MAP SHOWING SPATIAL AND TEMPORAL RELATIONS OF MOUNTAIN AND CONTINENTAL GLACIATIONS ON THE NORTHERN PLAINS, PRIMARILY IN NORTHERN MONTANA AND NORTHWESTERN NORTH DAKOTA

By
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INTRODUCTION

This report is an overview of limits of glaciations and glacial history in, and east and southeast of, Glacier National Park, Mont., and on the Northern Plains farther east in Montana and northwestern North Dakota (lat 47°–49° N., long 102°–114° W.). Glacial limits east of long 102° W., in the United States and also in adjacent Canada, are shown on published maps of the U.S. Geological Survey Quaternary Geologic Atlas of the United States (I–1420) (for example, Fullerton and others, 1995, 2000). The glacial-advance limits shown on this map are from data compiled for the Lethbridge, Regina, Yellowstone, and Big Horn Mountains 4° × 6° quadrangles. Limits of Laurentide glaciations shown on the map supersede those mapped by Colton and others (196l, 1963) and Soller (1993, 1994). An abbreviated version of this report is presented in Fullerton and others (2004).

This pamphlet is an expanded explanation of the map and supplemental illustrations. Reference to the map is required to follow this discussion. However, the reader can visualize spatial and temporal relations of the glacial limits and patterns of regional ice flow simply by study of the map and figures 1 and 2. The map and figures 1 and 2 depict ice-flow patterns and selected glacial limits in a very large geographic region. Comprehensive discussion of data from this region is beyond the scope of this report. In this pamphlet, we discuss selected glacial limits and selected stratigraphic, geomorphic, and sedimentologic data in a spatial and temporal context. In general, discussion of specific areas under a temporal subheading (for example, "Illinoian Glaciation") proceeds from west to east.

This pamphlet is a summary of published and unpublished information from field studies in Montana and northwestern North Dakota by D.S. Fullerton and R.B. Colton. Discussions of stratigraphic, spatial, and temporal relations of tills in Montana and North Dakota are primarily based on field observations, secondarily on regional synthesis. Discussion of these complex relations is intended for a scientific audience. Descriptions of physical characteristics of tills, discussion of stratigraphic, spatial, and temporal relations of specific till units, and discussion of the history of glaciation are provided for the reader who is interested in the glacial deposits and history in one or more specific geographic or physiographic areas and who also has some prior knowledge of Quaternary deposits and history in the region. We summarize distinguishing field characteristics of Laurentide tills on the plains in Montana in the appendix.

Depiction and discussion of selected glacial limits in adjacent Canada is necessary to summarize the history of glaciation in Montana and North Dakota. Discussion of geologic relations in Canada is restricted to topics relevant to interpretation of Quaternary stratigraphy, chronology, and history in Montana and North Dakota.

The term "Laurentide glacier" was applied by Chamberlin (1895) to a continental ice sheet east of the Rocky Mountains in North America. Flint (1943) referred to the same body of ice as the "Laurentide ice sheet." Fulton and Prest (1987) and Dyke and others (1989) proposed that the name "Laurentide ice sheet" be applied only to the ice sheet of Wisconsin age. Dyke and others (1989, p. 184) suggested that ice sheets of Illinoian and pre-Illinoian ages be referred to as "Laurentide ice," but not "Laurentide ice sheets." "Laurentide ice sheet" here refers to any Quaternary continental ice sheet east of the Rocky Mountains in the United States and Canada. "Laurentide till" refers to till deposited by a Laurentide ice sheet. A Laurentide ice sheet here is distinguished from a Cordilleran ice sheet (Chamberlin and Salisbury, 1906) in the Cordilleran region in parts of Washington, Idaho, and Montana in the United States and in adjacent Canada. Clague (1989) indicated that Cordilleran ice sheets formed several times during the Pleistocene. He did not restrict the term "Cordilleran ice sheet" to an ice sheet of Wisconsin age.

Numerical ages cited in reference to the astronomically tuned marine oxygen isotope time scale are from Martinson and others (1987), Bassinot and others (1994), Berger and others (1994), and Chen and others (1995). Most of the conventional radiocarbon (^{14}C) ages cited also are given as calibrated ages (CAL ages). The CAL ages (Bard and others, 1990a,b) cited were calculated by using the simple linear equation for calibration in Bard and others (1998):

$$CAL\ yr = 1.168\ [^{14}C\ yr] \qquad (1)$$

We summarize time terminology in table 1. Chronologic controls for selected marine oxygen isotope stages and marker horizons mentioned in the text are given in table 2.

Table 1. Time terminology

TIME SCALE (SIDEREAL YEARS)	INFORMAL NOMENCLATURE BASED ON MOUNTAIN GLACIATION	INFORMAL NOMENCLATURE BASED ON CONTINENTAL GLACIATION	INFORMAL TIME NOMENCLATURE	FORMAL TIME NOMENCLATURE	
≈11,680 yr —	—?—?—?—?—?—			HOLOCENE	QUATERNARY
≈35,000 yr —	PINEDALE	LATE WISCONSIN	LATE PLEISTOCENE	PLEISTOCENE	
≈55,000 yr —	—?—?—?—?—?—	MIDDLE WISCONSIN			
≈80,000 yr —		EARLY WISCONSIN			
≈128,000 yr —	—?—?—?—?—?—	SANGAMON			
≈310,000 yr —	BULL LAKE	ILLINOIAN	MIDDLE PLEISTOCENE		
≈778,000 yr —	—?—?—?—?—?—	PRE-ILLINOIAN	EARLY PLEISTOCENE		
≈1,806,000 yr —	PRE-BULL LAKE			PLIOCENE	TERTIARY

Table 2. Chronologic controls of selected marine oxygen isotope stages and marker horizons
[AST, astronomically tuned age; AA, $^{40}Ar/^{39}Ar$ age; yr, years; ≈, approximately]

Oxygen isotope stage or marker horizon	Age
2	AST ≈11,000 to ≈24,000 yr[1] AST ≈12,050 to ≈24,110 yr[2]
4	AST ≈71,000 to ≈57,000 yr[1] AST ≈73,900 to ≈58,960 yr[2]
6	AST ≈186,000 to ≈127,000 yr[1] AST ≈189,000 to ≈129,840 yr[2] AST ≈190,000 to ≈130,000 yr[3]
8	AST ≈242,000 to ≈301,000 yr[1]
12	AST ≈474,000 to ≈427,000 yr[1] AST ? to ≈427,000 yr[3]
14	AST ≈568,000 to ≈528,000 yr[1]
16	AST ≈659,000 to ≈21,000 yr[1] AST ? to ≈623,000 yr[3] Lava Creek B tephra—airfall during late stage 16: AA 639,000 ± 2,000 yr[4]
18	AST ≈760,000 to ≈712,000 yr[1] AST ? to ≈714,000 yr[3]
19	Matuyama-Brunhes geomagnetic polarity reversal during early stage 19: AST 775,000 ± 10,000 yr[1] AST 777,900 ± 1800 yr[6] AST 778,800 ± 2500 yr[6] AA 778,000 ± 4000 yr[7] AA 778,000 ± 10,000 yr[8,9] AA 778,200 ± 3500 yr[6] AA 778,700 ± 1900 yr[10]
20	Termination: AST ≈787,000 yr[1] AST ≈790,000 yr[3] Isotopic maximum: AST ≈793,000 yr[1] Midpoint: AST ≈794,000 yr[5]
22	AST ≈900,000 to ≈867,000 yr[3] AST ? to ≈865,000 yr[1]
65	Pliocene-Pleistocene boundary, in stage 65 sediments: AST ≈1,806,000 yr[11,12]
78	Midpoint: AST ≈2,078,000 yr[5]
82	Midpoint: AST ≈2,146,000 yr[5] Matuyama-Réunion geomagnetic polarity reversal during stage 82 (Sarnthein and Tiedemann, 1990): AST ≈2,149,000 yr[11] AST ≈2,150,000 yr[13] AA 2,150,000 ± 30,000 yr[14] AA 2,150,000 ± 40,000 yr[15] AA 2,190,000 ± 9000 yr[16]
104	Gauss-Matuyama geomagnetic polarity reversal during stage 104: AST ≈2,582,000 yr[11]

[1]Bassinot and others, 1994
[2]Martinson and others, 1987
[3]Berger and others, 1994
[4]Lanphere and others, 2002
[5]Chen and others, 1995
[6]Tauxe and others, 1996
[7]Pringle and others, 1995
[8]Spell and McDougall, 1992
[9]Renne and others, 1994
[10]Singer and Pringle, 1996
[11]Lourens, Antonarakou, and others, 1996
[12]Lourens, Hilgen, and others, 1996
[13]Hilgen, 1991
[14]Baksi and others, 1993
[15]McDougall and others, 1992
[16]Honey and others, 1998

Rank terms of formal stratigraphic units are capitalized (for example, Kennedy Formation; Medicine Hill Formation). Rank terms of informal stratigraphic units are not capitalized (for example, Havre till; Perch Bay till). We retain, as informal nomenclature, the stratigraphic names of Laurentide glacial deposits (tills) on the plains of Montana proposed by Fullerton and Colton (1986) (for example, Loring till; Crazy Horse till) (table 3). Those tills are allostratigraphic units (North American Commission on Stratigraphic Nomenclature, 1983), not lithostratigraphic units. Revisions of the stratigraphic and chronologic framework of the tills in Montana proposed by Fullerton and Colton are included in table 3. Stratigraphic units related to mountain glaciation in the Glacier National Park region (Kennedy till; Bull Lake till; Pinedale till) also are informal allostratigraphic units. The terms "Bull Lake" and "Pinedale" (tables 1 and 3) refer to informal time divisions represented by mountain glacial deposits. "Bull Lake till" refers to all mountain till in the Rocky Mountains deposited during the Bull Lake glaciation. "Bull Lake glaciation" is the time interval during which Bull Lake Till (a formal stratigraphic unit) was deposited in the type area in the Wind River Mountains in Wyoming (Richmond, 1965a). In the same sense, "Pinedale till" refers to all mountain till in the Rocky Mountains deposited during the Pinedale glaciation in the Wind River Mountains. "Early Pinedale till" refers to till deposited during the early stade of the Pinedale glaciation. Names of tills in Canada are not included in table 3.

The base for the map is simplified. Selected hydrographic features, selected towns and cities, selected physiographic features, and a grid of $1^{\circ} \times 2^{\circ}$ topographic quadrangles are included to aid the reader in location of the glacial limits and other features depicted here on other maps at different scales. Most of the geologic data were compiled at 1:250,000 scale.

VALLEY GLACIATION AND ICE-FIELD GLACIATION (MOUNTAIN GLACIATION)

Mountain glaciers and continental ice sheets interacted during the Bull Lake (≅Illinoian) and Pinedale (≅late Wisconsin) glaciations (figs 2 and 3, tables 1 and 3). Glacial lakes were dammed between mountain glacier and continental ice margins in some areas; the ice masses merged in some other areas. Stratigraphic and geomorphic relations of tills and glacial lake deposits east and southeast of Glacier National Park, Mont., uniquely document the temporal and spatial relations of the Bull Lake and Pinedale mountain glacial maxima and the Illinoian and late Wisconsin continental glacial maxima.

Piedmont glacial lobes fed by mountain glaciers and ice fields in the Rocky Mountains extended eastward and northward onto the plains in Montana and Alberta. Those derived primarily from coalesced valley glaciers here are referred to as "valley-glacier piedmont lobes." Lobes that were primarily outlet glaciers from the northern Montana ice field are referred to as "outlet-glacier piedmont lobes." The distinction is necessary because late Pinedale valley-glacier deposits (unit LP on the map) are younger than "late Pinedale" outlet-glacier deposits (unit P3).

VALLEY GLACIERS

The valley glaciers depicted on the map were confined to Glacier National Park and adjacent areas east of the Continental Divide in Montana. The limits of the valley glacier advances are from unpublished mapping by G.M. Richmond and D.S. Fullerton.

OUTLET-GLACIER PIEDMONT LOBES

The piedmont ice lobes southeast of Glacier National Park were primarily outlet glaciers from the northern Montana ice field in the Rocky Mountains. The "northern Montana ice cap" (Alden, 1932, 1953; Richmond and Mudge, 1965; Richmond, 1986a; Carrara, 1989; Locke, 1995) was an ice field that formed as a result of coalescence of valley glaciers east and west of, and across, the Continental Divide in the mountains west of the piedmont lobes shown on the map. Richmond (1986b) referred to the ice field as a "coalescent alpine glacier complex." The largest outlet glacier, represented by till of units P1, P2, and P3 in the Cut Bank $1^{\circ} \times 2^{\circ}$ quadrangle, funneled ice northeastward from Marias Pass to East Glacier Park village, and then northeastward and eastward as the Two Medicine piedmont lobe. The Birch Creek piedmont lobe was an independent ice lobe south of the Two Medicine lobe (Calhoun, 1906; Alden, 1932). However, because the lobes were confluent during the Pinedale 1 (P1) and Pinedale 2 (P2) glacial maxima, the term "Two Medicine lobe" here refers to the combined lobes, as illustrated on the map and in figures 1 and 3. The limits of outlet-glacier advances (limits of units P1, P2, P3) are from unpublished mapping by D.S. Fullerton and R.B. Colton.

KENNEDY GLACIATIONS

Willis (1902) used the term "Kennedy gravels" in reference to gravel of inferred fluvial origin. Alden (1932) concluded that some of the deposits described by Willis are glacial and glaciofluvial in origin, and he used the term "Kennedy drift" in reference to all of the glacial, glaciofluvial, and nonglacial fluvial deposits. Alden (1912, 1914, 1932), Alden and Stebinger (1913), Martin (1927), Horberg (1954, 1956), Karlstrom (1981, l987a, c, 1988a, 1991), and du Toit (l988) described till, glaciofluvial deposits, and nonglacial fluvial deposits on upland erosion surfaces in and near Glacier National

Table 3. Correlation of glacial deposits in Montana and North Dakota.

[Letters and numbers in parentheses, for example P3, KD, LW, correspond to units on colored map; Fm. = Formation; double line represents interglacial desposits]

<table>
<tr><th rowspan="2">Age of glacial maximum (sidereal yr B.P.)</th><th colspan="2">Mountain glaciation</th><th colspan="5">Continental glaciation</th></tr>
<tr><th>Piedmont outlet glacier lobes from northern Montana ice</th><th>Piedmont valley glaciers and valley glacier</th><th>Glacier National Park region</th><th>Western Montana plains</th><th>Central Montana plains</th><th>Eastern Montana plains</th><th>Northwestern North Dakota</th></tr>
<tr><td>≈17,580 —</td><td></td><td>Late Pinedale till (LP)</td><td rowspan="3">Boundary Creek till (LW)</td><td rowspan="3">Fort Assinniboine till (LW)</td><td rowspan="3">Loring till (LW)</td><td rowspan="3">Crazy Horse till (LW)</td><td rowspan="3">Till of Snow School Fm.;• "unit F" (LW)</td></tr>
<tr><td></td><td>"Late Pinedale till" (P3)</td><td rowspan="2">Middle Pinedale till (MP)</td></tr>
<tr><td>≈23,360 —</td><td>"Middle Pinedale" till (P2)</td></tr>
<tr><td>≈26,000 —</td><td>Early Pinedale till (P1)</td><td>Early Pinedale till (EP)</td><td colspan="5"></td></tr>
<tr><td>≈140,000 —</td><td></td><td>Late Bull Lake till (LB)</td><td>Emigrant Gap till (IL)</td><td>Heron Park till (IL)</td><td>Markles Point till (IL)</td><td>Kisler Butte till (IL)</td><td>Till of Horseshoe Valley Fm.; "unit E"</td></tr>
<tr><td>≈160,000 —</td><td></td><td>Early Bull Lake till (EB)</td><td colspan="5"></td></tr>
<tr><td>≈648,000 —</td><td></td><td>Till of Kennedy Fm. (KD)</td><td></td><td>Upper unit of Havre till (PI)</td><td>Upper unit of Perch Bay till</td><td>Upper unit of Archer till</td><td>Upper till of Medicine Hill Fm.; "unit B"</td></tr>
<tr><td>≈720,000 —</td><td></td><td>Till of Kennedy Fm. (KD)</td><td></td><td>Lower unit of Havre till</td><td>Lower unit of Perch Bay till</td><td>Lower unit of Archer till</td><td>Lower till of Medicine Hill Fm.</td></tr>
<tr><td>≈800,000 —</td><td></td><td>Till of Kennedy Fm. (KD)</td><td colspan="5"></td></tr>
<tr><td>≈875,000 —</td><td></td><td>Till of Kennedy Fm. (KD)</td><td colspan="5"></td></tr>
<tr><td></td><td></td><td>Tills of at least three late Pliocene glaciations</td><td colspan="5"></td></tr>
</table>

Park, Mont., and contiguous Waterton Lakes National Park, Alta. Horberg (1954, 1956) referred to the glaciogenic deposits as “Kennedy drift” and “Kennedy mountain drift.” All of the mountain till and associated stratified deposits of pre-Bull Lake age in the Cut Bank $1^{\circ} \times 2^{\circ}$ quadrangle here are assigned to the Kennedy Formation. The till is referred to informally as “Kennedy till.”

