Paleozoic Formations in the Wind River Basin Wyoming

GEOLOGICAL SURVEY PROFESSIONAL PAPER 495-B

Prepared in cooperation with the Geological Survey of Wyoming and the Department of Geology of the University of Wyoming as part of a program of the Department of the Interior for development of the Missouri River basin

Paleozoic Formations in the Wind River Basin Wyoming

By W. R. KEEFER *and* J. A. VAN LIEU

GEOLOGY OF THE WIND RIVER BASIN, CENTRAL WYOMING

GEOLOGICAL SURVEY PROFESSIONAL PAPER 495-B

Prepared in cooperation with the Geological Survey of Wyoming and the Department of Geology of the University of Wyoming as part of a program of the Department of the Interior for development of the Missouri River basin

UNITED STATES GOVERNMENT PRINTING OFFICE, WASHINGTON : 1966

UNITED STATES DEPARTMENT OF THE INTERIOR

STEWART L. UDALL, *Secretary*

GEOLOGICAL SURVEY

William T. Pecora, *Director*

For sale by the Superintendent of Documents, Government Printing Office Washington, D.C., 20402

CONTENTS

ILLUSTRATIONS

[Plates are in pocket]

TABLES

PALEOZOIC FORMATIONS IN THE WIND RIVER BASIN, WYOMING

By W. R. Keefer and J. A. Van Lieu

ABSTRACT

The Wind River Basin, occupying an area of 8,500 square miles in central Wyoming, is an extensive structural depression bounded on the south by the Granite Mountains, on the west by the Wind River Range, on the north by the Washakie Range and Owl Creek and Bighorn Mountains, and on the east by the Casper arch. These features were formed during Laramide deformation in Late Cretaceous and early Tertiary times. Paleozoic rocks underlie the basin and crop out along all the mountain flanks except the east.

During Paleozoic time central Wyoming was part of the interior stable shelf region that lay along the east side of the Cordilleran geosyncline. Rocks representing all the Paleozoic systems, except possibly the Silurian, were deposited across the present site of the Wind River Basin during repeated transgressions and regressions of epicontinental seas. Because greatest sedimentary accumulation was toward the west, the Paleozoic section is thicker and more complete in the western part of the basin than in the eastern part. Marked changes in the thickness and lithology of nearly all formations take place from west to east, and some units disappear eastward owing to truncation or nondeposition. Although most of the strata were deposited under normal marine conditions, the pattern of sedimentation was often influenced locally by slight but significant fluctuations in the base level of deposition caused by changes in sea level or by tectonic movements.

Paleozoic formations in the Wind River Basin include the Flathead Sandstone, the Gros Ventre Formation, and the Gallatin Limestone of Cambrian age; the Bighorn Dolomite of Ordovician age; the Darby Formation of Devonian and earliest Mississippian age; the Madison Limestone of Mississippian age; the Amsden Formation and the Tensleep Sandstone of Pennsylvanian age; the Park City Formation and its equivalents of Permian age; and the Goose Egg Formation of Permian and Triassic age. Subdivision into formations and members of formations is based primarily on the classification and nomenclature that have been established for the sequence in the northern part of the Wind River Range.

The Cambrian rocks are a well-defined transgressive sequence containing a basal quartzitic sandstone unit (Flathead), a middle predominantly shale unit (Gros Ventre), and an upper limestone unit (Gallatin). Thicknesses decrease from 1,200 feet in the northwestern part of the Wind River Basin to 50 feet at the southeastern corner. The sedimentary pattern is one of continuous transgression, beginning in Middle Cambrian time and reaching a maximum in Late Cambrian time, but the strata show that the transgression was interrupted many times by minor retreats of the sea.

The Bighorn Dolomite is divided into three members: the Lander Sandstone Member which occurs locally at the base, an unnamed massive cliff-forming dolomite in the middle, and the Leigh Dolomite Member at the top. The formation, 300 feet thick in the northwestern part of the basin, thins to a wedge edge in the central part. Most of the Bighorn is considered to be Late Ordovician in age, but the Lander Sandstone Member may be Middle Ordovician.

Silurian rocks have not been recognized in central Wyoming. If ever deposited in the region, they were removed by erosion before succeeding strata were laid down.

The Darby Formation of Devonian and earliest Mississippian age has the most limited distribution of any of the Paleozoic units in the area, being confined to the western third of the basin. The sequence, which has a maximum thickness of 200 feet, consists chiefly of dolomite in the lower part and of interbedded siltstone, shale, sandstone, and carbonate rocks in the upper. Locally at the top is a dark shale which is part of a widespread unit of latest Devonian and earliest Mississippian age that rests uncomformably on older Devonian rocks.

Mississippian strata are 700 feet thick in the western part of the Wind River Basin, but they thin to 175 feet in the southeastern part. These rocks, predominantly resistant massive to thin-bedded crystalline limestone, are all assigned to the Madison Limestone. Brecciated and cavernous zones and intraformational unconformities are present, particularly in the upper part. The formation rests on the Darby Formation with apparent conformity toward the west, but eastward it truncates successively older formations and rests on the Flathead Sandstone in the southeast corner of the basin. The bulk of the Madison is Early Mississippian, but in the northern and central Wind River Range and adjacent Washakie Range it includes about 100 feet of beds that are Late Mississippian.

The Amsden Formation is divided into a lower sandstone unit, called the Darwin Sandstone Member, and an upper unnamed unit of interbedded dolomite, shale, sandstone, and limestone. Thicknesses range from 355 feet in the northwestern part of the Wind River Basin to zero at the southeast corner. The formation is largely Early to Middle Pennsylvanian, but the Darwin Sandstone Member and immediately overlying beds are Late Mississippian. The Tensleep Sandstone, a uniformly massive to crossbedded cliff-forming sandstone, containing a few thin beds of chert and limestone, is one of the most conspicuous Paleozoic units. Thickness ranges from 200 feet to more than 600 feet. Most of the formation is Middle Pennsylvanian. At the southeast corner of the basin all Pennsylvanian rocks (mostly Upper Pennsylvanian), as well as some Lower Permian rocks, are included in the Casper Formation.

Permian rocks are interbedded chert, limestone, dolomite, siltstone, sandstone, and phosphorite in the western two-thirds of the Wind River Basin, and red beds and gypsum, with thin beds of limestone, in the eastern third. The chert and carbonate facies, 150–350 feet thick, is referred to the Park City Formation and equivalents, and the red-bed facies, 355–380 feet thick, to the Goose Egg Formation. The Park City is largely early Late Permian, although the basal few feet, in some places separated from the overlying beds by an unconformity, is Early Permian. The lower 270–300 feet of the Goose Egg is equivalent to the Park City, but the remainder intertongues with the Dinwoody

Formation of Early Triassic age. Uppermost Permian rocks have not been recognized in the basin area.

The Wind River Basin was remarkably stable throughout the Paleozoic Era. Although considerable time elapsed between deposition of successive formations, there are no recognizable angular discordances within the entire sequence of Paleozoic rocks. Tectonic movements were limited to broad warping along trends which show little direct relation to structural trends developed during Laramide deformation. One exception is along the southeast margin of the basin where an area coinciding roughly with the present-day Laramie Mountains and parts of the Granite Mountains was positive intermittently during much of Paleozoic time.

Paleozoic rocks, principally the Park City Formation and Tensleep Sandstone, are among the most important oil and gas reservoirs in the Wind River Basin. Much of the production has been from anticlinal traps along the basin margins, but conditions favoring stratigraphic entrapment also exist. The Park City contains extensive, though undeveloped, deposits of low-grade phosphate, especially in the western and northwestern parts of the basin. Raw materials for construction use are widespread. Pisolitic iron-bearing red shale occurs in the Amsden Formation, and small deposits of radioactive materials have been reported from Cambrian, Pennsylvanian, and Permian rocks at a few localities.

INTRODUCTION

Rocks of Paleozoic age underlie most of central Wyoming. Complete sequences are especially well exposed along the margins of the Wind River Basin and in the adjacent mountain ranges. Owing to both academic and economic interest, these rocks have been studied extensively since the earliest territorial surveys in the 1870's. As a result, many basic data on the distribution, thickness, lithology, paleontology, and depositional history of the various Paleozoic formations have been accumulated. These data, many of which were obtained by the U.S. Geological Survey during a program of detailed geologic mapping and stratigraphic studies in the Wind River Basin, begun in the early 1940's, are summarized in this report.

Many individuals have thus contributed valuable information and ideas to this study; specific contributions will be cited below. Two of the early investigators, however, merit special recognition for their fundamental work on the classification and nomenclature of Paleozoic strata in the Wind River Basin and ajacent areas. Darton (1906 a, b) mapped the distribution of the Paleozoic rocks in the Owl Creek and Bighorn Mountains along the north and northeast margins of the basin and formally named and defined many of the individual formations. Blackwelder, during geological investigations in central and northwestern Wyoming from 1910 to 1913, traced many of the Paleozoic units through the Teton and Gros Ventre Ranges into the northern Wind River Range along the west edge of the Wind River Basin. He (1918) modified some of Darton's earlier classifications and also named and defined several new formations and members of formations, particularly in the lower part of the sequence. The work by Darton and Blackwelder, though since modified, has formed the basis for all subsequent studies of the Paleozoic rocks in central Wyoming.

Thomas (1948, 1952, 1957) presented excellent summaries of the stratigraphy and depositional history of the Paleozoic rocks in central Wyoming and adjacent regions.

GENERAL GEOGRAPHIC AND GEOLOGIC SETTING

The Wind River Basin includes about 8,500 square miles in central Wyoming, occupying most of Fremont County and the western part of Natrona County (fig. 1). It is an extensive structural depression bounded on the south by the Granite Mountains, on the west by the Wind River Range, on the north by the Washakie Range and Owl Creek and Bighorn Mountains, and on the east by the Casper arch. (See fig. 2.) As thus defined, the outline of the structural basin is roughly that of a parallelogram, with a maximum length of about 180 miles from northwest to southeast and a maximum width of about 75 miles north to south. The structurally deepest parts are close to the Owl Creek Mountains and the Casper arch, which lie along the north and east margins, respectively. The basin and adjacent mountains were formed during Laramide deformation in Late Cretaceous and early Tertiary times.

Most of the Wind River Basin is also a conspicuous topographic depression which has the same western and northern limits as the structural basin, but it is restricted on the south by the Beaver Divide, a 1,000-foot erosional escarpment of Tertiary rocks, and on the east by a series of broad low ridges, sometimes called the Hiland Divide, extending north from the Rattlesnake Hills to the south end of the Bighorn Mountains. The Hiland Divide separates the Wind River drainage system from that of the Powder and North Platte Rivers to the east; the Beaver Divide separates the Wind River drainage system from that of the Sweetwater River to the south.

The floor of the Wind River Basin, underlain by flat-lying lower Eocene rocks, is a region of low relief with only a few scattered buttes and mesas rising to as much as 500 feet above the general landscape. In the northwestern part of the basin, the altitude of the floor is 7,000 feet above sea level, whereas farther to the east and southeast in the central part it is 5,000–5,500 feet. With few exceptions, Paleozoic rocks do not crop out within the basin proper, but they have been penetrated in many places near the margins by wells drilled for oil and gas.

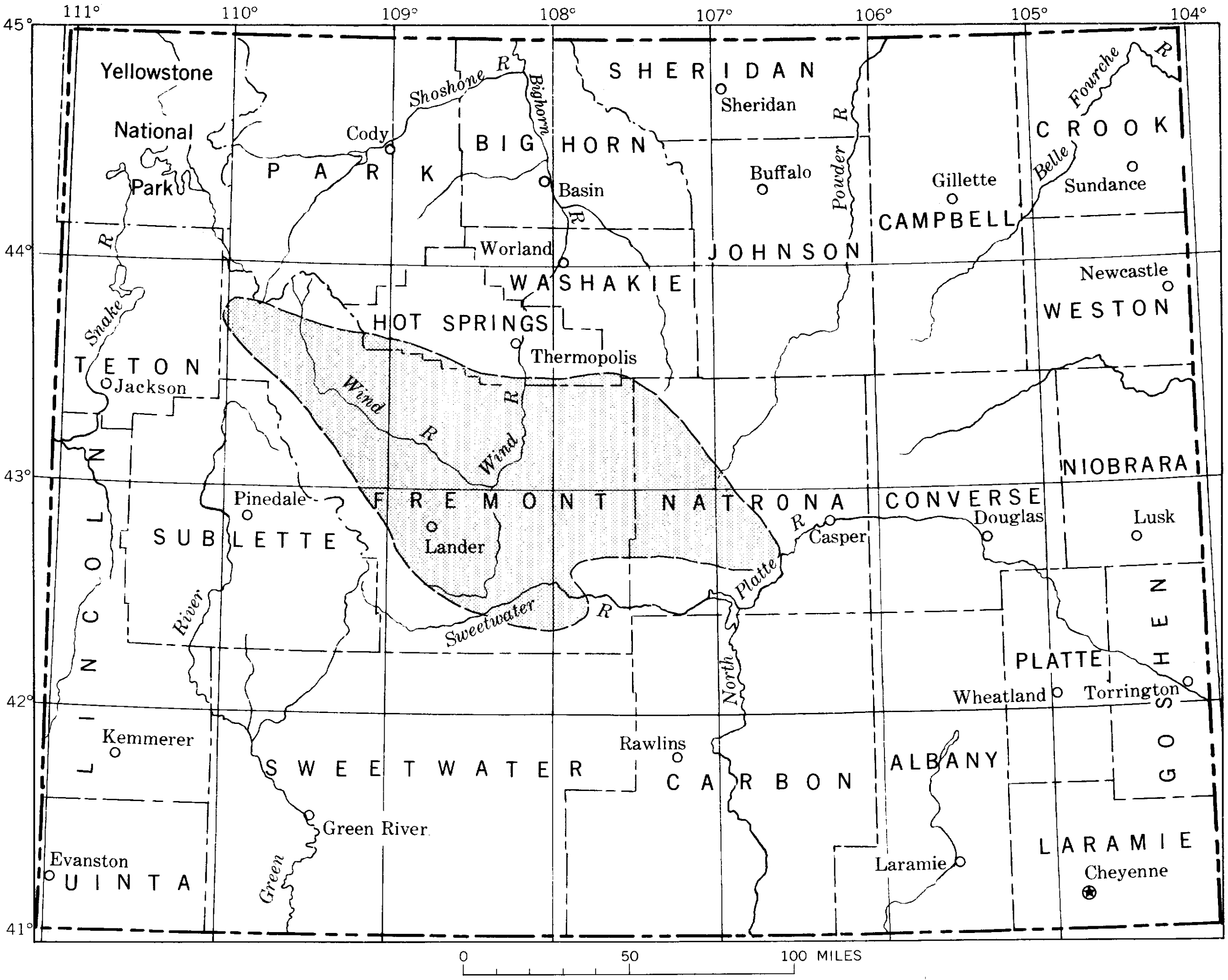


FIGURE 1.—Index map of Wyoming showing location of Wind River Basin (stippled).

The Granite Mountains are formed from a large anticlinal uplift that extends eastward from the south end of the Wind River Range for about 75 miles. The steep flank is to the south and southwest; the gentle north flank coincides with the south margin of the Wind River. The range was folded to mountainous proportions and deeply eroded during Laramide deformation, but later was almost completely buried by middle and upper Tertiary sediments. At present, the range is visible only as a series of prominent granite knobs projecting as much as 1,000 feet above a nearly level plain of flat-lying Miocene and Pliocene rocks, 6,500–7,000 feet above sea level, upon which the Sweetwater River drainage is established. Paleozoic rocks crop out in some places along the flanks of subsidiary folds, such as the Rattlesnake Hills and Conant Creek anticline, that project northwestward into the Wind River Basin from the main mass of the range (pl. 1).

The Wind River Range, culminating in 13,785-foot Gannett Peak, is the largest mountain uplift in Wyoming. The range trends northwest through the west-central part of the State for about 100 miles and has a maximum width of about 50 miles. Its west flank, bordering on the Green River Basin, is steep to overturned and broken by eastward-dipping reverse faults along which 35,000 feet or more of structural displacement has taken place (Berg, 1961). In contrast, the east flank, with which we are concerned in the present study, is characterized by moderately gentle dip slopes on Paleozoic rocks which descend uniformly eastward into the Wind River Basin at dips of 10°–20°. The structural continuity of the east flank is interrupted in a few places by high-angle normal and reverse faults, with displacements commonly less than 1,000 feet, and by sharp monoclinal folds. Precambrian crystalline rocks, chiefly massive granite and granite gneiss, form the core of the range, and Paleozoic rocks are well exposed in a nearly linear outcrop belt along the entire east flank (pl. 1). Complete sections of the Paleozoic rocks may be observed in many steep-walled canyons which have been cut transverse to the strike of the strata by eastward-flowing tributaries of the Wind River. (For example, see fig. 3.)

The Washakie Range, at the northwest corner of the Wind River Basin, is a series of faulted folds, in echelon, which extends northwest from Black Mountain (S½ T. 7 N., R. 4 W., pl. 1) for about 70 miles (Love, 1939, p. 5). This extensive uplift was completely buried by Eocene and Oligocene pyroclastic rocks of the Absaroka Range; it has now been partly exhumed. The south flank of the Washakie Range, generally steep and broken by northward-dipping reverse faults, is visible in several areas, and the Precambrian rocks exposed in some of the larger folds may represent part of its central core (Keefer, 1957, p. 205); the north flank is still buried. Altitudes of the exposed parts are as much as 10,000 feet above sea level. Discontinuous outcrops of Paleozoic rocks occur around the margins of many individual anticlines.

The Owl Creek Mountains rise abruptly along the north edge of the Wind River Basin. This large anticlinal uplift, separating the Wind River Basin from the Bighorn Basin to the north, extends nearly east-west across central Wyoming from the southeast edge of the Absaroka Range to the southwest end of the Bighorn Mountains, a distance of about 80 miles.[1] Rocks on the south flank of the range are steep to overturned, and have overriden the north margin of the Wind River Basin along an extensive system of reverse faults (Keefer, 1965, pl. 4). Strata on the north side of the range dip gently northward into the Bighorn Basin. Although the range as a whole is elongated nearly east-west, individual segments in the west half trend obliquely northwestward. Altitudes range from 4,800

[1] That part of the range lying east of Wind River Canyon was called the Bridger Range by many previous workers, but in recent years it has been a more common practice to refer to the entire range as the Owl Creek Mountains.

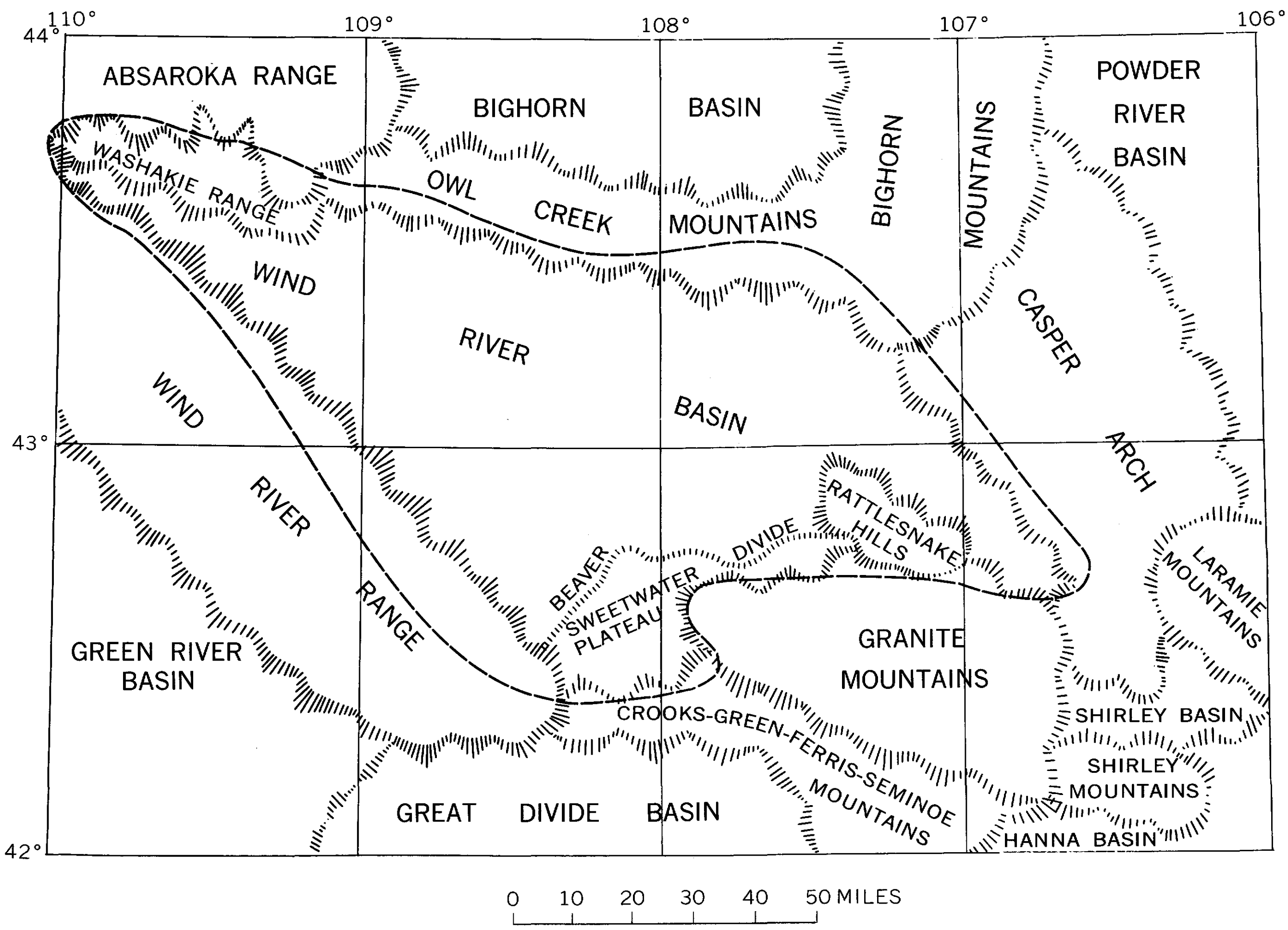


FIGURE 2.—Major physiographic and structural features in central Wyoming. Dashed line indicates approximate outline of Wind River structural basin.

205-540 O—66——2

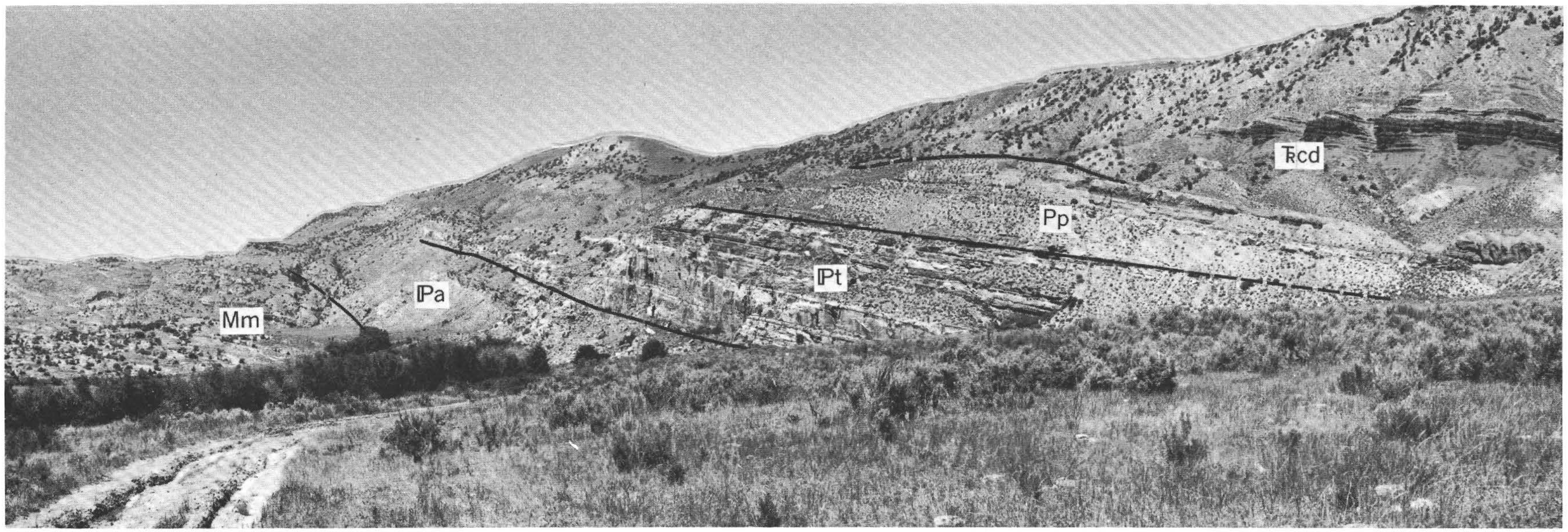


FIGURE 3.—Exposures on north wall of Dinwoody Canyon, northern Wind River Range. TRcd, Chugwater and Dinwoody Formations; Pp, Park City Formation and equivalents; IPt, Tensleep Sandstone; IPa, Amsden Formation; Mm. Madison Limestone.

to 6,000 feet above sea level along the south edge to 7,000 to 9,500 feet along the crest. Complete sequences of Paleozoic rocks crop out in many places; the best and most spectacular exposures are in Wind River Canyon which cuts across the central part of the range (fig. 4).

The southwest-plunging end of the Bighorn Mountains forms the northeast margin of the Wind River Basin. This part of the range has an extensive core of Precambrian crystalline rocks flanked on the south by Paleozoic strata that dip south about 15° (Tourtelot, 1953). These Paleozoic rocks have likewise been thrust south over the deep part of the Wind River Basin. The Bighorn Mountains are separated structurally from the Owl Creek Mountains to the west by a broad shallow synclinal area underlain for the most part by flat-lying Eocene rocks.

The Casper arch is a major, but not deeply eroded, structural upward whose steep to overturned and faulted west flank coincides with the east margin of the Wind River structural basin. The arch, which is underlain chiefly by Upper Cretaceous and Paleocene rocks, is characterized by low hills and ridges which slope gently eastward into the Powder River Basin. Altitudes range generally from 5,500 to 6,000 feet above sea level. No Paleozoic rocks crop out along the Casper arch, but all or part of the sequence has been penetrated by drill holes at many localities. Outcrops are present to the south, however, along the north edge of Casper Mountain at the northwest end of the Laramie Mountains (fig. 2). A complete sequence of Paleozoic strata is also exposed near Alcova Reservoir, at the extreme southeast end of the Wind River Basin.

GENERAL STRATIGRAPHIC FEATURES

During Paleozoic time central Wyoming was part of the interior stable shelf region that lay along the east side of the main Cordilleran geosyncline. Rocks representing parts of all the Paleozoic systems, except possibly the Silurian, were deposited across the present site of the Wind River Basin during numerous transgressions and regressions of epicontinental seas. Because the greatest sedimentary accumulation was toward the west, in the geosynclinal area of southeastern Idaho, the Paleozoic sequence is thicker and more com-

FIGURE 4.—Aerial view looking north into Wind River Canyon, central Owl Creek Mountains. Boysen Dam and north end of Boysen Reservoir are at bottom of photograph. PPu, Park City, Tensleep, and Amsden Formations; MOu, Madison Limestone and Bighorn Dolomite; €u, Cambrian rocks, undivided; p€r, Precambrian igneous and metamorphic rocks; f–f, Boysen normal fault. Faulted complex of Paleozoic and Mesozoic rocks, overlapped by Tertiary rocks, occupies lower half of view. Photograph courtesy of P. T. Jenkins and L. P. House.

plete in the western part of the Wind River Basin than in the eastern part. Marked changes in the thickness and lithology of nearly all formations take place from west to east across the basin, and some units disappear eastward owing to truncation or nondeposition.

