; 1951, p. 15–16) believes that the locally faulted and folded sediments which form the rolling upland between the flood plains of the Missouri and Milk Rivers and which form terraces along these valleys were laid down on and against stagnant ice in the Missouri River valley. He attributes the deformation, and the undrained depressions, to slumping during wasting of the ice. Somewhat similar phenomena, but on a smaller scale, are visible at the east end of the Poplar-Brockton terrace in the present problem area. If the ice-wasting interpretation is correct, then the preservation of the undrained depressions might be considered an argument against an age as old as the early Wisconsin(?). However, the surface depressions may be unrelated to the slumping. The depressions near Poplar, Mont., are so shallow that the possibilities of development by other processes, such as channel scour by running water, differential compaction, or deflation, cannot easily be eliminated.

The principle reason for believing that the drift in question is no younger than the Tazewell is the necessity of accounting for the two later drifts. The first of these later drifts—middle Wisconsin(?)—differs lithologically from the early Wisconsin(?) drift and is separated from it by a marked erosion interval. The second is lithologically similar to the early Wisconsin(?) drift but is separated from the middle Wisconsin(?) drift by another significant erosion interval. The differences in lithology and the magnitude of the erosional intervals seem to require more than just pulsational advances of the ice of a single substage.

SUMMARY OF AGE DISCUSSION

Extrapolation of data from Warren's area in South Dakota, plus considerations of regional history and fossil data, suggest that the drift is Wisconsin in age. The presence of two later drifts suggest that it is either Iowan or Tazewell.

If the conclusions regarding the ages of the drifts are valid, then the following geologic history seems indicated. The ancestral north-flowing drainage in northwestern North Dakota and northeastern Montana was diverted at the peneplain level to create the present course of the Missouri River, presumably in Kansan time; in Yarmouth time the Missouri trenched its diversion course to a depth of several hundred feet, down to the level of the Crane strath; invasion by the Illinoian ice brought the erosion during the Yarmouth stage to a close and resulted in deposition of a drift of which the exposure at Smoke Creek may be a remnant; Sangamon erosion then resulted in the deep, inner trench which extends far below present river level; the Sangamon cycle was interrupted by the advent of the first of the Wisconsin glaciers.

MIDDLE WISCONSIN(?) DRIFT

DISTRIBUTION

The middle Wisconsin(?) drift occupies a lobate area of about 800 square miles in northwestern North Dakota between the coteau belt and the Missouri River. The area narrows to the southwest and terminates near the junction of the Missouri and Yellowstone Rivers. Maximum width of the area is 25 to 30 miles; its length is 45 to 50 miles. The drift border shown in plate 1 probably represents the minimum extent of ice.

Except at its southwestern extremity, the drift is restricted to the drainage basin of southward-flowing Little Muddy Creek. A large part of the drainage basin consists of a broad swale in the Missouri Plateau peneplain; the swale probably favored the advance of the ice lobe. The drift has not been identified on the south side of the Missouri River valley.

PHYSICAL CHARACTERISTICS

Texture and structure.—The middle Wisconsin(?) till is generally siltier and has a greater concentration of pebbles, cobbles, and boulders than the other tills of the region. The silty facies is neither as dense nor as tough as the early Wisconsin(?) till on which it rests. Because of the large proportion of silt, exposed surfaces rarely exhibit polygonal mud cracks, and where the polygons do occur they are generally of large size, several inches to more than a foot across, In some exposures, thin sand seams as much as 1 inch thick are included in the till, and the deposit is characterized by a small-scale parting, consisting of wavy layers one-eighth to one-quarter of an inch thick and as much as 2 inches long. The parting is not due to megascopic textural variations, nor is such variation revealed in thin section.

The middle Wisconsin(?) till is not everywhere silty; in places it is a typical boulder clay with the same wavy parting exhibited by the silty facies. One easily accessible exposure is in a deep ditch along the Williston-Spring Brook road where the road ascends from Little Muddy Creek valley to the upland north of the Williston golf course (fig. 19). Typical clay till is also exposed in cuts along the north-south section road crossing the morainal area on the east side of Little Muddy valley 12 to 13 miles north of Williston; along U.S. Highway 85 where it crosses the large morainal area in Little Muddy valley 18 to 20 miles north of Williston; along U. S. Highway 2, a quarter of a mile east of the turnoff to Epping, N. Dak.; and in numerous roadside ditches in the eastern part of the drift area, as, for example, at locality 61, plate 5.

FIGURE 19.—Middle Wisconsin(?) till resting on early Wisconsin(?) till (below head of pick) in ditch along Williston-Spring Brook road on north side of Williston golf course. The middle Wisconsin(?) till at this locality is a typical boulder clay.

Color, oxidation, leaching, gypsum content.—Weathered surfaces of the middle Wisconsin(?) till are light gray, but the till itself is light brown or buff. The writer has nowhere observed a dark-gray unaltered phase. The reason for the complete alteration is probably two-fold: the generally silty texture which favors penetration by oxidizing waters, and the thinness of the drift.

Wherever the present soil is developed directly on the middle Wisconsin(?) till, a leached A horizon is present. Where the till underlies glaciofluvial sediments or loess, it is generally unleached. No exposures have been found in which the till is overlain by Mankato till.

Gypsum is rare in the middle Wisconsin(?) till in contrast with its abundance in the early Wisconsin(?) till.

Thickness.—Observed thicknesses of the middle Wisconsin(?) drift range from a few inches to 15 feet or more and probably average about 5 feet. The amount of relief in morainal areas suggests that the thickness locally exceeds 25 feet and may be considerably more.

Lithology.—The glacial-pebble content of four samples of the middle Wisconsin(?) till is shown in table 5.

TABLE 5.—*Glacial-pebble content of middle Wisconsin(?) till*

[Exposures stratigraphically above early Wisconsin(?) till]

Sample (table 1 and pl. 5)		Granitic rocks	Other plutonic rocks	Metamorphic foliated rocks	Limestone and dolomite
Number	Number of pebbles [1]				
43	173	31.2	1.7	23.7	43.4
50	99	35.3	0	23.1	41.5
52	189	27.0	1.1	25.9	46.0
61	158	33.0	.6	22.1	44.3
Average		31.6	.9	23.7	43.8

[1] After elimination of local types.

The lithology is unlike that of the underlying early Wisconsin(?) till which, in these exposures, averages 22 percent granitic rocks, 11 percent foliated rocks, and 64 percent limestone and dolomite.

The early recognition of this area of lithologically distinctive till prompted a regional study of till lithology (Howard, 1950, 1956). The study revealed that it is the underlying till which is lithologically similar to the till on either side of the middle Wisconsin(?) drift.

Possible alternative explanations of the distinctive lithology, based on the assumption that the till is merely an upper facies of the underlying early Wisconsin(?) till, are considered in the section on "Differentiation."

UNDERLYING TOPOGRAPHY AND GEOLOGY

The middle Wisconsin(?) drift mantles a topography of varied relief. It covers the early Wisconsin(?) drift not only on interstream uplands but on gently sloping valley sides and on terrace flats as well. In addition it is found on the floor of Little Muddy Creek valley below a terrace of early Wisconsin(?) drift (fig. 20) and probably extends well below stream level. The geomorphic relations in Little Muddy valley are described in detail in subsequent pages.

Middle Wisconsin(?) till may be observed resting directly on the early Wisconsin(?) till at a number of places, although generally its base is not revealed. Locally, it rests on glaciofluvial sediments, as in two road cuts just east of Epping, N. Dak., on U.S. Highway 2.

SURFACE TOPOGRAPHY

Nearly everywhere, the surface of the middle Wisconsin(?) drift is level to gently undulating and has thoroughly integrated drainage. Locally, however, the topography is morainal and includes a few lakes and swamps. The two most pronounced morainal areas are approximately 12 and 18 miles north of Williston. Some of the hills in the morainal areas stand 25 feet or more above depression floors.

In steep exposures where the middle Wisconsin(?) till directly overlies the early Wisconsin(?) till, the former is eroded to a more gentle slope than the bluff-forming till below.

Locally the middle Wisconsin(?) till is mantled by glaciofluvial sediments, the deposition of which contributed to the leveling of the till topography. The deposits are generally small but are exploited by farmers for sand and gravel. Ice-contact deposits form a few small hills on the floor of Little Muddy valley. The deposition of loess on the middle Wisconsin(?) drift has had little influence on the topography. The loess is rarely more than 2 to 3 feet thick and is fairly uniformly distributed over the region.

DIFFERENTIATION

The middle Wisconsin(?) till is nearly everywhere easily identified by its silty texture and by the gray color of its weathered surface. These characteristics, however, do not prove that the till differs in age from the clay-rich till below. The gray silty till, in the absence of contrary evidence, could be interpreted as being a superglacial facies of the same drift of which the clay-rich till is a lodgement facies, a possibility that is considered in subsequent pages.

The till is distinguished from the early Wisconsin(?) till on the basis of a number of lines of evidence, no one of which is probably conclusive. These are as follows:

1. Geomorphic relations
2. Distinctive lithology
3. Restricted distribution
4. Absence of similar deposits on Mankato till
5. Probable intervening soil
6. Intervening unconformity
7. Presence of boulder-clay facies

GEOMORPHIC RELATIONS

Throughout the area of study, long, smooth slopes descend from the divides and terminate as a terrace along the larger streams. Over the greater part of the glaciated area, this smooth surface is underlain by the brown—early Wisconsin(?)—till. North of Williston, the brown till is overlain by a thin veneer of gray—middle Wisconsin(?)—till. In places the gray till is absent, or so thin as to be indistinguishable. The surface of the brown till retains its smoothness under the gray till. The terrace alluded to earlier thus consists of brown till locally mantled by gray.

In many places the brown till of the terrace extends down to, and presumably below, stream level. Elsewhere, as at the mouth of the East Fork of Little Muddy Creek, 14 to 15 miles north of Williston, glaciofluvial sediments are exposed below the till. In still other places, the till mantles and drapes over a bedrock terrace, the surface of which is either remarkably level or highly irregular. Even where the bedrock surface is irregular, the surface of the till presents a smooth, graded appearance. It seems apparent, therefore, that whereas the smoothness of the till surface away from the stream valleys may be due to deposition of a uniformly thin veneer of till on a smoothly graded bedrock surface, the smooth surface, where the till buries an irregular bedrock surface, must indicate later grading. Only locally is the smooth surface of the till due to deposition of glaciofluvial sediments in slight initial irregularities; for the most part it has probably been smoothed by streams, slope wash, and mass movements, possibly periglacial.

The till terrace is present in Little Muddy Creek valley north of Williston. It is generally mantled by loess as much as 3 feet thick. In two places, the gray till—middle Wisconsin(?)—displays pronounced morainal topography below the level of the terrace (fig. 20). The probable stratigraphic relations are indicated in the structure sections of figure 20. The valleys, occupied by the Mankato outwash, need not be considered at present. The till relations are not easily explained on the basis of a single superglacial-lodgement till complex. The difficulty is not the failure to identify the gray till on the terrace; it could be present as a thin, not easily recognizable, veneer. The difficulty lies in the abrupt change from the level terrace surface to the steeply sloping frontal scarp against which the gray till of the morainal areas abuts.

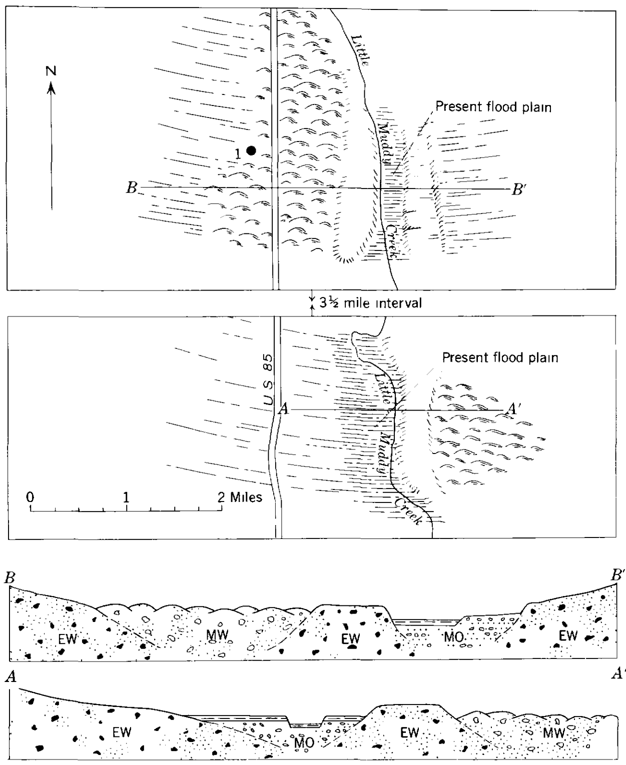


Figure 20.—Geomorphic relations of middle Wisconsin (*MW*) and early Wisconsin (*EW*) tills and Mankato outwash (*MO*) in Little Muddy Creek valley, N. Dak. The 2 sections are 12 and 18 miles north of Williston. A well 70 feet deep at locality 1 is entirely in till, but it is unknown whether more than 1 till is present.

The terrace surface could hardly have been fashioned after deposition of the gray till, because the undrained depressions of the latter lie at lower elevations. Nor is it likely that the terrace surface was graded while stagnant ice covered the morainal areas. It is conceivable that a small area might be graded during the lifetime of a small mass of stagnant ice, but it is hardly likely that the entire region of study, in and out of the area of the gray till, would have been so modified. Nor can it be claimed that the flat surface of the brown till is due to the presence of a flat bedrock surface immediately below a thin veneer of the till, because the till is exposed nearly to creek level in at least the southern terrace remnant.

All but one of the above difficulties inherent in the superglacial-lodgement hypothesis disappear if it is assumed that the gray till of the morainal areas is later than the till of the terrace. The sequence of events then indicated would be as follows:

1. Burial by the brown till of a landscape including a widespread rock terrace
2. Regional development of the graded surface on the brown till
3. Trenching of the graded till surface, creating a till terrace and locally exposing the rock terrace below, and
4. Invasion by a later ice, thicker in the trench than on the terrace, resulting in deposition of the gray moraines and diversion of the Little Muddy Creek to new paths across the terrace.

LITHOLOGIC EVIDENCE

The lithology of the gray—middle Wisconsin(?)—till is compared with that of the brown—early Wisconsin(?)—till in figure 18. The histograms of the gray till show three fairly prominent peaks, representing the granitic rocks, the foliated rocks and the limestone and dolomite. The histograms of the brown till are dominated by limestone and dolomite, with granite a poor second and the foliated rocks negligible. Comparison of the histograms reveal that, on the average, the gray till has only about half the quantity of limestone and dolomite contained in the brown till, but it has 4½ times the quantity of foliated rocks and nearly three-fifths again as much granite. The contrast in lithology, particularly in the content of limestone and dolomite, is not due to different degrees of weathering, because both tills are unaltered except for oxidation. Limestone and dolomite pebbles occur throughout the vertical extent of all exposures.

The histograms of samples 38 and 66 in plate 5 are intermediate in character between those of the middle Wisconsin(?) till and the neighboring tills. Some intermixing of the tills may be indicated.

The possibility was considered that the upper thin silty gray till is a superglacial facies of a single drift of which the lower thick clayey brown till is a lodgement facies and that the contrast in lithology is due to different rates of wear of the pebbles under the different modes of transport and deposition. Thus, more severe wear of the materials of the lodgement till might account not only for the smaller size of the pebbles but also for the increase in the number of limestone and dolomite pebbles. The increase in limestone and dolomite pebbles would in turn account for relative decrease in granite and foliated rocks. Unless, however, the rock types behave quite differently in a frozen condition and under the stresses encountered in a glacier, it is difficult to understand why the foliated rocks did not break up more readily than the limestone and dolomite. The effect of attrition of materials with a high limestone-dolomite content can be observed in the outwash from the latest drift in the area, the lithology of which is identical to that of the brown till. Pebble analyses of the outwash show practically no change in lithologic percentages after miles of stream transport; the histograms are practically identical to those of the till. This point is not stressed, because the behavior of pebbles in ice transport may be quite different from that in stream transport. It is interesting to note, however, that the lithology of the boulder-clay facies of the gray till is virtually identical to the lithology of the silty facies. If the silty facies is due to water-working, the water-working did not noticeably change the lithology.

The same arguments and observations make it seem unlikely that the different lithology of the gray till is due to selective waterwear of the upper part of the brown till in periglacial times.

Finally, the distinctive lithology of the gray drift is not due to local enrichment of the brown drift with crystalline rocks from a buried source (Howard, 1956). The distribution of the gray drift is such that the location of the contaminating crystalline mass could only be under the moraines of the coteau belt. Numerous wells, other excavations, and isolated exposures indicate that the Fort Union formation is more or less continuous under the coteau belt.

The information on the presence of Fort Union beneath the drift of the coteau belt comes from the following sources: R. C. Townsend (written communications, 1947); Lindvall and Hansen (1947); Simpson (1929); Waring and LaRoque (1949); Beekly (1912); and observations by the writer.

Additional evidence against a buried crystalline mass is the fact that neither the early Wisconsin(?) till, where it underlies the highly crystalline till, nor the Mankato till, where it abuts abruptly the highly crystalline till, show a concentration of crystalline rocks. Although the crystalline mass may have been completely buried by older drifts at the time of advent of the Mankato ice, it is unlikely that this was the condition at the time of invasion by the early Wisconsin(?) ice.

An alternative explanation for the distinctive lithology is that the gray till is a separate and distinct deposit rather than a weathered, superglacial, periglacial, or locally contaminated facies of the underlying brown till. The distinctive lithology may merely record a somewhat different avenue of approach of the middle Wisconsin(?) ice as compared with the early Wisconsin(?) ice. Figure 21 shows the location of the middle Wisconsin(?) drift with respect to the belt of Paleozoic limestone and dolomite which flanks the Canadian Shield to the northeast. Ice which crossed the belt obliquely or where the belt is broadest would presumably be more heavily charged with limestone and dolomite than ice crossing the belt by a direct route or where the belt is narrow.

ADDITIONAL ARGUMENTS AGAINST SUPERGLACIAL ORIGIN

RESTRICTED DISTRIBUTION

The distribution of the gray—middle Wisconsin(?)—till was mapped on the basis of lithology and other physical characteristics. The distribution is shown in plate 5, in which are plotted the histograms for most of the till samples analyzed. The histograms include only the granite, metamorphic-rock, and limestone-dolomite groups; the group including plutonic types other than granite rarely averages more than a few percent.

It will be noted that the gray till, as indicated by the distribution of its characteristic lithology, is largely limited to the drainage basin of Little Muddy Creek. This drainage basin forms a broad topographic low trending north-south. A remarkably similar topographic sag occurs immediately to the west in Montana. This sag, which marks the former northeasterly course of the Missouri River across the upland (pl. 8), extends northeastward from the edge of the bluffs overlooking Poplar, Mont., to beyond Medicine Lake. Big Muddy Creek follows the depression for many miles south of Medicine Lake. If the gray till were actually a superglacial facies of the brown till, then it is difficult to understand why its distribution is so restricted, particularly when almost identical topographic relations are present immediately to the west. It might be argued that the superglacial till elsewhere consists only of scattered boulders, hence is not readily identifiable. The problem still remains, however, as to why it should be so prominently developed in one area and so weakly developed in an almost identical topographic setting nearby.

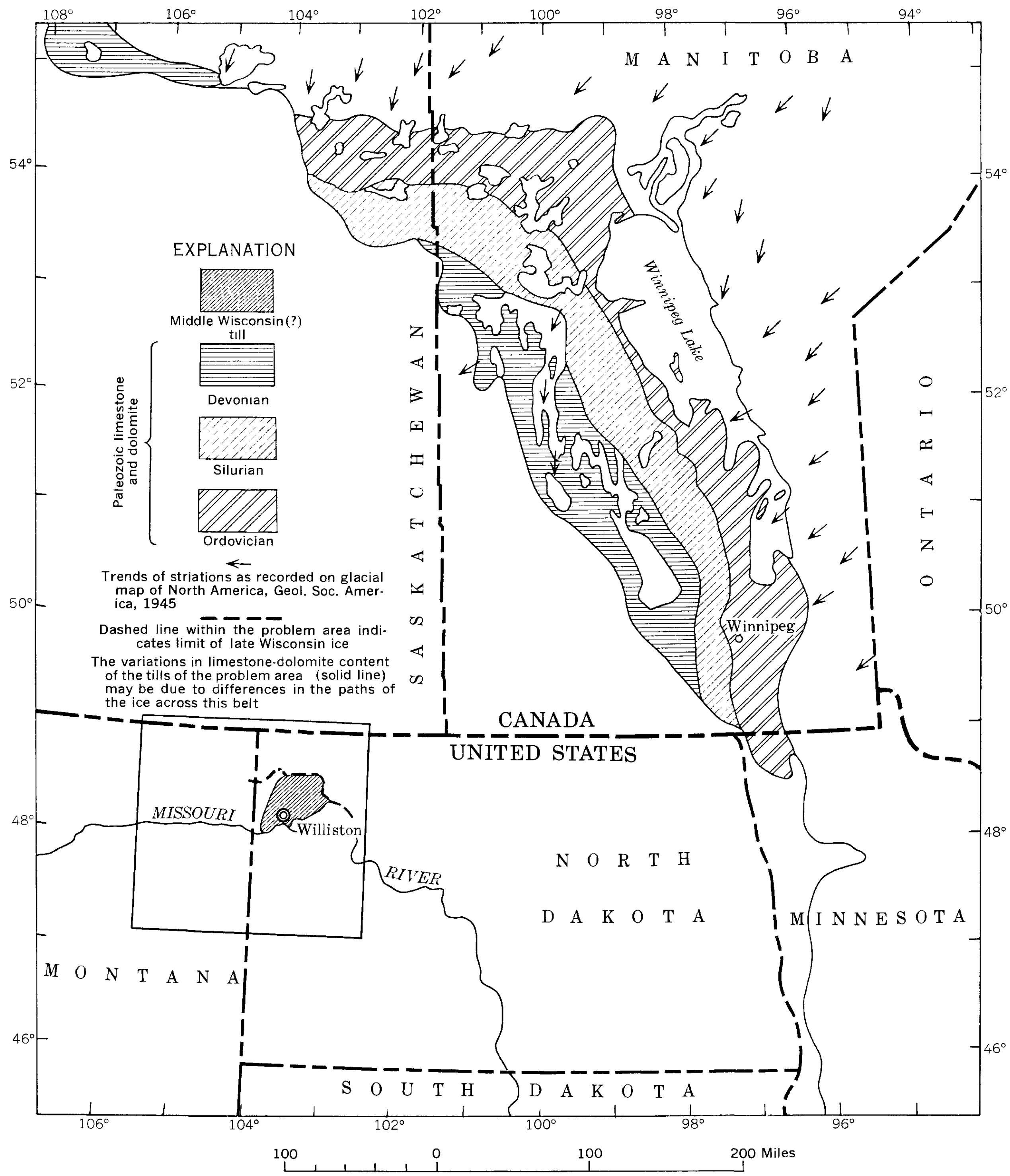


FIGURE 21.—The area of middle Wisconsin(?) till is shown in relation to the Paleozoic limestone and dolomite belt flanking the Canadian Shield to the northeast. The variations in limestone-dolomite content of the tills of the problem area may be due to differences in the paths of the ice across this belt.

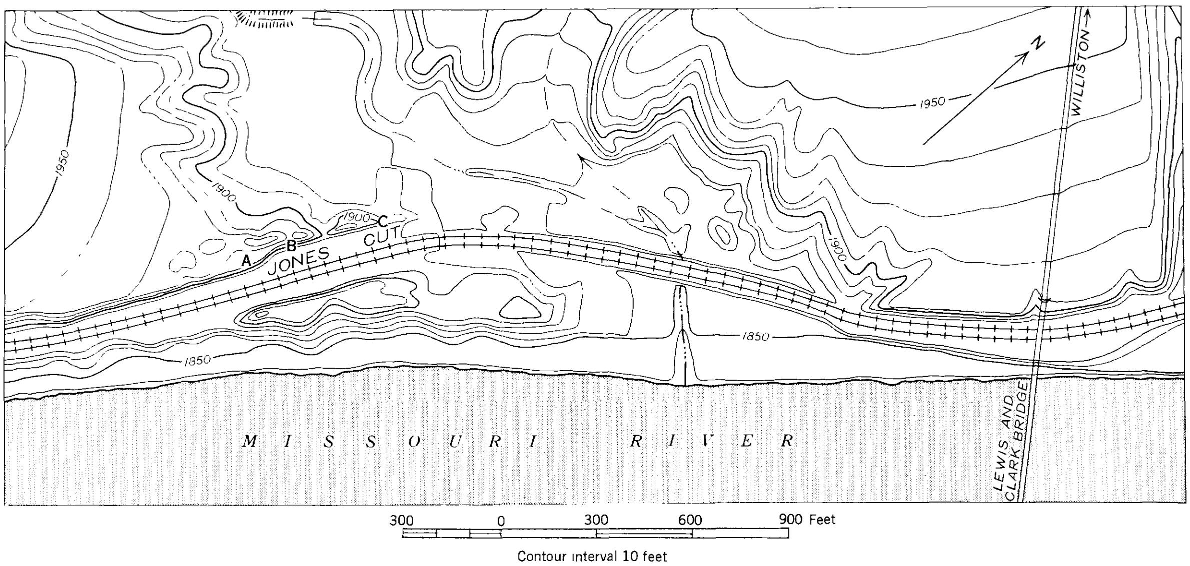


FIGURE 22.—Topographic map of Jones cut and vicinity, about 5½ miles southwest of Williston, N. Dak. *A*, location of the section shown in figure 24; *B–C* location of the section shown in figure 44.

ABSENCE OF SIMILAR DEPOSIT ON MANKATO TILL

The Mankato till fails to show a superglacial-lodgement stratification in topographical environments comparable with that of the gray—middle Wisconsin(?)—till. This is true even in the northern part of the identical lowland occupied by the gray till. It seems unlikely that superglacial moraine would have formed in only one relatively small area of one till sheet. The localization of the gray till presents no problem, however, if it is the deposit of a separate ice sheet, projecting from below a still younger drift.

PROBABLE INTERVENING SOIL

Traces of an intervening soil were found in two of the exposures in which the gray till overlies the brown. One of the exposures is in Jones cut (fig. 22) along the Great Northern Railway about 5½ miles southwest of Williston. The soil is developed at the top of the lower—early Wisconsin(?)—till in the southeast face of the cut (fig. 23). In 1947, James Thorp, then a regional director of the Division of Soil Survey of the Department of Agriculture, visited the area with a party of six Canadian soil scientists: W. E. Bowser, F. Bentley, and W. Odynsky of the Department of Soils, University of Alberta; and J. Clayton, W. Jensen, and H. C. Moss of the Department of Soils, University of Saskatchewan. All members of the party were inclined to agree that a soil was present, although some were less positive than others. At one place some members of the group identified both an A horizon about 1 foot thick and a lighter B horizon 2 to 3 feet thick. The B horizon contained seams of gypsum. Below the B horizon, the lower till had a thin sandy phase which, however, probably reflected glacial rather than soil-forming interglacial conditions. As the "soil" was traced westward along the face of the exposure, the A horizon disappeared and the upper till rested directly on the gypsiferous B horizon. Whether the erosion preceded the advance of the younger ice, or was caused by it, is not obvious.

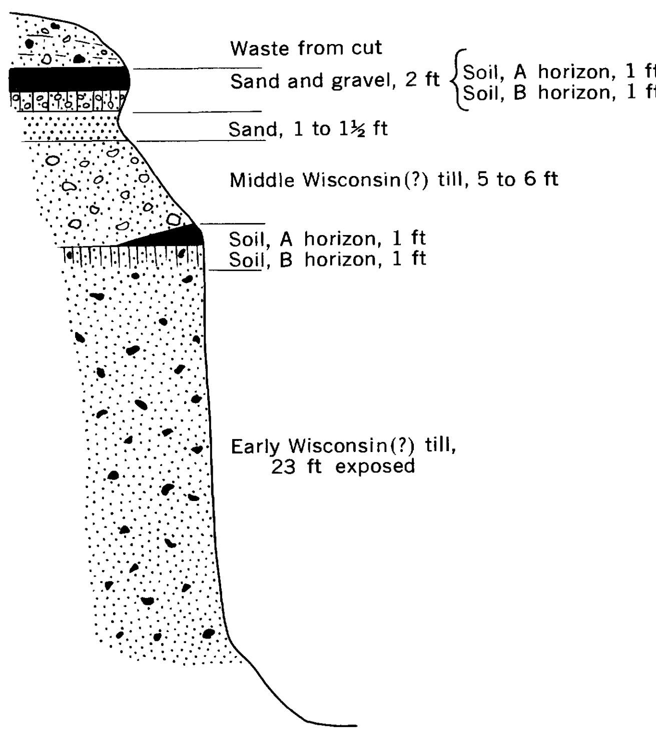


FIGURE 23.—Stratigraphy of southeast face of Jones cut along Great Northern Railway about 5½ miles southwest of Williston and half a mile west of Lewis and Clark Bridge. A pebble analysis of the middle Wisconsin(?) till appears as No. 43 in table 1; an analysis of the early Wisconsin(?) till appears as No. 44.

The party also examined the second exposure at the junction of U.S. Highways 2 and 85, about 10 miles north of Williston and only a few miles southwest of the southern morainal area shown in figure 20. Here, too, a definite break was found at the gypsum level, but no identifiable A horizon was observed.

INTERVENING UNCONFORMITY

A section in the northwest face of Jones cut (fig. 22, loc. *A*, and fig. 24) furnishes additional suggestive evidence that the middle Wisconsin(?) till is distinct from the early Wisconsin(?) till. Here, beneath a thin layer of middle Wisconsin(?) till (*MW*) a shear zone separates two facies of the underlying till. Facies 1 is darker than 2, is presumably more clayey as indicated by the abundant mud cracks, and is prominently jointed. In addition, it contains many more of the brown water-rounded Flaxville-type pebbles such as underlie the early Wisconsin(?) till in many places along this stretch of the Missouri River valley. Near the head of the shear zone, part of a flat-lying fold about 2 to 3 feet high and about 6 feet long is exposed in the till. The simplest explanation of the relations is that facies 2 represents a higher portion of the original till deposit and has slid down alongside 1. The fold could have developed at this time.

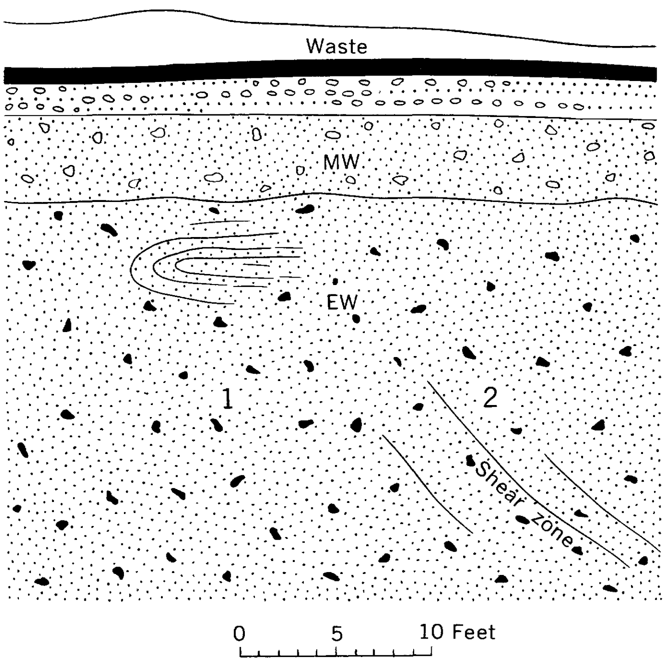


FIGURE 24.—Stratigraphic relations at the southwest end of Jones cut (fig. 22, loc. *A*). The till beneath the middle Wisconsin(?) till (*MW*) consists of two facies separated by a shear zone. A flat-lying fold about 6 feet long and 2 to 3 feet thick is exposed near the top of the lower deposit. Facies 2 probably represents a higher part of the early Wisconsin(?) till (*EW*) which slid down against facies 1. The dark band below the waste is the A horizon of a soil profile. Height of section is about 30 feet.

If the above interpretation is correct, then the contact between the middle Wisconsin(?) till and the till complex below is an erosional unconformity, because all surface irregularities due to the slumping have been effaced. The section does not conclusively prove the existence of more than one age of till because the deformation conceivably might have resulted from shove beneath the ice and also the moving ice may have leveled the surface of the disturbed till before the upper, undisturbed till was superimposed as ablation moraine.

PRESENCE OF BOULDER-CLAY FACIES

In some areas, the gray till is a boulder clay, quite unlike typical superglacial till. The writer prefers not to stress this as an argument against a superglacial origin, because ablation moraine on a glacier might not everywhere be subjected to the same degree of water-working and removal of fine material. However, the presence of a boulder-clay facies of the gray till weakens the "texture" argument for a superglacial origin.

There is at least one large exposure in which the gray—middle Wisconsin(?)—till shows much more water-working than usual. The exposure is in a cut along the Great Northern Railway in Stony Creek valley about 4 miles northeast of Williston (NW¼ sec. 10, T. 154 N., R. 100 W.). The stratigraphy is similar to that of the southeast face of Jones cut, discussed earlier. The sequence is shown diagrammatically in structure section No. 15, plate 7. The gray—middle Wisconsin(?)—drift, about 6 feet thick, includes sand lenses and a few gravel seams. The contact between the gray and the underlying brown—early Wisconsin(?)—drift is sharp but uneven. Locally, a channel in the surface of the lower till is occupied by sands and gravels (pl. 7, No. 15). If the gray deposit outside the channel area is till, then the conditions at the time of deposition were much more conducive to modification by glacial waters than they were at Jones cut. It is unknown whether (1) the channel deposits are contemporaneous with the gray deposit alongside, (2) whether at least part of the sand and gravels antedate the gray deposit, or (3) whether the channel deposits postdate the gray deposit.

SUMMARY OF DISCUSSION OF DIFFERENTIATION

In summary, the case for a separate identity of the gray—middle Wisconsin(?)—drift rests on the combined weight of a number of lines of evidence. The most compelling evidence is afforded by the geomorphic relations and by the presence of the intervening soil. The rest of the evidence varies from strong to weak. Each of the lines of evidence described is also a rather valid argument against the origin of the gray till by periglacial water-working of the upper part of the underlying till.

MIDDLE WISCONSIN(?) GLACIOFLUVIAL AND LACUSTRINE SEDIMENTS

The city of Williston is on a prominent terrace 20 to 25 feet above the Missouri flood plain and 40 to 45 feet above low-water river level. The western part of the terrace is underlain by middle Wisconsin(?) and early Wisconsin(?) till. The till is mantled by massive buff silts from 1 to more than 8 feet thick. The eastern half of the terrace is underlain by sands and gravel. The gravel is only moderately well sorted and many of the pebbles are subangular. In some layers the pebbles average more than 1 inch in size and there are scattered cobbles as much as 6 inches in size. The pebbles consist of glacial erratic types and brown water-rounded pebbles of local derivation. The deposit is nowhere overlain by till, hence it is later than the surrounding middle Wisconsin(?) till.