The surface topography on Kennedy till is very subdued. Lateral moraines were modified greatly by erosion and mass-movement processes. Most of the Kennedy drift (unit KD) shown on the map and in figure 1 is eroded till. Many small areas of till are not shown. The eastern limit of till (unit KD) east of the Saint Mary River, as shown on the map and in figure 1, is not everywhere the limit of pre-Bull Lake glaciation; in some areas, all of the till was removed by erosion. Data published by Karlstrom (1981, 1988a, 1991) indicate that the Kennedy drift in Montana and Alberta includes tills deposited during at least six pre-Bull Lake mountain glaciations. Three of those glaciations were late Pliocene in age (table 3).

Gravel on high bedrock benches (erosion surfaces) far east of the limit of Kennedy till on the map was interpreted by Karlstrom (1981, 1991), Richmond (1986a, and unpub. mapping), and du Toit (1988) to be outwash (glaciofluvial) deposits related to Kennedy glaciations. We interpret most of that gravel to be Pliocene (and older?) nonglacial alluvium. Gravel on the drainage divide on the east side of the Saint Mary River valley was interpreted to be outwash associated with the “lower till of the Kennedy Formation,” of early middle Pleistocene age (Richmond, 1986a). Samples from a lens of silt in the gravel had normal remanent geomagnetic polarity (Richmond, 1986a). The gravel has no plausible spatial or temporal relation to any of the Kennedy till units on upland surfaces; it probably is nonglacial alluvium deposited earlier than 2.6 Ma, during the Gauss Normal Polarity Chron (table 2). Most (or all) of the “pre-Illinoian alpine till on Number 1 Bench” (Kennedy till) mapped by Karlstrom (2000a, fig. 1) south of the North Fork Milk River in Montana (the river is shown in fig. 3) we interpret to be nonglacial fluvial gravel of Tertiary age.

Richmond (1957) described multiple till units and intervening buried soils in the St. Mary Ridge stratigraphic section (fig. 1) high on the east wall of the Saint Mary River valley southeast of Babb, Mont. In a subsequent reference to that exposure, Richmond (1965b) concluded that three pre–Bull Lake tills (which he referred to as Washakie Point Till, Cedar Ridge Till, and Sacagawea Ridge Till) and three buried soils were overlain by Bull Lake Till.[1] An additional stratigraphic unit at the base of the section, referred to as till 1 by Karlstrom (1981, 1988a, 1991), was not exposed when the landslide-headwall exposure was examined by Richmond. The Bull Lake till in the St. Mary Ridge exposure is Karlstrom’s till 5. Richmond (1986a) assumed that the three pre–Bull Lake tills he observed in the exposure were laterally equivalent to weathered tills on upland surfaces, described by Alden (1912, 1914, 1932), Alden and Stebinger (1913), Martin (1927), and Horberg (1954, 1956), and he assumed that there were only three Kennedy tills on the upland surfaces. Consequently, he designated the three pre–Bull Lake tills he observed in the St. Mary Ridge section as the lower, middle, and upper tills of the Kennedy Formation.

There is no demonstrated spatial or temporal relation of any of the pre–Bull Lake tills in the St. Mary Ridge stratigraphic section to any specific Kennedy till unit on upland erosion surfaces in Montana and Alberta. All of the tills in the exposure were deposited by valley glaciers long after the present northward course of the St. Mary River was initiated by piracy. Some of the Kennedy tills on upland surfaces were deposited by valley glaciers or piedmont glaciers prior to the drainage diversion; they are much older than the tills in the exposure. Karlstrom (1981, 1988a, 1991) identified five diamictons in the St. Mary Ridge section. The genesis of his till 1 is not certain. The intensely weathered diamicton in some places possibly is nonglacial alluvium; in some other places possibly it is till or alluvium overlain by till. Two pebble fabric measurements indicated to Karlstrom (Karlstrom, 2000b; Karlstrom and Barendregt, 2001) that the sediment he sampled is till. His till 1, till 2, and till 3 have reversed remanent geomagnetic polarity (Cioppa and others, 1995). Till 2 and till 3 probably were deposited during oxygen isotope stages 22 and 20, respectively, during the Matuyama Reversed Polarity Chron (table 2). The temporal relation of till 1 to the oxygen isotope record is problematic. Karlstrom and Barendregt (2001) indicated that till 4 has normal remanent polarity. However, according to Cioppa and others (1995), the buried soil in till 4 has normal polarity, but the polarity of till 4 was not determined. Till 4 probably was deposited during oxygen isotope stage 18 or 16, approximately contemporaneous with extensive pre-Illinoian Laurentide glaciation in the Cut Bank–Sunburst–Shelby region farther east in Montana (tables 2 and 3).

The stratigraphic record in exposures of Kennedy drift on upland erosion surfaces contrasts with the record in the St. Mary Ridge section. Martin (1927) described thick till overlying nonglacial alluvium in exposures on Two Medicine Ridge, a drainage divide north of the Two Medicine River north and northwest of East Glacier Park village, Mont. Karlstrom (l981, 1988a) and Cioppa and others (1995) identified four Kennedy “till” units (diamictons) and two intervening buried soils in one exposure on Two Medicine Ridge

[1]The Bull Lake till of Richmond (1957, 1965b) in the Glacier National Park region is not the same stratigraphic unit as the Bull Lake Till in its type area in Wyoming. The informal term "Bull Lake till" here refers to all mountain till deposited during the Bull Lake glaciation (see “Introduction”).

(fig. 1). Karlstrom's till 1 apparently is younger than the nonglacial alluvium described by Martin. Till 1 is overlain by three diamictons that are separated by buried soils. Pebble fabric measurements from his till 2 indicated it is outwash or ice-contact gravel, not till; till 1 and till 2 probably are products of a single glaciation (Karlstrom, 2000b). The intensity of development of the composite soil in the youngest Kennedy till (till 4 of Karlstrom, 1981, 1988a) indicates that till 1, till 3, and till 4 all are late Pliocene in age (table 3). A possible transition from reversed to normal remanent geomagnetic polarity in till 3 (Cioppa and others, 1995) is consistent with an interpretation that till 3 was deposited approximately 2.15 Ma, during the Matuyama-Réunion geomagnetic reversal (table 2). That reversal, possibly recorded in till and glacial lake sediments in Minnesota, Wisconsin, Missouri, and Iowa, occurred during oxygen isotope stage 82 (Sarnthein and Teidemann, 1990). Karlstrom and Barendregt (2001) stated that till 4 has normal remanent polarity. However, according to Cioppa and others (1995), the paleosol in the till has normal polarity but the polarity of the unaltered till was not determined. Till 4 on Two Medicine Ridge possibly was deposited approximately 2.078 Ma (Chen and others, 1995), during oxygen isotope stage 78 (table 2). The reversed-polarity till 1 and associated stratified deposits (till 2) possibly were deposited approximately 2.443 Ma, during oxygen isotope stage 96.

Karlstrom's till 2 and till 3 of the Kennedy drift on Mokowan Butte (fig. 1), a bedrock erosion surface remnant in southwestern Alberta (Karlstrom, 1981, 1987a,c, 1988a,b, 1991; Barendregt and others, 1991; Cioppa and others, 1995), probably are late Pliocene in age. Till 4 and till 5 on Mokowan Butte probably are middle Pleistocene in age. The youngest till (till 5) probably was deposited during oxygen isotope stage 18 or 16 (table 2). Till 1 and till 5 have normal remanent geomagnetic polarity, and till 3 has reversed polarity (Barendregt and others, 1991; Cioppa and others, 1995). Karlstrom (2000b) indicated that till 2 has normal remanent polarity. However, Barendregt and others (1991) stated that the polarity of till 2 could not be determined. Cioppa and others (1995) sampled the paleosol developed in the till, but not the unaltered till. They concluded that the normal remanent polarity of the paleosol possibly is a Brunhes Normal Polarity Chron overprint. Consequently, the detrital remanent polarity of till 2 is not known. The remanent polarity of till 4 also is not known. Karlstrom's till 1 is an extremely weathered diamicton; possibly it is Tertiary alluvium.

Two Kennedy diamictons were identified at one site on Milk River Ridge (fig. 1), a bedrock erosion surface remnant in Montana (Karlstrom, 1981, 1988a). Both of the diamictons were interpreted to be lodgement tills on the basis of pebble fabric analysis (Karlstrom and Barendregt, 2001). However, interpretation of the fabric from till 1 was not definitive. Samples from till 1 and the paleosol in till 2 had normal remanent polarity (Karlstrom and Barendregt, 2001). The paleosol in till 1 is intensely developed (Karlstrom, 1981, 1988a; Karlstrom and Barendregt, 2001). Till 1 possibly is late Pliocene in age. Till 2 probably was deposited during oxygen isotope stage 18 or 16, approximately contemporaneous with extensive pre-Illinoian Pleistocene continental glaciation on the plains farther east in Montana (tables 2 and 3).

The Albertan mountain till and the Labuma Laurentide till of Stalker (1963, 1976) and Stalker and Harrison (1977) are the oldest tills yet identified on the plains in southern Alberta. Stalker concluded that the Albertan piedmont-lobe till was buried by Labuma Laurentide till, and both tills were deposited during a single pre-Wisconsin glaciation. Jackson and others (1996) and Jackson and Little (2003) concluded that the Albertan and Labuma tills in exposures described by Stalker both are late Wisconsin in age. If that interpretation is correct, no till of pre-Wisconsin age is known to be present on the Northern Plains in southewestrn Alberta. On the basis of stratigraphic relations of tills in Montana and southeastern Alberta, we infer that the subsurface Albertan mountain till in the Monarch-Kipp area, Alberta, and possibly subsurface mountain till in the Oldman River valley as far west as the Piegan and Brocket sections of Stalker (1963), is pre–Bull Lake (pre-Illinoian) in age. We also infer that that till is laterally and temporally equivalent to the youngest Kennedy till on Mokowan Butte (fig. 1) in Alberta (Karlstrom, 1981, 1987a,c, 1988a,b, 1991), and that it was deposited during oxygen isotope stage 18 or 16 (tables 2 and 3).

The samples analyzed to calculate a uranium-trend age of 440±120 yr (DLM–3), attributed to the upper till of the Kennedy Formation by Richmond (1986a), were collected from till in a well-defined late Bull Lake lateral moraine, not from Kennedy till. Consequently, the uranium-trend age does not support interpretations by Richmond that (1) the youngest pre–Bull Lake till in the St. Mary Ridge stratigraphic section (fig. 1) was deposited during oxygen isotope stage 12, or (2) the youngest Kennedy till on upland surfaces in Montana was deposited during stage 12 (table 2). We are not aware of any evidence of extensive valley glaciation in Glacier National Park and Waterton Lakes National Park or adjacent areas in Montana and Alberta during oxygen isotope stages 14 or 12 (table 2). Laurentide continental glaciation during stage 12 apparently is not recorded in Montana, northwestern North Dakota, or southern Saskatchewan. However, Laurentide glaciation during stage 14 is recorded by till in southern Saskatchewan, and possibly it is recorded by subsurface till in extreme northeastern Montana, extreme northwestern North Dakota, and eastern Alberta adjacent to Saskatchewan (E.A. Christiansen, unpub. test hole data, 1999–2001; Fullerton and others, in press). The youngest pre–Bull Lake till in the St. Mary Ridge exposure, the younger Kennedy till on Milk River Ridge, and the youngest Kennedy till on Mokowan Butte (see previous discussion) probably were deposited approximately 720 ka or approximately 648 ka, during oxygen isotope stage 18 or 16 (tables 2 and 3). The reversed remanent geomagnetic polarity of till 2 and till 3 in the St. Mary Ridge section indicates possible deposition approximately 875 ka and approximately 800 ka, respectively, during oxygen isotope stages 22 and 20 (tables 2 and 3). Till 2 and till 3 on Mokowan Butte and the tills on Two Medicine Ridge are late Pliocene in age; all are older than 1.806 Ma (table 2).

A glaciofluvial origin of gravel of pre–Bull Lake age on bedrock benches and beneath terraces between, and east of, outlet-glacier piedmont lobes in the Choteau $1^{\circ} \times 2^{\circ}$ quadrangle (Richmond and Mudge, 1965) has not been demonstrated. Most of the deposits probably are nonglacial alluvium; the oldest deposits are Tertiary in age. Till of pre–Bull Lake age or Bull Lake age has not been identified beyond the limits of the early Pinedale piedmont lobes depicted on the map.

BULL LAKE GLACIATION

The Bull Lake till of Richmond (l965b) in the Glacier National Park region, Montana, is the “early Wisconsin” till of Richmond (1957) and till 5 of Karlstrom (1981, 1988a, 1991) and Cioppa and others (1995) in the St. Mary Ridge stratigraphic section. The Bull Lake till in that exposure (fig. 1) consists of two superposed till units that initially were distinguished by differences in color (Richmond, 1957) and locally are separated by a weakly developed buried soil (Karlstrom, 1981, 2000b). Some of the till mapped as “early mountain moraines” and “early Wisconsin mountain drift” by Horberg (1954) is Bull Lake till. Most of the Bull Lake till in Montana was mapped by Horberg as “mountain drift” and “undifferentiated mountain moraines.” The Bull Lake till in the Glacier National Park region was assigned to the St. Mary Ridge formation by Richmond (1986a). Lateral moraines of early Bull Lake (unit EB) and late Bull Lake (unit LB) ages east of the Saint Mary River in Montana were distinguished by Richmond (1965b, 1986a, and unpub. mapping). In most areas in the Glacier National Park region, the moraines and the till of Bull Lake age are undivided (unit BL), owing to absence of objective criteria for age distinction of the early stade and late stade moraines and deposits.

During the interstade between the early Bull Lake and late Bull Lake stades, the areal extents of the early Bull Lake valley-glacier piedmont glacial lobes in Alberta and Montana were reduced markedly. Active valley-glacier margins in the eastern part of Glacier National Park and east of the park retreated from the plains into the mountains. The interstade is recorded by a weakly developed paleosol in the St. Mary Ridge section (Karlstrom, 1981, 2000b) and by a weakly developed paleosol between early Bull Lake and late Bull Lake outwash deposits (G. M. Richmond, oral commun., 1984–86).

The valley glaciers and piedmont lobes subsequently were reactivated. A late Bull Lake valley-glacier piedmont lobe in the Saint Mary River drainage basin and the Illinoian Laurentide continental ice sheet coalesced at and north of the Hall Coulee spillway, northeast of Babb, Mont. Terraces underlain by late Bull Lake valley-glacier piedmont-lobe outwash and terraces underlain by Illinoian Laurentide (Emigrant Gap) outwash merge as a single depositional surface in the North Fork Milk River valley (Richmond, 1986a, unpub. mapping and oral commun., 1984–86), confirming that the late Bull Lake glacial maximum and the Illinoian continental ice sheet maximum were synchronous (table 3). It is not known if early Bull Lake valley-glacier piedmont lobes on the plains east and northeast of Glacier National Park and Waterton Lakes National Park interacted with the Illinoian Laurentide ice sheet margin.

If late Bull Lake valley-glacier piedmont lobes had not obstructed westward expansion of the Illinoian Laurentide ice sheet in the Saint Mary, Belly, Waterton, and Oldman River drainage basins in the Lethbridge $1^{\circ} \times 2^{\circ}$ quadrangle, the Laurentide ice margin would have advanced westward and southwestward far into the Alberta foothills and into (up) some of the larger mountain valleys (fig. 2*A*). Valley-glacier piedmont-lobe glaciation of pre-Pinedale (pre–late Wisconsin) age in those drainage basins in Alberta was recognized by Vernon (1962), Stalker (1963, 1976), Wagner (1966), Alley (1972, 1973), Alley and Harris (1974), Harrison (1976), Stalker and Harrison (1977), Jackson (1980), and Jackson and others (1989). More recently, however, all of the mountain tills and all of the Laurentide tills in exposures in, and northwest, north, northeast, and east of, Waterton Lakes National Park in those drainage basins in southwestern Alberta were interpreted by Jackson (1994), Jackson and others (1996), and Jackson and Little (2003) to be late Wisconsin in age (see discussion of Kennedy glaciations). We infer that the Laurentide Maunsell till of Stalker (1963, 1976) and Stalker and Harrison (1977) in some exposures in southern Alberta is Illinoian in age.

The presence of “Illinoian (?) mountain moraine” (Bull Lake till) at low altitudes in the North Fork Milk River valley in Montana east of the Hall Coulee spillway (Karlstrom, 1999, fig. 1) has not been verified. Those sediments also were termed “late Illinoian alpine till or outwash” (Karlstrom, 2000a, fig. 1). Unpublished maps by others indicate that the deposits in that area are Kennedy, Bull Lake, and Pinedale outwash, Kennedy till, landslide deposits, colluvium, and nonglacial alluvium. If mountain-provenance till is present in the area mapped by Karlstrom, it is early Bull Lake in age, and the ice moved eastward in the valley, not westward.

