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William J. Wayne
University of Nebraska-Lincoln, wwayne3@unl.edu

James S. Aber
Emporia State University

Sherry S. Agard
US Geological Survey, Denver

Robert N. Bergantino
Montana Bureau of Mines and Geology, Butte, Montana

John P. Bluemle
North Dakota Geological Survey

See next page for additional authors

Authors
William J. Wayne, James S. Aber, Sherry S. Agard, Robert N. Bergantino, John P. Bluemle, Donald A. Coates, Maurice E. Cooley, Richard F. Madole, James E. Martin, Brainerd Mears Jr, Roger B. Morrison, and Wayne M. Sutherland

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Chapter 15

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William J. Wayne
Department of Geology, University of Nebraska, Lincoln, Nebraska 68588
James S. Aber
Earth Science Department, Emporia State University, Emporia, Kansas 66801
Sherry S. Agard
U.S. Geological Survey, Box 25046, Denver Federal Center, Denver, Colorado 80225
Robert N. Bergantino
Montana Bureau of Mines and Geology, Butte, Montana 59701
John P. Bluemle
North Dakota Geological Survey, 600 E. Boulevard Ave., Bismarck, North Dakota 58205
Donald A. Coates
U.S. Geological Survey, Box 25046, Denver Federal Center, Denver, Colorado 80225
Maurice E. Cooley
Water Resources Division, U.S. Geological Survey, Cheyenne, Wyoming 82071
Richard F. Madole
U.S. Geological Survey, Box 25046, Denver Federal Center, Denver, Colorado 80225
James E. Martin
South Dakota Geological Survey and South Dakota School of Mines and Technology, Rapid City, South Dakota 57701
Brainerd Mears, Jr.
Department of Geology, University of Wyoming, Laramie, Wyoming 82071
Roger B. Morrison
Morrison and Associates, 13150 West 9th Avenue, Golden, Colorado 80401
Wayne M. Sutherland
Wyoming Geological Survey, Laramie, Wyoming 82071

INTRODUCTION

William J. Wayne, Roger B. Morrison, and Richard F. Madole

PHYSIOGRAPHIC AND PRE—QUATERNARY GEOLOGIC SETTING

The Great Plains physiographic province lies east of the Rocky Mountains and extends from southern Alberta and Saskatchewan nearly to the United States–Mexico border. This chapter covers only the northern part of the unglaciated portion of this huge region, from Oklahoma almost to the United States–Canada border, a portion that herein will be referred to simply as the Northern Great Plains (Fig. 1).

This region is in the rain shadow of the Rocky Mountains. Isoheyets are roughly longitudinal, and mean annual precipitation decreases from about 750 mm at the southeastern margin to less than 380 mm in the western and northern parts (Fig. 2). Winters typically are cold with relatively little precipitation, mostly as snow; summers are hot with increased precipitation, chiefly associated with movement of Pacific and Arctic air masses into warm, humid air masses from the Gulf of Mexico. Vegetation is almost wholly prairie grassland, due to the semiarid, markedly seasonal climate.

The Northern Great Plains is a large region of generally low relief sloping eastward from the Rocky Mountains toward the Missouri and Mississippi Rivers. Its basic bedrock structure is a broad syncline, punctuated by the Black Hills and a few smaller uplifts, and by structural basins such as the Williston, Powder River, and Denver-Julesburg Basins (Fig. 3). Its "surface" bedrock is chiefly Cretaceous and Tertiary sediments, with small areas of older rocks in the Black Hills, central Montana, and eastern parts of Wyoming, Kansas, and Oklahoma.

During the Laramide orogeny (latest Cretaceous through Eocene), while the Rocky Mountains and Black Hills were rising, synorogenic sediments (frequently with large amounts of volcanic ash from volcanic centers in the Rocky Mountains) were deposited in the subsiding Denver-Julesburg, Powder River, and other basins. From Oligocene to Miocene time, sedimenta-

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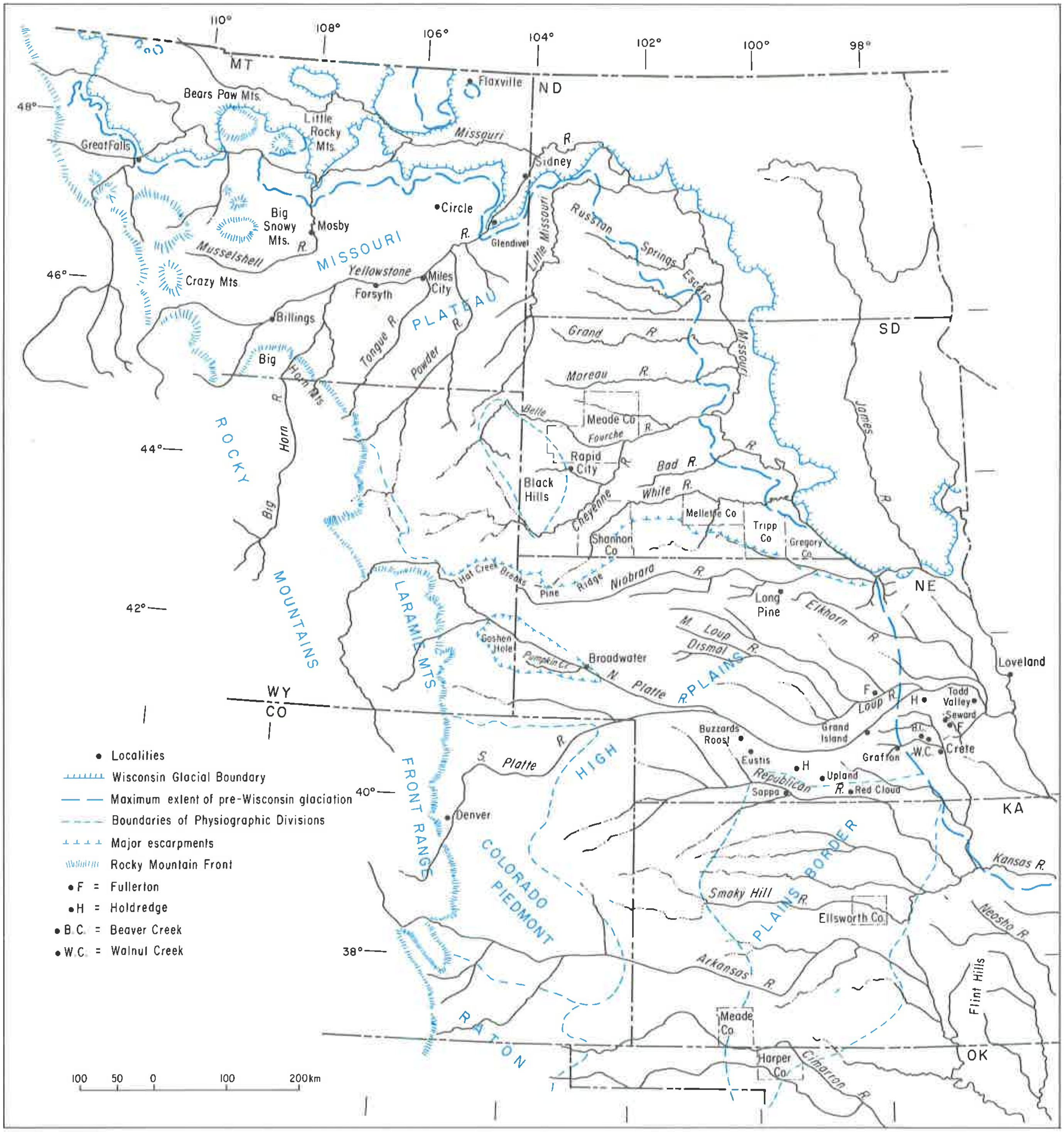

Figure 1. Map of the Northern Great Plains, showing locations of sites and regions mentioned in text.

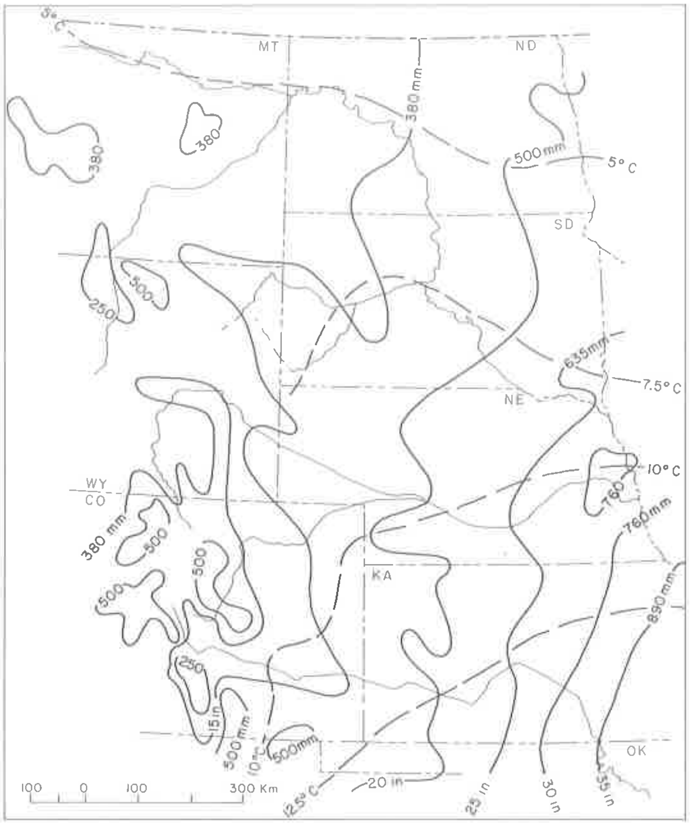

Figure 2. Mean annual isotherms and isohyets across the Northern Great Plains.

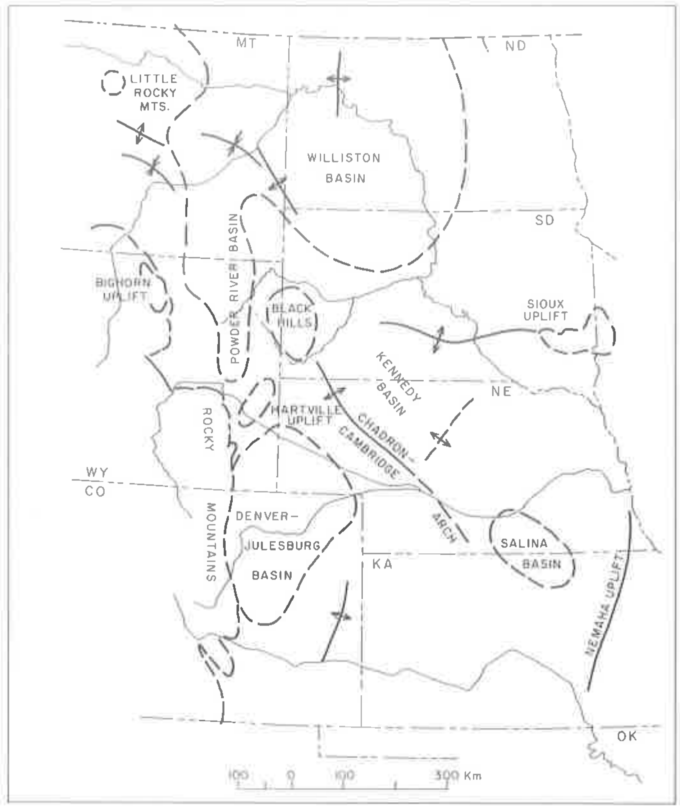

Figure 3. Major bedrock structures of the Northern Great Plains.

tion generally slowed with declining tectonism and volcanism in the Rocky Mountains. However, since the later Miocene, epeirogenic uplift, probably associated with the East Pacific Rise, affected the Great Plains and particularly the Rocky Mountains. During the last 10 m.y. the Rocky Mountain front has risen 1.5 to 2 km, and the eastern margin of the Great Plains 100 to 500 m (Gable and Hatton, 1983), with half to one-quarter of these amounts during the last 5 m.y. Thus, during the later Miocene the Great Plains became a huge aggrading piedmont sloping gently eastward from the Rocky Mountains and Black Hills, with generally eastward drainage, on which the Ogallala Formation and equivalents was deposited. The Ogallala underlies the High Plains Surface, the highest and oldest geomorphic surface preserved in this region. It has been completely eroded along some parts of the western margin of the region (e.g., the Colorado Piedmont), but eastward, it (and its equivalents, such as the Flaxville gravels in Montana) locally is preserved as caprock or buried by Quaternary sediments (Alden, 1924, 1932; Howard, 1960; Stanley, 1971, 1976; Pearl, 1971; Scott, 1982; Corner and Diffendal, 1983; Diffendal and Corner, 1984; Swinehart and others, 1985; Aber, 1985).

During the Pliocene, regional aggradation slowly changed to dissection by the principal rivers. In the western part of the region the rivers flowed eastward, but the continental drainage divide extended northeast from the Black Hills through central South Dakota, far south of its present position. The ancestral upper Missouri, Little Missouri, Yellowstone, and Cheyenne Rivers drained northeast to Hudson Bay, whereas the ancestral White, Platte, and Arkansas Rivers went to the Gulf of Mexico (Fig 4A). Their courses are marked by scattered surface and subsurface gravel remnants; in Montana and North Dakota, deposits of the preglacial Missouri River and its tributaries are buried deeply beneath glacial and other sediments (Howard, 1960; Bluemle, 1972).

NATURE OF THE QUATERNARY STRATIGRAPHIC RECORD

Various continental glaciations, starting about 2.5 Ma (Boellstorff, 1978; Richmond and Fullerton, 1986, p. 183–200 and Chart 1 summation), caused reorientation of the Missouri River system southeastward to the Mississippi River, resulting in many stream captures and other geomorphic changes (Fig. 4B). Each time the ice blocked eastward-flowing rivers, proglacial lakes formed, spilled across divides, and developed new courses around the glacial margin. The present course of the Missouri River through North and South Dakota is chiefly along a late Illinoian ice margin. The Platte River evolved through spasmodic uplift of the Chadron arch (Stanley and Wayne, 1972) and sev-

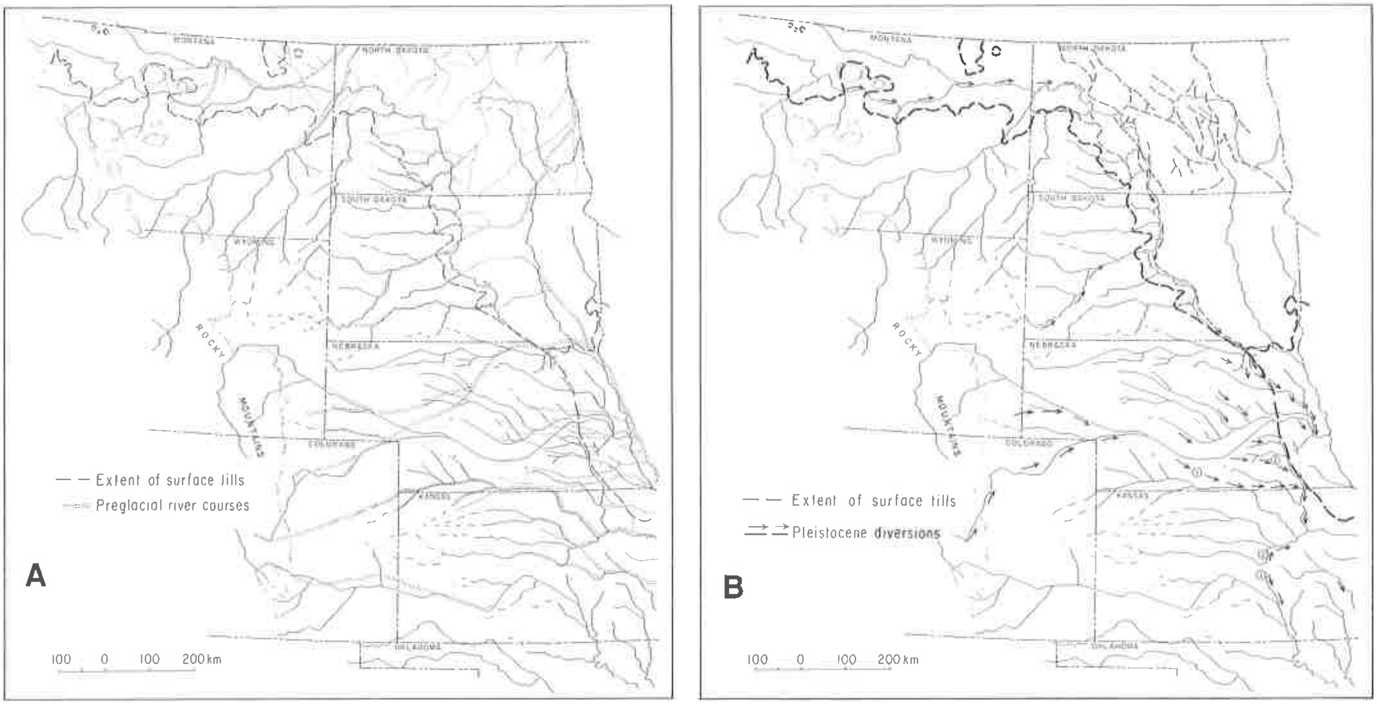

Figure 4. A. Principal preglacial drainage lines of the Northern Great Plains. B. Major Pleistocene drainage diversions in the Northern Great Plains.

eral early and middle Pleistocene glacial advances into eastern Nebraska/northeastern Kansas. In the middle Pleistocene, the Platte joined with the glacially diverted Missouri River and formed a wide alluvial plain across east-central Nebraska and northeast Kansas. The Cheyenne River captured many streams that formerly flowed to the White River, so now its system nearly encircles the Black Hills (Todd, 1902; Wanless, 1923). The Kansas and Belle Fourche Rivers extended headward, capturing many tributaries.

The Quaternary deposits of the unglaciated Northern Great Plains are chiefly fluvial sediments transported eastward from the Rocky Mountains, but in places include eolian sand and loess and lacustrine sediments deposited in ice-dammed and slackwater lakes. Previous regional discussions of these deposits are given in Lugn (1935), Condra and others (1947), Frye and Leonard (1952), Howard (1960), Reed and Dreeszen (1965), and Scott (1965). These workers used an interpretive model based on glacial-interglacial cycles that were then recognized in the Central Lowland (Kay, 1931; Leighton, 1931; Leighton and Willman, 1950; Frye and Leonard, 1952): the model had four glaciations (Nebraskan, Kansan, Illinoian, and Wisconsin) separated by three interglaciations (Aftonian, Yarmouth, and Sangaman). Their lithostratigraphic units typically were differentiated and correlated on the premise that coarse-grained fluvial sediments were coeval with glaciations, and fine-grained sediments with interglaciations (Lugn, 1935). One tephra layer, the "Pearlette volcanic ash," and a single paleosol, the "Sangamon," were used as key strata for correlation. Recent research has demonstrated that this model is far too simplistic: (1) The Laurentide ice sheet advanced into the Central Lowland seven or eight times between 2.2 Ma and the Wisconsin glaciation (Richmond and Fullerton, 1986, chart 1, summation). (2) The "Pearlette Ash" is a family of three similar tephra layers, dating 2.0, 1.2, and 0.62 Ma, respectively (Boellstorff, 1978; Izett, 1981). (3) Many more interglacial and interstadial paleosols are now recognized, and those formerly correlated with the Sangamon range in age from about 30 to 260 ka (Richmond and Fullerton, 1986, Chart 1, summation). One result of the new research is that the names Nebraskan, Afton(ian), Kansan, and Yarmouth(ian), formerly widely used, are now abandoned (Richmond and Fullerton, 1986, p. 14, 183–184).