The Williston terrace continues almost uninterruptedly for about 4 miles up Little Muddy Creek valley. Beyond this point scattered remnants are present as far as the mouth of Cow Creek, about 9 miles north of Williston. No remnants of the terrace have been found farther up Little Muddy valley. Deposits of comparable, though somewhat coarser, materials extend westward up Cow Creek valley, where they are exposed in terraces as much as 40 feet above stream level (fig. 42).

Mankato outwash from the north disappears under the alluvium of Little Muddy valley before reaching Cow Creek, hence at Cow Creek it is probably at least 40 feet below the middle Wisconsin(?) terrace. The topographic relations indicate that the terrace gravel is older than the Mankato outwash. In the author's opinion, the terrace gravel was deposited when the front of the middle Wisconsin(?) ice stood just north of Cow Creek. This phase of the middle Wisconsin(?) glaciation will be considered in detail under the heading "Geologic History."

Remnants of the Williston terrace are present on the east side of Little Muddy valley and downstream along the Missouri River. Some of the terrace gravels in the Missouri valley west of Williston may also be of this age.

Middle Wisconsin(?) glaciofluvial deposits locally mantle middle Wisconsin(?) till on the uplands north of Williston. Their deposition has contributed to the smoothness of the topography. The sediments near the top of the section in Jones cut have already been described. Several hundred yards west of Jones cut, where the bluffs stand about 150 feet above the flood plain, digging revealed a minimum of 8 feet of stratified loose sand and gravel at the crest of the bluffs. The layers dip gently valleyward at precisely the angle of slope of the upland surface. The probable conditions of deposition in this precarious topographic situation are considered later. Many sand and gravel pits within the area of the middle Wisconsin(?) drift are in glaciofluvial deposits of this age.

Glaciofluvial sediments also clog many of the valleys in this area. Those along Cow Creek are considered in detail on page 91.

Middle Wisconsin(?) ice-contact deposits include an eskerlike feature and several kames on the floor of Little Muddy valley just below the mouth of Cow Creek. Only two of the kames are shown in plate 1. The eskerlike feature is several hundred yards long and consists partly of stratified sands and gravels and partly of a pebble-boulder mixture. The feature is fairly straight and could be a crevasse filling or a lateral kame moraine.

The middle Wisconsin(?) ice apparently dammed many tributaries of Little Muddy Creek, as suggested by laminated or varvelike clays and silts. Lacustrine silts may be observed below middle Wisconsin(?) till near the mouth of East Fork of Little Muddy Creek (sec. 1, T. 156 N., R. 100 W.).

AGE OF DRIFT

The gray drift is probably Tazewell or Cary in age, herein referred to as middle Wisconsin(?). The dating is primarily based on the fact that the underlying brown drift is interpreted as being early Wisconsin(?), that is, Iowan or Tazewell, whereas the still younger drift of the coteau belt is interpreted as being Mankato.

MANKATO DRIFT

DISTRIBUTION

The youngest drift is largely restricted to the northeastern portion of the area (pl. 1). The drift border is irregular, consisting of alternating lobes and embayments. The largest lobe, 30 miles long and 25 miles wide, projects southwest to Medicine Lake, Mont., and will be referred to as the Medicine Lake lobe. The development of the Medicine Lake lobe was influenced by a broad southwest-trending swale in the topography. The eastern flank of the swale culminates in the divide between Big Muddy Creek in Montana and Little Muddy Creek in North Dakota. Bull Butte, with an elevation of 2,530 feet, stands several hundred feet above the upland and is part of this divide area. The western flank of the swale rises to the foot of the Flaxville Plateaus west of Big Muddy valley.

A smaller lobe extended below Zahl, in northwestern North Dakota, and will be referred to as the Zahl lobe. The present broad valley of Little Muddy Creek probably encouraged the development of this lobe.

The border of the Mankato drift west of the Medicine Lake lobe in Montana and east and southeast of the Zahl lobe in North Dakota is not well defined, and the influence of the underlying topography on its configuration is unknown. It is possible that the Mankato drift extends south of the Missouri River in the extreme eastern part of the area. The evidence for differentiating this drift from the others and for plotting its border in the position shown in plate 1 is considered later.

TERMINOLOGY

The Mankato drift includes the moraine of the coteau border of the Missouri Plateau. This moraine has generally been referred to as the Altamont moraine. Although it may well be the continuation of the Altamont moraine of southeastern South Dakota, the correlation has not been proved. Townsend and Jenke (1951, p. 850–851) propose that the name "Max moraine" be substituted for "Altamont moraine" to avoid the implications of regional correlation.[6] Unfortunately Townsend and Jenke apply the name "Max" over a distance of about 800 miles. This subjects the name to the very criticism leveled by Townsend and Jenke against the name "Altamont moraine," namely that it presupposes correlations which have not been established. The writer will refer to this moraine within the present map area by the noncommittal terms "coteau moraine" or "moraine of the coteau belt."

PHYSICAL CHARACTERISTICS

Texture and structure.—The Mankato till is calcareous and clay rich and has scattered pebbles, cobbles, and boulders. Boulders over 3 feet in diameter are not uncommon, the larger ones generally being limestone or dolomite. Small fragments of lignite coal are common, particularly in basal portions of the deposit, and limonite concretions, many of which are cylindrical, are interspersed through the oxidized zone. In many exposures the till is vertically jointed and is mud cracked.

The till locally includes lenses, irregular patches, or unoriented blocks of sand and gravel. In places it is interbedded with glaciofluvial sediments. A large railroad cut about half a mile east of White Earth, in Mountrail County, N. Dak., exposes three separate till horizons that are separated from each other and from the Fort Union formation below by glaciofluvial sediments (fig. 25).

[6] The name "Max" is derived from a small town in the morainal belt in north-central McLean County, N. Dak.

Color and oxidation.—Thick sections of the Mankato till are not exposed in the coteau belt, because dissection has been negligible. In the area north and east of the Missouri escarpment, however, erosion marginal to deep valleys, and numerous open-pit lignite mines, reveal thick sections. A light-gray to white lime-rich zone 2 to 4 feet thick underlies the dark A horizon of the surface soil. This is the caliche horizon, a regional feature. In this zone, and for a short distance below, the pebbles of the till generally bear encrustations of calcium carbonate on their undersides. The till below the caliche zone is oxidized and is generally buff or light brown. Many weathered surfaces appear gray. Lemke and Kaye (1953) have reported oxidation locally to depths of 20 to 40 feet. This

FIGURE 25.—Mankato till and interbedded sediments in cut along Great Northern Railway about half a mile east of White Earth, northwestern Mountrail County, N. Dak. *t*, till; *g*, gravel; *si*, silt; *sd*, sand.

approaches the maximum depth of oxidation in the early Wisconsin(?) till. In thick exposures the oxidized till passes below into dark, locally bluish-gray, unaltered till.

Thickness.—The maximum thickness of Mankato drift in the coteau belt is estimated to be between 150 and 300 feet (R. C. Townsend, written communication, 1947; Lindvall and Hansen, 1947; and Lemke and Kaye, 1953). North and east of the Missouri escarpment most of the ground moraine probably ranges in thickness from a few inches to about 20 feet, but greater thicknesses are common (R. C. Townsend, written communication, 1947). Glaciofluvial sediments mantle the till in many places and contribute to the levelness of the surface. The uniformity of thickness is revealed in open-pit lignite mines. Leonard, Babcock, and Dove report (1925, p. 78) that over a distance of 5 miles a particular coal bed lies only 10 to 27 feet below the surface. They report the same coal bed 15 miles to the northwest at a depth of 30 feet. Although the thickness of the till cover is not stated, it is presumably less than 30 feet over much of the area described.

Lithology.—The lithology of the Mankato till has been determined from 18 pebble analyses (table 1, samples 34, 35, 37, 38, 63–67, 69–77). Location of these samples is shown in plate 5. The content of glacial pebbles, based on 10 of these samples, all well north of the mapped drift border, is shown in table 6.

TABLE 6.—*Glacial-pebble content of Mankato till*

Sample (table 1 and pl. 5)		Granitic rocks	Other plutonic rocks	Metamorphic foliated rocks	Limestone and dolomite
Number	Number of pebbles [1]				
63	177	10. 7	0. 6	10. 7	78. 0
67	130	20. 0	5. 4	4. 6	70. 0
70	138	10. 1	. 7	8. 0	81. 2
71	103	12. 6	. 0	9. 7	77. 7
72	138	13. 8	4. 3	4. 3	77. 6
73	113	15. 9	1. 8	6. 2	76. 2
74	141	17. 0	2. 8	13. 5	66. 7
75	104	11. 5	3. 8	4. 8	79. 8
76	112	24. 1	3. 6	6. 3	66. 0
77	107	14. 9	1. 8	2. 8	80. 4
Average		15. 1	2. 5	7. 1	75. 4

[1] After elimination of local types.

The averages indicate that 75 percent of the pebbles consist of limestone and dolomite; 15 percent, of granitic rocks; 7 percent, of metamorphic foliated rocks; and 3 percent, of plutonic types other than granite. Histograms of the Mankato till which appear in figure 18 are also based on 10 analyses, but not precisely the same 10 as listed above. The average percentages, however, are about the same. The histograms of figure 18, therefore, may be accepted as reasonably accurate portrayals of the erratic-pebble content of the Mankato till. Similarity of the histograms representing the average glacial-pebble content of the Mankato and early Wisconsin(?) tills in figure 18 is striking. Lithology, therefore, is of no use in mapping the boundary of these two drifts. It will be noted in figure 18, however, that the glacial-pebble lithology of the middle Wisconsin(?) till is markedly different from that of the others. This contrast has assisted in plotting the middle Wisconsin(?)-Mankato boundary for a distance of more than 30 miles in northern Williams County, N. Dak. These matters will be considered in greater detail later.

UNDERLYING TOPOGRAPHY AND GEOLOGY

As one approaches the coteau morainal belt from the east, the illusion is strong that the entire belt is a massive deposit of till as much as 500 feet thick (fig. 26*A*). Actually, accumulating evidence indicates that the moraine is a surface mantle of variable thickness resting on a much-dissected bedrock high. The idea is not new; these relations were suggested by T. C. Chamberlin in 1883 (p. 396). Townsend and Jenke (1951, p. 845–849) have summarized historically the

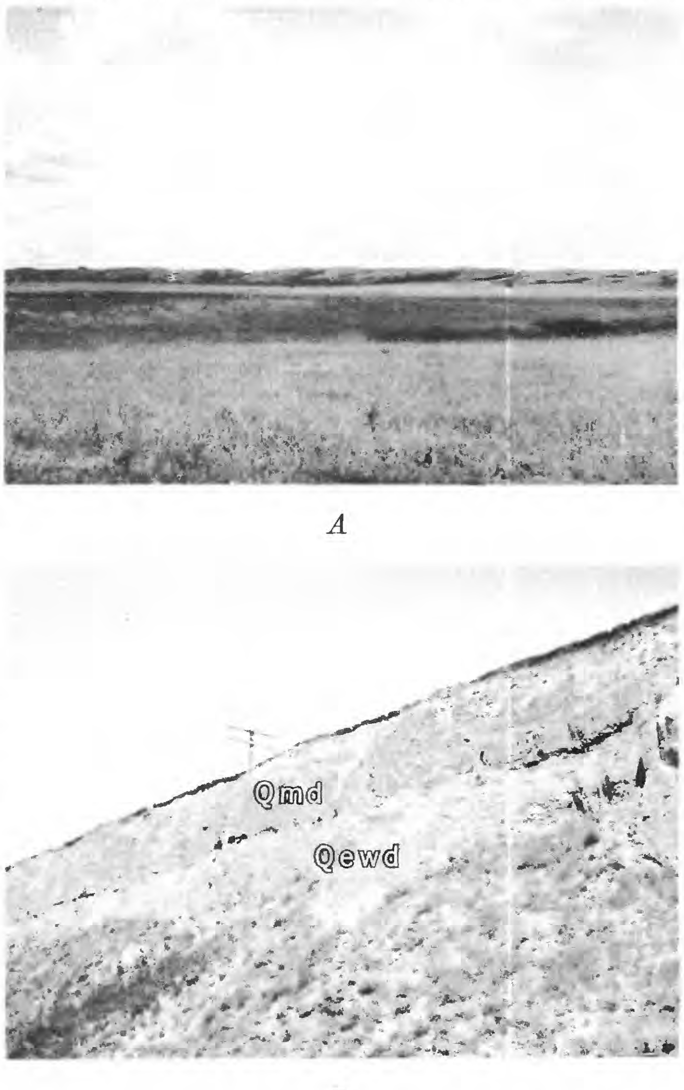


FIGURE 26.—*A*, View west toward the coteau moranial belt, southeast of Columbus, N. Dak. The entire area in the field of view is mantled by glacial deposits. Evidence indicates that the land form in the distance is not a massive accumulation of till, but rather the dissected till-mantled border of the Missouri Plateau. The scarp is the Missouri escarpment. Locally, the scarp may be of structural origin. *B*, Exposure along Great Northern Railway about 2½ miles east of Tioga, Williams County, N. Dak. The compact clayey till above is separated from the more silty till below by as much as 2 feet of sand locally including a disturbed soil(?). To the right, the lower till rises nearly to the surface. The upper till is tentatively identified as Mankato (*Qmd*); the lower, as early Wisconsin (*Qewd*).

contributions which have strengthened the above concept and there is no need to repeat the summary here. Suffice it to say that the contributions consist of discoveries of bedrock exposures at high levels in the morainal belt, as well as information on bedrock altitudes supplied by drilling in connection with the Missouri River basin development program. Reference has already been made to the swales which probably record buried valleys. One of these, marking the probable ancestral course of the Yellowstone River, is shown in figure 29. Waring and LaRocque (1949, p. 42) report that the rock floor of this channel, as revealed by drilling, is 400 to 500 feet below the surface near Crosby, N. Dak. Drilling near Froid, Mont., has revealed another smaller channel (Witkind, 1959, p. 33). Additional revelations of this sort may be expected in the future.

Only 2 or 3 exposures have been found in which the Mankato drift rests on a possibly older drift. These will be described in the section on "Differentiation."

SURFACE TOPOGRAPHY

The most pronounced and widespread morainal topography in the region occurs within the area of Mankato drift. The coteau belt itself is a morainal area in which drainage is completely disorganized. Lakes and swamps occupy many of the closed depressions. Many depressions are probably kettles; others probably indicate irregular deposition or reflect the influence of the bedrock topography below. The relief in parts of the coteau belt amounts to several hundred feet, but the bottoms of closed depressions are rarely more than 50 feet below the lowest part of the basin rims. The moraine reaches its maximum elevation of more than 2,500 feet southwest of Lignite, N. Dak., where it stands about 500 feet above the till plains to the north.

The northeastern border of the morainal belt is clearcut; it is at the edge of the Missouri escarpment. The southwestern border, however, is ragged and the raggedness is added to by outlying islands of moraine. Over large areas the morainal topography passes insensibly into the till plains beyond. Under these circumstances it is impossible to define precisely the southwestern border of the coteau morainal belt. The width of the belt, including the more subdued morainal topography of the Medicine Lake lobe, ranges from 15 to 50 miles.

Over most of the coteau belt the topography is coarse textured (fig. 27). Closed depressions more than one-eighth of a mile long average about 7 per square mile. Many are over 1 mile long, but these generally occupy outwash channels or swales along buried valleys. In the areas of coarse morainal topography, the knobs between depressions are broad and generally occupy a larger area than the depressions.

In several places, the topography is fine textured (fig. 28). The depressions are small, the greater number being less than 200 feet across. There may be between 200 and 300 per square mile. The intervening divides are generally low and are often little more than narrow ridges. Locally, however, the ridges are broad as compared with the depressions and are then flat topped. The topography resembles a biscuit board, but lacks the orderliness.

The three largest areas of fine-textured morainal topography are in northwestern North Dakota. The first is southwest of Crosby, in Divide County; the second is south of Alamo, in northern Williams County; and the third is in southwestern Burke County.

The origin of the areas of fine-textured topography is unknown. A few shallow road cuts through the ridges reveal till, but whether this is universal is unknown. The ridges themselves rise to approximately accordant summits, and the broader ones are flat topped. Except for the till exposures, the topography might be interpreted as pitted outwash. Perhaps the till is ablation moraine, not excessively water-worked, which has been lowered to the ground during wasting of areas of stagnant ice along the margin of the Mankato ice sheet.

The coteau belt includes a number of topographic swales, a few of which cross the entire belt. The larger of these are shown in plate 1. Most of the swales slope southwestward and generally contain outwash in their lower courses. Others contain lakes, some of which are linear and are as much as several miles long. Most of the swales become less pronounced toward their "heads." The largest extends 25 to 30 miles, between Crosby and Zahl in northwestern North Dakota (fig. 29). It is 15 to 20 miles wide and 200 to 300 feet deep. For several miles northeast of Zahl its floor is covered by outwash; beyond this point its floor is hummocky and underlain by till. Locally, the till forms ridges extending across the swale; these are probably recessional features. The swale probably represents the ancestral course of the Yellowstone River before the series of diversions which created the Missouri River. This matter is considered in detail under the heading "Geologic History."

North and east of the coteau belt and the Missouri Plateau, within the area of the ground moraine, the surface of the Mankato drift is remarkably level, although interrupted by glacial channels and by youthful valleys near major streams. The till mantles a

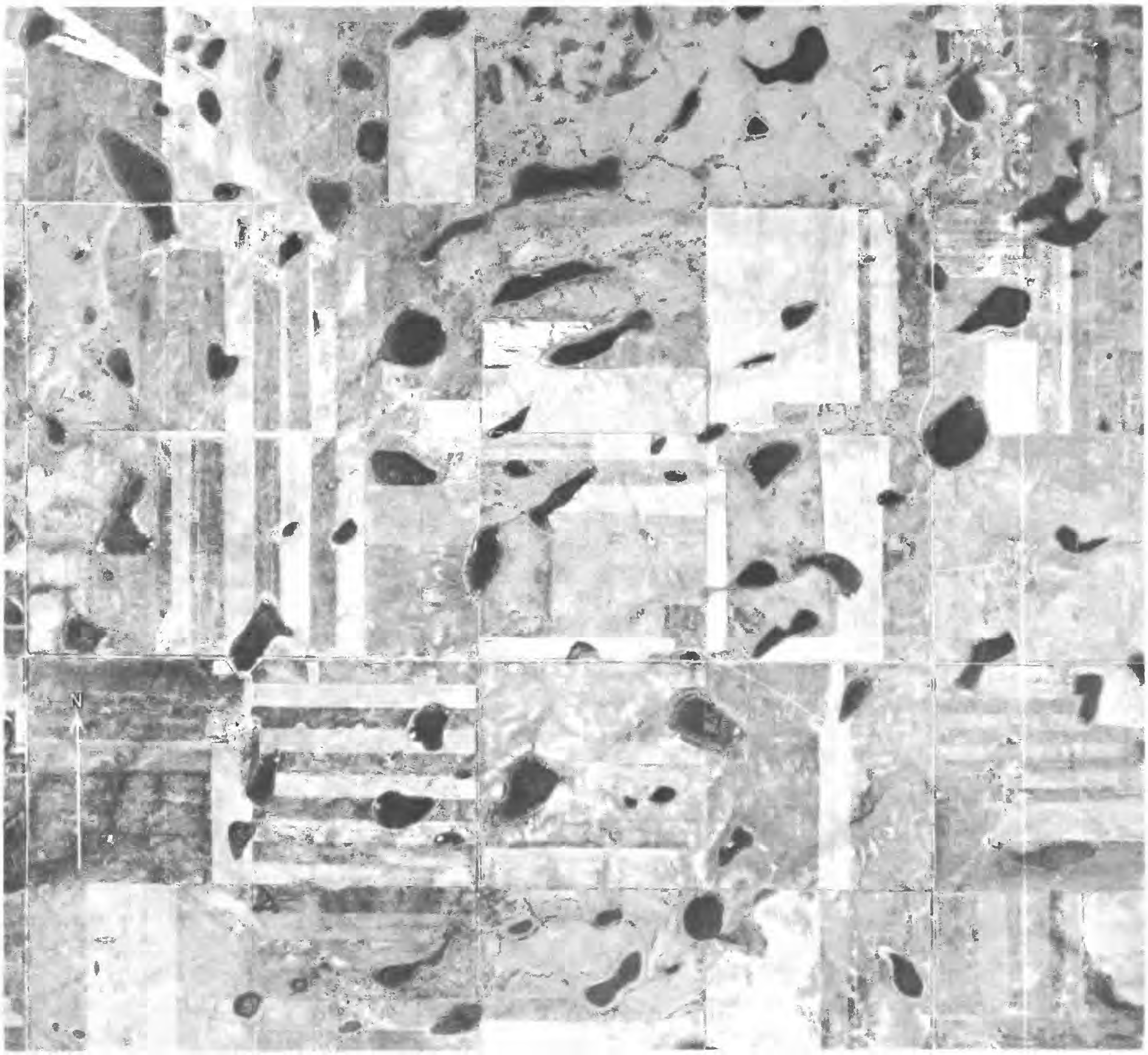


FIGURE 27.—Coarse-textured Mankato morainal topography characteristic of greater part of the coteau morainal belt in northwestern North Dakota and northeastern Montana. The area is about 4 miles northeast of Alamo in north-central Williams County, N. Dak. Roads are along section lines.

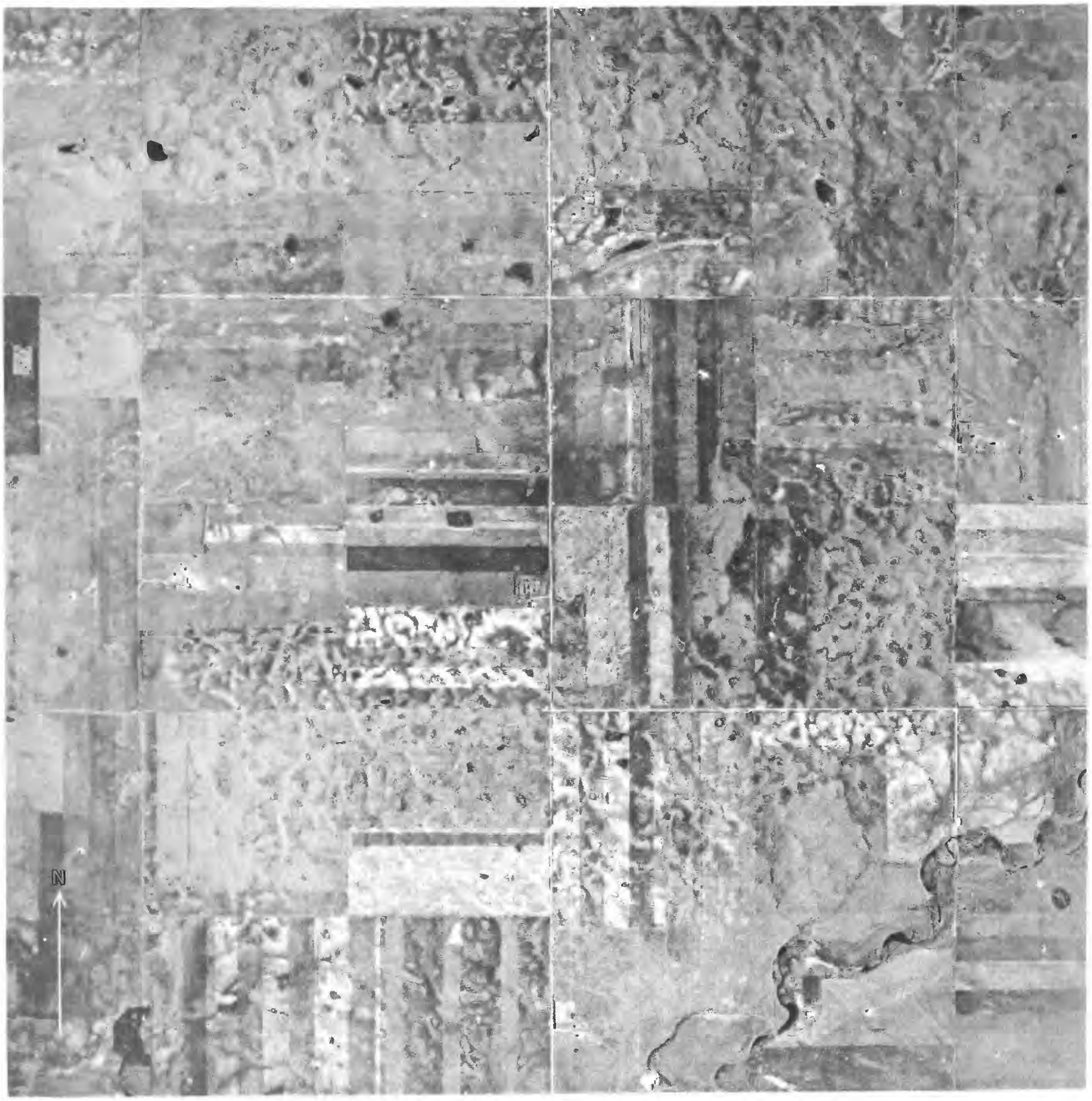


FIGURE 28.—Fine-textured Mankato morainal topography in the coteau morainal belt about 2 miles south of Alamo, north-central Williams County, N. Dak. The southern border of the Mankato drift is drawn along Little Muddy Creek in the southeast corner of the photograph. Roads are along section lines.

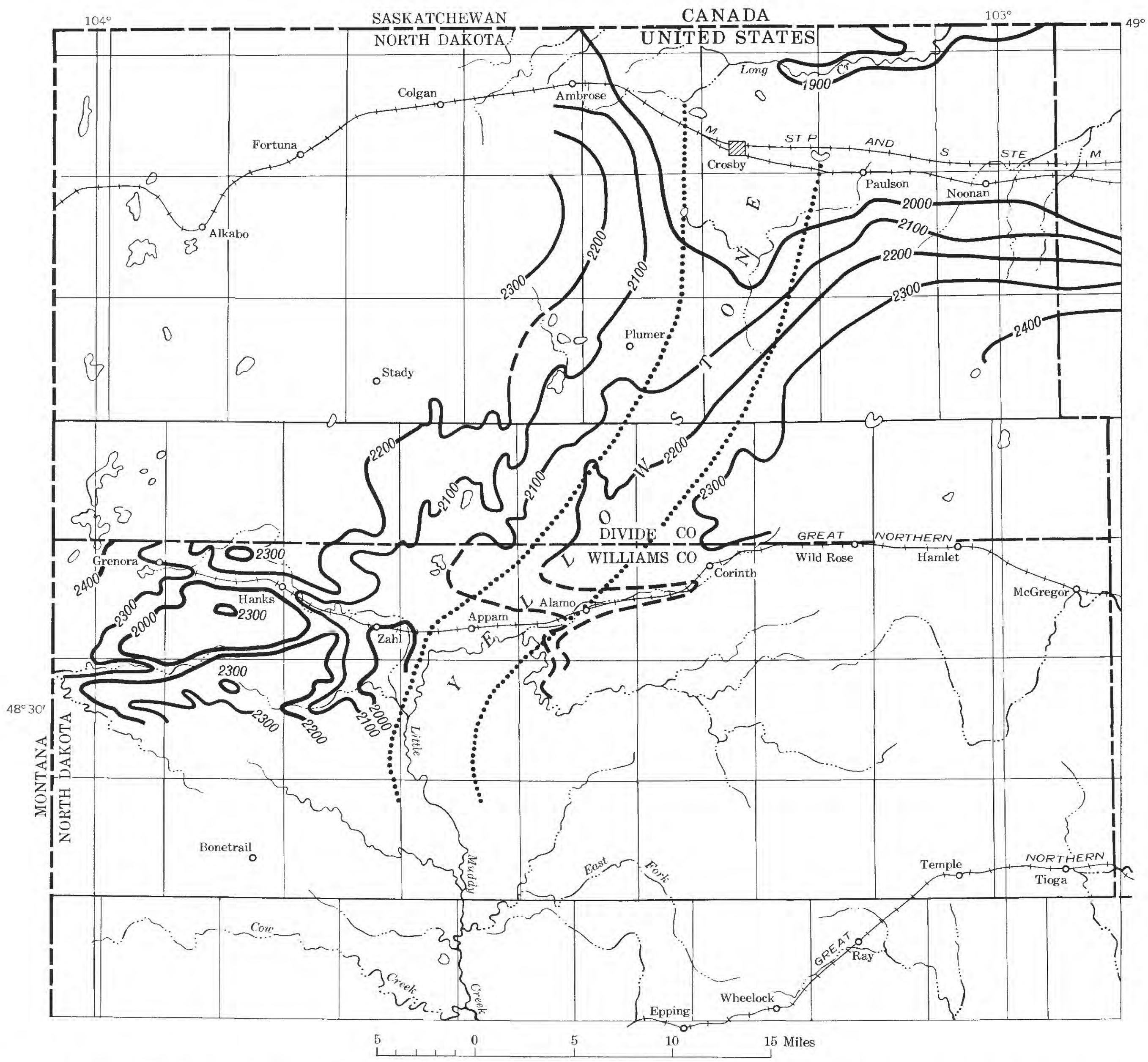


Figure 29.—Generalized contour map of swale through the coteau morainal belt, probably marking buried ancestral course of the Yellowstone River. The dotted lines indicate the path suggested by Alden (1932, pl. 1).

level surface which probably accounts in part for its own levelness. Numerous initial irregularities, however, have been effaced by deposition of glacial outwash.

In contrast with the other drifts, the area of the Mankato drift abounds in eskers, kames, and glacial channels. Some of the largest channels, however, those formed by the ice-marginal Missouri River, are in the area of the early Wisconsin(?) drift.

Nearly all the outwash-occupied channels back of the Mankato drift border contain kettles of various shapes and sizes. These are particularly profuse in the channel which extends southwest from Westby to Medicine Lake in Montana. They are also common in the outwash area west of Grenora, N. Dak., in the lower part of the channel which extends northeast to Stady, and in the broad outwash area near Zahl. Additional kettles indent the outwash of the channels in the far eastern part of the area. The kettles range from a few tens of feet to more than 2 miles in length and are commonly steep sided. Presumably the outwash was deposited around and over blocks of ice isolated from the main ice mass by differential wasting. Locally, as in the area east of Zahl, the outwash was trapped in areas between ice blocks and was deposited to levels above that of the surrounding plain.

Mankato outwash forms valley trains in Big Muddy, Little Muddy, and White Earth valleys, but it is scarce or absent in the Poplar valley. The latter may have had only limited contact with the Mankato ice.

DIFFERENTIATION

The Mankato drift has been differentiated from the others on the basis of a number of lines of suggestive evidence which are treated in the following order: (1) the lithologic contrast with the adjoining middle Wisconsin(?) till, (2) stratigraphic relations, (3) geomorphic relations, (4) topographic relations, and (5) the appearance of youth as indicated by the concentration of ice-contact features, valley trains, and glacial channels.

LITHOLOGIC CONTRAST WITH MIDDLE WISCONSIN(?) DRIFT

The middle Wisconsin(?) drift is everywhere surrounded by drift with a much higher content of limestone and dolomite. The drift with the high limestone-dolomite content which directly underlies the middle Wisconsin(?) drift is interpreted as being early Wisconsin(?). It is exposed both east and west of the area of middle Wisconsin(?) drift and extends south to the glacial limit. It is unlikely that the northern drift, herein mapped as Mankato, is an extension of the early Wisconsin(?) drift, because there would then be no rational explanation of the small "island" of drift which is relatively low in limestone and dolomite—middle Wisconsin(?)—in the midst of a sea of drift which is high in limestone and dolomite. Nor, for reasons presented earlier, is it likely that the "lithologic island" is a border facies of the northern—Mankato—drift. The drift to the north is probably younger than the middle Wisconsin(?) drift and covers the latter everywhere except in its present small area of exposure.

No exposures are known in which the Mankato drift rests on the middle Wisconsin(?) drift. This is not necessarily an argument against the reality of the middle Wisconsin(?) drift; exposures of multiple drifts are relatively rare throughout the glaciated regions. It is interesting to note that no exposures involving undoubted multiple drifts have been found by other workers in the coteau belt and in the region of the ground moraine to the north and east. Yet the writer is unaware of any prevalent opinion to the effect that only the Mankato ice invaded this region. An exposure will be described in the next section in which the Mankato till may actually overlie the early Wisconsin(?) till.

SUGGESTIVE STRATIGRAPHIC RELATIONS

RAILROAD CUT EAST OF TIOGA

Two layers of till, Mankato and possibly early Wisconsin(?), are exposed in a long cut along the Great Northern Railway about 2½ miles east of Tioga in easternmost Williams County, N. Dak. (pl. 7, loc. 22, and fig. 26*B*). The evidence favoring two distinct tills is as follows: (1) the presence of a probable soil profile at the top of the lower till; (2) intertill channeling; and (3) the physical differences in the two tills, the upper being more clayey and compact.

The exposure is about 800 feet long and is as much as 50 feet high. A sharp contact, 30 to 35 feet above track level, separates the till layers. The upper till, which is brown, compact, and clayey, is divided by vertical joints and displays small, hackly shrinkage polygons. It contains numerous small pebbles as well as scattered boulders of granite, gneiss, and basic igneous rocks as much as 2 feet in diameter.

The sharp break at the base of the upper till is due largely to an underlying layer of pebbly coarse sand as much as 2 feet thick. About 300 feet from the west end of the exposure the sand includes contorted seams of dark material which gives no acid reaction (pl. 7, loc. 22). All other materials in the exposure are calcareous. This dark, leached material probably represents the A horizon of an old soil disturbed by overriding ice.

About 50 feet east of the soil(?) locality, the intertill sand rests on about 3 feet of stratified gravels which occupy a 10-foot-wide channel in the lower till. The gravels are locally heavily oxidized. The channeling may be further evidence of a time break, although it can be explained otherwise.

The lower till displays a horizontal parting for several feet below the sand. Locally, paper-thin seams of dark clay follow the partings. At the west end of the cut the parting passes laterally into thin sedimentary bedding. The zone of parting and bedding apparently weathers more easily than the rest of the till below, for the latter protrudes from the slope. Even where the intertill sand is missing, the sharp contact of the two tills persists because of the more rapid weathering of this zone. The color of the layered zone is variegated buff and gray. Below the layered zone, the till is gray to buff and compact, but perhaps a little more silty than the till above. It is also less pebbly, has fewer large boulders, and generally is not so well jointed. At both ends of the cut, the lower till rests on glaciofluvial silts with layers of fine gravel. Similar sediments are exposed below till in the next railroad cut about one-quarter of a mile east. This exposure, however, adds nothing to the relations described above.

The probable soil, the intertill channeling, and the physical contrasts in the two tills suggest that the tills are of different age. If so, the similar high percentage of limestone and dolomite would indicate that the younger is Mankato and the older is early Wisconsin(?). The similarity in pebble composition is shown in the following table.

Glacial-pebble content of upper and lower tills in railroad cut 2½ miles east of Tioga, N. Dak.