Weathered surficial deposits mapped by Mudge (1967, 1972) in the Rocky Mountain foothills were interpreted to be till of Bull Lake age deposited by the Sun River outlet-glacier piedmont lobe (the largest piedmont lobe in the Choteau $1^\circ \times 2^\circ$ quadrangle). A glacial origin of those deposits has not been verified. If the deposits are till, a Bull Lake outlet-glacier piedmont lobe was much thicker and much larger than the early Pinedale piedmont lobe depicted on the map, and extensive deposits of till of Bull Lake age should be present far beyond the limit of the early Pinedale till (unit P1) on the plains. Till has not been identified beyond the limit of the early Pinedale Sun River piedmont lobe. Mudge's "Bull Lake till" in the foothills probably is a product of mass-movement processes. Richmond and Mudge (1965) stated that till of Bull Lake age underlies till of Pinedale age on the plains in the Sun River lobe area. However, the subsurface till was not described. Oxidized till of Bull Lake or pre–Bull Lake age, or both, apparently underlies unoxidized till of Pinedale age in the central part of the Two Medicine lobe area east of East Glacier Park village. It also has not been described.

Scattered residual limestone boulders and cobbles on eroded bedrock in the drainage basin of the Teton River, far east of the Teton River outlet-glacier piedmont lobe and west of Choteau, Mont., were cited as evidence of one or more pre-Pinedale piedmont glaciations (Chalmers, 1968). Till has not been identified anywhere in that region. The clasts probably are lag boulders and cobbles—residual products of erosion of limestone-rich, mountain-source, nonglacial alluvium of Tertiary age that formerly covered the bedrock surfaces.

Richmond (1986a) inferred that the early Bull Lake and late Bull Lake tills in the Glacier National Park region were deposited during oxygen isotope stages 8 and 6, respectively. However, we are not aware of any evidence of extensive valley glaciation during stage 8. Paleosols between two Bull Lake till units or between two Bull Lake outwash units, reported to be present at sites in Montana, are weakly developed soils. We interpret the Bull Lake deposits and moraines to represent two stades of a single glaciation (table 3), correlated with oxygen isotope stage 6 (table 2). The age of the Bull Lake deposits in this region is not closely constrained. As noted in the discussion of Kennedy glaciations, the samples analyzed to calculate a uranium-trend age of 440±120 ka (DLM–3), attributed to pre–Bull Lake Kennedy drift by Richmond (1986a), were collected from till in a well-defined late Bull Lake lateral moraine, not from Kennedy till. That age is rejected. On the basis of chronometric constraints in the Rocky Mountains elsewhere in Montana, Wyoming, and Colorado (Madole and Shroba, 1979; Rosholt and others, 1985; Richmond, 1986a; Sturchio and others, 1994; Chadwick and others, 1997; Phillips and others, 1997; Easterbrook and others, 2003; Benson and others, 2004), the age of the early Bull Lake lateral moraines in the Glacier National Park region is approximately 160 ka, and the age of the late Bull Lake lateral moraines is approximately 140 ka (tables 2 and 3).

PINEDALE GLACIATION

Glacial deposits of early Wisconsin age (oxygen isotope stage 4) have not been identified in the region of Glacier National Park and Waterton Lakes National Park. Early Wisconsin valley glaciers apparently were much less extensive than early Pinedale glaciers, and early Wisconsin deposits were eroded and buried during later Pinedale glacial advances.

The Pinedale till[2] or Pinedale drift of Richmond (1965b) in the Glacier National Park region is the "mountain drift of the Wisconsin stage" of Alden (1932) and the "late mountain moraines" and "late Wisconsin mountain drift" of Horberg (1954). It also includes some of Horberg's "early mountain moraines" and "early Wisconsin mountain drift." It was assigned to the Cutbank Creek formation by Richmond (1986a). The presence of "Wisconsin mountain moraine" (Karlstrom, 1999, fig. 1) or "Wisconsin alpine moraine" (Karlstrom, 2000a, fig. 1) at low altitudes in the North Fork Milk River valley east of the Hall Coulee spillway has not been verified. Unpublished maps by others indicate that the deposits in that area are Bull Lake outwash, Emigrant Gap (Illinoian continental) till and outwash, landslide deposits, colluvium, and nonglacial alluvium. If till is present near the valley floor and south of the river, as depicted by Karlstrom, it probably is Laurentide continental till (Emigrant Gap till) of Illinoian age.

The term "Albertan" apparently is applied to valley-glacier piedmont-lobe tills of different ages in different places in southern Alberta. The Albertan mountain till in many exposures in the Oldman, Waterton, and Belly River valleys north and northeast of Waterton Lakes National Park is the M1 till of Little (1995b, 1998b), Leboe (1996, 1998a,b), Holme (1998a,b), and Jackson and Little (2003), of early Pinedale age. As noted previously, the subsurface Albertan mountain till in the Monarch-Kipp area in the Oldman River valley west of Lethbridge, Alta. (Stalker, 1963, 1976; Stalker and Harrison, 1977) possibly is pre–Bull Lake (pre-Illinoian) in age. Jackson and others (1996) concluded that the mountain till in exposures in the latter area is the same age as (and laterally continuous with) the M1 (early Pinedale) till southwest of that area. On the basis of that conclusion, they concluded further that the M1 (early Pinedale) valley-glacier piedmont-lobe ice margin advanced 100 km from the mountain front onto the plains. The latter interpretation and their calculations of

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[2]The Pinedale till of Richmond (1965a,b) in the Glacier National Park region is not the same stratigraphic unit as the Pinedale Till in the type area in Wyoming. The informal term "Pinedale till" here refers to all mountain till deposited during the Pinedale glaciation (see "Introduction").

ice-surface gradients in the Oldman, Waterton, and Belly River valleys (Jackson and others, 1996, p. 170 and table 6) are dependent on the validity of their interpretation that the subsurface Albertan mountain till in the Monarch-Kipp area is early Pinedale in age. On the basis of the spatial relations of early Pinedale and middle Pinedale glacial limits in the Glacier National Park region and elsewhere in the Rocky Mountains in the United States, we consider it unlikely that early Pinedale piedmont lobes in Alberta were as extensive as those envisioned by Jackson and others (1996). The middle Pinedale limits of valley glaciers, the Yellowstone ice cap outlet glaciers, and other piedmont ice lobes in Montana, Wyoming, and Colorado are not far upvalley from the early Pinedale limits. The interpretation of Jackson and others (1996) apparently requires that an early Pinedale piedmont-lobe glacial limit in the Monarch-Kipp area, Alberta, was approximately 100 km north of the middle Pinedale limit in the Saint Mary River valley. We infer that the subsurface Albertan till in exposures in the Monarch-Kipp area, and possibly subsurface till in some other exposures in the Oldman River valley as far west as the Piegan and Brocket sections of Stalker (1963), is pre–Bull Lake (pre-Illinoian) in age (see discussion of Kennedy glaciations). We also infer that it is laterally and temporally equivalent to the youngest Kennedy till (Karlstrom's till 5) on Mokowan Butte (fig. 1) in Alberta (see Karlstrom, 1981, 1987a,c, 1988a,b, 1991), and it was deposited during oxygen isotope stage 18 or 16 (tables 2 and 3).

Early Pinedale valley glaciers east of the Continental Divide, in and east of Glacier National Park and in Waterton Lakes National Park, merged and formed valley-glacier complexes. In the Waterton, Belly, and Oldman River valleys in southwestern Alberta, the M1 valley-glacier complexes coalesced and formed large piedmont lobes or lobe complexes. Those complexes apparently were not obstructed by the late Wisconsin Laurentide ice sheet during maximum early Pinedale glaciation. The Laurentide continental ice sheet margin was northeast of the early Pinedale piedmont lobe limits. Most of the M1 drift of the early Pinedale lobe complexes on the plains in the part of southern Alberta shown on the map was overridden by Laurentide ice during late Wisconsin Laurentide glaciation.

The early Pinedale valley-glacier piedmont-lobe glacial maximum in, and northeast and east of, Glacier National Park and the Pinedale 1 outlet-glacier piedmont-lobe glacial maximum farther south along the mountain front were synchronous (table 3). The early Pinedale valley glacier in the Two Medicine River valley **merged with** the Pinedale 1 Two Medicine piedmont lobe.

Maximum early Pinedale valley-glacier piedmont-lobe glaciation occurred between approximately 26,160 and 25,700 CAL yr B.P. (22,400 and 22,020 ^{14}C yr B.P.), as indicated by chronometric controls elsewhere in the Rocky Mountains in Montana, Wyoming, and Colorado (Nelson and others, 1979; Madole, 1986, and oral commun., 2003; Richmond, 1986a; Sturchio and others, 1994; Gosse and others, 1995; Chadwick and others, 1997; Phillips and others, 1997; Easterbrook and others, 2003; Benson and others, 2004). The early Pinedale glacial maximum occurred during the later part of oxygen isotope stage 3, not during stage 2 (see table 2).

Prior to the middle Pinedale regional glacial readvance, the termini of early Pinedale valley glaciers in the eastern part of Glacier National Park and east of the park in Montana retreated unknown distances into the mountains (up the valleys from the limits of unit MP shown on the map). In some areas in Montana and southwestern Alberta, the early Pinedale piedmont lobes or lobe complexes wasted primarily by stagnation. We suggest that, although the areal extents of active piedmont lobes or lobe complexes on the plains were greatly reduced, areas of debris-covered, stagnant or dead, early Pinedale mountain ice remained throughout the interstade between the early Pinedale and middle Pinedale stades.

In some areas in southwestern Alberta, the surface landforms and the surface till of the valley glaciers and the valley-glacier piedmont-lobe complexes in the present landscape were inferred to be either early Pinedale or late Pinedale in age. Lateral and terminal moraines of middle Pinedale valley glaciers and terminal moraines of middle Pinedale piedmont lobes were inferred to be absent or missing in the landscape. We suggest that some of the surface landforms (for example, drumlins and flutings) and deposits associated with coalescence of mountain-glacier piedmont lobes and Laurentide continental ice in southern Alberta mapped by Harrison (1976), Bayrock and Reimchen (1980), and Shetsen (1987) are products of middle Pinedale mountain glaciation. When the piedmont lobes were rejuvenated, the middle Pinedale mountain ice overrode, eroded, and buried older early Pinedale deposits, and it also overrode and reactivated buried early Pinedale ice. Consequently, the early Pinedale deposits in some areas are not exposed. Upglacier from the limits of maximum early Pinedale/middle Pinedale (interstadial) recession of active piedmont lobe margins, the early Pinedale and middle Pinedale stades in most places in valleys and lowlands are represented by a single Pinedale till (those areas were covered by active ice throughout the interstade).

The encroaching Laurentide continental ice sheet eventually obstructed the readvancing middle Pinedale piedmont lobes on the plains in southwestern Alberta. In some areas, Laurentide ice overrode and buried early Pinedale deposits and landforms; in some other areas, middle Pinedale ice overrode and buried early Pinedale deposits and landforms. In some areas (for example, in the Belly River valley in Alberta), middle Pinedale mountain-glacier ice and the late Wisconsin Laurentide continental ice sheet coalesced, and temporary superglacial lakes formed on the sutures between the coalesced ice masses. In some other areas (for example, in the Saint Mary River valley in Montana), lakes were dammed between middle Pinedale valley glaciers and the Laurentide ice sheet (see figure 3).

As noted previously, early Pinedale mountain-glacier piedmont-lobe deposits and landforms in some areas were overridden, eroded, and buried by late Wisconsin Laurentide continental ice. Reactivation of buried early Pinedale ice by overriding Laurentide continental ice produced till in which mountain and continental matrix and clast compositions are juxtaposed or intermixed. Exposures of till characterized by those compositions were described by Wagner (1966) and Karlstrom (1987b, 1999). The juxtaposed or intermixed composition of till does not necessarily record confluence of mountain and Laurentide ice margins, as inferred by Wagner and Karlstrom.

Minor paleosols between early Pinedale valley-glacier piedmont-lobe till and overlying late Wisconsin Laurentide till were identified at several sites (Calhoun, 1906; Horberg, 1954; Karlstrom, l987b, 1999). The paleosols at some sites possibly are overlain by slumped Laurentide till (not by in-place till), and thus may not record soil development subsequent to the early Pinedale glacial maximum and prior to the middle Pinedale (≅late Wisconsin) glacial maximum (Michael Wilson, written commun., 2000; Lionel Jackson, written commun,. 2000). Several of the paleosol sites no longer are available for study.

Limits of maximum recession of early Pinedale valley glacier margins in the Beaver Mines region in southwestern Alberta, reconstructed by Holme (1998b), were depicted by Holme and others (2000, fig. 3c). The middle Pinedale (M2) readvance limit also was reconstructed (Holme and others, 2000, fig. 3d). The M2 till in southwestern Alberta (Little, 1995b, 1998b; Leboe, 1996, 1998a,b; Holme (l998a,b; Holme and others, 1998, 2000) is middle Pinedale in age. Cosmogenic ages for boulders on M2 till surfaces ranged from 13,200±700 to 19,200±1,800 ^{36}Cl yr (Jackson and others, 1999).

As noted previously, the early Pinedale valley glacier in the Two Medicine River valley in Montana merged with the Pinedale 1 Two Medicine outlet-glacier piedmont lobe northeast of East Glacier Park village (fig. 1), indicating that the early Pinedale and Pinedale 1 advance maxima were synchronous (table 3). Pinedale 1 till of the Two Medicine lobe unconformably overlies sediment deposited in Illinoian glacial Lake Cut Bank, and the till is overlain unconformably by sediment deposited in late Wisconsin glacial Lake Cut Bank. The Guardipee Lake spillway east of Browning, Mont. (map and fig. 1), the outlet of the highest Illinoian phase of glacial Lake Cut Bank (altitude approximately 1,256 m to approximately 1,232 m), was overridden and buried by Pinedale 1 deposits of the Two Medicine lobe. Guardipee Lake (Barnosky, Anderson, and Bartlein, 1987; Barnosky, Grimm, and Wright, 1987) is in an ice-block depression in Pinedale 1 (unit P1) drift that partly fills the buried lake spillway.

Maximum middle Pinedale valley glaciation in the eastern part of Glacier National Park and east of the park, maximum Pinedale 2 mountain glaciation in the Two Medicine outlet-glacier piedmont lobe, and maximum late Wisconsin Laurentide continental glaciation were synchronous (fig. 3, table 3). Maximum late Wisconsin Laurentide glaciation in this region is inferred to have occurred approximately 23,360 CAL yr B.P. (20,000 ^{14}C yr B.P.), during oxygen isotope stage 2, on the basis of chronometric constraints from Montana, Alberta, and Saskatchewan (see "Late Wisconsin Laurentide glaciation in Montana and northwestern North Dakota"). The chronologic constraints for the glacial maxima indicate that the middle Pinedale mountain glacier maximum and the late Wisconsin Laurentide continental glacial maximum occurred 2,300 to 2,800 CAL yr later than the early Pinedale valley-glacier piedmont-lobe and outlet-glacier maximum.

Glacial Lake St. Mary (Horberg, 1954; Richmond, 1965b, 1986a) fronted the middle Pinedale valley glacier at its maximum extent in the Saint Mary River valley at Babb, Mont. Lake sediment is interfingered with, and overlies, till on the distal side of the middle Pinedale terminal moraine, and lake sediment in some places is overlain by outwash deposits associated with formation of the moraine. The outwash was deposited when the highest late Wisconsin phase of the lake (altitude approximately 1,420 m) was extinguished (Richmond, 1986a, and oral commun., 1984–86). The lake was dammed by the late Wisconsin Laurentide ice sheet when the ice sheet was at its maximum extent, demonstrating that the middle Pinedale mountain valley glacier advance and the late Wisconsin Laurentide advance maxima were synchronous (fig. 3). Glacial Lake St. Mary drained southward to the North Fork Milk River by way of the Hall Coulee spillway. The North Fork flowed into glacial Lake North Fork (Horberg, 1954) in Alberta. Outflow from glacial Lake North Fork entered the highest late Wisconsin phase of glacial Lake Twin River (Horberg, 1954), at an altitude of approximately 1,268 m, in the Milk River valley northwest of Cut Bank, Mont., and southeast of Del Bonita, Alta. Lake Twin River also was dammed by the late Wisconsin Laurentide ice sheet when the ice sheet was at its maximum extent (fig. 3). Glacial Lake Twin River drained through the Antelope Coulee spillway to the highest late Wisconsin phase of glacial Lake Cut Bank north, west, and south of Cut Bank. The highest late Wisconsin phase of Lake Cut Bank (altitude approximately 1,207 m) also was dammed by the late

Wisconsin Laurentide ice sheet when the ice sheet was at its maximum extent (fig. 3). The duration of each high-level glacial lake phase (of Lakes St. Mary, North Fork, Twin River, and Cut Bank) was short.