The general lithologies, thicknesses, and stratigraphic relations of Paleozoic formations around the margins of the Wind River Basin and along the flanks of the adjacent mountain ranges are shown graphically on four correlation charts, plates 2–5. Individual measured surface sections or drilled subsurface sections were selected at 26 key localities, the distance between adjacent sections ranging from 6 to 28 miles (pl. 1). Localities and sources of data are listed in table 1; significant fossil collections are given in tables 2–8.

The stratigraphy is discussed according to the classification and nomenclature of strata in the northwestern part of the basin, where the Paleozoic record is most nearly complete. The rocks there are classified as follows:

Permian:
- Park City Formation and equivalents

Pennsylvanian:
- Tensleep Sandstone
- Amsden Formation

Mississippian:
- Madison Limestone

Devonian and lowermost Mississippian:
- Darby Formation

Ordovician:
- Bighorn Dolomite

Cambrian:
- Gallatin Limestone
- Gros Ventre Formation
- Flathead Sandstone

CAMBRIAN ROCKS

GENERAL FEATURES

A well-defined transgressive marine sequence, including a basal sandstone unit, a middle shale unit, and an upper limestone unit, was deposited across most of central Wyoming during an eastward advance of the epicontinental sea in Middle and Late Cambrian time (fig. 5). The major trend of transgression was interrupted repeatedly by minor retreats of the sea. At times certain areas emerged and were eroded, while adjacent areas were still submerged and receiving sediments; there is no evidence, however, that the sea withdrew completely from the region between the time that the initial sediments were laid down in Middle Cambrian time and the close of the period. With each new surge the sea encroached farther eastward; by Late Cambrian time all central Wyoming was inundated, and the sea spread as far as the Black Hills region in northeastern Wyoming and western South Dakota (Thomas, 1957, p. 4).

The Cambrian sequence varies greatly in thickness and lithology from one place to another. The tripartite division of sandstone, shale, and limestone is especially distinct in the mountains along the west side of the Wind River Basin; but toward the east the rocks become progressively more clastic, and many units lose their distinctiveness owing to changes in facies. Formation boundaries cross time lines and become progressively younger from west to east across the basin area. Thickness of Cambrian rocks decreases from a maximum of about 1,200 feet along the west edge of the basin to only 50 feet in the eastern most part (fig. 6). The decrease is due partly to nondeposition and partly to the pronounced erosional unconformity at the top of the sequence.

In the Owl Creek and Bighorn Mountains, Darton (1906a, p. 14–15; 1906b, p. 23–26) referred all Cambrian strata to the Deadwood Formation, a name he had given earlier (Darton, 1901, p. 505) to Cambrian rocks in the Black Hills region. In the northwestern Wind River Range, Eliot Blackwelder (U.S. Geol. Survey, 1918, unpub. field notes), distinguished three formations corresponding to the three major lithologic subdivisions and employed terms then in current use farther west—Flathead Sandstone, Gros Ventre Formation, and Gallatin Limestone (fig. 7). Blackwelder's classification has been used subsequently throughout much of the Wind River Basin, and is used in this report.

Because of the variations in thickness and lithology, however, alternate systems of classification and nomenclature have been proposed for the Cambrian sequence in some areas of the basin. In the central and eastern Owl Creek Mountains, for example, Miller (1936, p. 124) applied the name Depass Formation to the lower part of the Cambrian sequence, and Deiss (1938, p. 1104) applied the name Boysen Formation to the upper part. In the southwestern part of the Wind River Basin, Shaw (1957) referred the basal Cambrian unit to the Flathead Sandstone, but elevated both the Gros Ventre and Gallatin to group status and separated each group into several individual formations. Shaw (1954) also introduced the name Buck Spring Formation for all Cambrian rocks overlying the Flathead Sandstone in the south-central part of the basin. Additional discussions of the problems in classification and nomenclature are included below in the descriptions of formations.

TABLE 1.—*Locations of measured sections and the sources of data used in compilation of this report*

[An asterisk (*) indicates source is unpublished field notes, U.S Geol. Survey; blank space indicates formation absent]

<table>
<tr><th>Section</th><th>Flathead Sandstone</th><th>Gros Ventre Formation</th><th>Gallatin Limestone</th><th>Bighorn Dolomite</th><th>Darby Formation</th><th>Madison Limestone</th><th>Amsden Formation</th><th>Tensleep Sandstone</th><th>Park City Formation and equivalents</th></tr>
<tr><td rowspan="2">Warm Spring Creek</td><td colspan="2">Secs. 30 and 31, T. 42 N., R. 108 W.</td><td colspan="3">Sec. 1, T. 41 N., R. 108 W.</td><td>Sec. 31, T. 42 N., R. 107 W.</td><td>Secs. 14 and 15, T. 41 N., R. 107 W.</td><td>Secs. 9 and 10, T. 41 N., R. 107 W.</td><td>Secs. 11 and 14, T. 41 N., R. 107 W.</td></tr>
<tr><td colspan="2">Miller (1936, p. 131, 132)</td><td colspan="3">Keefer (1957, p. 166–168)</td><td>Weart (1948)</td><td colspan="2">Keefer (1957, p. 170–172)</td><td>Weart (1948)</td></tr>
<tr><td rowspan="2">Dinwoody Canyon</td><td colspan="2">Northeastern part, T. 4 N., R. 6 W.</td><td>Sec. 11, T. 4 N., R. 6 W.</td><td colspan="2">Sec. 12, T. 4 N., R. 6 W.</td><td>Northeastern part, T. 4 N., R. 6 W.</td><td>Sec. 1, T. 4 N., R. 6 W.</td><td colspan="2">Sec. 6, T. 4 N., R. 5 W.</td></tr>
<tr><td>J. F. Murphy*</td><td>J. F. Murphy*
Eliot Blackwelder*</td><td>J. F. Murphy*</td><td>J. F. Murphy*
Eliot Blackwelder*</td><td>J. F. Murphy*</td><td>J. F. Murphy*
Eliot Blackwelder*</td><td>J. F. Murphy*</td><td>Eliot Blackwelder*</td><td>Sheldon (1957, pl. 13)</td></tr>
<tr><td rowspan="2">Bull Lake Canyon</td><td colspan="5">Sec. 9, T. 2 N., R. 4 W.</td><td>Sec. 3, T. 2 N., R. 4 W.</td><td colspan="2">Sec. 2, T. 2 N, R. 4 W.</td><td>Sec. 35, T. 3 N., R. 4 W.</td></tr>
<tr><td colspan="9">Murphy and others (1956); Murphy and Richmond (1965)</td></tr>
<tr><td rowspan="2">Trout Creek</td><td rowspan="2">(Not measured)</td><td colspan="2">Sec. 1, T. 2 S., R. 3 W.</td><td colspan="2">Sec. 5, T. 2 S., R. 2 W.</td><td colspan="3">Secs. 4 and 5, T. 2 S., R. 2 W; secs 32 and 33, T. 1 S., R. 2 W.</td><td>Sec. 22, T. 1 S., R. 2 W</td></tr>
<tr><td colspan="2">Planetable traverse by W. R. Keefer and R. L. Koogle</td><td>Love and others (1947, p. 37)</td><td>J. F. Murphy*</td><td colspan="4">Love and others (1947, p. 36, 37)</td></tr>
<tr><td rowspan="2">Squaw Creek</td><td colspan="3" rowspan="2">(Not measured)</td><td colspan="2">Sec. 4, T. 33 N., R. 101 W.</td><td>Sec. 4, T. 33 N., R. 101 W.</td><td>Sec. 5, T. 33 N., R. 101 W.</td><td>Along North Fork Popo Agie River</td><td>Sec. 18, T. 33 N., R. 100 W.</td></tr>
<tr><td colspan="2">Burton (1952)</td><td>Strickland (1957, p. 21); Burton (1952)</td><td>Biggs (1951)</td><td>Burton (1952)</td><td>King (1947, p. 30–40)</td></tr>
<tr><td rowspan="2">Sinks Canyon</td><td colspan="3" rowspan="2">(Not measured)</td><td>Sec. 25, T. 32 N., R. 101 W.</td><td colspan="2">Sec. 18, T. 32 N., R. 100 W.</td><td>Sec. 18, T. 32 N., R. 100 W.</td><td>Sec. 20, T. 32 N., R. 100 W.</td><td>Sec. 25, T. 32 N., R. 100 W.</td></tr>
<tr><td>Hembree (1949)</td><td colspan="2">Strickland (1957, fig. 1)</td><td>Biggs (1951)</td><td>Hembree (1949)</td><td>Condit (1924, p. 24)</td></tr>
<tr><td rowspan="2">Beaver Creek</td><td colspan="2">Sec. 24, T. 30 N., R. 100 W.</td><td>Sec. 13, T. 30 N., R. 100 W.</td><td>Sec. 6, T. 29 N., R. 98 W.</td><td rowspan="2"></td><td>Secs. 8 and 18, T. 30 N., R. 99 W.</td><td>Sec. 9, T. 30 N., R. 99 W.</td><td>Secs. 9 and 10, T. 30 N., R. 99 W.</td><td>Secs. 11 and 12, T. 30 N., R. 99 W.</td></tr>
<tr><td colspan="4">Gooldy (1947)</td><td>Gooldy (1947)</td><td>Lower part, Shaw (1955, p. 61)</td><td>Gooldy (1947)</td><td>Condit (1924, p. 28, 29)</td></tr>
<tr><td rowspan="2">Sweetwater Canyon</td><td colspan="2">Secs. 27 and 28, T. 29 N., R. 97 W.</td><td colspan="2">Sec. 34, T. 29 N., R. 97 W.</td><td rowspan="2"></td><td colspan="3">Sec. 27, T. 29 N., R. 97 W.</td><td>Sec. 26, T. 29 N., R. 97 W.</td></tr>
<tr><td colspan="3">DeLand (1954)</td><td>Bell (1955)</td><td colspan="4">Bell (1955)</td></tr>
<tr><td rowspan="2">Windy Gap</td><td>Sec. 1, T. 42 N., R. 106 W.</td><td>Northeastern part, T. 43 N., R. 106 W.</td><td>Sec. 29, T. 43 N., R. 106 W.</td><td colspan="2">Sec. 36, T. 43 N., R. 106 W.</td><td>Sec. 7, T. 42 N., R. 105 W.</td><td>Sec. 19, T. 43 N., R. 106 W.</td><td>Sec. 18, T. 42 N., R. 105 W.</td><td>Sec. 13, T. 43 N., R. 107 W.</td></tr>
<tr><td>Love (1939, p. 15)</td><td>W. R. Keefer*</td><td>Love (1939, p. 16); W. R. Keefer*</td><td colspan="3">Love (1939, p. 18–25)</td><td>Biggs (1951)</td><td>Love (1939, p. 29)</td><td>Keefer (1957, p. 174)</td></tr>
<tr><td>Circle Ridge</td><td colspan="2">(Not measured)</td><td colspan="2">Secs. 30 and 31, T. 8 N., R. 2 W.; secs. 4 and 5, T. 42 N., R. 101 W.</td><td colspan="5">Sec. 13, T. 7 N., R. 3 W.</td></tr>
</table>

			Milton (1942)						
Sheep Ridge	Sec. 3, T. 6 N., R. 2 E.	Sec. 9, T. 7 N., R. 2 E.	Sec. 3, T. 7 N., R. 2 E.	Secs. 4 and 20, T. 7 N., R. 2 E.	Sec. 2, T. 7 N., R. 1 W.	Sec. 27, T. 7 N., R. 1 E.	Secs. 10 and 35, T. 7 N., R. 1 E.	Secs. 14 and 15, T. 7 N., R. 1 E.	Sec. 31, T. 7 N., R. 2 E.
	Phillips (1958)		Flanagan (1955)	Phillips (1958)	Thomas (1948, p. 85)	Phillips (1958)			
Blackrock Ridge	Sec. 33, T. 7 N., R. 3 E.	Sec. 29, T. 7 N., R. 3 E.	Secs. 23 and 24, T. 6 N., R. 3 E.	Sec. 24, T. 6 N., R. 3 E.		Secs. 24 and 25, T. 6 N., R. 3 E.	Sec. 25, T. 6 N., R. 3E.		Sec. 36, T. 6 N., R. 3 E.
	Powell (1957)					Powell (1957)			
Wind River Canyon	Sec. 4; T. 5 N., R. 6 E.			Sec. 33, T. 7 N., R. 6 E.		Sec. 32, T. 7 N., R. 6 E.	Sec. 30, T. 7 N. R. 6 E.	Sec. 19, T. 7 N., R. 6 E.	Sec. 28, T. 42 N., R. 94 W.
	Tourtelot and Thompson (1948)					Tourtelot and Thompson (1948)			Tourtelot (1953)
Eastern Owl Creek Mountains	Sec. 19, T. 40 N., R. 91 W.			Sec. 35, T. 41 N., R. 92 W.		Sec. 35, T. 41 N., R. 92 W.	Producers Oil No. 1 Barcklay, sec. 11, T. 40 N., R. 91 W.		Secs. 7 and 8, T. 40 N., R. 91 W.
	H. A. Tourtelot*			Lyon (1956)		Lyon (1956)	Petroleum Information, Inc., Denver, Colo.		Tourtelot (1953)
Badwater	(Not measured)					Sec. 6, T. 40 N., R. 90 W.	Sec. 20, T. 40 N., R. 89 W.	Sec. 22, T. 40 N., R. 89 W.	Secs. 16 and 21, T. 41 N., R. 89 W.
						H. A. Tourtelot*			Tourtelot (1953)
Deadman Butte	Sec. 26, T. 39 N., R. 88 W.					Sec. 3, T. 40 N., R. 86 W.; sec. 34, T. 41 N., R. 86 W.	Sec. 21, T. 40 N., R. 86 W.	Secs. 32 and 33, T. 40 N., R. 86 W.	Sec. 4, T. 38 N., R. 87 W.
	H. A. Tourtelot*					Woodward (1957, p. 259, 260)			Tourtelot (1953)
Conant Creek	Sec. 13, T. 32 N., R. 94 W.					Sec. 23, T. 32 N., R. 94 W.			Sec. 36, T. 33 N., R. 94 W.; sec. 31, T. 33 N., R. 93 W.; sec. 2, T. 32 N., R. 94 W.
	Van Houten and Weitz (1956); Shaw (1957, p. 12)					V. L. White*			
Muskrat	(Not penetrated)					(Not penetrated)	Sinclair-Wyoming Oil Co. No. 2, sec. 1, T. 33 N., R. 92 W.		
							E-log study by J. A. Van Lieu		
East Canyon Creek	(Not measured)	Sec. 26, T. 33 N., R. 89 W.				Secs. 23 and 26, T. 33 N., R. 89 W.			(Not measured)
		Van Houten and Weitz (1956); J. L. Weitz*				Van Houten and Weitz (1956)			
Rattlesnake Hills	Sec. 24, T. 33 N., R. 88 W.					Sec. 24, T. 33 N., R. 88 W.	Sec. 32, T. 33 N., R. 87 W.	Sec. 24, T. 33 N., R. 88 W.	Sec. 4, T. 32 N., R. 87 W.
	Shaw (1957, p. 12); Bogrett (1951)					Bogrett (1951)			Thomas (1934, p. 1673, 1674)

TABLE 1.—*Locations of measured sections and the sources of data used in compilation of this report*—Continued

Section	Flathead Sandstone	Gros Ventre Formation	Gallatin Limestone	Bighorn Dolomite	Darby Formation	Madison Limestone	Amsden Formation	Tensleep Sandstone	Park City Formation and equivalents
Fish Creek	(Not penetrated)					Atlantic Refining Co. No. 1, sec. 9, T. 31 N., R. 84 W.			
						E-log study by J. D. Love			
Alcova	Northwestern part, T. 29 N., R. 83 W.					Northwestern part, T. 29 N., R. 83 W.			Sec. 30, T. 30 N., R. 82 W.
	Lee (1927, p. 52)					Lee (1927, p. 52)			Thomas (1934, p. 1677)
Bates Creek	Sec. 5, T. 30 N., R. 79 W.					Secs. 5 and 6, T. 30 N., R. 79 W.			Sec. 6, T. 30 N., R. 79 W.
	Harrison (1938)					Harrison (1938)			Jenkins (1950)
Notches	(Not penetrated)					Trigood Oil Co. Cheyenne Lease No. 10, sec. 3, T. 37 N., R. 85 W.			
						E-log study by J. A. Van Lieu			
North Casper Creek	Pure Oil Co. No. 1, sec. 36, T. 37 N., R. 82 W.					Pure oil Co. No. 1, sec. 36, T. 37 N., R. 82 W.			
	Sample and E-log study by J. A. Van Lieu					Sample and E-log study by J. A. Van Lieu			
Casper Mountain	Sec. 9, T. 32 N., R. 79 W.					Sec. 9, T. 32 N., R. 79 W.	Northwestern part, T. 32 N., R. 80 W.		Sec. 12, T. 32 N., R. 81 W.
	Sims (1948)					Sims (1948)	Lee (1927, p. 49)		Burk and Thomas (1956, p. 5–7)

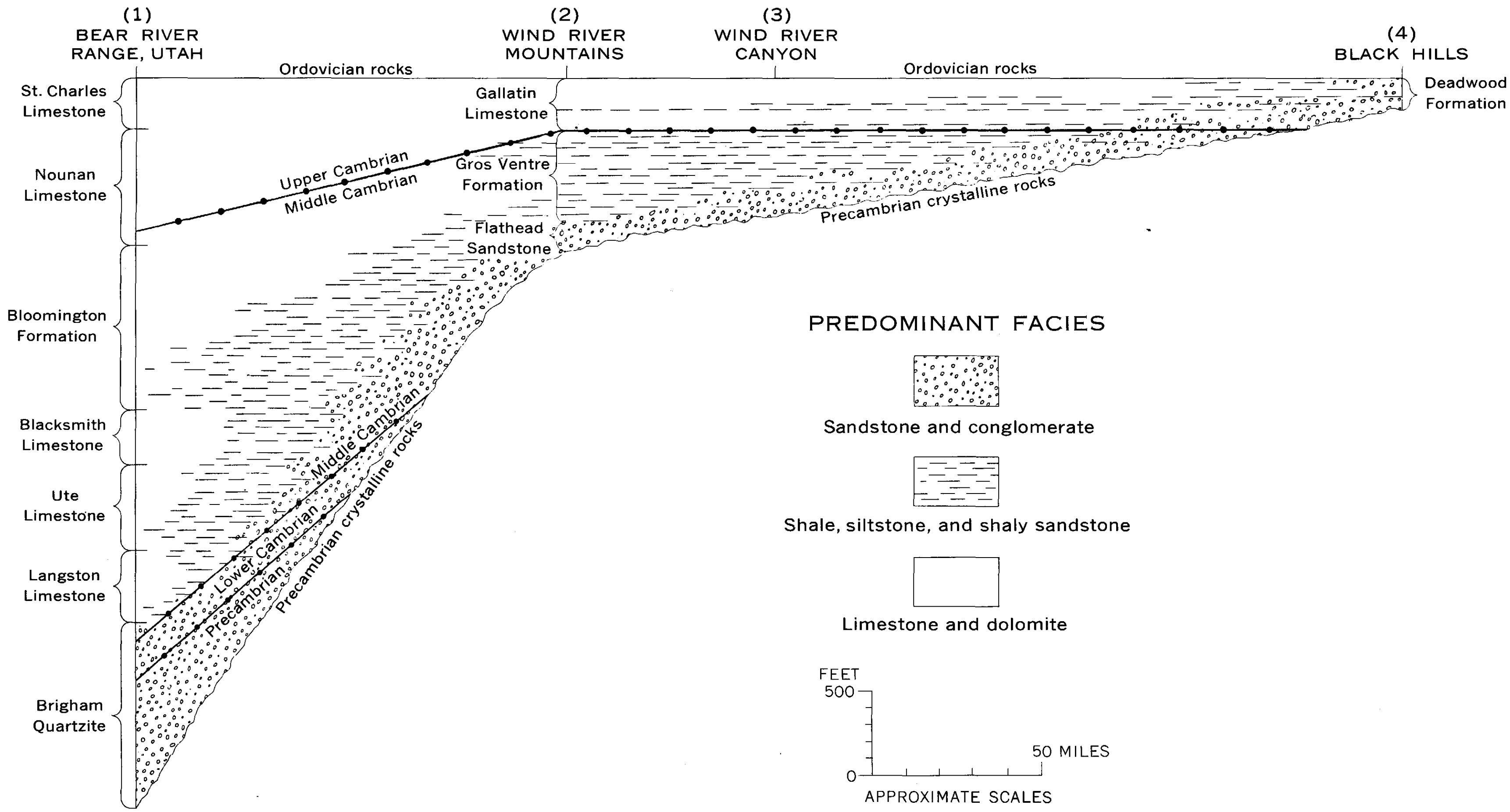


FIGURE 5.—Facies relations in Cambrian rocks from northern Utah to western South Dakota. Sources of data: (1) Stokes (1953), Deiss (1938), S.S. Oriel (oral commun., 1964); (2) Keefer (1957); (3) Tourtelot and Thompson (1948); (4) Darton and Paige (1925).

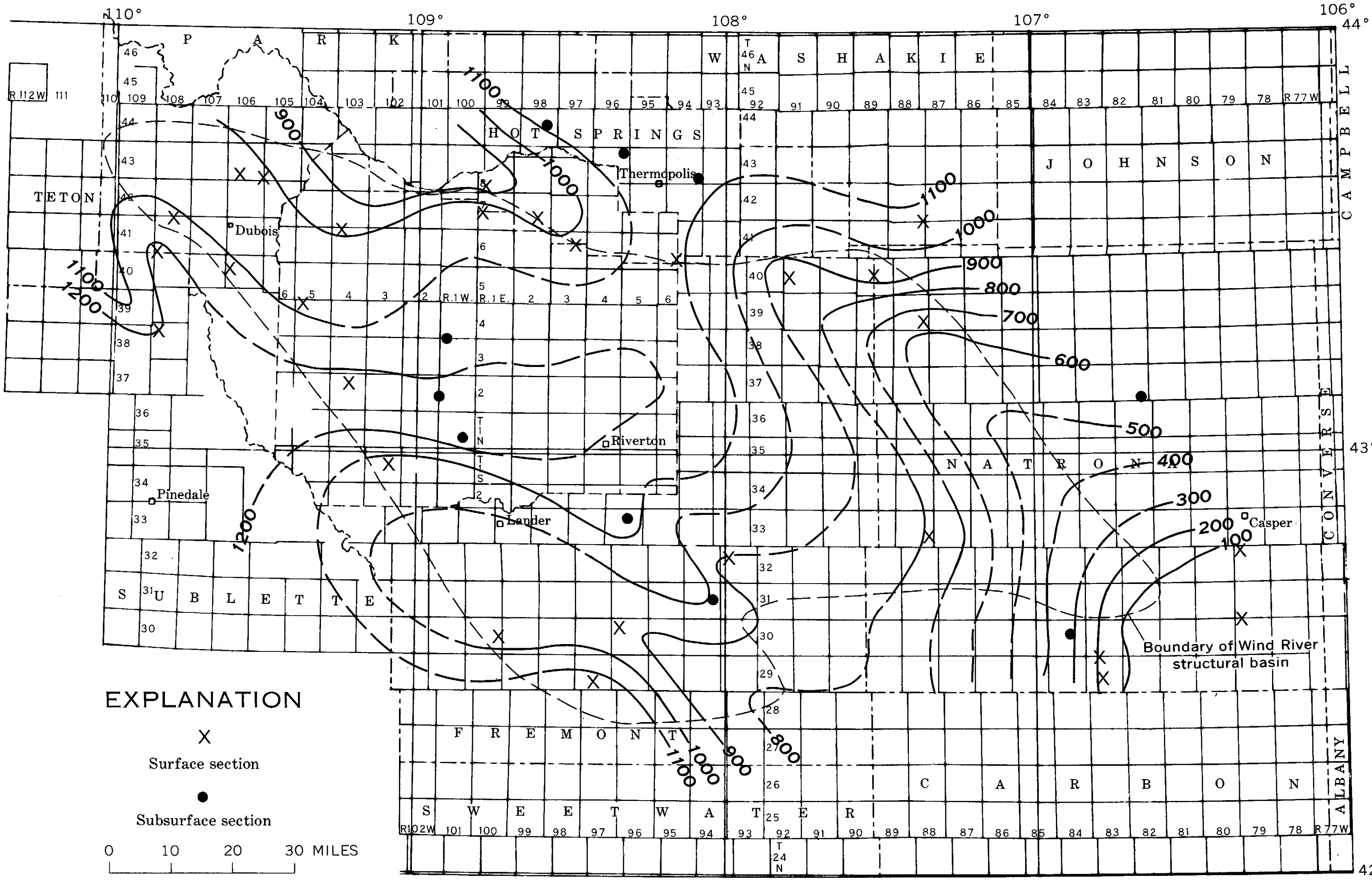


Figure 6.—Thickness map of Cambrian rocks in central Wyoming. Interval is 100 feet; isopachs are restored across mountain arches where rocks have been eroded; isopachs are dashed where control is inadequate.

205-540 O—66——3

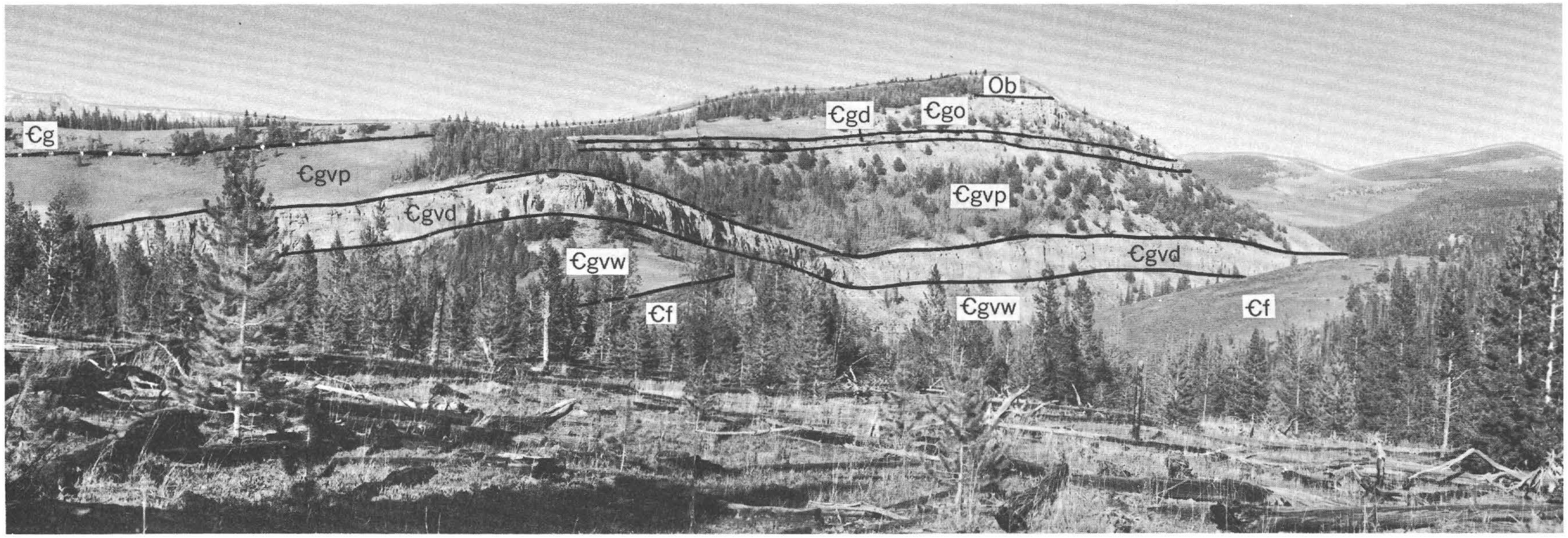


FIGURE 7.—Cambrian rocks along Warm Spring Creek, northern Wind River Range. Ob, Bighorn Dolomite. Єg, Gallatin Limestone; Єgo, Open Door Limestone Member, Єgd, type section of the Du Noir Limestone Member. Gros Ventre Formation; Єgvp, Park Shale Member, Єgvd, Death Canyon Limestone Member, Єgvw, Wolsey Shale Member. Єf, Flathead Sandstone. Note grass covered slopes of Gros Ventre Formation in right background and treeless dip slope of Flathead Sandstone in right middle ground.