[The complete analyses appear as Nos. 35 (upper till) and 36 (lower till) of table 1]

	Granitic rocks	Other plutonic rocks	Metamorphic foliated rocks	Limestone and dolomite
Upper till	17. 0	4. 7	4. 7	73. 6
Lower till	20. 0	2. 2	2. 2	75. 6

Simpson[7] has described two tills, Iowan(?) and Mankato, in the Garrison quadrangle, 50 to 60 miles southeast of the present map area. He remarks that the tills are alike lithologically, but that the fine fraction of each behaves differently when immersed in a 0.01*N* solution of sodium oxalate. The fine fraction of the Iowan(?) till went readily into suspension and remained so for weeks. The fine fraction of the Mankato till, however, flocculated and settled immediately. It would be interesting to see this test applied to the Tioga exposure and other exposures in the present problem area. The test may offer a possible means of differentiation and correlation of similar-appearing tills.

[7] Simpson, H. E., Jr., 1942, The Pleistocene geology of Garrison quadrangle, North Dakota: Ill. Univ., unpub. thesis, 40 p., 1 map.

WHITE EARTH RIVER VALLEY EXPOSURE

The large exposure along U.S. Highway 2 at the edge of the upland on the west side of White Earth River valley has already been described. A boulder band separates two distinct till facies. Neither the upper nor the lower till resembles the middle Wisconsin(?) till. A short distance from the exposure, the upper till displays morainal topography comparable in relief with that of the Mankato moraine of the coteau belt. If two distinct tills are present, they are presumably the Mankato and the early Wisconsin(?).

GEOMORPHIC RELATIONS

The geomorphic relations in Little Muddy Creek valley, North Dakota, have already been discussed in connection with the problem of the relative ages of the early Wisconsin(?) and middle Wisconsin(?) drifts. We may now consider the relation of the Mankato deposits to those previously described.

The middle Wisconsin(?) ice displaced Little Muddy Creek laterally at two places in its valley. The diversions paths (fig. 20), now occupied by Mankato outwash, were probably eroded in part in the middle Wisconsin(?)-Mankato interval. Two pebble analyses of the outwash reveal about 62 percent of limestone and dolomite, about 27 percent of granitic rocks, about 4 percent of other plutonic types, and about 7 percent of metamorphic foliated rocks. The analyses are similar to those of the Mankato till of the coteau morainal belt into which the outwash can be traced without a break.

TOPOGRAPHY

The pronounced morainal topography of the coteau belt contrasts strongly with the largely nonmorainic well-drained topography of the gray—middle Wisconsin(?)—drift. The contrast is similar to that between the Mankato and early Wisconsin(?) drifts and, by analogy, may also reflect differences in age.

CONCENTRATION OF ICE-CONTACT FEATURES AND VALLEY TRAINS

By far the greatest concentration of ice-contact features, such as kames, eskers, and crevasse fillings, as well as of undissected valley trains, lies within the area of disorganized drainage, that is, north of the line of topographic unconformity which forms the approximate boundary of the Mankato drift. These features are most numerous in the swales occupied by the Medicine Lake and Little Muddy ice lobes (pl. 1). These glacial features are much less common outside the coteau belt but may be found as far as the glacial limit, 100 miles to the south. The writer is inclined to attribute their scarcity south of the coteau belt to subsequent erosion, perhaps during the stage when the grading of the till surface took place. A few peculiar linear concentrations of boulders may represent the total remains of some of the elongate deposits. The few eskers which are present are located in areas that are not now being eroded. The esker near Lambert, Mont., furthermore, is composed of coarse and permeable materials, not readily susceptible to surface erosion. This is also true of the small pitted outwash plain west of the esker. The materials of this plain are so porous and pervious that none of the innumerable small depressions are occupied by lakes.

SUMMARY OF DISCUSSION OF DIFFERENTIATION

The differentiation of the Mankato from the older drifts rests on the collective weight of five lines of evidence. These are (1) contrast in lithology with the middle Wisconsin(?) drift; (2) presence of an underlying till and a probable intertill soil in the exposure east of Tioga, N. Dak.; (3) evidence of a middle Wisconsin(?)-Mankato erosion interval; (4) topographic unconformity at the mapped Mankato boundary; and (5) the appearance of glacial youth in large parts of the mapped Mankato area as indicated by the concentration of ice-contact features, valley trains, and glacial channels.

LOCATION OF DRIFT BORDER

If the existence of the middle Wisconsin(?) drift is accepted, then the boundary between it and the Mankato drift can be fairly closely defined. This is possible because of the physical differences in the two tills, particularly the differences in lithology. However, there is reason to believe that, in some places along its border, the Mankato ice left too meager a sprinkling of deposits to be distinguished from the middle Wisconsin(?) drifts. For example, the boundary between the Mankato and middle Wisconsin(?) drifts along Little Muddy Creek valley just below Zahl, N. Dak., has been plotted (pl. 1) not at the lithologic boundary but at a short distance to the south. The reason for this is as follows (Howard, Gott, and Lindvall, 1946): Middle Wisconsin(?) drift, with well-developed morainal topography, occupies the area within the right-angle bend of the Little Muddy Creek below Appam, N. Dak. This morainal area is hemmed in on the north and west by the Mankato valley train of Little Muddy Creek. The creek flows 20 to 25 feet below the surface of the outwash. The surface of the outwash is higher than large parts of the middle Wisconsin(?) moraine with its many closed depressions. The closed depressions preclude the possibility that the moraine has been exhumed from beneath the outwash. It seems reasonable to suppose that burial by outwash was forestalled by a protective mantle of ice.

The contrast between the pronounced morainal topography of large parts of the Mankato drift and the dominantly nonmorainic well-drained topography of the middle Wisconsin(?) drift has already been referred to. The contrast in topography extends east and west of the area of the middle Wisconsin(?) drift. If the contrast in the latter area is a function, at least in part, of a difference in age, then it seems reasonable to suppose that it may have a similar significance in the adjoining areas. The border of the Mankato drift in these adjoining areas, therefore, has been drawn at the boundary between the areas of disintegrated and integrated drainage. Except where this line of demarcation is marked by a deep glacial channel, as along both sides of the Medicine Lake lobe, it is probably only a rough approximation. This is due, first, to the lack of lithologic contrast between the Mankato and the adjoining early Wisconsin(?) drift, and second, to the fact that over large areas the morainal topography of the coteau belt flattens out very gradually toward the till plains to the southwest.

The writer has tentatively mapped as Mankato the drift of the channel area between Scobey, Mont., and the Canadian boundary. The evidence which suggests that the drift is Mankato consists of a network of fairly fresh looking channels and the presence of a fairly pronounced moraine on the small Flaxville plateau west of the channel network (pl. 1). If the Mankato ice was thick enough to reach the top of the plateau, however, it seems strange that it did not extend more widely over the lowlands. Perhaps the plateau moraine is pre-Mankato and the Mankato ice was confined to the lowlands. One exposure of probable outwash has been observed downstream in the Poplar River valley above the level of the alluvium. This exposure, plus the presence of the dune field leeward of the lower stretch of valley, suggests that outwash may underlie the valley alluvium. There is no direct evidence, however, that the outwash or the derived dune sands are Mankato in age.

The Medicine Lake lobe is defined on the southeast by Cottonwood channel and on the northwest by Coalridge channel, west of Coalridge, Mont. The Cottonwood channel is about 125 feet deep and about an eighth of a mile wide. Like the Grenora-Zahl channel to the north, it crosses the divide between the ancestral valleys of the Missouri and Yellowstone Rivers. The Coalridge channel is about a quarter of a mile wide (exaggerated in pl. 1), except where constricted by till. The channel trends southwest athwart the southeast-sloping flank of the divide between the Medicine Lake lowland on the east and the valley of Big Muddy Creek on the west. Although the channel is the boundary between well-integrated drainage to the west and disorganized drainage to the east, it seems likely that the ice may, for a time, have extended a few miles farther west. This is suggested by the presence of small channels across the divide to the west. The Cottonwood, Coalridge, and neighboring channels have been described in detail by Witkind (1959).

The Mankato border east of the area of the middle Wisconsin(?) drift has been drawn to include the pronounced morainal areas in the vicinity of Temple, N. Dak., and on both sides of the White Earth River valley south to the Missouri River. The boundary thus includes the railroad cut east of Tioga, N. Dak., described earlier, in which two(?) tills, rich in limestone and dolomite and separated by a probable soil, are exposed (fig. 26*B*). It was suggested that the upper till may be Mankato; the lower, early Wisconsin(?).

The Mankato boundary has been drawn to exclude the hairpin-shaped morainal area on the North Dakota-Montana boundary (pl. 1). This moraine is assigned to the early Wisconsin(?) drift primarily because of its location at the crest of the divide between the Little and Big Muddy Creek drainage basins. If the moraine had been deposited by the Mankato ice, the ice would have submerged both drainage basins and would almost certainly have extended at least as far south as the Missouri. Evidence presented earlier indicates that the Mankato ice never had so great an extent or lay so deep in these drainage basins. It might also be pointed out that much of the drainage of the hairpin morainal area is already integrated, and a region of integrated drainage separates the morainal area from the coteau belt to the north.

The probable conditions under which the hairpin moraine developed will be considered in the section dealing with the waning of the early Wisconsin(?) ice.

TERRACE SILTS—OLDER ALLUVIUM AND SLOPE WASH

Until now it has been tacitly assumed that the flood-plain alluvium is younger than the terrace silts. The possibility was considered, however, that the flood-plain alluvium is simply the basal, older part of a stratigraphic section of which the terrace silts are younger, higher members. According to this interpretation, the "flood plain" is actually a degradational surface cut in a single thick alluvial fill. However, the relative abundance of artifacts, principally arrowheads, in the flood-plain alluvium as compared with the terrace silts, and the reported presence of buffalo and Indian skeletons clearly indicate a more recent age for the flood-plain alluvium than for the terrace deposits bearing the *Bison antiquus*(?). It seems reasonable to conclude that the terrace silts were trenched below present stream level and that the trench was subsequently occupied by the alluvium of the present flood plain. It is interesting to note that a structure section across the Missouri River valley at Nashua, Mont., west of the map area, prepared by geologists of the U. S. Bureau of Reclamation on the basis of drill records, supports the stratigraphic relations suggested above. The silts at Nashua, however, are referred to as "high alluvium," and those at Wiota, a few miles to the southeast, as "lake beds."

DISTRIBUTION AND COMPOSITION

Patches of the older and higher alluvium are scattered along many of the valleys of the region (fig. 30 and 31). The alluvium is mantled by, and probably interfingers with, thick deposits of wash from the side slopes. In the Missouri and Yellowstone River valleys, the alluvium and slope wash contribute to the composition of most of the terraces mapped as composite terrace fill (Qt) (pls. 1 and 6). In the Little Missouri River valley the terrace alluvium has not been mapped separately from the recent alluvium. For convenience, the complex of alluvium and slope wash will be referred to as the terrace silts, although layers of sand and gravel are included. The pebbles of the gravel layers are generally smaller than pea size, but 8-inch cobbles are present close to the bluffs.

The terrace remnants vary greatly in size and shape. The remnant on which Trenton, N. Dak., is located is about 12 miles long and 2 miles wide (fig. 30). Several other remnants are nearly as large.

The older alluvium and slope wash which constitute the terrace silts are finely laminated at many localities. Some of the silt layers contain large numbers of fragile gastropod shells. Most of the shells belong to the family Lymnaeidae. A few representatives of the Planorbidae and Pupillidae also appear. The first two groups are fresh-water forms. The Pupillidae are terrestrial forms and may have been washed into the silts or buried during floods. Vertebrate bones are not uncommon but have been of no help in determining the precise age of the deposits. Teeth of an ancient bison have been more helpful and will be discussed later.

Several exposures of the terrace silts show fossil soils, indicating that deposition was discontinuous. At least two such fossil soils are exposed in a pit on the south side of the Missouri River valley about 1 mile west of Nohly, Mont., just west of the North Dakota line. The section is shown diagrammatically as section 7 in plate 7.

The silts clog every low-gradient valley outside the glacial limit and form the most abundant constituent of valley fills in areas lightly veneered with drift. Except in certain valleys, however, silt is not an important constituent of valley fills in areas thickly mantled by drift. Evidence will be presented in the section on "Geologic History" that indicates that at least part of the silts in and south of the Missouri River valley are penecontemporaneous with Mankato outwash to the north. The textural contrast would probably prevail, however, even if the silts were in part post-Mankato. In the unglaciated areas, or areas only lightly veneered with till, the clays and silts of the Tertiary deposits are the principal materials available for erosion and deposition, whereas in areas thickly mantled by drift a far greater textural variety of materials is available. Thus, the silt terraces of the southern areas may be represented in the north by

FIGURE 30.—Silts forming terrace along the Missouri River. View is east toward Trenton, N. Dak. (in background). Locally, 55 feet of silts is exposed in the terrace scarp.

FIGURE 31.—Terrace scarp 10 miles long on north side of the Missouri River valley. View is east from about 1 mile west of the White Earth River. In places the terrace scarp exposes silts from top to bottom; elsewhere the silts cover till and bedrock.

terraces in which sand and gravel are important constituents.

The older silts of the Missouri River valley are, in many places, banked against and mantle glacial sands and gravels. This suggests a separate and distinct episode of deposition. The possibility has already been suggested, however, that these sands and gravels are of middle Wisconsin(?) rather than Mankato age.

The Missouri flood plain at the close of deposition of the terrace silts stood only 10 to 15 feet above the present flood plain, as indicated by the height of the lowest, outermost, and flattest part of the terrace surface. The terrace surface, however, slopes riverward so that the height of the frontal scarp may stand 80 to 100 feet above the present flood plain where appreciable lateral erosion has taken place (fig. 31). The presence of terrace remnants 80 to 100 feet above the present flood plain, therefore, does not mean that the Missouri River valley was ever completely filled to that height.

STRATIGRAPHIC AND GEOMORPHIC RELATIONS

Most of the terrace remnants shown in plate 6 are underlain in whole or part by the silts (all sections except 2, 4, 14, 16, 18, 19, 25, 28, and 30). It is unnecessary to describe these sections in detail; the following summary of stratigraphic and geomorphic relations should suffice:

At localities 6, 8, 10, 12, 13, and 26 (pl. 6) the entire scarp consists of silt and associated fine gravels. Locally the silt is 70 to 80 feet thick. At locality 12 several feet of delicately laminated silts probably indicate deposition in stagnant or near-stagnant water.

At localities 11, 21, and 23 the silts completely bury a bedrock terrace that had previously been dissected to below present flood-plain level. The surface of the bedrock terrace rises pedimentlike toward the bluffs, and in places emerges from below the silts. In its flat distal part, it is almost certainly the Crane terrace.

At localities 1, 25, and 26 the silt veneers or buries a till terrace. In these and other sections the till is almost certainly early Wisconsin(?) in age.

At localities 5, 31, and 32 the silts rest on the nonglacial Flaxville-type Crane Creek gravel which probably mantles a bedrock terrace, as it does for many miles between Poplar and Wolf Point in Montana.

At localities 7, 9, and possibly 24, the silts rest on glacial sands and (or) gravels. In the area of the middle Wisconsin(?) drift at least, these gravels are probably middle Wisconsin(?) in age. In the eastern part of the area, coarser gravels are probably Mankato in age. Although early Wisconsin(?) glaciofluvial sediments are exposed, they are not directly overlain by the silts.

The above relations indicate the following:

1. Prior to silt deposition, the principal valleys of the region contained remnants of (1) the pre-early Wisconsin(?) gravel-veneered Crane terrace, (2) the early Wisconsin(?) till terrace, and (3) middle Wisconsin(?) and possibly Mankato gravel terraces.

2. Silt deposition probably began when the floors of the principal valleys were lower than they are now, as indicated by the presence of the silts down to present river level.

3. A large part of the silty material was contributed by wash from the valley sides and by side-stream deposition. This is suggested by the increasing thickness of the silts toward the valley sides and by the dip of the silt layers toward the center of the valley. Side-stream deposition was probably induced by aggradation in the major valleys. Locally, as in sections 13 and 20 (pl. 6), the terrace surface seems to truncate the edges of the silt layers, implying subsequent regrading of the surface.

4. The silts were locally deposited in quiet, but not necessarily standing, water. This is indicated by the presence of delicately laminated layers in some exposures. The restricted vertical and horizontal distribution of the lamination, as well as the presence of sand and gravel layers, suggests that quiet-water conditions were not prevalent throughout the region. The abundant fragile fresh-water and terrestrial gastropod shells in the silts require no different conditions for preservation than those that prevail today, because similar shells are present in the modern flood-plain silts.

AGE

Available evidence suggests that the terrace silts are Mankato in age. The best evidence is a set of fossil teeth found on the south side of the Missouri River valley near the southern abutment of Wolf Point Bridge (pl. 6, sec. 1). The exposure is along a dirt road about 200 feet east of the highway. The terrace silts cover an irregular till topography; the till in turn mantles a dissected rock bench (not shown in pl. 6, sec. 1).

The teeth have been tentatively identified by C. Bertrand Schultz, director, Nebraska State Museum, University of Nebraska, as being those of an extinct bison. He found that they compared in size and form with various examples of those of *Bison antiquus* from the Great Plains, with which Folsom dart points are associated. According to Schultz, *Bison antiquus* remains are frequently found in the terrace-2A fill of Nebraska (Schultz, Lueninghoener, and Frankforter, 1951, p. 8) which, according to carbon-14 tests, dates back 10,000 to 11,000 years.

Inasmuch as the identification of the fossil teeth as *Bison antiquus* is not positive, correlation of the terrace silts of this area with the terrace-2A fill of Nebraska may not be justified at the present time. It does seem safe, however, to suggest that the terrace silts are more closely related in time to the Mankato glaciation than to an earlier one.

A fragment of an artifact, probably part of an arrowhead, was found in the older alluvium about 100 feet from the fossil-teeth locality at a depth of about 12 feet below the terrace surface. Unfortunately, the fragment is too small to permit identification of the culture represented.

As for the age of the terrace-2A fill with respect to the Mankato climax, Schultz, Lueninghoener, and Frankforter (1951, p. 34–37) believe that at least the lower part of the fill antedates the Mankato climax in southeastern South Dakota by possibly several thousand years. Among the evidence they cite is the absence of the terrace-2A fill within the area of the Mankato till in southeastern South Dakota. They argue that if the fill were later than the Mankato climax, it ought to be preserved in that area, as is the post-Mankato terrace-1. Furthermore, according to these authors, the terrace-2 fill as a whole was eroded prior to deposition of a late Mankato fill.

Nebraska and southeastern South Dakota are far from the present area of study so that events and dates need not correlate exactly. For example, if the Mankato ice blocked the Missouri River at Sanish, N. Dak. (pl. 1), as postulated on page 93, deposition may have occurred upstream, possibly in part under lacustrine conditions, but there need not have been comparable deposition downstream. In view of the several episodes of glaciation in this area and the probable segmentation of the valley by ice lobes during advance of the ice and by ice lobes and stagnant masses of ice during recession, the terrace sequence may not be amenable to regional correlation. Unfortunately, the scope of the present study prevented detailed consideration of the exact age of the silts relative to the Mankato climax. However, the absence of till on the silt terrace in the area of Mankato drift suggests that at least the upper silts are later than the Mankato climax.

In summary, the terrace silts are largely Mankato in age, but deposition probably continued beyond the Mankato climax.

EOLIAN DEPOSITS

SAND

The several large sand-dune areas in eastern Montana north of the Missouri River (pl. 1) are composed of loose sand presumably winnowed from outwash deposits immediately to the northwest. The sand is fine and contains much quartz, feldspar, mica, and hornblende. The outwash bodies which supplied the large dune areas near Poplar and Froid are now presumably buried under valley alluvium. Much of the source area for the dunes southeast of Medicine Lake is submerged under lake waters. Two other small dune areas, one about 6 miles east of the Medicine Lake dune field and the other along Long Creek northeast of Crosby, N. Dak., were supplied from outwash in small glacial channels.

The dunes in some sections of the fields are largely bare and appear to have been recently active. In other sections the dunes are well anchored by vegetation. Sand forms a thin veneer over the areas bordering the dune fields, particularly to the southeast. Within a mile or two of the dune fields the sand is replaced by loess.

LOESS

Loess is not restricted to areas leeward of the sand regions, but it mantles the entire area of study, hilltop and lowland alike. It is missing only where active dissection in badland belts has removed all the surface mantle. Collier (1918, p. 413) believed that the valleys of the region were occupied by loess as much as 30 feet thick. The writer has nowhere found undisturbed loesslike material more than 5 feet thick, but thicker accumulations due to mass wasting occur locally along valley sides. Part of the stratified silt which chokes the valleys of the region may represent loess which was subsequently reworked by streams.

Numerous exposures reveal loess of several ages, separated by soil profiles. Many of the exposures are recorded in plate 7. Others have been omitted because of their proximity to practically identical exposures already recorded. Some of the sections in plate 7 have been described in the text in connection with the glacial drifts and the terrace silts. The others will not be described in detail. The following discussion will concern itself with the possible significance of the loesses and soils in terms of chronology and regional history. In view of the possibility that future detailed work may reveal additional loesses and soils, the present discussion is in the nature of a progress report.

Loess deposits separate the several Wisconsin drifts in some of the States to the southeast. Unfortunately, undoubted loess has not been found between tills in the few exposures in this area in which multiple tills are revealed. There is thus no direct evidence that any of the loesses are pre-Mankato in age. However, it is interesting to note that multiple loesses are restricted to the areas of pre-Mankato drifts. Thus, three loesses, each bearing a mature soil profile, overlie the early Wisconsin(?) drift at localities 1, 4, and 19 (pl. 7). Two loesses overlie the middle Wisconsin(?) drift at locality 9 and possibly at locality 13. At locality 13, however, the mantling deposits lack the fluffy, structureless character of the loess deposits elsewhere and may have a different origin. Only one loess overlies the Mankato drift at locality 12. The above observations suggest that the two older loesses are pre-Mankato in age; one, possibly of early Wisconsin(?) age, the other of middle Wisconsin(?) age. However, the restriction of the multiple loesses to the older drifts is based on very few observations. A far greater number is needed before the suggested relations can be accepted as generalities. As a working hypothesis, however, the writer suggests the following:

1. The lowest loess is early Wisconsin(?) in age. The loess was deposited at least in part after the glacial climax, because it mantles the early Wisconsin(?) till. Loess deposition was followed by a period of soil formation.

2. The middle loess is middle Wisconsin(?) in age and it, too, was deposited in part after the glacial climax. A period of soil formation followed.

3. The upper loess is Mankato in age. The presence of soil on underlying sands and gravels of Mankato age (pl. 7, sec. 12) suggests that, in some areas, loess deposition may not have begun immediately. The Mankato loess is widespread and locally reaches a thickness of 5 feet. It is leached to variable depth in all exposures tested. Near Froid, Mont., where the loess rests on a soil profile developed on the early

Wisconsin(?) till, the entire 4 feet of loess is leached (pl. 7, sec. 5, and fig. 32*A*). On the upland between Stony Creek and Little Muddy Creek northeast of Williston, N. Dak., the loess is leached to a depth of 2 to 2½ feet. This loess is not to be confused with that of historic dust storms, which show only an incipient soil profile. It is interesting to note that waste excavated from railroad cuts in 1887 shows little or no traces of soil.

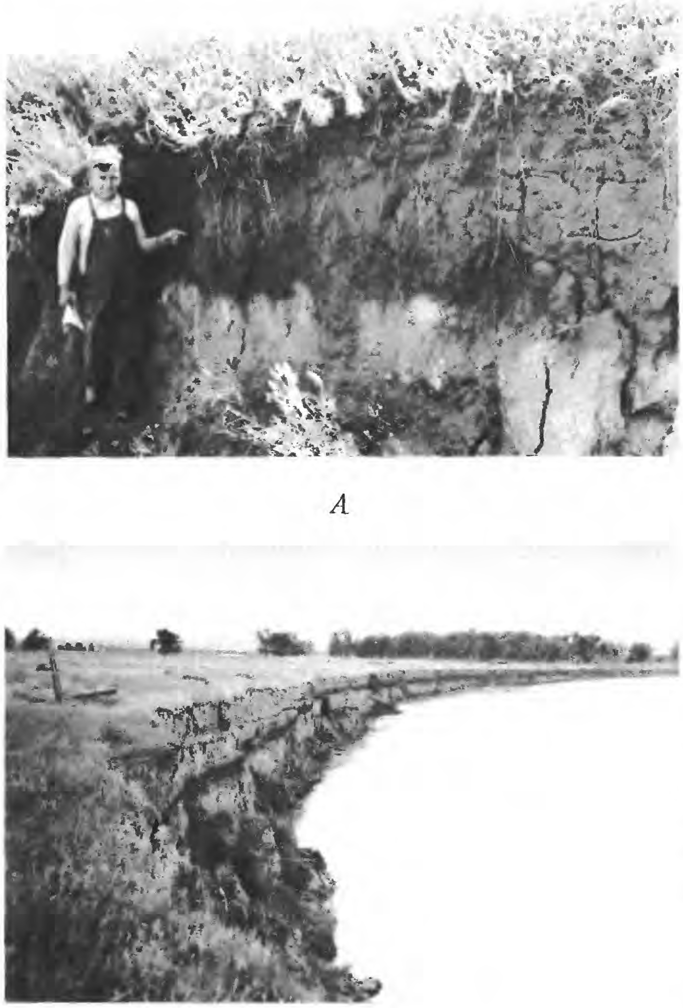


FIGURE 32.—*A*, Contact of leached loess (above) and soil-capped sand resting on till (below) in ditch on country road, 2 miles south of Froid, Mont. and 1 mile east of Montana State Highway 16. The boy is pointing to the top of a soil A horizon developed on sands resting on early Wisconsin(?) till. Below the A horizon is a white caliche zone. Above the dark soil is 3 to 4 feet of thoroughly leached dark-brown loess. *B*, Section of Missouri River flood plain at Snowden, Mont., about 2 miles west of the North Dakota boundary. The numerous dark bands probably indicate discontinuous aggradation. The cottonwood trees are characteristic of the flood plain.

The soil profiles on the loesses do not in themselves prove that the soil-forming intervals were of interglacial magnitude. Until specific stratigraphic or fossil evidence is available, the early Wisconsin(?), middle Wisconsin(?), and Mankato dates should be regarded only as suggestions.

RECENT ALLUVIUM

The flood plain of the Missouri River valley is immediately underlain by finely bedded muds, silts, and fine sands, with abundant gastropod shells. The alluvium conceals glacial deposits, including coarse gravels. It is impossible to make a sharp distinction in well borings between the Pleistocene and Recent alluvium, hence it is impossible to estimate the depth of the latter. Borings by the Great Northern Railway in the channel at Elbowoods, N. Dak., south of the problem area, revealed "quicksand" and river mud to a depth of 60 feet. The combined Pleistocene and Recent fill, including till, amounts in places along the Missouri River to more than 100 feet and to more than 200 feet a short distance up Big Muddy Creek valley. The alluvium alone in Big Muddy Creek valley is estimated to be 100 feet thick (Witkind, 1959, p. 28).

Available evidence indicates that the alluvium of the flood plain occupies an inner valley carved below the terrace silts. At numerous places, as at Snowden on the Montana-North Dakota line, several dark bands are exposed in the river bank (fig. 32*B*). Because of the steepness of the banks, these could not be examined in detail. It is unknown whether the bands are A horizons of buried soils or simply accumulations of organic matter on the flood plain. If they represent soil horizons, appreciable interruptions in aggradation are indicated. If they represent organic accumulations independent of the development of soil profiles, the interruptions may have been of relatively short duration.

COLLUVIUM

Under this heading are included all deposits which have accumulated as a result of mass movements. In places, as in the valley of the Little Missouri River due south of Watford City, N. Dak., landsliding has involved large segments of the valley wall. These segments form a belt of discontinuous, parallel ridges between the valley flat and the steep valley side. The internal structure of the blocks is generally undisturbed except for rotation and fracturing. In other localities, however, the sliding has apparently been more turbulent, and the deposits consist of a chaotic assemblage of "bedrock" fragments of all sizes in a fine-textured matrix. Where only one type of parent material, such as till, is involved, the appearance of the slide material closely resembles the original except for shear planes and local convolutions. Landsliding of this type, in contrast with the sliding en masse, gives rise to an irregular hummocky topography.

Gentle, concave-upward slopes between the valley flats and the valley sides are underlain by fine-textured and stratified wash deposits. More massive deposits, similar to those farther upslope, may have been emplaced by creep.

The colluvium has not been separately mapped in this investigation.

GEOLOGIC HISTORY

TERTIARY

Investigators are generally agreed that the Fort Union formation of Paleocene age was deposited over a low marshy and forested plain that sloped gently away from the Rocky Mountains. The coal beds and the floral assemblages indicate a mild, humid climate, a situation which apparently prevailed during parts of the Eocene as well. The Rocky Mountains are believed to have been low, permitting free access of westerly, moisture-bearing winds to the plains region. The change toward increasing aridity is attributed to progressive uplift of the mountains across the path of these winds.

There is no factual evidence bearing on the ancestral drainage of the northern Great Plains in early Tertiary time. Bauer (1915, p. 56–57) believes that the Missouri and Yellowstone Rivers date from the end of Cretaceous time. He believes that these rivers extended their courses across the level plain exposed by withdrawal of the Cretaceous seas and that the drainage at that time flowed into the predecessor of Hudson Bay or into the Arctic Ocean. He attributes deposition of the Fort Union sediments to these streams and their neighbors and believes that the period of deposition was terminated by an epeirogenic uplift accompanied by local folding. During the ensuing period of erosion, the Yellowstone and Missouri Rivers are believed to have maintained their courses across the rising Cedar Creek anticline and Poplar dome, respectively.

From the above, it would appear that the great age assigned to the Missouri and Yellowstone Rivers is based on the postulated antecedent relations of these streams to the early Tertiary structures mentioned. It seems equally possible, however, that the rivers were superimposed on the underlying structures from a higher surface, perhaps the depositional surface of the White River formation of Oligocene age, or from the Oligocene or Miocene degradation surface (Rimroad cycle). Such superposition would have had to antedate the Missouri Plateau peneplain because both the Missouri and the Yellowstone Rivers were apparently already established in approximately their present paths in Flaxville time, the cycle preceding that in which the peneplain was developed. This is indicated by the presence of the high-level gravel-capped Flaxville bench along these valleys.

The gross elements of the Oligocene or Miocene landscape are shown in figure 33. It will be noted that the Rimroad gravel, which is clearly fluvial and must once have occupied a valley floor, now caps the high divide between the Yellowstone River valley and Redwater Creek valley to the west. Its location suggests that, prior to Flaxville time (Miocene or Pliocene), the Yellowstone River lay farther to the west, perhaps beyond the termination of the Cedar Creek anticline.

Scattered buttes and mesas are capped by the White River formation of Oligocene age, which is only about 200 feet thick. Unless they were once appre-

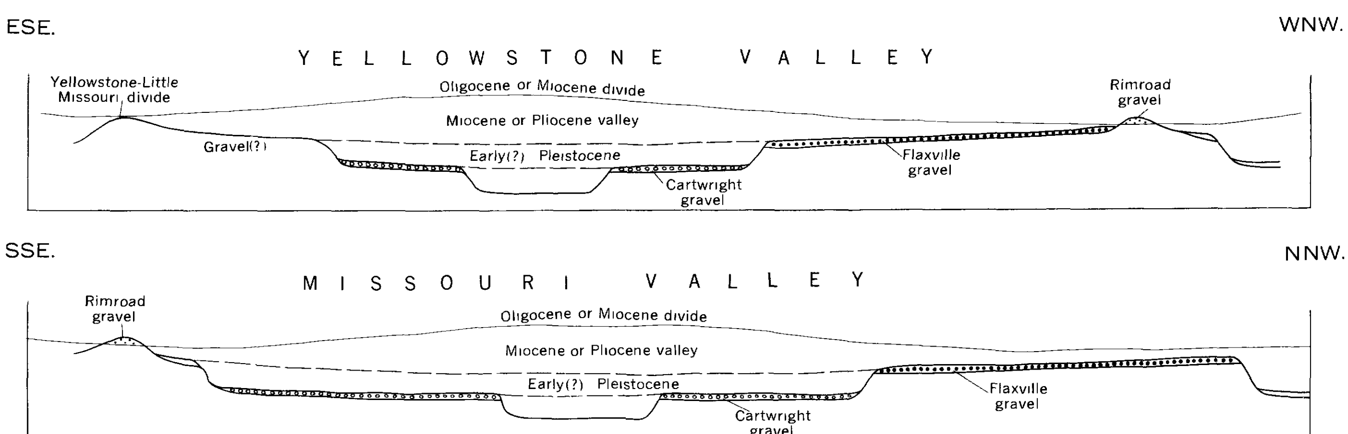


FIGURE 33.—Present and restored cross-profiles of the Yellowstone and Missouri River valleys. The profiles are oriented as they would appear to an observer looking upstream. The Rimroad gravel marks an Oligocene or Miocene valley bottom. The divide areas on either side were lowered more rapidly than the gravel-coated valley bottom to create the present valleys. The Flaxville gravel similarly retarded erosion of the Miocene or Pliocene valley floors, large parts of which now stand as high mesas. The Crane terrace within the inner valleys is not shown.

ciably thicker, their present elevations, together with the elevation of the Rimroad gravel, indicate a surface of low relief during the Rimroad cycle.

The Rimroad gravel and the gravels at lower elevations are strath gravels; they represent the normal complement of alluvium used by the rivers in erosion of their valleys. They do not record independent episodes of deposition. Thus, in the northern Great Plains, there is no record of aggradation subsequent to deposition of the White River formation of Oligocene age except for glacial and associated deposits. In the High Plains, however, in Nebraska and the States to the south and southwest, aggradation continued into Pliocene time (Ogallala formation).

The problem of why the northern plains suffered degradation while aggradation continued to the south is beyond the scope of this report. Possible explanations which have been generally recognized include climatic changes and differential tilting. If the degradation began in Oligocene or Miocene time, Pleistocene climatic changes alone are inadequate. The alternative to the climatic explanation is that at least part of the degradation resulted from stream rejuvenation due to differential uplift of the northern Great Plains during late Tertiary and Quaternary time.

Whatever the reasons, the northern Great Plains seem to have been undergoing degradation since the deposition of the White River formation of Oligocene age. The Rimroad gravel presumably records the first cycle of degradation. Alden (1932, p. 12) has correlated this gravel with the Cypress Hills gravel of southern Alberta. Mackin (1937, p. 871) finds that the Cypress Hills gravel actually consists of a thick older deposit of Oligocene age, for which he proposed retention of the name "Cypress formation," and a uniformly thin mantle of younger gravel as much as 40 feet thick of unknown age. For convenience in this report, the mantle gravel will be referred to as the summit gravel of the Cypress Hills. Inasmuch as the summit gravel occurs at higher elevations than the Flaxville gravel of Miocene or Pliocene age, it is probably older and may have been deposited at any time from the late Oligocene to Miocene or Pliocene time. The Rimroad gravel of the present report has some of the same characteristics as the summit gravel of the Cypress Hills: uniform thickness, absence of fossils, and exposure at higher elevations than the Flaxville gravel. Although these similarities are hardly valid evidence for correlation between widely separated areas, the Rimroad gravel may well be the equivalent of the summit gravel of the Cypress Hills.

The presence of the fluvial Rimroad gravel on the high divide west of the Yellowstone River valley indicates that a reversal of topography has taken place. The protective nature of the gravel is due to its high porosity and permeability which prevent the development of surface drainage. It is impossible to estimate either the breadth of the valley in which the Rimroad gravel was deposited or the location of the valley axis.