CRITERIA FOR AGE ASSIGNMENTS AND CORRELATION

Age assignments and correlations of Quaternary deposits usually are made using a combination of criteria that may include position in the landscape, superposition and crosscutting stratigraphic relations, intensity of weathering and degree of soil formation, stratigraphic position with respect to identified volcanic ash beds, and to a limited extent, ^{14}C ages, uranium-series ages, and magnetostratigraphy. The importance and applicability of the criteria vary with the age of the deposit. The degree of weathering and soil development in early and middle Pleistocene deposits typically is distinct, and it readily distinguishes them from markedly less weathered late Pleistocene and Holocene deposits. However, differences in weathering and soil formation generally are not useful for distinguishing between early and middle Pleistocene deposits, except in areas where there has been a progressive accumulation of secondary $CaCO_3$ (Gile and others, 1966; Machette, 1985). Exept for $CaCO_3$ content, the soil properties that change systematically with age are approaching or have already attained a steady state in deposits as old as middle Pleistocene. In addition, comparisons of soils in deposits as old as early and middle Pleistocene may be complicated by erosion and incomplete preservation of the original surface and soil profile. Consequently, fluvial deposits of early and middle Pleistocene age are correlated almost exclusively by height above present streams, and ages assigned to the deposits are crudely constrained by a small number of widely spaced deposits of Pearlette-family tephra layers. In contrast to early and middle Pleistocene deposits, late Pleistocene and Holocene deposits of several different relative-age classes are distinguished by differences in weathering and degree of soil formation. Also, because deposits of late Pleistocene and Holocene age are largely restricted to valley floors, criteria of superposition and crosscutting stratigraphic relations are widely applicable. Most of these deposits also are within the range of ^{14}C dating.

THE UNGLACIATED MISSOURI PLATEAU AND BLACK HILLS

The Laurentide and Cordilleran ice sheets repeatedly covered the northern part of the Missouri Plateau, diverting the Missouri River southeastward, starting as early as about 2.1 Ma (Richmond and Fullerton, 1986, Chart 1, summation; Fig. 1). The glacial deposits have received much more study than the extraglacial ones.

The unglaciated Missouri Plateau (punctuated by the Crazy, Big and Little Snowy Mountains, and Black Hills), descends from 1,800 to 1,000 m altitude adjacent to the Rocky Mountains and outlying mountain ranges, to about 600 m at the Montana–North Dakota border near Glendive, Montana, and about 150 to 200 m at the eastern edge of the plateau. Topographic relief ranges from a few tens of meters per kilometer in interstream areas and on Cretaceous shale, to 60 to 150 m per kilometer in badlands along major rivers; scattered outlier buttes of Eocene to Oligocene rocks rise 100 to 240 m above surrounding plains and as much as 440 m above nearby rivers.

YELLOWSTONE AND MUSSELSHELL DRAINAGE BASINS, MONTANA

Robert N. Bergantino

PEDIMENT, TERRACE, AND ALLUVIAL-FAN DEPOSITS NEAR THE MOUNTAINS

Flaxville gravels (late Miocene)

High-level gravel cappings on disjunct but widespread mesas in this region were named the "Flaxville gravels" by Collier and Thom (1918), who first studied them comprehensively. They are chiefly poorly sorted boulder to pebble gravel and pebbly sand, commonly 5 to 10 m and rarely as much as 40 m thick, and bear thick pedogenic carbonate. Alden (1932) correlated the highest terrace-gravel deposits flanking the Beartooth Mountains with the Flaxville.

The Flaxville has yielded only relatively nondiagnostic mammalian fossils, indicative of Miocene (or Pliocene, as formerly defined) age (Collier and Thom, 1918). Fission-track ages on zircon from tephra layers in this unit indicate that the unit is at least 9 to 6.5 m.y. old (Whitaker, 1980, p. 87–88).

The *Wyota Gravel* of Jensen (1951), consists of pediment/terrace-gravel deposits 20 and more meters below the Flaxville and similar in lithology. It is the most widespread of the "preglacial" gravel remnants (Jensen and Varnes, 1964).

Quaternary deposits

Ritter (1967) found five distinct terraces in each valley in this region, with similar elevations above principal streams, suggesting a regional sequence of erosional/depositional events. Thus, although he was unable to trace and correlate terraces from one valley to another, he concluded (1) that terraces of equivalent relative heights above principal streams are coeval, (2) most of their gravel was deposited during episodes of mountain glaciation, and (3) all the terrace units are Pleistocene.

ALLUVIAL UNITS DISTANT FROM THE MOUNTAINS

Isolated mesa-like remnants of the Flaxville gravel (and perhaps the Wyota), capping earlier Tertiary sediments, persist far eastward, into eastern Montana. A clinker boulder sample from gravel capping a ridge 30 km southwest of Forsyth, Montana, and about 365 m above the Yellowstone River, provided a fission-track age of 4.0 ± 0.7 Ma (Heffern and others, 1983). Alden (1932) believed this gravel to be coeval with the Flaxville, but it might be equivalent to the Wyota.

Investigations of Pleistocene terraces in this district were made by Alden (1932), Pierce (1936), Parker and Andrews (1939), Bryson (1959), and Bryson and Bass (1973). Johnson and Smith (1964) mapped eight terraces along the lower Musselshell River, and attributed them to downcutting after glacial diversion of the Missouri River westward across the Musselshell drainage basin. The highest terrace surface is 190 m above the Musselshell; deposits beneath some terrace surfaces are as thick as 20 m. They found erratic boulders on the surfaces of the six higher terraces (but no till in the terrace deposits), and no erratics on the two lowest terraces. Bass (1932) obtained mammoth (*E. columbi*) teeth and rib bones from terrace sediments about 7 m above the Holocene flood plain of the Tongue River, indicating a late Pleistocene age. No additional fossils have been reported.

The Lava Creek B tephra layer has been identified southwest of Miles City and near Sidney, 80 and 130 m, respectively, above the Yellowstone River, in uppermost deposits of the most extensive and best-preserved terrace along this river.

Just south of the Montana-Wyoming border, a stream flowing eastward captured the upper part of the Little Missouri River, beheading it sometime in the Pleistocene (Darton and O'Harra, 1905). Knechtel and Patterson (1962) noted six terrace levels along the upstream portion in Wyoming, but only the three highest levels in Montana. They named the strath left by the diversion the *Stoneville Surface.*

SEDIMENTS OF GLACIAL LAKE GREAT FALLS

Glacial lakes existed on the Montana plains during each glaciation (Colton and others, 1961), but few depositional sequences have been described. The deposits of Glacial Lake Great Falls, in the Missouri River valley, are more than 90 m thick upstream from the city of Great Falls (Lemke, 1977). They consist of a thin upper unit, probably of late Wisconsin age, and a lower, considerably thicker sequence that probably accumulated during an earlier ice advance into the area. The upper unit is chiefly fine sand and silt, and the lower unit is principally silt with minor amounts of clay and fine sand.

POWDER RIVER BASIN

Brainerd Mears, Jr., Sherry S. Agard, and Wayne M. Sutherland with a contribution on clinker by Donald A. Coates

The Powder River basin is a structural and topographic basin. Its southern boundary is Hat Creek Breaks; its western, the east edge of the Mesaverde Formation along the Casper arch; its northern, the front of the Bighorn Mountains; and its eastern boundary is chiefly the Dakota hogback along the Black Hills. North of the Black Hills the Powder River basin merges with the unglaciated Missouri Plateau in Montana.

The basin is floored by the Eocene Wasatch Formation, eroded into colorful badlands in places, and encircled by broad expanses of dissected gray sandstone and shale of the Paleocene Fort Union Formation and the Cretaceous Lance Formation and Pierre Shale (near the Black Hills). Oligocene and Miocene deposits are absent, except for isolated remnants of channel gravel of the White River Formation, capping Wasatch Formation in the Pumpkin Buttes, 275 m above adjacent terrain (Fig. 5).

By late Miocene or early Pliocene time, filling of the Powder River basin had ceased, and east- and northeastward-flowing streams, rejuvenated by regional uplift and climatic changes, began to excavate the basin, with many stream captures and alterations of stream courses (Fig. 6), continuing to the present. The eastward courses of the Belle Fourche and Cheyenne Rivers may be relicts of the mid-Tertiary drainage, as suggested by incised meanders where these superposed streams cross the Black Hills (Figs. 1 and 5). In the western third of the basin, however, the Powder River has eroded south and east, beheading the upper reaches of these rivers.

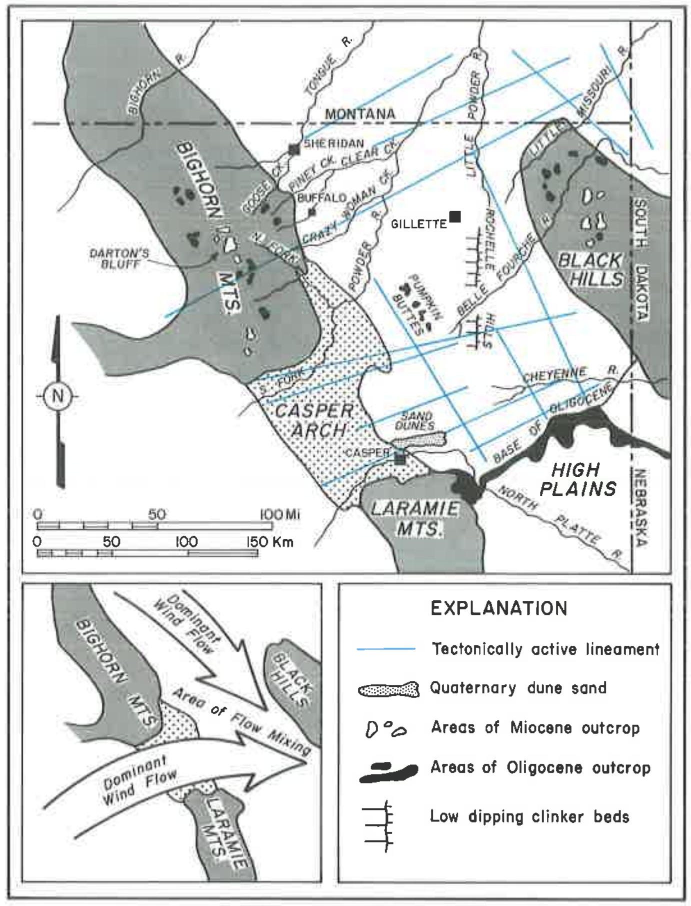

Figure 5. The Powder River Basin, showing features mentioned in the text and pattern of tectonically active lineaments.

EFFECTS OF STREAM PIRACY

Streams originating in basin and piedmont areas underlain by relatively soft, fine-grained sediments have lower gradients adjusted to transport such sediment, whereas streams originating in the mountains maintain steeper gradients to transport coarser gravel (Ritter, 1987). Occasionally, minor piedmont streams, eroding headward, breach drainage divides and capture streams carrying coarser loads. These events result in localized cutting or filling as the affected streams adjust to changed discharge, load, and base level. They are random in space and time. Therefore, they can confound correlations that assume terrace-forming events were synchronous across different drainage systems.

QUATERNARY FLUVIAL STRATIGRAPHY

The Quaternary fluvial history of the Powder River Basin has been discussed in (1) regional overviews (Alden, 1932; Sharp, 1949; Scott, 1965); (2) general geologic studies of specific areas

(Hose, 1955; Kohout, 1957; Mapel, 1959); (3) surficial-geologic studies, primarily in the central and western parts of the basin (Leopold and Miller, 1954; Albanese and Wilson, 1974; Ebaugh, 1976; Hinrichs, 1984; Reheis, 1987; Reheis and Coates, 1987); and (4) archeological studies (Albanese, 1974, 1982; Haynes and Grey, 1965; Reider, 1980, 1982, 1983). This work has not been integrated into a cohesive summary. The few detailed stratigraphic studies are of local areas, and most of the deposits have not been dated. Pleistocene tephra, common in other parts of Wyoming and Montana, are either lacking or not yet identified. Recent archeological work has yielded some radiocarbon dates and soil stratigraphic data for Holocene deposits (Albanese, 1974, 1982; Reider, 1980, 1982, 1983), and a few uranium-series ages have been obtained from carbonates from buried soils in late Pleistocene alluvium (Hinrichs, 1984). A coherent picture of basin history will require many more data.

The Quaternary fluvial history of the basin is recorded in terrace deposits, which represent relatively brief periods of stability or aggradation in an otherwise continuous erosion of the basin. This record, however, is not equally well preserved everywhere. The Belle Fourche and Cheyenne River drainage basins to the east and the Powder River drainage basin to the west represent contrasting styles of erosion and deposition, reflecting different geomorphic settings. The Belle Fourche and Cheyenne Rivers (Fig. 5) head in soft sediments in the center of the basin and lack coarse, resistant alluvium; therefore, few terrace remnants older than late Pleistocene and Holocene are preserved. Moreover, in this part of the basin, eolian deposits (mostly derived from the Wasatch Formation) have introduced a significant amount of fine sediment into the fluvial system. The best-preserved terrace sequences, as well as extensive pediments, are along the flanks of the basin where tributary streams heading in the mountains deposited gravel with resistant lithologies. The terrace sequences are particularly well developed along the western margins of the basin where tributaries of the Powder River were influenced by glaciation in the Bighorn Mountains.

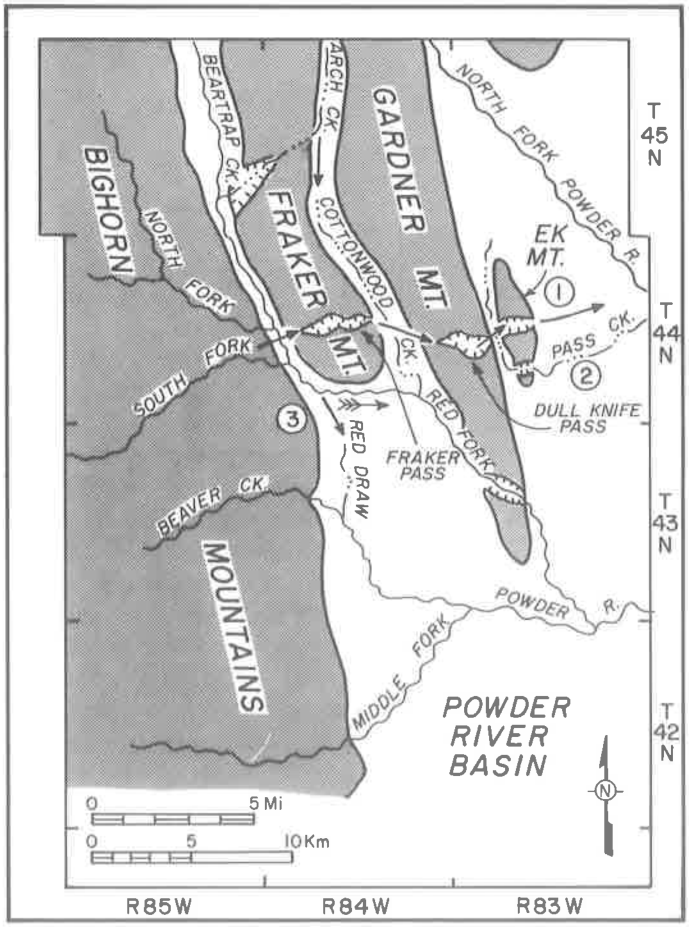

Figure 6. Map showing former courses (arrows) of Red Fork Powder River through abandoned passes (hachures) in the southeastern Bighorn Mountains and adjacent Powder River Basin. Numbers indicate probable sequence of stream courses.

The stratigraphic relations (relative positions) of terrace sequences in the Powder River Basin (Table 1) are based on approximate heights above modern streams. However, relative topographic position, though useful, is not a reliable criterion for correlation of terraces of different streams (Reheis and others, this volume). Streams may leave the mountains at different altitudes (Hose, 1955), transport different materials, and have different gradients.

Recent studies in the center of the basin (Reheis, 1987; Reheis and Coates, 1987) show that Holocene terrace sequences are more complex than (and do not correlate well with) the three Holocene terraces of Leopold and Miller (1954). At different sites, there are from three to five Holocene terraces. Such complexity probably is due to threshold events being influenced by local factors in addition to regional climatic/hydrologic changes.

STREAM AND EOLIAN PATTERNS

Streams, sand dunes, and eolian deflation basins in the Powder River Basin show strong northwest-southeast alignment. The stream alignments probably are influenced by a strong northwest-southeast regional master joint-and-fault pattern in the bedrock. The eolian features, however, are controlled by the predominantly northwesterly winds characteristic of this region; thus, both types of features have similar, coincident alignments that are due to different causes.

Late Pleistocene wind patterns in Wyoming are consistent with wind-flow patterns interpreted from Holocene features (Frison, 1974; Kolm, 1977; Marrs and Kolm, 1982; Gaylord, 1982). However, the predominantly stabilized dunes seen today indicate much stronger eolian activity in former times. In the Ferris dune field in central Wyoming, episodes of pronounced eolian activity prior to 4,560 B.P. are interpreted to represent a response to warm dry intervals, such as the Altithermal (Gaylord, 1982). Thus, the intensity of eolian activity changed during the late Quaternary, but wind-flow patterns remained relatively constant.

CLINKER

Natural burning of subbituminous coal beds in the Paleocene Fort Union Formation and Eocene Wasatch Formation caps many prominent landforms in the Powder River Basin in Wyoming and Montana. In most areas the coal-bearing formations are chiefly friable, easily eroded, clay-rich sandstone and siltstone. Beds of high-volatile coal commonly are ignited by spontaneous combustion or by range fires wherever they are above the water table and exposed to air.

Burning of the coal hardens, sinters, and in some places melts the overlying rocks, coloring them dominantly red but also black, gray, green, purple, orange, and yellow. These thermally metamorphosed rocks, collectively called clinker, differentially resist erosion, not only because they are harder than the unbaked rock, but also because heating and subsidence during burning produces high fracture permeability, allowing water to infiltrate and minimizing surface runoff. Thus, erosion leaves clinker caps on characteristic steep-sided, red-topped knobs, ridges, buttes, and mesas standing above the less-resistant unmetamorphosed bedrock. Clinker-capped topographic features are eroded chiefly by parallel slope retreat, dominated on steeper slopes by landsliding and other mass wasting and on gentler slopes by slopewash.

Clinker is useful in determining rates of landscape development. On the east flank of the Powder River Basin, the Wyodak coal zone in the Fort Union Formation contains 10 to 35 m of coal. Where it has burned north of Little Thunder Creek (Coates and Naeser, 1984), as much as 40 m of overlying rock has been metamorphosed to clinker. The resulting discontinuous, westward-dipping blanket of clinker forms an east-facing escarpment, the Rochelle Hills, the highest topography in the area (Fig. 5). During the last 700 k.y., the burning of the Wyodak coal by repeated fires has progressed irregularly downdip along a regional burn front as stream incision has lowered the water table. Because detrital zircons in sandstones overlying a burning coal bed are annealed by the heat of burning, fission-track (F-T) dating can determine the time of cooling. Detrital zircons in the clinker exposed along the divide north of Little Thunder Creek yield F-T ages showing about 8 km of westward downdip migration of the regional burn front during the last 700 k.y. (Coates and Naeser, 1984). This suggests about 100 m of stream downcutting in this area, or a rate of about 0.14 m/k.y. This rate is consistent with downcutting rates of the 0.12 to 0.19 m/k.y. suggested for the northern Bighorn Basin (Reheis and others, this volume). Similarly, near Ashland, Montana, in the northern Powder River Basin, F-T ages of clinker formed by burning of the Anderson and Knowlock coals in the Fort Union Formation show that the Tongue River has downcut about 300 m in the last million years.

Complex structures, produced in clinker during subsidence as underlying coal is burned, are systematically oriented in some places (Verbeek and Coates, 1982). Open fractures, produced as the rock subsides by rotational slumping, serve as conduits for combustible gases released from the heated coal. As the combustible gas rises and mixes with air, the resulting fire is hot enough to melt part of the adjacent rock. The resulting paralava welds surrounding fragments and forms resistant "chimneys" that commonly stand above the general ground level after erosion has lowered the surrounding surface.

Mineralogy of paralava is unique, a result of temperatures as high as 1,700°C combined with low pressures. Cosca and Peacor (1987), Cosca and others (1988), and Cosca and others (1989) describe two new minerals and show that paralava from the Fort Union Formation in the Powder River Basin is depleted in Si, Al, and K with relation to the surrounding rock but is enriched in Fe, Mg, and Ca.