FLATHEAD SANDSTONE

Definition

The name Flathead Formation was originally proposed by Peale (1893, p. 20–22) for exposures in Flathead Pass, southwestern Montana, and included a basal quartzite, the Flathead Quartzite, and an upper green shale sequence designated the "Flathead shales." In the Gros Ventre Range of northwestern Wyoming, Blackwelder (1918, p. 417) included the green shale in the lower part of his Gros Ventre Formation and restricted the Flathead to the basal quartzitic sandstone. The formation is now generally accepted in this restricted sense, and has been recognized over wide areas in western and central Wyoming.

Lithology and Thickness

Along the east flank of the Wind River Range the Flathead Sandstone has an average thickness of about 200 feet, except at the southeast end where it thickens to nearly 350 feet (pl. 2, Sweetwater Canyon section). The formation is predominantly pink, reddish-brown, tan, and gray fine- to coarse-grained sandstone. It is quartzitic for the most part, and contains glauconite and hematite in many places. Many of the beds are thinly laminated with pink and reddish-brown layers alternating with tan and gray layers. Some crossbedding is present. The basal beds of the formation are commonly conglomeratic and arkosic, containing angular fragments of quartz, feldspar, and mica. Granite pebbles as much as 1 inch across are locally disseminated in the lowermost strata. The upper beds of the Flathead are generally softer and are shaly, and have numerous partings of green and gray-green micaceous shale.

Topographically, the Flathead Sandstone usually forms a series of resistant blocky ledges which rise abruptly in steep slopes above the basal contact with the Precambrian crystalline rocks. Outcrops support heavy growths of trees in most places, but locally prominent treeless dip slopes on top of the uppermost resistant sandstone bed (fig. 7) occur. The sandstone commonly breaks into large slabs which slide short distances down the slopes; small rock slides are numerous along steep dip slopes.

Strata included in the Flathead Sandstone along the north and south margins of the Wind River Basin are similar lithologically to those in the Wind River Range. In the Washakie Range and western part of the Owl Creek Mountains, thicknesses remain fairly constant at 120–160 feet, but they increase eastward to 255 feet in the Wind River Canyon and to nearly 400 feet in the eastern Owl Creek Mountains (pl. 3, Badwater section). Along the south edge of the basin the Flathead ranges in thickness from 245 feet at the Conant Creek locality to more than 500 feet in the northern part of the Rattlesnake Hills (pl. 4). In the latter area the basal conglomerate is as much as 120 feet thick. The eastward thickening of the formation is at the expense of the overlying Gros Ventre Formation, with which it interfingers.

At the southeast end of the Wind River Basin, and in adjacent areas along the west side of the Laramie Mountains, 45–90 feet of coarse-grained quartzitic sandstone (pl. 5) was assigned to the Flathead Sandstone by Shaw (1954). These strata are overlain directly by clastic rocks forming the base of the Mississippian Madison Limestone, and Maughan (1963, p. C26) suggested the possibility that the sandstone now placed in the Flathead may represent a locally thick sandy facies of the basal part of the Madison Limestone.

Contact With Precambrian Rocks

The Flathead Sandstone unconformably overlies Precambrian igneous and metamorphic rocks. The contact is generally concealed by forest cover or talus. In a few places, however, such as in Wind River Canyon, the contact is continuously exposed for several hundred yards (fig. 8). The eroded surface of the Precambrian rocks apparently had some topographic relief; this relief probably accounts for some of the local variations in thickness of the Flathead.

Where the Flathead Sandstone directly overlies massive granite and granite gneiss, there is commonly a transitional zone of weathered arkose at its base. The weathered zone represents a regolith that was winnowed during the encroachment of the Cambrian sea.

Contact With the Gros Ventre Formation

The Flathead sandstone is overlain conformably by the Gros Ventre Formation; laterally the two formations interfinger. The formation boundary is usually placed at the top of the uppermost persistent quartzitic sandstone in the sequence, but in most places the change from quartzitic and, in part, shaly sandstone below to soft shale and shaly sandstone above is gradational. The contact is also marked locally by a topographic change from cliffs below to concave slopes above. On electric logs there is an easily recognized shift from the low resistivity and self-potential curves of the Gros Ventre to the more expanded curves of the Flathead (pl. 6).

Age

Few diagnostic fossils have been found in the Flathead Sandstone in the Wind River Basin (table 2); thus, the part of Cambrian time represented is not precisely known. From regional stratigraphic and paleontologic correlations, Miller (1936, p. 142) and Shaw (1957, p. 9) considered the formation to be early Middle

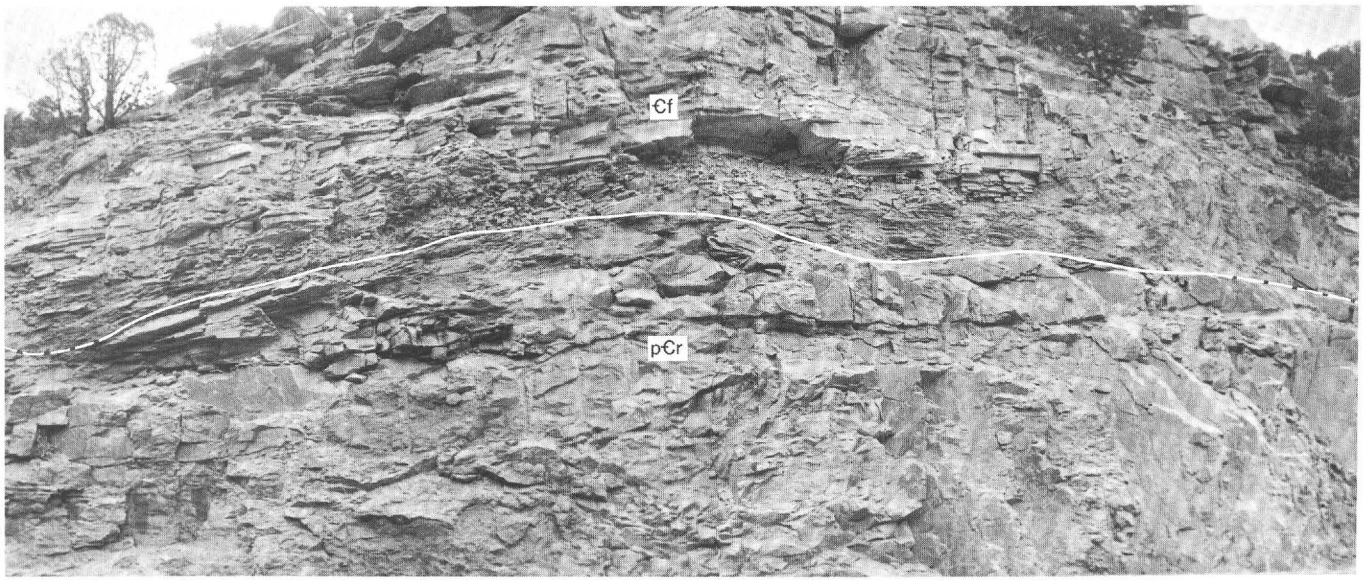


FIGURE 8.—Flathead Sandstone (€f) unconformably overlying Precambrian rocks (p€r) along east side of Wind River Canyon, central Owl Creek Mountains.

Cambrian in age. Owing to its transgressive origin, the sequence is somewhat younger in the eastern part of the basin than in the western part (fig. 5).

The Flathead Sandstone probably is lithologically continuous with the lower part of the Deadwood Formation of Late Cambrian and Early Ordovician age in the Black Hills and the Brigham Quartzite of late Precambrian and Early(?) and Middle Cambrian age in southeastern Idaho and northern Utah (fig. 5).

Conditions of Deposition

The Flathead is a classic example of transgressive sandstone, representing the initial deposit laid down during the eastward migration of the Cambrian seas into the interior stable shelf region (fig. 5). Debris eroded from the adjacent Precambrian terrane accumulated chiefly in a shore and near-shore environment. Periodic retreats and readvances of the sea caused intertonguing of the coarser sediments of the Flathead on the east with the finer grained sediments of the Gros Ventre Formation on the west.

GROS VENTRE FORMATION

Definition

The Gros Ventre Formation was named by Blackwelder (1918) from exposures on Doubletop Peak in the Gros Ventre Range, about 25 miles southwest of the Warm Spring Creek section (pl. 1). The sequence probably includes the lateral equivalents of the "Flathead shales" and the lower part of the Gallatin Formation of Peale (1893, p. 20–25) in southwestern Montana.

At the type section, the Gros Ventre Formation can be divided into three distinctive lithologic units: a lower shale and shaly sandstone unit; a middle limestone unit called the Death Canyon Member by Miller (1936, p. 119) and referred to in this report as the Death Canyon Limestone Member; and an upper shale and limestone unit. These units are readily distinguished in the western and northwestern parts of the Wind River Basin (fig. 7) but not in the central and eastern parts.

In th western part of the Wind River Basin, Shaw (1957) raised the Gros Ventre to group rank and, following the terminology of Weed (1900) in central Montana, referred to the upper unit as the Park Shale and to the lower unit as the Wolsey Shale. He (Shaw, 1957, p. 9–10) retained the Death Canyon Limestone for the middle unit. Although Shaw's terminology is used in this report, it is modified to the extent that the Gros Ventre is retained as a formation and the three subdivisions are classed as members. Gros Ventre strata are generally poorly exposed and, as a result, are mapped as a single unit.

Lithology and Thickness

In the Warm Spring Creek section at the northwest end of the Wind River Range (pl. 2) the Gros Ventre Formation is 747 feet thick and contains the threefold division mentioned above. The shale units are commonly grass covered and good exposures are rare; in contrast the Death Canyon Limestone Member forms vertical cliffs and is one of the most conspicuous units in the Cambrian sequence (fig. 7).

The Wolsey Shale Member is 107 feet thick along Warm Spring Creek and consists of greenish-gray, tan, and pink micaceous sandy shale and very fine grained shaly sandstone that is highly glauconitic in most places. The basal part contains some beds similar to those in the upper part of the Flathead Sandstone, so that the contact in places is gradational. The Death Canyon Limestone Member, 219 feet thick, is charaterized by cliff-forming gray thin-bedded crystalline limestone that is commonly mottled by inclusions of irregularly shaped tan granular limestone or dolomite masses a fraction of an inch to several inches across. The Park Shale Member consists of 421 feet of soft greenish-gray micaceous shale with varying amounts of thin-bedded gray limestone that increases in abundance toward the top. The limestone commonly contains beds of flat-pebble conglomerate, consisting of disk-shaped limestone fragments oriented mostly parallel to the bedding planes in a dense to granular limestone matrix. The flat pebbles are particularly conspicuous on weathered surfaces that transect the bedding planes.

TABLE 2.—*Fossils from Cambrian formations*

[Generic and specific names are quoted without modification from sources shown]

Stratigraphic section	Collection No. (pls. 2–5)	Species	References
Warm Spring Creek	1	*Lingulepis acuminatus*	Miller (1936, p. 132).
	2	*Cedarina* cf. *C. alberta*, *C. victoria*, *Bynumia* sp. indet., *Modocia*? sp., *Tricrepicephalus*? sp., *Semicircularia arcuata*.	Denson (1942).
	3	*Agnostus* sp., *Linnarssonella* sp. cf. *L. modesta*, *L. transversa*, *Obolus*? sp. cf. *O. zentus*.	Miller (1936, p. 132).
	4	*Kingstonia sublettensis*, *Blountia globosa*, *B. dunoirensis*, *B.* sp. cf. *B. amage*, *Crepicephalus* sp., *Cedaria* sp. aff. *C. prolifica*, *Arapahoia levis*, *Maryvillia* sp., *Agnostus* sp. aff. *A. tumidosus*.	Do.
	5	*Elvinia roemeri*, *Burnetia*? sp., *Eoorthis*? sp. cf. *E. iddingsi*	Miller (1936, p. 131).
	6	*Taenicephalus cordillerensis*, *Wilbernia* sp., *Billingsella coloradoensis*, *Billingsella*? sp. aff. *B. striata*, *Dicellomus* sp., *Lingulepis acuminata* var. *meeki*, *Acroteta microscopica* var. *tetonensis*, *Irvingella gibba*.	Do.
	7	*Briscoia schucherti*, *Synthrophia alata*, *Eoorthis* sp. cf. *E. iophon*	Do.
Dinwoody Canyon	1	*Obolus* sp	Eliot Blackwelder (U.S. Geol. Survey, unpub. field notes).
	2	Abundant fragments of Late Cambrian trilobites	Do.
	3	*Eoorthis* cf. *E. remnicha*	Do.
Sweetwater Canyon	1	*Glyphaspis*	Shaw (1957).
	2	*Bolaspidella*?	Do.
	3	*Bolaspis*, *Hyolithes*	Do.
	4	*Apachia convexa*, *Burnetia alta*, *Cliffia latagenae*, *Deadwoodia duris*, *Dellea saginata*, *Housia vacuna*?, *Iddingsia occidentalis*, *Irvingella* aff. *I. flohri*, *Xenocheilos* cf. *X. spineum*, *Elvinia* sp. (*Elvinia* zone).	Do.
	5	*Taenicephalus cordillerensis* (*Taenicephalus* zone)	Do.
	6	*Idahoia* sp., *Monocheilus*? sp., (*Idahoia* zone)	Do.
	7	*Ellipsocephaloides* sp., (*Idahoia* zone)	Do.
	8	Cranidia B and C of DeLand and Shaw (Trempealeauan)	DeLand and Shaw (1956, p. 561).
Windy Gap	1	*Billingsella coloradoensis*, *Eoorthis* cf. *E. remnicha*, *Lingulepis acuminatus*, *Acroteta microscopica* var. *tetonensis*, *Dicellomus* cf. *D. nanus*, *Hyolithes*? sp., *Prozacompsus* sp., *Idahoia* sp., *Prosaukia* sp., *Chariocephalus*? sp., *Maladia* sp. (*Ptychaspis-Prosaukia* zone).	Love (1939, p. 16).
Wind River Canyon	1	*Scolithus*	Miller (1936, p. 137).
	2	*Scolithus*	Deiss (1938, p. 1099).
	3	*Lingulepis*	Do.
	4	*Glyphaspis*	Deiss (1938, p. 1097, 1098).
	5	*Arapahoia spatulata*, *A.* sp. cf. *A. typa*, *Dicellomus nana*, *Linnarssonella tennesseensis*, *L.* sp. cf. *L. modesta*.	Miller (1936, p. 136); Deiss (1938, p. 1096).
	6	*Arapahoia spatulata*, *A. typa*, *Blountia dunoirensis*?	Deiss (1938, p. 1906).
	7	*Crepicephalus tripunctatus*, *Maryvillia* sp. cf. *M. ariston*, *Hyolithes* sp	Miller (1936, p. 136).
	8	*Blountia* sp., *Kingstonia* sp., *Tricrepicephalus* sp., *Aphelaspis* sp	Deiss (1938, p. 1095).
	9	*Taenicephalus cordillerensis*, *Saratogia* sp., *Billingsella coloradoensis*	Miller (1936, p. 136).
	10	*Elvinia* n. sp., *Pterocephalia* n. sp	Deiss (1938, p. 1094).
	11	*Dikelocephalus*	Deiss (1938, p. 1093).
Deadman Butte	1	*Peronopsis*?, *Hyolithes*?, *Elrathiella*, *Westonia* cf. *W. ella*	Woodward (1957, p. 222).
	2	*Tricrepicephalus tripunctus*, *Dicellomus nanus*	Do.
Conant Creek	1	*Tricrepicephalus* (*Crepicephalus* zone)	Shaw (1957).
	2	*Billingsella*, *Eoorthis*, *Glyptotrophia*, *Linnarssonella*, *Berkia*, *Bernia*, *Dellea*, *Parabolinoides* (*Elvinia* zone).	Van Houten and Weitz (1956); Shaw (1957).
	3	*Billingsella*, *Eoorthis*, *Huenella*, *Bemaspis*, *Orygmaspis*, *Taenicephalus* (*Taenicephalus* zone).	Van Houten and Weitz (1956).
Rattlesnake Hills	1	Approximate horizon of the type of *Westonia dartoni* of Walcott	Shaw (1957).
	2	*Lingulepis*; an undescribed trilobite of pre-*Cedaria* Middle Cambrian age	Do.

Southeast of Warm Spring Creek, along the east flank of the Wind River Range, the thickness of the Gros Ventre Formation ranges from a minimum of 300 feet to a maximum of about 760 feet (pl. 2). The three members can be recognized at least as far southeast as Bull Lake Canyon, but farther south the Death Canyon Limestone Member becomes progressively more shaly and sandy and loses its identity as a resistant cliff-forming unit. In Sweetwater Canyon at the southeast end of the range, the Gros Ventre is chiefly interbedded shale and sandstone with numerous thin beds of limestone and flat-pebble conglomerate in the upper one-third to one-half of the formation.

The three members of the Gros Ventre Formation are also easily identified in outcrops along the south side of the Washakie Range where their aggregate thickness ranges from 500 to 700 feet. Sufficient observations have not been made in the Circle Ridge area (pls. 1, 3) to determine the presence of the Death Canyon Limestone Member, but this unit seems to have disappeared by facies change at least as far west as the exposures near Merritt Pass in the western Owl Creek Mountains (T. 8 N., R. 1 W., pl. 1).

In the Owl Creek and southern Bighorn Mountains the Gros Ventre Formation is a nonresistant unit consisting largely of interbedded gray-green and tan glauconitic micaceous shale, sandstone, and siltstone and some beds of limestone and flat-pebble conglomerate in the upper part (fig. 9). Some of the sandstone is hematitic and red to reddish brown. Thickness ranges from 150 feet in the eastern part of this region to 575 feet in the central and western parts; variations seem to be largely the result of eastward interfingering with the upper part of the Flathead Sandstone (pl. 3). Because of this interfingering, Miller (1936, p. 123, 124) proposed that the Flathead and Gros Ventre be combined as a single unit and applied the name Depass Formation to all Cambrian rocks older than the Gallatin Limestone. Tourtelot and Thompson (1948), Tourtelot (1953), and Woodward (1957), however, differentiated the Flathead and Gros Ventre Formations on the basis of detailed mapping of the Cambrian rocks in this region.

Along the south margin of the Wind River Basin, strata assigned to the Gros Ventre Formation are predominantly reddish-brown and gray fine-grained sand-

FIGURE 9.—Cambrian, Ordovician, Mississippian, and Pennsylvanian strata exposed along west wall of Wind River Canyon, central Owl Creek Mountains. IPta, Tensleep and Amsden Formations; Mm, Madison Limestone; Ob, Bighorn Dolomite; Єg, Gallatin Limestone; Єgv, Gros Ventre Formation.

stone, siltstone, and a few thin beds of shale. In the Conant Creek area, beds of gray glauconitic flat-pebble limestone conglomerate, possibly equivalent to part of the Death Canyon Limestone Member, are present in the middle of the formation; some of these beds extend as far east as the East Canyon Creek section, but they seem to be entirely absent at the north end of the Rattlesnake Hills (pl. 4). Thickness of the Gros Ventre Formation decreases from 512 feet in Sweetwater Canyon to 287 feet in the northern Rattlesnake Hills. The decrease in thickness eastward is caused by interfingering with the Flathead Sandstone at the base of the formation and possibly in part by truncation at the top by the Madison Limestone. The Madison bevels eastward across successively older beds in the Cambrian sequence, overlaps the entire Gallatin Limestone in the south-central part of the basin, and rests on the Gros Ventre Formation at the north end of the Rattlesnake Hills (pl. 4 and fig. 15; Shaw, 1957, p. 12).

Inasmuch as the Cambrian rocks overlying the Flathead in the south-central part of the Wind River Basin, and adjacent regions to the south, are atypical of the Gros Ventre Formation farther west and northwest, Shaw (1954) introduced a new name, Buck Spring Formation, for this red facies. That name, however, has not yet been used widely in the Wind River Basin.

A well drilled in the North Casper Creek oil field along the east edge of the Wind River Basin (pl. 5) penetrated 205 feet of Cambrian strata between the Madison Limestone and the Flathead Sandstone. This sequence, consisting chiefly of purple and tan to white glauconitic micaceous siltstone in the lower part and purple and gray glauconitic dolomite in the upper part, may represent only the Gros Ventre Formation, or it may also contain some equivalents of the Gallatin Limestone. Farther south, however, at the west end of Casper Mountain, these particular strata are entirely cut out beneath the Madison Limestone which rests directly on the Flathead Sandstone.

Contact With the Gallatin Limestone

The contact between the Gros Ventre Formation and the overlying Gallatin Limestone, in those areas where the Du Noir Limestone Member is present at the base of the Gallatin, is marked by a sharp change from grass-covered shaly slopes below to cliff-forming limestone above (figs. 7, 9). In subsurface sections there is an easily recognized shift in the resistivity curves on electric logs (pl. 6). Where the Du Noir Limestone Member is absent or poorly developed, however, the formation contact is not so readily defined. At the south end of the Bighorn Mountains, Woodward (1957, p. 217) stated that "it is difficult if not impossible to differentiate the Gros Ventre from the overlying Gallatin northeastward from the point of disappearance of the basal Du Noir limestone member of the Gallatin." Similarly, along the south margin of the Wind River Basin, east of Conant Creek, Shaw (1957) recognized the possibility that the upper part of the Gros Ventre Formation (his Buck Spring Formation) contained some equivalents of the Gallatin Limestone.

An erosional unconformity between the Gros Ventre Formation and Gallatin Limestone was observed by Miller (1936, p. 122) in the Teton Range and by Blackwelder (1918) at the type section of the Gros Ventre in the Gros Ventre Range in northwestern Wyoming. Miller (1936, p. 119) reported an unconformity locally in the Warm Spring Creek area in the northern Wind River Range, and DeLand (1954) described a similar relation between the two formations in the Sweetwater Canyon area at the south end of the range. Elsewhere in the Wind River Basin the contact seems to be conformable.

In the southeastern part of the basin, strata assigned to the Gros Ventre Formation are overlain, and in some places completely overlapped, by the Madison Limestone.

Age

Fossils collected from the Gros Ventre Formation (table 2) indicate that over much of the basin area the formation is of middle to late Middle Cambrian age. The *Cedaria* zone is represented in the upper part of the formation; this zone is considered to mark the base of the Upper Cambrian Series (Howell and others, 1944).

Strata assigned to the Gros Ventre Formation are somewhat younger in the eastern part of the Wind River Basin that in the western part. Both in the Rattlesnake Hills and southern Bighorn Mountain areas (pl. 3, 4), east of the wedge edge of the Du Noir Limestone Member of the Gallatin Limestone, the stratigraphic and paleontologic relations suggest that the upper part of the Gros Ventre is in part the age equivalent of the Du Noir (Shaw, 1957, p. 13; Woodward, 1957, p. 223, fig. 4). The base of the Gros Ventre also becomes younger eastward because of intertonguing with the Flathead Sandstone (fig. 5).

Conditions of Deposition

Strata of the Gros Ventre Formation are offshore deposits that were laid down contemporaneously with the shore and near-shore sediments of the Flathead Sandstone as the Cambrian sea continued its eastward transgression across central Wyoming (fig. 5). The Death Canyon Limestone Member probably was deposited during a more easterly advance of the sea than was the overlying Park Shale Member; if so, there may have been a limited westward retreat of the shoreline, or the shoreline may have been nearly stabilized in one position, during the early part of the deposition of the Park. Later, the sea readvanced and limestone was deposited along with the shale over much of the wind River Basin area.

Shaw (1957, p. 13) cites evidence that the reddish-brown clastic rocks of the Gros Ventre Formation (his Buck Spring Formation) in the south-central part of the basin are partly marine and partly nonmarine. He concludes that this area, and the adjacent area to the south, were emergent periodically "either through local uplift to the south of Wyoming or simply by filling of the sea at a faster rate than the eustatic rise of sea level could submerge the sediments." Although the data are inconclusive, Shaw (1957, p. 13) favors the second interpretation, because (1) the strata generally lack coarse conglomerates and (2) the repeated intercalation of sandstones containing inarticulate brachiopods suggests a very close balance between depositional filling and marine transgression such as would occur on a low shelving shore.

In the western part of the Wind River Basin the Park Shale Member of the Gros Ventre Formation was partially eroded before the deposition of the Du Noir Limestone Member of the Gallatin Limestone. Farther east, however, sedimentation seems to have been continous.

The origin of limestone flat-pebble conglomerate beds has been discussed by several workers. McKee (1945, p. 65–70) presents an excellent summary of the different views bearing on this subject in connection with his detailed studies of the Cambrian rocks of the Grand Canyon region. He (McKee, 1945, p. 67–69) observed that the flat-pebble conglomerate beds in the canyon region are common in regressive deposits, but they do not occur in transgressive deposits, and that they developed exclusively along the shoreward margins of the limestone facies. Although most previous investigators believed that such conglomerate was formed near shore and in shallow water (beach or tidal-flat conditions), McKee (1945, p. 69) concluded that some were formed at least 100 miles seaward from the strand line. From the sedimentary relations, he further concluded that these beds originated at a time when basin sinking ceased or slowed down to such an extent that sedimentation built up the bottom, partly above the profile of equilibrium, after which the sea regressed and conglomerate was formed in favorable localities.

Regarding similar deposits in parts of Montana and Wyoming, Lochman-Balk (1957, p. 145) and Robinson (1963, p. 20) agree that the flat limestone pebbles formed largely through fragmentation of previously deposited limestone by wave and current action, although Robinson also suggests that some of the conglomerate may have formed as a result of different rates of lithification of thin-bedded strata. The distribution of flat-pebble conglomerate in the Cambrian sequence of the Wind River Basin seems to be very similar to that found elsewhere across the Rocky Mountain foreland region, but few studies have been made. In a brief discussion Shaw (1957, p. 13) suggests that the normal condition for the development of the conglomerate was one of shallow water with restricted circulation.

Glauconite, though absent in most of the Paleozoic rocks of central Wyoming, is abundant in many of the Cambrian units. Conditions facilitating the development of glauconite have been summarized by Cloud (1955) and Lochman-Balk (1957). These include: (1) marine waters of normal salinity, (2) slightly reducing environment resulting from decaying organic matter, (3) ingestion of sediment by bottom-dwelling organisms having low oxygen requirements, (4) presence of micaceous minerals (especially biotite) or bottom muds of high iron content, and (5) slow sedimentation rate to avoid rapid burial of bottom sediments. Presumably, many of these conditions existed during the deposition of much of the Paleozoic succession in this region. However, one important factor seems to set the Cambrian apart from the younger rock systems. The Cambrian seas advanced across an eroded surface of Precambrian crystalline rocks which doubtless furnished a greater supply of micaceous minerals than did any of the succeeding periods of the Paleozoic. Where other favorable conditions were met along the slowly submerging shelves, a significant quantity of these minerals was probably converted to glauconite.