The Rimroad gravel resembles the gravels at lower levels along the Yellowstone River valley in texture, lithology, and surface slope. It was presumably deposited by the ancestral Yellowstone at a level 1,200 to 1,300 feet higher than it is now. In succeeding erosion cycles, the Yellowstone was confined to approximately its present position between the Redwater-Yellowstone divide on the west and the Yellowstone-Little Missouri divide on the east. At the close of the Flaxville cycle, the river was about 500 feet below the Rimroad valley floor and 600 to 700 feet above the present valley floor. At this level, the floor of the Yellowstone valley was probably more than 20 miles wide. The valley floor bore a mantle of gravel as much as 40 feet thick, the Flaxville gravel. In the Yellowstone-Missouri interconfluence area, the gravel forms the caprock of isolated plateaus, and these plateaus, like those north of the Missouri, record a second reversal of topography.

There seems little reason to doubt that in Flaxville time the Missouri River flowed at the level of the Flaxville bench which flanks the present valley to the north. The Flaxville Plateaus still farther north cannot be definitely related to the Missouri. Their large surface area may indicate a confluence of two or more rivers from the west, a possibility which could not be investigated during the course of the present study.

The Tertiary history of the problem area may be briefly summarized as follows: In Paleocene and part of Eocene time, the area was a low, marshy, forested plain. Whether the Yellowstone and Missouri Rivers were established in approximately their present courses at that time is unknown. Regional erosion preceded deposition of the White River formation of Oligocene age. This erosion, and the subsequent deposition of more than 200 feet of Oligocene sediments, may have obliterated surface expression of the Glendive (Cedar Creek) and Poplar anticlines and other early Tertiary structures. The Rimroad gravel at this time was valley-bottom alluvium. Its coarseness and thickness indicate a stream which was at least periodically vigorous. The White River formation and its contained fauna indicate more arid conditions than those that prevailed during the Paleocene and early Eocene. This probably correlates with uplift of the Rocky Mountains and the denial of much moisture to the northern plains. The uplift of the mountains and the increasing aridity probably account for the large and coarse load supplied to the plains rivers. In consequence of this large and coarse load, the plains streams flowed on relatively steep gradients even though graded. The vertical distribution of the White River formation suggests that the Oligocene or Miocene surface was probably one of low relief. The Yellowstone and Missouri Rivers might have been superimposed from the surface of the Oligocene beds onto the structures below.

A complete reversal of topography took place during the Flaxville cycle. Uplift of the mountains and a slight increase in the easterly slope of the plains may have caused the master streams to incise themselves in their gravelly flood plains. Tributaries found downcutting much easier in the unprotected divide areas than in the now-elevated gravel-capped valley bottoms. In time the divide areas were worn lower than the former valley floors, whose detached remnants now form the highest parts of the Yellowstone-Redwater divide.

When the master streams again attained grade, they opened out broad valleys at the Flaxville level. The Yellowstone River valley at this time was 20 to 30 miles wide, and the Missouri River valley was probably wider. Slight downcutting during lateral widening of the valleys may account for the valleyward slope of the Flaxville surfaces, although the slope may have resulted from subsequent smoothing of an original terraced surface.

A second rejuvenation led to incision below the Flaxville valley floors. Again more rapid lowering of the divide areas resulted in the preservation of large parts of the former valley bottoms as gravel-capped uplands. Along the Yellowstone River valley these uplands form a high, broad bench rising westward toward the divide capped by Rimroad gravel. The distribution of the gravels (pl. 1) suggests that at the close of the Flaxville cycle the Yellowstone River was on the east side of its gravelly flood plain. Intrenchment, therefore, left large areas west of the river capped by "protective" gravel, whereas only small areas to the east were similarly protected. This may explain the greater dissection east of the river. In the Yellowstone-Missouri interconfluence area the Rimroad gravel divide is absent and the Flaxville valley floor is preserved as isolated mesas. North of the Missouri, the valley bottoms of Flaxville time are also preserved as high gravel-capped mesas. Those between the Missouri River and the valley of Cottonwood Creek to the north rise gently away from the Missouri. They apparently mark the high-level course of the Missouri River. The Flaxville Plateaus farther north are of great lateral extent and may represent broad interconfluence areas involving unknown streams.

In the absence of evidence to the contrary, it is assumed that the general paths followed by the Missouri and Yellowstone Rivers in Flaxville time across the map area were maintained continuously until disrupted during the Pleistocene.

Alden (1932, p. 31) suggests that the development of the Flaxville surface may not have been completed until Pleistocene time. This suggestion is based on the identification by Gidley (Collier and Thom, 1918, p. 180–181) of a *Camelops* tooth from the Flaxville gravel, as well as Alden's observation that the Flaxville Plain had not been greatly dissected prior to glaciation of the mountains. It should be noted, however, that Gidley was not entirely certain of his identification of the tooth and in addition referred to it only as "apparently Pleistocene." All the elements of the fauna indicate a Miocene or Pliocene age. Mackin (1937, p. 872) also comments on "the shallow depth of stream trenching between the Flaxville surface and lower terraces of known Pleistocene age, and the relative narrowness of the valley floors opened out in the weak rocks of the Plains by vigorous streams issuing from the mountains in post-Flaxville time." He suggests that in spite of the faunal assemblage in the gravels, the Flaxville surface may be Pliocene or early Pleistocene in age. Actually, in North Dakota at least, the first of the Pleistocene surfaces, the Missouri Plateau peneplain, is as much as 600 feet below the Flaxville valley bottoms and is quite extensive. At best, the problem of a Pleistocene versus a Miocene or Pliocene age for the Flaxville Plain must be considered unsettled.

PLEISTOCENE

MISSOURI PLATEAU PENEPLAIN

Alden (1932, p. 31) suggests that the presence of an old till on possible remnants of the Flaxville surface close to the Rocky Mountains indicates that the Flaxville Plain had not been greatly dissected at the start of glaciation. The trenching of the Flaxville Plain and the beginning of glaciation are inferred to have been initiated by regional uplift. Gidley (Collier and Thom, 1918, p. 181), however, had concluded that, with the exception of the "apparently Pleistocene" *Camelops* tooth, the gravel "cannot be older than Miocene or younger than lower Pliocene." If Gidley is correct, then dissection of the Flaxville Plain may have begun during the Pliocene. The statements of Alden and Gidley, however, are not necessarily contradictory. Uplift initiated in Pliocene time may not have reached sufficient magnitude to cause glaciation until Pleistocene time.

Reasons have already been presented indicating that the Missouri Plateau peneplain is probably as young as the Kansan. The Missouri and Yellowstone Rivers flowed across the peneplain at the time of the first ice invasion. In the western part of the area they were confined to valleys 500 to 600 feet below the Flaxville uplands. Farther east beyond the limits of the Flaxville Plateaus, the rivers occupied broad, relatively shallow valleys on the peneplain, and the total relief, exclusive of isolated buttes and mesas, amounted to only a few hundred feet.

Plate 8 illustrates the author's conception of the ancestral drainage of the map area in relation to major ice-marginal channels. The plotted courses of the major valleys differ little from those proposed, in whole or part, by previous investigators. The ancestral paths are buried under thick drift in the northeastern part of the map area. Indicated widths of the ancestral valleys of the Missouri, Yellowstone, and Little Missouri Rivers represent the valley floors and the narrow, irregular east-west connecting channels are ice-marginal channels now occupied in part by master streams. The trends of some tributaries of the ancestral master streams, shown by single heavy lines, coincide with parts of the ice-marginal channels and may have influenced the paths of the ice-marginal waters. A number of other streams which were glacially diverted are also shown.

The evidence for the ancestral courses of the master streams is reasonably reliable and will be considered in detail. The evidence for the ancestral paths of the minor streams is less reliable and will be discussed briefly. In some instances, it consists solely of certain peculiarities in the present drainage pattern.

ANCESTRAL DRAINAGE

The present path of the Missouri River is anomalous in that it crosses the region almost at right angles to the northerly regional slope. That this anomalous course is of ice-marginal origin is suggested by the fact that it parallels the almost continuous line of channels which mark the glacial limit to the south (pl. 8), and it approximately parallels a less continuous line of glacial channels to the north between Grenora and Powers Lake, N. Dak. In other words, the present valley of the Missouri River is one member of a set of east-west valleys, the others of which are unquestionably of glacial origin.

That the ancestral drainage of the region flowed down the regional slope to the north is suggested by broad swales in the Missouri Plateau peneplain north of the Missouri River. One such swale extends across the upland from the Missouri River valley at Poplar, Mont., and probably represents the ancestral path of the Missouri. Another continues north of Williston, N. Dak., along the present valley of Little Muddy Creek, and represents the former path of the Yellowstone River. The upland swale along the White Earth River valley (shown only near the mouth of the valley in pl. 8) may represent the former path of the Little Missouri River. The network formed by the north-trending upland swales on the one hand, and the east-trending valleys and glacial channels on the other, is remarkably similar to the drainage network of the north German plain. There, too, the east-west drainageways are regarded as being ice-marginal channels developed across the paths of north-flowing streams.

An ice-marginal origin for the Missouri River valley is also suggested by the abrupt variations in width downstream from Poplar (fig. 5). In some stretches, the valley is as much as 4 miles wide, whereas elsewhere its width decreases to less than a mile. The narrow segments cross the north-trending divides between the upland swales, whereas the broad segments appear where the present valley crosses the broad swales or locally coincides with them. The narrow segments are not due to greater rock resistance, because the rocks are all rather uniformly weak clays, silts, and sands. The narrower segments are readily explained, however, on the assumption that the present valley was developed along the path of ice-marginal waters. Thus, where the ice-marginal path crossed or coincided with the broad ancestral swales, the river had less material to remove in subsequent deepening of its valley than it had in high, divide areas. Under these circumstances, once grade was attained, the river was able to enlarge its valley more rapidly in these areas than in the divide areas.

MISSOURI RIVER

General G. K. Warren (1869, p. 311) suggested that the present course of the Missouri River was glacially determined. He wrote:

> There, then, on that limit [glacial limit] a river must have been formed to carry away the melting water from the glacier, and this limit was the Missouri River. * * *
> It cut along this glacial limit because all the streams west of it came from the mountains toward it, down the inclined plain, and there their old course was terminated.
> We see what lakes must have periodically formed here; what great barriers must have been formed and burst, one after another, and what deluges the lower valley must have experienced.

Warren's shrewd observations regarding the origin of the Missouri River in the north-central Great Plains have been seriously challenged only by Leonard (1916a, p. 295), who regards the present path of the Missouri as preglacial, that is, "older than the oldest ice-invasion of this region." All others, including the writer, believe that the evidence confirms Warren's original views, that is, with the possible exception of the "bursting dams."

As for the direction of flow of the preglacial drainage, Todd, in 1884 and again in 1914 and 1923, suggested that the major streams of eastern Montana and western North Dakota flowed northeastward into Hudson Bay. Prior to Todd's two latter papers, Beekly (1912) had already described a possible ancestral valley of the Missouri River in northeastern Montana north of the present valley. He wrote (p. 323):

> A broad, shallow depression, strongly suggestive of a wide preglacial river valley, extends from Medicine Lake northeastward to the border of the flat plains area and continues no less prominently across the hilly region on the northeast [coteau belt] * * * This long valley-like depression may have been either the preglacial channel of the Missouri or that of a large tributary to the Missouri.

Bauer (1915) directed attention to the broad hanging valley in the north wall of the Missouri River valley just east of Poplar, Mont. (fig. 34). The hanging valley extends northeast for 20 miles to Big Muddy Creek valley, with which it apparently coincides for 15 miles to the vicinity of Medicine Lake. Beyond here it continues as the broad swale noted earlier by Beekly. Thus, Bauer mapped the preglacial course of the Missouri across the upland from Poplar to the coteau belt, and he suggested that the Missouri once continued to Hudson Bay. Alden (1932, p. 58) and Witkind (1959) support the view that the preglacial Missouri originally flowed to the northeast, into Arctic drainage.

There seems little reason to doubt that the ancestral course of the Missouri continued as suggested by Beekly, Bauer, Alden, and Witkind past the area now occupied by Medicine Lake to at least as far as Grenora, N. Dak., on the south side of the coteau belt. The course beyond Grenora is a matter of speculation.

Alden (1932, p. 128) suggested that the ancestral valley of the Missouri crossed the present coteau belt along a topographic swale now followed by the railroad between Alkabo and Colgan in northwestern North Dakota (pl. 2). This swale, however, is only about 2 miles wide, a fraction of the width of the ancestral valley elsewhere; hence, if Alden is correct, it must be assumed that the valley was appreciably constricted here. Yet, the probable path of the ancestral Yellowstone River across this same area is marked by a swale fully as broad as the ancestral valley of the Yellowstone elsewhere. On a comparative basis, therefore, it seems unlikely that the narrow Alkabo-Colgan swale represents the ancestral path of the Missouri. Alpha[8] suggested that the ancestral valley of the Missouri continued north from Grenora, N. Dak., to Westby, Mont. (fig. 35), and then turned northwestward into Saskatchewan for 160 miles to beyond Johnstone Lake (fig. 38).

The dimensions of the surface swale support the view that the Missouri continued north to Westby. However, the greater part of the course in Saskatchewan is narrow and is more probably an ice-marginal channel. Although it is of no consequence to the immediate problem, Alpha[9] erroneously attributes to Alden (p. 129) the suggestion that this ancestral course "was subsequently used by the Saskatchewan River at the close of the late Wisconsin ice invasion." Alden was at that point referring to Big Muddy Creek valley and its continuation to the north.

YELLOWSTONE RIVER

The ancestral Yellowstone River flowed in a broad, shallow valley across the Missouri Plateau peneplain (pl. 8). Its course coincided with the present valley of the Missouri from the Montana-North Dakota line to Williston and continued north along the broad swale now occupied by Little Muddy Creek, as suggested by Alden (1932, p. 273). Concrete evidence for at least part of this course is to be found in the lithology of the high-level gravel (Cartwright gravel, pl. 1 and fig. 42) on the west side of Little Muddy Creek valley at the mouth of Cow Creek. The gravel contains numerous pebbles of silicified volcanic rocks, which are distinctive of the gravels at all levels above the Yellowstone River south of that river's junction with the Missouri River. The high-level gravels of the Missouri River, on the other hand, consist almost exclusively of quartzite, chert, and quartz.

Alden projected the ancestral course of the Yellowstone from Appam, N. Dak., in Little Muddy Creek valley, to Crosby, N. Dak., on the north side of the coteau belt. The deep, broad swale across the morainal belt, however, is slightly farther west of Alden's projected course (fig. 29). The swale has a width of about 15 miles and a depth of more than 300 feet, which corresponds closely with the estimated breadth of the Yellowstone River valley at the peneplain level south of the present Missouri valley. The depression is the surface reflection of a deep bedrock valley, the floor of which, at Crosby, is about 500 feet below the surface (Waring and LaRocque, 1949, pl. 6).

[8] Alpha, A. G., 1935, Geology and ground-water resources of Burke, Divide, Mountrail, and Williams Counties in North Dakota: N. Dak. Univ., unpub. thesis, p. 10.

[9] Op. cit., p. 11.

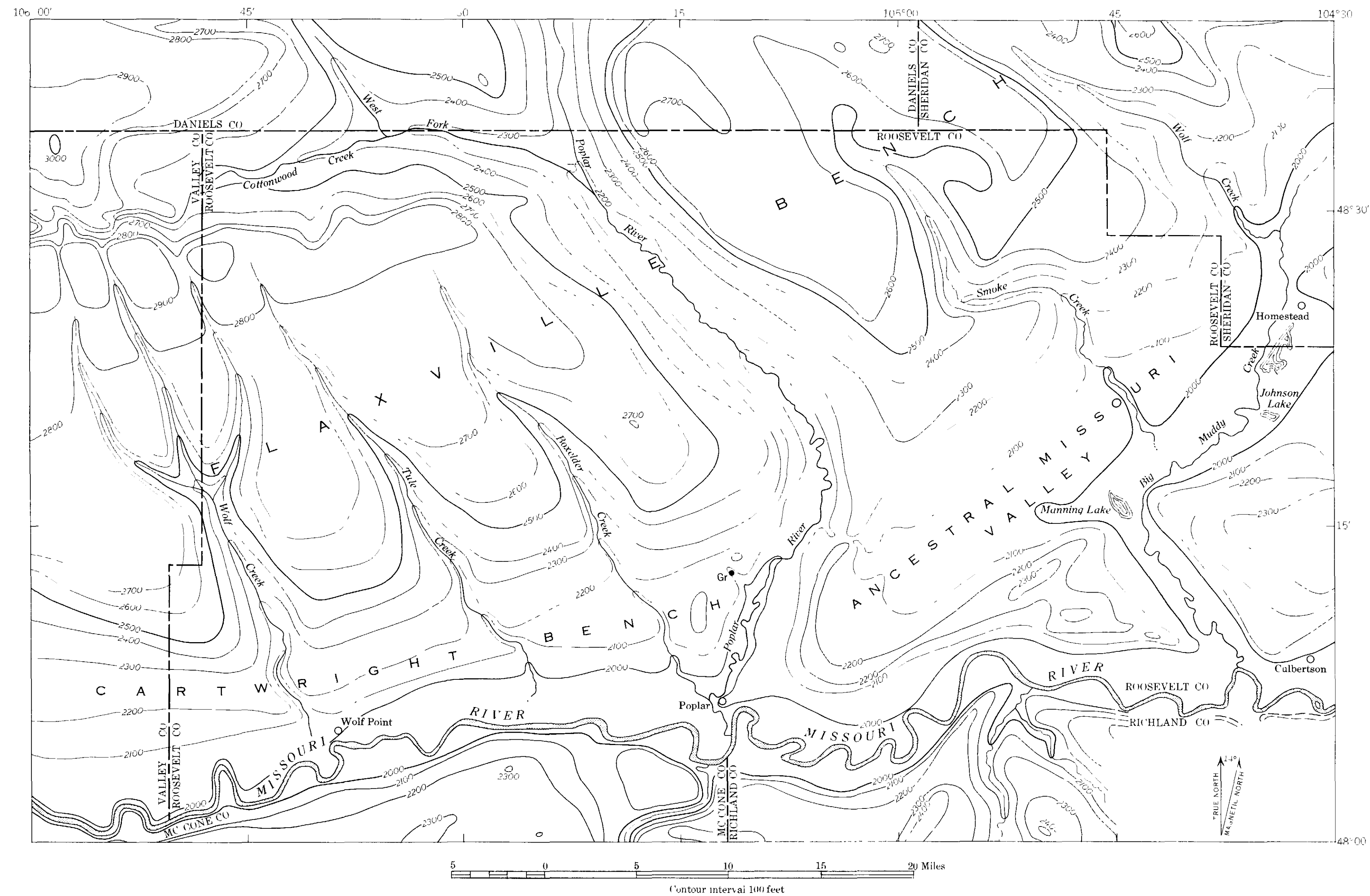


FIGURE 34.—Generalized contour map of the greater part of Roosevelt County, Mont., and small parts of adjoining counties. The "Ancestral Missouri Valley" marks the former northeastward path of the Missouri River at the level of the Missouri Plateau peneplain. The high Flaxville bench follows the same trend. The letters *Gr* indicate a high-level exposure of gravel that probably represents a remnant of the floor of the ancestral valley. The map was prepared from a preliminary U.S. Geological Survey topographic map of the Fort Peck Indian Reservation and from standard quadrangle maps. The meanders of the Missouri were corrected from photo indexes.

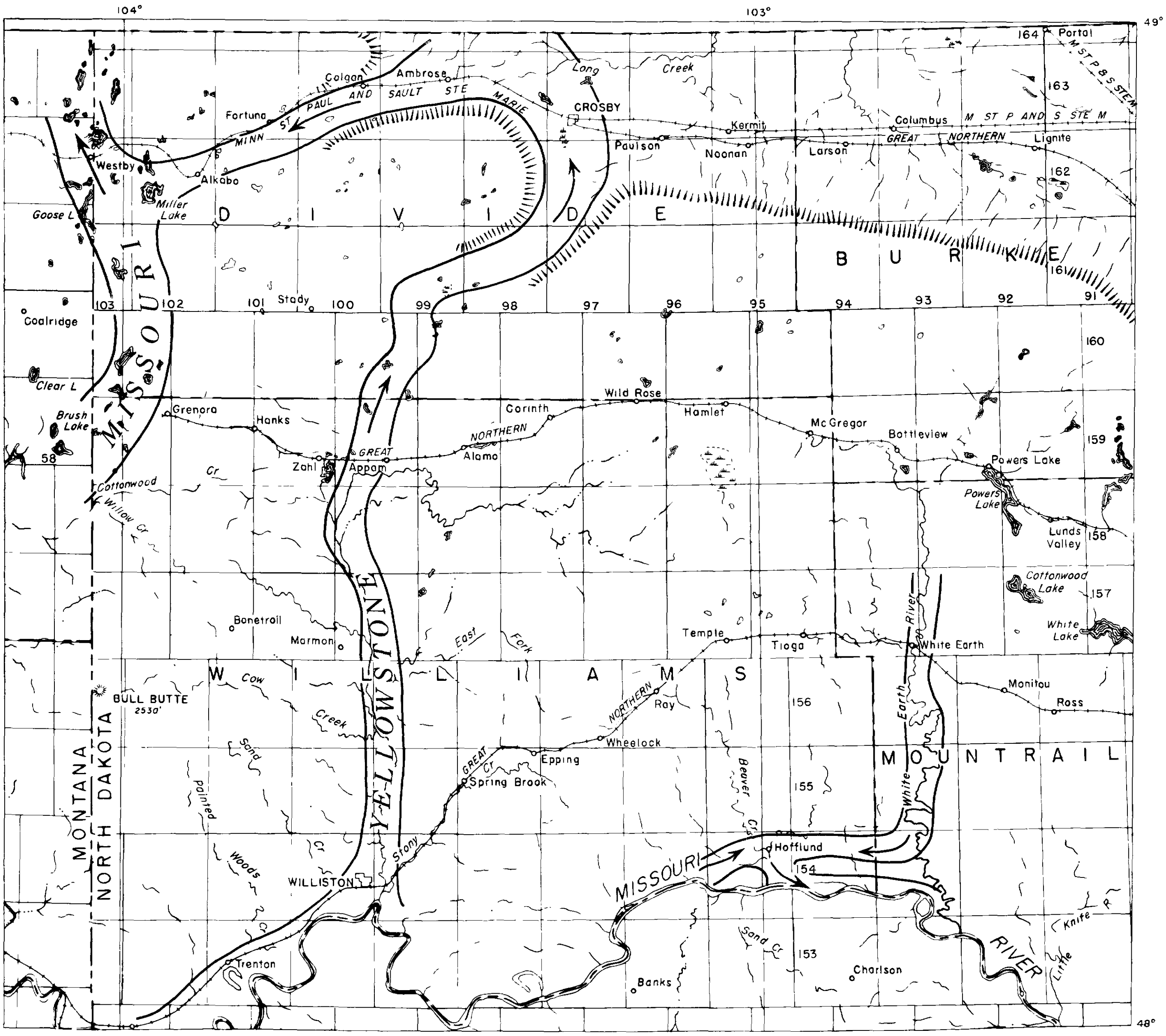


FIGURE 35.—Ancestral drainage of northwestern North Dakota, after Alpha.[10] Actual widths of the ancestral valleys are greater than shown by Alpha. The Yellowstone River is presumed to have turned back on itself near Crosby, N. Dak., to join the Missouri in Montana. In the southeast is shown Alpha's interpretation of the ancestral drainage responsible for the abandoned channels near Hofflund, N. Dak. Township and range lines are at 6-mile intervals.

[10] Op cit.

The above considerations indicate that Alden was essentially correct in his charting of the course of the ancestral Yellowstone River from Williston to Crosby (pl. 8).

The ancestral course beyond Crosby is unknown. Alpha [11] suggested that the Yellowstone turned back sharply to the southwest to join the Missouri at Westby, Mont. (fig. 35). Alpha was apparently influenced by the fact that as yet no buried valley has been reported in southern Canada north of Crosby. It is difficult to understand, however, why the Yellowstone, after having passed out of the plateau to the plains beyond, should have turned back and reentered the plateau. Furthermore, if the swale is too narrow to represent the former path of the Missouri, as suggested earlier, then it is too narrow to represent the former path of the Yellowstone. If we judge the paths of the ancestral drainage on the relative widths of the respective swales, it seems more reasonable to assume that the Missouri continued northwest from Westby, as suggested by Alpha, and that the narrow swale between Westby and the ancestral Yellowstone near Crosby is a pre-Mankato ice-marginal or proglacial channel. Further speculation is useless until more information on the buried topography becomes available.

DATE OF DIVERSION OF THE MISSOURI AND YELLOWSTONE RIVERS

Available evidence favors a Kansan date for the diversion of the ancestral Missouri and Yellowstone Rivers. The evidence for a pre-Wisconsin date will be presented first; the evidence for the precise date (Kansan) within the pre-Wisconsin will be considered later. The reasons for believing that the Missouri and Yellowstone were diverted at the level of the Missouri Plateau peneplain (No. 2 bench of Alden, 1932) are (1) the broad, shallow ancestral valleys are peneplain valleys, that is, together with the broad, low interfluves they form the Missouri Plateau peneplain, and (2) the absence of continuous trenches within the ancestral valleys suggests that the rivers did not persist in these valleys during the succeeding Crane or deep-trench erosion cycles.

A drilling program sponsored by the U.S. Geological Survey revealed a deep inner valley near Homestead, Mont., along the presumed ancestral course of the Missouri (pl. 8). The floor of the valley is at an elevation of 1,744 feet. Big Muddy Creek, which flows south, follows the ancestral valley of the Missouri for many miles, but in a direction opposed to the original flow of the Missouri. It leaves the ancestral valley near Manning Lake and follows a southeasterly course for about 10 miles to the Missouri. Drilling by the Bureau of Reclamation revealed that the rock floor, in this southeasterly stretch, is about 200 feet below the flood plain, or at an elevation of 1,690 feet. If the bedrock valley near Homestead is continuous with that along the present lower course of Big Muddy Creek, then the latter was flowing south while it occupied the buried valley. In other words, the presence of the bedrock valley at Homestead does not prove that the ancestral Missouri River continued to follow the ancestral course during a subsequent cycle of erosion. It is possible, of course, that the bedrock valley at Homestead is unrelated to that along the lower course of Big Muddy Creek valley. As Witkind (1959, p. 13) suggests, it may mark the path of an independent stream. For reasons to be presented later, its seems unlikely that the diversion could date back to the Nebraskan; a Kansan or later date seems more likely. Alden (1932, p. 44–45, 58) had suggested as evidence of the peneplain level of the diversion of the Missouri, the absence of the No. 2 (Cartwright) bench along the present diversion valley downstream from Poplar, Mont. However, the bench would be absent below Poplar even if the Missouri had continued along its ancestral course during the subsequent deep-trench erosion cycle.

EVIDENCE FOR PRE-WISCONSIN DATE

The presence of early Wisconsin(?) till within the present diversion valley of the Missouri could mean either that the diversion of the ancestral drainage was pre-early Wisconsin(?) in age or that the diversion trench was carved during a halt in advance and (or) retreat of the early Wisconsin(?) ice. If the trench was carved during a halt in retreat of the ice, the presence of the till within the trench requires a subsequent readvance. For the reasons presented below, it is believed that the diversion of the Missouri took place in pre-early Wisconsin(?) time.

If the entire depth of the present trench had been carved during a halt in advance of the early Wisconsin(?) ice, 300 to 500 feet of downcutting would have been required at this time, down to and below present river level. The presence of the Crane terrace, however, midway between the upland and the buried rock floor, indicates that the trenching did not take place during a single halt of the ice; two cycles of erosion are indicated. During the first cycle, erosion proceeded to the level of the Crane terrace, and the trench was opened out to widths of several miles. During the second cycle, the deep inner trench was eroded 100

[11] Op. cit., p. 12.

to 200 feet below the Crane valley floor. For the same reason, it is improbable that the trench was carved during a halt in retreat of the early Wisconsin(?) ice.

The assumption that the trench was carved in part during a halt in advance of the ice and in part during a halt in recession meets other objections. Let us assume that during advance of the ice the trench was eroded to the Crane level. The width of the trench at this level is about 4 miles, about 4 times greater than the width of the trench carved by the ice-marginal Missouri at the glacial limit. Presumably, then, the ice remained many times longer at the site of the present trench than it did at the glacial limit. The gravels that mark the Crane level of erosion, however, are nonglacial, a fact that belies the presence of an immediately adjacent ice front. A second objection is the presence of remnants of the Crane valley floor in valleys north of the Missouri trench. This northern area was presumably covered by ice while the ice front stood at the present Missouri trench. If erosion to the Crane level took place at this stage, then the Crane surface could not have been developed north of the trench.

Other objections arise when subsequent development of the trench is considered. The hypothesis requires that the ice front, in its recession from the glacial limit, halt again on the north side of the trench and remain there long enough to permit erosion of the inner trench to a depth of 100 to 200 feet below the Crane terrace. A readvance must then be postulated to explain the till which mantles the terrace, drapes over its edge, and extends below present river level. This involves more than a modest readvance of the ice because the early Wisconsin(?) till is found below the level of the Crane terrace far to the south in the Yellowstone River valley. The inner trench, below the Crane terrace, is also present in valleys north of the Missouri. According to the hypothesis under consideration, the inner trench could only have been eroded after the ice uncovered this northern area. The presence of the early Wisconsin(?) till within the trench in this region too, however, demands another extensive readvance of the ice. It might be argued that the ice front, in its retreat from the glacial limit, did not halt at the Missouri but continued to move back until it lay several tens of miles to the north. The inner trench might then have been eroded simultaneously north and south of the Missouri, and only a single readvance would be necessary to explain the till within the inner trench. This hypothesis, however, calls for a major readvance of the ice, for which there is no evidence, and fails to meet the other objections mentioned above.

The theory that the Crane terrace and inner trench were formed prior to the early Wisconsin(?) glaciation avoids the above objections. According to this theory, the diversion that led to the present ice-marginal course of the Missouri took place in pre-early Wisconsin(?) time. The original ice-marginal channel need only have been eroded to a depth necessary to prevent the ancestral streams from returning to their former northerly courses. Erosion to the Crane level took place in a subsequent interglacial interval, as indicated by the essentially nonglacial Crane Creek gravel. Later, erosion of the deep inner trench left the Crane valley floor standing as the Crane terrace. This multicycle topography was then overridden by the early Wisconsin(?) ice and blanketed by the early Wisconsin(?) drift.

A pre-early Wisconsin(?) age of the Missouri diversion is further supported by evidence indicating that the Poplar River, Big Muddy Creek, and Little Muddy Creek were already flowing southward into the present Missouri trench prior to advent of the early Wisconsin(?) ice. The evidence consists of exposures of early Wisconsin(?) till in these valleys more or less continuously to their junction with the Missouri. The till not only veneers the valley sides but descends to creek level in many places, and at least in Big Muddy Creek valley continues below flood-plain level. This suggests that the pre-early Wisconsin(?) Missouri River valley was deeper than it is now. Erosion of the south-sloping valleys cannot be attributed to melt waters flowing south from the ice during its advance, because if the Missouri valley had not yet been in existence these waters would have merely ponded in front of the ice and spilled sideways along the ice front. Nor can the depth of the valleys be attributed to ice erosion, because (1) at least parts of these valleys lie athwart the presumed direction of ice movement, and (2) it is extremely unlikely that the ice, riding up the regional slope, would erode valleys sloping in the reverse direction toward a Missouri trench which had not yet come into existence. These and earlier considerations suggest that the diversion of the Missouri and Yellowstone Rivers to form the present east-trending valley was pre-early Wisconsin(?) in age.

EVIDENCE FOR KANSAN DATE

There then remains the problem of when, in pre-early Wisconsin(?) time, the diversion took place. The amount of dissection of the walls of the Missouri trench where it crosses the former divides is slight except near large tributary streams. This is true even of those parts of the trench that were not affected by later diversions as at Culbertson, Hofflund, and Sanish. The badland belts are only a few miles wide. The trench itself in these newer reaches may be a mile or less in width, as compared with valley widths of 3 to 4 miles elsewhere (fig. 5). In the newer stretches, the valley is only in early maturity; it is not yet wide enough to accommodate fully developed meanders. The small amount of subsequent dissection and valley widening suggest that the diversion is probably not as ancient as the Nebraskan.

Warren (1952) concluded from his studies in the vicinity of the junction of the White and Missouri Rivers in South Dakota (fig. 1) that the diversion of the ancestral drainage in that area took place in Illinoian time. The relations of the till(?) at Smoke Creek to the ancestral valley of the Missouri, however, suggest an earlier date for the diversion in North Dakota. These relations will be briefly considered.

The till(?) at Smoke Creek occurs within the ancestral valley of the Missouri (pl. 8). The base of the till is at an elevation of about 2,000 feet. The surface on which the till(?) at Smoke Creek rests is 30 to 70 feet below the projected floor of the ancestral valley, depending on which set of figures is used in restoring the original profile between Poplar and Homestead, Mont. This suggests that the ancestral valley antedates the advent of the Smoke Creek ice.

The gravels below the till(?) at Smoke Creek show imbrication toward the south, which is the present direction of flow of this part of Smoke Creek. This suggests that Smoke Creek was responsible for development of the gravel-covered rock floor which is now preserved as a strath terrace. The Missouri, therefore, had already abandoned its ancestral valley prior to the development of the Smoke Creek strath. The till(?) at Smoke Creek overlies the strath gravels, hence it is also later than the Missouri diversion. The till(?) in turn is overlain by early Wisconsin(?) till, hence it is pre-Wisconsin(?). If the till(?) at Smoke Creek is Illinoian, the date of the Missouri diversion was probably Kansan.

Future surveys may reveal additional exposures of till of Smoke Creek age. The vertical distribution of such occurrences should help to settle the question of the date of the Smoke Creek glaciation relative to the date of abandonment of the ancestral valley. Actually, a few additional deposits of possible pre-Wisconsin age were discovered during the course of the present investigation. These have already been described. They consist of much-altered glaciofluvial deposits. It is interesting to note that these ancient-appearing deposits occur below the level of the Missouri Plateau peneplain in a valley tributary to the present Missouri trench. If these deposits are actually equivalent to the till(?) at Smoke Creek, then their presence below the level of the ancestral valley also implies that the diversion of the Missouri took place at a higher level as a result of a pre-Smoke Creek glaciation. However, the intense alteration of these glaciofluvial deposits may be due to rapid groundwater percolation favored by a high degree of permeability and a favorable topographic setting. Perhaps the deposits are early Wisconsin(?) in age.

In summary, available information favors a Kansan date for the diversion of the ancestral Missouri River. The Kansan date is supported by the histories of other valleys to be described in subsequent sections of this report.

LITTLE MISSOURI RIVER

Evidence herein presented indicates that the ancestral Little Missouri River was also blocked and diverted at the level of the Missouri Plateau peneplain at the site of the present Missouri trench (pl. 8). The river continued to flow north into the Missouri, however, during the Crane and deep-trench erosion cycles until permanently diverted to the east in southern McKenzie County, N. Dak., by the early Wisconsin(?) ice. The ancestral Missouri and Yellowstone Rivers were likewise diverted along the margin of the early Wisconsin(?) ice, but on recession of the ice these rivers resumed their former paths.