Early explorers reported abundant coal fires in the region (Moulton, 1987, p. 45), but in this century most have been extinguished and new fires have not been permitted to become established. The large areas of clinker represent enormous quantities of coal burned during Quaternary time. The burning has been an intermittent process, leaving a datable physical record in the form of annealed zircon grains and remagnetized rocks (Jones and others, 1984), and the hardened clinker forms a resistant landscape element; thus studies of clinker will continue to shed light on geomorphic evolution of the Powder River Basin.

THE UNGLACIATED MISSOURI PLATEAU IN THE DAKOTAS

John P. Bluemle and James E. Martin

At the time of the first late Cenozoic glaciation (Dunn glaciation; Clayton, 1969) about 2.1 Ma, the Missouri Plateau was an undulating plain sloping from about 975 m altitude bordering the Black Hills to about 180 m in the northeast. This surface, called the Missouri Slope (Moran and others, 1976), now is deeply buried by Pleistocene glacial, glaciofluvial, and glaciolacustrine sediments in the northern and northeastern parts, but in the southwest, remnants of its late Pliocene or early Pleistocene sediments are exposed at altitudes between about 700 and 900 m.

The thickest deposits of Quaternary alluvium in the unglaciated part of the Missouri Plateau are valley fills in trenches carved by glacial meltwater during early glaciations. However, alluvium, which is neither glacial nor glaciofluvial in origin, also occurs in some narrow valleys and may exceed 75 m in thickness. The sediments consist mainly of poorly sorted, obscurely bedded silt and clay with lenses of sand. In near-surface exposures of these materials, weak paleosols, scattered mammal bones and teeth, terrestrial snail shells, and wood fragments are common. Their stratigraphy is still poorly understood, and only a broad outline of the Quaternary history can be presented at this time. Both erosional surfaces and river terraces generally are underlain by sediments related to the fluvial activity that produced them, but until more detailed stratigraphic study is undertaken, morphostratigraphy provides the principal basis for our present knowledge.

Some time after the earliest identified glaciation, much of the Missouri slope was lowered 30 to 60 m to the widespread

Green River stability surface (Moran and others, 1976; Clayton, 1969). This erosion surface nearly coincides with the modern surface over much of the area between the Little Missouri Badlands and the Russian Springs Escarpment (Fig. 1). Following formation of numerous eolian deflation depressions, renewed fluvial erosion lowered the area northeast of the Green River stability surface by several tens of meters and ended with formation of the Knife River stability surface (Moran and others, 1976), which is separated from the Green River surface by a northeast-facing erosional scarp, the Russian Springs Escarpment. The *Cartwright Alluvium* of Howard (1960), probably was deposited at this time. It occurs as terrace gravel remnants about 75 m above the level of the Yellowstone and Missouri Rivers near the Montana border; it is equivalent to Alden's (1932) number 2 bench.

A period of pedimentation followed this stability episode. Pediment remnants as much as 1 km^2 in area are common on the Missouri slope. They are capped by 2 to 3 m of fluvial gravel that accumulated when hillslope activities provided a large supply of gravel, possibly during periods of periglacial conditions.

Four terrace surfaces are present along streams that emerge from the Black Hills. Also, gravel remnants with clasts from the Black Hills are present in many places in western South Dakota. The informally designated *Medicine Root* gravel in Shannon County, South Dakota, considered to be "Nebraskan" in age by Harksen (1966), is believed to be a deposit of a large, high-energy, east-trending river that was pirated by the Cheyenne River. Harksen (1969) termed this ancient river the ancestral Medicine Root River and traced its course by gravel remnants from the southern edge of Cuny Table to northwestern Mellette County, a distance of nearly 130 km. Remains of fossil vertebrates were assigned a medial or late Blancan Land-Mammal Age (Martin, 1973; Pinsof, 1985).

In Gregory County, southernmost South Dakota, just west of the Missouri River, Stevenson (1958) named a coarse gravel unit the *Herrick Formation* (Fig. 1). Mammal fossils from the Burke Gravel pit (Stevenson's paratype section) indicate a probable Blancan age (Pinsof, 1985); however, a younger Irvingtonian taxon, *Rangifer tarandus* (Green and Lillegraven, 1965) also is present. Both faunas indicate much cooler climate than the present.

Also, just north of the Nebraska border are three other fossil localities to which Pinsof (1985) assigned a Blancan age. All have been correlated to the western-provenance alluvium of Flint (1955, p. 45), which appears to have been partly derived from the Black Hills. The similar composition, age, and east-west alignment of the gravel deposits suggest that they all may be correlative with the Herrick Formation. Moreover, these sites generally are in line with the trend of the Medicine Root River, which appears to have flowed easterly during Blancan time.

To the north in Meade County, Harksen (1969) correlated another gravel remnant west of Enning to the "Nebraskan," based only on stratigraphic position. From this one exposure he postulated a second major paleodrainage subparallel to the Cheyenne and Missouri Rivers, flowing easterly. Just south of the confluence of the Cheyenne and Missouri Rivers, Crandall (1958) described western-derived gravels at Standing Butte. A horse tooth was identified by C. W. Hibbard (in Crandall, 1958) as a "Nebraskan" or possibly early "Yarmouthian" fossil; Pinsof (1985, p. 257) tentatively assigned the fragmentary specimen to *Equus* (perhaps *Dolichohippus*).

LATE PLEISTOCENE AND HOLOCENE

Most of the younger terraces (late Pleistocene–Holocene; Fillman, 1929; Plumley, 1948) have been traced along creek drainages until they intersect the Cheyenne River. A mammoth-kill site along the White River north of Ogallala (Lange-Ferguson locality) associated with the Clovis Cultural Tradition appears to be within the oldest late Pleistocene unit in this area. A humic zone above the fossil bed was dated at 10,670 ± 300 B.P. (I-11, 710; Hannus, 1982). White and Hannus (1985) postulated two periods of Holocene alluviation in the Badlands, based on ten radiocarbon ages (980 to 2,520 B.P.) from hearths along Fog Creek.

Two other units have been named, the *Red Dog Loess* (Harksen, 1968) and the *Oahe Formation* (Clayton and others, 1976). Harksen (1969) stated that the loess occurs on Red Dog Table and on Babby Butte, 8 km to the southeast. The unit may correlate with one of the eolian units in northern Nebraska. The Oahe Formation, an upland silt in North Dakota, is late Wisconsin and Holocene in age. The Sand Hills Formation, an extensive late Quaternary eolian unit in Nebraska, has been extended into southern South Dakota (Collins, 1959; Sevon, 1960).

BLACK HILLS

James E. Martin

The Black Hills lie on the South Dakota–Wyoming border and extend northward barely into Montana (Figs. 1, 5). The ovate uplift dates from the Laramide orogeny (Noble, 1952; Lisenbee, 1978). The southern part of its core is Precambrian Harney Peak Granite, intruded into metasedimentary and metaigneous rocks; a northern block consists of Tertiary trachyte-phonolite intrusions. Dipping away from the core is an almost complete succession of Paleozoic and Mesozoic strata (Gries and Martin, 1986) on which differential weathering and erosion have produced a series of ridges and valleys encircling the Black Hills. The surrounding plains are underlain by Upper Cretaceous black shale units and by nearly horizontal fine-grained Tertiary rocks. Clasts from the Black Hills are widely distributed in time and space in Cenozoic deposits of the Missouri Plateau to the east.

The Cheyenne River drains the southern part of the Black Hills and its tributary, the Belle Fourche River, drains the northern part. Both rivers cut across northerly trending structures and appear to be antecedent. Many other east-west tributaries also are discordant to major structures and are perpendicular to the strike of Paleozoic and Mesozoic strata that surround the uplift, producing a subradial drainage pattern.

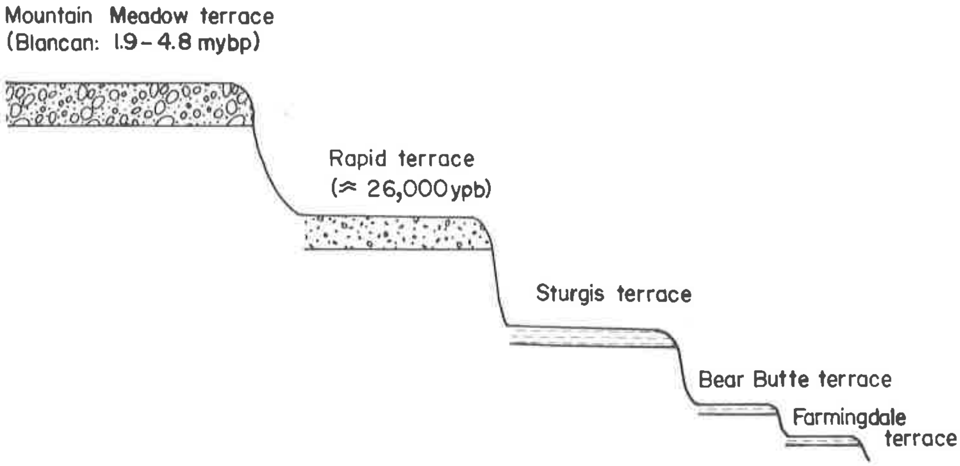

Figure 7. Diagram of the erosional surfaces in and near the Black Hills.

QUATERNARY STRATIGRAPHY

Morphostratigraphic methods have dominated Quaternary stratigraphic studies of the Black Hills. Most of the preserved Pleistocene deposits occur along creeks on the eastern margin of the Black Hills. Named surfaces and terraces include the underlying sediments.

Mountain meadow surface

The first comprehensive study of terrace deposits in the Black Hills was by Fillman (1929), who named three units (Fig. 7). The oldest and topographically highest unit is the *Mountain Meadow surface.* Precambrian siliceous clasts dominate the gravel beneath this surface. Fillman (1929, p. 29) correlated the gravel with the White River Formation, of medial Oligocene (Orellan) age. Plumley (1948) questioned this correlation, and Meyerhoff and Olmstead (1937) noted geomorphic features indicating Miocene or Pliocene age. A camel ankle bone from the Mountain Meadow gravel just south of Rapid City was assigned to *Megatylopus* by Green and Gries (1963), but was reidentified as *Gigantocamelus,* a common Blancan camelid (Harrison, 1985). The Mountain Meadow terrace may be related to the hypothesized rejuvenated uplift of the Black Hills starting about 4.5 Ma (Harksen, 1969).

Late Pleistocene surfaces and deposits

The *Rapid terrace* (Fillman, 1929) is topographically lower than the Mountain Meadow surface, and its sediments contain more locally derived clasts. It is 55 m above Rapid Creek, near Rapid City, and has been traced along the eastern flank of the Black Hills uplift (Plumley, 1948). Kempton (1980) correlated the Rapid terrace with her T4 terrace, which is associated with the Hot Springs mammoth site, a natural elephant trap formed by a karst depression containing a warm spring (Laury, 1980). A radiocarbon date from mammoth bone apatite of 26,075 ± 880 B.P. (GX-5895-A; Laury, 1980; Agenbroad, 1985) provides a minimum age for the site.

The lowest terrace surface is the *Sturgis terrace* (Fillman, 1929), 21 m above Bear Butte Creek near the town of Sturgis (Plumley, 1948; Fig. 7). Kempton (1980) correlated the thick conglomerate in the Hot Springs area, her terrace T3, with deposits that veneer this terrace. Along Fall River, the gravel in this terrace is well cemented, at least 30 m thick, and dominated by Minnekahta Limestone (Permian) clasts. The ages of this terrace and the sediments beneath it have not been established.

Along Bear Butte Creek, Plumley (1948) noted two terraces below the Sturgis terrace. The higher is the *Bear Butte terrace,* a major terrace along the Belle Fourche River and along Bear Butte Creek as far upstream as Bear Butte. The lower *Farmingdale terrace* extends from the town of Farmingdale along Rapid Creek, east to the Cheyenne River. Plumley correlated this terrace with a lower unnamed terrace along Bear Butte Creek. He also identified another unnamed terrace between the Rapid and Sturgis terraces southeast of Rapid City. The Rapid and younger terraces have been attributed to damming by glacial ice along the Missouri River (Fillman, 1929, p. 44; Plumley, 1948, p. 537; Kempton, 1980). However, the presence of hydrothermal springs in the southern Black Hills suggests that local tectonism may have been a contributing factor.

THE HIGH PLAINS IN WYOMING

Brainerd Mears, Jr.

The High Plains surface in Wyoming slopes gradually eastward from an altitude of 2,725 m against the Laramie Mountains to about 1,580 m at the state line near Pine Bluffs, beyond which the surface continues across western Nebraska. The lowest point in the Wyoming High Plains, 1,250 m, is at the Nebraska state line along the North Platte River in Goshen Hole, a lowland eroded into the plains surface (Fig. 8). Along the Colorado state line, the High Plains are bounded by the Chalk Bluffs, a 300- to 400-m-high scarp that descends in steps to the floor of the Colorado Piedmont. On the north, the High Plains surface ends at the Hat Creek Breaks (Pine Ridge Escarpment of Fenneman, 1931), an erosional scarp with a rim that stands 300 m or higher above the floor of the adjacent Powder River Basin.

STRATIGRAPHIC FRAMEWORK

The geomorphic features of the High Plains are products of events that began in Tertiary time. Paleocene and Eocene deposits—although widespread to the north in the Powder River Basin and to the south in the center of the Colorado Piedmont—have not been observed between the Hat Creek breaks and the Chalk Bluffs. The absence of early Tertiary deposits probably reflects nondeposition on a topographically positive region or uplift and erosional removal of thin deposits (Ahlbrandt and Groen, 1987).

Oligocene and early Miocene strata—predominantly bentonitic mudstones ("ash clays") in the White River Group and fine to very fine sand in the Arikaree Group—record a major episode of regional aggradation. Shallow stream channels, whose clasts were largely derived from the adjacent Laramide uplifts, were repeatedly formed and subsequently engulfed by airborne pyroclastics that drifted into the region from distant volcanic eruptions in Colorado or the Basin and Range (Stanley, 1976).

The late Miocene Ogallala Formation reflects a major change in depositional environment. Massive infalls of airborne pyroclastics had ceased, and the older pyroclastic materials were reworked and deposited as local lenses in alluvium. The dominantly fluvial Ogallala deposits, which overlie a pronounced erosion surface, are mostly relatively coarse clastics in deep channel fills of shifting stream systems. Some of the clasts were derived from the distinctive pink Sherman Granite, dark anorthosites, and other Precambrian rocks in the Laramie Mountains; however, a notable component is Precambrian rock types from the Colorado Front Range. The most distinctive clasts are rhyolitic cobbles, pebbles, and granules from late Oligocene volcanic fields in Colorado. The rhyolitic clasts are compelling evidence for a regional surface of subdued relief in the late Miocene. At that time, streams transported gravel northeastward from distant source areas at higher elevations in Colorado, across planated Precambrian mountain cores and sediment-filled basins, to the Wyoming High Plains.

The ongoing episode of Pliocene and Quaternary regional erosion, beginning about 5 m.y. ago (Swinehart and others, 1985), has etched out the existing landscape of the Wyoming High Plains. The southern part, the Cheyenne Tableland, is a moderately dissected, rolling surface generally underlain by the Ogallala Formation. North of the latitude of the east-flowing segment of Horse Creek (Fig. 8), the Ogallala is preserved adjacent to the mountain front but has been extensively stripped from the central and eastern parts of the Wyoming plains (Figs. 9A, B).

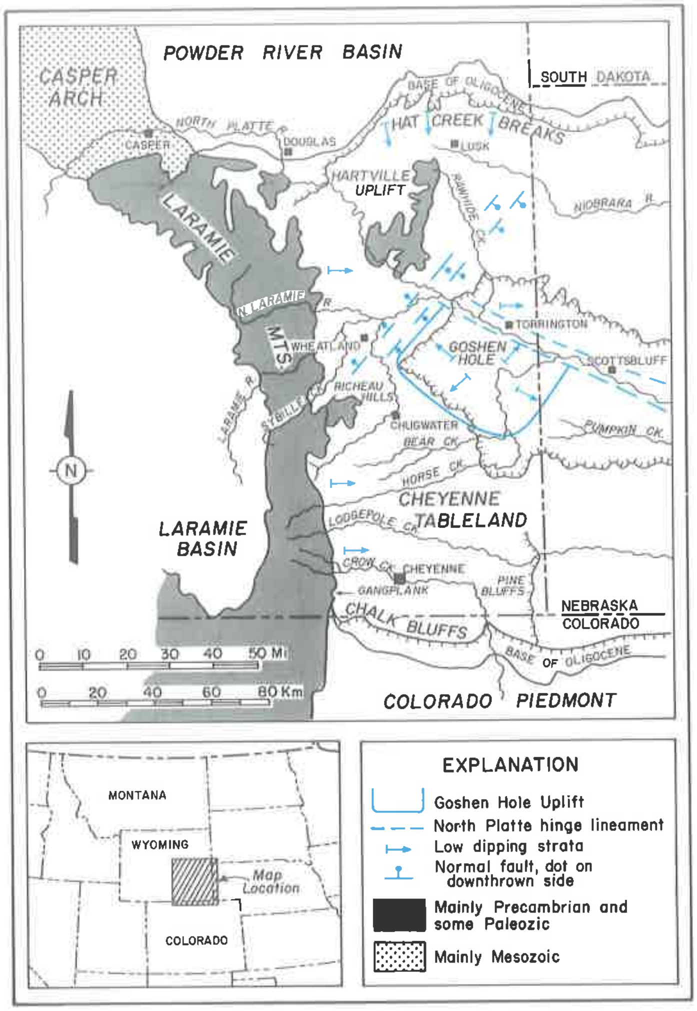

Figure 8. The High Plains in Wyoming and adjacent areas, showing features mentioned, including major structures and dominant lineaments.

Here broad uplands are capped by sandstone of the Arikaree Group, which, where breached by erosion, forms escarpments and mesas. They become increasingly common in the more dissected areas bordering larger streams, such as the Laramie and North Platte Rivers, and the bordering zones along the major lowland of Goshen Hole. Pliocene and Quaternary alluvial deposits have been largely restricted to terraces and flood plains along incising stream valleys. Quaternary eolian activity is recorded by shallow deflation hollows and in areas of stabilized dune sand, which in easternmost Wyoming are outliers of the Sand Hills region in Nebraska.

TECTONIC AND RELATED FEATURES

The Gangplank (Moore, 1959), a small relict of the late Miocene landscape that grades westward from Ogallala deposits onto the granite upland of the Laramie Mountains (Figs. 8, 9A), was preserved by its fortuitous position between the North and

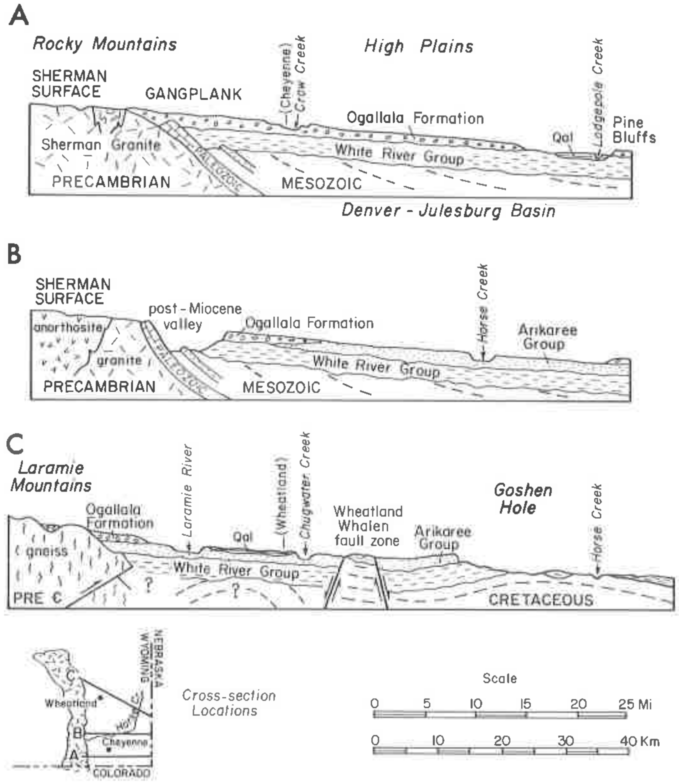

Figure 9. Representative cross sectional diagrams through the Wyoming High Plains.