The Pinedale 2 readvance maximum of the Two Medicine outlet-glacier piedmont lobe also occurred during the existence of the highest late Wisconsin phase of glacial Lake Cut Bank (fig. 3). Lake sediment unconformably overlies Pinedale 1 mountain till (unit P1) and outwash (valley train) deposits in the Two Medicine River, Cutbank Creek, Spring Creek, and Badger Creek valleys in Montana. The Pinedale 1 melt-water streams in those valleys did not flow into glacial lakes to the east when the valley trains formed (the Laurentide ice margin did not reach the western part of the Montana plains during Pinedale 1 time). Iceberg-rafted erratic boulders and cobbles from the Canadian Shield are scattered on Pinedale 1 till (unit P1) surfaces. In some places, thick sediment of late Wisconsin glacial Lake Cut Bank unconformably overlies Pinedale 1 till. The lake sediment interfingers with till on the distal side of the Pinedale 2 piedmont-lobe terminal moraine (unit P2). Ice-rafted erratic boulders and cobbles from the Canadian Shield are present on distal slopes of that moraine at altitudes lower than the highest late Wisconsin phase of glacial Lake Cut Bank, but are not present at higher altitudes on the distal slopes or at any altitude on proximal slopes of the moraine. Eugene Stebinger noted the presence of iceberg-rafted erratics from the Canadian Shield on the Pinedale 2 terminal moraine as early as 1912 (Alden, 1932, p. 106). The middle Pinedale valley glacier in the Two Medicine River valley **merged with** the Pinedale 2 Two Medicine outlet-glacier piedmont lobe northeast of East Glacier Park village (fig. 3), demonstrating that the middle Pinedale and Pinedale 2 glacial advance maxima were synchronous.

Prior to the late Pinedale regional glacial readvance, the termini of valley glaciers in the eastern part of Glacier National Park and east of the park in Montana retreated unknown distances into the mountains (up the valleys from the limits of unit LP shown on the map). As noted previously, the early Pinedale and middle Pinedale valley glaciers in the Two Medicine River valley merged with the Two Medicine outlet-glacier piedmont lobe. However, the terminal moraine of the late Pinedale valley glacier (unit LP) in the Two Medicine River valley north of East Glacier Park village, described by Martin (1927), overlaps the Pinedale 3 deposits and landforms (unit P3) of the Two Medicine lobe (fig. 1). The active terminus of the Pinedale 3 outlet glacier that funneled ice from the northern Montana ice field across Marias Pass to East Glacier Park village, and then eastward as the Two Medicine piedmont lobe, retreated into the mountains southwest of East Glacier Park prior to culmination of the late Pinedale valley-glacier readvance. Possibly, it retreated as far into the mountains as Marias Pass. The spatial relation of unit P3 till and landforms and unit LP till and landforms north of East Glacier Park village (map and fig. 1) indicates that the "late Pinedale" outlet-glacier drift (unit P3) is significantly older than the late Pinedale valley-glacier drift of unit LP (table 2).

Mount Saint Helens set Jy tephra and Glacier Peak G tephra were deposited in the vicinity of Marias Pass after the pass was deglaciated (Carrara and others, 1986; Carrara, 1989). Wood beneath the older tephra yielded an accelerator ^{14}C age (AA–9530) of 12,195±145 yr B.P. (Carrara, 1995). Hummocky topography between East Glacier Park village and Marias Pass, identified on aerial photographs, has not been examined in the field. If the morainal topography marks the limit of a post-Pinedale 3 outlet-glacier readvance synchronous with the late Pinedale (unit LP) valley-glacier readvance in the Two Medicine River valley northwest of East Glacier Park village, the late Pinedale glacial readvance maximum occurred much earlier than 14,245 CAL yr B.P. (12,195 ^{14}C yr) B.P.

Evidence of a regional readvance of valley glaciers contemporaneous with the Pinedale 3 (unit P3) outlet-glacier piedmont-lobe readvance has not been reported. The early Pinedale, middle Pinedale, and late Pinedale valley-glacier advances in the Glacier National Park region probably reflect large-scale climatic changes throughout the Rocky Mountain region. The Pinedale 3 outlet-glacier advance occurred when the northern Montana ice field still was very large and when a very substantial discharge of ice still was funneled across Marias Pass to the outlet glacier. Ice-surface profiles and ice thickness in the Two Medicine lobe were very low (Locke, 1995). Therefore, relatively minor changes of ice thickness and (or) ice regimen in the ice field would have resulted in major fluctuations of theTwo Medicine lobe ice margin. In contrast, the late Pinedale valley-glacier readvance (unit LP) occurred after the volume of ice in the valleys had been reduced greatly and after the active ice margins had retreated long distances up the valleys toward the cirques. The Pinedale 3 readvance of the Two Medicine outlet-glacier piedmont lobe probably was driven primarily by mass-balance changes in the northern Montana ice field associated with dissipation of the ice field, rather than by the regional climatic factors that were responsible for the late Pinedale readvance of valley glaciers.

Maximum late Pinedale valley glaciation occurred approximately 17,640 to 17,520 CAL yr B.P. (15,100 to 15,000 ^{14}C yr B.P.), during oxygen isotope stage 2, based on chronometric constraints elsewhere in the Rocky Mountains. It was not contemporaneous with any known regional glacial readvance maximum of the Laurentide continental ice margin in Montana, North Dakota, Saskatchewan, or Alberta.

LAURENTIDE CONTINENTAL GLACIATION

Outcrops and excavations on the plains in Montana generally are small and widely scattered, and commonly only one till unit is exposed. Some scientists accustomed to a “layer cake” vertical and lateral distribution of till units of different ages in other regions assume that the surface till observed in exposures and excavations in Montana and northwestern North Dakota is late Wisconsin in age **simply because it is the surface till**. This is an erroneous assumption. The observed physical properties of the surface till in different places (actually, tills of pre-Illinoian, Illinoian, and late Wisconsin ages in different places) all are assumed erroneously to be diagnostic properties of till of late Wisconsin age. The surface till, thus recognized, has a very broad spectrum of field characteristics because those properties are the combined characteristics of all of the tills in the region. The distinguishing characteristics of the pre-Illinoian, Illinoian, and late Wisconsin tills (outlined in the appendix) are not recognized, and the complexity of the glacial history is not apparent.

In the past, some observers attributed all oxidized zones in subsurface till and all oxidized zones in inter-till sediments to groundwater flow or seepage, thereby denying a possibility that any subsurface oxidized zones record intervals of subaerial weathering. Weathered zones within till units in exposures nearly everywhere separate till units that are characterized by other contrasting physical properties. Those properties are diagnostic characteristics of different stratigraphic units (see appendix). Most oxidized zones more than 30 cm thick in subsurface till are subaerial weathering horizons. Oxidation of till by groundwater flow or seepage is common in zones a few centimeters thick at contacts between till and overlying stratified sediments, but it has not been identified within till.

Also, the complexity of the till stratigraphy is recognized by some observers only where strongly developed buried soils that have thick B horizons are present between till units. Paleosols between pre-Illinoian tills at several sites are buried organic horizons (O horizons) or thin, truncated B horizons. Buried paleosols in the youngest pre-Illinoian till in exposures commonly are truncated B horizons. Buried paleosols in Illinoian till generally are either truncated B horizons or truncated calcic soils. Landscapes were intensely eroded during interglacial intervals. Soils (chiefly truncated paleosols) were preserved only rarely, where they were buried by proglacial lake sediments or till. In most regions, paleosols were destroyed by subglacial deformation. Paleosols overridden by ice were consumed as part of the deforming bed that lubricated movement of the overriding ice. All of the continental tills on the Montana plains have sedimentologic properties indicative of deforming glacial beds (Alley, 1991; Boulton, 1996a,b; Clark, 1994), and ice-surface profiles and ice thickness were extremely low during each of the Pleistocene glaciations.

LATE PLIOCENE PRE-ILLINOIAN GLACIATION

Evidence of late Pliocene continental glaciation on the Northern Plains in Montana, North Dakota, and southern Saskatchewan is preserved only on the highest (oldest) erosion surfaces (not in the deepest or topographically lowest “preglacial” buried valleys). In Montana, nearly all of the Pliocene continental glacial deposits were removed by erosion during a nonglacial interval longer than 1.4 m.y., prior to middle Pleistocene glaciation (during the interval between oxygen isotope stages 82 and 22; see table 2).

W.T. Pecora reported the presence of large residual erratic blocks of garnet gneiss and pegmatite, presumably from the Canadian Shield, on drainage divides in the western Bearpaw Mountains south and southeast of Rocky Boy, Mont., in the southwest quarter of the Havre 1° × 2° quadrangle (Knechtel, 1942, p. 921). The blocks are 2.2–6.5 km east and southeast of the mapped limit of till. They are in rugged mountains at altitudes more than 300 m higher than any known occurrence of till. Their presence cannot be explained by rafting in an ice-dammed lake. They must have been transported into the mountains by ice during one or more late Pliocene continental glaciations. The erratic blocks were preserved because they are on flat or gently sloping drainage divides, and thus they escaped removal and burial by mass-movement processes.

North of Glasgow, Mont., in the Glasgow 1° × 2° quadrangle, a cobble of granite from the Canadian Shield was extracted from the lower part of a deposit of calcium carbonate-cemented (indurated) colluvium or till that probably is late Pliocene in age (Fullerton and others, in press). The colluvium or till underlies two tills of middle Pleistocene pre-Illinoian age that are separated by a paleosol. Farther north, in the vicinity of Opheim, Mont., cobbles from the Canadian Shield were pressed by ice into thick petrocalcic soil carbonate horizons of Tertiary (late Miocene?) age that overlie gravel (Fullerton and others, in press). Till associated with the ice-pressed erratics has not been identified. North of Opheim, residual glacial erratic boulders are scattered on a dissected bedrock landscape (unit GB). The limit of unit GB shown on the map is the limit of known residual glacial erratic boulders, not the limit of till exposures.

The part of unit GB in the northwestern part of the Wolf Point 1° × 2° quadrangle was mapped by Howard (1960) as “till and small bodies of glaciofluvial sediment.” However, he did not observe any till in that region. His limit of glaciation was the limit of residual glacial erratic boulders on bedrock surfaces (Howard, 1960, p. 29). Subsequent searches for till in that region by other field parties also were futile. Redeposited pebbles and cobbles from the Canadian Shield are scattered in alluvium and colluvium in the areas mapped as unit GB in Montana. Some of the redeposited erratic boulders and cobbles were rafted by icebergs in Pleistocene ice-dammed lakes. However, many are at altitudes too high to be attributed to rafting. Apparently, they are residual evidence of late Pliocene continental glaciation (Fullerton and others, in press).

Residual glacial erratic boulders and cobbles also are present in colluvium and alluvium in the area mapped as unit GB in southern Saskatchewan (Klassen, 1992a,b, 2002). Till is not preserved in that region, mapped by Klassen as "bedrock terrain with residual drift." Boulders of igneous and metamorphic rocks from the Canadian Shield that were pressed by ice into Tertiary gravel and colluvium on pediments are thoroughly decomposed (Klassen, 1992b, fig. 14, p. 372; Klassen, 2002, fig. 15).

The western part of the region mapped as pre-Wisconsin till and glacial erratic boulders (unit PW) in North Dakota is intensely dissected bedrock terrain. South of the Little Missouri River and on the north margin of the Killdeer Mountains, scattered residual glacial erratic boulders on some bedrock surfaces at high altitudes possibly are lag boulders derived from erosion of till of late Pliocene age. Dissected till of Pleistocene age is present on some drainage divides north of the Little Missouri River, and residual glacial erratic boulders are scattered on bedrock surfaces. Between the mapped limit of late Wisconsin glaciation and the Little Missouri River, isolated outcrops of eroded till of Illinoian age, till of pre-Illinoian age, or both, are present on some drainage divides. Owing to intense dissection of the tills and bedrock, the limits of those glaciations in that region cannot be accurately delineated The area mapped as unit PW includes residual erratic boulders of possible late Pliocene age, till of pre-Illinoian Pleistocene age, till of Illinoian age, and possibly some till of late Wisconsin age. The Dunn glaciation (Clayton, 1970) and the Verone glaciation (Bickley, 1972) were proposed on the basis of the presence of scattered residual glacial erratic boulders on bedrock surfaces northwest, west, and south of Dunn Center, N. Dak. A single area of till attributed to the Dunn glaciation was reported by Moran and others (1976). Clayton (1970) concluded that the regional bedrock surface has been lowered more than 60 m by erosion since the Dunn glaciation. Till, glaciofluvial deposits, and residual glacial erratic boulders on high-altitude bedrock surfaces west of the Missouri River in South Dakota (southeast of the region shown on the map) also are attributed to late Pliocene continental glaciation (Fullerton and others, 1995).

EARLY PLEISTOCENE PRE-ILLINOIAN GLACIATION

Till of early Pleistocene age, deposited during oxygen isotope stages 22 and 20 (table 2), has not been identified on the plains in Montana or northwestern North Dakota. However, till of the Mennon Formation, of probable early Pleistocene age, was identified in a test hole approximately 1.3 km north of the International Boundary, in Saskatchewan northwest of Crosby, N. Dak. (E.A. Christiansen, written commun., 1999; Fullerton and others, in press). The presence of Mennon Formation till in the buried ancestral valley of the Missouri River (Meneley and others, 1957) at that site in the Weyburn $1^{\circ} \times 2^{\circ}$ quadrangle indicates that till of early Pleistocene age possibly is present in the buried ancestral valleys of the Missouri and Yellowstone Rivers in extreme northeastern Montana and extreme northwestern North Dakota. The Mennon Formation (Christiansen, 1992) includes two till members that locally are separated by a weathered zone (E.A. Christiansen, written commun., 1999). A study of remanent geomagnetic polarity of Mennon Formation tills is in progress (R.W. Barendregt, oral commun., 2003).

MIDDLE PLEISTOCENE PRE-ILLINOIAN GLACIATION

The limit of middle Pleistocene pre-Illinoian glaciation in most regions in Montana has not been delineated in detail in the field. In most areas, the pre-Illinoian till is concealed by Illinoian till (unit IL) and (or) late Wisconsin till (unit LW). In some other areas, the pre-Illinoian till has been exhumed by erosion. It is exposed in isolated, eroded outcrops in dissected bedrock terrain. An apparent limit of pre-Illinoian Pleistocene till (unit PI) is mapped in two valleys on the north margin of the Bearpaw Mountains. A buried limit of middle Pleistocene pre-Illinoian glaciation in the vicinity of Great Falls, Mont., shown on the map, is the southern limit of known outcrops of the Havre till in that region. The southern limit of known exposures of till of pre-Illinoian age is not necessarily the limit of pre-Illinoian Pleistocene glaciation. Till deposited beyond that limit may have been removed by erosion or may be covered by Illinoian till or by mass-movement deposits.

In general, the Pleistocene pre-Illinoian tills on the western and central plains in Montana are much thicker than the younger tills. Where pre-Illinoian and younger tills are superposed in exposures, pre-Illinoian till commonly constitutes 60 percent to more than 80 percent of the total till thickness. Where Illinoian and late Wisconsin tills were removed by erosion (for example, in the vicinity of Great Falls, Conrad, Shelby, and Havre, Mont., and elsewhere in the Teton, Marias, and Missouri River valleys in the Great Falls, Shelby, and Havre $1^{\circ} \times 2^{\circ}$ quadrangles), pre-Illinoian till is the surface till in most of the large exposures and excavations.

The Havre till on the western plains in Montana, the Perch Bay till on the central plains, and the Archer till on the eastern plains (Fullerton and Colton, 1986) are laterally continuous allostratigraphic units of middle Pleistocene pre-Illinoian age (table 3). Each stratigraphic unit consists of two compositionally similar till units that locally are separated by a paleosol, a weathered zone, or both. The Lothair till of Smith and others (1959) is the Havre till. The upper unit of the Archer till is laterally continuous with unit B (Salomon, 1974, 1976) or till of the Medicine Hill Formation (Fulton, 1976) east of the Yellowstone River and south of the Missouri River in westernmost North Dakota (table 3). Till laterally continuous with the lower unit of the Archer till is present in the buried ancestral valleys of the Yellowstone and Missouri Rivers in northwestern North Dakota; it has not been identified south of the present Missouri River in North Dakota. The field properties of the Labuma till in southeastern Alberta (Stalker, 1963, 1976) are identical to those of the Havre till in Montana (A.M. Stalker, oral commun., 1980), and the two stratigraphic units apparently are laterally equivalent. The "black" till of Stalker (1969, 1976), Stalker and Churcher (1972), and Stalker and Wyder (1983) or till of the Twin Cliffs formation (Stalker, 1976) in southeastern Alberta is laterally equivalent to the Labuma till (Stalker, 1976). The Dundurn Formation of Christiansen (1992), in southern Saskatchewan, includes two till members that locally are separated by a zone of subaerial weathering (E.A. Christiansen, written commun., 1999; Fullerton and others, in press). The Dundurn Formation till has normal remanent geomagnetic polarity, and it is laterally equivalent to the Archer till in northeastern Montana and till of the Medicine Hill Formation in northwestern North Dakota (Fullerton and others, in press). The upper till member of the Dundurn Formation in Saskatchewan is overlain by the Wascana Creek ash, the normal-polarity Lava Creek B tephra that originated in the Yellowstone National Park region (Westgate and others, 1977; Izett and Wilcox, 1982). The airfall tephra was deposited in ponds on stagnant ice during late Dundurn deglaciation (E.A. Christiansen, written commun., 1990, 1999; Fullerton and others, in press).