The ancestral valley of the Little Missouri River is well defined topographically (pls. 2 and 8). Attention was first directed to this valley by Wilder (1905), and the valley was described in detail by Leonard (1916a). The Little Missouri River (pl. 2) follows a northerly course into McKenzie County, N. Dak., but about a dozen miles north of the county line it turns abruptly eastward. At this point, a deep hanging valley appears in the north wall of the canyon and extends as a broad, well-defined valley 55 miles northeast to the Missouri River near Hofflund, N. Dak. This trunk valley is now occupied by several independent streams (pl. 1). Bowline and Redwing Creeks occupy small parts of the valley at the south end; Cherry Creek follows it to the northeast for about 20 miles to beyond Watford City, N. Dak.; and Tobacco Garden Creek occupies the valley for the remainder of the distance to the Missouri River. Except at either end, the floor of the ancestral valley is monotonously smooth, and the divides between Redwing, Cherry, and Tobacco Garden Creeks are broad and low. At and near the extremities, however, the valley floor has been eroded in response to downcutting by the master streams.

Gravels are exposed in the ancestral valley between Redwing Creek and the present trench of the Little Missouri (pl. 1). The best exposures are on the broad divide between Redwing and Bowline Creeks. The gravels are nonglacial, consisting almost exclusively of waterworn Flaxville-type pebbles such as chert, quartzite, and silicified igneous rocks, but they contain a few pebbles of local bedrock types. The pebbles average more than 1 inch in diameter, but they include individuals as much as 5 inches in diameter. The gravels occur in two levels, the lower of which is the most extensive and is the surface below which Redwing and Bowline Creeks have eroded their wide inner valleys. The two terraces are 150 and 200 feet (estimated) above the level of the Little Missouri River at the point where the ancestral valley starts across the upland.

In many places the back-slope of the higher terrace is gentle and merges with the upland. The terrace probably represents the floor of the Little Missouri River valley across the Missouri Plateau peneplain, and it is tentatively correlated with the Cartwright terrace of the Missouri and Yellowstone River valleys. The lower terrace is tentatively correlated with the Crane terrace of the Missouri and Yellowstone valleys.

If the correlations are correct, the Little Missouri River continued to occupy its ancestral valley at least as late as the Crane cycle of erosion. Independent evidence indicates that it occupied its ancestral valley even during the subsequent deep-trench cycle. This evidence is discussed below.

Nonglacial waterworn Flaxville-type gravels cap a remnant of the Crane terrace and are in turn covered by early Wisconsin(?) till in a long cutbank about 1 mile above the point of entry of Tobacco Garden Creek into the Missouri River valley (pl. 1, and fig. 36). The level gravel-capped bench is exposed intermittently for a distance of 200 to 250 feet. The gravels are coarse, comparable in texture with those of the gravel-capped terraces to the south. Unless little Tobacco Garden Creek, which now transports only silt, was far more vigorous during the Crane cycle, it could hardly have deposited these coarse gravels. The gravels were probably deposited by the same stream (ancestral Little Missouri River) which deposited the gravels of the terraces to the south.

If, for the sake of argument, we assume that the till-covered gravels were actually deposited by Tobacco Garden Creek, we must then seek a pre-early Wisconsin(?) cause for the diversion of the Little Missouri River. The assumption that an earlier ice reached and diverted the river involves the coincidence that two distinct ice sheets extended to precisely the same limit. Furthermore, if the terrace gravels near the mouth of the ancestral valley were deposited subsequently by Tobacco Garden Creek, it is strange that they do not contain glacial pebbles derived from the older drift.

Thus, available evidence indicates that the Little Missouri was diverted from its ancestral valley during or after the deep-trench erosion cycle. Diversion by piracy seems unlikely because of the remarkable coincidence of the course of the captor stream and the glacial limit and the difficulty of visualizing any advantages the captor stream would have had over any of its eastward-flowing neighbors.

FIGURE 36.—Early Wisconsin(?) till (*Qewd*) on gravel-capped bedrock terrace in Tobacco Garden Creek valley about 1 mile upstream from the Missouri River valley (sec. 2, T. 153, N., R. 97 W.). *Qcg*, Crane Creek gravel; *Tfu*, Fort Union formation with coalbed. Height of bluff is about 50 feet. To the right of this view, the till drapes over the front edge of the terrace, indicating that the terrace was already in existence prior to the advent of the early Wisconsin(?) ice.

The strongest argument for diversion of the Little Missouri River by the early Wisconsin(?) ice is the faithfulness with which the river follows the drift border. To the west, the glacial border is marked by a string of deep channels which are alined with the east-trending course of the Little Missouri River valley (pl. 1). The principal difference between these channels and the trench of the Little Missouri is that the former are abandoned whereas the latter is still in use.

It is concluded from the preceding considerations that the Little Missouri River flowed north into the Missouri River until the advent of the early Wisconsin(?) ice. Prior to that time, the Little Missouri River had eroded its ancestral valley down to and below the level of the gravel-capped bench at the mouth of Tobacco Garden Creek. This is indicated by the fact that the early Wisconsin(?) till not only mantles the bench but extends over its edge and continues below the flood plain. The evidence indicates, therefore, that the Little Missouri River continued to flow north for a long time after the diversion path of the Missouri had been established.

CRANE EROSION CYCLE

After acquiring its easterly path across the upland surface, the Missouri River cut down to the level of the Crane terrace. That the downcutting was fairly continuous is suggested by the absence of intervening terraces. The river reached grade about 40 feet above the present low-water level, at approximately the minimum height of the edge of the rock bench veneered by the Crane Creek gravel. The downcutting during this cycle of erosion thus extended about 200 feet below the ancestral valley floors and more than 400 feet below the ancestral divides. Whether the erosion was accomplished entirely in Sangamon time, as suggested by Warren for South Dakota, or whether it represents erosion during Yarmouth time, or both, is not definitely known. If, however, the writer's interpretation of the Pleistocene sequence is correct, the necessity of providing for another glaciation (Smoke Creek) and another erosion cycle (deep-trench) prior to the advent of the early Wisconsin(?) ice seems to require that the Crane erosion cycle be no younger than the Yarmouth interglacial stage.

The Missouri and Yellowstone River valleys were opened out to widths of 3 to 4 miles at the Crane level. The greatest widths along the Missouri were attained in areas where the diversion course coincided with the ancestral courses of the Missouri and Yellowstone, as, for example, west of Poplar, Mont., and between the mouth of the Yellowstone River and Williston, N. Dak. (fig. 5). At this level, the Missouri River probably flowed through the Culbertson-Bainville and Hofflund channels. The channel opposite Wolf Point, Mont., was probably not yet in existence.

The Crane erosion cycle has not yet propagated itself throughout the drainage system; the headwaters of many streams are still flowing at the Missouri peneplain level.

SMOKE CREEK GLACIATION

The Crane erosion cycle was interrupted or terminated by advent of the Smoke Creek ice (Illinoian?), which is recorded by a single exposure along Smoke Creek in Montana. If the kames southwest of Bainville, Mont. (pl. 1), are the same age, then the Smoke Creek ice extended at least as far as the Missouri River.

In Smoke Creek valley, the Crane Creek gravel is overlain by sediments including lacustrine clays (fig. 13). The valley, therefore, was dammed before being overridden by the ice. This requires that the ice encroach from the southeast. Evidence will be presented shortly indicating that during advance of the early Wisconsin(?) ice, a lobe occupied the Big Muddy Creek lowland and expanded laterally to the northwest and southeast. Presumably, the Smoke Creek glacier expanded similarly.

DEEP-TRENCH EROSION CYCLE

Following the Smoke Creek (Illinoian?) glaciation, a new erosion cycle (deep-trench) was initiated in Sangamon time (interglacial between the Illinoian and Wisconsin stages). Whether the erosion was initiated in response to regional tilting, climatic change, or a combination of both is unknown. In consequence of this rejuvenation, the valleys of the master streams, and of the tributaries for varying distances upstream, were incised 140 feet or more below the Crane Creek gravel or more than 100 feet below present low-water river level. The Crane valley floor now appears as the Crane terrace. The evidence for the depth of erosion consists of drilling records, some of which are recorded in figure 37.

Drilling programs for dam sites supply the most significant subsurface data, because the drill holes are usually closely spaced and numerous. Prior to final selection of a damsite, exploratory drilling may be carried on at other sites. Although the density of drilling at these rejected sites is less than at the selected site, much additional information is made available.

At Fort Peck Dam, about 30 miles west of the map area, the average maximum depth to bedrock revealed in 11 lines of drill holes was 142 feet. Actual depths ranged from 115 to 175 feet. The variations in depth could be due to localized river or glacial scour, to possible misidentification of bedrock, or to the possibility that the drill hole spacing in some profiles failed to reveal the deepest part of the bedrock valley. The average of 142 feet compares with depths of 103 and 120

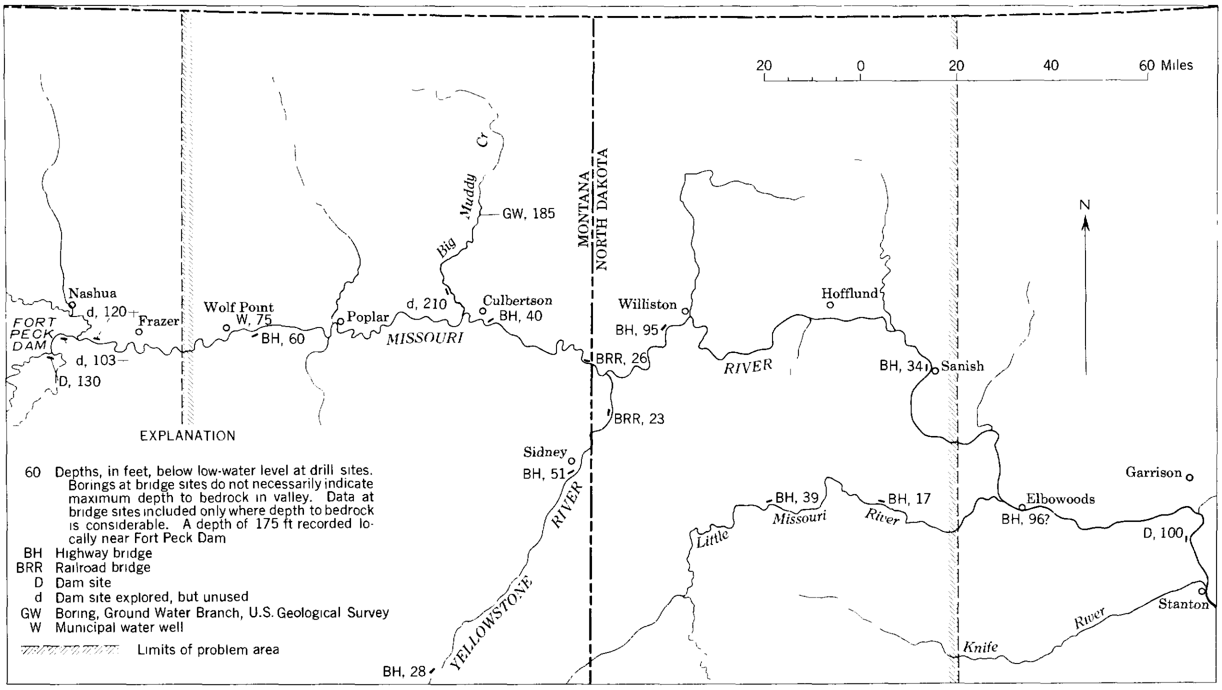


FIGURE 37.—Depth to bedrock in major valleys of the region. Information is from the Corps of Engineers, Bureau of Reclamation, State Highway Commissions of Montana and North Dakota, Ground Water Branch, U.S. Geological Survey, Great Northern Railway, local municipalities, and private individuals.

feet revealed at drilling sites 6 and 12 miles downstream, respectively.

No other precise information on maximum depths to the rock floor is available for 300 miles downstream or 80 miles downvalley from the eastern edge of the map area. At Garrison Dam site the bedrock floor is about 100 feet below present river level.

Information obtained in drilling for bridge sites is unfortunately of limited value because the line of drill holes is largely confined to the river crossing regardless of the width of the valley floor, and drill holes generally are driven only far enough to reveal sound foundations for the bridge piers. At most or all the bridge sites, the deepest part of the bedrock floor has probably not been found, as indicated by the slope of the floor revealed in the drilling. The depths of 95 feet at Lewis and Clark Bridge over the Missouri River near Williston, and of 96 feet at Elbowoods, N. Dak., a short distance east of the map area, probably indicate that the present position of the Missouri on its flood plain at these sites is not far removed from the position of the streamline of the bedrock valley below.

The only other place where comprehensive drilling has been carried on is in Big Muddy Creek valley several miles above its junction with the Missouri. Here a bedrock depth of 210 feet below low-water level was revealed. The great depth may be due to glacial scour, a possibility which is discussed later in more detail.

The writer knows of no deep drill records along the Yellowstone River valley within or close to the problem area. Shallow drilling for the bridge sites at Glendive and Sidney, Mont., revealed bedrock at depths of 28 and 51 feet, respectively (fig. 37). However, the deepest parts of the buried rock floor may not have been found.

The only evidence for bedrock depth in the Little Missouri River valley is also at bridge sites. A depth of 39 feet was found at the bridge along U.S. Highway 85, due south of Watford City, N. Dak., and a depth of 17 feet was found at the bridge along North Dakota Route 22. Greater depths may prevail away from the river. Engineers of the Northern Pacific Railway drilled a deep hole at Medora, N. Dak., about 10 miles south of the map area. The upper 30 feet of the core, above probable bedrock, is listed as sand and may represent valley fill. This one hole, however, does not exclude the possibility of greater depths to bedrock elsewhere across the valley.

Deep holes drilled in search of water by municipalities and individuals are generally too poorly recorded, in terms of the deposits penetrated, to supply precise information on the level of the bedrock floor. Colton (1951, p. 37) remarks that various wells in and around the towns of Wolf Point and Poplar, Mont., penetrate more than 100 feet of alluvium below the flood plains. Shallow holes, which stop in glacial deposits, supply only a minimum estimate of the depth to bedrock. None of these have been recorded in figure 37, because more significant data are available nearby and are already plotted.

Thus, at the two widely distant damsites where comprehensive drilling has been carried on, the buried rock floor of the Missouri River valley is 100 feet or more below present low-water level, or 120 feet or more below the present flood plain. At two bridge sites in the intervening area, depths of 95 and 96 feet to bedrock were recorded, but these are probably not maximum depths. It seems safe to conclude, therefore, that the Missouri River, and presumably its principal tributaries, eroded inner trenches to a depth of 140 feet below the lowest parts of the Crane terrace prior to deposition of the oldest of the valley fills. In the absence of evidence to the contrary, it is assumed that the oldest of the valley fills is early Wisconsin(?) in age. As additional information on the deep fill and the bedrock profile becomes available, it may develop that the deep trench was only partly eroded prior to the early Wisconsin(?) glaciation and was completed subsequently. Present drill data, however, give no evidence of such an intervening erosion cycle.

The Missouri River, during the deep-trench stage, presumably followed the Culbertson-Bainville and Hofflund channels (pl. 2 and fig. 5). The Culbertson-Bainville channel is 20 miles long, 1½ to 3 miles wide, and it is separated from the present Missouri trench by a plateau remnant that is about 50 square miles in area. The Hofflund channel is 6 to 8 miles long, 2 to 3 miles wide, and it is separated from the Missouri trench by a much smaller plateau remnant. Unfortunately, there are no deep well records along these channels, and the depth to bedrock is unknown.[12] Early Wisconsin(?) till, however, is exposed on both flanks of these channels and apparently extends below the alluvium in the center.

MINOR DRAINAGE CHANGES PRIOR TO EARLY WISCONSIN(?) GLACIATION

In addition to the major drainage changes which preceded the early Wisconsin(?) glaciation, there were a number of minor ones. Four of these, involving the Poplar River, Big Muddy Creek, Little Muddy Creek, and the White Earth River, will be considered in the following pages. The histories of the Poplar River and Big Muddy Creek are considered together inasmuch as both are related to the history of the ancestral Missouri River.

[12] Mr. Asbury, chief of investigations of the Wolf Point office of the Bureau of Reclamation, reports that only auger holes have been bored in the Culbertson-Bainville channel and that these went to depths of only 24 ft or less.

POPLAR RIVER AND BIG MUDDY CREEK VALLEYS

At the level of the Missouri Plateau peneplain, the Poplar River probably entered the ancestral Missouri River directly; the southwesterly-trending lower course had not yet been acquired (pl. 8). At this level, too, Hardscrabble Creek probably continued north across the present site of the Missouri River valley and along the southeasterly-trending lower course of Big Muddy Creek into the ancestral valley near Manning Lake. Farther west, East Charley Creek behaved similarly, entering the ancestral valley by way of a wind gap northwest of Brockton, Mont. This wind gap was later used as an outlet for glacial lake waters.

The relations of Big Muddy Creek to the ancestral Missouri River valley are more complex than those of the Poplar River and are more significant with regard to geomorphic history. Hence, Big Muddy Creek is discussed in greater detail. Available data indicate that Big Muddy Creek valley, for the greater part of its length, probably originated as an ice-marginal channel in a glacial age prior to the early Wisconsin(?).

The geology of Big Muddy Creek valley in the vicinity of the town of Antelope, Mont., has been described by Witkind (1959, p. 28–31). The town is located in a north-south swale (pls. 1 and 2), which is the surface expression of a drift-buried valley. The buried valley is continuous in trend with the present valley to the north and south and clearly represents a segment of the original valley from which Big Muddy Creek was displaced to the west.

Glacial gravel is exposed beneath the early Wisconsin(?) till at the north and south ends of the buried segment and is presumably continuous under the till cover. Glacial gravels within the present diversion course, however, are not capped by till and are believed to be younger than the till. Witkind interprets the history as follows:

1. Big Muddy Valley [in Montana north of Medicine Lake] was formed as an ice-marginal channel probably during withdrawal of a pre-early Wisconsin(?) ice sheet.

2. The early Wisconsin(?) ice invaded the area. During a halt in withdrawal of this ice, a lobe extended a short distance west of the town of Antelope, and Big Muddy Creek carved the new trench around the end of this lobe. This new channel was maintained after withdrawal of the ice.

Witkind concludes that the buried valley could only have been formed in pre-early Wisconsin(?) time or during a halt in advance of the early Wisconsin(?) ice because of the presence of early Wisconsin(?) till in the buried valley. He does not believe that the buried segment could have been formed during recession of the early Wisconsin(?) ice because no other ice advanced to this locality that could be held responsible for the displacement of Big Muddy Creek to the west. However, it could be argued that a temporary advance during recession of the early Wisconsin (?) ice caused the displacement. The "temporary advance," however, would have had to have been regional, because early Wisconsin(?) till occurs down to and below flood-plain level all along Big Muddy Creek valley, yet Big Muddy Creek underwent only one minor displacement.

Obviously, the best evidence for a pre-early Wisconsin(?) age would be the presence, within the original valley, of the nonglacial, pre-early Wisconsin(?) Crane Creek gravel. A thickness of 11 to 15 feet of this gravel is exposed on the west side of Big Muddy valley across and slightly downstream from the town of Homestead, Mont. A pebble analysis of the gravel revealed only one pebble of granite and none of limestone or dolomite. In contrast, these rock types are abundant in glacial gravels. The gravel is overlain by early Wisconsin(?) till, which extends below the flood plain. The evidence suggests that Big Muddy valley had been in existence during at least two cycles of erosion (Crane and deep-trench) prior to advance of the early Wisconsin(?) ice.

Unfortunately, the exposure of Crane Creek gravel is located within the area of intersection of the northeastward-trending ancestral valley of the Missouri River and the present south-trending valley of Big Muddy Creek. This raises the important question of whether the gravel was deposited before or after Big Muddy valley came into existence. The gravel could have been deposited before Big Muddy valley came into existence, either by a stream following the ancestral valley to the northeast at the level of the Crane terrace, or by a tributary entering the trenched ancestral valley. Or, it could have been deposited by Big Muddy Creek, already established in its south-flowing course. If the gravel was deposited by a northeast-flowing stream before Big Muddy valley came into existence, then Big Muddy valley postdates the Crane cycle. If the gravel was deposited by the south-flowing Big Muddy, the valley antedates the Crane cycle. The gravel exposure merits further study for imbrication to determine the direction of flow of the depositing stream. Eventually, it may be possible to show that, outside the area of intersection, the Crane Creek gravel is restricted either to Big Muddy valley or to the ancestral Missouri River valley, in which case the problem can be solved.

Although the evidence is not conclusive, the writer is inclined to agree with Witkind that the valley of Big Muddy Creek was initiated in pre-early Wisconsin(?) time, possibly during the Smoke Creek glaciation (Illinoian?). However, the mode of origin north of the ancestral valley of the Missouri was apparently different from that to the south. These contrasted modes of origin are considered below. Between Plentywood and Medicine Lake, Mont. (fig. 38), Big Muddy Creek follows a path that is extremely barbed in relation to the course of the ancestral Missouri. Actually, this portion of the path of Big Muddy Creek is almost at right angles to the paths of other tributaries of the ancestral valley, such as Smoke and Wolf Creeks. North of Plentywood, and for many miles into Canada, the valley is not occupied by through drainage. The situation is described by Rose (1916, p. 18–19) as follows:

> There are in the area [southern Saskatchewan] large coulees or valleys which at first sight seem to mark the courses of large rivers, but which contain no through-flowing streams. These old channels can be traced for many miles and a former drainage system worked out. The bottoms of the valleys have become silted up and are now occupied by scattered saline lakes, by streams of greatly diminished size, or for short distances by local streams from nearby springs. The water of these streams and springs sinks into the silt or empties into the saline lakes and evaporates.
>
> A good example of one of these old valleys is Big Muddy valley. It is tributary to the Missouri river and in its lower course carries a small stream, Big Muddy Creek. The head of this creek barely reaches the International Boundary, but the large valley can be traced for 100 miles farther. It is cut to an average depth of 250 feet below the prairie level and has a width varying from 1 to 1½ miles. A line of terraces about 200 feet above the valley bottom marks a break between a mature upper valley and a main younger valley and indicates that there were at least two stages in its development. The valley crosses into Canada in Tp. 1, range 22, W. 2nd mer. Springs in that vicinity form the headwaters of Big Muddy creek. After crossing a slight divide, the first of a series of saline lakes, Big Muddy lake is met. This lake occupies the old valley bottom for a distance of 19 miles and for 15 miles is less than 1 mile wide. In Tp. 4, range 25, W. 2nd mer., Big Muddy valley divides into two branches of about equal size. The north branch is occupied along its course by Willowbunch lake, 21 miles long, and by Lake of the Rivers, 25 miles long. It heads in the flat country about Lakes Johnston and Chaplin, and doubtless at one time drained that area. The south branch passes close to the town of Willowbunch, along a number of small saline lakes, and finally through Twelvemile lake, a lake over 15 miles long, but in a few places more than one mile wide. It heads in the country immediately north of Wood Mountain plateau which is now occupied by the headwaters of Wood river.

The northwesterly trend of the Big Muddy channel is at right angles to the northeasterly regional slope. It seems reasonable to suppose that the valley is an ice-marginal feature formed in the same way as the

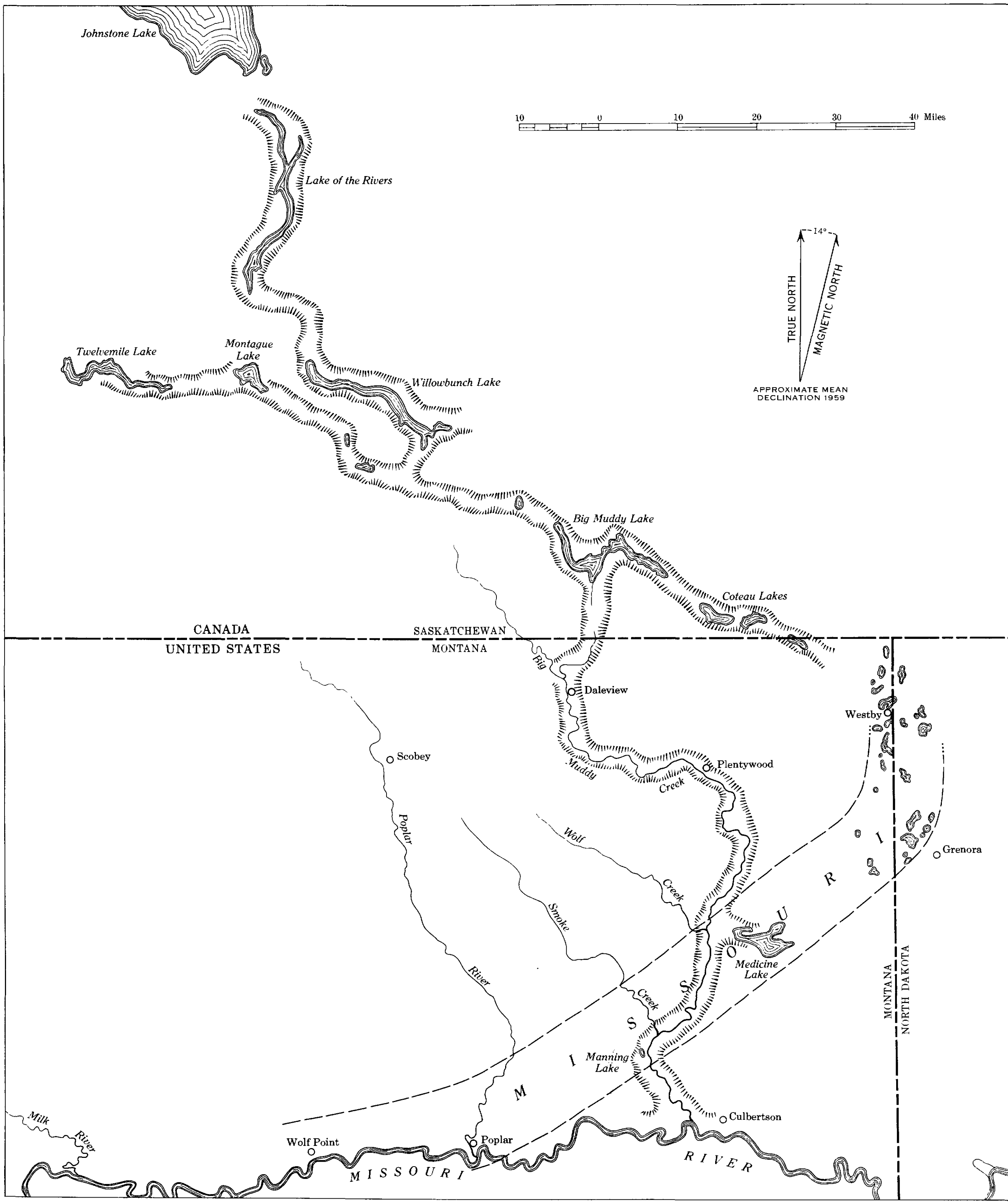


FIGURE 38.—Relation of Big Muddy Creek valley to the valley of the ancestral Missouri River and to the prominent glacial channels to the north.

Missouri trench and other diversion channels to the south. Big Muddy channel, however, is not necessarily the same age throughout. Where occupied by early Wisconsin(?) till, the channel may antedate the early Wisconsin(?) ice or may be early Wisconsin(?) in age. Other parts may have formed in later intervals. That the channel may have been followed by glacial and (or) interglacial waters on more than one occasion is suggested by the two-level profile described by Rose. The presence, near Antelope, Mont., of glacial gravels immediately under early Wisconsin(?) till suggests that the channel was used, at least in part, in early Wisconsin(?) time. The Montana part of the channel was also used by Mankato glacial waters, as indicated by the presence of late glacial gravels which can be traced back to the coteau morainal belt (pl. 1).

The lower course of Big Muddy Creek trends southeast from Manning Lake, within the ancestral valley of the Missouri, to the present Missouri trench (fig. 38). As suggested earlier, this lower course probably marks the former path of northwestward-flowing Hardscrabble Creek into the ancestral Missouri. The reversal of drainage in this lower course may have been effected as follows:

After the ancestral Missouri valley was abandoned, the valley floor still drained to the northeast at least as far as the site of Manning Lake. This is indicated by the continuous descent of the valley floor in that direction and the preserved path of a former stream between the Poplar River valley and Manning Lake (pl. 1). The course of the above-mentioned stream begins in a col in the divide between the ancestral Missouri valley and the Poplar valley, suggesting that it was the Poplar River that flowed through the col and down the slope of the ancestral valley.

Big Muddy Creek valley between the present Missouri trench and Manning Lake was at that time occupied by Hardscrabble Creek, whose valley floor presumably sloped to the northwest. If, as was true for the early Wisconsin(?) ice, a proglacial lake was impounded in the ancestral Missouri valley, this lake would have had an arm extending southeastward up Hardscrabble Creek valley to the present Missouri trench. Erosion by the escaping lake waters may, therefore, have caused a northward migration of the lake outlet and a reversal of gradient of the valley between the outlet and the Missouri. If the lake existed for only a relatively short time, the reversal of the gradient may have been completed by headward erosion on the part of a stream occupying the southeast-draining lower valley.

The Poplar River no longer enters the ancestral valley of the Missouri; it follows a southwesterly course into the present Missouri. The abrupt change in course may be due to piracy on the part of a vigorous stream eroding headward from the present Missouri. In any event, the present lower courses of the Poplar River and Big Muddy Creek were established long before arrival of the early Wisconsin(?) ice, as indicated by the presence of early Wisconsin(?) till down to and below flood-plain level in both these streams. The presence of Crane Creek gravel along the lower course of the Poplar River suggests that this stream, at least, was diverted from the ancestral valley of the Missouri as early as Yarmouth(?) time.

LITTLE MUDDY CREEK VALLEY

Little Muddy Creek, in North Dakota, flows south in a valley trenched below the north-sloping ancestral valley of the Yellowstone River (pl. 8). The presence of early Wisconsin(?) till at or near creek level, and the occurrence below the till of glacial gravels with imbrication indicating a south-flowing current, indicate that Little Muddy Creek was already established in its present course prior to advance of the early Wisconsin(?) ice. The conclusion is confirmed by drill records of the U.S. Bureau of Reclamation and private contractors. Drilling reveals that the sands and gravel that underlie the early Wisconsin(?) till at the junction of Little Muddy and the Missouri valleys extend well below flood-plain level.

WHITE EARTH RIVER VALLEY

At least in its lower course, the present trench of the White Earth River is incised below a broad upland sag which probably represents an ancestral peneplain valley. That the direction of flow had already been changed from north to south prior to early Wisconsin(?) time is suggested, but not proved, by analogy with the histories of Poplar, Big Muddy, and Little Muddy valleys. However, because the White Earth valley is within the border of the Mankato drift, it is impossible to determine whether the till exposures on the valley floor are early Wisconsin(?) or Mankato. The tills are lithologically similar. It seems reasonable to assume, however, that reversal of the gradient of the White Earth valley began soon after the diversion (Kansan?) which created the present Missouri River.

An old till-covered terrace of glacial gravel at the town of White Earth, N. Dak. (fig. 39, left side) indicates that the White Earth River valley dates back at least one glaciation prior to the Mankato. The gravel is heavily oxidized and thoroughly leached; the overlying till(?) is unleached. Gravels of Mankato age form a lower terrace in the valley and are nowhere till covered. The weathered, till-covered

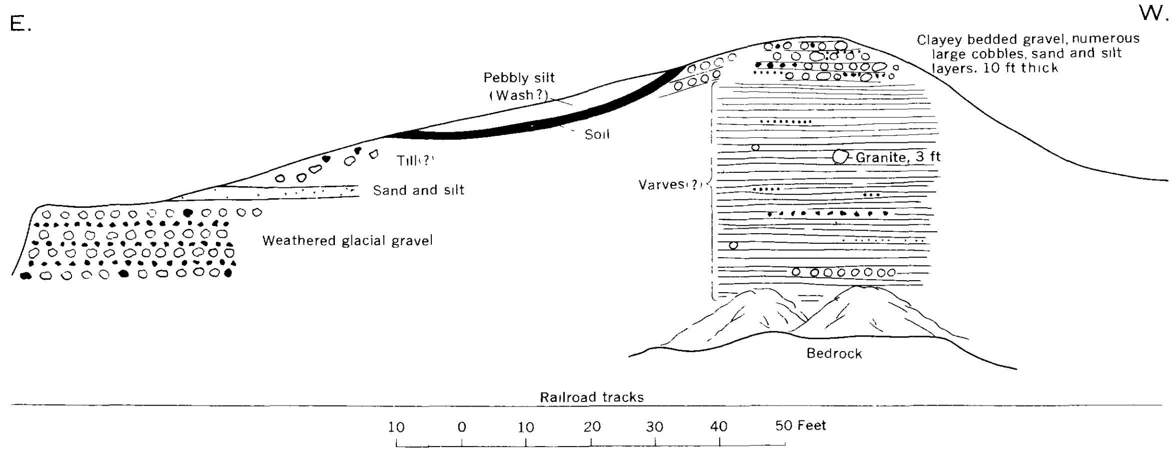


FIGURE 39.—South face of railroad cut at north edge of the town of White Earth, Mountrail County, N. Dak. Height of exposure is about 60 feet. The weathered (and leached) glacial gravel to the left is believed to be appreciably older than both the unleached till(?) above and the thick succession of sediments to the right. The varved(?) clays were probably deposited in a local lake.

glacial gravels are interpreted as being outwash deposited by pre-Mankato melt waters.

The complicated network of present and past drainageways between Tobacco Garden Creek and the White Earth River cannot be deciphered with assurance on the basis of the few data available. A few comments on the possible significance of this channel network, however, may be in order.

Alpha[13] suggested that the White Earth River once flowed westward through the channel between the White Earth valley and Hofflund, N. Dak. (fig. 35).[14] He believes that the Missouri River at that time occupied the curved channel between the mouth of Tobacco Garden Creek and Hofflund. Alpha made no attempt to explain the extremely barbed relationship of the White Earth River to the Missouri.

The White Earth-Hofflund channel is about a mile wide, is several hundred feet deep, hangs 75 to 100 feet above the floor of the White Earth valley, slopes westward, is filled with drift (Wood, 1903), and contains a prominent terrace of pebbly glacial gravel. The channel was almost certainly used by glacial waters either prior to development of the lower few miles of the White Earth valley, or after development of the lower course of the White Earth valley at a time when that valley was blocked by ice. The position of the Mankato drift border (pl. 1) suggests that the lower course of the White Earth valley was blocked by Mankato ice and that melt waters followed the channel to the Missouri.

As a working hypothesis, the writer suggests the following drainage history for this area. The ancestral Little Missouri River crossed the site of the present trench of the Missouri at the level of the Missouri Plateau peneplain and followed the line of the White Earth valley for an unknown distance to the north. For some time after the glacial diversion (Kansan?) which created the present Missouri River, the river followed the Hofflund-White Earth channel and the lower course of White Earth valley. The Missouri persisted in this course during several hundred feet of downcutting. During this interval, the gradient of what is now White Earth valley, but which was previously the valley of the northward-flowing ancestral Little Missouri River, became reversed. The reversal may have been effected in part by the melt waters of the ice which diverted the Missouri and in part by headward erosion of a stream draining south into the deepening Missouri trench. Displacement of the Missouri from the Hofflund-White Earth channel to its present trench to the south probably took place in early Wisconsin(?) time, as suggested by the presence of till of this age in the channel. However, the presence of the short channel extending south from Hofflund suggests that the diversion took place in two stages, which may or may not have been confined to early Wisconsin(?) time. The lower part of White Earth valley was blocked by Mankato ice, and melt waters flowed west through the channel system in a roundabout course to the Missouri.