South Platte Rivers. The Richeau Hills and Hartville uplift (Fig. 8) are Laramide-faulted anticlines (with cores of Mesozoic, Paleozoic, and some Precambrian rocks) that have been exhumed from a cover of gently dipping Tertiary strata. The Wheatland-Whalen zone of high-angle normal faults (Figs. 8, 9C), with displacements of as much as 200 m (McGrew, 1963), trends northeastward from the Richeau Hills across the plains adjacent to the Hartville uplift. Goshen Hole (Figs. 8, 9C), traversed by the North Platte River, is the major lowland cut into the High Plains. Structurally a broad uplift, it is rimmed by Arikaree Sandstone scarps and floored by the White River Formation and, in places, Late Cretaceous rocks.

STREAM DIVERSIONS AND PIRACY

Stream courses on the Wyoming High Plains have been strongly affected by late Cenozoic epeirogenic uplift and the associated regional degradation. The influence of warping is evident from the northward deflection of Chugwater Creek (Fig. 8) around the west end of the Goshen Hole uplift (Ahlbrandt and Groen, 1987). Stream piracy has been frequent; many creeks that initially flowed eastward from the Laramie Mountains have been captured by headward erosion of tributaries to the North and South Platte River systems in the Colorado Piedmont and Goshen Hole lowlands. Except for the Niobrara River, right-angle elbows of capture are common along stream courses.

FLUVIAL HISTORY OF GOSHEN HOLE

Remnants of five higher surfaces, ranging from 275 to 60 m higher than the present North Platte River in Goshen Hole, previously have been interpreted to be pediments graded to former valley floors (Rapp and others, 1957). These relicts were preserved because of armoring of the underlying erodible bedrock by terrace-gravel caps. The surfaces are prevalent in the northern part of Goshen Hole but generally absent to the south, possibly reflecting deposition of the gravels along former courses of the North Platte River. Studies of paleovalleys in adjacent Nebraska (Swinehart and others, 1985) indicate that the North Platte River has flowed southeastward during and since creation of the late Miocene Ogallala surface. However, successive Pliocene and Pleistocene channel courses were displaced southward during the excavation of the Goshen Hole lowland. This history merits further investigation where the lowland extends into Wyoming.

Within Goshen Hole, the North Platte has two prominent alluvial terraces whose deposits and heights above the river suggest Wisconsin and pre-Wisconsin ages (Rapp and others, 1957), but the terraces have not been precisely dated. Pleistocene alluvium extends as much as 60 m below the present river, a depth exceeding that along other major rivers in central and eastern Wyoming. For example, in both the Laramie and Powder River Basins, the channels of the Laramie and Powder Rivers are just above bedrock. Such contrasting fluvial histories pose an unresolved problem possibly involving tectonic warping.

EFFECTS OF LATE CENOZOIC CLIMATES

Regional uplift is the currently accepted mechanism for the post-Miocene dissection of the High Plains; the impact of late Cenozoic glacial and interglacial episodes has been debated. In Nebraska, Swinehart and others (1985) stress the overriding dominance of structural movements; on the other hand, Stanley and Wayne (1972) infer appreciable roles for both climatic change and epeirogenic and tectonic deformation. Early Pleistocene stream gravels are coarser than late Tertiary gravels and are enriched in the less resistant rock types. Stanley and Wayne attribute the change to increased stream discharges, reflecting lowered evaporation rates, and possibly increased precipitation. They determined that many of the eastward-flowing streams had headwaters in the Laramie Mountains, which were not glaciated but were subject to periglacial conditions that produced abundant fresh clasts. Late Pleistocene periglacial conditions in Wyoming now are confirmed by abundant localities of relict periglacial ice and sand wedges, including the Cheyenne area (Mears, 1987). Thus, the sequence of gravel caps protecting relict surfaces and the pre-Holocene alluvial terrace deposits may reflect times when the general post-Miocene valley deepening was interrupted by influxes of cobbles and gravel during glacial-periglacial climatic episodes.

THE CHEYENNE TABLELAND

M. E. Cooley

The Cheyenne Tableland (Fig. 8) is developed on late Miocene Ogallala deposits whose erosional history during the Pliocene and Quaternary is recorded by benches, terraces, and abandoned valleys. In the Gangplank and Cheyenne areas (Fig. 9A, B), the oldest post-Ogallala deposits, informally called the "reddish brown gravels," were deposited in shallow valleys that now are as much as 55 m above the present streams. This gravel unit, as much as 20 m thick, is widely distributed, from just south of Horse Creek to the brink of the Chalk Bluffs and from the Gangplank-Cheyenne area as far east as the surface of the Pine Bluffs. Like the underlying conglomerates in the Ogallala Formation, its rounded and subrounded clasts were derived from Precambrian crystalline rocks and Oligocene volcanics in the Front Range of Colorado and transported northeastward into Wyoming. A southwestern source for these fluvial deposits is confirmed by cross bedding and imbrications. Although not yet dated by fossil or radiometric means, the "reddish brown gravels" are considered to be Pliocene in age because of their topographic situations and the degree of development of paleosols they bear.

In Quaternary time the northeastward-flowing stream system that had existed throughout most of the Tertiary in the Cheyenne Tableland was disrupted as the South Platte River excavated the Colorado Piedmont and tributary canyons in the Colorado Front Range and the existing system of east-flowing streams was established. Stages in the dissection of the tableland during the Quaternary are recorded by bench and terrace gravel units that were largely derived from Sherman Granite and associated bedrock in the Laramie Mountains, as well as some redeposited clasts from Ogallala and Pliocene(?) conglomerates. Gravel-capped surfaces about 55 m, 45 m, 35 to 30 m, 9 m, and 6 m higher than present flood plains are designated as T5 (oldest) to T1. They are considered to be Pleistocene because of topographic position and paleosols (Fig. 10, 11; Table 1).

The wide variety of soil types (Table 2), in addition to the presence of periglacial ice-wedge casts, suggests broad variation of temperature and moisture on the High Plains during the late Tertiary and Quaternary, but grassland seems to have predominated. Many ancient soils are buried and/or truncated and commonly have younger soils developed on them. The lowest terrace deposits, 1.5 to 4 m high, near Pine Bluffs, are correlated with T4 and T2 deposits in the Cheyenne area.

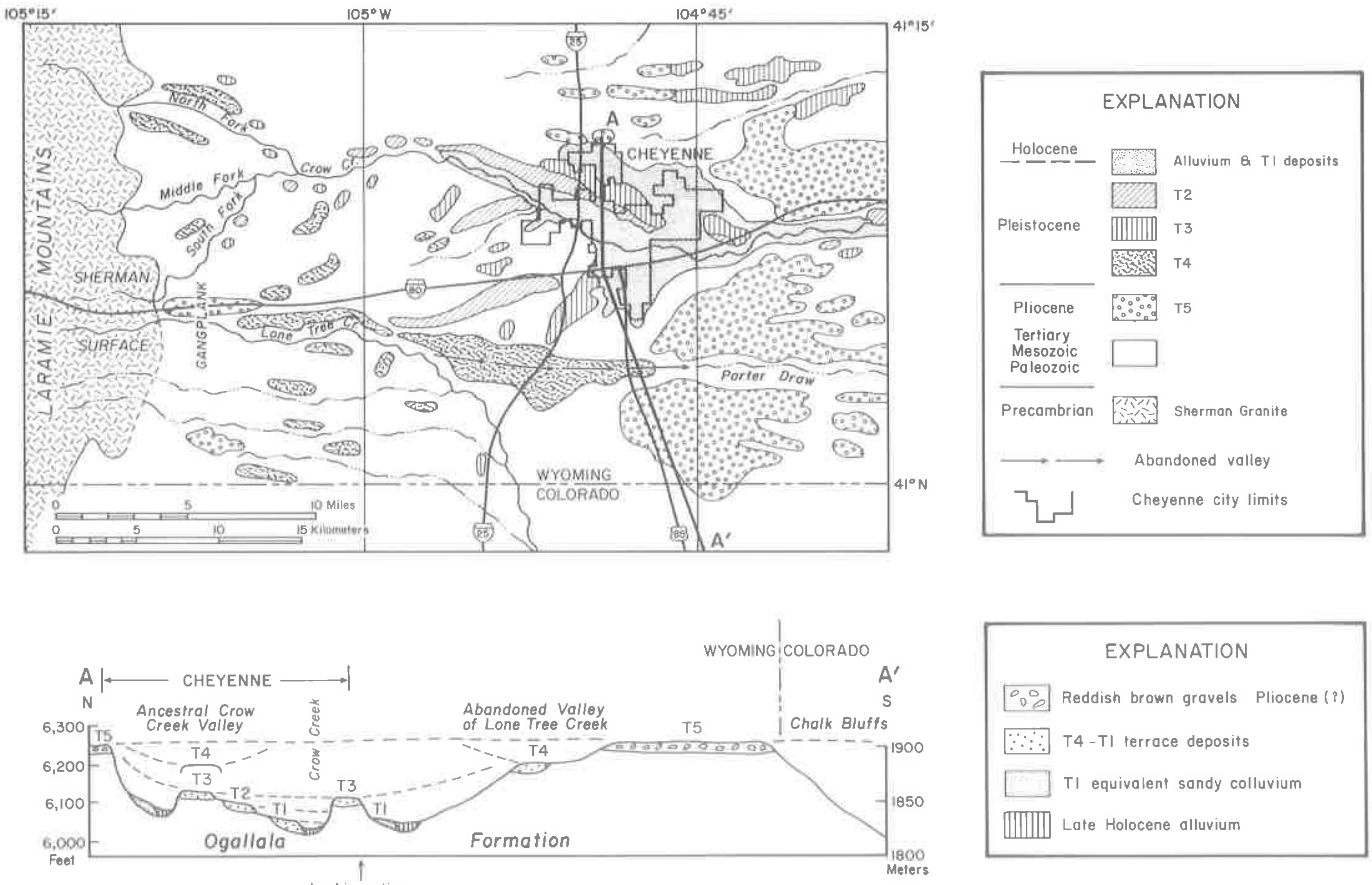

Figure 10. A. Map of late Cenozoic deposits and principal streams in the Gangplank and Cheyenne areas. B. Profile showing late Cenozoic benches, terraces, the present flood plain, and associated deposits from Cheyenne to the Chalk Bluffs.

TABLE 1. SELECTED TERRACE SEQUENCES IN THE POWDER RIVER BASIN*

	Powder, Belle Fourche North Platte Rivers Clear Creek (Leopold and Miller, 1954)	Upper Powder River, Middle Fork Powder River (Kohout, 1957)	Tongue River Goose Creek, Piney Creek, Clear Creek (Alden, 1932; Sharp, 1949)	Crazy Woman Creek (Hose, 1955)	North Platte River (Albanese and Wilson, 1974)	Clear Creek (Mapel, 1959)	Little Goose Creek (Ebaugh, 1976)
Holocene	Lightning terrace 4 to 7 ft; 1–2 m				T1 5 ft; 1.5m		Lightning terrace 0 to 3 ft; 0–1 m
	Moorcroft terrace 8 to 12 ft; 2–4 m	Terrace 1 6 to 12 ft; 2–4 m				Qt1 9 to 25 ft; 3–8 m	
		Terrace 2 15 to 25 ft; 4–8 m	No. 4 25 ft; 8 m	Qtg 1 5 to 40 ft; 2–12 m	T2 30 ft; 9 m		Kaycee terrace 6 to 50 ft; 2–15 m
	Kaycee terrace 20 to 50 ft; 6–15 m	Terrace 3 30 to 50 ft; 9–15 m					
Pleistocene		Terrace 4 60 to 80 ft; 18–24 m		Qtg2 40 to 90 ft; 12–27 m	T3 80 to 100 ft; 24–30 m		Qg3 50 to 100 ft; 15–30 m
		Terrace 5 90 to 110 ft; 27–34 m		Qtg3 90 to 130 ft; 27–40 m			
		Terrace 6 120 to 150 ft; 37–46 m	No. 3 115 ft; 35 m		T4 120 to 150 ft; 37–46 m	Qt2 130 to 160 ft; 40–49 m	Qg4 100 to 220 ft; 30–67 m
		Terrace 7 215 to 235 ft; 66–72 m	No. 2 265 ft; 81 m	Qtg4 130 to 300 ft; 40–92 m	T5 200 ft; 61 m	Qt2+ 170 to 220 ft; 52–67 m	
			No. 1 285 ft; 87 m			Qt3- 230 to 275 ft; 70– 84 m	Qg5 240 to 300 ft; 73–92 m
				Qtg5 300+ ft; 92+ m		Qt3 300 to 350 ft; 92–107 m	Qg6 300 to 400 ft; 92–122 m
						Tt 450 to 600 ft; 137–183 m	Qg7 500 to 600 ft; 153–183 m

*The relative positions of terraces in this chart are based on approximate heights above the modern streams but do not imply accurate correlation. Little work has been done on relative or absolute dating of these deposits and surfaces. Because of the variable fluvial setting of these streams (trunk stream vs. tributary; mountains vs. piedmont), correlation based solely on relative position in the landscape can be inaccurate and misleading.

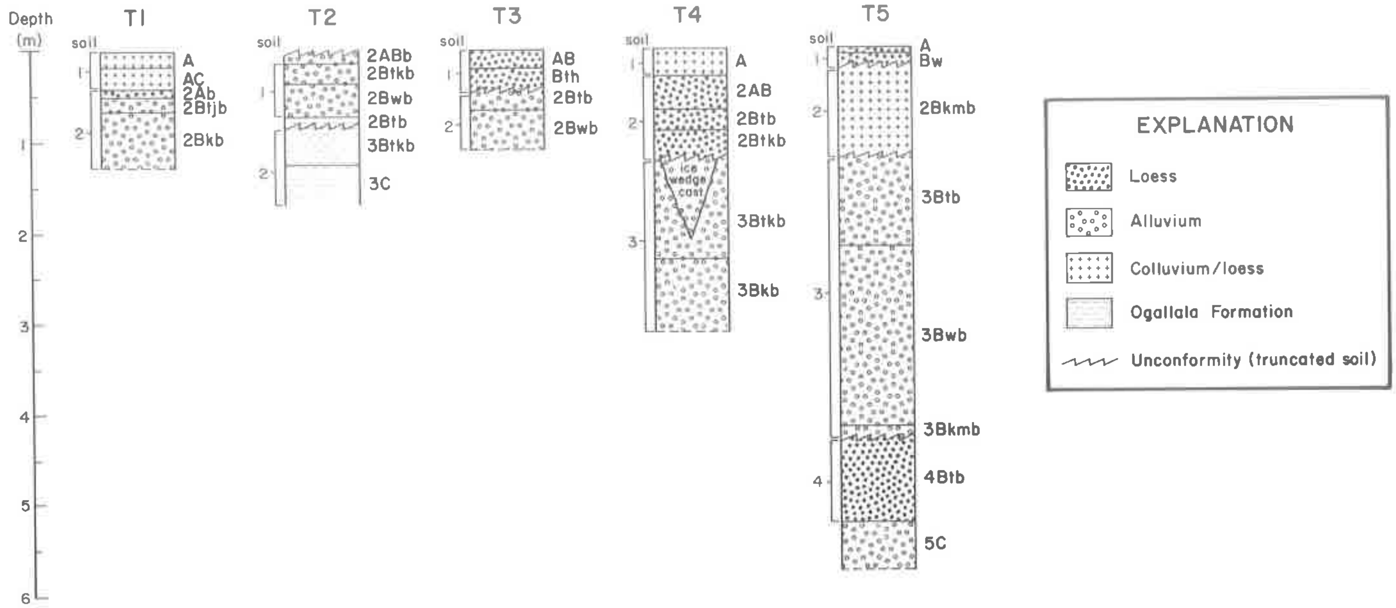

Figure 11. Selected soil profiles developed in deposits of the terrace sequence in the Cheyenne area (prepared by E. T. Karlstrom and R. G. Reider).

TABLE 2. PROVISIONAL CLASSIFICATION, HORIZON THICKNESSES, DEPTH OF LEACHING, CARBONATE STAGE, AND MAXIMUM COLOR FOR OXIDATION OF SOILS OF TERRACES AND HIGH SURFACES NEAR CHEYENNE

Terrace	Soil	Soil Classification*	Horizon Thicknesses (cm) A	Bw	Bt	Bk	DOL (cm)†	$CaCO_3$ Stage§	Maximum Color of Oxidation**
1	1	Haploboroll	18	0	0	0	40	—	10YR 4/3d
1	2	Haplargid?	6	0	18	62+	24	II+	10YR 5/3d
2	1	Argiboroll	9	37	25	25	12	-I	7.5YR 5/4d
2	2	Argiboroll?	0	0	48	48	—	II	10YR 5.5/4m
3	1	Arbigoroll	20	23	0	0	43	—	10YR 4/4m
3	2	Argiboroll?	0	55+	22	0	67+	—	7.5YR 5/5m
4	1	Haploboroll	30	0	0	0	—	—	10YR 3.5/3d
4	2	Argiboroll	38	0	52	30	60	I	10YR 6/5d
4	3	Paleustoll	0	0	110	180+	180+	III	5YR 4.5/6m
5	1	Cryoboroll	4	12	0	0	18	—	10YR 4/5m
5	2	Peleargid (Paleustoll?)	0	0	0	100	0	III+	7.5YR 7/4m
5	3	Paleudoll	0	0	100	200	300+	—	5YR 5/5m
5	4	Paleudoll	0	0	95	0	95+	—	5YR 4/6m

*Soil Survey Staff (1975).
†Depth of leaching.
§Birkeland (1984).
**D = dry; m = moist.

COLORADO PIEDMONT SECTION

Richard F. Madole

INTRODUCTION

The Colorado Piedmont section of the Great Plains (Fig. 12) is an erosional inlier. It is distinguished by having been stripped of the Miocene fluvial rocks (Ogallala Formation) that cover most of the adjoining Great Plains and by having a surface that is topographically lower than the surrounding regions. Prominent bluffs separate the Colorado Piedmont from the High Plains of Wyoming to the north, and low, less definite escarpments separate it from the High Plains to the east. The southern boundary of the Colorado Piedmont is arbitrary; slopes between it and the much higher lava-capped landscapes of the Raton section (Figs 1, 12) are transitional over a distance of several tens of kilometers.

The Miocene and Pliocene rocks and the underlying Paleocene, Eocene, and Oligocene rocks have been eroded from most of the Colorado Piedmont by the South Platte and Arkansas Rivers and their tributaries. Paleocene through Oligocene rocks remain, however, on the higher part of the interfluve between the South Platte and Arkansas Rivers. The largest part of the Colorado Piedmont is underlain by shale, claystone, and siltstone, mostly Upper Cretaceous, which give rise to low and rolling topography. Maximum relief is about 450 m, which is the maximum height of the divide between the South Platte and Arkansas Rivers. As much as 300 m of relief exists between the South Platte River and the line of bluffs that forms the northern boundary of the region.

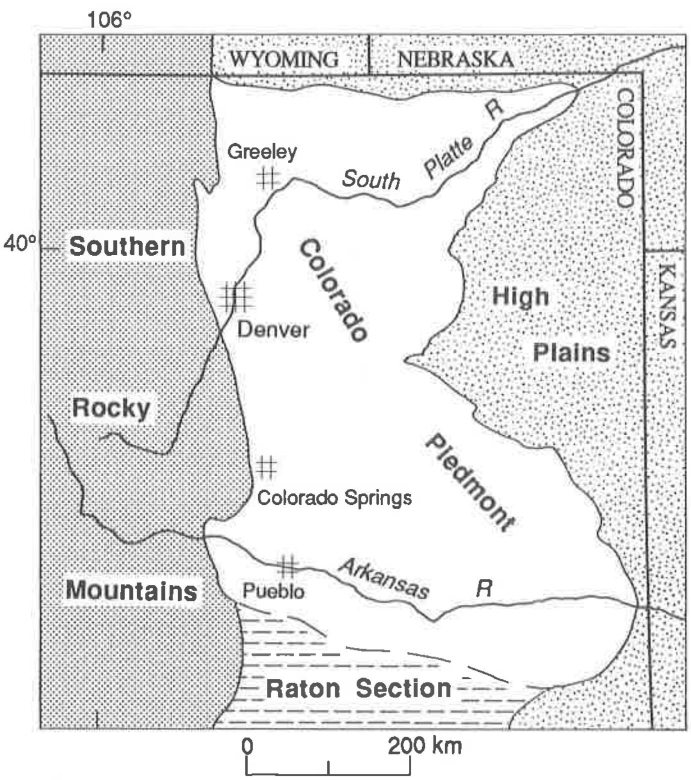

Figure 12. Map showing the Colorado Piedmont and surrounding physiographic subdivisions. Piedmont boundary is dashed where arbitrary. Ogallala Formation (Miocene) caps most of the High Plains, but has been eroded from the Colorado Piedmont.