The areal distribution of pre-Illinoian till on the western Montana plains indicates that regional ice flow during both of the glacial maxima (recorded by two laterally persistent Havre till units) was generally similar to that during Illinoian glaciation (illustrated in fig. 2*A*). We are not aware of any evidence of southeastward regional movement of Pleistocene pre-Illinoian ice from south-central Alberta into Montana. Carbonate clasts in the till are predominantly eastern-source limestone and dolomite from Saskatchewan and Manitoba (Shetsen, 1984), not western-source rocks from the Rocky Mountains and foothills in Montana and Alberta. Most of the western-source clasts in the till were incorporated from older gravel deposited by eastward-flowing, nonglacial streams that originated in the Rocky Mountains in Montana.

Southwest of Cut Bank, Mont., pre-Illinoian till is overlain by till of Illinoian age and by thick sediments of Illinoian and late Wisconsin phases of glacial Lake Cutbank. The areal extent of buried pre-Illinoian till north, northwest, and west of Cut Bank is not known. The buried pre-Illinoian and surface Illinoian glacial limits probably are nearly coincidental west of Shelby, Mont., east of the area where early Pinedale till of the Two Medicine piedmont lobe (unit P1) overlaps the Illinoian till (unit IL). Pre-Illinoian till has not been identified in the dissected bedrock terrain south of Lake Frances and northeast of Bynum. Most of the area mapped as Illinoian till (unit IL) in the Bynum-Choteau area is covered by thick sediment of glacial Lake Choteau, and the limit of pre-Illinoian glaciation is concealed.

The Shonkin Sag channels (see map), a plexus of bedrock channels excavated (and, in some cases, re-excavated) by catastrophic glacial-lake outburst floods during four glaciations, are major features of the landscape on the north flank of the Highwood Mountains. Pre-Illinoian till (Havre till) nearly fills the northernmost and topographically lowest catastrophic flood channel. The pre-Illinoian till in that channel in some places is overlain by tills of Illinoian and late Wisconsin age. Apparently, there were two Havre glaciations along the north margin of the Highwood Mountains, and the second glaciation was more extensive than the first. Two superposed, compositionally similar, units of pre-Illinoian (Havre) lodgment till are exposed in many places in that region and the Fort Benton–Loma region, in the Teton River and Missouri River valleys. An intervening paleosol has not been identified. Nonetheless, the two Havre till units are interpreted to be products of two distinct glaciations, and to be laterally equivalent to the lower and upper units of the Havre till at Havre, Mont. The two till units at Havre locally are separated by a paleosol.

Prior to the earlier of the two middle Pleistocene pre-Illinoian glaciations, the ancestral Missouri River flowed northeastward from the vicinity of Great Falls to Havre, and then eastward in the lowland now occupied by the Milk River. During both of the glaciations, Missouri River drainage was diverted south of the Bearpaw Mountains. The paleovalley northeastward to Havre was reoccupied during each deglaciation.

Middle Pleistocene pre-Illinoian till is exposed locally in valleys on the north margin of the Bearpaw Mountains (unit P1). Till of Illinoian age has not been identified in those valleys; however, discontinuous Illinoian till possibly also is present. Hillslopes in the Bearpaw Mountains south of the mapped limit of late Wisconsin glaciation are very steep. Till of pre-Illinoian and (or) Illinoian age possibly was deposited farther south in the mountains and was removed by erosion or is concealed by thick debris deposited by mass-movement processes. Pre-Illinoian till is present beneath till of Illinoian age (unit IL) on the margins of the Sweetgrass Hills nunataks in the Shelby $1^{\circ} \times 2^{\circ}$ quadrangle.

The pre-Illinoian Perch Bay till on the central plains in Montana, the Havre till on the western plains, and the Archer till on the eastern plains (Fullerton and Colton, 1986) are laterally continuous allostratigraphic units (table 3). The Sturgeon Bay till of Fullerton and Colton (1986) at that time was interpreted to have reversed remanent geomagnetic polarity, indicating deposition during the Matuyama Reversed Polarity Chron (deposition prior to approximately 778 ka; see table 3). Subsequent field investigations indicated that the Sturgeon Bay till is a compositional facies of the normal-polarity lower unit of the Perch Bay till. Reassessment of the paleomagnetic data by Mark Hudson (U.S. Geological Survey, written commun., 1991) confirmed that the Sturgeon Bay facies has normal remanent polarity, indicating deposition later than 778 ka. Both of the Perch Bay till units were deposited by ice that flowed southward across the Wood Mountain upland in Saskatchewan, to the Malta–Glasgow–Fort Peck region in Montana (Fullerton and others, in press). A paleosol separates two pre-Illinoian Perch Bay till units at several sites.

Middle Pleistocene pre-Illinoian till is exposed beneath Illinoian till and late Wisconsin till on buried escarpments on all sides of the Boundary Plateau, and also on the eastern and western parts of the plateau surface in the Havre and Glasgow $1^{\circ} \times 2^{\circ}$ quadrangles. Till of pre-Illinoian age is not present on the central part of the plateau, indicating that that region was ice free (a nunatak) during both of the middle Pleistocene pre-Illinoian glaciations. During both of the glacial maxima, a suture between Havre and Perch Bay ice lobes extended from the Boundary Plateau to the Bearpaw Mountains.

The limits of deposits of both of the pre-Illinoian Perch Bay glaciations are south of the Fort Peck Lake dam. Two normal-polarity till units, separated by a well-developed paleosol, are present locally near Fort Peck, on the north shore of the lake. The lower unit is the Sturgeon Bay facies of the lower unit of the Perch Bay till, discussed previously. Outcrops of pre-Illinoian till are widespread north of the Missouri River and the lake, and also west and southwest of Sun Prairie in the Jordan $1^{\circ} \times 2^{\circ}$ quadrangle and south of Landusky in the Lewistown $1^{\circ} \times 2^{\circ}$ quadrangle. The till has not been identified in the intensely dissected bedrock terrain south of the Missouri River (north of Roy and Valentine) or south of Fort Peck Lake south and southeast of Sun Prairie. Pre-Illinoian till is overlain by till of Illinoian age in much of the region mapped as Illinoian till (unit IL) between the International Boundary and the Milk River in the Glasgow $1^{\circ} \times 2^{\circ}$ quadrangle. Much of the region beyond the limit of late Wisconsin glaciation farther south (between the Milk River and Fort Peck Lake in the Glasgow and Jordan $1^{\circ} \times 2^{\circ}$ quadrangles) is dissected bedrock terrain. Isolated small outcrops of exhumed, eroded, pre-Illinoian till are present in some areas. Small outcrops of extremely weathered, exhumed till of pre-Illinoian age are widespread in dissected bedrock terrain northwest and west of Circle and south and east of the east arm of Fort Peck Lake.

The earliest middle Pleistocene, pre-Illinoian, northeast-source Archer glaciation in northeastern Montana was not extensive, and the early Perch Bay and early Archer ice lobes did not coalesce in the Missouri River valley south of the Peerless Plateau and west of Poplar, Mont. The lower unit of the Archer till is known to be present only in the buried valley of the ancestral Missouri River. The buried eastern limit of the lower unit of the Perch Bay till in the Missouri River valley apparently is in the vicinity of Wolf Point (Fullerton and others, in press). The Sprole Silt (Colton, 1963a,b) was deposited in a lake bordered by early Perch Bay ice (to the west) and dammed by early Archer ice (to the northeast). Sediment deposited in a contemporaneous (possibly confluent) lake in the Yellowstone River valley has normal remanent geomagnetic polarity (Fullerton and others, in press), indicating deposition later than 778 ka (table 2). The “till(?) along Smoke Creek” (Witkind, 1959) or “till(?) at Smoke Creek” (Howard, 1960) is the upper unit of the Archer till. Much of the “early Wisconsin(?) till” in stratigraphic sections described by Howard (1960) is Archer till or Medicine Hill Formation till. The “early Wisconsin(?) till” mapped by Howard is not a stratigraphic unit. In some places, it is exhumed Archer till or till of the Medicine Hill Formation, of pre-Illinoian age; in other places, it is till of Illinoian age or late Wisconsin age, or till of both ages.

Data from drill holes, wells, and exposures indicate that an arcuate “end moraine” mapped by Colton and others (1961) northeast of Poplar, Mont., is a buried pre-Illinoian terminal moraine. The core of the moraine, composed of the lower unit of the Archer till, is overlain by the upper unit of the Archer till and younger tills of Illinoian and late Wisconsin age. The topographic form of the broad, low-relief “end moraine” is inherited from a massive, buried early Archer terminal moraine that crosses the bedrock valley of the ancestral Missouri River. Initial diversion of the ancestral Missouri and Yellowstone Rivers eastward into North Dakota occurred when early Archer and early Medicine Hill ice obstructed northeastward flow of the ancestral Missouri River northeast of Poplar and northwest of Brockton, Montana, and obstructed northward flow of the ancestral Yellowstone River southwest of Williston, N. Dak. (Fullerton and others, in press). The Missouri River was diverted eastward through the Culbertson Sag buried valley(see map) at Culbertson, Mont., described by Alden (1932) and Howard (1960), and the combined

Yellowstone-Missouri discharge was diverted eastward into North Dakota through the Charbonneau Sag buried valley (see map) northeast of Cartwright, N. Dak. During the following interglaciation (the early Archer–late Archer interglaciation), eastward drainage of the Missouri–Yellowstone River system continued in the Charbonneau Sag. An alluvial fill, graded to a (presently) hanging bedrock sill at Cartwright, was deposited in the Missouri River valley in Montana and in the Yellowstone River valley in Montana and North Dakota. The sill at the head of the (presently) buried Charbonneau diversion channel served as a base-level control for the Yellowstone and Missouri Rivers to the west. The fills in the Charbonneau Sag and in the Missouri and Yellowstone River valleys west of the sill were overridden by ice during the late Archer or late Medicine Hill glaciation. Temporary diversion channels farther south, eastward across the Yellowstone–Little Missouri drainage divide, were excavated during late Archer or late Medicine Hill glaciation and deglaciation, but the Missouri-Yellowstone drainage did not reoccupy the Charbonneau Sag during the subsequent interglaciation. Diversion of the Missouri River eastward across a drainage divide southeast of Williston, N. Dak., during late Medicine Hill deglaciation initiated the present eastward course of the river in westernmost North Dakota.

Pre-Illinoian late Archer glaciation was very extensive. The ice margin advanced southwestward **at least as far as Glendive, Mont.**, in the Yellowstone River valley, and late Perch Bay and late Archer ice lobes coalesced in the Missouri River valley southeast of the Peerless Plateau and west of Poplar, Mont. The upper unit of the Archer till is present (in small isolated outcrops) in all of the areas mapped as Illinoian till (unit IL) in the Glendive $1^{\circ} \times 2^{\circ}$ quadrangle. The inferred limits of Illinoian glaciation and pre-Illinoian glaciation are approximately coincidental in those areas. Late Archer glaciation in the Yellowstone River valley in Montana and late Medicine Hill glaciation in the Little Missouri River valley in North Dakota (in the region mapped as unit PW) possibly was more extensive than the later Illinoian glaciation in those regions. The late Medicine Hill glacial limit probably is south of the course of the diverted Little Missouri River in some places.

Initial diversion of the Little Missouri River eastward in North Dakota in its present course occurred during maximum late Medicine Hill glaciation. The single till unit in the Medicine Hill Formation in its type section (Ulmer and Sackreiter, 1973) east of the region shown on the map is laterally equivalent to unit B of Salomon (1974, 1976), to the single till of the Medicine Hill Formation described by Fulton (1976), and also to the upper unit of the Archer till in Montana. Fullerton and Colton (1986) inferred that the Lava Creek B tephra was deposited during the early Archer–late Archer interglaciation. Data collected during a subsequent field investigation demonstrated that the tephra is younger than the upper unit of the Archer till (see the following discussion of the Lava Creek B tephra).

The areal distribution of Perch Bay till and Archer till in Montana indicates that regional ice flow during the two middle Pleistocene pre-Illinoian glaciations was similar to that during Illinoian glaciation (illustrated in fig. 2*A*). However, as noted previously, regional ice flow from the northeast into northeastern Montana was not as extensive during early Archer glaciation as during the late Archer, Illinoian, and late Wisconsin glaciations.

Where they were not protected by overlying, relatively impermeable lake sediment, the Pleistocene pre-Illinoian tills on the Northern Plains in Montana and northwestern North Dakota are intensely weathered. Oxidation of till in northeastern Montana to depths of 11–15 m , reported by Howard (1960), was in exposures of Archer till and Medicine Hill Formation till of pre-Illinoian age, not in exposures of till of Illinoian age or late Wisconsin age. At depth in many exposures, the oxidation is in joints that penetrate deeply into the till. Till a few centimeters away from the joints is not oxidized. Oxidation of till to depths greater than 21 m, reported by Jensen and Varnes (1964), was in exposures in which oxidized tills of pre-Illinoian age and oxidized tills of Illinoian and (or) late Wisconsin age were superposed. Those depths included oxidation of pre-Illinoian till in joints much deeper than the oxidation in the till matrix between joints. Where two units of Havre till or Perch Bay till are superposed in exposures, commonly the weathering profile developed in the upper unit extends through that unit and penetrates (or extends through) the underlying till. However, unoxidized "late" pre-Illinoian till overlies weathered "early" pre-Illinoian till in some exposures. It is clear that the upper part of the lower till unit was oxidized by subaerial weathering prior to deposition of the upper unit. Subaerial weathering of the upper units of the Havre, Perch Bay, and Archer tills in some places represents an interglacial interval of approximately 500,000 yr, prior to the Illinoian glaciation (the interval between oxygen isotope stages 16 and 6; table 2).

The two units of the Havre till, the two units of the Perch Bay till, and the two units of the Archer till in Montana have normal remanent geomagnetic polarity and are younger than ≈778 ka, the $^{40}Ar/^{39}Ar$ age and astronomically tuned age of the Matuyama-Brunhes paleomagnetic reversal (table 2). The upper units of the three tills are laterally continuous. Northwest of Intake, Mont. (see map), the normal-polarity airfall Lava Creek B tephra (Izett and Wilcox, 1982) was deposited in a pond that was dammed by an alluvial fan in an abandoned meltwater channel. The channel had been incised into the upper unit of the Archer till and the underlying terrace gravel during late Archer deglaciation. Published $^{40}Ar/^{39}Ar$ ages of the Lava Creek B Tuff in the Yellowstone National Park region are 600±10 ka and 610±10 ka (Obradovich and Izett, 1991), 602±2 ka (Gansecki and others, 1998), and 639±2 ka (Lanphere and others, 2002). Ages for the distal airfall tephra are 660±10 ka and 670±10 ka (Izett and others, 1992), 630±15 ka (Gansecki and others, 1998), and 635±7 ka (Lanphere and others, 2002). The most reliable $^{40}Ar/^{39}Ar$ age for the Lava Creek B Tuff is 639±2 ka (Lanphere and others, 2002). That age (table 2) is a minimum age for the Archer, Perch Bay, and Havre tills.

The ages of the two Havre glaciations, the two Perch Bay glaciations, and the two Archer glaciations in Montana, the two Medicine Hill glaciations in North Dakota, and the two Dundurn glaciations in Saskatchewan are constrained by the age of the Matuyama-Brunhes paleomagnetic reversal (≈778 ka; oxygen isotope stage 19) and the age of the Lava Creek B tephra (≈639 ka; late oxygen isotope stage 16) (see table 2). The presence of a well-developed paleosol between the two Havre till units and between the two Perch Bay till units indicates that two distinct glaciations are represented. Those two glaciations can be correlated only with middle Pleistocene marine oxygen isotope stages 18 and 16 (table 2).

Till of the Warman Formation in southern Saskatchewan (Christiansen, 1992) overlies pond sediment that contains airfall Lava Creek B tephra (E.A. Christiansen, written commun., 1990, 1999; Fullerton and others, in press). It was deposited during oxygen isotope stage 14, contemporaneous with deposition of northwestern-source and northern-source pre-Illinoian continental till in South Dakota, Iowa, Minnesota, and Missouri. The buried zone of subaerial weathering in Warman Formation till is the most intensely developed weathered zone in the complex sequence of tills and intertill sediments in southern Saskatchewan (E.A. Christiansen, written commun., 1990, 1999; Fullerton and others, in press). It represents subaerial weathering during an interglacial interval of 380,000–400,000 yr duration, prior to Illinoian glaciation (the interval between oxygen isotope stages 14 and 6; see table 2). Till laterally equivalent to till of the Warman Formation has not been identified in Montana or northwestern North Dakota. However, Warman Formation till was identified in a test hole approximately 1.9 km north of the International Boundary, in Saskatchewan northwest of Crosby, N. Dak. (E.A. Christiansen, written commun., 1999; Fullerton and others, in press). The presence of Warman Formation till in the buried ancestral valley of the Missouri River (Meneley and others, 1957) at that site in the Weyburn 1° × 2° quadrangle indicates that till of the same age possibly is present in the buried ancestral valleys of the Missouri and Yellowstone Rivers in extreme northeastern Montana and extreme northwestern North Dakota (Fullerton and others, in press).