MISCELLANEOUS STREAMS

Possible ancestral paths of several small streams are indicated in plate 8. These are based on (1) the assumption that prior to the development of the present course of the Missouri all drainage was into the ancestral master streams; (2) the presence of wind gaps, such as those which mark the former entrances

[13] Alpha, A. G., 1935, Geology and ground-water resources of Burke, Divide, Mountrail, and Williams Counties in North Dakota: N. Dak. Univ., unpub. thesis, p. 13.

[14] Wood (1903, p. 104) had made a similar suggestion much earlier, but he did not identify the stream as the White Earth River.

of the Poplar River and East Charley Creek into the ancestral Missouri River valley; (3) the alinement of opposed tributaries across the present valleys of the master streams as, for example, the alinement of Hardscrabble Creek and the lower, anomalous course of Big Muddy Creek across the Missouri valley, and the alinement of Bennett Creek and the lower course of Redwing Creek across the Little Missouri valley; and (4) the alinement of tributary streams and parts of the diversion paths of the Missouri and Little Missouri as, for example, the alinement of Charlie Bob and Bear Creeks with contiguous parts of the valley of the Little Missouri. The diversion courses of the Missouri and Little Missouri were probably determined in part by such favorably disposed tributaries.

SUMMARY OF DRAINAGE HISTORY

The ancestral master drainage of the Missouri Plateau peneplain was to the northeast (pl. 8). All three major streams of the region, the Missouri, Yellowstone, and Little Missouri, flowed across the site of the present trench of the Missouri, possibly into Hudson Bay drainage. The valley of the ancestral Missouri departs from the present valley at Poplar, Mont., and strikes northeastward across the upland to at least as far as Westby in extreme northeastern Montana. The course of the ancestral Missouri beyond Westby is unknown.

The course of the ancestral Yellowstone River coincides with that of the present Missouri between Buford and Williston, N. Dak. At Williston it turned north along Little Muddy Creek valley and crossed the coteau belt to Crosby, N. Dak. Its course beyond Crosby is unknown.

The ancestral Little Missouri River crossed the site of the present Missouri valley and followed the course of the White Earth valley for an unknown distance to the north. Somewhere north of the present Missouri trench, the ancestral Little Missouri veered again to the northeast along one of several possible paths across the coteau belt.

The ancestral northeasterly courses were blocked by Kansan(?) ice to establish the present course of the Missouri. The route followed by the impounded waters was influenced in part by favorably oriented stretches of the ancestral valleys and in part by favorably disposed tributary valleys. The Missouri, at the time of the diversion and for a long time afterward, followed the now-abandoned Culbertson-Bainville and Hofflund channels.

Reversal of the original northerly courses of the larger tributaries of the ancestral rivers was probably completed during the Crane cycle of erosion. The reversals may have been started by Kansan(?) melt waters and completed by headward erosion from the deepening Missouri trench.

EARLY WISCONSIN(?) GLACIATION

Except in the area of the later drifts, the present topography differs little from that overridden by the advancing early Wisconsin(?) ice. The drainage pattern was virtually as it is now. The notable exceptions were that Little Missouri River still flowed north in the channel now occupied in part by Tobacco Garden Greek, and the Missouri River still occupied the Culbertson-Bainville, Hofflund, and Sanish channels (pl. 2). The major valleys and the lower courses of their tributaries, however, were more than 100 feet deeper than they are now.

The early Wisconsin(?) ice presumably came from the northeast. The drift is rich in limestone and dolomite derived from the formations which flank the Canadian Shield. The broadest areas of these rocks are around Lake Winnipeg in Manitoba. With one exception, glacial striations were not observed in the present map area. Rocks lithologically suited to the preservation of glacial markings are scarce in the soft Cretaceous and Tertiary deposits. In addition, the glacial drift forms an almost unbroken veneer on the rocks below. Clifford Kaye of the U.S. Geological Survey discovered a set of striations on sandstone in a roadside ditch 4½ miles north of Culbertson, Mont., and half a mile west of Montana Highway 16. The striae were on a block about 10 feet across and only slightly shifted from the parent outcrop. Allowing for the shift, the original strike was probably between N. 20° and 30° E. This, however, probably records only the direction of movement in this restricted area as determined by topographic influences.

ADVANCE OF THE ICE

On entering the map area, a lobe of ice apparently pushed southwestward up the ancestral valley of the Missouri, below the floor of which the present southward-flowing drainage was already established. This lobe, which will be referred to as the Poplar lobe, is outlined by marginal channels which will be discussed shortly. It seems reasonable to suppose that a similar lobe followed the trend of the ancestral Yellowstone valley, north of Williston, below whose floor Little Muddy Creek was trenched.

When the expanding Poplar lobe reached the vicinity of Manning Lake, part was deflected southeastward down the lower course of Big Muddy valley. Erosion by this deflected ice stream may have contributed to the great depth to bedrock in this stretch of valley.

A comprehensive drilling program by the Bureau of Reclamation in connection with a possible damsite in Big Muddy valley several miles above its mouth revealed a bedrock depth of 210 feet below low-water level. Gravel is present in one core at a depth of 203 feet. Average size of the well-rounded pebbles is less than one-half inch, although pebbles over 1 inch are included. The gravel consists of quartzite (43 percent), chert (21 percent), silicified volcanic rocks (11 percent), granite and related rocks (9 percent), quartz (6 percent), limestone and dolomite (4 percent), and miscellaneous types (6 percent). None of the 84 pebbles examined were striated. However, the presence of granite, limestone, and dolomite indicates that, although the gravel itself may not be outwash, the depositing stream had access to previously deposited drift.

As the front of the Poplar lobe moved southwest from Manning Lake, still following the swale marking the ancestral Missouri valley, it must necessarily have impounded a lake against the rising slope. This lake, and its counterpart, during recession of the ice, is herein referred to as glacial Lake Brockton (pl. 1). The lake probably first drained westward into the Poplar River by way of the wind gap which marks the former point of entry of the Poplar River into ancestral Missouri valley. Subsequently, when this outlet was blocked by the advancing ice, the lake drained by way of the wind gap through the eastern divide just northwest of Brockton, Mont. This wind gap was the former point of entry of Charley Creek into the ancestral valley.

As the Poplar lobe moved southwestward, it also expanded laterally up the gentle flanks of the ancestral Missouri valley. Some of the ice-marginal channels which scar the slopes are occupied by till. The presence of the till may mean that some of the channels antedate the early Wisconsin(?) ice, or they were carved during halts in advance of this ice or during halts in retreat followed by readvance. When the western edge of the Poplar lobe reached the foot of the Flaxville Plateaus, at an elevation of 2,350 to 2,400 feet, further lateral expansion was prevented for a long period of time. The blocked waters of the long, southeast-draining streams, such as Smoke and Wolf Creeks, eroded trenches across the intervening divides to form the long deep channel extending from Big Muddy valley near Archer, Mont., southwestward to the lower course of the Poplar River (pl. 2). When the snout of the Poplar lobe submerged the lower course of Poplar valley, the Poplar River in turn was impounded, and its waters, together with those coming from the northeast, escaped into Boxelder Creek by way of a deep, tortuous channel cut in part through a plateau promontory.

Simultaneously, the Poplar lobe was expanding eastward. There were no steep slopes to retard expansion in this direction, nor were there large streams to be diverted along the ice margin. As a result, there are no channels of the magnitude of those to the west. Faint channels, some of which are much more clearly visible in aerial photographs than on the ground, furrow the long slope which rises gently from Medicine Lake to the north-south drainage divide on the Montana-North Dakota line. The divide descends to the north into the valley of Cottonwood Creek (pl. 1). The slope here is steeper than it is to the west, and the deep closely spaced glacial channels which encircle the north end of the upland indicate slow movement of the ice front. The channels are as much as a quarter of a mile wide and 150 feet deep. Maximum channel cutting may have taken place during recession of the ice.

The eastern divide, topped by Bull Butte (2,530 feet), is about the same elevation as the foot of the Flaxville Plateaus to the west. It seems likely, therefore, that the lateral expanding margin of the Poplar lobe reached the crest of the eastern divide at about the same time as it reached the foot of the plateaus. Prior to this time, the Bull Butte divide undoubtedly separated the Poplar lobe from the Little Muddy lobe in Little Muddy Creek valley to the east.

The slope descending eastward from the Bull Butte divide is very gentle, steepening only slightly as it approaches Little Muddy valley, and it is not drained by large streams. As a result there are no deep ice-marginal channels. It has not been possible to distinguish, within the area of the later middle Wisconsin(?) drift, older channels from those formed by the middle Wisconsin(?) ice. Assumptions regarding expansion of the Little Muddy lobe of the early Wisconsin(?) ice are therefore based on the inferred topographic influence on the advancing ice front.

During expansion of the frontal lobes of the early Wisconsin(?) ice, the Big Muddy-Little Muddy and Little Muddy-White Earth divides, being hundreds of feet lower than the Flaxville Plateaus west of the Poplar lobe, were probably submerged before the plateaus. The western margin of the Poplar lobe probably remained fixed at the steep front of the Flaxville Plateaus while elsewhere the ice thickened hundreds of feet and spread in other directions. The small northeastern outliers of the Flaxville uplands, southwest of Plentywood, Mont., were probably the first plateau elements to be overtopped, the ice previously having filtered among them.

Sometime during its expansion, the ice reached and submerged the Culbertson-Bainville and the Hofflund channels, which were still being used by the Missouri River. The Missouri was shunted to the south. Part of the erosion of the present channel of the river may have occurred at this time. In the Culbertson-Bainville channel, however, a large part of the erosion apparently took place during recession of the ice.

The topography of the western half of the map area was largely responsible for the manner in which the Poplar lobe continued to expand. The Missouri River valley, in its easterly course, cuts obliquely across a 50-mile-wide, northeast-trending lowland, one side of which is formed by the Flaxville Plateaus on the northwest and the other, by the Yellowstone-Missouri divide on the southeast. As the Poplar lobe pushed southwestward and westward, it expanded laterally to the northwest and southeast. That the ice front actually expanded northwestward over the large Flaxville upland west of the Poplar River and north of the Missouri is indicated by evidence to be presented shortly. The ice surface, therefore, must have been higher to the south and southeast along the axial line of the lowland. South of the axial area, the ice surface presumably descended again toward the southern margin of the lobe where it rested against the Missouri-Yellowstone divide. The ice reached elevations of 2,500 to 2,600 feet on this divide. On the Flaxville upland, 50 miles to the northwest, the ice stood at elevations of 2,700 to 2,800 feet. A simple profile based on these two figures would obviously not give the true surface slope of the ice because of the intervening high along the lobe axis.

During its southward expansion, the ice displaced the Missouri River from its valley and crowded it southward. There are no prominent channels between the Missouri valley and the glacial limit. There is thus no reason to believe that there was any sustained halt in either advance or recession of the ice across this area.

North of the Missouri, the thickening ice crowded higher and higher up the marginal slopes of the Flaxville Plateaus until all but the highest parts were submerged (pl. 1). That the ice expanded northwest over the plateaus is indicated by the following evidence:

Deep (east-west) channels cross the southern part of the Wolf Point plateau. The channel segment between Tule Creek and East Fork of Wolf Creek (pl. 1) ranges in depth from 20 to more than 60 feet and averages about a quarter of a mile in width. Mounds of till, scattered closed depressions, and a profusion of glacial boulders are found south of the channel. Till is exposed in most of the road cuts in the area south of the channel, and a morainal area, with 25 feet of relief, occurs 2 to 3 miles to the southwest.

No till was found north of the channel, although boulders are present in decreasing numbers in that direction. Faint closed depressions were found, but these may not be of glacial origin; they may have originated by deflation, differential compaction, or other causes. Low mounds rising above the flat surface are composed of Flaxville gravel or brown loessic silts. The former may represent old river bars; the latter, loess dunes.

The presence of the channel athwart the southeastward-sloping surface is itself clear evidence that the ice expanded northward. Had it moved southward, the melt waters would have flowed down the slope to the south; no channel could have formed. The northward expansion is also indicated by the widespread occurrence of till south of the channel and its absence to the north. It should be pointed out that the gravel-capped Flaxville Plateaus are mantled by a brown, loessic deposit as much as 2 feet thick. The loess does not mantle till; however, numerous exposures reveal the Flaxville gravel directly beneath the silt.

The channel across the Wolf Point plateau descends to the west and becomes progressively deeper. Just outside the map area, where the channel is occupied by Todd Lakes, it is about 80 feet deep. Ice-marginal waters clearly moved westward across the plateau connecting lakes impounded in south-draining valleys.

The glacial boulders north of the channel were probably ice-rafted on lakes impounded in the glacially dammed valleys. Boulders are also found on the flat interstream divides, suggesting that at times the valley lakes merged along the ice front.

Expansion of the ice east of the Poplar lobe and north of the Missouri River was less eventful because of the absence of prominent topographic highs. The low divides were overtopped early and a wide uninterrupted expanse of ice resulted. As the ice crowded across the Missouri, however, it found two favorable avenues for penetration to the south, the broad clearly defined upland swale within which the Yellowstone River valley is trenched, and the valley of the ancestral Little Missouri River which was still occupied by the Little Missouri. A lobe of ice, the Yellowstone lobe, pushed its way south up Yellowstone valley, driving the impounded waters of the Yellowstone River before it. To the east, a lobe protruded along ancestral Little Missouri valley, driving the impounded waters of the Little Missouri ahead of it. This will be referred to as the Watford lobe, after Watford City which lies within the ancestral valley.

The ice eventually reached the limits shown in plate 1. These limits mark the southernmost occurrences of glacial boulders, neglecting those which may have

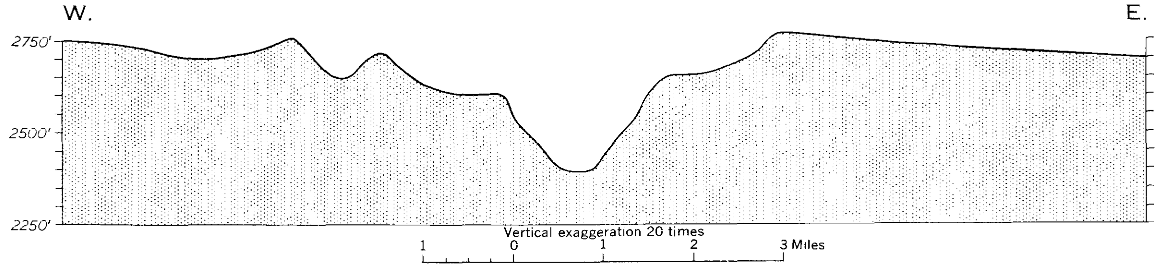


FIGURE 40.—Profile across outlet of glacial Lake Lambert. The original outlet level was probably 2,600 feet or less. The presence of ice-rafted boulders within the lake basin to an elevation of at least 2,525 feet suggests that the original outlet was at least that high. For purposes of mapping, a lake elevation of 2,550 feet has been assumed.

been rafted southward on glacial lakes. In general the limit agrees with that mapped by Alden (1932, pl. 1).

When the early Wisconsin(?) ice stood at the glacial limit, it dammed the north-draining streams. Channels marking the probable paths of water from the west into the Redwater Creek drainage basin are indicated in the western part of plate 1. Redwater Creek valley itself was occupied by glacial Lake Circle. Farther east, in the vicinity of Enid and Lambert, Mont., east-draining streams were blocked along the flank of the Yellowstone lobe to create glacial Lake Lambert. The absence of a clear channel at maximum lake level between glacial Lake Circle and Lake Lambert suggests that the lakes were confluent and controlled by a single outlet. This outlet, about 4 miles southwest of Lambert, is now a broad wind gap between the Fox Creek drainage to the north and North Fork of Burns Creek to the south.

The outlines of glacial Lake Circle and Lake Lambert shown in plate 1 are approximate. There are no recognizable shorelines. Downcutting at the outlet into North Fork of Burns Creek presumably started at a level above the present floor of the col. A profile across the gap, based on the Glendive topographic sheet but corrected for absolute altitude,[15] is given in figure 40. According to the profile, downcutting began at or below the 2,600-foot level, depending on how low this part of the divide had been eroded prior to early Wisconsin(?) time. It is this inability to fix the initial outlet level which, in the absence of undoubted shorelines, makes it impossible to map precisely the maximum extent of the lake.

Parker (1936, p. 137) also recognized the possibility of a lake in Redwater Creek valley and attributed outlying erratics to ice-rafting. The highest erratic found by the writer in the few traverses made for the purpose of fixing the lake level is at an elevation of about 2,525 feet. The erratic is a cobble of gray granite, about 6 inches in diameter, along Montana State Highway 18, about 3 miles east of Circle, Mont. As it is unlikely that the highest of the rafted boulders has been found, the shoreline of confluent Lake Circle and Lake Lambert has been mapped at an elevation of 2,550 feet (pl. 1). The configuration of the shoreline is also approximate, not only because of uncertainty regarding the exact level of the lake but also because of the lack of topographic maps for the western part of the area.

The channel connecting Lake Lambert with Lake Glendive in the Yellowstone River valley is among the most impressive in the area (pl. 1). The southeast-trending section coincides with preglacial North Fork of Burns Creek so that part of its depth is preglacial. Some idea of the amount of glaciofluvial deepening that has taken place may be gained by a comparison of gradients of this stretch of North Fork of Burns Creek with its companion streams to the south. These streams flow down the same slope, but their valleys were not occupied by glacial waters. Middle and South Fork of Burns Creek and North Fork of Thirteen Mile Creek have gradients of about 30 feet per mile for a distance of 15 miles from their heads. The present gradient of North Fork of Burns Creek in its 10-mile-long southeasterly course below the Lake Lambert outlet is only about 15 feet per mile, indicating considerable lowering of its profile. A comparison of the valley-floor elevations of these streams, at comparable distances below their headwaters, suggests that the valley of North Fork of Burns Creek was deepened 200 to 300 feet just below the outlet. The writer favors the smaller figure, because if the original level of the col between Lake Lambert and North Fork of Burns Creek had been 300 feet above the valley floor it would probably not have been the lowest escape route available.

Ten miles southeast of the outlet, North Fork of Burns Creek turns to the southwest, following a channel which is at right angles to the surface slope and which cuts across the divides between the southeast-

[15] According to Parker (1936, p. 124), the altitudes shown on the Glendive topographic sheet (scale, 1:250,000, contour interval, 50 ft) are 30 to 50 ft too high in the Lambert region. In constructing the profile, the value of the contour lines has been lowered an even 50 ft. Elevations are, therefore, accurate only within 50 ft or so.

trending streams. The channel here has precisely the same relations to the Flaxville uplands to the west as the channel at the foot of the Flaxville uplands north of the Missouri River. Both were eroded by ice-marginal waters crowded against the steep slopes of the plateaus.

The southwest-trending segment of Burns Creek channel is locally more than a mile wide and is trenched 200 feet or more below the surface at the foot of the plateaus. Burns Creek leaves this southwest-trending segment to resume a southeasterly course into the Yellowstone River near Burns. The continuation of the channel, however, can be traced southwestward across the divide between Burns Creek and Thirteen Mile Creek, where its floor is 2,375 feet in elevation (barometer). This is higher than the floor of the channel to the northeast, suggesting that this southwesternmost segment was occupied only until withdrawal of the ice exposed the southeast-draining valley now occupied by lower Burns Creek. The fact that the channel does not continue farther south probably means that it here entered Lake Glendive, an ice-impounded lake in the Yellowstone valley. At this point the channel itself is 50 to 75 feet deep, suggesting that the highest level of the lake was 2,425 to 2,450 feet. The higher figure is probably more nearly correct, as indicated by the elevation of the lake outlet across the Yellowstone valley to the east.

The outlet channel cuts across the divide between Dry Creek and Smith Creek about 7 or 8 miles east-southeast of Burns (pl. 1). The upland is approximately 2,500 feet in elevation on the east side of the channel and 2,450 feet on the west side. The channel floor is about 2,350 feet. The exact elevation of the upland at the outlet site before it was eroded by the escaping lake waters is unknown, but the depth of the channel indicates about 100 feet of downcutting. It seems reasonable to assume that the lake waters first overtopped the divide at approximately 2,450 feet. The outline of Lake Glendive shown in plate 1 was therefore mapped on the assumption that its maximum elevation was about 2,450 feet. This figure agrees very closely with that of Alden (1932, p. 82), who estimated the lake level on the basis of the elevation of boulders interpreted as having been ice-rafted. The writer agrees that, with rare exceptions, boulders are not found above about 2,450 feet in the Yellowstone River valley south of Intake, Mont.

Alden (1932, p. 82–83) reported "* * * a few small granite boulders on the edge of a high bench 10 miles west of Burns, 800 feet above the river, or 2,000 feet above sea level * * *"[16] As an alternative to the unlikely possibility that the ice overtopped this summit, Alden suggested that the boulders may have been hauled to their present site "as they were seen near other stones that had evidently been collected for the purpose of making a foundation for a building." The suggestion is reasonable; ranchers haul cobbles and boulders for long distances for various purposes, the most common of which is either for weighting down barbed-wire fences that cross depressions or for strengthening corner posts.

If the elevation of 2,450 feet is accepted as an approximation of the maximum level of Lake Glendive, the lake extended about 110 miles up the Yellowstone River valley to the vicinity of Hathaway, Mont., 20 miles south of Miles City, Mont. The map, plate 1, shows only the northern 15 miles of the lake, or about one-seventh of its linear extent.

The writer has found no undoubted shore features of Lake Glendive. This may be due in part to the reconnaissance nature of the fieldwork. The writer was reluctant to spend the amount of time necessary to visit each faint valley-side bench in order to determine whether it might be wave cut rather than structural. However, thick deposits of silts, including delicately laminated layers, are present in many places along the valley. Two of these have been described in detail in the section on "Terrace Silts." Some of the material may be lacustrine and part of the deposits of Lake Glendive.

From the outlet of Lake Glendive, the escape waters followed the ice border to the Little Missouri River. Northwest-draining tributaries of the Yellowstone River were dammed and numerous small lakes were linked by channels across intervening divides. Most of the channels are ½ to ¾ of a mile wide and several hundred feet deep, and they maintain these dimensions across the intervening divides (fig. 6). Some are flat floored; others have sloping sides consisting of slope wash and fan deposits.

The ice border between the Yellowstone River valley and the Little Missouri River (pl. 1) is not located as precisely as it is elsewhere. The ruggedness of the area and the scarcity of roads made detailed investigation difficult. It is unlikely, however, that the border is anywhere in error by more than a few miles. Near the ice border, some of the higher buttes and mesas probably projected above the surface of the ice. It will be noted in plate 1 that the ice-marginal channel did not everywhere hug the ice front. Locally, the most favorable paths for the eastward-flowing waters lay as much as 6 miles to the south.

East of the Yellowstone lobe, the ice front advanced southward against the drainage of the ancestral Little Missouri which, until now, still flowed north into the

[16] The 2,000-ft figure is obviously a slip. The river itself is at 2,000 ft. The high bench referred to is about 2,800 ft in elevation.

Missouri. A lake must have been imponded in this valley during both advance and retreat of the ice.

The lake which occupied the valley at the glacial maximum will be referred to as glacial Lake Mikkelson, after a settlement in the valley near the southern border of the map area.

Lake Mikkelson may have been several hundred feet deep at the ice front. The floor of the ancestral valley at its point of departure from the present trench stands 100 to 200 feet above present river level. Inasmuch as the upland, across which the waters of Lake Mikkelson escaped eastward, stands 350 to 450 feet above the ancestral floor, Lake Mikkelson may have been that deep. Actually the level of the lake was probably lower, because the outlet waters undoubtedly followed drainageways below the level of the upland. The escaping waters picked their way eastward, in some places hugging the ice front and elsewhere wandering some distance afield.

The fact that the Little Missouri River did not forsake its newly acquired eastward course on retreat of the early Wisconsin(?) ice indicates that it had eroded its newly acquired course below the level of the ancestral valley or down to within 100 to 200 feet of present river level. This indicates that the ice front remained stationary for a long time at the site of the present trench of the Little Missouri River. During this downcutting, the level of Lake Mikkelson steadily dropped. When erosion reached the level of the floor of the ancestral valley, Lake Mikkelson disappeared.

The present valley of the Little Missouri River was probably eroded 50 to 150 feet below the level of the floor of the ancestral valley while the early Wisconsin(?) ice stood at, or close to, the glacial limit. This is suggested by the fact the ice-marginal channel occupied by Hay Creek, which enters the Little Missouri from the west, hangs only 40 to 50 feet above the present river level. The channel could not have been eroded so low by glacial waters unless the ice-marginal trench of the Little Missouri was also excavated to that depth. Post-early Wisconsin(?) lowering of Hay Creek channel by tiny Hay Creek is unlikely because of the great dimensions of the broad, flat-floored channel and the fact that it retains these great dimensions across the divide separating Hay Creek from Bennie Pierre Creek to the west.

The course of the outlet waters of Lake Mikkelson was undoubtedly influenced by the trends of valleys incised in the upland. Some of these trends are suggested in plate 8. Near the western boundary of the Fort Berthold Indian Reservation, about 15 miles from the eastern border of the map area, the early Wisconsin(?) ice crossed the present site of the Little Missouri River valley and deflected the escape waters of Lake Mikkelson southward along a course normal to the gentle eastward slope of the upland.

WANING OF THE ICE

As the front of the early Wisconsin(?) ice withdrew northward, Lake Circle, Lake Lambert, and Lake Glendive expanded in that direction until lower outlets were exposed. Lake Mikkelson, however, had already disappeared, having been drained by downcutting of its eastern outlet prior to or during early stages of withdrawal of the ice. The paths followed by the ice-marginal waters between the lakes also changed during glacial recession. Some of these changes will be briefly considered.

As the Yellowstone lobe withdrew from its maximum point of advance near Intake, Mont., the marginal waters on the west side of the lobe found more direct routes into Lake Glendive. Thus, when the ice front had withdrawn to the vicinity of Burns, the marginal waters abandoned the southwesternmost segment of the marginal channel in favor of the present lower course of Burns Creek.

Lake Lambert continued to drain through Burns Creek channel and into Lake Glendive throughout the time that the level of Lake Glendive dropped from 2,450 feet to possibly 2,200 feet or lower. This is suggested by the fact that the floor of Burns Creek channel, which heads in the outlet of Lake Lambert, descends to an elevation of 2,200 to 2,250 feet near the point of entry into Lake Glendive.

During wasting of the ice, Lake Lambert expanded not only northward, but eastward. As the margin of the Yellowstone lobe receded downslope, it exposed lower escape routes for the waters of Lake Lambert. The outlet south of Lambert, Mont., was abandoned in favor of the channels between Dunlap and Fox Creeks west of Crane, Mont. These also led into Burns Creek channel but by a more indirect route across the low divide at the head of Dunlap Creek.

The drop in level of Lake Glendive from 2,450 feet to 2,200 to 2,250 feet implies a lowering of the outlet of Lake Glendive on the east side of the Yellowstone River valley or else that lower outlets became available. The depth of the outlet between Dry and Smith Creeks indicates 100 to 150 feet of deepening during use of the outlet, from approximately 2,450 feet to between 2,300 and 2,350 feet. Even assuming a 30- to 50-foot error in elevations on the topographic map, the present floor of the outlet is above the designated level of Lake Glendive at the time of abandonment of the Burns Creek channel. In other words, the

Burns Creek outlet of Lake Lambert continued in use after abandonment of the high-level outlet of Lake Glendive.

The ice-marginal channel east of Lake Glendive crosses successive divides before reaching the Little Missouri (pl. 1). A typical view of the channel, between Smith and Shadwell Creeks, is shown in figure 6. The channel here is more than 200 feet deep and a mile or more across.

The channel descends to the north and is connected to the Yellowstone valley by several valleys which drain westward through the intervening upland. In view of these topographic relations, the level of Lake Glendive must have dropped intermittently as the front of the Yellowstone lobe withdrew northward. Consider, for example, the situation which must have prevailed when the ice front withdrew north of Smith Creek. The waters of Lake Glendive would have inundated Smith Creek valley and extended to the ice-marginal channel at its head. The elevation of the channel here, however, is lower than the elevation at the original lake outlet. The lake level, therefore, would have dropped to the level of the channel at the head of Smith Creek, leaving the Dry Creek outlet abandoned. Similarly, when the receding ice front exposed Shadwell Creek valley, the control for the lake level would have been the channel floor at the head of Shadwell Creek. The elevation of the channel here is still lower and undoubtedly resulted in another drop in lake level. By this method, the level of Lake Glendive dropped intermittently until the lake waters flooded Bennie Pierre valley and spilled down the Hay Creek channel to the Little Missouri River. At this stage the entire stretch of marginal channel to the south, being above the level of the reduced lake, lay abandoned, and only the Bennie Pierre-Hay Creek section was in use.

Meanwhile, the level of Lake Mikkelson, in the Little Missouri River valley, was also being lowered, but by progressive erosion of its eastern outlet. Reasons were presented earlier for believing that Lake Mikkelson was completely drained before Hay Creek channel was eroded to its present depth. If the reasoning is valid, Lake Glendive persisted in the Yellowstone River valley long after Lake Mikkelson disappeared.

Between the glacial limit and Missouri River valley to the north there are no channels of either the dimensions or the continuity of those which mark the outer border of the early Wisconsin(?) ice. The absence does not seem to be attributable to differences in topography; it seems more likely that the ice receded to the Missouri without any sustained halts. The relatively short and shallow channel segments which are present were apparently carved by local melt waters or by the escape waters of small ice-impounded lakes.

The evolution of Lake Circle and Lake Lambert beyond the stage shown in plate 1 is largely unknown. A lake must have occupied Redwater Creek valley throughout the withdrawal of the ice to the Missouri River. Lake Lambert, however, was unique in that it was located directly over a low part of the Yellowstone-Missouri drainage divide to which it was confined by ice on the east, north, and west and by rising ground to the south. As the ice shrank away from the divide area, Lake Lambert was probably split as its level fell below that of the divide. The formerly continuous lake thus probably separated into several smaller lakes confined to the valleys draining radially away from the divide. These lakes must necessarily have spilled across intervening divides from one to the other until they reached Lake Glendive.

Southwest of Lake Lambert, between Lane, Mont. and Richey, Mont., a continuous strip of alluvium passes across the divides separating several northwest-flowing streams (pl. 1). The strip of alluvium connects the headwaters of East Redwater Creek on the west with Fox Creek on the east. The channel occupied by this alluvium could not have been carved when Lake Circle and Lake Lambert were at their maxima, for these areas were then flooded. The channels were presumably carved after the dismembering of Lake Lambert, when individual lakes were confined to each of the northwest-draining valleys.

Lake Glendive must have continued to occupy the Yellowstone River valley during withdrawal of the ice to the Missouri River valley. It is unknown whether the Bennie Pierre-Hay Creek channel served as the outlet during the entire period of withdrawal, whether the lake water escaped into the Missouri valley around the margin of the Yellowstone lobe, or whether the escape routes were used successively.

East of the Yellowstone valley the ice front remained at the glacial limit long enough to allow the ice-marginal waters to erode below the level of the ancestral valley of the Little Missouri, or to within 100 to 200 feet of the present valley bottom. At this time the ice covered all the region east of the Killdeer Mountains, and the drainage from the west swung south along the ice front. The present trench of the Little Missouri, from the Killdeer Mountains to the Missouri, had not yet come into existence. As the ice front receded down the eastern slope, it exposed numerous east-draining valleys. One of these was selected as the escape route for the diverted waters of the Little Missouri, thereby completing the course as it is today.

As the ice withdrew northward, a proglacial lake (Upper Cherry Creek Lake, pl. 1) was confined within the abandoned, northward-draining ancestral valley. This lake, too, expanded northward as the ice receded; it drained south into the Little Missouri where the old valley hung above the new.

It was suggested earlier that the southeast-trending segment of Redwing Creek, between the ancestral and modern valleys of the Little Missouri, represents the former lower course of Bennett Creek when the latter was tributary to the ancestral river (pl. 8). Redwing Creek was presumably captured by a tributary of the modern Little Missouri eroding headward along the abandoned former lower course of Bennett Creek.

While Upper Cherry Creek Lake was expanding in the ancestral valley of the Little Missouri, another glacial lake, Lower Cherry Creek Lake, may have occupied what is now the southeast-trending lower course of Cherry Creek. At that time, however, lower Cherry Creek valley sloped to the north and was occupied by a former northwest-trending tributary of the ancestral Little Missouri River (pl. 8). Upper and Lower Cherry Creek Lakes expanded northward in the wake of the receding ice; their outlets presumably remained at their distal ends, at the Little Missouri. The channel west of Watford City, N. Dak., may indicate a temporary halt in the recession of the ice. The channel was presumably carved by the outlet waters of a small lake impounded in the headwaters of Tobacco Garden Creek.

When the ice front reached the junction of the ancestral Little Missouri River valley and the tributary valley from the southeast, Upper and Lower Cherry Creek Lakes merged to form a single lake, which we will refer to simply as glacial Cherry Creek Lake. The outlet across the divide at the distal end of the southeast arm of the lake was probably lower than that at the end of the southwest arm, as suggested by the direction of the present drainage of the lake basin. Rapid headward erosion by the outlet waters at the southeast end of the lake may have contributed to the complete reversal of drainage in lower Cherry Creek valley. In contrast, the drainage has been reversed in only the distal portion of the basin of Upper Cherry Creek Lake.

Cherry Creek Lake expanded as the ice front withdrew north. Inasmuch as there are no other possible outlets between Cherry Creek and the Missouri River valley, 20 miles to the north, the subsequent evolution of Cherry Creek Lake depended on the rate of recession of the ice and the rate of lowering of the outlet. As the outlet stream into the Little Missouri valley eroded headward into the northwest-sloping floor of Cherry Creek valley, the level of Cherry Creek Lake must have lowered. If the migration of the outlet had been slow relative to the rate of recession of the ice, Cherry Creek Lake may have extended far to the north along Tobacco Garden Creek valley and may have had long branches to the southeast and southwest. If migration of the outlet had been relatively rapid, the two branches of Cherry Creek Lake may have been completely drained before the ice had receded very far north of the Cherry Creek-Tobacco Garden Creek divide. Under these circumstances Cherry Creek Lake may never have had a great extent. In either situation, the lake eventually became restricted to the valley of Tobacco Garden Creek, and its level was determined by the elevation of the Cherry Creek-Tobacco Garden Creek divide.

At some stage during retreat of the early Wisconsin(?) ice, the front of the Watford lobe in the ancestral valley of the Little Missouri stood at the recessional(?) moraine which forms the present divide between Cherry and Tobacco Garden Creeks, and its east flank stood at the site of the Keene moraine (pl. 1). The northwesterly trending channel south of the Keene moraine was presumably carved by escape waters from a glacial lake in the head of Bear Den Creek valley. This suggests that the Keene-Charlson upland was surrounded by ice except in the southwest. The channel waters probably escaped south by way of Cherry Creek Lake and the Cherry Creek outlet into the Little Missouri River. A second, less conspicuous channel descends into Clear Creek valley and may have a similar history.