QUATERNARY STRATIGRAPHY

The Quaternary stratigraphy of the Colorado Piedmont is dominated by fluvial deposits of the South Platte and Arkansas Rivers and their tributaries and by eolian sediments derived in large part from deposits of those streams. Most previous studies have been concentrated in a zone of 10 to 30 km wide along the mountain front. There, the Quaternary fluvial stratigraphy is conspicuously expressed in landforms, such as pediments, benches, mesas, and stream terraces. The benches and mesas are capped by coarse gravel, generally boulder gravel. Many benches and mesas mark the axes of paleochannels now inverted in the landscape. Topographic inversion is common along the mountain front because boulder and cobble gravels deposited during Quaternary time have been more resistant to erosion than the older sedimentary rocks upon which the gravels were deposited. The geomorphically prominent Quaternary fluvial units have created an angular landscape that dies out eastward a few tens of kilometers from the mountain front, largely because the caliber of sediment underlying the surfaces decreases rapidly eastward. Where pebbles and small cobbles are, or were, the dominant framework clasts, the deposits have been much less resistant to erosion than where boulders and large cobbles are the dominant framework clasts, as they are along the mountain front. Also, beginning about 10 km east of the mountain front, eolian sediment blankets most of the interstream areas, imparting a smooth appearance to the landscape that contrasts with the angular landscape of the bench-and-mesa zone along the mountain front.

Alluvial deposits and lithofacies

The vertical sequence of lithofacies in terrace and valley-floor alluvium varies according to whether the source stream headed in the mountains or drained largely or wholly on the piedmont. Alluvial deposits of mountain streams are mostly clast-supported gravel composed chiefly of Precambrian igneous and metamorphic rock, whereas alluvial deposits of piedmont streams are predominantly sand and silty sand. Basal gravels are common in the valley fills of piedmont streams, but are thin and generally composed of Upper Cretaceous sedimentary rock and/or Precambrian igneous and metamorphic rock reworked from Tertiary gravel. Major streams heading in the mountains include the South Platte and Arkansas Rivers and several tributaries to the South Platte north of Denver that head in the glaciated part of the Front Range (Fig. 13). The principal streams heading on the Colorado Piedmont are creeks, most of which flow only intermittently, that trend north or south, or northeast or southeast, to the east-flowing

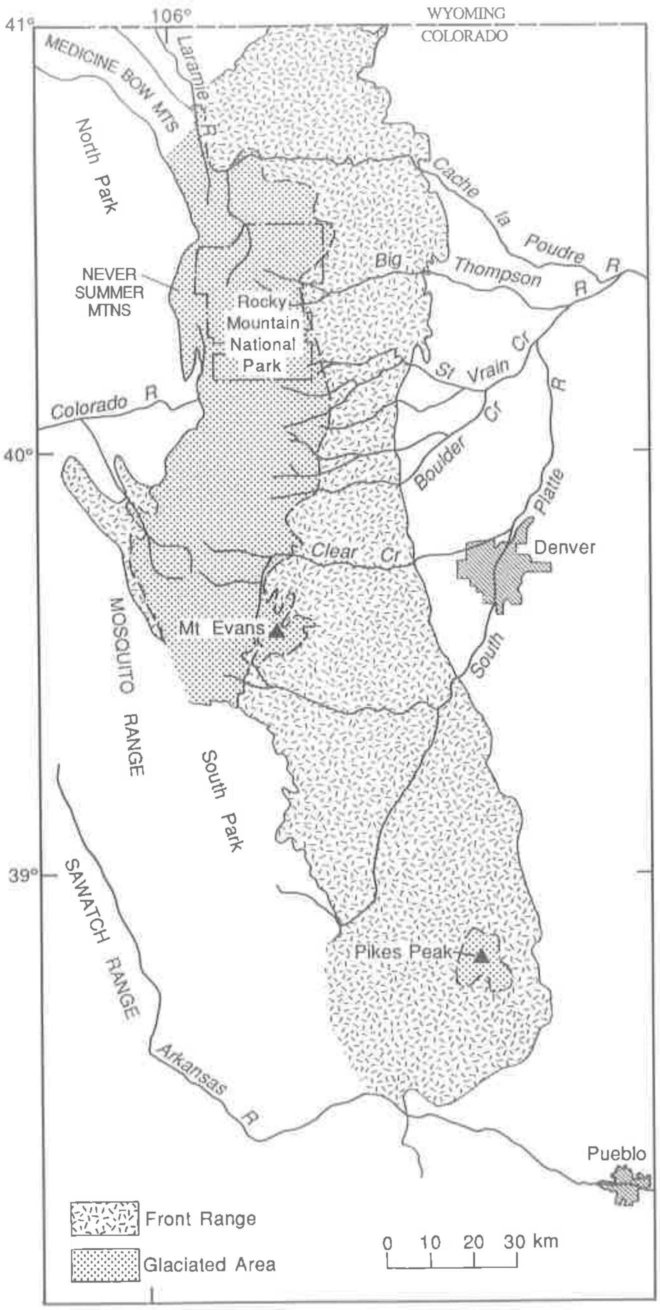

Figure 13. Generalized map of the glaciated area of the Front Range. The boundary of the glaciated area connects the downvalley limits of individual Pleistocene glaciers along a line drawn normal to valley trends. Only about a third of this area was ice covered; the rest, mostly interfluves, was higher than the glaciers. The boundary is omitted on the north and west where Front Range glaciers merged with glaciers from other ranges.

South Platte and Arkansas Rivers. The creeks tend to be underfit; commonly, they are only 5 to 10 m wide, yet they occupy flat-floored valleys as much as 1 to 3 km wide.

For at least 20 to 30 km east of the mountain front, the terrace deposits and valley-floor alluvium of streams that head in glaciated areas consist mainly of vertical sequences of superposed longitudinal bars. The bars were deposited by braided streams of the Scott type (Miall, 1977, 1978). Along the South Platte and Arkansas Rivers, terrace deposits and valley-floor alluvium typically are 15 to 30 m thick; locally, especially in the eastern (downvalley) part of the region, they are more than 60 m thick. South Platte tributaries that head in glaciated areas have valley fills that are much thinner than the alluvial fill in the main valley. In most tributaries, valley fills range in thickness from 8 to 10 m near the mountain front to as little as 2 to 3 m about 20 to 30 km east of the mountain front. The mean size of framework clasts in the clast-supported gravel of these valley fills also decreases eastward. Small boulders and large cobbles are common within 1 to 3 km of the mountain front, whereas pebbles are the dominant framework clast 20 to 30 km farther east.

Ground-water studies have been made in most of the drainage basins that head on the Colorado Piedmont. Although these studies generally do not provide detailed information about alluvial stratigraphy and chronology, they do document the distribution and thickness of alluvial deposits. Drill-hole data show that the alluvium beneath the floors of the larger piedmont valleys is as much as 12 to 30 m thick (Babcock and Bjorklund, 1956; Hershey and Schneider, 1964; McGovern, 1964; Smith and others, 1964; Weist, 1965; Coffin, 1967), which is notably thicker than the valley fills of South Platte tributaries heading in glaciated areas. Holocene alluvium is widespread on the floors of the piedmont valleys, but is probably a veneer over thicker sections of Pleistocene sediment. Alluvium of early and middle Pleistocene age also is present on terraces and uplands adjacent to the piedmont streams (Soister, 1972).

Alluvial sequence

Nine named alluvial units that range in age from Pliocene and Pleistocene(?) to Holocene are present in the Colorado Piedmont (Fig. 14). The Nussbaum Alluvium is of Blancan age and may be entirely Pliocene (Scott, 1982). Alluvium of early and early middle Pleistocene age (Rocky Flats Alluvium and Verdos Alluvium) is preserved mainly in a narrow band (3 to 5 km wide in most places) along the mountain front and in locations flanking the South Platte and Arkansas Rivers (Sharps, 1976; Scott, 1982; Scott and others, 1978). The Nussbaum Alluvium also is present locally along the South Platte and Arkansas Rivers, but is most extensive far to the east of the mountain front on the south flank of the divide between the Arkansas and South Platte Rivers (Sharps, 1976; Scott and others, 1978; Tweto, 1979). Near the mountain front, the Rocky Flats and Verdos Alluviums commonly cap ridges, buttes, and mesas. Previous workers regarded these two alluvial units as pediment deposits

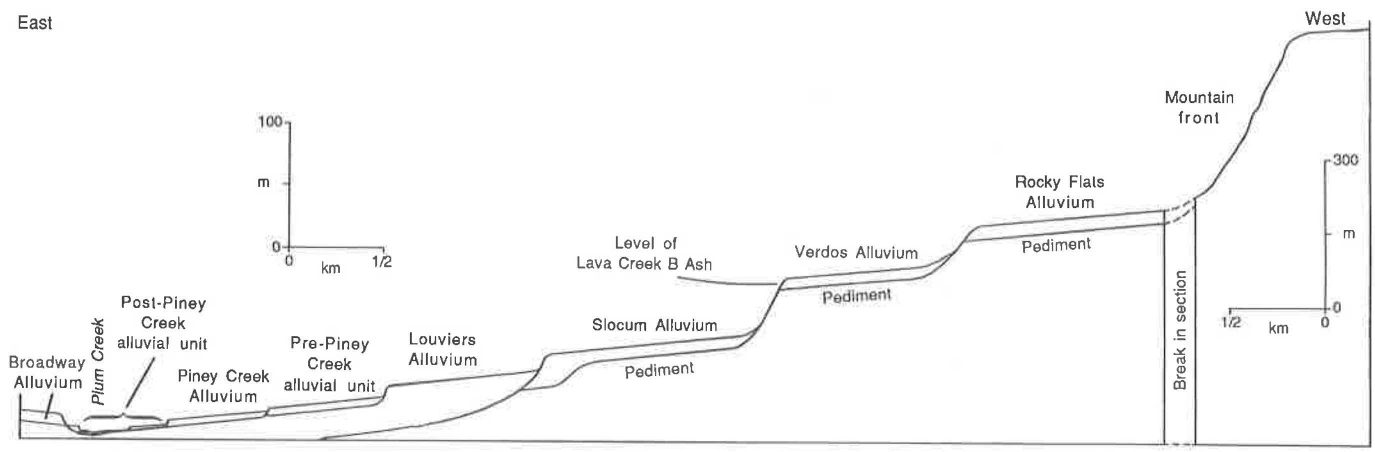

Figure 14. Diagrammatic cross section showing the stratigraphic relations of alluvial units in the Denver area, modified from Scott (1963b). Broadway Alluvium is shown as overlying Louviers Alluvium (east end of section) based on work along the South Platte River and Plum Creek. These units have a different spatial relationship in valleys heading in the glaciated part of the Front Range. There, Broadway Alluvium underlies the valley floor and Louviers Alluvium is on a terrace above the valley floor.

(Fig. 14), but some of the alluvial remnants included in these units may be simply isolated paleochannel deposits. Unlike the Rocky Flats and Verdos deposits, the Slocum is widespread both on pediments and in stream terraces, although its type locality is on a pediment (Scott, 1960, 1962).

Nussbaum Alluvium. The Nussbaum Alluvium consists of sand and gravel as much as 45 m thick. It is present mainly in the southern part of the Colorado Piedmont and is particularly extensive along the south flank of the divide between the Arkansas and South Platte Rivers (Scott, 1963a; Scott and others, 1978; Soister, 1967; Sharps, 1976). The Nussbaum Alluvium is on surfaces that project to altitudes slightly lower than the altitude of surfaces overlain by the Ogallala Formation. Gilbert (1897), who named the Nussbaum Alluvium, regarded it as pre-Pleistocene in age, as did Scott (1982, p. 7). Scott considered the Nussbaum to be early and middle Blancan in age on the basis of mammal remains and geomorphic position. These same criteria are used to correlate the Nussbaum with the Broadwater Formation of western Nebraska. The Broadwater contains fossils that indicate it is equivalent in age to the Blanco Formation of Texas (Scott, 1982; Dalquest, 1975). The Blanco Formation is mostly Pliocene but may include early Pleistocene beds, depending on the age assigned to the Pliocene-Pleistocene boundary (Holliday, 1988). Small gravel remnants in a few places along the mountain front may be Nussbaum Alluvium because they are higher in the landscape than remnants of the Rocky Flats Alluvium, the next younger deposit in the sequence. The average height of remnants of Nussbaum Alluvium above present streams is about 140 m.

Rocky Flats Alluvium. The Rocky Flats Alluvium consists mostly of coarse, clast-supported gravel, generally 3 to 12 m thick. Remnants are common along the mountain front and along the South Platte and Arkansas Rivers. The alluvium is deeply weathered and contains a strongly developed paleosol that includes Bt/Btk/K horizons (Fig. 15). In most places, the entire section of Rocky Flats Alluvium is oxidized, and hues as red as 5YR are common in the upper few meters. The larger remnants of this alluvium are at the west edge of the Colorado Piedmont adjacent to mountain valleys that head in unglaciated drainage basins. Initially, the Rocky Flats Alluvium may have been extensive where all major streams emerged from the mountains, but along the largest valleys, which head in glaciated areas, intense erosion has apparently removed most of the alluvium. The large remnants of Rocky Flats Alluvium adjacent to unglaciated valleys show that not all coarse gravel deposits in the region are glaciofluvial. Nevertheless, most of the coarse gravel deposits in unglaciated valleys probably accumulated during glacial times when a lowering of timberline and expanded periglacial activity caused an increase in runoff and sediment yield that favored stream aggradation.

Deposits of Rocky Flats Alluvium commonly are 100 ± 10 m higher than present streams and about 30 m higher than remnants of the Verdos Alluvium, the next younger deposit in the sequence (Fig. 14). Rocky Flats Alluvium is probably early Pleistocene in age because it is topographically higher than the Verdos Alluvium, which is early middle Pleistocene. The presence, in places, of the Lava Creek B volcanic ash in the fine-grained upper part of the Verdos Alluvium establishes the age of the Verdos as early middle Pleistocene. Izett and Wilcox (1982) considered the Lava Creek B ash to be about 620 ka on the basis of the K-Ar age of the source-area tuff.

Verdos Alluvium. The Verdos Alluvium is similar in character, thickness, and distribution to the Rocky Flats Alluvium, but it is about 70 ± 5 m higher than present streams (Scott, 1960, 1962). The Verdos Alluvium is deeply weathered and has a strongly developed surface paleosol (Fig. 15). Beds and lenses of Lava Creek B volcanic ash are present in the Verdos Alluvium at the type locality and a few other places in the Denver area (Hunt, 1954; Scott, 1963b). This 620-ka ash in the fine-grained upper part of the Verdos establishes the age of the alluvium as early middle Pleistocene.

Slocum Alluvium. The Slocum Alluvium is present both as pediment deposits (Scott, 1960) and stream-terrace deposits. The alluvium is mostly clast-supported gravel that ranges in thickness from 3 to 27 m. It is deeply oxidized and generally contains abundant decomposed clasts. A strongly developed paleosol that includes Bt/Btk/K horizons is developed in the alluvium (Fig. 15). The Slocum Alluvium is 6 to 30 m higher than streams, depending on the size of the stream—about 30 m higher along the South Platte and Arkansas Rivers and 6 m higher along tributaries of the South Platte River that head in the glaciated part of the Front Range (Fig. 13).

The Slocum Alluvium is middle Pleistocene in age. On the basis of position in the landscape, it is probably much closer in age to the Louviers Alluvium than to the Verdos Alluvium because it is 40 to 65 m lower than the Verdos surfaces and only 6 to 12 m higher than the Louviers surfaces. A bison horn core from the lower part of the gravel facies of the Slocum Alluvium near the Arkansas River yielded a uranium-series age of 160 ± 60 ka (Scott and Lindvall, 1970), which later was corrected to 190 ± 50 ka (Szabo, 1980). If the corrected age is valid, then the Slocum Alluvium, given the evidence for the age of the Louviers Alluvium discussed next, probably is about 240 ka, the maximum indicated by the analytical error of the uranium-series age.

Louviers Alluvium. The Louviers Alluvium, as described initially by Scott (1960, 1963b), included two facies: a coarse facies (cobble and pebble gravel overlain by sand and minor silt) adjacent to large streams, and a fine-grained facies (thin, basal pebble gravel overlain by much thicker sand and silt) adjacent to small streams. This dichotomy illustrates differences described earlier between deposits of streams heading in the mountains and deposits originating on the piedmont. The type locality of the Louviers Alluvium is the only one in Scott's (1960) sequence (Fig. 14) that is not adjacent to a stream originating in the mountains. Thicknesses of the two facies are similar to those of the other named alluvial deposits.

Along Plum Creek, the Louviers Alluvium is 12 to 18 m higher than the stream and 3 to 10 m higher than the late Wisconsin Broadway Alluvium. The spatial relations of the two deposits are different from those adjacent to tributaries to the South Platte that head in the glaciated part of the Front Range. In

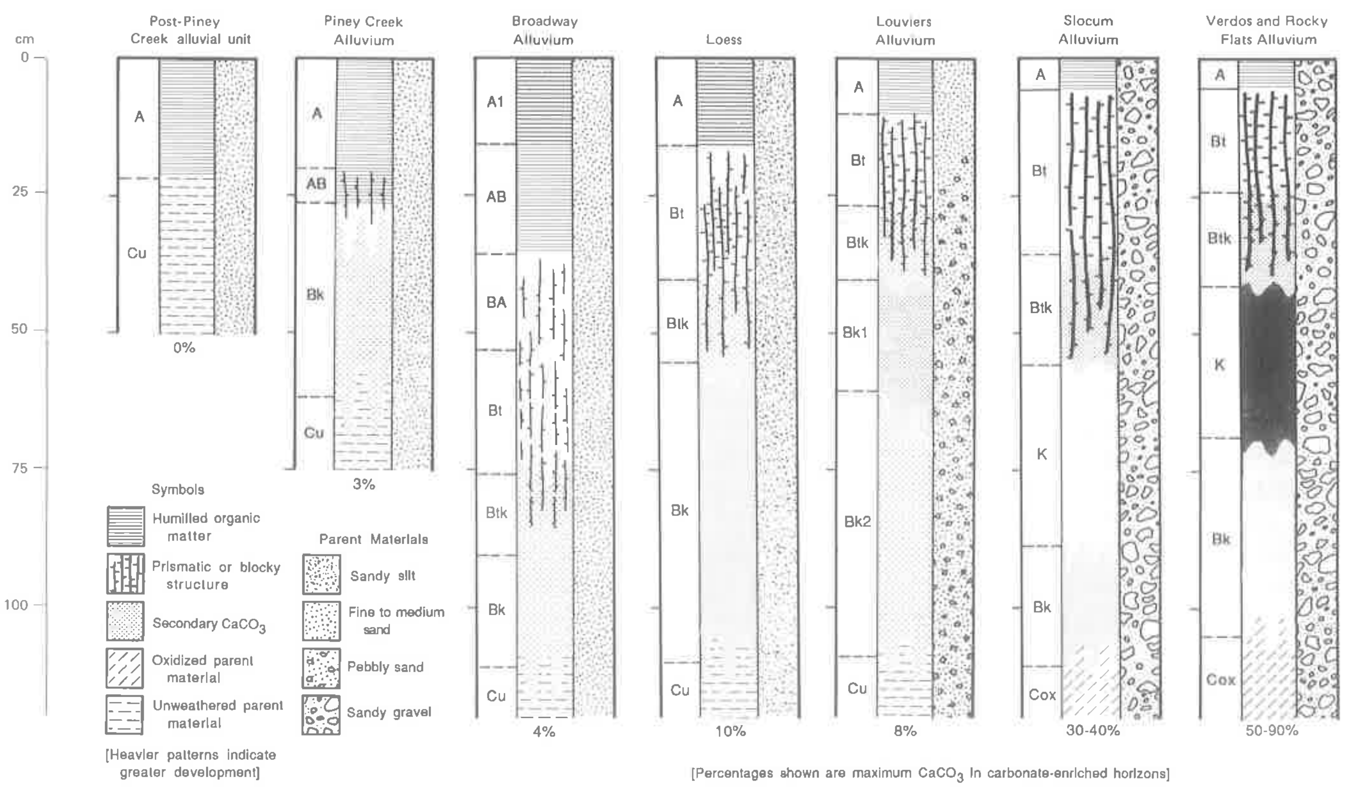

Figure 15. Diagram comparing the maximum soil development in parent materials of different ages in the Lafayette Quadrangle, Colorado, modified from Machette (1977). This quadrangle does not have streams that head in the glaciated area; hence, soil parent materials in the Louviers Alluvium and Broadway Alluvium do not include clast-supported gravel. Soils in the Slocum, Verdos, and Rocky Flats deposits are similar, except for development of calcic horizons. The thin A horizons in these older soils were developed on surfaces eroded in older B horizons. Soil-horizon nomenclature follows Birkeland (1984).

valleys that head in the glaciated area, both the Louviers Alluvium and the Broadway Alluvium 10 to 15 km downstream from the mountain front are less than 10 m higher than the streams. Locally near the mountain front, these deposits are separated by a vertical interval of only 1 to 3 m. Farther downstream, the Louviers Alluvium is associated with a terrace as much as 10 m higher than the streams, but in most places the geomorphic expression of the terrace is subdued, if not obscured, by younger eolian deposits and alluvium from small side-valley streams.