ILLINOIAN GLACIATION

On most of the Northern Plains in Montana, the Illinoian glaciation was the most extensive Pleistocene glaciation (see unit IL on the map). An interglacial interval of nearly 500,000 yr duration followed deposition of the middle Pleistocene pre-Illinoian tills and preceded deposition of the Illinoian tills on most of the plains in Montana. That interval was the interval between oxygen isotope stages 16 and 6 (table 2).

Although two or more till units of Illinoian age are superposed in exposures in some areas in Montana, nowhere has a paleosol or weathered zone been identified between till units. Consequently, all of the till units of Illinoian age are interpreted to have been deposited during a single glaciation (correlated with oxygen isotope stage 6; see table 2). A single till unit is observed in most exposures. That till typically is thin (relative to tills of other ages), and commonly it is discontinuous. The field characteristics of the Illinoian tills differ markedly from those of the pre-Illinoian tills (see appendix), and the Illinoian and pre-Illinoian tills are distinguished easily in most exposures and excavations. The field characteristics of Illinoian tills in some areas are similar to those of late Wisconsin tills, however. Consequently, some observers in the past erroneously assumed that the Illinoian and late Wisconsin tills were deposited during a single glaciation. The Pondera till of Smith and others (1959) includes Herron Park till of Illinoian age and Fort Assiniboine till of late Wisconsin age. Most of the Illinoian till (unit IL) in the Havre, Glasgow, and Jordan 1° × 2° quadrangles and the eastern part of the Lewistown quadrangle was credited to a "Wisconsin 1" or "early Wisconsin" glaciation by Colton and others (1961) and Lemke and others (1965). That age assignment was derived from interpretation of aerial photographs, not from stratigraphic investigations.

In most areas beyond the limit of late Wisconsin glaciation (the limit of unit LW on the map), the Illinoian till is thin and discontinuous, and bedrock outcrops are extensive and abundant. Prior to Illinoian glaciation, areas of dissected bedrock terrain, including badlands, formed in Montana and western North Dakota and the pre-Illinoian till was removed or intensely dissected. The Illinoian ice margins advanced into dissected bedrock terrain (badlands) in many areas. The rugged erosional topography was smoothed by the ice, and the relief of the bedrock surface was reduced. Till was plastered as a thin veneer directly on the bedrock surface and on dissected remnants of pre-Illinoian till. The general badland topographic grain was retained. During and after Illinoian deglaciation, streams reoccupied the valleys and largest ravines and removed much of the Illinoian till. Badlands were resurrected, and the Illinoian till was preserved primarily in broad valleys, on flat or gently sloping upland bedrock surfaces, and on surfaces armored by older till or gravel.

The Illinoian Emigrant Gap till (Richmond, 1986a, unpub. mapping and oral commun., 1984–86) east of Glacier National Park (unit IL) is thin and discontinuous. As noted previously, late Bull Lake valley-glacier piedmont-lobe ice obstructed the Emigrant Gap continental ice north of Hall Coulee. Beach gravel east of Hall Coulee, in Montana, deposited in an ice-dammed lake 20 m higher than the limit of late Wisconsin till, and also ice-rafted Canadian Shield erratics 38 m higher than the limit of the late Wisconsin till east of Hall Coulee, were ascribed to a pre-Wisconsin (Illinoian?) glaciation by Karlstrom (1999, 2000a). Apparently, a local lake was bordered by the late Bull Lake mountain ice in the vicinity of Hall Coulee and by the Illinoian Emigrant Gap continental ice farther east. (See discussion of Bull Lake mountain glaciation.)

The Herron Park till on the western plains in Montana, the Markles Point till on the central plains, and the Kisler Butte till on the eastern plains (Fullerton and Colton, 1986) are laterally continuous allostratigraphic units (table 3). The Kisler Butte till is laterally continuous with unit E of Salomon (1974, 1976) or till of the Horseshoe Valley Formation (Fulton, 1976) east of the Yellowstone River and south of the Missouri River in westernmost North Dakota (table 3). Apparently, the Herron Park till is laterally equivalent to the Maunsell till (Stalker, 1963, 1976) and to the "dark gray" till of Stalker (1969, 1976) and Stalker and Wyder (1983) in southeastern Alberta. The Markles Point till on the central plains in Montana and the normal-polarity Kisler Butte till on the eastern plains are laterally equivalent to the normal-polarity lower till member of the Floral Formation of Christiansen (1968, 1992) in southern Saskatchewan (Fullerton and others, in press).

The Herron Park till is thin and discontinuous in the northeastern part of the Cut Bank 1° × 2° quadrangle. Thick sediment of Illinoian and late Wisconsin phases of glacial Lake Cut Bank overlies it, and it is exposed only locally. The Illinoian continental glacial limit (the limit of unit IL) is overlapped by Pinedale 1 till (unit P1) of the Two Medicine outlet-glacier piedmont lobe southwest of Cut Bank (see map and fig. 1). Illinoian till is very thin and discontinuous in the southeastern part of the Cut Bank quadrangle. Residual erratic boulders and cobbles from Canada are scattered on dissected bedrock surfaces. In some areas, neither till nor lake sediment has been identified, and it is not certain whether the erratics are lag clasts derived from eroded till or they were rafted by ice in glacial lakes. Both origins are tenable. The Illinoian glacial limit (limit of unit IL) in that region (between the unit P1 overlap of unit IL and Pendroy on the map) is the western limit of erratic boulders, not the limit of known till exposures. Sediment deposited in Illinoian and late Wisconsin phases of glacial Lake Choteau overlies the till north of Choteau in the Choteau 1° × 2° quadrangle. The Illinoian till is continuous on some upland surfaces east and southeast of Choteau. It is discontinuous from Power to Great Falls, and it is covered by sediment of glacial Lake Great Falls and by sheetwash alluvium in much of that region.

In bluffs adjacent to the Missouri, Teton, and Marias Rivers and their major tributaries in the northeast quarter of the Great Falls 1° × 2° quadrangle and in the Shelby and Conrad areas in the western part of the Shelby quadrangle, the Illinoian and late Wisconsin tills in many areas were removed by erosion. In those areas, the Illinoian till and late Wisconsin till commonly are exposed only in the heads of gullies on hillslopes high above and far away from the major streams. The sediments exposed in the spectacular bluffs adjacent to major rivers are chiefly exhumed pre-Illinoian till units (Havre till) and thick alluvial and lacustrine deposits of middle Pleistocene age.

The Illinoian till limit is south of the Shonkin Sag catastrophic-outburst-flood channel complex on the north margin of the Highwood Mountains (see map). The channels were blocked and overridden by Illinoian ice, and then some of the channels were reoccupied by flood discharge from Illinoian glacial Lake Great Falls. During the Illinoian glaciation, the Missouri River was permanently diverted from the ancestral valley to its present eastward course south of the Bearpaw Mountains. Extensive landslide deposits conceal the till limit between the Highwood Mountains and the Bearpaw Mountains.

The region mapped as Illinoian till (unit IL) south and east of the eastern arm of Fort Peck Lake is primarily dissected bedrock terrain. The till is very thin and discontinuous, and outcrops generally are either eroded Illinoian till or exhumed pre-Illinoian till. Superposed pre-Illinoian and Illinoian tills rarely are exposed. The mapped limit of Illinoian glaciation in that region is the limit of known exposures of Illinoian till. In some areas, the limit of Illinoian glaciation or the limit of pre-Illinoian Pleistocene glaciation, or both, possibly is south of the limit shown on the map. The region at the reentrant northeast of Richey, Mont., (in the Glendive 1° × 2° quadrangle), also is primarily dissected bedrock terrain characterized by outcrops of thin and discontinuous till of Illinoian or pre-Illinoian age, or both ages.

Till of Illinoian age is present as a discontinuous fringe in some places along the margins of the Sweetgrass Hills nunataks (in the Shelby 1° × 2° quadrangle), and it mantles most of the surface of the Boundary Plateau farther east. Regional ice flow was due southward across the Boundary Plateau during maximum Illinoian glaciation (fig. 2*A*). The active ice margin retreated from the plateau surface and then readvanced from the north and northwest. The readvance limit (overlap) on the Boundary Plateau is indicated on the map. A paleosol has not been identified between the two Illinoian till units on the Boundary Plateau, and we interpret the till units to represent two stades during a single glaciation.

Till of Illinoian age is thin and discontinuous on the east margin of the Bearpaw Mountains and on the north margin of the Little Rocky Mountains. However, east of Landusky it is thick and continuous locally. Much of the region mapped as Illinoian till (unit IL) west and south of Fort Peck Lake and between the lake and the Milk River farther north (in the Lewistown, Jordan, and Glasgow 1° × 2° quadrangles) and east of the reservoir (in the Jordan and Glendive quadrangles) is dissected bedrock terrain. Areas of badland topography are extensive. In those regions, the till is very thin and discontinuous; in some areas, it is preserved only as thin patches on flat or gently sloping drainage divides or on gentle hillslopes. North of the Milk River (in the Glasgow 1° × 2° quadrangle), in some areas the till is a thick, continuous blanket overlying pre-Illinoian till and Tertiary gravel (Fullerton and others, in press).

Illinoian till veneers pre-Illinoian till and Tertiary gravel on bedrock benches on the southeast margin of the Peerless Plateau in the Wolf Point 1° × 2° quadrangle. During maximum Illinoian glaciation, regional ice flow was southeastward, and then eastward and northeastward, around the west and south sides of the plateau (fig. 2*A*). A suture between northern-source ice (from western Saskatchewan) and northeastern-source ice (from Manitoba and southeastern Saskatchewan) extended south-southeastward from the Flaxville Plateau nunatak toward the reentrant south of Culbertson, Mont. (fig. 2*A*). The limit of Illinoian glaciation in the Scobey region, Montana, and in the Willow Bunch Lake 1° × 2° quadrangle in Saskatchewan is buried; it was overridden by ice during late Wisconsin glaciation. Till of Illinoian age has not been identified beyond the limit of late Wisconsin glaciation (the limit of unit LW) on the east margin of the Peerless Plateau or on the west margin of the Flaxville Plateau. However, a small area of till that possibly is Illinoian in age (unit IL) is mapped north of the Flaxville Plateau. The till in part of the region in the Wood Mountain 1° × 2° quadrangle that Klassen (1992a) characterized as "patchy ground moraine" here is interpreted to include areas of discontinuous till of Illinoian age (see Fullerton and others, in press).

The northeast-source Kisler Butte till (Fullerton and Colton, 1986), identified in scattered exposures, is overlain and concealed by thick, continuous late Wisconsin till in many parts of the Wolf Point and Glendive 1° × 2° quadrangles. In some areas, however, Illinoian till is at the surface. In most areas, the Illinoian till and late Wisconsin till cannot be distinguished on the basis of surface topographic expression. In areas depicted as Illinoian till (unit IL) in the Yellowstone River drainage basin west and south of Intake, Mont., the till generally is preserved only where it overlies flat or gently sloping, gravel-capped terrace or bench remnants. Commonly, pre-Illinoian (Archer) till is present between the gravel and the Illinoian till. Lake sediment and till of late Wisconsin age commonly overlie the Illinoian till. The terrain between terrace remnants underlain by till and gravel in the Yellowstone River valley is chiefly dissected bedrock or late Wisconsin till overlying bedrock. Owing to intense hillslope erosion and stream dissection in that region, the limit of Illinoian glaciation east and west of the Yellowstone River cannot be accurately delineated. In the Glendive 4° × 6° quadrangle northwest of Intake, Mont., in some places the Illinoian Kisler Butte till overlies (and contains clasts of) Lava Creek B tephra (see previous discussion of the relation between pre-Illinoian Archer till and the Lava Creek B tephra.)

Thin and discontinuous till of Illinoian age is present west and north of the Little Missouri River in the region mapped as unit PW in the Watford City 1° × 2° quadrangle. Unit E of Salomon (1974, 1976) or till of the Horseshoe Valley Formation (Fulton, 1976), in northwestern North Dakota, and the Kisler Butte till in adjacent Montana are laterally continuous stratigraphic units (table 3). The southern limit of till of the Horseshoe Valley Formation, as mapped by Fulton (1976), does not conform to expected topographic controls on ice flow, and outcrops of the till are present south of that limit. The Illinoian glacial limit in that region cannot be accurately delineated on the basis of current information (see discussions of late Pliocene pre-Illinoian glaciation and middle Pleistocene pre-Illinoian glaciation).

Regional ice-flow patterns during maximum Illinoian glaciation are depicted in figure 2*A*. Ice from the dominant northern (Keewatin) dispersal center flowed nearly due south, from western Saskatchewan onto the plains in central Montana. Flow into northeastern Montana from that dispersal center was blocked by ice from a subordinate northeastern dispersal center. The latter flow was generally westward and southwestward, from southern Manitoba and southeastern Saskatchewan into northwestern North Dakota and extreme northeastern Montana (Fullerton and others, in press). The reconstruction of regional ice movement during maximum Illinoian glaciation (fig. 2*A*) is based on the spatial distribution of till of Illinoian age (unit IL on the map), directions of flow of ice-marginal meltwater during maximum Illinoian glaciation, and compositional characteristics of the Illinoian tills (Fullerton and others, in press). Terrain in the Cypress Hills region in Canada, characterized by thin and discontinuous drift, was mapped as (1) "bedrock with patches of glacial drift" (Westgate, 1965b, 1968), (2) "reworked bedrock and rare glacial erratics" Klassen (1991), and (3) "bedrock terrain with drift" or "bedrock with scattered erratics" (Klassen, 1992b). The areal distribution of Illinoian till on the western and central plains in Montana indicates that the areas mapped as Illinoian till (unit IL) in the Cypress Hills region and in the Wood Mountain 1° × 2° quadrangle were glaciated at that time (fig. 2*A*).

Regional ice flow to the regions mapped as Illinoian till (unit IL) in the Cut Bank, Choteau, and Great Falls 1° × 2° quadrangles was west-southwestward, southwestward, and southward (fig. 2A). Ice flowed southward across the East Block and Central Block of the Cypress Hills, across the Wood Mountain Upland, and across the Boundary Plateau (in the Cypress Lake, Wood Mountain, Havre, and Glasgow 1° × 2° quadrangles). It flowed west-southwestward between the West Block of the Cypress Hills and the Sweetgrass Hills, and southward between the West Block and the Bearpaw Mountains. As noted previously, southward-moving ice crossed the Wood Mountain Upland and then flowed southeastward along the west margin of the Peerless Plateau (in the Glasgow 1o × 2° quadrangle). The Wood Mountain ice flowed northeastward on the southeastern edge of the plateau (in the Wolf Point quadrangle). In that region, the ice margin declined in altitude northeastward, and melt-water flow in channels at the glacial limit (north of Wolf Point) was northeastward (Fullerton and others, in press). A glacial lake was dammed between the Markles Point and Kisler Butte ice lobes in the vicinity of Scobey, Mont., and the temporary lake drained southeastward (englacially or superglacially) to the Yellowstone River valley. Extensive areas of glacial lake sediment were mapped by Colton and others (1978) west of Scobey in the region mapped as unit GB.

EARLY WISCONSIN GLACIATION

Laurentide continental till of early Wisconsin age, deposited during oxygen isotope stage 4, has not been identified in Montana. The upper till member of the Floral Formation in southern Saskatchewan (Christiansen, 1968, 1992), of early Wisconsin age, was identified in a test hole approximately 1.9 km north of the International Boundary (E.A. Christiansen, written commun., 1999; Fullerton and others, in press). Its presence in the buried ancestral Missouri River valley (Meneley and others, 1957) northwest of Crosby, N. Dak., in the Weyburn 1° × 2° quadrangle, indicates that till of early Wisconsin age possibly is present in the buried ancestral valleys of the Missouri and Yellowstone Rivers in extreme northeastern Montana and extreme northwestern North Dakota (Fullerton and others, in press). The upper till member of the Floral Formation has not been identified in the southern halves of the Cypress Lake and Wood Mountain 1° × 2° quadrangles (E.A. Christiansen, written commun., 1999). The Brocket till (Stalker, 1963, 1976) and the "light brown till" of Stalker (1969, 1976) in some exposures in southeastern Alberta possibly are early Wisconsin in age.

The youngest Laurentide continental till in the Saint Mary River valley and tributary valleys in Montana east of Glacier National Park (unit LW) was named the "Boundary Creek till" by Richmond (1986a). It was inferred by Karlstrom (1981, 1982, 1984) to be early Wisconsin in age on the basis of two ^{14}C ages for "charcoal" in alluvium overlying the till [42,170±2140 ^{14}C yr B.P. (A–1998) and >38,000 ^{14}C yr B.P. (A–1997)]. The "charcoal," also collected by three other field parties, contains secondary crystals of pyrite in growth position, an indication that it is Cretaceous or Tertiary lignite (Fullerton and Colton, 1986). A calcium carbonate soil horizon at a depth of 93 cm in the alluvium yielded a ^{14}C age of 4950±80 yr B.P. (Beta–10458) (Karlstrom, 1999). The Boundary Creek till is late Wisconsin in age.