GALLATIN LIMESTONE

Definition

The Gallatin Limestone, the youngest Cambrian formation in central Wyoming, was named and described by Peale (1893, p. 22, 23) from exposures in the Gallatin Range of southwestern Montana. At the type section it originally included in its lower part limestone and shale of Middle Cambrian age, which probably constitute the upper part of the Gros Ventre Formation as that unit was later defined by Blackwelder (1918), and limestone of Late Cambrian age in its upper part. Inasmuch as the Gallatin Limestone, as used in the Wind River Basin, is not completely representative of the type Gallatin, Deiss (1938, p. 1104) substituted a new name, Boysen Formation, for all Upper Cambrian rocks in the Wind River Canyon area and subdivided it into the Maurice, Snowy Range, and Grove Creek Members (pl. 3). The Boysen, although extended westward into the Gros Ventre Range by Foster (1947, p. 1547) and Wanless and others (1955, p. 13), has not gained wide acceptance in central Wyoming.

Along the east flank of the Wind River Range the Gallatin Limestone can be divided into three units: a thin lower limestone unit, a thin middle shale unit, and a thick upper limestone unit. The basal limestone was named the Du Noir Member (referred to as the Du Noir Limestone Member in this report) by Miller (1936, p. 124, 125); the section along Warm Spring Creek at the northwest end of the range was designated as the type (fig. 7). Shaw (1957, p. 13–15) classed the Gallatin as a group in this region, raised the Du Noir to formational status, and named the remainder the Open Door Limestone (Shaw and DeLand, 1955, p. 38) from exposures along the east wall of Granite Canyon, just below the topographic feature known as the "Open Door" in the Gros Ventre Range in northwestern Wyoming. The Open Door Limestone included the thin shale unit at the base that was referred to as the Dry Creek Shale Member by Shaw (1957, p. 14) following the redefinition of that unit by Lochman (1950) in central Montana. Although Shaw's terminology is used in this report, it is modified to the extent that the Gallatin is retained as a formation, and the Open Door and Du Noir are classed as members, because they are generally mapped as one unit.

Lithology and Thickness

The Gallatin Limestone is a resistant sequence that forms conspicuous cliffs above the shale slopes of the Gros Ventre Formation. It forms the lowermost series of ledges in many canyons. The formation is 265–365 feet thick in the Wind River Range. In the Washakie Range, and Owl Creek and southern Bighorn Mountains the minimum thickness is about 235 feet and the maximum is 455 feet (pl. 3), although Darton (1906a, p. 14, pl. IV–b) described a section in Owl Creek Canyon (T. 8 N., R. 3 W.) in which the Bighorn Dolomite apparently rests directly on shale of the Gros Ventre Formation with no intervening Gallatin. Along the south edge of the Wind River Basin the formation can be traced from Sweetwater Canyon eastward only as far as the Conant Creek area, where it is 209 feet thick. Farther east the Du Noir Limestone Member at the base merges with clastic sediments of the Gros Ventre Formation (pl. 4). In the eastern and southeastern parts of the basin the Gallatin Limestone is entirely cut out by the erosional unconformity at the base of the Madison Limestone.

Along Warm Spring Creek, at the northwest end of the Wind River Range, the Du Noir Limestone Member is a prominent 48-foot cliff-forming unit (fig. 7) of gray thin-bedded highly glauconitic and oolitic limestone with some flat-pebble conglomerate. The overlying lower shale unit of the Open Door Limestone Member is 54 feet thick and consists chiefly of soft greenish-gray shale with a minor amount of thin-bedded gray limestone. The upper 263 feet of the Open Door is a hard, resistant series of gray thin-bedded to massive limestone and a few flat-pebble conglomerate beds. The uppermost part of this member is commonly mottled with tan small irregular masses of granular limestone. Some of the exposed surfaces weather to a rough pitted surface similar to the overlying Bighorn Dolomite.

The Gallatin Limestone in many places contains cavities filled with red earthy deposits which, where exposed on the sides of the cliffs, contribute much red staining to the outcrops. Because of this, the outcrops, when viewed from a distance, appear to contain much red material.

The characteristic lithologies of the two members of the Gallatin are readily distinguished in most other sections southeast along the east flank of the Wind River Range. Contacts between members are not drawn through some of the localities shown on plate 2, but this is chiefly because of poor exposures and (or) the lack of adequate observations. The Du Noir Limestone Member is 82 feet thick in Sweetwater Canyon at the southeast end of the range, which is the maximum observed for this region. The overlying shale unit of the Open Door at the same locality (pl. 2) is only 23 feet thick and contains a few thin white sandstone and red siltstone beds in addition to limestone.

Shaw (1957, fig. 3) has traced the Du Noir Limestone Member as far east as the Conant Creek area along the south margin of the Wind River Basin, where he assigned 110 feet of limestone and sandstone to the unit. Farther east, the limestone wedges out into clastic strata of the Gros Ventre Formation and the member becomes unidentifiable (pl. 4).

The subdivisions of the Gallatin Limestone are easily recognized in the Washakie Range. The Du Noir Limestone Member can be traced eastward along the entire south side of the Owl Creek Mountains and into the southern Bighorn Mountains, but there the limestone disappears because of facies change (Woodward, 1957, fig. 4). The Open Door Limestone Member also becomes progressively more shaly and sandy eastward (pl. 3).

In subsurface sections in the western part of the Wind River Basin the Gallatin Limestone is easily identified from electric logs (pl. 6).

Contact With the Bighorn Dolomite

An erosional unconformity separates the Gallatin Limestone from the Bighorn Dolomite. In the Warm Spring Creek area, at the northwest end of the Wind River Range, Miller (1936, p. 131) noted 100 feet of beds at the top of the Gallatin on Warm Spring Mountain (W½ sec. 35, T. 42 N., R. 108 W., pl. 1) that are absent 3 miles farther west. At the southeast end of the range, Bell (1955) observed that the contact relief is as much as 3 feet, and that debris from the uppermost Cambrian rocks is incorporated in the Lander Sandstone Member of the Bighorn. In Owl Creek Canyon, at the northwest corner of the Owl Creek Mountains, the Gallatin was apparently cut out by pre-Bighorn erosion (Darton, 1906a, p. 14). In the Wind River Canyon area, Tourtelot and Thompson (1948) reported a sharp uneven contact at the top of the Gallatin and fragments of limestone locally in the basal 3–5 feet of the overlying Bighorn. In many other places, however, the formation contact shows little evidence of the considerable length of time (Late Cambrian to Middle or Late Ordovician) that transpired between the deposition of the Gallatin and that of the Bighorn Dolomite.

Both the Gallatin and the Bighorn commonly form cliffs, but the massive character of the Bighorn is generally in sharp contrast to the more pronounced banded appearance of the Gallatin (fig. 9). The formation boundary is easily detected on electric logs (pl. 6).

In the eastern Owl Creek and southern Bighorn Mountains and in the Conant Creek area along the south edge of the basin, the Gallatin Limestone is overlain directly by the Madison Limestone.

Age

Fossils collected in the Wind River Basin and adjacent regions (table 2; Miller, 1936, p. 140, 141; Shaw, 1957, p. 14, 15) indicate the following zonation for the Gallatin Limestone: Du Noir Limestone Member—*Cedaria*, *Crepicephalus*, and possibly *Aphelaspis* zones; Open Door Limestone Member—*Aphelaspis* and *Elvinia* zones in the lower shale unit and *Elvinia* and other zones of Late Cambrian and possibly earliest Ordovician age in the upper limestone unit. Thus, the formation is largely, if not wholly, Late Cambrian in age.

Regionally, Lochman-Balk and Wilson (1958, p. 334) have described an extensive faunal change between the *Aphelaspis* and *Elvinia* zones. They attribute this change to widespread emergence and shallowing of the seas across much of the North American cratonic and miogeosynclinal areas. The boundary between these two zones corresponds closely to the contact between the Du Noir and Open Door Limestone Members of the Gallatin. As noted by Shaw (1957, p. 14), and as discussed below, sedimentary features suggest a break in sedimentation between these two units in the western part of the Wind River Basin.

Conditions of Deposition

Before the Gallatin Limestone was deposited the sea had apparently withdrawn from the western part of the Wind River Basin area, but it had remained in the central and eastern parts. At the beginning of the depositional period represented by the Du Noir Limestone Member, the entire region was again submerged and carbonate sediments accumulated in the central and western parts of the basin area and clastic sediments in the eastern part. The flat-pebble conglomerate in the upper part of this member, and the shale in the overlying Open Door Limestone Member, suggest that regression, and perhaps emergence, took place at this time. In Late Cambrian time, which is represented by the upper limestone of the Open Door Limestone Member, the sea expanded eastward beyond any of its previous limits and extended across at least the northern half of Wyoming and into the Black Hills region. Cavities filled with red earthy material at some horizons within the Open Door Limestone Member may indicate that parts of central Wyoming were periodically emergent during Late Cambrian time.

At the end of Cambrian time, or shortly thereafter, the sea withdrew completely from central Wyoming and the entire region was exposed to erosion. There is, however, little evidence of this widespread regression in the uppermost Cambrian rocks.

ORDOVICIAN ROCKS, BIGHORN DOLOMITE

Definition

The Bighorn Dolomite, defined by Darton (1904, p. 394–396) from typical exposures along the east flank of the Bighorn Mountains, includes all the rocks of Ordovician age in central Wyoming. Because it invariably forms cliffs, the Bighorn is perhaps the most conspicuous Paleozoic formation in the mountains along the west and north sides of the Wind River Basin (figs. 4, 9).

In the type area, Darton (1906b, p. 26, 27) recognized three lithologic subdivisions of the Bighorn (which he called the Bighorn Limestone): a lower light-gray sandstone, a middle massive buff limestone, and an upper thin-bedded white to gray limestone containing, locally, a red-weathering fossil-rich zone at the top. Later work by Kirk (1930, p. 460–465) in the southern Bighorn and eastern Owl Creek Mountains indicated that the basal sandstone is in reality two units separated by an erosional unconformity. Kirk concluded that (1) the lowermost sandstone, which contains abundant fish remains, does not extend south and west of the eastern end of the Owl Creek Mountains and is the equivalent of the Middle Ordovician Harding Sandstone of central Colorado; and (2) the upper sandstone, which has few, if any, fish remains but a large invertebrate fauna, is of much wider distribution and forms the base of the Bighorn proper. Thomas (1952, p. 35) and Ross (1957a, p. 17) concur in this interpretation.

In the Wind River Range the basal sandstone (post-Harding Sandstone equivalent) of the Bighorn Dolomite was called the Lander Sandstone Member by Miller (1930, p. 196) from outcrops along the Middle Popo Agie River southwest of Lander (near the Sinks Canyon section shown on pl. 2). The upper thin-bedded dolomite was referred to the Leigh Dolomite Member by Blackwelder (1918), the name being derived from the canyon of that name in the Teton Range of northwestern Wyoming. The middle massive dolomite has not received formal designation.

Lithology and Thickness

The Lander Sandstone Member occurs only locally along the east flank of the Wind River Range (pl. 2). It is a lenticular gray to tan, in part red, fine- to coarse-grained sandstone generally less than 5 feet thick. Fragments of the underlying Gallatin Limestone are commonly incorporated in the basal part. Because the member is thin and lenticular and easily concealed by talus, it may have a wider distribution than is apparent from known outcrops.

The unnamed middle member of the Bighorn Dolomite is a remarkably uniform sequence of very resistant cliff-forming buff to light-gray massive granular dolomite. The lower part is rather limy in places and commonly darker gray, but it still retains its granular texture. The dolomite weathers to a very rough pitted surface (fig. 10) that consists of a randomly branching network of slightly raised ridges a fraction of an inch in height and width. In Dinwoody Canyon, J. F. Murphy (oral commun., 1963) found a conglomerate lens, 1 inch thick, 11 feet above the base of the massive dolomite, which contains pebbles of igneous and metasedimentary rocks ⅛–¼ inch across. Farther south, in Bull Lake Canyon, Murphy and others (1956) described beds of dolomite breccia at the top of the member, and cave fillings of thin-bedded dolomite in the upper part. Thickness of the middle member ranges from about 230 feet in the central and northern Wind River Range to only about 15 feet in the southern part. The decrease in thickness southward is the result of beveling at the top by Mississippian and (or) Devonian strata, and perhaps the result in part of nondeposition.

The Leigh Dolomite Member consists chiefly of white, gray, and pink thin-bedded to platy dolomite. The dolomite is dense, porcelaneous, and weathers chalky

FIGURE 10.—Weathered surface of Bighorn Dolomite showing typical "fretwork" pattern. Outcrop along east side of Horse Creek, south-central Washakie Range.

white. It generally forms a conspicuous recessed white band along the cliff faces. Beds of massive dolomite similar to those in the underlying middle member of the Bighorn are in the upper part of the Leigh at many localities. The member is 85 feet thick at the northwest end of the Wind River Range, but thins southward and disappears, owing to the erosional unconformity at the top.

The middle and upper members of the Bighorn Dolomite, which have a combined thickness of 210–300 feet, are readily identified in the Washakie Range; the Lander Sandstone Member is only locally present, but in some sections it is as much as 7 feet thick. In the Circle Ridge area the Bighorn is 260 feet thick, but because of the unconformity at the top, it thins progressively eastward to a wedge edge at the east end of the Owl Creek Mountains (fig. 11; Thomas, 1952, p. 35), approximately at the east edge of T. 40 N., R. 92 W. Sandy beds at the base of the formation in Wind River Canyon and in the eastern Owl Creek Mountains probably represent the Lander Sandstone Member rather than the Harding Sandstone equivalent. The Bighorn Dolomite is not present in the Badwater and Deadman Butte sections in the southern Bighorn Mountains (pl. 3), but it reappears beneath the Madison Limestone a few miles farther north and thickens along the east flank of the range toward the type area.

The Bighorn Dolomite is absent along the east and south margins of the Wind River Basin (fig. 11), although in the Conant Creek area the lower part of the Madison Limestone is a hard ragged-weathering siliceous dolomite that has been mistaken for the Bighorn.

Contact With Overlying Rocks

The Bighorn Dolomite is overlain unconformably by the Darby Formation in the central and northern parts of the Wind River and Washakie Ranges, and western Owl Creek Mountains and by the Madison Limestone in the eastern Owl Creek Mountains (fig. 9) and southern Wind River Range. Along Warm Spring Creek in the northern Wind River Range the erosional unconformity at the top of the Bighorn is marked by earthy beds having layers and cavities filled with calcite crystals. The upper contact has as much as 20 feet of relief and at one place in the north wall of Warm Spring Canyon a large mass of breccia is present in the top of the Bighorn Dolomite. As discussed below, this breccia may represent a channel filled with Devonian sediments similar to those that have been found in the Dinwoody Canyon and Bull Lake Canyon areas. The contact is generally irregular and shows several feet of local relief at many other localities; in some places, however, there is little physical evidence of the long hiatus between the Bighorn and next overlying strata. The Darby Formation is more easily eroded than the Bighorn, and generally forms slopes above the massive dolomite cliffs.

Where the Bighorn Dolomite is overlain by the Madison Limestone, the contact is likewise sharp and irregular in most places. At the south end of the Wind River Range, Bell (1955) observed that a weathered surface occurs at the top of the Bighorn and that vertical fractures or channels filled with red siltstone extend down several feet into the underlying dolomite. Where the basal part of the Madison is dolomitic, however, the contact is not always apparent.

Age

The Bighorn Dolomite contains sparse poorly preserved fossils, including corals, mollusks, and crinoids, in the western and northern parts of the Wind River Basin (table 3). Probably the most characteristic form is the chain coral *Catenipora* (formerly a subgenus of *Halysites*). The largest collections of fossils were obtained by Miller (1930). At the type locality of the Lander Sandstone Member (Sinks Canyon section, pl. 2), for example, he (Miller, 1930, p. 198–201) found a total of 135 different species, mostly mollusks. Along the southeast edge of the Bighorn Mountains (northeast of the Deadman Butte section shown on pl. 3), Amsden and Miller (1942) collected several species of conodonts from both the Harding Sandstone equivalent and the next younger Lander Sandstone Member.

There seems little doubt that the Harding Sandstone equivalent is Middle Ordovician. However, despite the variety of fossils found in the younger parts of the Bighorn Dolomite, including the Lander Sandstone Member, agreement has not yet been reached regarding the exact stages of Ordovician time that are represented by the formation. The problems have been discussed in many papers (for example, Ross, 1957b; Thomas, 1952, p. 35; 1948, p. 83; Love, 1939, p. 20). No new information for dating the formation was obtained during this investigation; the main issues are summarized below.

Darton (1906b, p. 29) considered the upper part of the Bighorn Dolomite in the Bighorn Mountains as Late Ordovician (Richmond) and the lower part as Middle Ordovician (Trenton), based on fossil identifications by E. O. Ulrich. This interpretation prevailed in central Wyoming until Miller (1930, 1932) found that 25 of the relatively diagnostic species from the Lander Sandstone Member seem to indicate Middle Ordovician (Trenton and possibly Black River) age and 22 Late Ordovician (Richmond) age. He (Miller, 1930, p. 203, 204; 1932, p. 203, 209) concluded that the appearance of Richmond fossils is sufficient to fix the age of the lower part of the Bighorn Dolomite (Lander Sandstone Member) as very Late Ordovician. Subse-

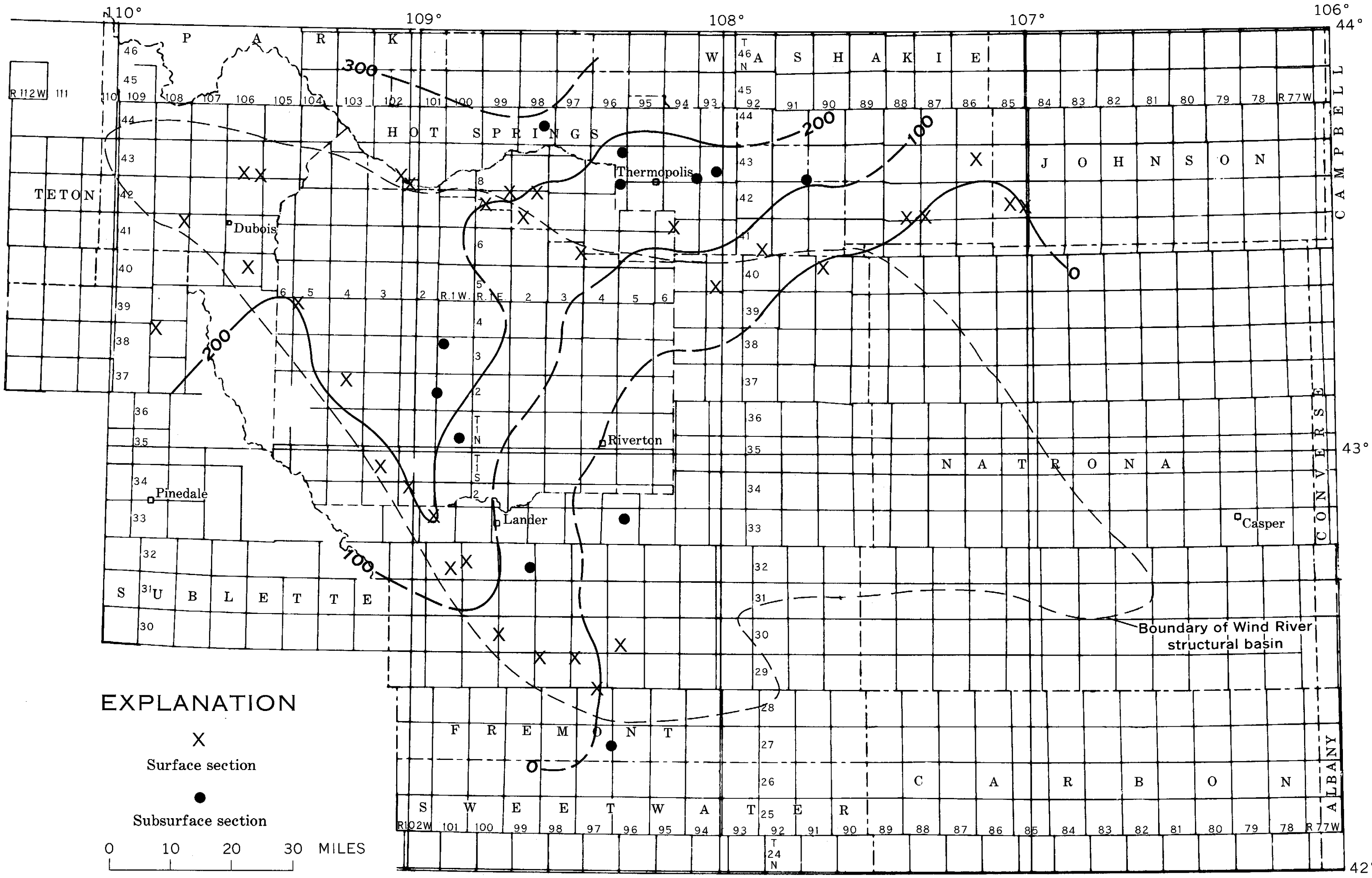


Figure 11.—Thickness of map of Bighorn Dolomite in central Wyoming. Interval is 100 feet; isopachs are restored across mountain arches where rocks have been eroded; isopachs are dashed where control is inadequate.

TABLE 3.—*Fossils from the Bighorn Dolomite*

[Generic and specific names are quoted without modification from sources shown]

Stratigraphic section	Collection No. (pls. 2-5)	Species	References
Dinwoody Canyon	4	*Catenipora gracilis, Streptelasma* aff. *S. foerstei, Grewinkia* sp., *Calapoecia* sp., *Nyctopora* sp., *Zygospira* cf. *Z. jupiterensis, Paucicrura* sp., *Sowerbyella* or *Thaerodonta* sp., *Platystrophia* sp., *Hesperorthis* sp., *Cyclendoceras*? sp.	R. J. Ross, Jr., to J. F. Murphy (written commun., 1955).
	5	*Streptelasma* sp., *Euomphalopterus*? sp., *Fusispira* cf. *F. inflata, Halysites gracilis, Drepanodus subrectus, Panderodus gracilis.*	J. W. Huddle to J. F. Murphy (written commun., 1963); Eliot Blackwelder (U.S. Geol. Survey, unpub. field notes); identification by E. O. Ulrich.
Bull Lake	1	*Halysites* (*Catenipora*) cf. *H. gracilis, "Holophragma"* sp.; unidentifiable streptelasmoid coral, brachiopods, and cephalopods.	Murphy and others (1956); J. W. Huddle to J. F. Murphy (written commun., 1963).
	1a	*Drepanodus curvatus, Oistodus* sp., *Panderodus* cf. *P. gracilis, Cordylodus* sp., *Ozarkodina* sp. A, *Ozarkodina* sp. B, *Ligonodina* sp., *Trichonodella*? sp., *Belodina compressus, Prioniodina* sp.	
Sinks Canyon	1	Composite fauna from the Lander Sandstone Member in Sinks Canyon and vicinity, collected and described by Miller. The following species are included: *Receptaculites*, 1; *Cyclocrinites*, 1; coral, 7; worm, 1; brachiopods, 27; pelecypods, 13; gastropods, 34; cephalopods, 50; trilobites, 1.	Miller (1932, p. 204–206).
	2	Composite fauna from the middle massive dolomite and Leigh Dolomite Members in the central Wind River Mountains, collected and described by Miller. Includes *Receptaculites arcticus, Streptelasma corniculum, S. robustum, S.* sp., *Columnaria alveolata, C. halli, Palaeophyllum stokesi, Calapoecia borealis, C. canadensis, Halysites gracilis, Dinorthis*? sp., *Rhynchotrema capax, Catazyga* cf. *C. headi borealis, Maclurina*? sp., *Liospira*? sp., *Allumettoceras*? sp., *Spyroceras rarum, Trochonema umbilicatum.*	Miller (1932, p. 206, 207).
Sweetwater Canyon	9	*Receptaculites* sp., *Halysites gracilis, Halysites* cf. *H. microporus, Hesperorthis*? sp., *Resserella tersa*?, *Rhynchotrema increbescens*?, tetracorals indet.	Bell (1955).
Windy Gap	2	*Calapoecia* sp., *Streptelasma* sp., *Receptaculites* sp., *Halysites* sp.	Love (1939, p. 20);
	3	*Catazyga* sp., abundant crinoid stems	Keefer (1957; p. 167).

quent investigations of North American Ordovician cephalopods by Flower (1946, 1952) have shown that forms typical of the Bighorn occur both in rocks of the Upper Ordovician Series in southern Ohio and northern Kentucky and in beds of definite late Middle Ordovician age in New York and Ontario.

Reporting on more recent fossil discoveries at several horizons within the Bighorn Dolomite in the southeastern Bighorn Mountains, Hose (1955, p. 45–48, 111, 112) concluded, from identifications by Josiah Bridge, that the upper part of the basal sandstone (Lander Sandstone Member) is Middle Ordovician and the upper two members of the formation are Late Ordovician. Ross (1957b, p. 468) considered the fossil-rich red-weathering zone at the top of the formation in the central Bighorn Mountains to be Late Ordovician. Wanless and others (1955, p. 15, 16) expressed the opinion, based on the presence of favositid corals, that the Leigh Dolomite Member in the Teton Range might possibly be of Silurian age, though corals of this type are also known in some places from Upper Ordovician rocks.

From the evidence at hand it seems likely that at least the upper part of the Bighorn is Late Ordovician. Furthermore, in Ross' opinion (1957b, p. 463) "the paleontologic evidence for the Late Ordovician age of the [Lander Sandstone Member] is still impressive." Duncan (1956, p. 220) stated "General investigations of American Ordovician corals have led the writer to believe that the relatively diversified and advanced fauna in the lower part of the Bighorn dolomite and equivalent strata must be indicative of Late—probably early Late—Ordovician age."

Conditions of Deposition

The Ordovician sea first invaded central Wyoming in Middle Ordovician time and deposited sandstone equivalent to the Harding Sandstone in a shore and near-shore marine environment. The original distribution of this sandstone is not known, but Ross (1957a, p. 18) suggested the possibility that the Harding equivalent was deposited across at least part of the Wind River Basin, south and west of its present area of outcrop, and that it was completely removed by later erosion and reworking into the next younger Lander Sandstone Member. Ross (1957a, p. 17) also pointed out, on the other hand, that the irregular occurrence of the Harding Sandstone equivalent may be due as much to irregular original deposition as to any subsequent events. After the deposition of the Harding Sandstone equivalent, the region seems to have been raised above the level of the sea again and to have eroded until later Middle Ordovician time or until Late Ordovician time depending on the age that is assigned to the Lander Sandstone Member.

The deposition of the Lander Sandstone Member marked the beginning of widespread marine sedimentation in central Wyoming. Miller (1930, p. 213) concluded from faunal evidence that a shallow sea transgressed from the Arctic southeastward into the region. The source of the sand is not known. It is known however, that clastic rocks of Cambrian age were exposed in some parts of the region during the transgression of the Ordovician sea. The nearest such area, along the present southeast margin of the Wind River Basin, could have contributed some of the clastic debris in the Lander Sandstone Member. The evidence is inconclusive, however, because much of the record was destroyed by pre-Mississippian erosion. No conclusions can be drawn regarding the irregular distribution and thinness of the Lander; there is no evidence of erosion or reworking of the sands before the carbonate sediments were deposited over them. On the contrary, the contact between the Lander Sandstone Member and the overlying massive dolomite appears to be conformable, and, in many places, gradational (Ross, 1957b, p. 462).