The intensity of weathering and soil formation in Louviers Alluvium is much greater than that in Broadway Alluvium, and the difference is comparable to the difference in soil development observed in till of Bull Lake age and till of Pinedale age on the east slope of the Front Range (Madole, 1969, 1976a; Shroba, 1977; Madole and Shroba, 1979). Consequently, these two sets of deposits (till of Bull Lake age and Louviers Alluvium on the one hand and till of Pinedale age and Broadway Alluvium on the other) were correlated before numerical ages were obtained for any of the deposits. These correlations cannot be verified directly because deep, narrow canyons are present between the glaciated part of the Front Range and the west edge of the Colorado Piedmont. Terminal moraines in the main glaciated valleys commonly are 17 to 40 km west of and 750 to 1,200 m higher than the Colorado Piedmont. During glaciations, the deep canyons between the glaciated area and the piedmont acted more as conduits than as depositories for glaciofluvial sediments. Not much glaciofluvial sediment is present in the canyons; most apparently was flushed through them and deposited at the west edge of the piedmont.

On the east slope of the Front Range, till of Bull Lake age is correlated with deposits of marine oxygen-isotope stage 6 (200 to 130 ka) on the basis of (1) a uranium-trend age of a relict paleosol of 130 ± 40 ka (Shroba and others, 1983), and (2) weathering and morphologic criteria (Madole and Shroba, 1979). The weathering and morphologic criteria justify correlation with other deposits of presumed Bull Lake age, including those at West Yellowstone, Wyoming, to which Pierce and others (1976) assigned an obsidian-hydration age of 140 ± 15 ka. Szabo (1980) calculated uranium-series ages for samples of bone from two localities of Louviers Alluvium near Denver; the ages determined for the two alluvial deposits were 129 ± 10 ka and 86 ± 6 ka.

Hunt (1954) believed that the gravel fill, later named Louviers Alluvium, was outwash from Front Range glaciers, and that view continues to be widely accepted. Baker (1974), in a study of Louviers Alluvium along Clear Creek west of Denver, concluded on the basis of engineering hydraulics that the coarse gravel could have been transported only by "large quantities of meltwater." This conclusion followed from calculations demonstrating that transport of the coarse gravel in this area would have required sustained discharges ten times greater than that of the largest flood during 120 years of record. Of course, the Louviers Alluvium may not be temporarily equivalent to the Bull Lake till; more of the alluvium may have been deposited in late Bull Lake time than during early Bull Lake time, and part of the alluvium may be younger than the glaciation because sediment yield may have been greatest during and immediately after deglaciation (Church and Ryder, 1972). On the other hand, glacial meltwater may not have been the only source of the increased discharges that were required for the transport of the coarse gravel, as was noted in the discussion of the Rocky Flats Alluvium.

Broadway Alluvium. Scott (1960) applied the name Broadway Alluvium to the younger of two alluvial deposits underlying the Broadway terrace of Hunt (1954) in the Denver area. The Broadway terrace is a well-defined surface, nearly 1 km wide, on the east side of the South Platte River about 7.5 to 12 m higher than the river. Hunt (1954) interpreted the gravel of the Broadway terrace and the gravel underlying the flood plain of the South Platte River to be parts of a single alluvial fill. He believed that the upper part of the fill (the deposits underlying the Broadway terrace) was late Wisconsin in age and that the part underlying the flood plain possibly was early Wisconsin in age. The late Wisconsin age assignment was based on the degree of soil development on the Broadway terrace surface, the remains of extinct fossil mammals (horse, mammoth, and camel), the association of fossil mammals with what were believed to be man-made artifacts, and the likelihood that the gravel fill is glacial outwash. Scott (1960) interpreted the lower part of Hunt's (1954) "Wisconsin" gravel fill to be Louviers Alluvium, and at that time he inferred that the gravel was early Wisconsin in age.

The topographic position and stratigraphic relations of the Broadway Alluvium in valleys heading in the main glaciated part of the Front Range (Fig. 13) are somewhat different than those along the South Platte River. In the valleys heading in the glaciated area, Broadway Alluvium is not present in terraces above the valley floor; instead, at least near the mountain front, it underlies the valley floor, generally beneath a cover of Holocene alluvium. Eastward from the mountain front, Holocene deposits become constrained to a swath that occupies only a part of the valley floor; the rest of the valley floor is underlain by Broadway Alluvium and, locally, Louviers Alluvium.

The Broadway terrace is exceptionally continuous adjacent to the South Platte River from about 15 km southwest of Denver to about 25 to 30 km east of Greeley, a distance of approximately 110 km. Prior to Hunt's (1954) work in the Denver area, the Broadway terrace had been studied in the vicinity of Greeley (Bryan and Ray, 1940), where it had been named the Kersey terrace, and the deposits underlying it had been interpreted to be Wisconsin in age. Recently, Holliday (1987) reinvestigated the Kersey terrace and also the younger and lower Kuner and Hardin terraces. He discussed evidence for correlating the deposits of these respective terraces with the Broadway Alluvium and the Holocene Piney Creek Alluvium and post–Piney Creek alluvial deposits of the Denver area. On the basis of geoarcheological evidence, Holliday (1987) concluded that alluvium underlying the Kersey terrace had ceased to be deposited by 11 to 10 ka. The evidence includes a Folsom site (occupation 11 to 10 ka) in dune sand overlying the Kersey terrace and Clovis artifacts (occupation 11.5 to 11 ka) in the upper part of the terrace alluvium.

Although the Broadway Alluvium apparently is young enough to be dated by the radiocarbon method, no ^{14}C ages have been reported. Charcoal from near the base of the Broadway Alluvium in an excavation along Big Dry Creek, a small stream that heads on the piedmont and crosses the northern edge of the Denver metropolitan area, yielded a ^{14}C age of >38,440 B.P. (Machette, 1977). The alluvial fill overlying the charcoal was mapped as Broadway Alluvium and correlated with till of the Pinedale glaciation on the basis of the weakly developed surface soil formed in the fill and also topographic and stratigraphic relations with the Louviers Alluvium (Machette, 1975a, 1977). This correlation is reasonable, but the ^{14}C age of the charcoal is older than the Pinedale glaciation, which, following Blackwelder's (1915, p. 324–325) definition, I consider to be equivalent to late Wisconsin glaciation. Pinedale glaciation on the east slope of the Front Range probably began shortly after 30 ka and ended prior to 12 ka (Madole, 1986a).

The ^{14}C age of the charcoal on Big Dry Creek provides an imprecise maximum age for the Broadway Alluvium. The charcoal may be significantly older than 38,400 B.P.—the initial charcoal assay yielded an age of >45,910 ± 2,295 B.P. (SMU-136; Machette, 1975a, 1975b)—and an unconformity, although not well expressed stratigraphically, is probably present within the alluvial fill above the charcoal. Well-dated alluvial sequences associated with archeological excavations indicate that unconformities are not always distinct in the alluvium of minor streams.

Holocene alluvial deposits. Even though a detailed discussion of Holocene alluvial stratigraphy is beyond the scope of this volume, note that three named Holocene alluvial units are present in the Colorado Piedmont. The units are referred to as pre–Piney Creek, Piney Creek, and post–Piney Creek (Fig. 14). Although the Holocene stratigraphic record may be more complex than the nomenclature suggests, the three-fold division has been useful for mapping because stratigraphic breaks other than those used to differentiate the three alluvial units are indistinct. The possibility that other stratigraphic breaks are present within the Holocene alluvial sequence is suggested by artifacts and ^{14}C ages from archeological excavations.

Eolian deposits

Eolian deposits blanket nearly a third of the plains of eastern Colorado, including a large part of the Colorado Piedmont. Loess deposits, at least in part correlated with loess deposits in the midcontinent, are widespread, as are eolian sand deposits. Eolian deposits generally are poorly exposed, and information about the number, ages, and areal distributions of units is limited. The similarity of topographic expression within dune fields tends to give a deceptive impression of stratigraphic uniformity. Where stratigraphic relations are exposed, a relatively complex stratigraphy generally underlies a seemingly uniform surface (Madole, 1986b).

Loess. Loess deposits of at least three ages are present on the Colorado Piedmont. The oldest recognized loess is middle Pleistocene in age, and the most abundant loess, at least at the surface, is late Pleistocene. Holocene loess probably is also widespread, but it is neither areally nor volumetrically significant according to regional geologic maps (Scott, 1968, 1978; Sharps, 1976, 1980; Scott and others, 1978; Colton, 1978; Trimble and Machette, 1979a, 1979b; Bryant and others, 1981). Thin Holocene loess is ubiquitous in the higher parts of the Front Range, where stratigraphic relations with a variety of surficial deposits of known ages show that loess was deposited at several different times during the Holocene (Benedict, 1981, 1985; Madole, 1976b, 1984). Except in lakes and ponds, the Holocene loess in the Front Range generally is less than 25 cm thick. Given that multiple episodes of Holocene eolian activity are represented on the Colorado Piedmont by sand, it is reasonable to expect that Holocene loess also is present here.

Late Pleistocene loess, which is by far the most extensive loess on the Colorado Piedmont, is pale brown to yellowish brown and varies considerably in thickness, both locally and regionally. Where present within 20 km of the mountain front, the loess is generally 0.1 to 4 m thick; farther east it may be as much as 40 m thick in places (Scott, 1978). Terrace deposits of the Louviers Alluvium (late middle Pleistocene) are commonly blanketed by late Pleistocene loess. Soil profiles developed in this loess contain A/Bt/Bk/C horizons in which the Bt and Bk horizons are comparable to or are more developed than those in the Broadway Alluvium and less developed than those in Louviers Alluvium (Machette, 1977; Fig. 15). In northeastern and eastern Colorado, the late Pleistocene loess has been called Peoria Loess (Hill and Tompkin, 1953; Scott, 1978; Sharps, 1980), but may include correlatives of other midcontinent late Pleistocene loess units.

Deposits of middle Pleistocene loess commonly are buried by younger deposits. Consequently, little information is available about the character, thickness, and distribution of middle Pleistocene loess; typically it is reddish brown and much more weathered than the late Pleistocene loess. The stratigraphic position and intensely weathered character of the reddish brown loess indicate that it is of middle Pleistocene age. In the Kassler area, southwest of Denver, middle Pleistocene loess overlies Slocum Alluvium (Scott, 1962, 1963b). The distribution of middle Pleistocene loess probably was similar to that of late Pleistocene loess.

Eolian sand. In the Colorado Piedmont, eolian sand generally is associated with streams and it is especially prevalent along the downwind side of the South Platte River. Similarly, a belt of eolian sand, commonly 1.5 to 7 km wide, flanks the south side of the Arkansas River through the Colorado Piedmont and across the Great Plains of Kansas (Simonett, 1960). Sediments deposited by these rivers and smaller streams have been a primary source of eolian sediment (Gilbert, 1896; Toepelman, 1924; Smith and others, 1964; Coffin, 1967; Reheis, 1980; Muhs, 1985), and residuum derived mainly from Cretaceous and Tertiary rocks has been a secondary source. Dune forms and the orientation and distribution of dunes with respect to sediment sources indicate that eolian sand was transported and deposited by northwesterly winds. Decreases in grain size and changes in mineralogy away from sediment sources also indicate transport by northwesterly winds (Coffin, 1967; Reheis, 1980; Muhs, 1985). Prevailing winds are northwesterly at present, except during the summer months when south-southeasterly winds are common. Most dune sand presently is stable and covered with vegetation. Parabolic dunes are dominant, although the concentric portion of the dunes is not well preserved everywhere. Sharp-crested ridges, remnants of the trailing dune arms, generally are common to abundant.

Stratigraphic relations (exposed in widely scattered blowouts and stream cutbanks) and differences in degree of soil formation and roughness of dune topography indicate that eolian sand was deposited at several different times during the Pleistocene and Holocene. Deposits of late Holocene age are widespread and probably overlie older Pleistocene eolian sand in most places. In spite of their abundance, the eolian deposits have received relatively little study, and a chronology comparable in detail to that of fluvial deposits has not been developed.

Eolian sand of Holocene age is characterized by weakly developed A/C or A/AC/C soil profiles and a rough surface topography. Judging from differences in dune-surface roughness, dunes formed during more than one interval of Holocene time. Late Pleistocene eolian sand commonly underlies topographically low areas between and within clusters of Holocene dunes. In places, most notably in a broad belt east of the South Platte River in the Denver area, late Pleistocene eolian sand is dominant. The older sand is characterized by a more subdued surface morphology and moderately developed A/Bt/C and A/Bt/Bk/C soil profiles.

The ^{14}C age control for times of Holocene dune formation is sparse, and most is from studies on the High Plains east of the Colorado Piedmont. Also, most ^{14}C ages are from material that existed either before or after eolian activity and may or may not closely date the beginning or end of sand deposition. Radiocarbon ages delimit episodes of Holocene eolian sedimentation at localities in western Nebraska and eastern Wyoming and Colorado (Brice, 1964; Maroney and Swinehart, 1978; Gaylord, 1982; Madole, 1986b), but the episodes delimited are not necessarily the same at each locality. Combining ^{14}C ages from several localities in the Nebraska Sand Hills and selected Rocky Mountain basins, Ahlbrandt and others (1983) presented evidence of four phases of Holocene eolian activity. The ages of the four phases are early Holocene, middle Holocene, late Holocene, and recent. In the Southern High Plains, Gile (1979, 1981) also recognized four ages of Holocene dune sand (11.5 to 7 ka, 7 to 4 ka, 4 to 0.1 ka, and less than 0.1 ka). Furthermore, Gile identified six older eolian sands, of which two are middle Pleistocene, two may be late middle or late Pleistocene, and two are late Pleistocene in age (about 25 to 15 ka and 13 to 11 ka). Additional work may show that the stratigraphic record of eolian activity in the northern Great Plains is comparable in length and complexity to that in the southern Great Plains.

HIGH PLAINS AND PLAINS BORDER SECTIONS IN NEBRASKA, KANSAS, AND OKLAHOMA

William J. Wayne and James S. Aber

INTRODUCTION

The High Plains Section extends from the Pine Ridge Escarpment in Nebraska and South Dakota south through Kansas and the Oklahoma panhandle (Fenneman, 1931; Fig. 1). It is closely coincident with the area underlain by the late Miocene Ogallala Group, but in places it extends beyond the limits of this unit. On the east it is bordered by the Central Lowland Province (Chapter 17, this volume). In Nebraska and northeastern Kansas, the boundary is placed arbitrarily at the western limit of surface tills. The Plains Border section is a transition zone across central Kansas.

The early glaciations caused radical drainage changes in the northeastern part of the High Plains. The ice sheets repeatedly blocked the eastward-flowing drainage of all rivers north of 39°N, filled most of the existing valleys, and supplied meltwater that was directed southward along the ice margins. In Nebraska a series of southeastward-trending valleys marks positions of the ice margin as it melted (Wayne, 1985), and in Kansas, other trenches were cut in the bedrock.

STRATIGRAPHY

The names of many of the stratigraphic units used in the Kansas and Nebraska part of the High Plains Quaternary were applied originally as much to conceptual units as to valid lithostratigraphic units. As knowledge increased, interpretations of stratigraphic units in some type sections changed. In some cases, new units and type sections were designated according to new conceptual models. Figure 16 shows selected units whose stratigraphic placement and utility can be reasonably well documented by recent studies.

Upper Miocene stratigraphy

The Ogallala Group in Nebraska and Kansas, the Rexroad Formation in western Kansas, the Delmore Formation in central Kansas, and the Laverne Formation in Oklahoma, all composed of arkosic gravel, sand, silt, and clay, were deposited by rivers that built an alluvial apron eastward from the Rocky Mountains. Long regarded as Pliocene (Bayne and O'Connor, 1968), these units now are considered to be middle to late Miocene (Boellstorff, 1976; Gutentag and others, 1984; Swinehart and others, 1985). The Miocene was a time of aridity and extensive development of grasslands, which culminated in development of a thick paleosol with a resistant K horizon still preserved as a caprock on uplands of the western High Plains.

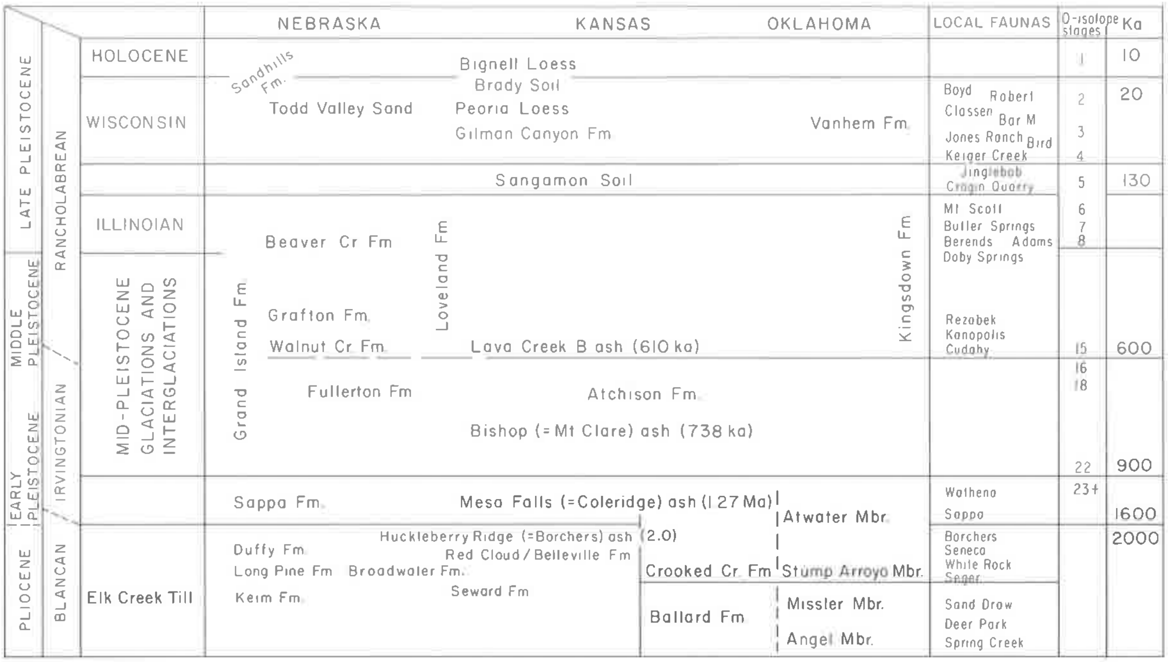

Figure 16. Correlation chart of late Pliocene to Holocene stratigraphic units in the Nebraska-Kansas High Plains.