LATE WISCONSIN GLACIATION

Late Wisconsin Laurentide glaciation in southern Alberta

Multiple Laurentide continental glaciations in southern Alberta were recognized on the basis of stratigraphic investigations (Vernon, 1962; Stalker, 1963, 1969, 1976, 1983; Alley, 1972, 1973; Stalker and Churcher, 1972; Stalker and Harrison, 1977; Stalker and Wyder, 1983; Jackson and others, 1989). Subsequent studies in west-central Alberta indicated that (1) only one Laurentide glaciation is recorded by till at and northwest of Edmonton, and (2) the single glaciation in that region (far north of the part of Alberta shown on the map) was late Wisconsin in age (Liverman and others, 1989; Young and others, 1994). Because ice from the Edmonton region moved southeastward through southern Alberta into Montana during late Wisconsin glaciation (fig. 2*B*), and because Laurentide tills of pre-Wisconsin age have not been identified in the Edmonton region, one might assume that all of the Laurentide tills in southwestern Alberta also are late Wisconsin in age. That assumption is not supported by stratigraphic data from Montana and southeastern Alberta.

As indicated in figure 1, and discussed previously in relation to Illinoian and pre-Illinoian Pleistocene glaciations, regional ice movement from the Edmonton region southeastward into Montana occurred **only during late Wisconsin glaciation**. Regional Laurentide ice flow in southern Alberta during pre-Wisconsin glaciations possibly was southwestward (fig. 2*A*) at times when the Edmonton region was not glaciated. Keewatin ice dispersal centers during pre-Illinoian and Illinoian glacial maxima possibly were south of the location(s) of the Keewatin center during late Wisconsin glaciation, and flow of ice from the Keewatin dispersal center was not obstructed by westward-flowing ice from an eastern dispersal center.

As noted previously, recent studies of till stratigraphy in southwestern Alberta (Jackson, 1994; Jackson and others, 1996; Jackson and Little, 2003) concluded that all of the mountain tills and Laurentide continental tills in exposures in southwestern Alberta, including the tills described by Stalker (1963, 1976), Stalker and Harrison (1977), and Alley (1972, 1973), are late Wisconsin in age. Stratigraphic units in southeastern Alberta, distinguished by Vernon (1962), Stalker (1963, 1969, 1976, 1983), Stalker and Churcher (1972), and Stalker and Wyder (1983), were not traced westward into southwestern Alberta in an effort to integrate stratigraphic sequences in southeastern and southwestern Alberta, and stratigraphic units and glacial landforms in Montana were not examined in an effort to construct an integrated stratigraphic, geomorphic, and chronologic framework of glaciation in southern Alberta and adjacent Montana.

Late Wisconsin Laurentide glaciation in, and north and northeast of, Waterton Lakes National Park, Alta., currently is interpreted to be represented by tills of two glacial advances (see map): a C1 advance or "maximum glacial advance" and a C2 readvance (Little, 1995a,b, 1998a–c; Leboe, 1996, 1998b; Jackson, 1998; Holme, 1998a, Holme and others, 1998, 2000; Jackson and others, 1999; Jackson and Little, 2003). In general, the C2 readvance is the Kimball readvance of Horberg (1954), Wagner (1966), and Karlstrom (1987b). A readvance limit in southeastern Alberta, the same age as the Kimball limit in southwestern Alberta, has not been delineated. The C2 glacial limit overlaps the C1 glacial limit in the Waterton Lakes National Park region (Little, 1995b, 1998b). West of long 114° W., the C2 limit apparently is the limit of late Wisconsin Laurentide continental glaciation and the older C1 limit is concealed (Holme, 1998a,b; Holme and others, 1998, 2000).

Distinctive blocks and boulders of the Foothills erratics train in western Alberta (Stalker, 1956; Morgan, 1969; Tharin, 1969; Jackson, 1993; Jackson and others, 1997, 1999) are present on surfaces of till deposited during the C2 (Kimball) readvance. The erratics also are present on till surfaces in the United States as far south as Conrad, Mont., in the Shelby 1° × 2° quadrangle (A.M. Stalker, written commun., 1980–81; R.W. Barendregt, oral commun., 2003). It is important to note that the Foothills erratics train was traced into Montana by Stalker on the basis of the presence of (1) **subangular blocks** of quartz-granule conglomerate and (2) **subangular blocks** of limestone from the Rocky Mountains. Jackson's "Foothills erratics" include **rounded boulders and cobbles** of bedrock types that Stalker did not believe were necessarily part of the train of **superglacially transported** erratics.

Extensive glacial lake deposits overlying late Wisconsin till were mapped in southwestern Alberta (Horberg, 1954; Wagner, 1966; Alley, 1972; Shetsen, 1980, 1987; Leboe, 1998a,b; Little, 1998a–c; Holme, 1998a). However, reconstruction of lake sequences, changes of lake levels, and relations of ice-margin positions to specific lake levels and lake spillway channels in Alberta was attempted only by Horberg (1954), Wagner (1966), and Alley and Harris (1974). Horberg attempted to relate some ice-margin positions, lakes, and channels in Alberta to ice-margin positions, lakes, and channels in Montana.

All of the rivers east of the Continental Divide in central and southern Alberta flowed eastward to glacial lakes dammed by the Laurentide ice sheet or by confluent Laurentide and mountain ice, and the Laurentide and mountain glacial meltwater also entered the lakes. Each lake drained to a lower lake to the south or southeast. Ultimately, all of the combined runoff and meltwater from southwestern Alberta entered Montana. Correlation of late Wisconsin ice-margin positions in southwestern Alberta and Montana is dependent on determination of the glacial lake history. The glacial and deglacial lake drainage history is an important tool in reconstruction of late Wisconsin deglaciation. Most of the western and central plains in Montana was deglaciated prior to formation of extensive deglacial lakes at altitudes lower than 1,220 m in southern Alberta.

A catastrophic subglacial megaflood in Canada was hypothesized to have delivered phenomenal volumes of water and sediment to the western Montana plains in the Shelby and Havre 1° × 2° quadrangles during, or very soon after, maximum late Wisconsin glaciation. Drumlins in the Livingstone Lake region in northern Saskatchewan (far north of the part of Saskatchewan shown on the map) were interpreted to have formed as a result of subglacial melt-water flow (Shaw and Kvill, 1984). The flood water of the catastrophic "Livingstone Lake event" in Saskatchewan was interpreted to have flowed southwestward and southward **beneath the Laurentide ice sheet** in Alberta. Runoff and meltwater from mountain glaciers in Alberta was added to the flood. Rains and others (2002, p. 64) stated, "It is very likely that meltwater stored in, under, and ice-marginal to the Cordilleran Bow–Kananaskis valley glaciers and piedmont tongue was released simultaneously with the theorized Livingstone Lake event underburst . . . ." The combined flow entered Montana (Rains and others, 1993, 2002; Shaw, 1996, 2002; Shaw and others, 1996, 2000; Beaney, 2002; Munro-Staciuk and Shaw, 2002).

According to Shaw and others (1989) and Shaw (1996), the estimated mean depth of the subglacial flood in the Livingstone drumlin field in Saskatchewan was 20–40 m, the estimated width of flow was ≥150 km, and the calculated duration of flow was 16–162 days. The calculated maximum (peak) volume of catastrophic subglacial melt-water flow in **northern Saskatchewan** was 84×10^3 km^3 (Shaw, 1989, 1996; Shaw and others, 1989, 1996; Rains and others, 1993), and the calculated maximum (peak) flood discharge was 60×10^6 m^3/s (Shaw and others, 1989; Shaw, 1996). That discharge is much larger than the minimum peak discharge (17×10^6 m^3/s) of the largest single event of glacial Lake Missoula floods in the Channeled Scabland in eastern Washington, calculated by O'Connor and Baker (1992). The volume of water released by the hypothetical subglacial flood in the drumlin field in **northern Saskatchewan** is equivalent to a global eustatic sea level rise of 23 cm (Shaw, 1989, 1990, 1996, 2002; Shaw and others, 1989). The volume of water in the hypothetical subglacial drainage system would have increased greatly southward in Alberta and Montana, owing to progressive addition of water (Shaw, 1990, 2002). The volume and discharge of water beneath the ice in Montana would have been truly phenomenal, and landforms produced by the flow should be conspicuous. Rains and others (1993, p. 326) stated that the hypothetical flood **in Saskatchewan** "removed many thousands of cubic kilometers of sediment and bedrock." Shaw (1996, p. 229) stated, "At least 10^4 km^3 of sediment are estimated to have been eroded **from Alberta alone** during the Livingstone event" (our emphasis). The flood also would have eroded sediment in Montana.

Tunnel channels that cross a bedrock divide in the southeast corner of Alberta and extend into Montana were interpreted to have been formed by catastrophic flood water associated with the hypothetical Livingstone Lake event (Beaney, 1998, 2002; Beaney and Shaw, 2000; Beaney and Hicks, 2000; Shaw, 2002). We are not aware of any convincing evidence that those channels in Montana were formed by catastrophic subglacial meltwater flow. A subaerial origin, as spillway channels of deglacial lakes (see Westgate, 1964, 1965a,b, 1968; Shetsen, 1987), is compatible with known field data, and most (or all) of the drumlins and flutes in Montana apparently were formed subglacially by ice, not by subglacial catastrophic floods. Meltwater from Alberta did enter Montana, and some drumlins apparently are composed in part of ice-molded glaciofluvial and glaciolacustrine deposits. However, neither the surface landforms nor the stratigraphic sequence of glacial and interglacial deposits are compatible with occurrence of a postulated Livingstone Lake event in Montana. Processes of formation of meltwater channels, drumlins, and flutings in Alberta proposed by Evans and Campbell (1990, 1992), Evans (1994, 1996, 2000), and Evans and others (1999) are compatible with morphologic and stratigraphic data in Montana.

Shaw and others (1996, p. 1163) stated, "Generalized maps of Laurentide landform associations . . . show large zones of drumlinized and fluted till, along with low-relief hummocky terrain, in parts of northern Montana, North Dakota, South Dakota, and farther eastward in the United States. We believe that these suites were moulded in part by subglacial sheet-flood action, particularly the Livingstone Lake event . . . ." Those suites are on, or are composed of, stratigraphic units of different ages in different places. Shaw (2002, p. 15) indicated that the path(s) of meltwater flow through Montana, North Dakota, South Dakota, and other states to the Gulf of Mexico illustrated by Shaw and others (1996, fig. 9) were identified from the digital elevation model, at 1:3,500,000 scale, of Thelin and Pike (1991). Apparently, no attempt was made to integrate detailed stratigraphic, chronologic, and geomorphic data in Montana and the Dakotas with the flood path(s) interpreted from the digital elevation model.

During maximum late Wisconsin glaciation in Montana, the hypothetical floodwater flowed beneath eastward-flowing, northwest-source, ice of the Havre lobe. Glacial Lake Jordan occupied the lowland between Fort Assiniboine ice of the Havre lobe and westward-flowing, northeast-source, Crazy Horse ice of the Glasgow lobe (see map and fig. 1*B*). All of the sediment transported to the ice margin must have entered Lake Jordan. Lake density-current underflow-fan deposits, delta deposits, and (or) other lake deposits that can be interpreted to be products of the megaflood in Saskatchewan and Alberta have not been identified in the basin of glacial Lake Jordan, in the basin of any other glacial lake, or in the basin of any extraglacial lake in Montana. The lake sediment is thin (generally less than 2 m thick, over Illinoian till or bedrock) and very discontinuous; there is no lake fill.

Morphologic and sedimentologic evidence of a "supercolossal, catastrophic, subglacial megaflood" on the Montana plains has eluded us. The floodwater and the sediment produced by the subglacial flood erosion during the hypothetical Livingstone Lake event in Saskatchewan and Alberta did not cross the International Boundary. Some of the meltwater channels in Montana were formed by subglacial drainage. However, they are not large enough to have accommodated the Livingstone Lake event flood, and many are filled or partly filled with pre-Illinoian, Illinoian, and (or) late Wisconsin till. Many of the meltwater channels near, at, and beyond the limits of ice lobes in Montana are too small to have accommodated the flood and (or) they contain thick deposits of till. The areal distributions, volumes, thickness, and sedimentologic properties of the sediments deposited in the late Wisconsin proglacial and deglacial lakes on the western and central Montana plains are consistent with an interpretation that the sediments were derived primarily from melting ice and river inflow in those regions.

Late Wisconsin Laurentide glaciation in Montana and northwestern North Dakota

The limits of late Wisconsin glacial readvances and positions of ice-margin stillstands on the map were selected to illustrate the general pattern of net deglaciation. Many minor readvance limits or temporary ice-margin positions are not shown. Some of the limits are delineated by stratigraphy (a distinct till unit was deposited as a result of a glacial readvance) or by discrete till ridges (end moraine ridges), or both. Some are delineated by cross-cutting relations of landforms. Others are delineated by side-valley ice-marginal meltwater channels that nearly parallel, rather than transect, topographic contours. Some of the positions in Montana are the distal ridges of sets of arcuate washboard moraines, ice-crack moraines, or minor moraines that are composed of till. Ice-molded landforms (flutings, drumlins, and rock drumlins), particularly in the Shelby, Havre, and Jordan $1^{\circ} \times 2^{\circ}$ quadrangles, are not mapped.

The distal limits of selected glaciotectonic (ice-thrust) deposits or structures are shown on the map. The glaciotectonic materials are masses of bedrock and surficial deposits that were transported, thrust, stacked, and (or) deformed by glacial ice. Glaciotectonic deposits or structures in the ancestral Missouri River valley (in the Brockton–Medicine Lake region in the Wolf Point $1^{\circ} \times 2^{\circ}$ quadrangle) were overridden by late Wisconsin ice. The limits of those buried glaciotectonic deposits and structures (Fullerton and others, in press) are not shown on the map.

On the plains in Montana, hummocky moraine generally is confined to uplands or other areas of high-altitude bedrock surfaces or areas of relatively high bedrock relief. Hummocky moraine nearly everywhere is in areas of compressive ice flow. In some areas, glaciotectonic (ice-thrust) deposits and structures are associated with hummocky topography. Much of the end moraine mapped by Colton and others (1961, 1963) and Lemke and others (1965) is hummocky stagnation moraine that includes few or no discrete, oriented till ridges. Munro and Shaw (1997) proposed that hummocky terrain in south-central Alberta was produced by subglacial meltwater erosion, not by stagnation of debris-covered glacial ice, and Munro-Stasiuk (1999a,b, 2000) proposed that in some places in Alberta it is composed of subglacial lake deposits, not till. We are not aware of evidence that extensive areas of hummocky moraine in Montana were formed by subglacial meltwater erosion or by subglacial lake deposition. We favor an interpretation that (at least in most regions) it was formed by collapse and gravitational loading of sediment associated with stagnant ice (Gravenor and Kupsch, 1959; Stalker, 1960; Evans, 1994; Eyles and others, 1999; Mollard, 2000; Boone and Eyles, 2001).

Surfaces on late Wisconsin till in Montana generally are smooth where modified by waves and currents in glacial lakes. Thick, continuous lake sediment generally overlies the till only at low altitudes in major valleys. Typically, the deglacial lake sediment is predominantly silt and fine sand less than 6 m thick. Proglacial lakes dammed in major drainage basins during maximum late Wisconsin glaciation (Colton and Fullerton, 1986) were areally extensive, and individual phases or "levels" of those lakes were of short duration. During maximum late Wisconsin glaciation, glacial Lake Jordan was dammed on the east by Crazy Horse ice and bordered on the west by Loring and Fort Assiniboine ice. The lake sediment overlies till of Illinoian age in the southern part of the Glasgow $1^{\circ} \times 2^{\circ}$ quadrangle and the northern part of the Jordan quadrangle. Delta deposits, density-current underflow-fan deposits, and (or) other lake deposits that can be attributed to a "supercollosal, catastrophic, subglacial megaflood" that was inferred to have occurred in northern Saskatchewan (discussed previously in relation to Laurentide glaciation in southern Alberta) have not been identified in the basin of glacial Lake Jordan or in any other lake basin in Montana.

Temporal and spatial relations between late Wisconsin Laurentide glaciation and Pinedale mountain glaciation in the Glacier National Park–Waterton Lakes National Park region are outlined in the discussion of Pinedale glaciation. Early Pinedale valley-glacier piedmont-lobe and outlet-glacier piedmont-lobe ice margins advanced from the mountains onto the plains in Montana and Alberta. During the subsequent (early Pinedale–middle Pinedale) interstade, active mountain ice margins retreated and areas of buried, stagnant or dead ice remained on the land surface. A middle Pinedale regional advance of mountain ice margins followed. In many areas, early Pinedale deposits were buried by middle Pinedale mountain glacier deposits or by Laurentide continental ice sheet deposits. Buried early Pinedale ice in some places was overridden and reactivated, as a result of loading by the middle Pinedale mountain ice or Laurentide continental ice. Reactivation of buried early Pinedale mountain ice by overriding Laurentide ice produced till in which mountain and Laurentide till matrix and clast compositions are juxtaposed or intermixed.