The thick carbonate sequence that forms the main body of the Bighorn Dolomite was probably deposited in a warm shallow sea, but far enough from shore so that very little clastic debris was carried into the main area of accumulation. The factors influencing sedimentation must have been constant for a long time, because the strata over large areas are so uniform in composition, color, and texture. Little is yet known, however, about the time or mode of dolomitization. Blackwelder (1913) was the first to consider this problem in detail; Ross (1957a) summarized some of the more current views on the subject. Both investigators conclude that the material was deposited originally as calcium carbonate; but, whereas Blackwelder (1913, p. 621, 622) believed that the limestone was dolomitized before lithification, Ross (1957a, p. 19) considered it more likely that dolomitization took place after lithification had begun. According to Robinson (1963, p. 23, 24), the possibility that dolomite was deposited directly from sea water should not be dismissed, even though dolomite is not being deposited in quantity in modern open seas.

The peculiar branching structures that are so characteristic of exposed surfaces of the Bighorn Dolomite (fig. 10) are clearly the result of differential weathering related to variations in the physical and (or) chemical composition of the dolomite. Darton (1906c) believed that the raised parts of the network are siliceous and hence less easily weathered than the surrounding dolomite. Blackwelder (1913, p. 615, 624) pointed out that the rock contains very little silica and concluded that the network is more likely the result of differential weathering of compact fine-grained (more resistant) dolomite embedded in a matrix of more coarsely crystalline and more porous (less resistant) dolomite. He (Blackwelder, 1913, p. 624) further concluded that the branching structures are of organic origin, possibly calcareous algae whose original organic structures were obliterated in the process of dolomitization. Ross (1957a, p. 18) cited evidence, on the other hand, that the structures could not have been secreted by algae above the ocean floor and suggested that the raised parts of the network were either minute solution channels or passages of burrowing organisms filled by carbonate mud.

The presence of the small pebbles of igneous and metasedimentary rocks in the basal part of the middle massive dolomite member in Dinwoody Canyon is enigmatic; so far as is known this is their only occurrence, but they could have been easily overlooked in other outcrops. The pebbles, though somewhat rounded, show little sign of having been extensively weathered before reaching their ultimate depositional site (J. F. Murphy, oral commun., 1963), indicating, perhaps, that transport had not been for any great distance. There is no evidence, however, to suggest that any part of the Wind River Range was emergent and being eroded during deposition of the Bighorn. Some parts of the Granite Mountains might possibly have been eroded to the Precambrian at this time, but these mountains lie 75 miles southeast of Dinwoody Canyon (pl. 1).

Much additional study is needed before more reliable conclusions can be made regarding the origin of the Bighorn Dolomite.

At the close of Ordovician time the sea withdrew from central Wyoming and a period of erosion ensued, possibly lasting through all of the Silurian Period and part of the Devonian Period.

DEVONIAN ROCKS, DARBY FORMATION

Definition

The name Darby Formation was applied by Blackwelder (1918) to a variable sequence of dolomite and shale lying between the Madison Limestone and Bighorn Dolomite in northwestern Wyoming. Blackwelder (p. 420) derived the term from the canyon of Darby Creek on the west slope of the Teton Range, but he gave a measured section from Sheep Mountain on the west slope of the Wind River Range (about 20 miles south of the Warm Spring Creek loc., pl. 1) as the type. He considered the Darby to be Devonian and equivalent to the Three Forks Shale and the upper part of the Jefferson Limestone in southwestern Montana as described by Peale (1893, p. 27–32).

In the Wind River Basin the Darby Formation, as mapped, likewise includes all strata lying between the Bighorn Dolomite and Madison Limestone. Regional stratigraphic studies by Andrichuk (1951), Sandberg (1963, oral commun., 1964), and Sandberg and McMannis (1964) have shown, however, that this sequence is very complex, and that it includes strata correlative with the Beartooth Butte Formation of Early Devonian age, the Maywood and Jefferson Formations of Late Devonian age, and a dark shale unit of latest Devonian and earliest Mississippian age.

Strata assigned to the Darby Formation have the most limited distribution of any of the Paleozoic units in the Wind River Basin, being restricted to the flanks of the northern and central Wind River Range, the Washakie Range, and western half of the Owl Creek Mountains (fig. 12).

Lithology and Thickness

The Darby Formation is 193 feet thick at the northwestern end of the Wind River Range, but thins progressively southeastward to a wedge edge near Sinks Canyon (pl. 2). Except in canyon walls, these rocks are poorly exposed, commonly forming topographic saddles or benches between the resistant Madison Limestone and Bighorn Dolomite. Where well exposed, however, they form a very distinctive unit because of their contrast in color and in bedding characteristics with both the overlying and underlying strata.

In the northwestern Wind River Range the Darby consists chiefly of resistant buff, gray, and brown thin- to thick-bedded ledgy dolomite in the lower part and greenish-gray and red soft siltstone, shale, and sandstone interbedded with white, pink, and brown limestone and dolomite in the upper part. The most characteristic beds are dark-brown granular dolomite which emits a strong petroleumlike odor when freshly broken; hence, the term "fetid" is commonly applied to them. The granular texture of some beds is so conspicuous that they are often mistaken for sandstone. The dolomites are readily distinguished from the carbonate rocks in the Bighorn Dolomite below and the Madison Limestone above, although some of the lower beds of the Madison are superficially similar in places. The sandstone of the Darby is fine to coarse grained and commonly contains abundant large rounded frosted grains. Such grains are also scattered through some of the dolomite beds.

The Darby Formation, as mapped in exposures extending southeast from Warm Spring Creek to Bull Lake, includes at the top 5–30 feet of thin-bedded siltstone, shale, and silty dolomite. According to Sandberg (1963), this unit is in part equivalent to a dark shale unit of latest Devonian and earliest Mississippian age that is widespread in northern Wyoming and southern Montana and that overlies all older Devonian strata with marked regional unconformity. In parts of Montana these strata were assigned to the basal beds of the Madison Limestone (Sloss, 1952, p. 66) rather than to the underlying Three Forks Shale.

In the Bull Lake Canyon and Dinwoody Canyon areas Murphy and others (1956) reported local channels of sandy to conglomeratic siltstone and dolomite at the base of the Darby that have cut into underlying Bighorn Dolomite. Some of the channels contain Early Devonian fossils and represent the Beartooth Butte Formation, whereas others contain early Late Devonian fossils and correlate with the Maywood Formation (C. A. Sandberg, oral commun., 1964). The large masses of breccia at the top of the Bighorn Dolomite at the Warm Spring Creek locality (see discussion above) may also represent either one or both of these formations, but no fossils have been collected.

The Darby Formation has been traced southeast along the east flank of the Wind River Range as far as Sinks Canyon (pl. 2), where about 20 feet of tan crystalline dolomite imbedded with large fragments of white medium- to coarse-grained quartzitic sandstone was reported by Strickland (1957, p. 27). Sections along the north edge of the Wind River Basin are similar to those in the Wind River Range; the sequence becomes more sandy and finally wedges out in the central Owl Creek Mountains (pl. 3). Strickland (1957, p. 25) noted that toward the limits of its distribution, the lower part of the sequence seems to be overlapped by the upper part. Correlation of the upper strata—as seen in sections near the wedge edges—with the dark shale unit of the northwestern Wind River Range is yet to be determined. Maughan (1963, p. C26) discussed the possibility, however, that clastic rocks assigned to the basal part of the Madison Limestone in the northern Laramie Mountains are in part equivalent to the dark shale unit.

Contact With the Madison Limestone

The dark shale unit seems to be transitional with the basal beds of the Madison Limestone in outcrops extending from Bull Lake Canyon northwest to Warm Spring Creek (C. A. Sandberg, oral commun., 1964). Farther east and south, where the dark shale unit is absent or poorly developed, an erosional unconformity may separate the Madison and Darby Formations, but

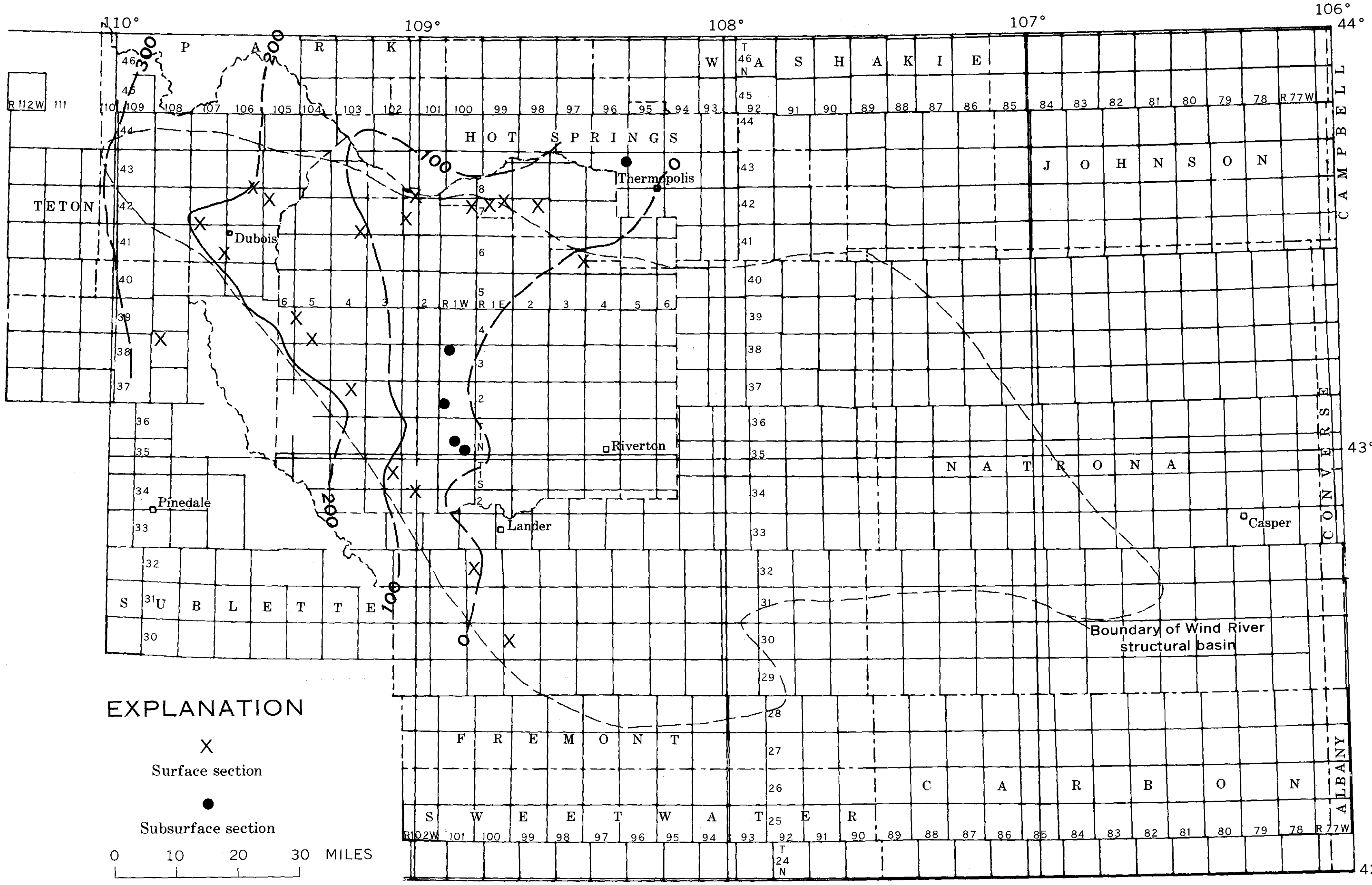


FIGURE 12.—Thickness map of Darby Formation in central Wyoming. Based in part on C. A. Sandberg (written commun., 1963). Interval is 100 feet; isopachs are restored across mountain arches where rocks have been eroded; isopachs are dashed where control is inadequate.

in most places observations have been too inadequate to determine the nature of the contact in detail.

Age

Fish remains have been found in channellike deposits at the base of the Darby Formation and associated strata in Bull Lake Canyon and nearby sections along the east flank of the Wind River Range (table 4; Denison, 1951, p. 234). These fossils indicate that the lowermost channels are Early Devonian in age, whereas the next overlying ones are early Late Devonian (Sandberg and McMannis, 1964; C. A. Sandberg, oral commun., 1964). Conodonts from the dark shale unit in Dinwoody and Bull Lake Canyons are latest Devonian (table 4; Klapper, 1958, p. 1083) and those from Warm Spring Lake are earliest Mississippian (C. A. Sandberg, oral commun., 1964). Most of the rocks assigned to the Darby Formation are, therefore, Late Devonian in age.

Conditions of Deposition

No positive record that rocks of Silurian age were deposited in the Wind River Basin area has been found. However, Silurian strata have been recognized in the northern part of the Powder River Basin (C. A. Sandberg, oral commun., 1963) and also in isolated patches in southeastern Wyoming (Chronic and Ferris, 1961); thus, Silurian seas may have covered at least parts of central Wyoming.

The region was emergent during most of Early and Middle Devonian time, although minor flooding occurred in at least the western part of the Wind River Basin area during the early part of the period. In early Late Devonian time the sea advanced eastward, and perhaps southward, into the region, first filling only narrow channelways but later covering all the land surface. However, the sea at this time probably did not extend much beyond the present known limits of the Darby Formation (fig. 12). Thomas (1948, p. 84) concluded, from the abundance of frosted sand grains in the formation in sections near the wedge edge, that the Darby in central Wyoming was deposited fairly close to shore and that the sand was blown into the sea. Presumably, therefore, a broad landmass extended across the eastern half of the Wind River Basin area during this part of Late Devonian time.

The sea withdrew from central Wyoming before the end of Devonian time, and a period of intense erosion followed. Thus, much of the erosion commonly attributed to the period immediately preceding the deposition of the Madison Limestone may have occurred instead before the end of the Devonian. In latest Devonian and earliest Mississippian time, the sea readvanced and the dark shale unit was laid down across the truncated edges of rocks ranging in age from Late Devonian in western and northern Wyoming (Sando and Dutro, 1960, p. 123; Sandberg, 1963, p. C19) to Precambrian in east-central Wyoming (C. A. Sandberg, oral commun., 1963). Sedimentation seems to have been continuous into Madison time in at least the western part of the Wind River Basin area, but the sea may have withdrawn from the central and eastern parts for a short interval before Madison deposition.

MISSISSIPPIAN ROCKS, MADISON LIMESTONE

Definition

The Madison Limestone was named and described by Peale (1893, p. 32) for the Madison Range in the Three Forks region of southwestern Montana. The age of the formation in the type area was designated as "Lower Carboniferous." In southeastern Idaho, Mansfield (1927, p. 60–71) subdivided the Mississippian rocks into the Madison Limestone at the base and the Brazer Limestone (named earlier by Richardson, 1913) at the top; the ages were considered to be Early Mississippian and Late Mississippian, respectively.

The name Madision has long been used to designate all the rocks of Mississippian age in central Wyoming (Darton, 1906a, b; Blackwelder, 1918). However, as early as 1918 Branson and Greger (1918) described fos-

TABLE 4.—*Fossils from the Darby Formation*

[Generic and specific names are quoted without modification from sources shown]

Stratigraphic section	Collection No. (pls. 2–5)	Species	References
Dinwoody Canyon	6	Fish remains	J. F. Murphy (U.S. Geol. survey, unpub. field notes).
Bull Lake Canyon	2	*Bothriolepis* plates; teeth and scales of unidentifiable crossopterygians	Do.
	3	*Apatognathus varians, Hindeodella* cf. *H. acuta, Icriodus darbyensis, I. mehli, Icriodus* sp., *Neoprioniodus* sp., *Ozarkodina regularis, Palmatolepis gonioclymeniae, P. rugosa, Polygnathus inornata, P. nodocostata, P. triangularis, P. varinodosa, P. wyomingensis, Spathognathodus aculeatus, S. jugosus.*	Klapper (1958).

sils of Late Mississippian age from the upper part of the formation in Bull Lake Canyon. The lower approximately 50 feet of this sequence, bearing what was considered to be a Ste. Genevieve fauna, was named the Sacajawea Formation, and the overlying beds, believed to be of possible Chester age, were included in the Amsden Formation by C. C. Branson (1937; see fig. 13). More recently other geologists have attempted to subdivide the Mississippian sequence in the Wind River Basin area; the various classifications, many of which follow the terminology established in central Montana (Collier and Cathcart, 1922, p. 173; Sloss, 1952; Easton, 1962) are shown on figure 13.

None of the proposed classifications, however, has been demonstrated to be wholly satisfactory for geologic mapping; many are based on lithologic and (or) paleontologic differences that may be easily recognized locally but not regionally. Denson and Morrissey (1952), for example, separated the Madison on the basis of insoluble residues, but stated (p. 40) that "Divisions of the Madison formations are frequently recognized only with difficulty in the field." Therefore, in this report all Mississippian rocks above the dark shale unit of the Darby Formation and below the Amsden Formation are included in the Madison Limestone because they constitute a natural lithologic unit for geologic mapping. For convenience in stratigraphic and paleontologic discussion, however, the formation has been divided into upper and lower members.

In the summer of 1964, W. J. Sando collected fossils and studied the stratigraphy of the Mississippian rocks at several localities along the east flank of the Wind River Range. The preliminary results of his work were made available to us during the final stages of preparation of the present report (W. J. Sando, written commun., 1965).

Lithology and Thickness

The Madison Limestone includes 600–700 feet of resistant cliff-forming carbonate strata in the central and northern Wind River Range and adjacent parts of the Washakie Range. The lower member, 500–600 feet thick, consists mainly of bluish-gray to gray massive to thin-bedded crystalline limestone and dolomitic limestone. The limestone is mottled by inclusions of tan granular limestone or dolomite similar to those in the Gallatin Limestone and the Death Canyon Limestone Member of the Gros Ventre Formation. At most localities the basal part contains thin beds of buff granular dolomitic limestone similar to some of the beds in the upper part of the Darby Formation. The lower strata also contain masses of breccia consisting of angular limestone fragments in a red earthy matrix; weathering and erosion of the breccia causes conspicuous red staining on many outcrops. Massive to bedded chert layers as much as 15 feet thick are present at several localities, and chert nodules are abundant throughout. Caverns are common in the limestone, especially in the more granular beds.

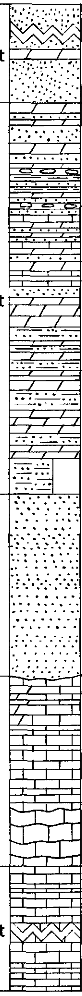
The upper member of the Madison Limestone, about 100 feet thick, is chiefly thin to massive and irregularly bedded gray, tan, and yellowish-tan dolomite and limestone. Thin beds of red shale and siltstone are locally present. The carbonate rocks in some zones are highly fractured, brecciated, and cavernous and contain much red clastic material. Conspicuous disconformities also occur at a few horizons.

The limestone and dolomite vary in lithology and texture; these rocks have been described in some detail by Andrichuk (1955) and Flanagan (1962).

Thickness of the Mississippian sequence ranges from 235 to 535 feet in the southern Wind River Range (pl. 2) and from 335 to 635 feet in the Owl Creek and southern Bighorn Mountains (pl. 3). Along the south edge of the Wind River Basin the thickness is 175–400 feet, along the east edge it is 245–275 feet (pls. 4, 5). Madison strata generally thin from northwest to southeast across central Wyoming (fig. 14).

Lithologies similar to those described above characterize the Madison Limestone in other areas around the Wind River Basin. The lower member is everywhere present, although Denson and Morrissey (1952, figs. 7, 8) indicated that the basal part (their Lodgepole Formation) is absent along the south margin. Tourtelot and Thompson (1948) observed lenses of sandstone at the base of the lower member in the Wind River Canyon area and, at the southeast edge of the basin, Jenkins (1950) included about 60 feet of quartzitic sandstone and conglomerate at the base. As noted above, these clastic rocks, as well as those that have been assigned to the underlying Flathead Sandstone, are all included by Maughan (1963, p. C26) in the basal part of the Madison Limestone in the northern Laramie Mountains.

The distribution of the upper member of the Madison Limestone is more restricted than that of the lower member, owing to post-Madison erosion; its exact limits have not yet been worked out in detail. Denson and Morrissey (1952, fig. 4) showed the upper member to be present in the Wind River Canyon area, whereas Andrichuk (1955, fig. 5) and Sando (written commun., 1965) do not extend it that far east in the Owl Creek Mountains. Correlation in the southern Wind River Range is also in doubt, but the unit may extend as far southeast as Sinks Canyon, or beyond. East of Sweetwater Canyon, however, the Amsden Formation seems to rest directly on the lower member of the Madison, and the upper member is absent.



This report		Murphy and others (1956)			Branson (1937)		Branson and Branson (1941)	Strickland (1957)			Denson and Morrissey (1952)		Andrichuk (1955)		Flanagan (1962)
Tensleep Sandstone		376 ft	Tensleep Sandstone		Tensleep Sandstone		Tensleep Sandstone	(Not studied)			(Not studied)		(Not studied)		(Not studied)
Amsden Formation	Upper part	197 ft	Amsden Formation		Amsden Formation	Upper Amsden									
	Darwin Sandstone Member	92 ft		Darwin Sandstone Member											
Madison Limestone	Upper member	96 ft	Brazer(?) Limestone equivalent			Lower Amsden (Beds of possible Chester age)	Beds of Chester(?) age	Madison Group	Upper Madison	Sacajawea Member	Madison Group	"Upper" Madison	Madison Group	Upper unit	"Upper" Madison
					Sacajawea Formation		Sacajawea Formation								
	Lower member	580 ft	Madison Limestone		Madison Limestone		Madison Limestone			(Unnamed part)		Mission Canyon and Lodgepole Formations		Middle and lower units	"Middle" and "lower" Madison
									Lower	Lodgepole and Mission Canyon					

FIGURE 13.—Nomenclature of Carboniferous strata, east flank of Wind River Range. Section shown is from Bull Lake Canyon.

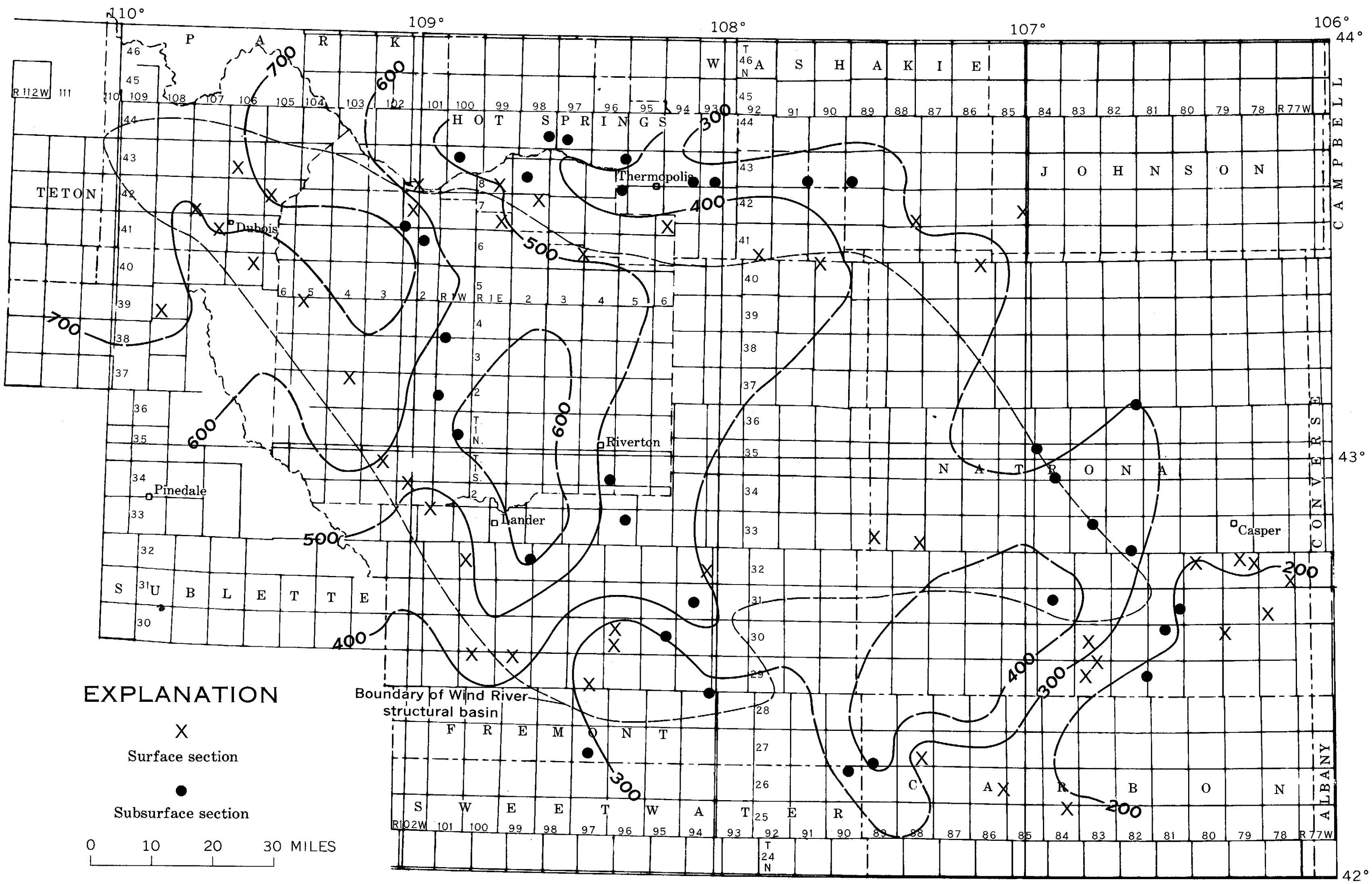


Figure 14.—Thickness map of Madison Limestone in central Wyoming. Interval is 100 ft; isopachs are restored across mountain arches where rocks have been eroded; isopachs are dashed where control is inadequate.

Basal Unconformity

The unconformity at the base of the Madison Limestone is one of the most prominent in central Wyoming; however, as stated above, much of the truncation may have taken place in latest Devonian time, before the deposition of the dark shale unit of the Darby Formation. The general distribution of various formations that directly underlie the Madison across the region is shown on figure 15. In the western part of the Wind River Basin the formation overlies the Darby Formation with apparent conformity, but eastward it rests on successively older beds and, in the southeast part of the basin, directly overlies the Cambrian Flathead Sandstone. The contact shows several feet of relief in many places, and basal strata of the Madison commonly contain sandstone and siltstone in which fragments reworked from the underlying formations are incorporated. In the Bates Creek section (pls. 4, 5) the basal quartzitic unit was probably derived principally from the Flathead Sandstone.

Contact With the Amsden Formation

The Madison Limestone is separated from the overlying Amsden Formation by a conspicuous erosional unconformity. The basal strata of the Amsden, generally sandstone, were deposited on a very irregular surface that shows several feet of local relief. The sandstone commonly fills cracks which extend downward into the underlying limestone. The sharp lithologic change between the Madison and Amsden Formations is readily recognized in most outcrops, and the contact is also easily identified on electric logs (pl. 6).