Late Pliocene stratigraphy

Streams that spread a veneer of gravel east of the Rocky Mountains to produce the Nussbaum Alluvium of the Colorado Piedmont (Madole, this chapter) carried sediments east into Nebraska and Kansas. North Platte River gravels, characterized by anorthosite and abundant clasts of coarse-grained Sherman Granite, were named the *Broadwater Formation* in western Nebraska (Schultz and Stout, 1945). Traced northeastward in the Sandhills Region, it is exposed along the Middle and North Loup Rivers and the Niobrara Valley (Stanley and Wayne, 1972). There it was named the *Long Pine Formation* by Skinner and Hibbard (1972). Skinner also named an underlying silt unit the *Keim Formation* and an overlying one the *Duffy Formation.* All of these sediments contain vertebrate fossils of Blancan age. A nonfossiliferous gravel, the *Pettijohn Formation,* overlies the Duffy Formation (Fig. 16).

The *Red Cloud Gravel* (Fig. 16) is a sediment of Laramie Range provenance that was deposited in central southern Nebraska; its relation to other Pleistocene stratigraphic units has been uncertain since the unit was named (Schultz and others, 1951). Clasts include quartzite, Sherman-type granite like that in the North Platte River valley, and a smaller percent of anorthosite. The original type section is now very poorly exposed, but similar gravels are exposed nearby.

The Red Cloud Gravel accumulated as a result of an early glacial diversion of the stream that later became the Platte River (Fig. 4b). It probably represents the route into which that stream was shunted as a result of tectonic activity along the Chadron arch, rather than blockage by a continental ice sheet margin (Stanley and Wayne, 1972), although the diversion may have taken place contemporaneously with an early Pleistocene glaciation.

The *Holdrege Formation* was defined by Lugn (1934, p. 342, and 1935, p. 92) as gravel of Rocky Mountain origin deposited during the "Nebraskan glaciation." Reed and Dreeszen (1965, p. 18, 26, 51) regarded the part of the well record Lugn had used to define the Holdrege gravel to be more likely Ogallala Formation; they retained the name and concept of the Holdrege Formation but designated another test hole 200 km to the northeast as a substitute type section (Fig. 1). The river that deposited the Red Cloud Gravel followed a course southeastward through the region near Holdrege, so the Holdrege gravel, as conceived by Lugn, may be the same unit as the Red Cloud.

Although most subsequent authors (Lugn, 1934, 1935; Miller and others, 1964; Condra and others, 1947) have either disregarded or overlooked it, the *Belleville Formation* (Wing, 1930) probably is the same sedimentary unit as the Red Cloud. Fine-grained sediments in the upper part of the Belleville have yielded late Blancan Mammalian fossils (Eshelman, 1975). The name Belleville Formation has priority over both Red Cloud and Holdrege and probably should replace them, but we have included both names (Fig. 16).

In southwestern Kansas and the Oklahoma Panhandle, two sheets of alluvium, *Ballard Formation* and *Crooked Creek Formation* above, were deposited by the ancestral Arkansas River (Zakrzewski, 1975). Each consists of a basal gravel (Angell Member and Stump Arroyo Member, respectively). The sediment came from local erosion of Ogallala and Rexroad deposits as well as from the Rocky Mountains. The Huckleberry Ridge volcanic ash bed (Pearlette type B or Borchers, fission-track dated at 2.0 Ma [Boellstorff, 1976] and correlated with the Huckleberry Ridge Tuff in the Yellowstone area, K-Ar dated 2.1 Ma [Izett, 1981]) is in the Atwater Member. Milder summers and warmer winters than today are indicated by both vertebrate and molluscan remains in local faunas of Blancan age in southwestern Kansas (Rexroad, Borchers, Sanders, Deer Park) and northern Nebraska (Seneca, Sand Draw; Hibbard, 1970; Taylor, 1960).

Condra and others (1947, p. 15–16) applied the name *Seward Formation* to fine-grained sediments as much as 70 m thick in the lower parts of valley fills on bedrock in southeastern Nebraska, but designated no type section. They considered it to be a distal facies of the Ogallala. Frye and Leonard (1952) did not include it in the Pleistocene units of Kansas. However, Reed and Dreeszen (1965) assigned an early Pleistocene age to the Seward Formation and designated a type section. Wayne (1982) restudied the type section and determined that it consists of two distinctive materials; the upper part is loess containing cool-climate land snails. He correlated it with the Pliocene(?) Elk Creek Till (Fig. 16), chiefly on the basis of stratigraphic position and degree of development of the strong paleosol it bears.

Early Pleistocene stratigraphy

The name *Sappa Formation* was introduced by Condra and Reed (1950; also see Frye and others, 1948) to replace the name, Upland Formation, which Lugn (1935) had given to a unit composed of stratified to massive silt that he inferred to be "Yarmouth" in age. No description of a type section appeared until much later (Reed and Dreeszen, 1965, p. 56). At the type section of the Sappa Formation in Harlan County, Nebraska, a volcanic ash bed, long called "the Pearlette ash," is present near the top of about 10 m of fine-grained sediments. On the basis of this ash, the Sappa (Upland) was assigned a "late Kansas" or "early Yarmouth" age (Condra and others, 1947; Condra and Reed, 1950; Frye and Leonard, 1952). Sediments were identified as Sappa Formation if an ash bed was in or at the top of a sequence of grayish silt beds. The "Pearlette ash" was regarded as "the most precise key horizon of the entire . . . sequence" (Reed and others, 1965, p. 197).

The ash bed in the type section of the Sappa (Pearlette type S) has a fission-track age of approximately 1.2 Ma (Naeser and others, 1973; Boellstorff, 1976, 1978) and is the airfall equivalent of the Mesa Falls Tuff in the Yellowstone National Park area (Izett, 1981). (The Coleridge ash of Boellstorff [1976] in northeastern Nebraska is the same ash) Now the Sappa Formation is restricted to deposits more or less coeval with the Mesa Falls tephra layer and therefore is early Pleistocene rather than "post-Kansan" in age (Fig. 16).

Quaternary deposits capped by a paleosol are preserved beneath till at Wathena in northeastern Kansas (Bayne, 1968; Einsohn, 1971; Dort and others, 1985), where fine silty sand to silty clay lake beds about 3.5 m thick overlie chert gravel. The lake sediments, which accumulated in a shallow pond or abandoned channel of the ancestral Missouri River, have yielded a large fauna of fish, amphibians, reptiles, birds, and mammals. The Wathena local fauna is judged to be slightly younger than the Sappa local fauna (Martin and Schultz, 1985). These Irvingtonian faunas indicate that the early Pleistocene continued to be mild climatically.

The Atwater Member of the Crooked Creek Formation in southwestern Kansas is correlated with the Sappa Formation in Nebraska. The Borchers local fauna is from silt that overlies the Huckleberry Ridge ash bed. It represents warm climatic conditions and contains a large variety of mammals, snakes, and the tortoise *Geochelone.* Martin and Schultz (1985, p. 177) regard this to be "one of the earliest known Irvingtonian faunas." In one Meade County site, the "upper Borchers ash" is 6.5 m above the type Borchers (Huckleberry Ridge) ash. This younger ash has been fission-track dated at 1.2 Ma, but is compositionally different from the Mesa Falls (Coleridge) ash; the presence of biotite phenocrysts suggests a Jemez Mountains, New Mexico, source (Boellstorff, 1976).

Middle Pleistocene stratigraphy

The Quaternary patterns of sedimentation are somewhat more complex than those Lugn (1935) envisioned, in which each glaciation was represented by a gravel-sand unit overlain by an interglacial silt-clay unit. In both Nebraska and Kansas, major streams were blocked repeatedly by glacial ice, and fine-grained alluvial and lacustrine sediments accumulated in the proglacial lakes.

Lugn (1934, p. 344, and 1935) named the fine-grained sediment overlying his Holdredge gravel the *Fullerton Formation* (Fig. 16) and considered it to be an "interglacial" sediment. Reed and Dreeszen (1965, p. 25, 26) believed that the relationships of the unit were unclear in the type section, so they described both an outcrop and a test hole as substitute type sections that they thought illustrated the concept better. The Fullerton Formation in the outcrop substitute type section is ripple-cross-laminated silt that was deposited as fallout from a density flow in a proglacial lake (Stanley, 1974). It is overlain, with no evidence of a weathered zone, by the youngest till in east-central Nebraska. Similar sediments in northeastern Kansas constitute the Atchison Formation, which is associated with the Lower Kansas till of normal polarity (Aber and others, 1988). The sediments in the test hole substitute type section of the Fullerton represent accumulation in an ice-dammed valley, and may be part of a different depositional sequence from the materials in the outcrop substitute type section.

In east-central Nebraska, a molluscan faunal assemblage (Wayne, 1981) from proglacial fine-grained alluvium interbedded with lacustrine clayey silt provisionally identified as Fullerton Formation contains mostly species that now live from southern South Dakota to central Minnesota, but it also contains a few species that now inhabit warmer areas. Plant remains include spruce needles. Glacial ice probably was at least 100 km away, so the faunule indicates only that conditions were cooler and moister than the present at that distance from the ice margin. Microtine rodent remains support the climatic interpration but suggest a Rancholabrean age (Michael Voorhees, letter, 19 December 1986).

Frye and Leonard (1952, p. 52, 59) used the names Holdrege and Fullerton in their conceptual sense to refer to coarse and fine-grained phases of the sediments they identified as Blanco Formation. Fission-track dating of volcanic ash, however, demonstrated that the Blanco Formation of north Texas is late Pliocene and transitional to earliest Pleistocene in age (Holliday, 1988). Although the name Fullerton Formation has become well established in eastern Nebraska on the basis of observations in the substitute type section, this stratigraphic unit should be renamed (Goodwin, 1986).

Lugn (1934, 1935) chose the name *Grand Island Formation* (Fig. 16) for one of the more widespread sedimentary units in east-central Nebraska. It is a thick unit of cross-bedded gravelly sand that underlies loess and other silt throughout most of the region between the Platte and Republican Rivers east of Lexington. In the Blue River basin it is ubiquitous.

The Grand Island gravel is dominated by clasts derived from mountains in southeastern Wyoming and carried eastward by the North Platte River. The most characteristic clasts are pebbles of Sherman granite and anorthosite, but varicolored chert and quartzite are common, and occasionally volcanic clasts are present. Lugn (1935) thought the Grand Island gravel to have been deposited during "the Kansan Glaciation" by sediment from the Rockies that became mixed, in eastern Nebraska, with continental glacial meltwater sediment. Clasts characteristic of Laurentide glacial outwash are notably absent, however. Because the regional slope is eastward toward the glaciated region, Laurentide glacial outwash was deposited only immediately adjacent to the former ice margins.

The type section of the Grand Island is an overgrown gravel pit; the unit is exposed, however, in many places, particularly along stream cuts and in gravel pits south and east of Grand Island. The highwall of an inactive gravel pit about 9 km east of York exposes cross-bedded gravel of Platte River provenance beneath 2 m of finer grained, light gray alluvial overbank sediment with a weakly developed paleosol, which in turn, is overlain by brown (7.5 YR 5/6) alluvium capped with loess.

Several aspects of the Grand Island gravel suggest that part of it, including the type section, may be younger than Lugn believed it to be. The land surface underlain by these gravels is an undissected loessal plain, and the surface of the alluvial plain beneath the loess is little dissected. The fine-grained gray to olive sediment that covers the gravel sheet contains shells of aquatic snails in some places, and the paleosol that has formed in it is not strongly developed. That paleosol in turn is buried beneath brown alluvial sediments that are overlain by loess.

Gray, fine-grained sediments overlie cross-bedded gravelly sand in many exposures of the Grand Island Formation in southeast-central Nebraska and northeastern Kansas. They rarely are more than 2 to 3 m thick, may be calcareous but generally are not, range in lithology from sandy loam to silty clay, contain fossil snail shells in some exposures, and generally have a weakly to moderately developed paleosol at the top. This unit, Lugn (1935, p. 119–127) named the Upland Formation, and he believed it to have been deposited in a nonglacial environment during "the Yarmouth interglaciation." When Condra and Reed (1950; also see Frye and others, 1948) renamed it the Sappa Formation, they assigned it a "late Kansas" age, because in the type section of the Sappa a "Pearlette" ash bed overlies a sequence of fine-grained sediments.

Reed and Dreeszen (1965, p. 34) tried to identify exposures of extraglacial fluvial and lacustrine sediments that could be correlated with tills in subsurface studies; the stratigraphic units they erected, with some redefinition, remain useful. One of these is the *Walnut Creek Formation* (Reed and Dreeszen, 1965, p. 32–33, 55), which consists of a fining-upward alluvial sequence that contains a thin ash bed, the Lava Creek B (Pearlette type O). Although in the original description the ash and overlying silty fine sand were included in the Sappa, they are younger than the Sappa Formation as redefined. The Walnut Creek Formation is nonglacial alluvium of middle Pleistocene age (Fig. 16).

The fine-grained alluvial and lacustrine sediments that cap the Grand Island at its type section, as well as throughout the plains south of the Platte River and east of Kearney, are more complex than are the sediments exposed at Lugn's (1935) type section of the Upland Formation. Two units described by Reed and Dreeszen (1965), the *Grafton* and *Beaver Creek Formations,* encompass the Upland. The Grafton Formation is alluvial sand and overlying clayey silt, at the top of which is a paleosol with moderately strong development. The Beaver Creek Formation is bedded silt and silty clay nearly everywhere and in some exposures contains shallow-water gastropods that can tolerate wide climatic conditions. It may be Illinoian in age (Fig. 16).

The *Loveland Loess* (Shimek, 1909, 1910) is a pale brown silt that underlies the late Wisconsin Peoria Loess and overlies a "Kansas" till along the bluffs of the Missouri River at Loveland, Iowa. Since that time, the term Loveland has undergone considerable modification in meaning. Today, the name is applied to a reddish brown to brown fine-grained sediment that lies beneath the Peoria Loess from the Ohio Valley through south-central Nebraska and northern Kansas. In its type section (Daniels and Handy, 1959) and in many other localities in eastern Nebraska and Iowa, the Loveland Loess rests on a paleosol developed on a till classically regarded as "Kansan," and in most localities the Loveland bears a distinctive paleosol, the Sangamon soil, beneath Wisconsin-age sediments. Leighton and Willman (1950) stated that it is loess of Illinoian age, and it has been so correlated since then (Frye and Leonard, 1952; Wayne, 1963; Reed and Dreeszen, 1965; Ruhe, 1969). Commonly the Loveland contains calcium carbonate concretions but only rarely fossil snail shells. Originally, some water-laid and mass-wasting sediments were included in the unit. Lugn (1935) regarded both loessal and fluvial sediments to be part of the Loveland in Nebraska, although only loess is present at the type section. Condra and others (1947, p. 24), however, separated the sand and gravel from finer grained sediments and named them the Crete Formation. Currently, the name Crete is applied to sheet-like gravel in the Blue River basin. It probably includes the gravel in the type section of the Grand Island Formation, as well as other sand and gravel that underlies the fine-grained brown sediment.

In some exposures the finer grained deposits overlying the Crete sand are sandy and clayey silt that accumulated as overbank sediment rather than as a loess, yet these commonly are called, uncritically, Loveland Loess, and only granular material at the base of the unit is regarded as Crete. In outcrops, granular channel deposits overlain by finer flood-plain sediments are readily distinguished, but in subsurface samples such distinctions are difficult.

Available data suggest that the names Grand Island and Crete may have been applied to the same body of sediment; one of these names should be formally abandoned. Because the term Grand Island Formation has priority, we use it rather than Crete in this report (Fig. 16).

In many sections south of the Platte River, the Loveland Formation contains two strongly developed paleosols (Fredlund and others, 1985; Morrison, 1987). Schultz and Martin (1970) applied the name Loveland to the entire sequence exposed at the Buzzards Roost exposures near Lexington, Nebraska. Following Reed and Dreeszen (1965, p. 62), they used the names Grafton and Beaver Creek for the two lower units of the Loveland and proposed a new name, Gothenburg Member, for the upper part. They also named the first major buried soil beneath the Sangamon soil the Buzzards Roost paleosol, and a second one lower in the section the Ingham paleosol. The normally magnetized, 620-ka, Lava Creek B ash (Kukla, 1978) separates the base of the Loveland from underlying sediments. At the old Eustis silica mine and in at least one other exposure in east-central Nebraska, brown Loveland silt grades downward into an ash bed. Thus the entire Loveland Formation as presently recognized in Nebraska accumulated between 0.62 Ma and the beginning of Sangamon time, about 130,000 years ago (Fig. 16).

Although Schultz and Martin (1970) and Fredlund and others (1985) treated the paleosols they recognized in the Loveland as "Illinoian interstadial soils," at least two, the Buzzards Roost and Ingram paleosols, are more argillic than the younger Sangamon profile, have a thicker B2t horizon with more strongly developed structure, visible clay skins on the columnar peds, and a weaker calcic horizon. These features suggest a long period of stability, perhaps even longer than that involved in formation of the Sangamon profile at the top of the Loveland.

In addition to the weathered zones they recognized within the Loveland Formation in the Eustis exposure, Fredlund and others (1985) identified grass phytoliths in all samples collected through the section. In nearly all samples, the relative abundance of different groups of grass phytoliths suggested that the loess accumulated in a climate similar to but somewhat drier than at present. The character of the two strongly developed paleosols within the unit, though, indicates relatively moist conditions while they formed.

If the Loveland Formation began to accumulate soon after deposition of the Lava Creek B ash bed and continued, with breaks, until development of the Sangamon Soil at its surface, a reexamination of the correlation of the formation is forced. Both Schultz and Martin (1970) and Fredlund and others (1985) have stated that it was deposited during the Illinoian Glaciation, as have most other geologists. If this correlation should be correct, then the Illinoian Glaciation would have begun about 620 ka and lasted for half a million years. Although this interpretation was adopted by Beard and others (1982), Zakrzewski (1975), and Kurtén and Anderson (1980), we regard only isotope stages 6 and 8 to be correlative with the Illinoian glaciation, as do Richmond and Fullerton (1986). If the three loess bodies within the Loveland in south-central Nebraska are related to glaciations, they may be a midcontinent record of the cold periods between oxygen isotope stages 14 and 6 of Shackleton and Opdyke (1973).

Nearly everywhere the top of the Loveland Formation is marked by a distinctive and well-recognized paleosol, the *Sangamon soil.* Follmer (1978) reviewed the history of the Sangamon Soil in its type area from its first recognition by Leverett (1899, p. 125–130), through development of concepts of horizonation, to the modern usage of the term (Willman and Frye, 1970). Where the Sangamon soil is fully developed and has not been truncated by erosion, it differs little from the soil in its type area in Illinois.

Late Pleistocene stratigraphy

During the Wisconsin glaciation, eolian sediments accumulated beyond the limits of the ice sheet, and fluvial sedimentation along streams that carried large volumes of glacial meltwater resulted in base-level changes along all of their tributaries. Three successive bodies of late Quaternary loess have been recognized in the northern part of the High Plains and overlying the older glacial deposits to the east. The origin and stratigraphy of these loess units were argued extensively by Lugn (1962, 1968), Reed (1968), Ruhe (1968, 1970, 1976), Frye and Leonard (1952), and Willman and Frye (1970), and controversy still surrounds some aspects of these sediments.

The basal loess. A dark-colored zone of noncalcareous silt or clayey silt overlies the Sangamon soil from the lower Wabash Valley in Indiana to west-central Nebraska and Kansas. Once considered to be the main part of the Sangamon soil, it was for many years referred to as the "*Citellus* zone" in Nebraska because it yielded vertebrate fossils, among them a ground squirrel, *Citellus.* Thorpe and others (1951) suggested that this unit represented slow accretion of loess, attenuating upward the A horizon of the Sangamon soil. Fredlund and others (1985) noted a significant increase in moisture-loving grasses while it accumulated.