The middle Pinedale mountain glacial maximum and the late Wisconsin Laurentide glacial maximum were contemporaneous (fig. 3, table 3). Chronometric constraints from the Northern Plains in Montana, Alberta, and Saskatchewan (Christiansen, 1968, 1971; Fullerton and Colton, 1986; Fullerton and others, 1995, 2000; Jackson and others, 1997, 1999) and the Rocky Mountains in Montana, Wyoming, and Colorado (see previous discussion of age controls for Pinedale glaciation) indicate that the Laurentide continental glacial maximum and the middle Pinedale mountain glacial maximum occurred approximately 23,360 CAL yr B.P. (20,000 ^{14}C yr B.P.), during oxygen isotope stage 2. That glacial maximum was approximately 2,300–2,800 CAL yr later than the early Pinedale mountain glacial maximum.

Richmond (1986a) named the Laurentide continental till (unit LW) east of Glacier National Park (in the Cut Bank 1° × 2° quadrangle) the Boundary Creek till. The "outer continental" till of Horberg (1954) or the "maximum glacial advance" till or C1 till of Little (1998a–c), Jackson and others (1996), and Jackson and Little (2003) and the Boundary Creek till are the same stratigraphic unit. Glacial Lake St. Mary (Horberg, 1954; Richmond, 1965b, 1986a) was dammed by the Boundary Creek ice margin in the Saint Mary River valley. The middle Pinedale valley-glacier ice margin at its maximum extent, at Babb, Mont., was fronted by glacial Lake St. Mary at that time (fig. 3). The sequence of glacial lakes was discussed previously in relation to middle Pinedale glaciation. Glacial Lake St. Mary drained to the North Fork Milk River and the North Fork flowed into glacial Lake North Fork in Alberta. Outflow from the highest late Wisconsin phase of Lake North Fork entered the highest late Wisconsin phase of glacial Lake Twin River southeast of Del Bonita, Alta. Glacial Lake Twin River was dammed by the Laurentide ice sheet in the Milk River valley in Montana during maximum late Wisconsin glaciation. Drainage from Lake Twin River flowed into the highest late Wisconsin phase of glacial Lake Cutbank. The highest late Wisconsin phase of Lake Cutbank also was dammed by the Laurentide ice sheet during maximum late Wisconsin glaciation (fig. 3). The Pinedale 2 glacial maximum of the Two Medicine outlet-glacier piedmont lobe in the Cut Bank 1° × 2° quadrangle also was contemporaneous with the late Wisconsin Laurentide glacial maximum (fig. 3, table 3).

The Fort Assiniboine till on the western plains and most of the central plains in Montana and the Loring till north and east of the Boundary Plateau (Fullerton and Colton, 1986) are laterally continuous allostratigraphic units (table 3). The Fort Assiniboine and Loring ice lobes merged in the Milk River valley between the Boundary Plateau and Hinsdale, southeast of the Boundary Plateau (Fullerton and others, in press), and also in Saskatchewan (Klassen, 1991, 1992b, 2002) between the East Block of the Cypress Hills and the Boundary Plateau (fig. 2*B*). The Fort Assiniboine till is laterally continuous with the "outer continental" till of Horberg (1954) and with Buffalo Lake till and "contorted" till of Stalker (1963, 1969, 1976), Stalker and Harrison (1977), and Stalker and Wyder (1983) in southeastern Alberta. The Loring till on the central plains in Montana and the Crazy Horse till on the eastern plains (Fullerton and Colton, 1986) are laterally continuous with till of the Battleford Formation of Christiansen (1968, 1971, 1992) in southern Saskatchewan (Fullerton and others, in press). The Crazy Horse till also is laterally continuous with unit F of Salomon (1974, 1976) or till of the Snow School Formation of Fulton (1976) east of the Yellowstone River and south of the Missouri River in western North Dakota (table 3).

The late Wisconsin Fort Assiniboine till on the western Montana plains (Fullerton and Colton, 1986), was deposited by ice associated with regional flow from central Alberta southeastward into Montana (fig. 2*B*). **That ice dispersal pattern** (Dyke and others, 1982; Dyke and Prest, 1987a,b; Andrews and Fulton, 1987) **did not occur during any previous Pleistocene glaciation on the Montana plains**. Regional ice flow during maximum late Wisconsin glaciation in northeastern Montana and northwestern North Dakota was southwestward (fig. 2*B*). A tongue of ice extended northward between the Flaxville Plateau and the Peerless Plateau (in the Wolf Point 1° × 2° quadrangle). At that time, apparently the ice margin farther north (south of Twin Valley and Conrach, Sask.) was on a northwest-southeast trending bedrock physiographic high (plateau). That limit was overridden during a subsequent glacial readvance (see map).

The limit of a major regional glacial readvance that culminated approximately 16,350 CAL yr B.P. (14,000 ^{14}C yr B.P.) is mapped in the Williston, Wolf Point, Willow Bunch Lake, Wood Mountain, Foremost, and Lethbridge 1° × 2° quadrangles (fig. 2*C*). After collapse of the predominant northeastern ice dispersal center depicted in figure 2*B*, the Keewatin dispersal center(s) became the predominant source of regional ice flow in southeastern Alberta, southern Saskatchewan, southwestern Manitoba, northeastern Montana, and northwestern North Dakota. The limit of the readvance from north of Scobey, Mont., northwestward to the eastern part of the Wood Mountain 1° × 2° quadrangle is the southernmost limit of late Wisconsin glaciation (the earlier late Wisconsin limit was overridden and buried in that region). The readvance limit in Montana is marked by the Medicine Lake and Scobey moraines. In southernmost Saskatchewan, it is marked by the East Poplar moraine (Parizek, 1961, 1964, 1969; Whitaker, 1974) and the Thomson Lake moraine (Whitaker, 1965, 1967; Klassen, 1992a) in the Willow Bunch Lake and Wood Mountain 1° × 2° quadrangles. Farther north and west in Saskatchewan, the readvance limit is associated with the Aikins moraine (Christiansen, 1959; Whitaker, 1970) and the Verlo and Fox Valley moraines (David, 1964). In southeastern Alberta, it is marked by the Etzikom moraine (Westgate, 1964, 1965a,b, 1968) and the Lethbridge moraine (Horberg, 1952; Stalker, 1962, 1977; Shetsen, 1980, 1987, and unpub. mapping, 1980). The late Wisconsin till south, southwest, or west of that readvance limit in northeastern Montana, northwestern North Dakota, and nearly all of southern Saskatchewan was deposited as a result of regional ice movement from a northeastern ice-dispersal center (fig. 2*B*). The youngest till north of that limit was deposited as a result of regional ice movement from a Keewatin dispersal center (fig. 2*C*). The readvance limit in North Dakota and Montana in many places is the southern limit of extensive belts of stagnation moraine (Fullerton and others, in press).

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APPENDIX. FIELD CHARACTERISTICS OF LAURENTIDE TILLS ON THE MONTANA PLAINS

INTRODUCTION

The field characteristics of pre-Illinoian, Illinoian, and late Wisconsin Laurentide tills are sufficiently distinctive that they are the primary tools for identification of tills on the plains in Montana. Textural (grain-size) data are not diagnostic. In most areas, all of the tills, regardless of age, were derived in large part from the same bedrock materials. The matrix of each till generally is calcareous clay loam or loam. The matrix textures locally are atypical, however, in that compositional facies can reflect incorporation of older surficial materials by ice. For example, where glacial ice incorporated large volumes of Tertiary quartzite-pebble and argillite-pebble gravel, the matrix of till of any age is loamy sand or sand and erratic bedrock clasts from Canada are extremely rare. Where ice incorporated valley-fill alluvium, commonly the till matrix is sandy loam and sandy clay loam. In some areas where ice incorporated lake sediment, the matrix varies abruptly from clay to sand, laterally and vertically. Colors (Munsell Products, 1973) of unoxidized tills also are not diagnostic. They reflect colors of incorporated bedrock and older surficial materials.

The most useful distinguishing field characteristics of tills on the plains in Montana are (1) dry and moist matrix colors of oxidized till; (2) extent and dry and moist colors of secondary iron oxide and manganese oxide stains, flakes, scales, crusts, and fillings; (3) nature and extent of secondary calcium carbonate and selenite (gypsum) accumulation; (4) compaction and induration; (5) matrix structure, fracture, and parting; (6) absence or presence and character of joints; and (7) abundance of clasts, particularly distinctive erratic boulders and cobbles. Identification of both oxidized and unoxidized till(s) in an exposure or excavation generally is based on multiple field characteristics.

PRE-ILLINOIAN TILLS—HAVRE, PERCH BAY, AND ARCHER TILLS OF FULLERTON AND COLTON (1986)

The matrix of till of pre-Illinoian age generally is calcareous clay loam and loam. Locally, it is loamy sand, sandy loam, sandy clay loam, or silty clay loam. Dry matrix colors (Munsell Products, 1973) of oxidized till are hues 10YR, 2.5Y, and 5Y. Value and chroma generally are dark grayish brown to light grayish brown, brown and pale brown, yellowish brown and light yellowish brown, light brownish gray, light olive gray, pale olive, and pale yellow. In strong sunlight and from a distance, the till in some exposures appears to have a yellow or orange tint, owing to a film of iron oxide particles on the surface. From a distance, unoxidized till in some exposures appears to be black; however, value and chroma are very dark gray and dark gray. Stains and thick scales, crusts, and fillings of secondary iron oxide or manganese oxide, or both, are common on joint surfaces and parting surfaces. Hues of oxide colors generally are 10YR, 7.5YR, 5YR, and 5Y. Value and chroma typically are strong brown, dark brown and brown, dark yellowish brown, yellowish red, reddish yellow, and olive. The till generally is overconsolidated and indurated. Typically, it is massive. Commonly, the till resists penetration by field excavating tools–a pick or mattock leaves only a dent in the till matrix. Locally, unoxidized till at depth is soft. In some areas, the till contains abundant pods, lenses, stringers, and deformed beds of stratified materials, or masses of bedrock, indicative of glaciotectonic deformation. Structure in oxidized till commonly is prismatic or platy, and the till commonly breaks into bladelike or angular pieces 2.5–10 cm in length. Where the matrix is extremely silty, in some places it has conchoidal fracture. Where it is massive and very hard, the oxidized and unoxidized till commonly breaks around (not through) cobbles, pebbles, and large granules, producing molds in the excavated till. Clayey till is sticky when moist. The till is extremely resistant to erosion. It forms hoodoos and spires in some areas of intense dissection. In large exposures adjacent to major streams (for example, in stream cuts adjacent to active meanders), the steep basal slopes above bedrock (and, in some places, above so-called "preglacial" gravel) commonly are underlain by pre-Illinoian till. Polygonal joints generally are very conspicuous where the till is thick. In large exposures, the till commonly appears to be oxidized uniformly to great depths. However, the apparent oxidation at depth in many places is related to oxidation in, and adjacent to, vertical joints in the till–the till is unoxidized only a few centimeters from the joints. In some exposures, individual crystals of selenite (gypsum) 6–50 mm in length, clusters of crystals, or crusts or irregular masses of selenite are conspicuous in joints. Scales, crusts, or plates of selenite are present on parting surfaces in some places. Slickensides are abundant on parting surfaces in some exposures of clayey till. Where the till is more than 10 m thick, in some areas it is very sparingly pebbly; boulders and large cobbles are extremely rare or absent in exposures, and the till contains few inclusions or clasts of stratified sediments. In some areas, the ice incorporated large quantities of gravel composed of quartzite and argillite clasts (for example, in the Milk River valley between Glasgow and Wolf Point, Mont.). The sandy till in those areas is glutted with quartzite and argillite pebbles and cobbles, and erratic clasts of limestone, dolomite, and igneous and metamorphic rocks from Canada are absent in many small exposures.

ILLINOIAN TILLS—HERRON PARK, MARKLES POINT, AND KISLER BUTTE TILLS OF FULLERTON AND COLTON (1986)

The matrix of till of Illinoian age generally is calcareous clay loam and loam; locally, it is sandy loam, sandy clay loam, or silty clay loam. Dry matrix colors (Munsell Products, 1973) of oxidized till are hues 10YR and 2.5Y. Typically, 10YR is predominant in Illinoian tills in the northern part of the glaciated plains and 2.5Y is predominant farther south. Value and chroma generally are dark grayish brown to light grayish brown, brown to very pale brown, light yellowish brown, and light brownish gray. Where tills of different glaciations are superposed, at a distance the Illinoian till commonly is lighter in color than the pre-Illinoian till or late Wisconsin till in the same exposure. Stains, flecks, or scales of secondary iron oxide and (or) manganese oxide are common on joint and parting surfaces (however, the thick oxide scales, crusts, and joint fillings that are characteristic of pre-Illinoian tills are not present in Illinoian tills). Hues of oxide colors also are 10YR and 2.5Y; value and chroma generally are very dark grayish brown, very dark brown, very pale brown, dark yellowish brown and yellowish brown, brownish yellow, and yellow. The till generally is very compact, but not extremely dense and heavy. Unoxidized till typically is very tough–it resists penetration by a pick or spade when moist or dry. Commonly, weathered till breaks through granules and small pebbles. Generally, the till is massive or it has blocky structure; locally, it has platy structure. Parting in some places is irregular or prismatic; commonly, the till breaks into small irregular pieces. Clayey till is sticky when moist. The till generally is stable on hillslopes; creep and slump are not obvious in most places. Faces in exposures (for example, stream cuts) do not slump readily. Because the till in exposures generally is less than 1.8 m thick and it is commonly partly covered or obscured by slumped debris, jointing in the Illinoian tills has not been examined closely. Polygonal joints are conspicuous in some exposures. Vertical joints commonly are spaced 0.6–10.2 cm apart. Joints and parting surfaces in some places are coated or filled with secondary white, powdery calcium carbonate or gypsum, or both. Selenite (gypsum) crystals 1–4 mm in length, or clusters of crystals, locally are common on joint surfaces. Calcium carbonate or gypsum filaments are common in pedogenic soil profiles developed in the till. Buried soils in Illinoian till at some sites are truncated petrocalcic soils; truncated calcic pendants extend downward into the till. Truncated petrocalcic soils have been observed only in tills of Illinoian age (in contrast, none of the truncated buried soils in tills of pre-Illinoian age examined was a petrocalcic soil). In most regions, the clasts are predominantly granules and pebbles, and they are not abundant. In many areas south of the limit of late Wisconsin glaciation, the Illinoian till is intensely dissected or the surface has been intensely deflated, or both. In those areas, lag (residual) erratic boulders and cobbles are common to abundant on till surfaces; however, boulders and large cobbles typically are rare within the till in exposures. Erratic boulders are abundant in Illinoian till in some places–for example, superposed units of very bouldery and cobbly Illinoian till are exposed on the southeast margin of the Boundary Plateau in the Glasgow $1^\circ \times 2^\circ$ quadrangle. South of the Shonkin Sag flood-channel complex in the Great Falls $1^\circ \times 2^\circ$ quadrangle, in one area the Illinoian till limit is represented by a ridge of boulders and cobbles. The matrix between the boulders and cobbles in some places is chiefly wind-blown sediment. The limit of an Illinoian glacial readvance on the western part of the Boundary Plateau shown in the Havre $1^\circ \times 2^\circ$ quadrangle is an end-moraine ridge containing abundant boulders. Paleosols have not been identified between superposed Illinoian till units anywhere on the plains in Montana.

LATE WISCONSIN TILLS—FORT ASSINIBOINE, LORING, AND CRAZY HORSE TILLS OF FULLERTON AND COLTON (1986)

The matrix of till of late Wisconsin age generally is calcareous clay loam and loam; locally, it is sandy loam, sandy clay loam, or silty clay loam. Dry matrix colors (Munsell Products, 1973) of oxidized till nearly everywhere are restricted to 10YR hue. Value and chroma generally are very dark grayish brown to grayish brown, dark brown to pale brown, and light brownish gray. Intense stains of secondary iron oxide and manganese oxide are uncommon; there are no thick oxide scales, crusts, or fillings. The till generally is weakly compact to compact, massive, and friable. In general, the till is soft where clayey, firm where silty, and loose to firm where sandy. In stagnation moraine, it is commonly soft and loose. Crumb or popcorn structure is common in the upper part of the till in some exposures; irregular blocky structure is common at depth. Parting typically is irregular or prismatic. Oxidized till commonly is platy and fissile. Clayey till is sticky when moist. The till moves by creep on hillslopes; it tends to slump in exposures. Joints are inconspicuous or absent. Joints commonly are confined to the B horizon of the pedogenic soil developed in the till. Joints and parting surfaces in some places are coated by secondary white, powdery calcium carbonate or gypsum, or both. Calcium carbonate or gypsum filaments are common in the pedogenic soil profile. Selenite (gypsum) crystals longer than 2 mm are uncommon on joint and parting surfaces. Erratic pebbles, cobbles, and boulders from Canada typically are more abundant in late Wisconsin till than in older tills. Also, the average diameter of boulders typically is larger than that in older tills. Surface boulders are not deeply embedded in the till surface.