Age

The Madison Limestone is sparsely to abundantly fossiliferous in many zones (table 5). Recent paleontologic studies by Sando (written commun., 1965) indicate that the lower member is largely late Early Mississippian (late Osage) in age, and that the upper member is early Late Mississippian (early Meramec) in age; thus, the bulk of the formation seems to correlate with the Mission Canyon Limestone of southwestern Montana. Basal strata of the lower member may be equivalent in part to the Lodgepole Limestone (Kinderhook and early Osage); as noted previously, C. A. Sandberg (oral commun., 1964) has observed evidence that the contact between the Madison and Darby Formations is transitional in places. Sando found no fossils to indicate that any of the rocks here assigned to the Madison are as young as latest Mississippian (Chester), as had been suggested by Branson (1937).

Conditions of Deposition

In Early Mississippian time the sea may have withdrawn from all but the western third of the Wind River Basin area; strata as old as the Flathead Sandstone were exposed to erosion in the southeastern part. The sea advanced from the north and northwest at the beginning of Madison deposition, and a thick succession of carbonate sediments were laid down across the entire

TABLE 5.—*Fossils from the Madison Limestone*

[Generic and specific names are quoted without modification from sources shown]

Stratigraphic section	Collection No. (pls. 2–5)	Species	References
Dinwoody Canyon	7	*Syringopora* sp	Thomas (1948, fig. 3).
Bull Lake Canyon	4	*Zaphrentis amsdenensis, Spirifer pellaensis, S. shoshonensis, Composita trinuclea, Eumetria marcyi, Pugnoides ottumwa, Spiriferina browni, Orthotetes kaskaskiensis, Chonetes chesterensis, Tetracamera subcuneata, Phillipsia* sp.? *Homalophyllites*?, *Caninophyllum*?, *Lithostrotion*?, *Diphyphyllum*?.	Branson and Greger (1918, p. 312); Murphy and others (1956); Helen Duncan (written commun., 1954).
Sinks Canyon	4	*Triplophyllites* sp., *Cyathophyllum* sp., *Syringopora* sp., *Diphyphyllum* sp., *Barbouria* sp.	Biggs, C. A. (1951).
Beaver Creek	1	*Spirifer* cf. *S. madisonensis*	Shaw (1955, p. 63).
Sweetwater Canyon	10	*Syringopora surcularia, Spirifer* sp	Bell (1955).
	11	*Spirifer madisonensis*?, *S. centronatus*, "*Orthotetes*" sp. indet	Do.
Windy Gap	4	*Lithostrotion*? sp., *Syringopora* sp., *Diphyphyllum*?, *Spirifer striatus* var. *madisonensis, Spirifer* cf. *S. increbescens, S.* sp., *Composita* sp. indet., crinoid columnals, fenestellid bryozoans.	Love (1939, p. 26); Keefer (1957, p. 169).
Conant Creek	4	*Spirifer centronatus*?	V. L. White (U.S. Geol. Survey, unpub. field notes).
Alcova	1	*Camarotoechia* aff. *C. sappho, C.* sp., *Spirifer* aff. *S. keokuk, Spiriferina solidirostris*?, *Cleiothyridina crassicardinalis.*	Lee (1927, p. 52).
	2	*Productus parviformis, Spirifer centronatus, Spiriferina solidirostris, Cleiothyridina crassicardinalis.*	Do.
Bates Creek	1	*Spirifer centronatus, Composita* sp. indet	Jenkins (1950).
Casper Mountain	1	*Spirifer centronatus, Syringopora surcularia, Composita humilis*	Lee (1927, p. 48).

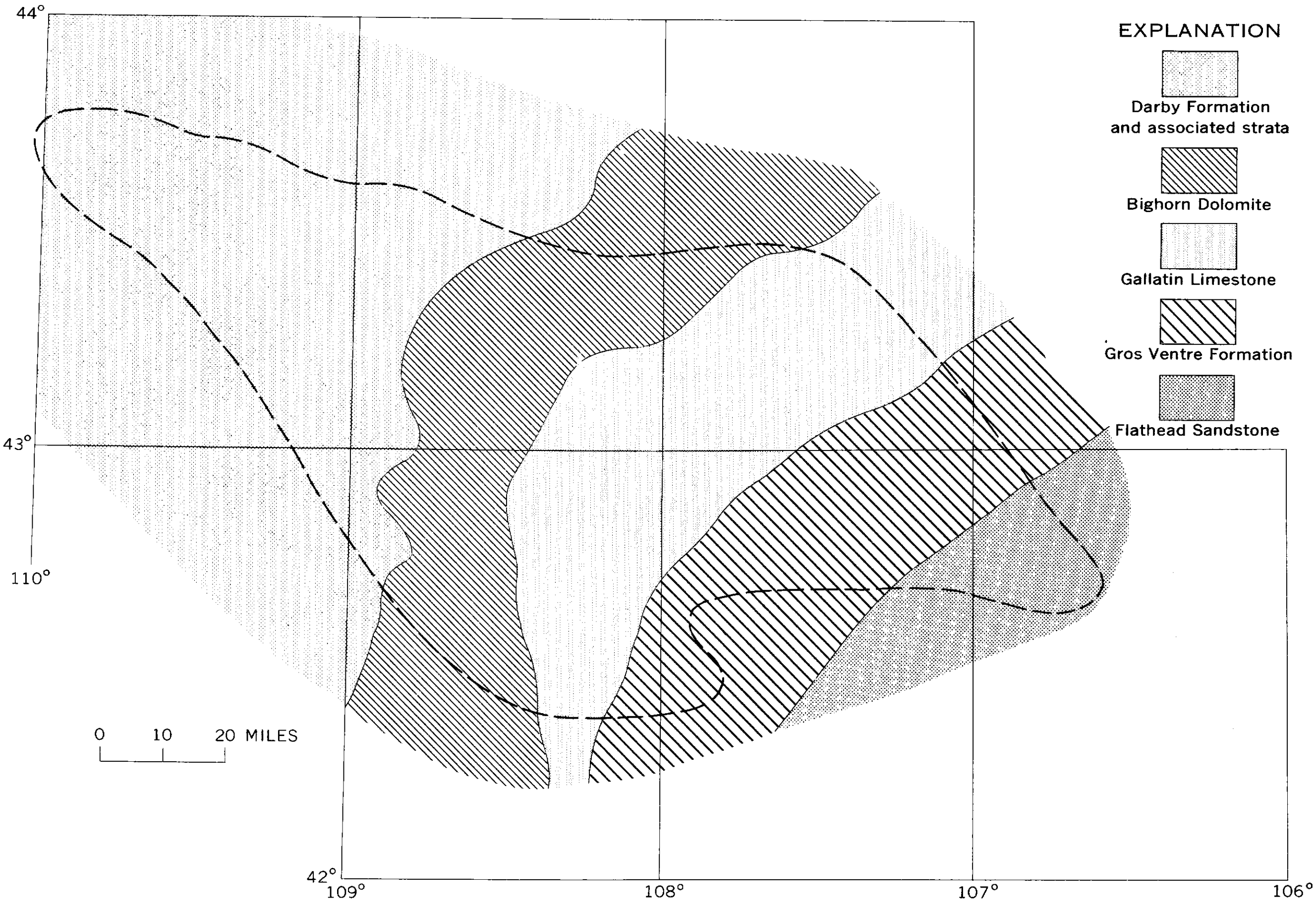


FIGURE 15.—General distribution of formations underlying Madison Limestone in central Wyoming. Dashed line indicates approximate outline of Wind River structural basin.

region. Variations in thickness and type of carbonate facies deposited at any given time or place probably reflect minor tectonic movements and consequent changes in sea level. In the eastern and southeastern parts of the basin, debris from the underlying rocks was reworked into the basal strata of the Madison, but it was spread only a short distance into the main basins of depostion to the north and west.

Toward the end of the depositional period represented by the lower member of the Madison, evaporitic sediments accumulated in south-central Montana. Andrichuk (1955, p. 2198) attributed this accumulation to extreme shoaling conditions which hindered free circulation of marine waters. As the circulation became more restricted, particularly in areas of deeper water surrounded by shoals, the water became progressively more saline and eventually evaporites were precipitated. Conditions did not favor evaporite deposition farther south in northern and central Wyoming during Early Mississippian time, but shoaling across the major shelf area is indicated by the presence of oolitic and fragmental limestone and dolomite in some places (Andrichuk, 1955, p. 2198).

Marine sedimentation also took place during early Late Mississippian time in at least the western and northwestern parts of the Wind River Basin area. To what extent the sea may have covered the eastern and southeastern parts of the basin during this part of the period is not known. If the Late Mississippian sea did extend into the southeastern part of this region, the sediments deposited were probably thin and hence destroyed by post-Madison pre-Amsden erosion. Evaporites accumulated in westernmost Wyoming (Sando and Dutro, 1960, p. 124; Wanless and others, 1955, p. 28, 29) and in northern Wyoming (Andrichuk, 1955, p. 2204) in Late Mississippian time. These deposits do not extend into central Wyoming, but the conspicuous brecciated and cavernous zones in the upper member of the Madison Limestone may represent evaporite beds that were leached during periods of emergence, probably in Late Mississippian time.

PENNSYLVANIAN ROCKS

AMSDEN FORMATION [2]

Definition

The Amsden Formation was named and described by Darton (1904, p. 396) from exposures in the northern part of the Bighorn Mountains; in it he included all the strata lying between the typical blue-gray limestone of the Madison and the massive crossbedded sandstone of the Tensleep. In the type area he described a red shale sequence, normally with a basal sandstone, in the lower part of the formation and interbedded limestone and sandy and cherty beds in the upper part. Darton (1906a, b) traced these units throughout the Bighorn and Owl Creek Mountains. Blackwelder (1918) identified the Amsden Formation in the Gros Ventre Range and northern and central Wind River Range, and named the basal sandstone, which is thick in this region, the Darwin Sandstone Member from Darwin Peak in the Gros Ventre Range, about 25 miles southwest of the Warm Spring Creek locality (pl. 2). Darton and Blackwelder believed the formation to be largely Pennsylvanian, but both suggested that the lower part might be Mississippian.

Although the Amsden Formation can be recognized with certainty throughout most of central Wyoming, an anomalous situation at the southern end of the Wind River Range has given rise to considerable confusion and controversy regarding its age and correlation. In 1918, Branson and Greger (1918) reported a Late Mississippian fauna from red ferruginous shale and sandstone overlying the Madison Limestone south of the canyon of the Little Popo Agie River (west edge of T. 31 N., R. 99 W.). They (Branson and Greger, 1918) referred these beds to the Amsden Formation, but because the exposures are poor (the fossils were collected from slope wash), they presented a measured section of the Upper Mississippian rocks (upper member of the Madison Limestone of this report) in Bull Lake Canyon as a lateral equivalent, noting (p. 310) that "The iron, sandstone, and shales of the [Amsden in the] Popo Agie region do not appear [in Bull Lake Canyon] and the rock is mainly dolomite and limestone." They failed to recognize that the prominent sandstone overlying the Upper Mississippian rocks in the Bull Lake area was actually the basal sandstone of the Amsden (Darwin) rather than the Tensleep Sandstone. This miscorrelation probably arose for two reasons: (1) the red shale beds are not present in Bull Lake Canyon; and (2) the Darwin Sandstone Member is either absent or poorly represented at the locality south of the Little Popo Agie River, and the first massive sandstone upward in the section is the Tensleep Sandstone.

[2] Since the preparation of this report, W. J. Sando (written commun., 1965) has collected fossils from the Amsden Formation in the western Wind River Basin which indicate that the lower part is Late Mississippian in age; see discussion of age of Amsden Formation.

C. C. Branson (1937) later referred the basal part of the Upper Mississippian sequence (bearing a Ste. Genevieve fauna) in the Bull Lake region to the Sacajawea Formation and the overlying limestone of possible Chester age to the "lower Amsden." The "upper Amsden" (fig. 13), according to Branson, "consists of 80 feet of sandstone [Darwin Sandstone Member] overlain by 96 feet of limestone, sandstone, and shale." Still later it was proposed (Branson, 1939; Branson and Branson, 1941) that the name Amsden be abandoned in this region, and that all strata of Pennsylvanian age be referred to the Tensleep Sandstone. No indication is given that the authors recognized the Darwin. The controversial nature of these proposals has fostered extensive investigation and discussion of the Amsden Formation in central Wyoming. Details may be found in many published papers, including Love (1939, p. 23, 24; 1954), Thomas (1948, p. 86, 87), Burk (1954, p. 2–4), Andrichuk (1955, p. 2178), Shaw and Bell (1955), Keefer (1957, p. 172), Strickland (1957, p. 23–25), and Easton (1962, p. 21, 22), and in unpublished theses by Biggs (1951), Bell (1955), Gorman (1962), and Bearce (1963).

The problems regarding the age and correlation of the Amsden Formation have been compounded because the Darwin Sandstone Member is unfossiliferous, because it is poorly exposed in critical areas in the southern Wind River Range, and because it thins southward and cannot be identified specifically in the vicinity of Little Popo Agie canyon where Branson and Greger (1918) first reported Late Mississippian fossils from the lower part of the Amsden. In 1955, however, Shaw and Bell (1955) trenched a section of the controversial red beds along Cherry Creek (sec. 19, T. 31 N., R. 99 W.), south of the Little Popo Agie canyon, described the rocks in detail, and made extensive fossil collections (fig. 16). It is apparent from Shaw and Bell's study that the controversial section, about 70 feet thick, contains Pennsylvanian fossils in the upper part and Mississippian fossils in the lower part. However, they (Shaw and Bell, 1955) believed that the Mississippian part correlated temporally with the upper member of the Madison Limestone at Bull Lake, whereas Sando (written commun., 1965) interpreted the fossils from Cherry Creek to be younger than those at Bull Lake. (See discussion below of age relations within Amsden Formation.)

Lithology and Thickness

In the northwestern Wind River Range and adjacent Washakie Range the Amsden Formation ranges in thickness from 210 to 355 feet. The Darwin Sandstone Member at the base consists of cliff-forming red, gray,

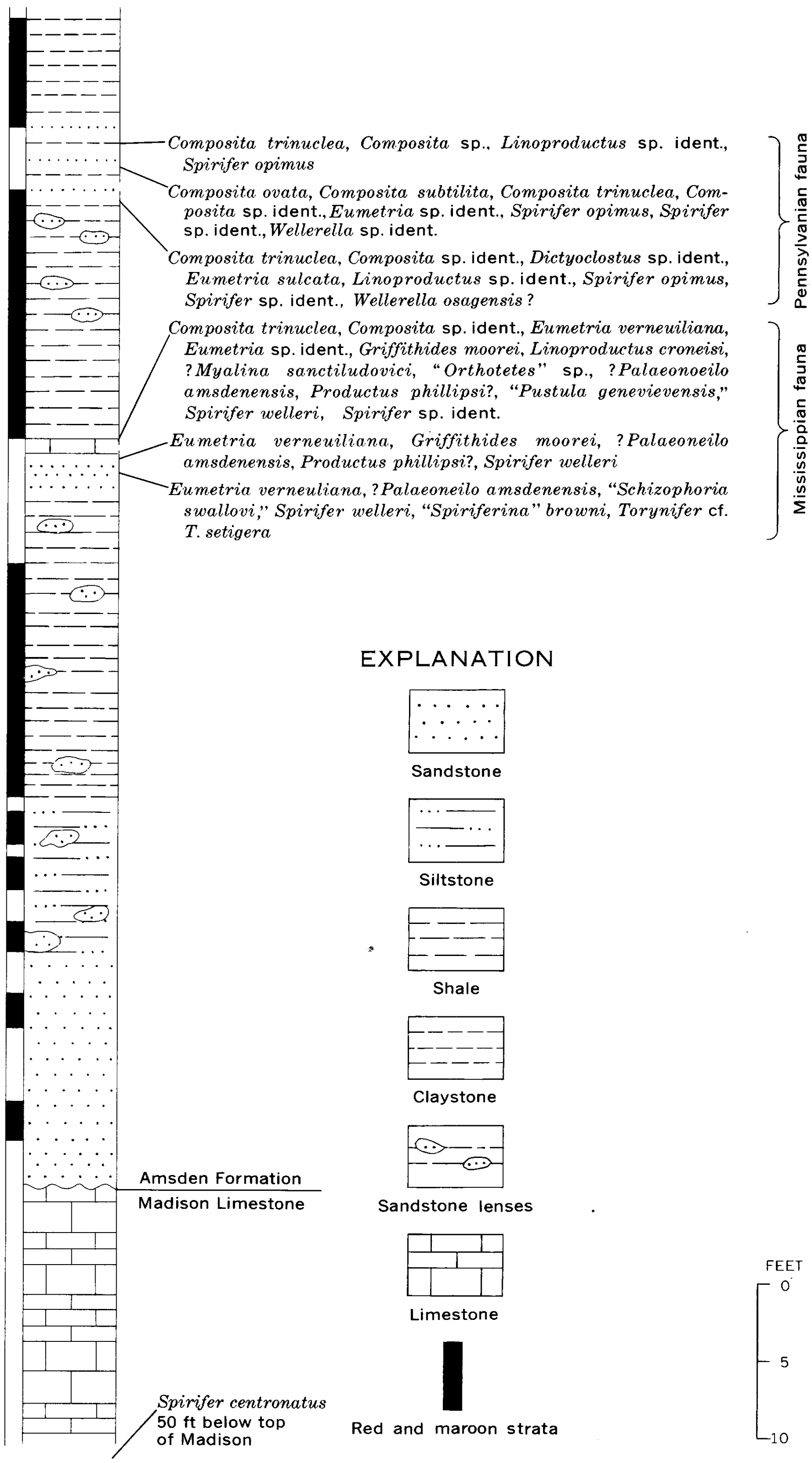


Figure 16.—Section of lower part of Amsden Formation at Cherry Creek, sec. 19, T. 31 N., R. 99 W. Based on Shaw and Bell (1955, p. 334).

and white fine- to medium-grained crossbedded to massive moderately porous and friable sandstone. In lithology and outcrop appearance it is strikingly similar to the Tensleep Sandstone. Thickness of the Darwin has a considerable range, from a minimum of about 30 feet to a maximum of 170 feet. The range in thickness, however, seems to have no uniform trend; it probably is due chiefly to the irregular erosion surface on which it was deposited.

The upper part of the Amsden Formation is a variable sequence of dolomite, shale, sandstone, and limestone ranging in thickness from 180 feet to about 300 feet. Colors most common are red, green, gray, and buff. The unit is generally poorly exposed because most of the strata are nonresistant, but it produces characteristic red, tan, and yellow soils. In many places, directly overlying the Darwin Sandstone Member, is a red shale about 40 feet thick that contains many hematitic nodules or pisolites. It forms a conspicuous red zone along the weathered slopes. Thin beds of sandstone occur mostly in the upper part of the formation and are typically fine grained, hard, and quartzitic. The dolomite and limestone are massive to thin bedded and are generally cherty. Slight disconformities occur locally within this upper sequence.

Southeast along the east flank of the Wind River Range the Amsden Formation is 120 feet to about 400 feet thick. The Darwin Sandstone Member, averaging about 75 feet in thickness, can be traced as a relatively well defined unit as far southeast as Canyon Creek, directly south of Lander in T. 31 N., R. 100 W. There the member is about 30 feet thick, unconformably overlies typical Madison Limestone, and conformably underlies red ferruginous shale. Farther southeast, in sections from the north rim of Little Popo Agie canyon (T. 31 N., R. 99 W.) to Sweetwater Canyon, strata assigned to the lower part of the Amsden Formation are chiefly interbedded red shale, siltstone, and sandstone. A 15- to 30-foot basal sandstone is present in most sections, but it is generally thin bedded and easily eroded, and hence forms a much less distinct unit than the Darwin Sandstone Member farther north. Although many workers have considered this basal sandstone to represent the Darwin, specific correlations of the various sandstone units are still very uncertain.

The upper part of the Amsden Formation can also be traced throughout the central and southeastern parts of the Wind River Range. Toward the southeast end of the range, however, the sequence contains a higher proportion of sandstone and red beds than it does in the central and northwestern parts.

The Amsden Formation is 190–250 feet thick in the Owl Creek and southern Bighorn Mountains. In most sections it can be separated into a basal gray to buff sandstone (generally correlated with the Darwin Sandstone Member), a middle red shale and siltstone, and an upper light-gray, buff, and tan cherty limestone and dolomite. In some places in the southern Bighorn Mountains, however, carbonate rocks are scanty, and the tripartite division cannot readily be made.

Along the south margin of the Wind River Basin the basal sandstone of the Amsden, also generally referred to as the Darwin Sandstone Member, is 20–50 feet thick and is overlain by 80–130 feet of chiefly red and gray shale and siltstone and locally a few thin beds of sandstone and cherty limestone and dolomite. In the Bates Creek and Casper Mountain sections (pl. 5), at the southeast edge of the basin, all Pennsylvanian rocks have customarily been included in the lower part of the Casper Formation. Lee (1927, p. 49) correlated 85 feet of red shale at the base of the Casper on Casper Mountain with the lower red shale of the Amsden Formation. However, it is now known that the lower part of the Pennsylvanian sequence in this area is younger than the Amsden (fig. 18) and hence is not equivalent in a time-stratigraphic sense.

Contact With the Tensleep Sandstone

The contact between the Amsden Formation and the overlying Tensleep Sandstone is marked by a sharp change in topography from weathered talus-covered slopes below to nearly vertical cliffs above. There is little evidence of a sedimentary break between the two formations, except locally (for example, Love, 1939, p. 28), and in many places the contact is gradational where thin beds of sandstone typical of the Tensleep are interbedded with cherty carbonate beds. Agatston (1954, p. 515) proposed that the top of the Amsden be drawn at the base of the lowermost sandstone typical of the Tensleep, even though thin carbonate, shale, and sandstone beds might be included in the Tensleep. Because of the uncertainty of the contact, the Tensleep and Amsden Formations have been mapped as a single unit in some regions, particularly in northwestern Wyoming (Love and others, 1951). For the same reason the two formations are combined on the isopach map (fig. 17).

Age

Fossils have not been found in the Darwin Sandstone Member of the Amsden Formation, but several zones above the Darwin are fossiliferous, particularly along the east flank of the Wind River Range (table 6). Until recently, all the formation above the Darwin was considered to be Pennsylvanian (probably Morrow and Atoka; fig. 18) in age (Burk, 1954, p. 5; Shaw and Bell, 1955, p. 335; Shaw, 1955, p. 62; Agatston, 1957, p. 32; Gorman, 1962). Furthermore, despite the fact that red strata assigned to the lower part of the Amsden in

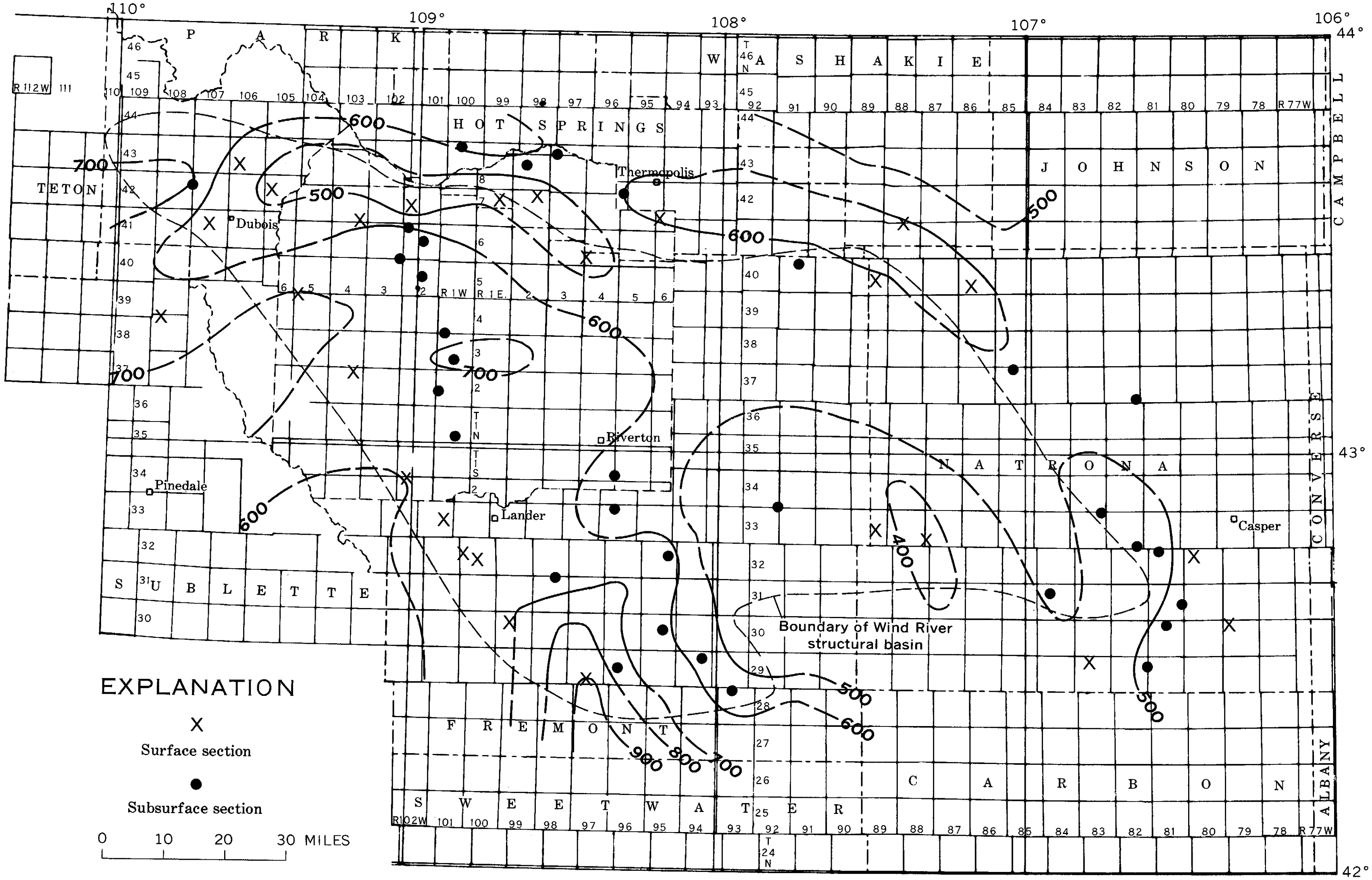


FIGURE 17.—Thickness map of Tensleep and Amsden Formations in central Wyoming. Includes Casper Formation in southeastern part of Wind River Basin. Based in part on W. W. Mallory (written commun., 1963). Interval is 100 feet; isopachs are restored across mountain arches where rocks have been eroded; isopachs are dashed where control is inadequate.

the southern Wind River Range contain Mississippian fossils, many workers believed that the conspicuous erosional unconformity at the base of the Darwin Sandstone Member marked the Pennsylvanian-Mississippian boundary. Such a boundary would correspond to the widespread unconformity between the two systems in most other regions across the continent (Moore and others, 1944, p. 663).