Two channels are mapped in northwestern McKenzie County, N. Dak. (pl. 1). The easternmost crosses the shallow divide between Charbonneau and Timber Creeks. It was presumably carved by waters flowing northeast. This would require a lake in Charbonneau Creek valley, which, in turn, would require that the Yellowstone ice lobe still occupy Yellowstone River valley, into which Charbonneau Creek flows. The second channel, north of Cartwright, N. Dak., descends to the south and was probably formed by melt waters along the margin of the ice when the ice had almost completely withdrawn from the upland in this area.

Eventually the early Wisconsin(?) ice withdrew to the Missouri River valley. It seems inconceivable that the receding ice front should have reached the Missouri valley simultaneously throughout the entire 200 miles of winding valley within the map area. It seems more likely that the several ice lobes, which are assumed to have arrived first at the Missouri during advance of the ice, would be the last to withdraw. Thus, ice lobes may have blocked the valley in some places while intervening stretches were ice free. This would imply separate ice-dammed lakes. If such lakes existed, their deposits cannot now be distinguished from other valley deposits. The situation, however, was probably even more complicated. Part of the thick ice in the Missouri valley probably remained as relic ice long after the active ice front had withdrawn to the north. On the south side of the Missouri valley opposite Wolf Point, Mont., is a marginal channel about 5 miles long and one-half to 1 mile wide. It is separated from the Missouri trench by a linear bedrock mass several square miles in area. About 4 miles to the west, another much narrower ridge separates the continuation of this channel from the main valley. The bedrock ridge is missing in the intervening area and the "channel floor" appears as a broad terrace. At least this part of the channel may have been eroded when the valley was occupied by stagnant ice, and part of the fill of the Missouri valley may have been deposited on or against masses of such ice.

Prior to the early Wisconsin(?) glaciation, Redwater Creek, which now enters the Missouri valley opposite Poplar, Mont., used to enter the Missouri to the west at the mouth of Nickwall Creek (pl. 1). A broad, marshy swale marks its former, more direct path. Presumably, the early Wisconsin(?) ice blocked this direct route into the Missouri, forcing Redwater Creek into its present path. The new route was eroded below the level of the old while ice still blocked the former course. An esker at the head of the abandoned valley suggests that the ice which blocked the old channel was stagnant rather than active.

It will be recalled that prior to the advance of the pre-early Wisconsin(?) ice, the Missouri flowed through the Culbertson-Bainville channel. The apparent absence(?) of glacial till within the present trench to the south of the channel suggests that it was formed during withdrawal of the early Wisconsin(?) ice. According to this interpretation, the Culbertson-Bainville channel and part of the islandlike upland to the south remained ice covered for a long period of time. During this time, the ponded waters of the Missouri overflowed along their present route and carved a new channel below the level of the old one. The location of the new route was probably determined in part by the presence of a long, northwest-draining prediversion valley (pl. 8). The kames shown in plate 1 on the north side of the present trench are in a tributary valley below the upland level. This does not mean that ice revisited this area after erosion of the trench. In the writer's opinion the small valley preceded the trench and was tributary to a northwest-draining valley whose orientation and direction determined the orientation and direction of the present trench.

A similar sequence of events took place at the Hofflund channel; the present trench was eroded as an ice-marginal channel while the broad channel to the north was occupied by ice.

The narrowest stretch of the Missouri valley is below Sanish, N. Dak., near the east border of the map. Geomorphically, the valley is here in very early maturity, whereas over the greater part of the map area it is in full maturity. Extending southeast from Sanish, however, is a broad swale which has been described by several observers and is clearly a former course of the Missouri. For reasons to be discussed later, it is believed that the displacement of the river from this channel took place in Mankato, rather than early Wisconsin(?) time.

While the early Wisconsin(?) ice was withdrawing from the glacial limit south of the Missouri, it was at the same time receding down the slopes of the Flaxville Plateaus north of the river. Valleys draining southeast were once again occupied by lakes and once again the lakes overflowed southwestward along channels marginal to the ice.

At about the time that the wasting ice surface stood at the level of the deep channel at the foot of the Flaxville Plateaus, the Big Muddy-Little Muddy divide began to reappear above the ice. Reappearance of the divide separated the Poplar and Little Muddy ice lobes. A moraine, which may be interlobate, caps the divide.

The moraine (pl. 1) is hairpin shaped, 15 to 16 miles long in a northeast direction, and about 5 miles wide. The dimensions apply to the area of pronounced morainal topography; the topography becomes progressively more subdued outward, especially to the northeast, east, and southeast. The most prominent morainal topography is at the southern terminus of the eastern prong of the hairpin. Here the relief approaches 50 feet. Boulders as much as 6 feet in diameter, mostly of granite, are present. The western prong of the hairpin also shows pronounced morainal topography and includes one esker that is half a mile long and 20 to 30 feet high. The low area within the hairpin is divided into two lowlands by a morainal crossbar. Till is exposed in the basins as well as in the morainal uplands, but widespread glaciofluvial and possibly glaciolacustrine deposits mask the morainal topography. The overall characteristics of the moraine seem to be reasonably explained on the basis of an interlobate origin.

An interesting boulder, roughly disc shaped and about 10 feet in diameter and probably 6 to 7 feet thick, was found in sec. 13, T. 30 N., R. 58 E., between the two branches of Sand Creek. The boulder, which is now in four fragments, consists of a coarse basic igneous rock. The surface, although irregular, is remarkably smooth. It was at first thought that the boulder was a meteorite and that the smooth surface was due to fusion. However, the smooth surface, which continues around the edges of the four fragments, is probably due to wind polishing. The high degree of polishing, if it took place after deposition of the boulder, may be independent evidence of the pre-Mankato age of the moraine.

In general, the pattern of withdrawal of the early Wisconsin(?) ice north of the Missouri River was largely a reversal of the pattern of advance. Inasmuch as the major valleys of the northern area drain south toward the Missouri, none were occupied by proglacial lakes. The southeast-trending stretches of the upper Poplar River and upper Big Muddy Creek, however, were dammed along the western margin of the Poplar lobe, as were the other southeast-flowing streams.

Lake Brockton (pl. 1), mentioned in the discussion of the advance of the early Wisconsin(?) ice, undoubtedly reappeared during recession of the ice. Its presence is demanded by the topography. The lake was confined by the ice to the northeast-sloping floor of the ancestral Missouri valley. The lake outlet, northwest of Brockton, Mont., is at an elevation of about 2,100 feet, but the channel was probably deepened 50 feet or more. The maximum level of Lake Brockton was, therefore, about 2,150 feet. At a lower level, the lake drained west into the Poplar River by way of the channel along which the latter once entered the ancestral Missouri valley. When the ice withdrew far enough to expose the southeast-trending course of Big Muddy Creek, the lake was completely drained.

As the Poplar lobe withdrew to the northeast and shrank down the flanks of the ancestral Missouri valley, it left in its wake a complex array of melt-water channels. Many of these have been described by Witkind (1959), and the reader is referred to his report for details.

The northern and northeastern parts of the map area are covered by later drifts. In general, the extent to which the topography of the early Wisconsin(?) drift is reflected in the later topography is unknown. However, many if not all of the large swales that cross the coteau belt were probably present in early Wisconsin(?) time. The same may be true for east-west channels, such as the one extending east from Grenora, N. Dak.

If the author's mapping is correct, the eastern half of the Grenora-Zahl channel in northwestern Williams County is within the area of early Wisconsin(?) drift. The channel is steep sided, relatively flat floored, from a quarter to half a mile wide, and 100 to 150 feet deep. The presence of early Wisconsin(?) till in the eastern part of the channel seems to indicate that the channel is at least as old as the early Wisconsin(?) glacial episode. The channel is a distinct topographic feature for more than 50 miles within the area of the Mankato drift. Deposition of the later drift has, therefore, failed to obliterate expression of this channel.

RATE OF RECESSION

A remarkable succession of low, parallel ridges north of the Missouri River extends from a point about 20 miles west of the map area northeastward to Coalridge in Sheridan County, Mont., a distance of 110 miles (pl. 1). These ridges, which we shall refer to as "bars" to forestall confusion with larger glacial features, are low, probably averaging less than 10 feet in height, although some are 25 feet. They may be several hundred feet across at the base and as much as 2 miles long. The average length is probably a quarter of a mile or less. The bars trend northwest and tend to be subparallel, although individuals may diverge 20° or more from the general trend. In general, the trend of the bars changes gradually from about N. 45° W. at the northeast end of the belt to N. 30° W. or less at the southwestern end of the belt. In the Oswego quadrangle just to the west of the map area, the orientation is locally nearly north-south. The bars are in general confined to a belt 8 to 10 miles wide on the floor of the broad swale which influenced the development and the direction of movement of the Poplar lobe of the early Wisconsin(?) ice.

Jensen (1951) has mapped in detail the bars of the Frazer quadrangle west of the present map area at the southwest end of the belt. Here, the bars locally form a conjugate network with both northwest and northeast members. Conjugate bars have been mapped by Colton (1954) in the Chelsea and Poplar quadrangles. Northeast-oriented bars are less common in the present map area.

Regardless of the detailed origin of the bars, they are clearly depositional rather than erosional features. Near modern valleys, ravines have expanded headward along the swales between the bars, but the greater part of the belt presumably preserves the original till surface.

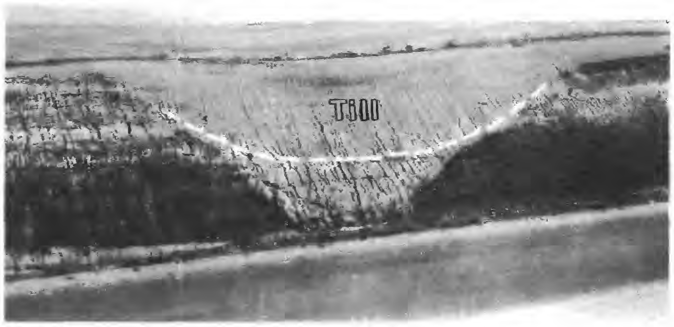
In the writer's opinion, the bars along the floor of the ancestral Missouri River valley are recessional features, such as have been recognized in Finland for four decades. Gwynne (1942) has described comparable features in Iowa which outline in detail the Mankato lobe in that region. Following Gwynne, the writer interprets these bars as being annual recessional features, although it is doubtful that the sequence of bars furnishes an unsullied record of retreat of the ice. There may have been halts of longer duration than a year, and readvances of the ice may have destroyed some of the bars previously deposited. The fact that the bars are relatively short is no argument against their origin as ice-marginal features. Conditions conducive to accumulation of marginal deposits vary considerably along ice fronts, depending on the local rates of forward motion of the ice, the debris content of the ice, and the local rates of backwasting. Furthermore, as noted above, slight readvances may obliterate parts of bars deposited earlier. The bars are not found in areas of irregular topography where presumably conditions were not conducive to linear deposition.

The recessional-bar theory does not in itself account for the conjugate pattern noted by Jensen west of the map area. Jensen (1951, p. 26; 1952, p. 85) has suggested that the conjugate pattern reflects a fracture pattern in the ice created by regional deformation during ice occupation of the area. Thus, the bars are "crevasse" fillings. One difficulty with this hypothesis is that the conjugate pattern is largely restricted to the western end of the 110-mile-long belt of bars. A second difficulty is the fact that all the bars examined by the writer and by others working in the area consist of till, whereas crevasse deposits should show evidence of water-working. Thirdly, the orientation of the bars changes to conform to the orientation of the broad swale followed by the Poplar lobe. It is difficult to understand why manifestations of regional deformation should change orientation to conform to surface topography. It would seem that the crevasses ought to maintain a regional orientation regardless of the trends of areas of stagnant ice. Fourthly, comparable though less regular bars in the area of the Mankato drift are arranged in sweeping curves conforming to the pattern of topographic highs and lows.

As the writer has not examined the conjugate-bar system in the area described by Jensen, the following explanations are only speculative. In the first place, a local conjugate arrangement of crevasses is not unusual at the termini of glacial lobes where both transverse and longitudinal crevasses may be present. The crevasse deposits, lowered to the ground during wasting of the ice, might therefore give rise to the conjugate bars under consideration. The bars, however, should consist largely of glaciofluvial deposits, just as they should under Jensen's hypothesis. Possibly the northeast-southwest bars represent longitudinal crevasse deposits let down on the ground below as the ice front was wasting back and depositing northwest-southeast recessional till ridges. This would require that at least the northeast-southwest bars consist of glaciofluvial debris. The writer does not know whether this is the case or not. Perhaps the northeast-southwest bars were deposited during advance of the ice as attenuated drumlins, whereas the northwest-southeast bars were deposited as ice-marginal recessional features. This would explain the till composition of the bars but would require an explanation of the scarcity of the northeast-trending bars except in this distal area. In any event, deformation is not required to explain intersecting crevasses at the margin of a glacial lobe.

On the assumption that the bars are annual recessional features, the writer has attempted to determine the rate of recession of the Poplar lobe of the early Wisconsin(?) ice. A line was drawn on aerial mosaics, down the center of the belt of ridges dividing the belt into two equal halves. Ridges on either side of the medial line were projected onto this line. In spite of the difficulty of correlating some bars across the medial line, and in spite of the problems created by bifurcating bars, it is believed that an estimate of about 7 bars per mile is reasonable. This suggests that, for a large part of the 110-mile belt, the rate of retreat has been about 1 miles in 7 years. This, however, is a maximum rate, because (1) it is by no means certain that all the bars were identified on the small-scale photo-index sheets used by the writer, (2) the ice probably lingered for several years at some of the larger bars, and (3) slight readvances of the ice may have obliterated earlier bars. On the other hand, the presence of gaps in the long succession of bars may indicate stages of very rapid recession, although the writer doubts that the ice front moved back 6 miles in 1 year, as indicated by the width of 1 gap. Gwynne estimated a rate of recession of 1 mile in about 15 years for the Mankato lobe in Iowa.

Assuming, then, that the bars are actually recessional features, it took a minimum of about 770 years for the front of Poplar lobe to withdraw 110 miles



A

B

FIGURE 41.—*A*, Early Wisconsin(?) till completely filling small valley in bedrock. The approximate contact is indicated by the broken line. The light deposit below the broken line is clayey silt washed down from the till above. U.S. Highway 2 between Big Muddy and Boxelder Creek valleys, Roosevelt County, Mont. *B*, Pebble and cobble concentrate on till, just north of U.S. Highway 2, about 2 miles east of Brockton, Mont.

from a point about 20 miles west of the map area to the vicinity of Coalridge in Sheridan County, Mont.

LANDSCAPE MODIFICATIONS

The drainage changes effected by the early Wisconsin(?) ice have already been discussed. The landscape modifications will now be considered briefly.

The principal effect of the early Wisconsin(?) glaciation was to subdue relief. With few exceptions, the broad divides of the Missouri Plateau peneplain are mantled with a thin, often patchy veneer of till. Bedrock is exposed in many road cuts at the crests of these divides. The major valleys, however, were deeply buried, and many small valleys were completely obliterated (fig. 41*A*). Level areas were generally unmodified by the ice, except for deposition of a layer of drift of uniform thickness. In a few places the early Wisconsin(?) ice increased the relief. These are the morainal areas, including the Keene moraine, the "hairpin" moraine on the Montana-North Dakota border north of the Missouri River, a moraine on the Flaxville upland west of the Poplar River in Roosevelt County, Mont., a moraine on the upland across the Poplar River to the northeast, and a few smaller areas (pl. 1). Morainal areas, however, are relatively rare in the area of early Wisconsin(?) drift, although faint closed depressions are scattered over the level uplands.

It is expectable that the landscape should have been progressively less affected toward the glacial limit. Actually, within several miles of the drift border only boulders record the former presence of the ice. There are no marked contrasts in topography across the glacial border. The topography north of the Little Missouri River, for example, is identical with that south of the river, and in many places it is impossible to determine exactly where the glacial limit should be drawn.

The most impressive characteristic of the surface of the early Wisconsin(?) drift is its graded appearance. Long, smooth slopes descend from the divides toward the valley bottoms, generally to terminate as terraces above the present flood plains. Over large areas, the smoothness of this surface is due in part to the smoothness of the underlying surface and in part to the uniform thickness of the overlying till. Two additional factors are probably involved. The first is elimination of depressions by deposition of glaciofluvial sediments, by accumulation of slope wash and vegetation, and by creep. The second is lowering of morainal hummocks and ridges by slope wash, mass movements, and deflation. The lowering of the surface is suggested by the presence of a widespread lag concentrate of pebbles and cobbles on the till itself. The concentrate locally reaches 2 to 3 feet in thickness (fig. 41*B*).

The smooth till surface extends either to the very edge of the bluffs overlooking the major valleys or descends to lower levels and terminates as a terrace above the valley floors. The lowest and flattest part of the terrace is about 20 feet above the present flood-plain level in major valleys, or about 40 feet above low-water river level. Where the terrace consists entirely of till, or of till overlying an irregular topography below, appreciable grading of the terrace surface must be presumed to have taken place. In the writer's opinion, this grading was accomplished at a time when the streams were flowing at the terrace level. Possibly the valleys of the region were clogged with glacial drift to this height. The Missouri River, at the level of the valley fill, would presumably serve as the base level with respect to which the till surface would be graded by side streams, slope wash, mass movements, and creep.

In some places where the graded till surface on the upland extends to the edge of the bluffs overlooking the Missouri River valley, the till is capped by glaciofluvial sediments. It is inconceivable that stream-laid deposits could have been deposited at the brink of such precipitous slopes. There has been ample opportunity for backwearing of the bluffs since early Wisconsin(?) time, however, and what seems like precarious situations now may not have been precarious then.

A large part of the grading of the surface of the early Wisconsin(?) drift probably took place during the early-middle Wisconsin(?) interval. This is suggested by the apparent continuation of the surface under the middle Wisconsin(?) drift at Jones cut and elsewhere and by the geomorphic relations in Little Muddy Creek valley described in the section dealing with the early-middle Wisconsin(?) erosion interval. However, the surface of the middle Wisconsin(?) drift and the surface of the Mankato drift over wide areas also appear to be graded. This raises the question of whether there were three episodes of grading, one after each glacial substage, or whether the level parts of the surfaces of one or more of the drifts merely reflect uniform deposition on level surfaces below. The application of this latter suggestion to the middle Wisconsin(?) drift is discussed later.

If the scattered, faint depressions on the uplands are original morainal depressions, their presence suggests that, at least in some topographic environments, streams were relatively unimportant in the grading process. Solifluction and other periglacial processes may have been more important. The area covered by the annual recessional bars has apparently undergone little modification of any sort.

SUMMARY OF EARLY WISCONSIN(?) GLACIATION

At the beginning of early Wisconsin(?) time, the major valleys of the region were more than 100 feet deeper than they are now, and the Little Missouri River still flowed north in its ancestral valley to join the Missouri near Hofflund, N. Dak. The Missouri itself was already established in its present course except that it still occupied the Culbertson-Bainville, Hofflund, and Sanish channels.

The Poplar and Little Muddy ice lobes advanced up the ancestral valleys of the Missouri and Yellowstone Rivers, respectively, and eventually merged by overtopping the intervening divide. The ice completely overrode the Missouri River valley and extended 25 to 30 miles to the south. At the height of the glaciation, the highest parts of some the Flaxville Plateaus north of the Missouri projected above the ice.

North-draining valleys were blocked at the glacier terminus to form lakes. Redwater Creek valley was occupied by glacial Lake Circle; the Yellowstone River valley, by glacial Lake Glendive; and the Little Missouri River valley, by glacial Lake Mikkelson. A low part of the Yellowstone-Redwater divide, almost completely enclosed by ice, was submerged by the water of glacial Lake Lambert. The lakes were connected across intervening divides by deep ice-marginal channels. During withdrawal of the ice to the north, lakes continued to occupy the north-draining valleys until the Missouri trench became exposed.

The Little Missouri River retained its ice-marginal path to the east at the glacial limit. An ice-dammed lake, Cherry Creek Lake, occupied its abandoned valley floor as the ice withdrew northward. The other north-draining rivers forsook their ice-marginal courses to resume their northward flow into the Missouri River. The Missouri itself was permanently displaced from the Culbertson-Bainville and Hofflund channels.

During withdrawal of the Poplar lobe north of the Missouri, Lake Brockton was confined to the north-draining ancestral valley of the Missouri. The rate of retreat of the Poplar lobe appears to have been about 1 mile in 7 years.

The principal effect of the early Wisconsin(?) ice on the landscape was to reduce relief by concentrating its deposits in the valleys. Relief was locally increased, however, by deposition of moraines. The surface of the early Wisconsin(?) drift is smoothly graded over much of its extent. The grading was fashioned, at least in part, prior to the middle Wisconsin(?) glaciation. Solifluction and other periglacial processes may have contributed to the grading.

EARLY-MIDDLE WISCONSIN(?) EROSION INTERVAL

After development of the graded surface on the early Wisconsin(?) drift and prior to advance of the middle Wisconsin(?) ice, the rivers of the region began to trench their debris-cluttered valleys. Although the trenching is assumed to have been regional, the evidence comes only from the valley of Little Muddy Creek north of Williston. Elsewhere, this stage of dissection cannot be distinguished from the dissection which took place during the middle Wisconsin(?)-Mankato and the post-Mankato intervals.

The graded surface of the early Wisconsin(?) till terminates as a terrace in Little Muddy Creek valley. The terrace itself consists of till. In two places, 12 and 18 miles north of Williston, pronounced moraines of middle Wisconsin(?) drift occupy part of the floor of the valley (pl. 1 and fig. 20). The moraines include many undrained depressions, some of which are occupied by lakes and marshes. The surface of the early Wisconsin(?) till terrace stands 40 to 50 feet above the level of Little Muddy Creek and is higher than most, if not all, of the depressions in the immediately adjacent middle Wisconsin(?) moraines. The terrace surface could not have been formed at its present level above the morainal areas alongside nor, if it is a regional feature, could it have been developed while the middle Wisconsin(?) ice shielded these local morainal areas. It seems more probable that the graded surface was formed prior to middle Wisconsin(?) time, although it may have undergone additional smoothing later. The graded till surface was trenched by Little Muddy Creek prior to deposition of the middle Wisconsin(?) moraines. A water well in the northern morainal area, which is on the west side of the valley, was sunk 70 feet without reaching bedrock. The rock floor of the valley at this place, therefore, is more than 40 to 50 feet below present flood-plain level. It is precisely here, where the rock floor is deep, that the morainal topography of the middle Wisconsin(?) drift is found. It seems probable that Little Muddy Creek was diverted from the deep part of the valley under the middle Wisconsin(?) moraine to its present site, either by ice or by glacial deposits. The writer favors diversion by ice, because casual inspection seems to indicate that the moraine itself is not high enough everywhere across the former valley to account for the displacement of the stream to the east. The ice, furthermore, apparently remained in the old valley long enough for the displaced stream to cut its new channel below the level of any possible escape route through the present moraine.

The same sort of diversion probably took place at the southern morainal area, but with displacement of the stream to the west.

The conclusion seems justified that Little Muddy Creek valley was incised after deposition of the early Wisconsin(?) till and after development of the graded till surface. The 70 feet of till in the northern morainal area is not necessarily an indication of the depth of stream trenching; part of the depth may be due to glacial scour. The presence of middle Wisconsin(?) drift about 25 feet below the till terrace indicates at least that much entrenchment below the graded till surface prior to advance of the middle Wisconsin(?) ice; the actual amount may have exceeded 70 feet.

MIDDLE WISCONSIN(?) GLACIATION

Following trenching of the graded surface of the early Wisconsin(?) till, the area was invaded by the middle Wisconsin(?) ice. The latter apparently extended only into the northeastern part of the map area. A lobe of the glacier, the Williston lobe, pushed southwest to the Missouri River valley, its direction of advance in large part determined by the trend of the ancestral valley of the Yellowstone River. In most exposures, the till is only 4 or 5 feet thick and has scarcely modified the earlier topography. Locally, as noted earlier, it may exceed 70 feet.

The area of middle Wisconsin(?) drift in plate 1 represents the minimum area covered by the ice. The thin edge of the ice may have extended beyond the borders mapped without leaving identifiable deposits.

The Williston lobe apparently reached the Missouri River valley at Williston and may have flooded the valley nearly to the mouth of the Yellowstone River. Unless there was space between the ice and the southern bluffs for the Missouri to escape around the ice margin, the valley must have been dammed. Damming at this point would have flooded the Yellowstone valley as well. Inasmuch as these valleys were probably deeper than they are now, because of the premiddle Wisconsin(?) episode of erosion, the lake surface may have been below present flood-plain level. Its deposits, therefore, would be included in the complex valley fill. Even if the lake level had stood higher than the present valley floor, the separation of the lake deposits from the complex assemblage of sediments which are exposed in patches along both valleys would be difficult.

The presence of glaciofluvial sands and gravels on middle Wisconsin(?) till at the edge of the bluffs of the Missouri valley at elevations that exceed 150 feet above river level does not necessarily mean that the Missouri valley was ever completely filled to that height. Had so thick a fill ever existed, remnants would probably be much more abundant throughout the drainage system, especially in view of the relatively short time available for subsequent erosion. Some of the deposits were almost certainly deposited on the upland when the edge of the bluffs was farther riverward. Other deposits may have been laid down marginal to masses of stagnant ice within the valley.

During advance and recession, the Williston lobe blocked valleys tributary to Little Muddy Creek. Marginal channels were eroded along the flanks of the Williston lobe during its withdrawal and possibly in part during its advance. Some of these are shown in plate 1 on both sides of Little Muddy valley north of Williston. The deepest of the channels is about a quarter of a mile wide and possibly 100 feet deep. The two channels north of East Fork of Little Muddy Creek were probably proglacial spillways, because they occupy the floor of a broad swale.

If future investigation should reveal early Wisconsin(?) till on the floors of some of these channels, the presumption would be that the middle Wisconsin(?) melt waters locally made use of preexisting channels.

Available evidence indicates that the front of the middle Wisconsin(?) ice halted a short distance north of Cow Creek for an appreciable length of time. The evidence is as follows:

1. The glaciofluvial gravels of the Williston terrace extend up Little Muddy Creek valley as far as the mouth of Cow Creek, but not beyond. Similar gravels, however, extend up Cow Creek valley, where they form a series of terraces, the highest of which (25 to 40 feet) is comparable in height with the Williston terrace (fig. 42). The gravels are nowhere overlain by the middle Wisconsin(?) till, hence they could be either middle Wisconsin(?) recessional outwash or outwash from the Mankato ice. However, the Mankato ice did not come within 15 miles of Cow Creek nor did its melt waters have access to this valley. Hence the gravels within Cow Creek valley cannot be due to direct deposition by Mankato melt waters. Nor can they be attributed to deposition in Cow Creek valley in response to Mankato deposition in Little Muddy Creek valley, because the Cow Creek deposits stand 40 feet or more above the buried Mankato outwash in this part of Little Muddy valley. A middle Wisconsin(?) age for the terrace gravels is indicated. The lower terraces in Cow Creek valley are believed to represent halts in degradation of a single sedimentary fill, because there seems to be no logical explanation for successive trenching and filling in the middle Wisconsin(?)-Mankato interval.

2. All the small south-draining tributaries of Cow Creek that were examined are clogged to their heads with coarse, cobbly, poorly sorted glacial gravel. At least one of these valleys rises in an area of subdued moraine. The relations suggest that outwash was poured into these valleys and on into Cow Creek valley from an ice front immediately to the north.

3. The terrace-forming gravels in Little Muddy and Cow Creek valleys were deposited while ice still covered the morainal area in Little Muddy valley 2 to 3 miles above the mouth of Cow Creek. This is indicated by the presence of undrained depressions in the morainal area below the level of the terrace sediments to the south. It is hardly likely that undrained depressions could be exhumed from beneath a sedimentary cover. It is more likely that overlying ice protected the depressions from burial under the sediments. Inasmuch as the moraine consists of middle Wisconsin(?) till, it was the middle Wisconsin(?) ice which lay to the north of Cow Creek valley when the sediments were deposited.

The above observations suggest that the receding front of the middle Wisconsin(?) glacier remained stationary for a long time just north of Cow Creek. During this interval, a thick valley train was deposited to the south in Little Muddy valley. At the same time, the ice front west of Little Muddy valley stood a few miles north of Cow Creek valley and fed large quantities of glacial debris down south-flowing tributaries to account for the thick sequence of coarse deposits in Cow Creek valley.

As the ice front resumed its withdrawal to the north, proglacial waters in Little Muddy valley were probably blocked by the high outwash fill to the south, and a lake may have occupied the valley until such time as the barrier was breached. Another halt in the ice is indicated by the moraine near Marmon, N. Dak., but no thick outwash deposits comparable with those at and below Cow Creek have been observed.

Nothing more than has been outlined is known of the middle Wisconsin(?) episode of glaciation. North of the headwaters of Little Muddy valley, the region is buried under the Mankato drift, the product of the latest glaciation. Prior to advent of the Mankato ice, however, there was another episode of erosion which will be considered shortly.

SUMMARY OF MIDDLE WISCONSIN(?) GLACIATION

A lobe of the middle Wisconsin(?) glacier extended south along the ancestral valley of the Yellowstone River and reached the Missouri River between Williston and the present mouth of the Yellowstone. The ice may have dammed the Missouri, but no direct evidence for this has been found. During both advance and withdrawal, the ice lobe dammed tributaries of Little Muddy Creek, and ice-marginal channels were eroded across intervening divides.

The ice front remained fixed for some time just north of Cow Creek. Large quantities of outwash were poured into Cow Creek valley and into Little Muddy valley below Cow Creek. These deposits form the eastern part of the terrace on which Williston is located.

In only two places were thick deposits of middle Wisconsin(?) drift laid down. These form moraines in Little Muddy valley 12 and 18 miles north of Williston. Elsewhere, the ice spread a rather uniform layer of drift 4 to 5 feet thick.

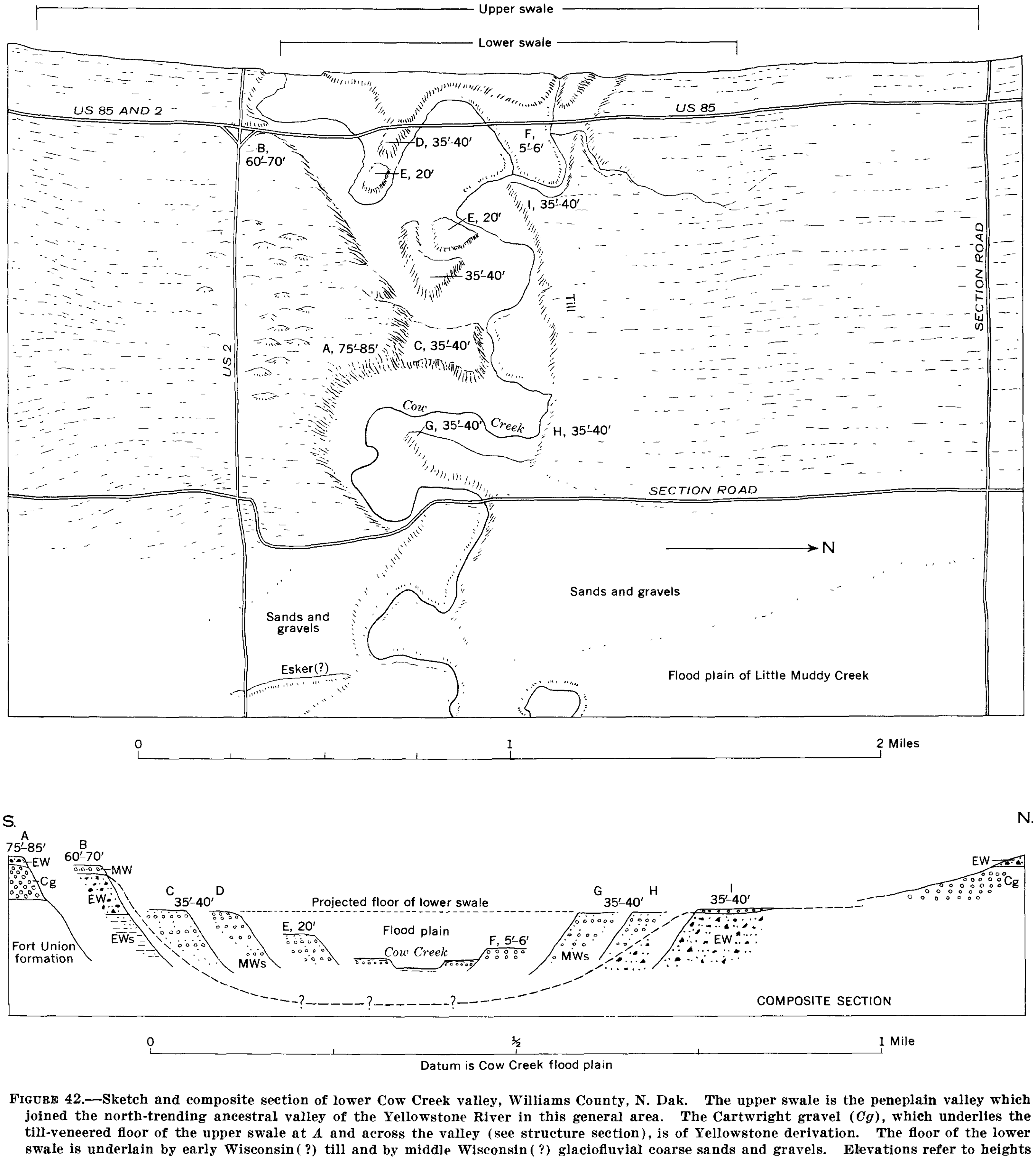


FIGURE 42.—Sketch and composite section of lower Cow Creek valley, Williams County, N. Dak. The upper swale is the peneplain valley which joined the north-trending ancestral valley of the Yellowstone River in this general area. The Cartwright gravel (*Cg*), which underlies the till-veneered floor of the upper swale at *A* and across the valley (see structure section), is of Yellowstone derivation. The floor of the lower swale is underlain by early Wisconsin(?) till and by middle Wisconsin(?) glaciofluvial coarse sands and gravels. Elevations refer to heights above Cow Creek. *MWs*, middle Wisconsin(?) outwash younger than the till (at *C*, *D*, *E*, *F*, *G*, and *H*); *MW*, middle Wisconsin(?) till; *EW*, early Wisconsin(?) till; *EWs*, early Wisconsin(?) outwash older than the till; *Cg*, Cartwright gravel.

MIDDLE WISCONSIN(?)-MANKATO EROSION INTERVAL

Evidence in Little Muddy Creek valley suggests that a minimum of 55 feet of erosion took place during the middle Wisconsin(?)-Mankato interval. The evidence is as follows:

At the site of the middle Wisconsin(?) moraine east of Marmon, N. Dak., (pl. 1), Little Muddy Creek flows in a broad inner trench 40 to 50 feet deep (fig. 20, section *C–D*). The flood plain consists of fine-textured alluvium, above which stands a gravel terrace 16 to 18 feet high. The terrace-forming gravels can be traced continuously upstream to the Mankato drift border and obviously comprise a Mankato valley train. The lithology of the gravels and the till are similar. Near its head, the valley train is pitted by kettles of all shapes and sizes.