Rarely more than a meter thick and generally thinner, this basal zone, recognized in Nebraska and Kansas, was named the *Gilman Canyon Formation* (Reed and Dreeszen, 1965, p. 42). Dreeszen (1970) reported radiocarbon ages on organic material near the base of the unit of 34,900 B.P. (I-2190) and 23,000 B.P. (I-2191) from the top. Additional dates from the base and top of the bed in sediments beneath a shallow ephemeral pond near York, Nebraska (Krueger, 1986) are 28,350 ± 690 B.P. (Beta-12274) and 20,940 ± 240 B.P. (Beta-12273). Farther west, however, along the Platte valley, lenses containing organic matter interbedded with eolian sediments have yielded ages as great as 34,880 ± 1060 B.P. (Beta-23495; May and Souders, 1988). Samples from a test hole near the exposure in southern Nebraska illustrated by Miller and others (1964, p. 44) date the top and middle of the unit at 19,770 ± 590 (Beta 33940) and 29,870 ± 1650 B.P. (Beta 33941; Souders and Kuzila (1990).

Todd Valley Sand. Sand in the upper part of an abandoned trough once occupied by the Platte River in eastern Nebraska, called Todd Valley, was named the *Todd Valley Sand* by Lugn (1934, 1935). The Todd Valley Sand underlies the Peoria Loess with no intervening buried soil; it is Wisconsin in age. Although Todd Valley surely was eroded originally as an ice marginal drainage line during an early glaciation, the Platte River abandoned it and now flows in the trench with the Elkhorn River. Wayne (1985, 1987, 1988) suggested that continental ice diverted part of the Missouri River into the Elkhorn valley during the late Wisconsin glaciation, raising base level enough to divert the Platte across Todd Valley. The Todd Valley sand accumulated at that time; after the Platte abandoned the Todd Valley route, the sand was blanketed by Peoria loess. Correlative sediments underlie Platte and Loup River terraces upstream from Fremont to beyond Kearney.

Peoria Loess. The late Wisconsin *Peoria Loess* veneers older sediments on much of the northeastern High Plains. It is coarsest and thickest adjacent to its source areas, which include the Missouri and Platte Rivers (that carried outwash) and the Nebraska Sandhills Region (Lugn, 1962). The Peoria Loess is recognized from Ohio to eastern Colorado, and has been studied extensively. Lugn (1962, 1968) proposed that both it and the dune sand of the Nebraska Sandhills Region were derived from the Ogallala Group, which underlies the dune region but is largely buried beneath fluvial sediments; he regarded the sand and loess to be coarse and fine-grained facies of the same stratigraphic unit. He also suggested (Lugn, 1968) that the eolian sediments were generated and deposited during times of extended drought. However, Frye and Leonard (1952), Reed (1968), and Ruhe (1969, p. 28–54) pointed out that the Peoria Loess was deposited during the maximum extent of the late Wisconsin glaciation and that it thins systematically downwind from aggrading glacial meltwater rivers. They concluded that it was derived almost wholly from dust blown from flood plains of those rivers. It also thins southeastward from the Nebraska Sandhills.

The Peoria Loess was deposited on topography that varied from a flat plain to rolling, even hilly, terrain. Ruhe and others (1967, p. 72–76) demonstrated that considerable erosion took place after deposition of the lower Wisconsin loess and before deposition of the Peoria Loess in western Iowa. In eastern Nebraska, the unconformity at the base of the Peoria cuts across both the Gilman Canyon Formation and the underlying Loveland Formation in some places. In much of the "Loess Hills" region of east-central Nebraska, present topography has a relief as great as 50 m. The thickness of the Peoria Loess, where it has not been stripped by erosion, varies from 1 or 2 m to more than 25 m. Some of the relief is caused by Peoria draped over the underlying erosional topography. A distinctive and unexplained topographic feature in this region is "parks," which are large, closed, flat-bottomed depressions.

In many exposures, the Peoria Loess contains shells of land snails that lived in vegetation growing on the surfaces where it accumulated (Leonard, 1952; Frankel, 1957). The shells are relatively abundant at sites that were moist and probably sheltered at the time of deposition, but are much more sparse at drier upland sites. Snails recovered indicate cool, moist, relatively open conditions.

Brady soil and Bignell loess. In central and western Nebraska and Kansas, a distinctive grayish brown zone separates the Peoria from a thin overlying loess cap. Named the *Brady soil* (Schultz and Stout, 1948), it has a dark A horizon (10YR 4/2 dry) 20 to 30 cm thick and either no B horizon or a B with very weak blocky structure, often with filaments of calcium carbonate (Ruhe, 1970). This geosol can be recognized only where it is buried beneath the thin *Bignell loess* (Schultz and Stout, 1945), the thickness of which at their type section is about 3 m, including the post-Bignell soil profile developed in the top part. At that section, both the Brady and post-Bignell soils seem to show about equal development.

The age of the Bignell and the time represented by the Brady soil have not been resolved. The Brady was thought originally to have developed during a late Wisconsin interstade. Radiocarbon ages 9,160 ± 250 B.P. (W-234); and 9,850 ± 300 B.P. (W-1767) from the top of the soil at the type section suggest that the Bignell loess was deposited during the latest part of the Wisconsin glaciation. Ages from shell carbon of 12,550 ± 400 B.P. (W-231) and 12,700 ± 300 B.P. (W-233) suggest greater antiquity, but shell ages commonly are older than wood ages from the same beds. Organic matter from the Brady soil sampled in a test hole near the Republican River recently was dated at 10,130 ± 140 B.P. (Beta 33939; Souders and Kuzila, 1990). Thus, Bignell loess deposition began about 9,000 to 10,000 years ago. Based on its development, the Brady soil may have formed during 2,000 to 3,000 years, under conditions generally similar to those of today. Snail shells are rare in the Bignell loess, but Leonard (1952) reported essentially a modern faunule from sites in Kansas.

The Sandhills Region of Nebraska

The Nebraska Sandhills Region is about 50,000 km^2 of stabilized dunes (Figs. 1, 17), the largest such region in the Western Hemisphere (Smith, 1965). A fully adequate explanation of the origin of this large region of dunes has not yet been offered, although three hypotheses have been proposed. The Loup, the Niobrara, and the Platte Rivers drain the region and have eroded trenches that expose the sediments beneath the dunes. They have few tributaries, so exposures are limited to bluffs along those rivers. The exposed sediments, the fossils they have yielded, dune morphology, and radiocarbon ages from organic materials beneath dune sand are the bases for hypotheses about their origin.

The Ogallala Group is exposed in a few places, but generally it is overlain by younger, mostly fluviatile sediments. The Broadwater Formation, which contains Blancan vertebrate remains, is exposed along the Middle Loup and Dismal Rivers and represents sediments carried by the Platte River before it was diverted to its present course. It is overlain by silts and diatomaceous sediments that have yielded late Blancan to Irvingtonian vertebrate fauna (Martin and Schultz, 1985; Bennett, 1979), mollusks, and diatoms (Maroney, 1978). These sediments represent ponding along the course of the river, with diatom spectra suggesting shallow, alkaline to brackish, warm lakes, ponds, and sloughs.

Lugn (1935, p. 161, and 1968, p. 146) proposed that both the dunes, which he named the *Sandhills Formation,* and the Peoria Loess had been deflated from the Ogallala Group when desert conditions existed. Reed (1968, p. 25) disagreed, primarily on the basis of the mollusks in the Peoria Loess. He suggested that unconsolidated subdunal alluvium was a more likely source for the sand, and that most of the Peoria Loess was blown from the aggraded flood plain of the Platte River.

H.T.U. Smith (1965, 1968) noted that most of the modern analogs of the dune forms seen in the Nebraska Sandhills Region are present in modern deserts, such as the Grand Erg Oriental in North Africa. He pointed out that the Nebraska dunes show evidence of several phases of activity and stability and proposed a chronology of three major periods of dune building. His "first series," which he thought might be early Wisconsin in age, consists of large transverse barchanoid dunal masses that form virtually continuous ridges, particularly in the northern and western parts of the region. Smaller elongate, parallel sand ridges, most of which are in the southeastern part of the region, make up the "second series," which he inferred to be late Wisconsin in age. The "third series" is represented by many blowouts on the dunes of the older series and by smaller dunes in the eastern and southern parts of the region. Smith (1965, p. 504) assigned them to dryer times during the Holocene. His age assignments were based on "Illinoian" vertebrate remains from sediments beneath dune sand of the "first series" dunes and Peoria loess beneath dunes of the younger series.

Stanley and Wayne (1972) pointed out that a predecessor of the North Platte River was diverted southeastward during early Pleistocene time and suggested that flood-plain sediments of that river may have served as a major source of sand for an early phase of dune development, later reworked repeatedly.

A third hypothesis regarding the age of the dunes was developed after radiocarbon dates became available for subdunal beds

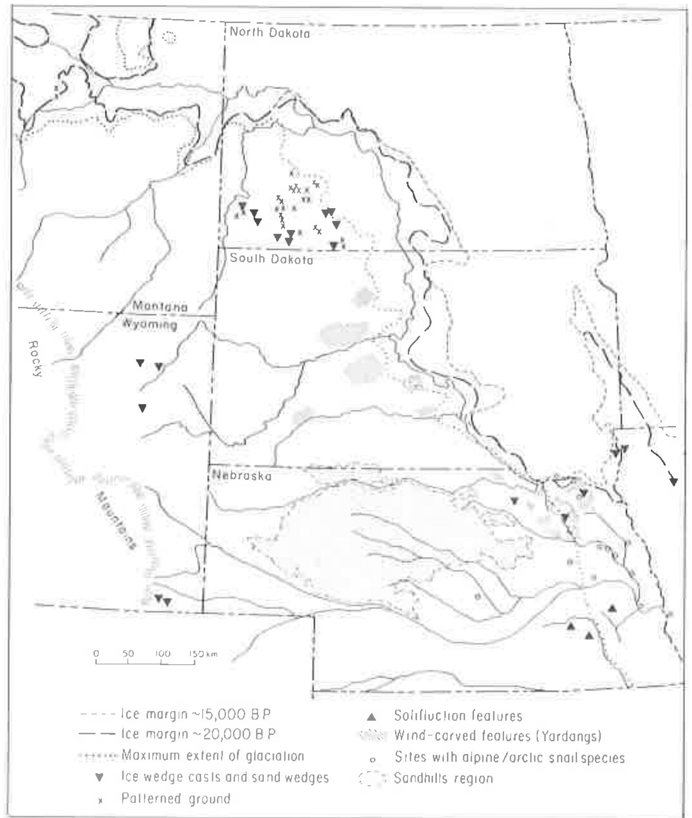

Figure 17. Distribution of relict periglacial features in the Northern Great Plains on which paleoclimatic interpretations are based.

of dark-colored alluvial sand exposed along the Middle Loup, Dismal, North Loup, and Snake Rivers (Maroney, 1978; Ahlbrandt and others, 1983, p. 393). All of the samples were Holocene in age and fall into three groups: early (10 to 7.2 ka), middle (5.0 to 3.4 ka), and late (1.6 ka). Ahlbrandt and others (1983) concluded that the dunes have formed during the past 10,000 years, as they did in other dated dune fields in northern Colorado and Wyoming. Similar conclusions, based on soil-profile development (Muhs, 1985) and dated mollusk-rich pond sediments (Madole, 1986b), were reached for the Wray dune field along the South Platte River.

Wright and others (1985) noted that pollen in sediments between dunes in the northern Sandhills shows the presence of spruce trees nearby 12,000 to 13,000 years ago and questioned the conclusion that the dune field is wholly Holocene in age.

Each of the hypothses answers some questions, yet leaves others unanswered. Both internal structures (Ahlbrandt and Fryberger, 1980) and external form (Warren, 1976) indicate that the winds that produced the dunes came dominantly from the northwest. Wayne (1990) noted evidence of intense northwest-southeast winds along the Wisconsin ice margin in South Dakota and northeastern Nebraska and suggested that significant dune activity could have taken place at that time. The Holocene dates were obtained from exposures along the rivers. Elsewhere in the region, test holes have penetrated fluvial silt and sand beneath the dune sand; one hole in the south-central part of the Sandhills Region encountered, at 61 m, a peat dated at 32,130 +1280/−1520 B.P. (Dic-2513) between Broadwater-type gravel and overlying alluvial sand (Swinehart and Diffendal, 1989). Only the top 6 m was dune sand. A 1-m-thick peat bed within alluvial sand beneath dunes exposed along Birchwood Creek northwest of North Platte yielded an age of 38,010 ± 1,260 B.P. (Beta 16606; Swinehart, 1988).

Southwestern Kansas and northwestern Oklahoma

The late Tertiary and Quaternary stratigraphic units in southwestern Kansas and the Oklahoma panhandle are different from those of the remainder of Kansas and southern Nebraska (Hibbard, 1958). Bayne (1976) outlined the history of the development of terminology used in this region. He points out that, through much of the period of study of the deposits of the area, the "Pearlette ash" was assumed to be a single key bed in the stratigraphic sequence. The deposits were fitted into the classical scheme of four glaciations and three interglaciations in the Midwest, based on the thermal and moisture implications of the faunal remains they contained. Faunal assemblages containing several taxa that require cool summers but none that required mild winters were interpreted to be "glacial" faunas; mild winter taxa were considered to be interglacial assemblages (Bayne, 1976). They were placed in a relative stratigraphic framework, based on the degree of evolutionary development of the faunal elements that were known to change in a particular direction through time, and on the extinction or first appearance of certain taxa.

Sediment accumulation in the region was chiefly fluvial, by the Cimarron River and its tributaries. Subsidence on the Crooked Creek fault resulted in deposition of considerable thicknesses of sediment, and development of sinkholes over salt beds has resulted in further complications in the stratigraphy.

Considerable disagreement developed over the years as to the identification of some of the stratigraphic units in southwestern Kansas. In particular, the sediments called Blanco Formation and Meade Formation by Frye and Leonard (1952, p. 58–68, 84–104) were renamed the *Rexroad, Ballard,* and *Crooked Creek Formations* by Hibbard (1958).

Each of these formations consists of fluvial gravel, sand, and silt. Caliches (K horizons), erosional unconformities, and volcanic-ash beds provide the principal guides to recognition in the field, but all of these units are so similar in lithology that differentiation must be supported by identification of fossil vertebrates and mollusks. The most recent correlations, based on studies by Bayne (1976), Miller (1976), and Zakrzewski (1975), are summarized on Figure 16.

Paleoclimates in the Northern Great Plains

W. J. Wayne

Most of the evidence that has been used to evaluate and understand the climatic fluctuations of the Pleistocene is based on vertebrate and invertebrate fossil assemblages recovered from the sediments. Spruce wood has been recovered from the basal till in Nebraska and Kansas. In the Northern Great Plains, pollen records are scarce (Wright, 1970; Baker and Waln, 1985) and are limited to the late Pleistocene and Holocene, although ice-marginal conditions are poorly known because faunal assemblages from that environment have not been preserved. In the central Great Plains, during glaciations, summers evidently were cooler and more humid than they were during interglaciations, and species that now live farther north and east were able to coexist with some of the more southerly species. During warmer and drier interglaciations, these taxa could not survive in those areas (Miller, 1966, 1975; Wayne, 1981).

Polygonal patterns in southwestern North Dakota thought to be of thermal contraction origin and recognized from air-photo analysis include some that extend to the margins of some of the buttes and probably are pre-Wisconsin in age. Most of the patterned ground in the area undoubtedly formed during the last glaciation, however (Bluemle and Clayton, 1986). Residual ventifacts common in the Flint Hills of east-central Kansas may relate to strong wind erosion beyond one of the mid-Pleistocene ice sheets.

During at least part of Sangamon time, the eastern part of the Northern Great Plains evidently was more moist than it is today, and it may also have been warmer. The Sangamon soil in that region has the charateristics of a forest soil and changes little between the type area in Illinois and eastern Nebraska (Ruhe, 1974). Conditions became drier toward the west, though, perhaps even more abruptly than they do today. Kapp (1965) reported that no pollen was preserved as a record of mid-Sangamon vegetation in southwestern Kansas and that conditions there probably were dry. The Jinglebob local fauna in southwestern Kansas indicates a return to more moist conditions; vertebrate and molluscan remains suggest cooler summers with winters no more severe than those of today, with significantly greater precipitation (Van der Schalie, 1953; Hibbard, 1955; Hibbard and Taylor, 1960). Both spruce and pine pollen are associated with the Jinglebob fauna, so it may represent an early Wisconsin accumulation (Kapp, 1965, 1970).

Far more information is available to interpret climatic conditions beyond the ice margin during the late-Wisconsin glaciation than during all of the pre-Wisconsin Pleistocene in the northern Great Plains. Not only are biotic materials more abundant, but many of the geomorphic features that resulted from severe climatic conditions near the ice margin have not been destroyed by surficial processes.

Both vertebrate and molluscan remains from ^{14}C dated deposits in Kansas and Oklahoma (Schultz, 1969; Taylor and Hibbard, 1955; Miller, 1976) indicate that cool conditions existed during the main part of late Wisconsin time, 20,000 to 12,000 B.P. Summer temperatures may have been comparable to those in North Dakota today, with winter temperatures no colder than those today in northern Kansas. Conditions became more severe northward, however. In deposits along the Republican River, Corner (1977) identified remains of caribou and both woodland and tundra muskox, and Johnson and others (1986) noted both fauna and flora indicating considerably moister and cooler conditions than present associated with 14,000-year-old spruce wood. Boreal steppe species dominate the Wisconsin faunal material (both microvertebrates and mollusks) recovered from central and northern Nebraska (Voorhies and Corner, 1985). The only studied late Pleistocene locality in western South Dakota, the Lange/Ferguson Clovis kill site (10,670 ± 300 B.P.), yielded a vertebrate fauna that indicates greater effective moisture than today in the Badlands, which supported a more varied fauna and probably a denser and more diverse vegetational mosaic (Martin, 1987).

Ice-wedge casts and extensive areas of patterned ground in south-western North Dakota (Clayton and Bailey, 1970; Bluemle and Clayton, 1986) indicate a permafrost belt that was at least 125 km wide. Sand wedges and ice-wedge casts were reported from southeastern South Dakota (Tipton and Steece, 1965), and sand-filled thermal-contraction wedges have been found in northeastern Nebraska (Wayne, 1990). Ice-wedge casts are widespread in eastern Wyoming (Mears, 1981, 1987) and indicate that permafrost existed there, with temperatures of −5°C or colder. These features (Fig. 17) provide evidence that the mean annual temperatures at the times they formed probably were at least 13°C lower than the present in North Dakota and 20°C lower in northeastern Nebraska. The most reasonable time for the permafrost to have existed is about 20 ka, contemporaneous with formation of similar features in Illinois (Johnson, 1986, 1990) and Indiana (Wayne, 1967).

Strongly developed northwest-southeast linear patterns that characterize the zone just southwest of the Wisconsin glacial limit in northeastern Nebraska and parts of central South Dakota (Crandall, 1958) may be relict yardangs (Wayne, 1990). Ventifacts, common along the surface beneath the Peoria Loess and associated with both sand-filled wedges and the linear topographic patterns provide further evidence of intense wind activity. Strong winds created linear deflation basins ("blowouts") along the White River in south-central South Dakota, and the "rainbasin depressions" of southeast-central Nebraska resulted from deflation about 20 ka (Krueger, 1986).

From 18,000 to 13,000 B.P., spruce trees were present, declining in abundance in northeastern Kansas (Caspall, 1972); soon after, prairie grassland replaced them in the western part of the province (Baker and Waln, 1985). The modern oak-hickory-prairie environment developed about 5,000 B.P. (Grüger, 1973). Farther north, little is known about the late-glacial conditions in eastern Nebraska, but from 12,500 to 11,000 B.P. spruce dominated in western Iowa (Van Zant, 1979) and was present in the northern Sand Hills, central South Dakota, and central North Dakota (Wright, 1970).

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