W. J. Sando (written commun., 1965) has obtained evidence to show that the above reasoning is erroneous; fossil collections obtained by him from a locality along Horse Creek north of Dubois (near Windy Gap section, pl. 3) indicate that the systemic boundary actually occurs within the red shale sequence above the Darwin. Sando therefore interprets the Darwin Sandstone Member and overlying red shale (at least as much as 73 ft

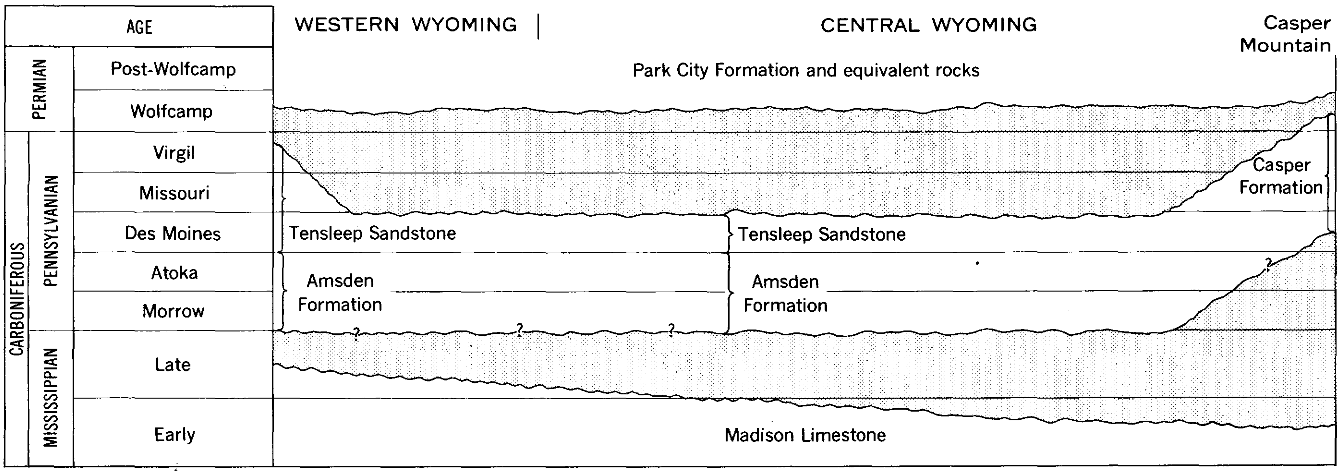


FIGURE 18.—Schematic diagram showing approximate intervals of time represented by Mississippian and Pennsylvanian formations in central and western Wyoming. Interpreted from data in Thomas and others (1953), Love (1954), and Agatston (1954).

TABLE 6.—*Fossils from the Amsden Formation*

[Generic and specific names are quoted without modification from sources shown]

Stratigraphic section	Collection No. (pls. 2–5)	Species	References
Dinwoody Canyon	8	*Pteria* sp., *Astartella* sp., *Entomis* sp., *Paraparchites*? sp.	Eliot Blackwelder (U.S. Geol. Survey, unpub. field notes); identification by G. H. Girty.
Bull Lake Canyon	5	*Climacammina* sp., *Endothyra* sp., *Bradyina* sp., *Textrataxis* cf. *T. millsapensis*, *Fusulinella gephyraea*?, *Spirifer* cf. *S. opimus*, *Spirifer* cf. *S. rockymontanus*, *Neospirifer*? n. sp., *Phricodothyris*? sp. indet., *Composita subtilita*, *Mesolobus striatus*, *Antiquatonia* n. sp., *Marginifera splendens*, *Punctospirifer*? sp. indet., *Linoproductus prattenianus*, *Pharkidonotus percarinatus*, *Plagioglypta*? sp.	Murphy and others (1956); Henbest (1956, p. 61); Love (1954).
Beaver Creek	2	*Orbiculoidea wyomingensis*, *Wellerella* n. sp., *Cancrinella boonensis*?, *Cleiothyridina*? sp., *Linoproductus* n. sp., *L. croneisi*, "*Marginifera*" n. sp., *Spirifer welleri*, *Allorisma*.	Burk (1954, p. 6–7); modified by Shaw (1955, p. 63).
	3	*Composita* sp. (not like *Composita* in the *Spirifer welleri* zone below it)	Shaw (1955, p. 63).
Sweetwater Canyon	12	*Spirifer occidentalis*, *Composita* sp	W. G. Bell (1955).
Windy Gap	5	*Spirifer* "*opimus*," *Linoproductus* n. sp., *Composita* "*trinuclea*," *Bellerophon* sp., *Strophostylus*? sp.	Love (1939, p. 28); Shaw (1955, p. 62).
	6	*Orthotetes*? sp., *Dictyoclostus portlockianus*, *Linoproductus prattenianus*, *Spirifer* sp., *Composita subtilita* var. *subtilita*, *C. subtilita* var. *trinuclea*, *C. subtilita* var. *ovata*, *C.* sp., *Cleiothyridina* sp., *Wellerella osagensis*, *Allorisma terminale*.	Burk (1954, p. 7–8).
	7	*Fusulinella* sp	Love (1939, p. 28); Shaw (1955, p. 62).
Conant Creek	5	*Chaetetes* cf. *C. milleporaceous*	Love (1954).
East Canyon Creek	1	*Allorisma* cf. *A. terminale*	Do.
Alcova	3	*Productus cora*, *P. hermosanus*, *Composita subtilita*, *Parallelodon*? sp., *Deltopecten*? sp., *Pteria* sp., *Plagioglypta*? sp., *Bellerophon crassus*?.	Lee (1927, p. 52).
	4	*Allorisma terminale*, *Pteria longa*?, *Deltopecten aviculatum*, *Parallelodon*? sp., *Schizodus*? sp., *Pleurophorus*? sp., *Bellerophon* n. sp., *Euphemus carbonarius*?, *Naticopsis* aff. *N. nana*.	Do.

above the top of the Darwin) to be latest Mississippian (Chester) in age, and infers partial or complete correlation with the controversial red-bed sequence at the base of the Amsden Formation along Cherry Creek in the southern Wind River Range. As stated above, previous workers (Branson and Greger, 1918; Morey, 1935; Shaw and Bell, 1955) had considered this latter sequence to be early Late Mississipian (Meramec) in age and hence equivalent to the upper member of the Madison Limestone at the Bull Lake Canyon locality. Sando (written commun., 1965), on the other hand, considers the Late Mississippian faunas in the basal part of the Amsden at Cherry Creek to be of Chester age and hence younger than the early Late Mississippian (Meramec) faunas from the upper member of the Madison at Bull Lake. It should be noted that studies of ostracodes from the Mississippian sequence in Illinois (for example, Croneis and Funkhouser, 1938, p. 334; Coryell and Johnson, 1939, p. 214) suggested a Chester age for forms similar to those identified by Morey (1935) as Meramec from the Cherry Creek locality.

Based on the evidence obtained by Sando (written commun., 1965), the following conclusions seem warranted; (1) The Amsden contains strata of both Mississippian and Pennsylvanian ages; (2) the basal beds of the Amsden are approximately the same age throughout the Wind River Range, but are younger than the upper member of the Madison Limestone; and (3) the upper member of the Madison Limestone was cut out by post-Madison pre-Amsden erosion in the southern Wind River Range. In the southern Wind River Range, it is still a matter of conjecture whether the Darwin Sandstone Member of the Amsden Formation disappears by facies change, by nondeposition, or by erosion. Precise correlation of the various units of the Amsden in the Wind River Basin with those of the type section along Amsden Creek in the northern Bighorn Mountains is also problematic. Gorman (1962) concluded, from a study of fusulinids at the type section, that at least the beds immediately overlying the unfossiliferous basal sandstone (generally considered to be a Darwin equivalent) are Pennsylvanian and that the basal sandstone is probably also Pennsylvanian.

Conditions of Deposition

After the Madison Limestone was deposited, central Wyoming emerged, and the exposed carbonate rocks were eroded into a very irregular surface. In the central and southeastern parts of the Wind River Basin area, erosion cut down into Lower Mississippian strata. A broad positive area occupied southeastern Wyoming and extended as far northwest as the region along the southeastern margin of the Wind River Basin (Mallory, 1963, fig. 195.2; Agatston, 1954, fig. 9); in some places its Precambrian core was breached.

At the beginning of Amsden deposition a shallow sea spread across all but the southeastern part of the Wind River Basin, probably from the west and northwest, and an extensive deposit of sand was laid down. Because much of eastern and southeastern Wyoming was underlain either by Mississippian carbonate rocks or Precambrian crystalline rocks at this time, the source of the highly quartzose debris of the Darwin Sandstone Member is not known. W. W. Mallory (oral commun., 1964) suggests that the sand may have come from north and east of Wyoming, but the evidence is not conclusive. Sand deposition was followed by the widespread accumulation of red clay and silt that was spread across the shallow shelf areas from sources also as yet unknown. Presumably the highly ferruginous clay was derived from a region in which regolithic soils were being developed, perhaps in some parts of southeastern Wyoming where Precambrian rocks are known to have been exposed to weathering and erosion at that time.

Fluctuations in sea level, probably in response to minor structural movements within the depositional area, gave rise to an alternating sequence of limestone, dolomite, shale, and sandstone in the upper part of the Amsden Formation. Local disconformities within this sequence, especially in the upper part (for example, Agatston, 1957, p. 29; Gorman, 1962), suggest that certain areas were periodically raised above the level of the sea and briefly eroded; the southeastern part of the basin area remained emergent. The gradational contact between the Amsden and overlying Tensleep indicates that there was no significant withdrawal of the sea at the close of Amsden time.

TENSLEEP SANDSTONE

Definition

The Tensleep Sandstone was defined by Darton (1904, p. 397) as the thick sandstone overlying the Amsden Formation in the southwestern part of the Bighorn Mountains. The formation can be traced throughout central and northwestern Wyoming, although on Casper Mountain at the southeastern edge of the Wind River Basin equivalent rocks are included in the lower part of the Casper Formation (Darton, 1908). Branson (1939) included all the Amsden Formation in the Tensleep Sandstone in the central Wind River Range, but this usage was not widely adopted.

Lithology and Thickness

The massive cliff-forming Tensleep Sandstone is one of the most distinctive Paleozoic formations in the region (figs. 3, 19). Along the east flank of the Wind River Range the formation is predominantly buff, tan, cream-colored, and white fine-grained massive to cross-

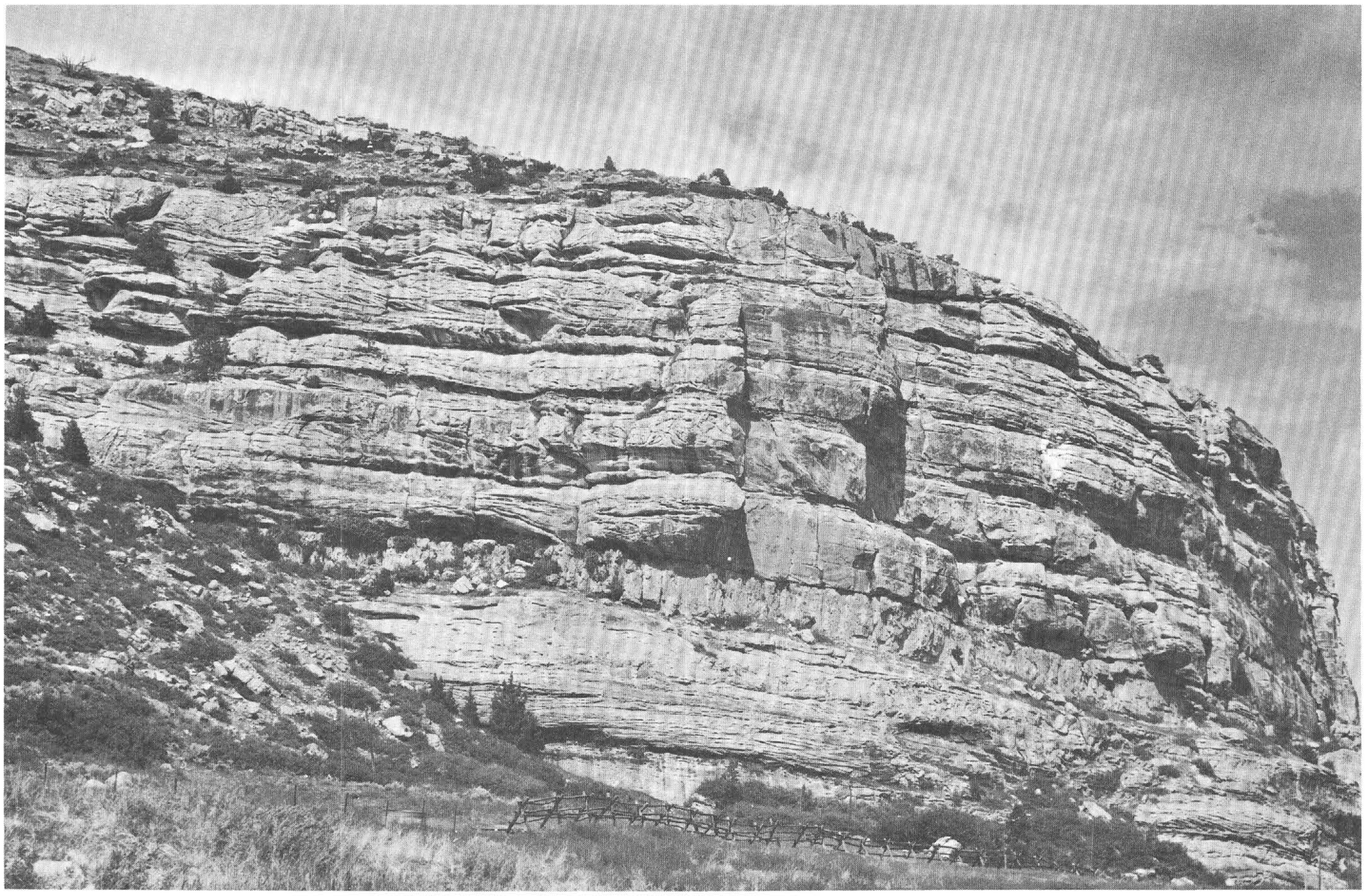


FIGURE 19.—Cliff of Tensleep Sandstone along north side of Middle Popo Agie River, Sinks Canyon. Note crossbedding in left half of photograph.

bedded sandstone (fig. 19). Most of the rock is rather porous and friable, but some beds are hard and quartzitic. Thin irregular beds of chert are common, and because of the gradational contact with the underlying Amsden, the basal part of the Tensleep may include a few thin beds of limestone or dolomite. Carbonate beds also occur higher in the formation. Many of the outcrops weather brown and rusty brown, and, viewed from a distance, some appear to be nearly black. Coarse talus generally accumulates at the base of the cliffs and obscures the lower part of the formation as well as much of the underlying Amsden. The Tensleep Sandstone is only about 215 feet thick at the northwest end of the range, but it thickens gradually to more than 600 feet at the southeast end (pl. 2).

The lithology of the Tensleep Sandstone is fairly uniform. Thickness along the north edge ranges from 200 to 440 feet, the thickest sections being in the eastern Owl Creek Mountains and the southwestern Bighorn Mountains (pl. 3). Along the south margin of the basin the formation is 220–350 feet thick (pl. 4).

In the Bates Creek and Casper Mountain areas (pl. 5), all Pennsylvanian rocks are placed in the Casper Formation (Darton, 1908, p. 418). Sandstone beds generally similar to the Tensleep Sandstone occur in the middle and upper parts of the Casper, although these beds are younger. (See fig. 18).

Contact With Park City and Equivalent Rocks

A conspicuous erosional unconformity separates the Tensleep Sandstone from the overlying Park City Formation and equivalent rocks of Permian age in nearly all exposures around the margins of the Wind River Basin. The contact is distinctly uneven and shows from a few inches to several feet of relief in many places. The basal beds of the Permian sequence are commonly conglomeratic. Blackwelder (U.S. Geol. Survey, unpub. field notes) reported that in the Dinwoody Canyon area as much as 37 feet of sandstone at the top of the Tensleep was cut out underneath the unconformity. The time span represented by the unconformity over much of the basin area includes Late Pennsylvanian and earliest Permian times. The contact is easily defined on electric logs (pl. 6).

Age

Fossils have been collected from several different zones within the Tensleep Sandstone at many scattered localities in central Wyoming (table 7). They indicate that most, if not all, of the formation within the Wind River Basin area is Middle Pennsylvanian (Des Moines) (Love, 1954; Agatston, 1954, p. 529; Henbest, 1954; Hoare and Burgess, 1960). Love (1954) suggested that in westernmost Wyoming the upper part is Late Pennsylvanian (Missouri and Virgil). In the Casper Mountain area the lowermost part of the Casper Formation may be Middle Pennsylvanian (Agatston, 1954, table 5), but fusulinids of Late Pennsylvanian age have been found near the base and in the middle part of the sequence, and Early Permian (Wolfcamp) fossils in the upper part (Thomas and others, 1953, fig. 2 and table 2). An interpretation of the regional age relations is shown on figure 18.

TABLE 7.—*Fossils from the Tensleep Sandstone*

[Generic and specific names are quoted without modification from sources shown]

Stratigraphic section	Collection No. (pls. 2–5)	Species	References
Warm Spring Creek	8	*Derbyia crassa, Linoproductus prattenianus, Dictyoclostus hermosanus, Spirifer opimus, Neospirifer cameratus, Composita* sp., *Straparolus (Euomphalus)* cf. *S. (E.) plummeri, Dentalium (Plagioglypta)* sp., *Ameura* sp., *Millerella*? sp., fusiform fusulinid, either *Fusulina* sp., or *Triticites* sp.	Hoare and Burgess (1960, p. 711–715).
Dinwoody Canyon	9	*Derbyia*? sp., *Productus cora, P. semireticulatus, P.* sp., *Spirifer rockymontanus, Composita subtilita, Aviculipecten* sp., *Pseudomonotis hawni*?, *Pleurophorous*? sp., *Dentalium* sp., *Euphemus carbonarius, Bellerophon percarinatus, Pleurotomaria* sp.	Eliot Blackwelder (U.S. Geol. Survey, unpub. field notes); identification by G. H. Girty; Branson (1939, p. 1210).
Trout Creek	1	*Calcitornella* sp., *Climacammina* sp., *Endothyra* sp., *Bradyina* sp., *Fusulina leei*?, *Wedekindellina* sp.	Henbest (1954, p. 52); Love (1954).
Sinks Canyon	5	*Dictyoclostus hermosanus, Linoproductus prattenianus, Squamularia perplexa, Composita subtilita.*	Branson (1939, p. 1208).
Wind River Canyon	12	*Idiognathodus* sp., *Streptognathodus* sp., *Prioniodus* sp., *Hindeodella* sp., *Helodus* sp., *Cladodus* sp., *Fusulina* n. sp., *Staffella* sp.	Branson (1939, p. 1207).
Bates Creek	2	Fusulinids	C. E. Jenkins (1950).
Casper Mountain	2	*Triticites kellyensis*, projected into section from about 6 miles farther east.	Thomas and others (1953).

In the Mayoworth area along the east flank of the Bighorn Mountains, about 40 miles northeast of the Deadman Butte section (pl. 1), Verville (1957) described fusulinids and Rhodes (1963) described conodonts, of Wolfcamp age in sandy limestone beds which they assigned to the uppermost part of the Tensleep Sandstone. According to E. K. Maughan (oral commun., 1963), however, these strata are lithologically similar to, and probably correlate with, the Nowood Member of the Phosphoria Formation (McCue, 1953) farther west in the southwestern Bighorn Mountains (pl. 3) and hence form no part of the Tensleep as we have used it in the Wind River Basin area.

Conditions of Deposition

Following the deposition of the Amsden Formation a shallow sea remained in central Wyoming, although some moderately unstable areas were at times raised above the base level of deposition, thereby producing local disconformities within the boundary zone between the Tensleep and Amsden Formations. At the beginning of Middle Pennsylvanian (Des Moines) time, quartz detritus was spread across the basin of deposition. The source of the sand is not apparent; Agatston (1954, p. 564) suggested that it might have been supplied by uplifts west and south of Wyoming. Thin beds of limestone, thickening toward the main area of carbonate deposition in eastern Wyoming (Agatston, 1954, p. 564), were deposited along with the sandstone as the sea advanced and retreated. The only land that stood above the sea in central Wyoming was in the Casper Mountain area, where sedimentation did not occur until later in late Middle Pennsylvanian or Late Pennsylvanian time. In southeastern Wyoming a broad positive area furnished sediment until it was submerged in Early Permian time (Agatston, 1954, p. 565).

If sedimentation occurred in the Wind River Basin, exclusive of the Casper Mountain area, during Late Pennsylvanian time, all record of it was destroyed by post-Tensleep erosion. The distribution of thickness and facies within the Pennsylvanian sequence (Love, 1954; Agatston, 1954), however, suggests that (1) much of central Wyoming was raised above the level of the sea at the beginning of Late Pennsylvania time; (2) the sea retreated eastward into eastern Wyoming and westward toward the major geosynclinal area in southeastern Idaho; and (3) detritus eroded from the exposed Tensleep Sandstone was redeposited in basins of deposition that flanked the uplift. Marine sedimentation continued along the southeast edge of the Wind River Basin and areas to the east through all Late Pennsylvanian time, and in westernmost Wyoming at least until near the close of that epoch.

PERMIAN ROCKS

NOMENCLATURE

Rocks of Permian age comprise one of the most complex Paleozoic systems in central Wyoming. Because it contains extensive phosphate deposits and is one of the most important reservoirs for oil and gas in the region, this system has also been one of the most widely studied. Marked lithologic changes occur from west to east across the Wind River Basin, and many problems exist in nomenclature and correlation. The regional relations, summarized below, have been discussed in detail by Thomas (1934), Tourtelot (1952), McCue (1953), McKelvey and others (1956; 1959), and Weichman (1958).

Darton (1906a, p. 17, 18) applied the name Embar Formation to a "prominent series of limestone and chert beds lying between the Tensleep sandstone and the Chugwater red beds" along the north slope of the Owl Creek Mountains. The type section is about 6 miles north of the Sheep Ridge section (pls. 1, 2). Darton (1906b, p. 35, 36) extended the term into the Bighorn Mountains, but limited it to a lower limestone and shale unit only and did not include any of the overlying red shale and gypsum. Condit (1916, p. 263), however, recognized that the marine strata graded eastward from the type area into red shales containing many gypsum beds. He (p. 263) stated "In the Bighorn Mountains * * * the transformation of the Embar beds is so complete that they can only with difficulty be distinguished from the overlying red beds composing the Chugwater formation [Triassic]."

In an early report on the northern Wind River Range, Blackwelder (1911, p. 475) used the name Embar, but he later (1918) separated the sequence into two formations. The upper unit, consisting of greenish-gray shale, siltstone, and sandstone now considered to be of Triassic age, he called the Dinwoody Formation from exposures in Dinwoody Canyon. The lower part, predominantly limestone, dolomite, chert, and phosphate rock, he referred to the Park City Formation, a name used earlier by Boutwell (1907, p. 443) in the Park City mining district of Utah. Condit (1924, p. 11) concluded that the beds assigned to the Park City by Blackwelder were more nearly correlative with the Phosphoria Formation as that unit had been defined by Richards and Mansfield (1912) in southeastern Idaho. The Phosphoria was then widely accepted in the central and western parts of the Wind River Basin for the rocks lying between the Tensleep Sandstone and the Dinwoody Formation. Limited use of the name Embar continued, but the term has long since been considered obsolete (Thomas, 1934, p. 1670, 1671) and was abandoned (McKelvey and others, 1956, p. 2835).

In 1947 the U.S. Geological Survey began a systematic investigation of the western phosphate field, which includes western Wyoming, northwestern Utah, eastern Idaho, and southwestern Montana. This program resulted in a comprehensive study and review of the stratigraphy, nomenclature, and correlation of the Permian rocks (McKelvey and others, 1956, 1959). It was proposed that the sequence be divided into three major lithologic facies, each being assigned to a different formation; the general distribution of the various facies is shown on figure 20. Because considerable intertonguing occurs between the different rock types, the individual members of one formation in any given vertical section may be directly overlain or underlain by members of another formation. (See Dinwoody Canyon section, pl. 2.) Therefore, for geologic mapping, it was further proposed that the name of the formation bearing the dominant lithology be given preference.

Within the Wind River Basin all the major types of facies are present—chiefly red beds of the Goose Egg Formation (see following paragraph) in the eastern part; carbonate rocks of the Park City Formation in the central part; and mudstone, chert, and phosphorite of the Phosphoria Formation and variable amounts of sandstone of the Shedhorn Sandstone in the western part (fig. 20). The distinction between the red-bed and carbonate rock facies in the eastern third of the basin is relatively clear cut, but in the western third the separation of facies is often very difficult because of the intricate intertonguing of many different rock types. The principal lithology over much of the area, however, is carbonate rock; consequently, the name Park City Formation and equivalents is used in this report rather than the name Phosphoria Formation which has been used more commonly in the past.

Thomas (1934) studied the intertonguing of limestone with red beds in central and southeastern Wyoming, and named several of the eastward-projecting tongues of the Phosphoria Formation. In a later report, Burk and Thomas (1956) proposed the name Goose Egg Formation for the red shale and gypsum sequence containing thin beds of limestone in the extreme southeast corner of the Wind River Basin (fig. 21). The name was derived from the Goose Egg Post Office along Wyoming Highway 220 at the west end

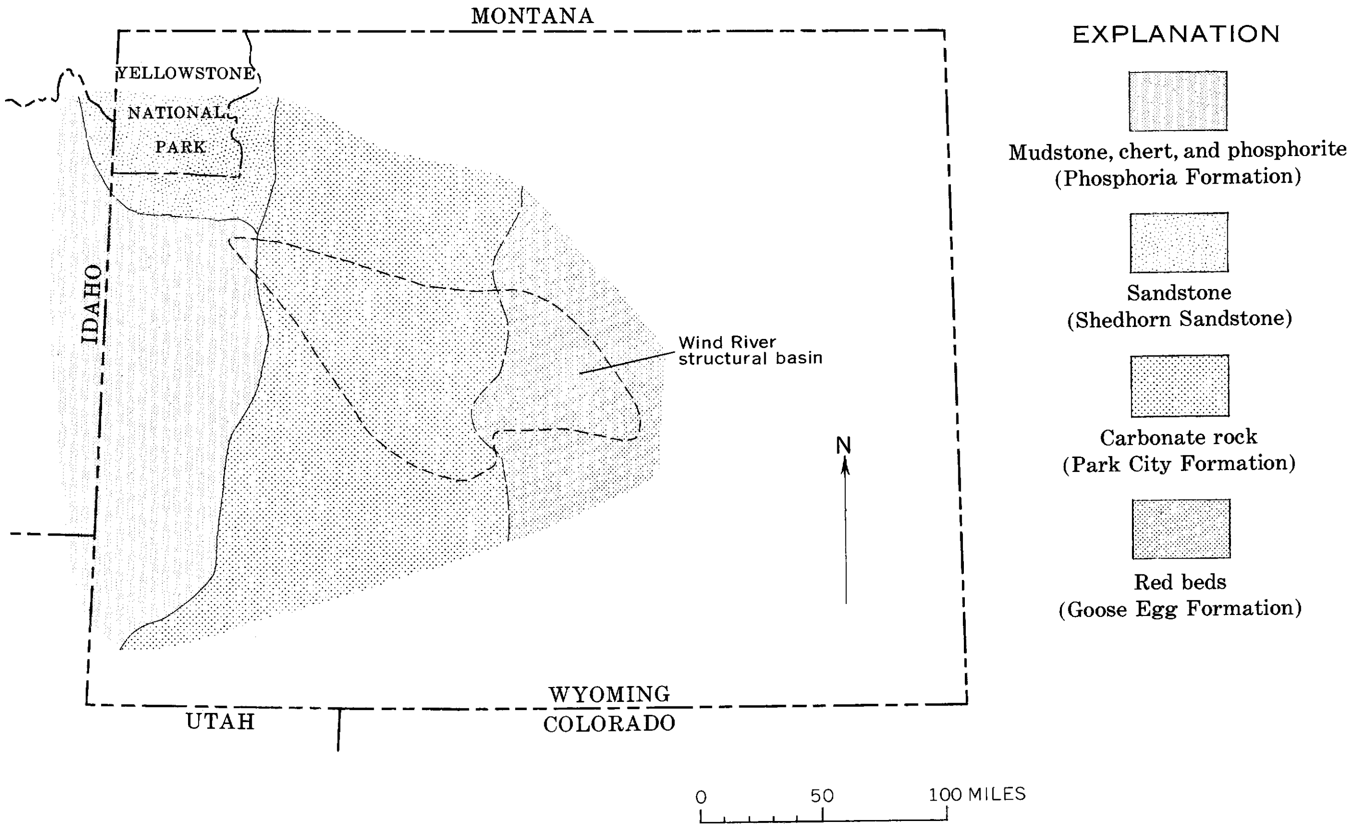


FIGURE 20.—Distribution of major facies in Permian rocks in central and western Wyoming. Based on McKelvey and others (1959, p. 4).

of Casper Mountain; the type section is shown on plate 5. In this report the Goose Egg is extended northward to include similar strata in the southern Bighorn Mountains.