Near Zahl and Appam, N. Dak., at the Mankato drift border, the Mankato valley train stands 25 to 30 feet above the flood plain of Little Muddy Creek. This interval decreases steadily to the south until the outwash passes below the alluvium a short distance below the mouth of East Fork of Little Muddy Creek. The outwash continues south under the alluvium and is locally revealed in the banks of the creek.

The gradient of the surface of the valley train averages about 10 feet per mile near the drift border and diminishes downvalley. The slope of the Little Muddy flood plain averages only about 4 feet per mile, which explains the downvalley disappearance of the outwash under the alluvium. Even if it be assumed that the gradient of the outwash decreases under the alluvium, which seems reasonable in view of the decrease in pebble size, the outwash would probably still lie below river level at the Missouri River valley. If this inference is correct, the surface of the Mankato outwash lies a minimum of 55 feet below the surface of the middle Wisconsin(?) gravel of the Williston terrace.

The minimum depth of the channel occupied by the Mankato outwash at the morainal area in Little Muddy valley east of Marmon is probably 40 to 50 feet, the distance between the surface of the till terrace and the lowest exposure of Mankato outwash within the inner valley. Both here and at the Missouri valley, the depth of pre-Mankato erosion was probably greater than the estimates given.

MANKATO GLACIATION

The Mankato ice was the last to invade the region. The Mankato drift, like the early Wisconsin(?) drift, contains an abundance of limestone and dolomite, suggesting that the Mankato ice also came from the northeast.

The Missouri escarpment, at the northeastern edge of the coteau morainal belt (pl. 1), must have retarded advance of the ice. It seems reasonable to suppose, however, that glacial tongues filtered ahead along favorably disposed valleys. Eventually the ice flooded over the escarpment and resumed its advance southwestward across the Missouri Plateau. In the western part of the map area, the Medicine Lake lobe protruded southwest along the broad swale marking the ancestral course of the Missouri River and reached the southwest end of Medicine Lake. No other lobes of comparable dimensions developed, as far as the writer is aware. A small lobe may have extended south nearly to Scobey, Mont., in the northwestern corner of the map area. The small Zahl lobe invaded Little Muddy Creek valley in North Dakota, but it occupied only a small part of the area which had been earlier occupied by the Williston lobe of the middle Wisconsin(?) ice. Farther east, a broad lobe with pronounced morainal topography extended beyond Ray, in eastern Williams County, N. Dak. This lobe is unusual in that, as mapped, it coincides with a broad, low divide rather than with a lowland. The glacial border in this area, however, is more speculative than it is elsewhere.

Only one possible diversion of the Missouri River resulted from the Mankato glaciation, and that was the displacement from the Sanish swale (pl. 2). The Sanish swale cuts across the semicircular curve of the Missouri known as Big Bend. The swale is about 15 miles long and 1 to 4 miles wide. Moraine covers part of the floor and west flank. That the Mankato ice was responsible is suggested by the fact that the present trench coincides with the border of the Mankato drift, and the trench is narrower here than anywhere else in the region.

It will be recalled that the Missouri River valley had been eroded more than 100 feet deeper than it is now prior to the early Wisconsin(?) glaciation. The Missouri presumably occupied the Sanish swale at that time (deep-trench erosion cycle) and may have occupied it up to the Mankato glaciation. If so, the bedrock floor of the swale should lie at considerable depth, below a thick sequence of Pleistocene deposits. Simpson (1929, p. 177) reports a well in section 35, T. 151 N., R. 91 W., which obtains water from gravel at a depth of 67 feet. No other data on the depth of the fill are available.

The pre-Mankato drainage was completely obliterated north of the drift border, although some of the outwash channels probably coincide with buried valleys. For example, the outwash-floored swale north of Zahl, N. Dak., is comparable in size with Little Muddy Creek valley to the south, whereas the present narrow, west-trending headwater part of Little Muddy valley is not. The swale north of Zahl almost certainly represents the pre-Mankato continuation of Little Muddy valley.

The longest swale in the area of Mankato drift is that which starts above Daleview, in Sheridan County, Mont., and continues northward into Saskatchewan for 100 miles. The swale, which is occupied by a string of long, narrow lakes (fig. 38), is comparable in size with Big Muddy Creek valley south of Daleview. Evidence was presented earlier for believing that this swale and Big Muddy valley were formed as an ice-marginal channel prior to the early Wisconsin(?) glaciation. The Mankato drift has dismembered the channel but has not effaced it. The influence of the earlier middle Wisconsin(?) ice on the channel is unknown.

When the Mankato ice was at its maximum, streams of outwash flooded down favorably disposed valleys. The outwash in Little Muddy Creek valley has already been discussed. In Big Muddy Creek valley, two ages of outwash have been identified (Witkind, 1959, p. 28): an older outwash, exhumed from beneath early Wisconsin(?) till, and a younger outwash, overlying the till. The latter has been dated as Mankato on the basis of the accordance in level of its remnants with undoubted Mankato outwash near the town of Medicine Lake, Mont.

The most continuous and extensive terraces in Big Muddy valley are above Medicine Lake. A structure section across Big Muddy valley a few miles above its mouth has been prepared by geologists of the U.S. Bureau of Reclamation from drill records made in connection with a possible damsite. The section shows one, and possibly two, deposits of sand and gravel below the 100 feet of alluvium. It seems reasonable to assume that the uppermost of these, directly under the alluvium, is Mankato outwash. The sands and gravels, however, are shown up to the level of the flood plain on either side of the thick alluvial fill. If it were not for this, the situation would be precisely analogous to that in Little Muddy Creek valley where the Mankato outwash disappears downstream under the later alluvium. The Big Muddy drill records are susceptible to a variety of interpretations, however, depending on the assumed number, age, and stratigraphic relations of the sedimentary deposits. For example, it is conceivable that the sands and gravels which are shown at flood-plain level on either side of the thick alluvial fill are older valley-side remnants below the level of which the younger (Mankato?) sands and gravels were deposited. Unless the possibility of such alternative explanations is admitted, it must be concluded that the Mankato outwash coming down Big Muddy valley reached the Missouri above present river level, whereas that coming down Little Muddy valley reached the Missouri below river level. Furthermore, it would have to be assumed that more than 100 feet of erosion of the Mankato outwash took place prior to deposition of the alluvium in Big Muddy valley. It should be emphasized that the introduction of Mankato outwash in the Missouri valley at different levels by different streams is not impossible. It does, however, add complications which are circumvented by assuming analogous situations in Big Muddy and Little Muddy valleys. The problem merits additional investigation.

Little is known about the late Pleistocene history of the Poplar River valley. Only one doubtful occurrence of outwash has been observed, about 1 mile north of the Roosevelt County line (pl. 1). Here, below the gently sloping, graded till surface on the west side of the valley, is an ill-defined, low terrace about 100 yards wide. The gentle frontal slope of the terrace is covered by float with abundant glacial pebbles, but no exposures were found. Across the river, in a bank about 20 feet high, dominantly nonglacial gravels are exposed. These are mapped as possibly Crane Creek gravel (pl. 1). Perhaps the terrace on the west side is also underlain by Crane Creek(?) gravel, and the glacial float represents materials derived from till. In any event, except for this doubtful occurrence, no terraces of Mankato outwash have been observed along the length of the valley. Although Mankato outwash may be buried under valley-side slope wash, none is revealed in the few places where the Poplar River has undermined the side slopes. Loess is widespread, but it is so thin that it could hardly conceal prominent outwash terraces such as are found in other valleys which had access to Mankato melt waters. It is unlikely that the fine-textured silt-covered gravels of the present flood plain are Mankato outwash. They are found throughout the various branches of the Poplar River, even in branches such as the West Fork of the Poplar River which had no access to Mankato melt waters.

It seems possible, therefore, that the Poplar valley received little or no Mankato outwash. This is one reason why the lobe of Mankato drift north of Scobey, Mont., has been mapped as tentative only. Yet the sands of the dune field east of the Poplar valley near its mouth most likely came from deposits within the valley. Either Mankato or older outwash is buried under the alluvium, or the source materials have been removed by erosion.

In the eastern part of the map area, where the Mankato ice reached the Missouri River valley, the glacial terrace gravels are coarse and poorly sorted. Some may have been deposited marginal to masses of stagnant ice.

If the Mankato ice displaced the Missouri from the Sanish swale to Big Bend, the glacial sands and gravels which form the terrace remnant within Big Bend about 8 miles below Sanish, N. Dak., are Mankato outwash. The terrace remnant is about 2 miles long and a mile wide at the maximum and occupies a reentrant in the east valley wall. Its edge stands about 60 feet above the flood plain, which in turn is 20 feet or so above the low-water plain. The terrace surface rises toward the bluffs and bears a mantle of brown silt which may be partly slope wash. Sands and gravels are exposed below the silts in pits around the terrace edge. The sediments are loose, but locally oxidized. In some layers, the average pebble size is about 2 inches, but all variations down to sand size are present. Nowhere is the terrace mantled by till.

The writer has been unable to determine whether the terrace represents a remnant of a once-continuous valley fill or whether it formed as an isolated deposit between the valley side and a mass of stagnant ice within the trench. If the latter is true, it must be assumed that after the present trench was carved along the margin of the Mankato ice, the ice advanced into the newly formed trench and stagnated there. The writer has not observed till at low levels within the trench, but the survey of this part of the valley was hurried. Nor has the writer found evidence of slumping and subsidence, such as might have occurred when the wasting ice withdrew support from the overlapping glacial sediments. However, such slump features might have been confined to the outer part of the terrace which was subsequently destroyed by lateral shifting of the Missouri. The high degree of sorting and stratification in the deposit, however, is unusual for kame-terrace deposits.

Actually a valley-side origin for the terrace would remove a difficulty which besets the alternative explanation that the patch is a remnant of a continuous valley fill. The difficulty is that according to evidence presented earlier, the Mankato outwash was deposited below present river level farther up the Missouri valley. However, inasmuch as the Mankato ice probably blocked the Missouri valley between the White Earth River and Sanish, the history of the valley above the ice barrier need not be identical to that below. For example, the low-level outwash may have been deposited upvalley before the ice reached the Missouri between Sanish and White Earth River and while the Missouri River was still flowing through the buried valley of the Sanish swale. When the ice blocked the Missouri valley and the Sanish swale, it must have created a lake upstream. Some of the "older alluvium" upstream may include lacustrine deposits of this stage. After erosion of the present ice-marginal channel of the Missouri west of the Sanish swale, but while the ice still blocked the valley upstream, a valley train of coarse glacial debris may have been deposited in the present trench below the ice barrier.

An additional element which must be considered in the history of this part of the Missouri River valley is the sloping rock-cut bench within the present trench. The bench truncates the bedrock strata; it is not a structural feature. Its remnants are generally narrow and terminate at various elevations from 100 to 200 feet above the river. Possible explanations for future investigation are (1) the rock bench represents the floor of, or a terrace in, a valley which existed here before the glacial Missouri River came into existence (pl. 8); (2) the rock bench was eroded by side streams graded to the level of a Mankato outwash fill in the center of the valley; and (3) the rock bench was eroded by side streams graded to an ice fill in the valley. The writer is inclined to favor the first possibility. Perhaps the possibility should be reconsidered that the present trench, in spite of its unusual narrowness, is pre-Mankato in age and that the terrace dates from the middle(?) or even the early Wisconsin(?). Many of the granite and metamorphic pebbles of the terrace gravel crumble in the hand, but the disintegration is physical, not chemical, and takes place readily in certain rock types in environments where wetting and drying and freezing and thawing alternate rapidly.

In spite of the inconclusive observations noted above, the author is inclined to believe that the bend of the Missouri below Sanish is Mankato in age, that the bend itself was determined in large part by the fortuitously situated paths of tributaries of the Missouri when the Missouri was flowing through the Sanish swale, that the terrace is Mankato in age and is a remnant of a once-continuous outwash fill, and that it is probably unrelated to terrace remnants farther upvalley.

The White Earth River valley, because of its deep and winding course, was probably occupied by stagnant masses of ice during deglaciation. Although some of the valley deposits may have been laid down marginal to relic ice masses, the accordant elevations of the outwash terraces, particularly above Battleview, N. Dak., suggest that most of the deposits were laid down as a valley train. The 15 miles of valley between White Earth and Battleview were examined in only a few places, so that the apparent scarcity of glacial sands and gravels (pl. 1) may be more apparent than real.

For reasons presented earlier, the terrace silts of the region are believed to have been deposited during Mankato time, with deposition continuing beyond the glacial climax. Deposition of the Missouri silts may have been initiated by obstructions in the valley, such as moraines, outwash fans, ice lobes, or masses of stagnant ice. Aggradation in the Missouri valley would account for aggradation throughout its tributary system.

Stages in the withdrawal of the Mankato ice are recorded by ice-marginal channels and recessional till ridges (pl. 1). The patterns indicate that lobes of ice lagged behind in the valleys, but they fail to reveal to what extent stagnant ice littered the area. For a discussion of an interesting succession of ice-marginal channels in the vicinity of Appam, N. Dak., the reader is referred to Gott, Lindvall and Hansen (1947).

The Mankato ice modified the topography principally by deposition. In some flat upland areas, the wasting ice deposited a fairly uniform veneer of till, and the surface is still remarkably level. One such area is north of Wild Rose in northeastern Williams County, N. Dak. Over the greater part of the region between the drift border and the Missouri escarpment to the north, however, the Mankato ice left a complex assemblage of deposits which remain today virtually unmodified. The deposits form the pronounced morainal topography of the coteau belt. The abundance of eskers in some areas indicates that the Mankato ice locally stagnated while the ice border elsewhere withdrew northward.

Eventually the ice front withdrew north of the Missouri escarpment, leaving in its wake a level till plain. The general smoothness of the plain is probably due in part to the smooth bedrock surface below, in part to the even distribution of the till, and in part to smoothing of irregularities by deposition of glaciofluvial deposits. In some areas, however, the ground moraine displays a subdued morainal topography.

A system of channels across this northeast-sloping surface records stages in recession of the ice front. The channel pattern is, on a small scale, a replica of that between the Missouri escarpment and the glacial border 50 to 100 miles to the south. Wherever the ice front halted in its withdrawal down the gentle northeast slope, the northeast-draining streams were diverted along the ice front and they eroded ice-marginal channels. As the ice front receded to lower positions, the northeast-flowing streams prolonged their courses downslope, thereby linking the successive channels.

SUMMARY OF MANKATO GLACIATION

The Mankato ice crowded over the dissected border of the Missouri Plateau and extended 20 to 50 miles to the southwest. The greatest penetrations in that direction were made by lobes which, with one exception, followed favorably oriented lowlands. The largest of the lobes reached Medicine Lake in Montana.

The Mankato ice reached the Missouri River in the eastern part of the area and diverted the river from the Sanish swale to the Big Bend. The blocking of the Missouri may account for some of the lacustrine members of the terrace silts upstream.

Channels were eroded at the ice front, not only at the Mankato limit but during halts in retreat. The greatest influence of the Mankato ice on the topography, however, resulted from deposition. The plateau, including the dissected border, was heavily blanketed with drift, completely disorganizing the earlier drainage and giving rise to the coteau morainal belt. Valley trains were deposited in the Big Muddy, Little Muddy, and White Earth valleys, but it is not certain that a valley train was formed in the Poplar valley.

The cluttering of the floor of the Missouri valley with till and outwash and the obstructions offered by lingering ice lobes and masses of relic ice may have initiated the cycle of silt deposition.

POST-MANKATO TIME

Following withdrawal of the Mankato ice, the Missouri River cut a new channel below the terrace silts. The silts stand 10 to 15 feet above present flood-plain level in the center of the valley, but their base is not exposed. Hence, the depth of the post-Mankato trenching is unknown. A drill hole sunk by the U.S. Bureau of Reclamation on the floor of the Missouri River valley about 3 miles southeast of Williston (NW¼ sec. 4, T. 153 N., R. 100 W.) penetrated 50 feet of "clay" before reaching gravel. The surface elevation of 1,840 feet suggests that the hole was located on the low-water plain. The "clay" is probably the alluvium of the present report, but whether it includes the lower part of the section of terrace silts is impossible to say. Numerous other drill records present the same problem. The alluvium of Big Muddy Creek valley in Montana extends 100 feet below the

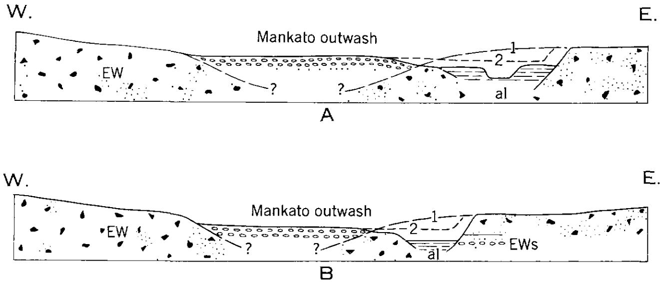


FIGURE 43.—Generalized sections across Little Muddy Creek valley, about 14 miles north of Williston, N. Dak. Section *A* is one-quarter of a mile above the mouth of East Fork of Little Muddy Creek; section *B* is two-thirds of a mile below East Fork. Little Muddy Creek was presumably superimposed on the till surface at the level of the Mankato outwash. Broken line 1 in each section represents the profile of the till surface prior to deposition of the Mankato outwash. Broken line 2 represents lateral erosion prior to downcutting to present stream level. *EWs*, Early Wisconsin(?) pre-till sands and gravels; *EW*, Early Wisconsin(?) till; *Al*, Alluvium of Little Muddy flood plain. Each section is about half a mile long.

present flood plain. How much of this is Recent alluvium, however, is unknown. The only conclusion possible at present is that the older silts were trenched to an unknown depth below present valley bottoms prior to deposition of the Recent alluvium. The latter forms an unbroken sequence from the surface of the flood plain down to the low-water plain of the Missouri.

Little Muddy valley contains evidence of post-Mankato erosion. The Mankato outwash in Little Muddy valley disappears under the valley-floor alluvium a short distance below the mouth of East Fork of Little Muddy Creek. For a few miles above this point, the present channel of Little Muddy Creek is eroded, not in the outwash but in the till alongside (fig. 43). Little Muddy Creek was presumably superimposed onto the till surface from the surface of the gravels. If the slope of the till was as shown by profile 1 (fig. 43, *A* and *B*), Little Muddy Creek could not have been superposed as far east as the present channel. Some side cutting at the level of the gravel fill, or at a halt during downgrading, must be assumed (fig. 43, *A* and *B*, profile 2). In any event, after deposition of the Mankato valley train, Little Muddy Creek eroded a new valley down to, and probably below, present stream level.

The flood-plain alluvium of the Big Muddy and White Earth valleys also occupies a trench eroded in Mankato outwash. The situation is particularly clear above Plentywood, Mont., in Big Muddy valley, and above Battleview, N. Dak., in the White Earth valley.

The post-Mankato trenching was followed by aggradation up to the present flood-plain level. The presence of intercalated soils(?) in the alluvial deposits of the Missouri River valley (fig. 32*B*) indicates that deposition was discontinuous. Perhaps the soils(?) are evidence of post-Mankato cycles such as have been described in Nebraska. The margin of the present flood plain, where it has not been impinged upon recently by the flood channel of the river, is generally mantled by slope wash or alluvial fans. Where the river swings laterally into these deposits, terracelike features of noncyclic significance result. Locally, the river has eroded laterally to the heads of the fans, leaving only a fill in the valley beyond. Headward erosion, starting at the mouth of the valley, results in a paired terrace in the valley above, but this terrace is also without cyclic significance.

As for the coteau morainal belt, the short span of post-Mankato time, probably less than 10,000 years, has resulted in very little modification of the topography. Eskers, kames, valley trains, kettles, and channels are still remarkably fresh. Integrated drainage is found only where deposition of the Mankato drift failed to obliterate pre-Mankato valleys, or locally at the margins of the drift border. Some of the morainal lakes, however, have been converted to meadows. In the area of the ground moraine, north of the coteau belt, post-Mankato dissection is largely restricted to the neighborhood of the Missouri escarpment and to the margins of glacial channels. The overall impression in the area of Mankato drift is one of extreme topographic youth.

No direct evidence of postglacial warping, such as tilted shorelines, has been found. The asymmetric position of the master streams in their valleys may, however, be symptomatic. The Missouri hugs the south side of its valley for long distances, and the broadest remnants of its former valley floors at both the Cartwright and Flaxville levels are north of the river. The situation along the Yellowstone is similar, except that the river hugs the east side of its valley and the remnants of the older and higher valley floors are almost entirely to the west.

To displace the Missouri to the south and Yellowstone to the east by tilting, the area would have had to be raised in the northwest. It seems unlikely that such tilting could be due to deglaciation because the ice center was to the north or northeast, and the Missouri and Yellowstone were already laterally displaced in preglacial time, at the Flaxville level.[17] If tilting is involved, therefore, it is probably due at least in part to causes other than glacial readjustment.

Alternative explanations include (1) deflection of the drainage to the right, owing to rotation of the

[17] The second part of this argument would be negated should it be demonstrated that development of the Flaxville surface continued into Pleistocene time.

earth (Ferrel's law); (2) excess deposition by tributaries from one side; (3) homoclinal shifting; and (4) combinations of factors.

The extent, if any, of deflection due to terrestrial rotation seems impossible to evaluate. Excess deposition from one side cannot be the complete explanation. Even though glacial streams entering the Missouri from the north undoubtedly brought in much glacial sediment and may have displaced the river southward, these glacial sediments are buried under younger silts. There is no reason to believe that more silt was brought in from the north than from the south, for there is no significant contrast in number and size of the streams north and south of the Missouri. In any event, outwash-laden streams could not have displaced the Yellowstone laterally because the tributaries did not head at the ice front and those on one side had no advantage over those on the other side. Finally, glacial events cannot explain the asymmetrical location of both the Missouri and Yellowstone on the Flaxville surface. Homoclinal shifting is an inadequate explanation because the master streams are independent of structure, except locally.

Perhaps the displacements of the Missouri and Yellowstone are not representative of the region as a whole. Study of many other valleys both in and out of the glaciated area may assist in appraising the various possibilities. For example, if the streams of the region are consistently displaced to the right regardless of their direction of flow, then deflection due to terrestrial rotation would seem an important influence. If, on the other hand, the streams display a favored azimuth of displacement, then tilting would be more likely. The problem merits further study.

JONES CUT, AN EXPOSURE OF MULTIPLE DRIFTS

Jones cut is a deep excavation along the Great Northern Railway about 5½ miles southwest of Williston and about half a mile west of Lewis and Clark Bridge over the Missouri (pl. 1 and fig. 22). The exposure is one of the most informative yet easily accessible of those in which multiple drifts are displayed.

Alden made no reference to this exposure in his 1932 report, in which he suggested that only 1 drift (Illinoian or Iowan) was present in the region between the glacial limit and a point about 20 miles north of Williston. In 1935, Alpha [18] reported that 2, and possibly 3 drifts were present in the southeast face of the cut. Alpha recognized that the uppermost material might be waste from the excavation, hence he suggested only that the middle stratum was Alden's Illinoian or Iowan till and that the lower stratum was Kansan in age. The relations exposed here support evidence from elsewhere indicating that 2 drifts are present. The 2 drifts, however, are Wisconsin in age.

Because part of the evidence for multiple drifts is displayed in this cut, the writer conducted 5 or 6 parties of glacial geologists and 1 party of soil scientists to the locality. The exposure will undoubtedly be visited many times in the future. As a matter of record, therefore, the stratigraphy of this cut is herein described in detail.

STRATIGRAPHY OF THE SOUTHEAST FACE

The stratigraphy of the southeast face of Jones cut is shown in figure 23. The uppermost till-like material, above the prominent buried soil profile, forms a hummocky topography on each side of the cut. Alpha [19] suspected that this material was waste from the railroad cut, but he did not discard the possibility that it might be till. That the deposit is not till became evident early in the present investigation because of the following observations:

1. The deposit lacks a soil profile and there is no obvious caliche zone. Both are well developed even in postglacial alluvium.
2. The hummocky topography is restricted to the immediate neighborhood of the cut.
3. Calcium carbonate crusts are restricted to the undersides of pebbles in the till below the prominent buried soil horizon, whereas in the hummocky deposit the pebbles are in haphazard orientation.
4. The stratigraphy of the hummocky deposit above the prominent soil horizon is a vertical reflection of that below. At one place, for example, about 2 inches of white lime-rich material, similar to that below the prominent soil, rests directly on this soil. The lime-rich material is followed upward by 2 to 3 feet of grayish cobbly "till" similar to that in the section below the prominent soil. The cobbly material is in turn followed by a brownish "till" similar to that forming the bluffs below the gray till. In each of the till-like deposits above the soil, the carbonate-crusted pebbles are in random orientation and are concentrated near the base of each deposit, in contrast with their position near the top of the actual tills themselves.

These facts suggest that the cut was excavated originally from the top rather than from the ends, and that part of the waste was piled on the surface. This would account for the reversal of the stratigraphic sequence in the waste deposits. That the cut was excavated as

[18] Alpha, A. G., 1935, Geology and ground-water resources of Burke, Divide, Mountrail, and Williams Counties in North Dakota: N. Dak. Univ., unpub. thesis, 63 p.

[19] Op. cit.

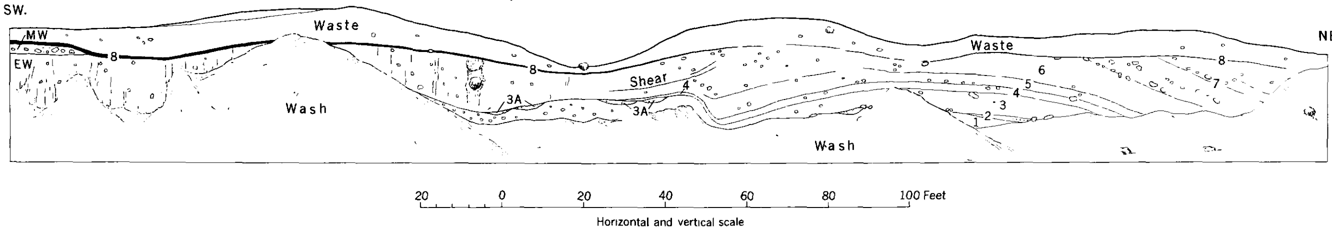


FIGURE 44.—Part of the northwest face of Jones cut (fig. 22, *B–C*). The entire section below the waste probably consists of a single drift, the early Wisconsin(?), complicated by slumping. Numbers are explained in text. Vertical scale is the same as the horizontal scale.

surmised was partially verified through correspondence with Mr. H. J. Seyton, chief engineer of the Great Northern Railway Company. Mr. Seyton advised (written communication, June 5, 1946) that the cut was originally excavated in 1887 with men and teams working down from the top and that much of the material was deposited to either side. Mr. Seyton understands, however, that there were hummocks on the surface before excavation was begun, but he knows nothing of their origin. The hummocks referred to may be the undulations underlying the waste shown in figure 44. These are clearly erosional phenomena antedating the soil beneath the waste. Mr. Seyton further reports that when the cut was widened for double tracking in 1929, the north face was cut back by steam shovel at track level and the debris was hauled away by trains.

There is thus little doubt that the debris above the prominent soil horizon is waste.

The prominent dark band below the waste (fig. 23) represents the almost thoroughly leached A horizon of a pebbly soil. The B horizon, rich in calcium carbonate and light-gray to white in color, is largely restricted to the sandy gravel immediately below. The pebbles of the gravel average about one-half inch in size, although some as much as 2 inches in size are present. Locally, below the gravel, there is a layer of medium to coarse sand as much as 1½ feet thick.

Beneath the sands is the middle Wisconsin(?) till of this report. The latter averages between 5 and 6 feet thick, is gray at the surface and buff within, is silty in texture, and contains a few seams of silt and sand, some of which are more than 1 inch thick. The deposit contains many cobbles and boulders, some of large size. Because of the low clay content, the surface is only locally mud cracked and the cracks are from several inches to more than a foot apart. The deposit is characterized by a small-scale wavy parting. A pebble analysis of the till appears as No. 43 in table 1.

Below the middle Wisconsin(?) till, and forming vertical bluffs, is the early Wisconsin(?) till, compact, brown, clayey, its surface characterized by an intricate lacework of hackly shrinkage polygons that range from a fraction of an inch to several inches in diameter. The surface is case hardened by lime carbonate, so that the collection of a hand specimen or the loosening of a pebble is extremely difficult. The deposit is vertically jointed, the joints being spaced from 1 foot to less than an inch apart. The till contains only scattered pebbles, most of them less than 4 inches in size although a few large boulders appear. Locally, tabular pebbles lie flat. This may indicate accretion of the deposit by the plastering process which gives rise to lodgement till.[20] A pebble analysis of the till appears as No. 44 in table 1. The soil at the top of this till has been described earlier. It consists locally of an A horizon about 1 foot thick and a B horizon 2 to 3 feet thick. Elsewhere the A horizon is missing.

STRATIGRAPHY OF THE NORTHWEST FACE

The geology of the northwest face of Jones cut (fig. 44) is more complicated than that of the southeast face, and the relations are not clearly understood. The prominent soil horizon (8) and the underlying middle Wisconsin(?) and early Wisconsin(?) tills in the extreme left of the figure are identical to those across the cut. They were obviously continuous before the cut was excavated. For several hundred feet southwest of the area shown in figure 44, the prominent soil horizon is remarkably level. Within the confines of the figure, however, it caps a rolling topography beneath the waste cover. Here, too, the soil is locally thin, churned up, or absent, probably as a result of the operations attending the transfer of the waste away from the cut.

The deposits shown in the right (northeastern) half of figure 44 are coarsely and indistinctly stratified. They form a broad, low anticline, the strike of which is unknown. Horizons 1, 3, and 5 are similar and consist of tough vertically jointed mud-cracked oxidized but unleached clayey till with scattered pebbles and cobbles. Slickensides are present in horizon 1, and there is a seam of gray clay about 1 foot above the

[20] The possibility that the pebbles were oriented during compaction, as mica flakes in the clay-shale transition, could be eliminated if a pebble orientation study revealed a consistent parallelism of the long axis of the pebbles.

base of horizon 3. Horizon 2 consists of a few inches to 2 feet of interbedded clay and sand. The clay ranges in color from dark gray to buff. The sands are friable and coarse and locally pebbly. Pebbles, cobbles, and boulders are scattered throughout this layer. Horizon 4 is the most persistent layer in the deposit. It is a 1-foot bed of coarse well-consolidated buff sand with many pebbles. The latter are generally 1 inch or less in size, but some as much as several inches in size are present. At least one boulder is included within the layer, but most boulders are below it or rest on it. Some of the boulders of the underlying till project up into the sand. The till of horizon 6 is superficially similar to that of horizons 1, 3, and 5, but it lacks the mud-cracked surface and the vertical joints, and seems to have fewer pebbles, cobbles, and boulders. The absence of mud cracks may indicate a higher silt content than the underlying till facies. The till of horizon 7 is more bouldery than that of horizon 6, and it is jointed; it contains several seams of dark clay and thin stringers of sand which contribute to the bedded appearance. In addition, pebbles and cobbles are strung out parallel to the bedding, and the long directions of the exposed pebbles are parallel to the dip. The anticline is truncated by the irregular erosion surface marked by the gravelly soil (8), which is more prominent on the left side of the figure where it mantles the middle Wisconsin(?) till.

In the central part of the exposure, the 1-foot layer of compact buff sand (horizon 4) thins abruptly to about 1 inch on the flank of a steep minor fold and pinches out a short distance to the southwest. Near its termination, this compact sand locally overlies a loose crossbedded sand (3A). The contact relations differ considerably: the compact buff sand may rest directly on the crossbedded sand below; it may be separated from the loose sand by 1 to 2 inches of till similar to horizons 1, 3, and 5; or it may be interbedded or interfolded with the loose sand. Patches of the loose sand are found for 50 feet farther west. A similar relationship between 2 tills and patches of intervening sand was observed in a roadside ditch in north-central Williams County about 7 miles south of Alamo, N. Dak. (pl. 5, loc. 61–62).

The deposit to the left of the minor flexure and above the swelling and pinching friable sand is similar to the till of horizon 6 to the east. Only locally does it show mud cracks, and its pebble content is less than that of the till below. Many of the tabular pebbles lie flat. The till below the friable sand is similar to that of horizons 1, 3, and 5 to the east. A pebble analysis of the till above the friable sand horizon is listed as No. 45 in table 2, and the analysis of the lower till is listed as No. 46. The samples agree closely in the proportions of 2 out of 3 of the glacial erratic types (limestone and dolomite and granite), but they disagree on the third (other plutonic types). However, single pebble analyses do not always supply a true picture of the lithology of a deposit.

It is interesting to note that the till above the friable sand, in the area between the minor fold and the long slope of wash, differs from the bluff-forming till across the railroad cut (fig. 23) in its siltier texture, as indicated by the scarcity of mud cracks. It is the till below the friable sand which more closely resembles the bluff-forming till across the way. Yet the sequence of deposits at the extreme left of figure 44 is an exact duplicate of the section in the southeast face of the cut. If the railroad cut, instead of cutting obliquely through the spur shown in figure 22, had been excavated at right angles to the trend of the spur, the deposits on the two sides of the cut would correspond far more closely. The long slope of wash on the left side of figure 44 conceals information which is necessary to a correct understanding of the relations between the deposits on either side. It does seem clear, however, that the section has been deformed. This is indicated not only by the major anticline with its continuous 1-foot-thick sand layer, but by the slickensides in horizon 1, the asymmetric minor fold with a dip of 60° on the east flank, and the shear above the minor fold. The direction of asymmetry of the minor fold and the attitude of the shear above suggest that the movement involved in the deformation was from the southwest. The ice, on the other hand, came from the northeast. Unless, therefore, one can demonstrate that underthrusting has occurred, the reasonable interpretation is that the deformation is due to slumping.

In the writer's opinion, the following deposits represent the upper part of an original thick till deposit: at least the upper deposits of the broad anticline of figure 44: the till above the friable sands west of the minor fold in figure 44; and facies 2 of figure 24. The exposure shown in figure 24 is just southwest of the limits of figure 44.

The lower part of the original deposit is believed to be represented by the following: the lower strata of the broad anticline; the till below the friable sands; facies 1 of figure 24; and the bluff-forming till in the southeast face of Jones cut.

The conditions of deposition that caused the stratification of the deposits in the broad anticline are unknown. Except for the persistent 1-foot sand layer, the stratification is faint, and is due to the following:

slight textural differences which affect the appearance of the weathered surfaces; presence of seams of clay; variations in boulder content; and parallel arrangement of lines (planes?) of boulders. Whatever the cause of the stratification, it was local, because the equivalent deposits west of the anticline are not stratified except for the gross contrast above and below the unconformity marked by the friable sands.

In summary, the complex sequence of deposits below the middle Wisconsin(?) till in the northwest face of Jones cut is believed to represent a single till, the early Wisconsin(?) till. The complications are believed to be due to a stratification, the cause of which is unknown, and to large scale landsliding and flowage. Part of the movement may have been along the zone of friable sand at the unconformity.

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