PARK CITY FORMATION

Lithology and Thickness

In the northern Wind River Range and adjacent parts of the Washakie Range, the Park City Formation consists of 150–260 feet (fig. 22) of interbedded chert, limestone, dolomite, siltstone, and very fine grained sandstone and a few thin beds of shale and phosphate rock. Colors are predominantly tan, gray, and buff; the phosphatic beds are dark gray to brown and black. The basal beds of the formation, 5–15 feet thick, are commonly conglomeratic and contain angular fragments of white and gray siltstone, sandstone, and crystalline quartz, and rounded pebbles and granules of black chert. The matrix of the conglomerate ranges from pink and gray dense sandy limestone to tan and pink coarse-grained sandstone. Chert occurs in thin beds but more commonly as tubular twisted masses in thick-bedded limestone and dolomite. The twisted chert masses, which are particularly abundant in the upper half, are the most characteristic lithologic features of the formation. Phosphate rock, generally black and oolitic, is in beds only a foot or two thick. Some other rock types contain abundant phosphatic grains or pellets, and many fossil shells are highly phosphatic.

Similar lithology characterizes the Park City Formation toward the southeast end of the Wind River Range, except for a gradual increase in limestone and dolomite. Also, little sandstone or basal conglomerate is present at most localities. Detailed descriptions have been published by King (1947–1957) and Sheldon (1957), who separated these rocks into several units for convenience in evaluating the phosphate deposits. Thickness ranges from about 180 to 325 feet (pl. 2). The upper massive chert and carbonate units erode to very prominent dip slopes along the entire east flank of the range (fig. 23). The dip slopes are grass covered, but they are conspicuously devoid of trees in contrast to the underlying Tensleep Sandstone.

Along the south margin of the Wind River Basin the Park City can be traced as far east as the Conant Creek area, where it is about 350 feet thick (pl. 1). Farther east, toward the north end of Rattlesnake Hills, the carbonate and chert facies grade into the red-bed facies of the Goose Egg Formation (Weichman, 1958). Thin tongues of limestone project eastward into the red beds, however, and are present throughout the eastern part of the Wind River Basin (fig. 21; see discussion of Goose Egg Formation).

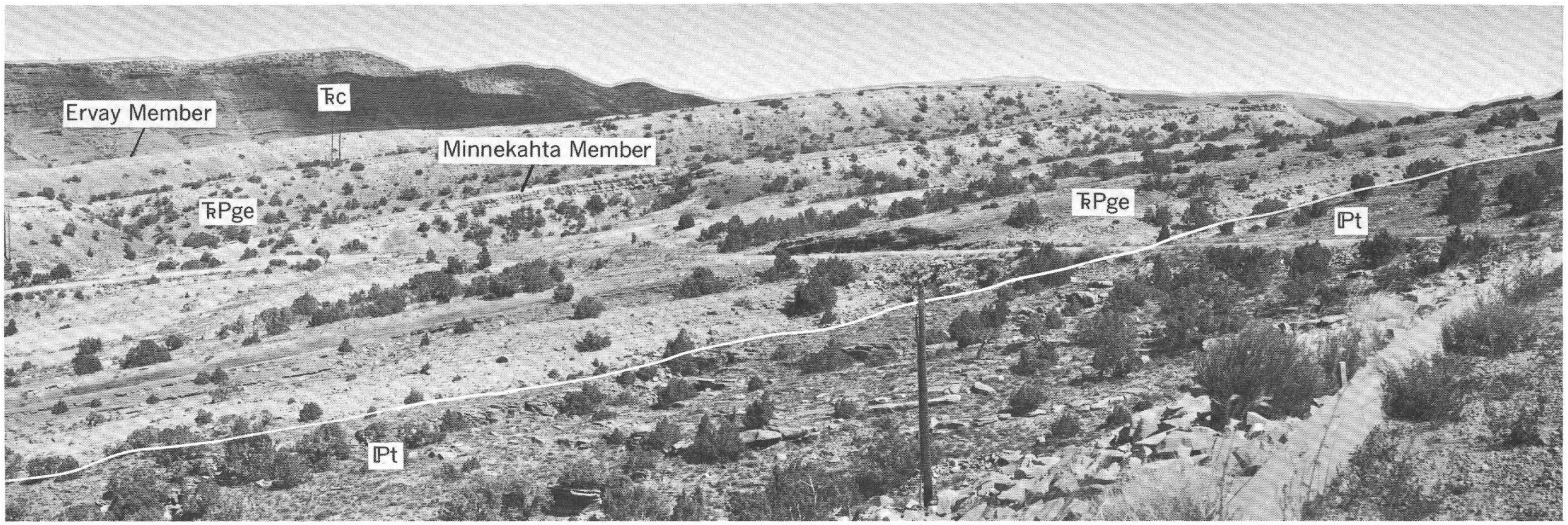


FIGURE 21.—Exposures of Goose Egg Formation (TRPge) at Alcova Dam, southeastern Wind River Basin. TRc, Chugwater Formation; IPt, Tensleep Sandstone.

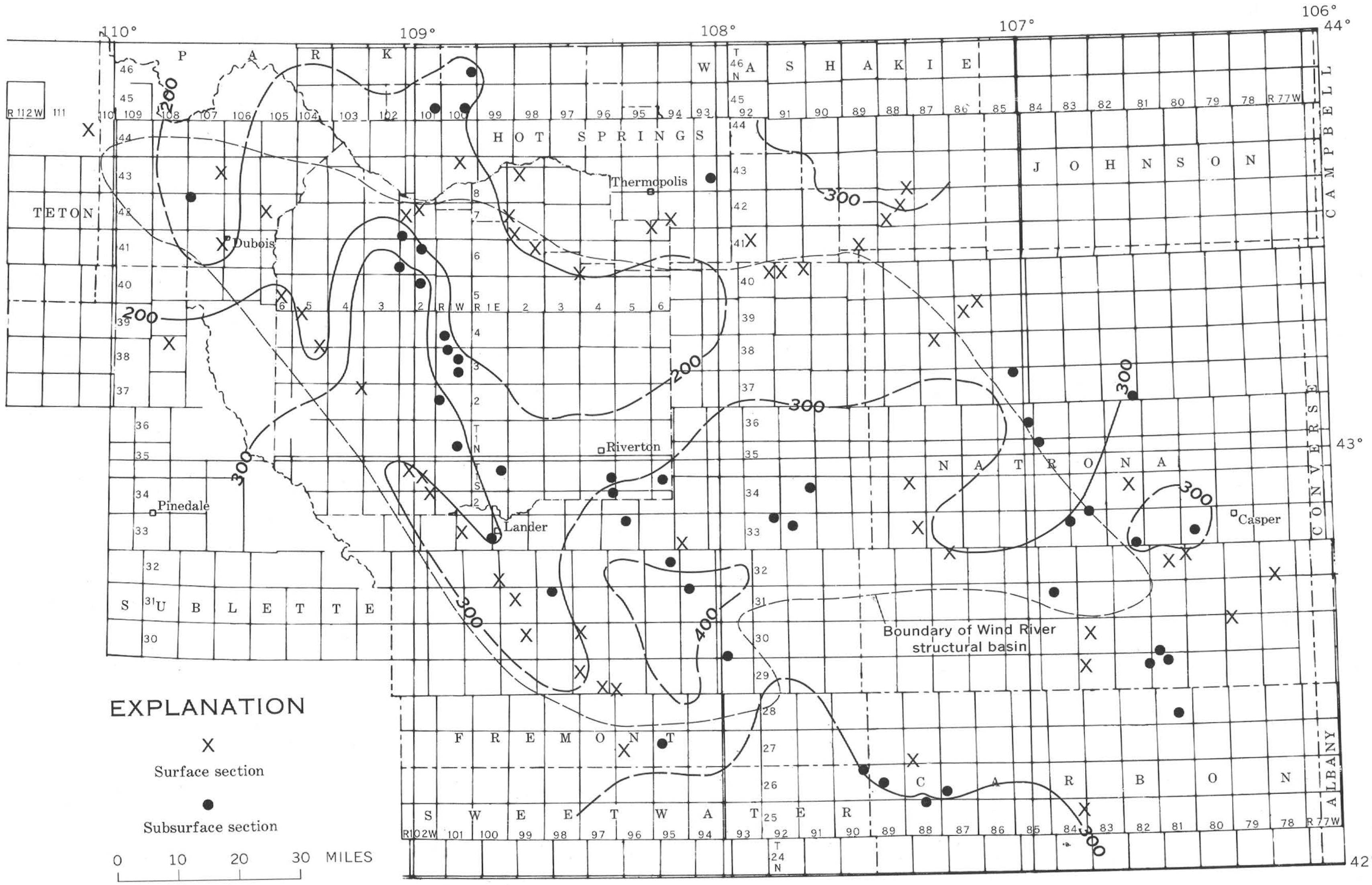


FIGURE 22.—Thickness map of Park City Formation and equivalents, in central Wyoming. Interval is 100 feet; isopachs are restored across mountain arches where rocks have been eroded; isopachs are dashed where control is inadequate.

FIGURE 23.—Prominent dip slopes ("flatirons") of Park City Formation along east flank of Wind River Range, near Fort Washakie. Tree-covered slopes in middle background are on Tensleep Sandstone.

The Park City Formation is 200–265 feet thick in the Owl Creek Mountains (pl. 3), and consists chiefly of limestone, dolomite, dolomitic siltstone, and chert. These beds persist to the east end of the range, but farther east, in exposures at the west end of the Bighorn Mountains, red beds and gypsum (of the Goose Egg Formation) are the dominant rock types (Tourtelot, 1952, p. 50). At the base of the Permain sequence in the southwestern Bighorn Mountains is a thin resistant dolomite that unconformably overlies the Tensleep Sandstone and which is, in turn, unconformably overlain by the lowermost red beds (pl. 3). This dolomite has been referred to as the Nowood Member of the Phosphoria Formation by McCue (1953). It can be traced westward into the Owl Creek Mountains (E. K. Maughan, oral commun., 1963) and probably correlates with the Grandeur Member of the Park City Formation in the northern Wind River Range (McKelvey and others, 1959, p. 12–15).

Contact With the Dinwoody Formation

The contact between the Park City Formation and the overlying Triassic Dinwoody Formation appears to be nearly everywhere conformable, even though a hiatus, possibly representing all Late Permian (Ochoa) time, occurs between the two formations. There is, however, a sharp change in lithology at the contact. In contrast to the ledge-forming massive gray and tan chert, limestone, and dolomite in the upper part of the Park City, the Dinwoody is characterized by slope-forming thin-bedded slabby siltstone and very fine grained sandstone that weather to a distinctive yellowish tan in most places. The contact is readily defined on most electric logs.

Age

The Park City Formation is abundantly fossiliferous, containing a predominance of brachiopods and bryozoans (table 8). Coquina beds consisting almost entirely of large specimens of a single fossil species occur locally, as do conodonts and fish and shark remains. Yochelson (1963) found that each major rock facies has its own characteristic fossil assemblage and stated that "The nearly complete limitation of each assemblage to a particular facies is striking."

Despite the variety and abundance of fossils, specific correlation with the standard reference section of Permian rocks in West Texas is still somewhat uncertain. According to Thomas (1948, p. 90), the faunas in the Permian rocks of central Wyoming are provincial and not well represented in the Texas Permian. The available data indicate, however, that most of the Park City correlates with the Word Formation of Guadalupe (Early and Late Permian) age (Thomas, 1948, p. 90; Williams, in McKelvey and others, 1959, p. 39, 40). The basal beds (Nowood or Grandeur Member) are as old as Leonard and possibly Wolfcamp (Early Permian), and the uppermost part probably is at least as young as late Guadalupe. Fossils of definite Ochoa (Late Permian) age have not been recognized.

Conditions of Deposition

At the beginning of Permian time a broad north-south-trending highland occupied all the Wind River Basin area. Rocks of Middle Pennsylvanian age were exposed in the central and western parts of the basin, some of Late Pennsylvanian age in the eastern part. The seas lay both east and west of the highland and, in Wolfcamp time, began to encroach upon it from both sides. Possibly by late Wolfcamp time the highland was breached by shallow throughgoing seaways that persisted into Leonard time, and a thin deposit of limestone was laid down across the northern and western parts of the basin. Sand, eroded from the exposed Tensleep Sandstone, was deposited along the east edge during Wolfcamp time. Then the seas regressed, at least from the eastern part of the basin area, and a short period of erosion followed. In late Leonard time, seas again encroached from either side of the highland, and by Guadalupe time the entire region was flooded. In the east the sea remained shallow, and red shale and evaporite sediments accumulated, but toward the west, sedimentation took place in deeper waters. Periodically the sea expanded eastward, and thin but widespread tongues of limestone were deposited over the westward-projecting tongues of red shale. In Thomas' opinion (1948, p. 91) the relation between the red beds and the limestone is one of intertonguing rather than of facies change, suggesting that the eastward expansion of the sea was fairly rapid.

The sea may have withdrawn by the end of Guadalupe time, for there is no positive record that sedimentation occurred in the basin area during latest Permian time.

TABLE 8.—*Fossils from Park City and Goose Egg Formations*

[Generic and specific names are quoted without modification from sources shown]

Stratigraphic section	Collection No. (pls. 2–5)	Species	References
Warm Spring Creek	9	*Bellerophon* sp., *Euphemites subpapillosus*, *Strophostylus* sp., *Plicatoderbya magna*, *Punctospirifer pulchra*, *Neospirifer pseudocameratus*, *Waagenoconcha montpelierensis*, *Aulosteges* cf. *A. hispidus*, *Pustula nevadensis*.	Love (1939, p. 34).
	10	*Stenopora* (3 sp.), *Lioclema* sp., *Phyllopora* sp., *Rhombopora* sp., *Meekella* sp., *Chonetes* aff. *C. geinitzianus*, *Aulosteges* sp., *Spiriferina pulchra*, *Aviculopecten* sp., *Pseudomonotis* aff. *P. hawni*, *Bellerophon*? sp., *Patella* sp., *Nautilus* sp.	Eliot Blackwelder (U.S. Geol. Survey, unpub. field notes); identification by G. H. Girty.
	11	*Orbiculoidea* sp.	Weart (1948).
Dinwoody Canyon	10	*Orbiculoidea* sp. indet., fish remains	E. L. Yochelson (written commun., 1964).
	11	*Orbiculoidea* (*Roemerella*) *utahensis*, *Schizodus* sp. indet. or *Scaphellina* sp. indet., *Sanguinolites*? sp. indet. or *Wilkingia*? sp indet., fish remains, bryozoans.	Do.
	12	*Orbiculoidea* (*Roemerella*) *utahensis*, *Derbyia magna*, *Wellerella* cf. *W. osagensis*, *Sphenosteges hispidus*, *Echinauris subhorridus*, *Bathymyonia nevadensis*, *Spiriferina pulchra*, *Aviculopecten* sp. indet., *Nucluopsis poposiensis*, *Polidevcia bellistriata*, *Stutchburia* sp. indet., fish remains, abundant bryozoans.	Do.
	13	*Composita subtilita*, *Chonetes* sp. indet. *Sphenosteges hispidus*, *Derbyia magna*, *Bathymyonia* sp. indet., *Spiriferina pulchra*, *Aviculopecten* sp. indet., *Hustedia phosphoriensis*, *Nucula pulchella*, *Cyrtorostia varicostata*, *Camptonectes*? sp. indet., *Pseudomonotis* sp. indet., *Conocardium* sp. indet., *Schizodus canalis*, *Stutchburia*? sp. indet., *Sanguinolites*? sp. indet. or *Wilkingia* sp. indet., *Stearoceras*? sp. indet., *Euphemitopsis subpapillosus*, fish remains, abundant bryozoans.	Do.
Bull Lake Canyon	6	*Orbiculoidea* sp. indet., *Scaphellina concinnus*, *Schizodus* cf. *S. wheeleri*, *Euphemites* sp. indet., *Sanguinolites*? sp. indet. or *Wilkingia* sp. indet., *Cyrtorostia*? sp. indet., fish remains, abundant bryozoans.	E. L. Yochelson (written commun., 1964).
	7	*Orbiculoidea* sp. indet., *Echinauris subhorridus*, "*Liosotella*" sp. indet., *Bathymyonia nevadensis*, *Spiriferina pulchra*, *Muirwoodia multistriatus*, abundant bryozoans.	Do.
	8	*Composita subtilita*, *Spiriferina* cf. *S. kentuckensis*, *Hustedia phosphoriensis*.	Do.
	9	*Wellerella* cf. *W. osagensis*, *Sphenosteges* sp. indet., *Composita subtilita*, *Derbyia* sp. indet., *Neospirifer pseudocameratus*, "*Liosotella*" sp. indet., *Bathymyonia nevadensis*, *Muirwoodia multistriatus*, *Spiriferina pulchra*, *Hustedia phosphoriensis*, *Dielasma* sp. indet., *Myalina* cf. *M.* (*M.*) *wyomingensis*, *Cyrtorostia varicostata*, *Aviculopecten* cf. *A. kaibabensis*, *Aviculopecten* sp. indet., *Plagioglypta canna*, abundant bryozoans.	Do.
Trout Creek	2	*Composita subtilita*?, *Solenomya* sp., *Pteria* sp., *Allorisma terminale*?, *Sanguinolaria*? sp., *Pleurophorus* sp., *Pleurophorus* aff. *P. subcostatus*, *Astartella*? sp., *Schizodus*? sp., *Plagioglypta canna*, *Patellostium*? sp., *Pleurotomaria sp.*	Eliot Blackwelder (U.S. Geol. Survey, unpub. field notes); identification by G. H. Girty.
Squaw Creek	1	*Orbiculoidea* sp.	King (1947, p. 38–39).
	2	*Composita* sp., *Neospirifer pseudocameratis*, *Punctospirifer pulcher*	King (1947, p. 32).
Sinks Canyon	6	*Orbiculoidea* sp., *Allorisma* sp.	Condit (1924, p. 24).
Beaver Creek	4	*Orbiculoidea* sp.	Condit (1924, p. 29).

TABLE 8.—*Fossils from Park City and Goose Egg Formations*—Continued

[Generic and specific names are quoted without modification from sources shown]

Stratigraphic section	Collection No. (pls. 2–5)	Species	References
Windy Gap	8	*Sphenosteges* sp. indet., *Composita* sp. indet., *Derbyia magna, Neospirifer pseudocameratus, Echinauris subhorridus, "Liosotella"* sp. indet., *Bathymyonia nevadensis, Spiriferina* sp. indet., *Nucula* sp. indet., *Polidevcia bellistriata, Schizodus* sp. indet. or *Scaphellina* sp. indet., *Sanguinolites*? sp. indet. or *Wilkingia* sp. indet., fish remains, abundant bryozoans.	E. L. Yochelson (written commun., 1964).
	9	*Bathymyonia nevadensis, Derbyia magna, Cyrtorostia varicostata,* fish remains, abundant bryozoans.	Do.
	10	*Conularia* sp. indet., *Orbiculoidea* sp. indet., *Sphenosteges hispidus, Bathymyonia nevadensis, Derbyia magna, Spiriferina pulchra, Spiriferina* cf. *S. kentuckensis, Hustedia phosphoriensis, Aviculopecten* sp. indet., *Schizodus* sp. indet. or *Scaphellina* sp. indet., *Sanguinolites*? sp. indet. or *Wilkingia* sp. indet., *Costatoria secradiata, Stutchburia*? sp. indet., fish remains, abundant bryozoans.	Do.
Circle Ridge	1	*Hustedia phosphoriensis, Spiriferina pulchra, Aviculopecten* sp., *Pustula nevadensis, Orbiculoidea* sp.	Milton (1942).
Sheep Ridge	1	*Plagioglypta, Myalina*?	McCue (1953).
	2	*Plicatoderbya, Punctospirifer, Batostomella, Aviculopecten, Hustedia, Stenopora*?, *Pleurotomaria, Horridonia*?, *Dictyoclostus, Ammodiscus.* (Collection obtained from type section of "Embar" Formation.)	Do.
Wind River Canyon	13	*Stenopora* n. sp., *Phyllopora* sp., *Derbya* aff. *D. multistriata, Spiriferina pulchra, Solenomya*? n. sp., *Chaenomya*? n. sp., *Deltopecten* aff. *D. vanvleeti, Pseuodmonotis* aff. *P. sublevis, Pseudomonotis* aff. *P. hawni, Oxytoma*? n. sp., *Myalina wyomingensis*?, *Myalina* sp., *Schizodus* aff. *S. deparcus, Schizodus* aff. *S. subcircularis, S. ferrieri*?, *Pleurophorus* n. sp., *Plagioglypta canna, Bellerophon* aff. *B. crassus, Euphemus*? sp., *Naticopsis*? sp., *Omphalotrochus*? sp.	Lee (1927, p. 57).
Rattlesnake Hills	3	*Plagioglypta*? sp	Thomas (1934, p. 1674).
	4	*Bryozoans* (many species), *Punctospirifer pulchra, Punctospirifer* aff. *P. kentuckyensis, Composita* cf. *C. mexicana, Derbya* aff. *D. plicatella, Tentaculites* sp., *Plagioglypta*? sp. (small), *Myalina* sp. (large), *Myalina* aff. *M. perattenuata, Pleurophorus pricei, Pleurophorus* (4 sp.), *Leda obesa, Cyrtorostra*? sp., *Aviculopecten* sp., *Deltopecten vanvleeti, Parallelodon* aff. *P. tenuistriatus, Edmondia*? sp., *Schizodus* cf. *S. wheeleri, Schizodus*? sp. (large), *Schizodus* sp. (small), *Pinna peracuta, Euphemus subpapillosus, Bellerophon* sp., gastropod (high-spired form), *Coelogasteroceras thomasi, Coloceras*? sp.	Thomas (1934, p. 1675).
Alcova	5	*Plagioglypta*? sp	Thomas (1934, p. 1678).
	6	*Myalina permiana, Pinna* cf. *P. peracuta, Pleurophorus* sp., gastropods (high-spired form; several species).	Do.
Bates Creek	3	*Dentalium* (*Plagioglypta*) cf. *D.* (*P.*) *canna, Pleurophorus occidentalis,* gastropods (high-spired form).	Jenkins (1950).
Casper Mountain	3	*Pinna paracuta, Schizodus ferrieri, Pteria*? sp., *Plagioglypta canna*	Lee (1927, p. 49).

PERMIAN AND TRIASSIC ROCKS, GOOSE EGG FORMATION

Lithology and Thickness

At its type locality (Casper Mountain section, pl. 5) the Goose Egg Formation consists of 380 feet of red gypsiferous shale and siltstone interbedded with distinctive gray and purple dense platy limestone. The formation can be divided in ascending order into Opeche Shale, Minnekahta Limestone, Glendo Shale, Forelle Limestone, Difficulty Shale, Ervay, Freezeout Shale, and Little Medicine Members (Burk and Thomas, 1956, p. 6; Maughan, 1964). The red-bed units range in thickness from 40 to about 100 feet, and the limestone beds from 6 to 22 feet. The Minnekahta, Forelle, and Ervay Members are eastward-projecting tongues from the Park City Formation, and the Little Medicine Member is an eastward-projecting tongue from the Lower Triassic Dinwoody Formation. Thus, only that part of the Goose Egg Formation lying below the top of the Ervay is considered to be the equivalent of the Park City Formation; these strata have a thickness of 304 feet (pl. 5).

The individual members of the Goose Egg Formation can be traced in subsurface sections northwest along the Casper arch to the south end of the Bighorn Mountains. In this area they are 355 feet thick; of this thickness the Park City equivalents comprise 270 feet (pl. 3). Westward into the Owl Creek Mountains the red beds and gypsum gradually intertongue with carbonate rocks and chert of the Park City Formation (pl. 3) and with dolomitic siltstone of the Dinwoody Formation. Another member, the Nowood Member of the Phosphoria

Formation of McCue (1953), occurs locally at the base of the Goose Egg in the southwestern Bighorn Mountains (pl. 3).

Although typical lithologies of the Goose Egg Formation can be observed in exposures at the north end of Rattlesnake Hills (pl. 4), it is difficult to distinguish the Forelle Limestone Member, because several thin limestone beds occur in the middle part of the formation. Some of the red-bed units appear to extend westward along the south margin of the Wind River Basin nearly as far as the Conant Creek area, but exposures are too poor in the intervening area to show where individual tongues wedge out.

Contact with the Chugwater Formation

The Goose Egg Formation is overlain conformably by the Triassic Chugwater Formation. Because both formations consist predominantly of red shale and siltstone, it is commonly difficult to determine the contact in surface sections. According to Burk and Thomas (1956, p. 7), the Goose Egg beds are generally darker red, sandier, more gypsiferous, and somewhat more resistant than those in the Chugwater. In addition, the Goose Egg contains thin beds of limestone, but the lower part of the Chugwater does not. The Goose Egg Formation generally shows higher resistivity on electric logs than does the Chugwater; the contact, therefore, is easily identified in subsurface sections.

Age

The Minnekahta Limestone and Ervay Members have yielded fossils at several localities in the southern and southeastern parts of the Wind River Basin (table 8). A large assemblage from the Ervay in the Rattlesnake Hills indicates correlation with the upper part of the Park City Formation and equivalents of the Wind River Range (Thomas, 1934, p. 1676). It is concluded, therefore, that the Goose Egg Formation below the top of the Ervay Member is probably Early and early Late Permian. The upper two members of the formation (Freezeout Shale and Little Medicine Members), on the other hand, are lateral equivalents of the Dinwoody Formation which is considered to be of Early Triassic age (Newell and Kummel, 1942). There is no evidence that rocks of latest Permian age exist in the region, despite the seeming conformity between the Ervay Member and overlying strata.

Conditions of Deposition

While normal marine sedimentation took place in the western half of the Wind River Basin during late Early and early Late Permian times, red clastic sediments accumulated in the shallow areas of the seas in the eastern part. The continuity and uniformity of the red beds suggest that the red beds were deposited under water, but this is still uncertain. The source of the red debris is not known, but it was probably low landmasses that lay principally to the east and south of Wyoming (Thomas, 1934, p. 1691, 1692). Increase in the depth of water, caused either by eustatic rise in sea level or cessation of uplift in the landmasses, resulted periodically in the deposition of limestone across the entire region. In the eastern part of the basin area, some arms of the sea became isolated, or circulation became restricted, to the extent that evaporites were deposited.

Sedimentation apparently ceased in central Wyoming during latest Permian time, but there is no evidence in the strata of the Goose Egg Formation to suggest that the sea withdrew.

PALEOZOIC STRUCTURE

Throughout the Paleozoic Era the present site of the Wind River Basin was a region of remarkable crustal stability. Although hiatuses—some of them representing long periods of time—occur between many of the formations, no angular discordances within the Paleozoic rocks were recognized. The thickness maps (figs. 6, 11, 12, 14, 17, and 22) suggest that tectonic movements were limited to broad upwarping and downwarping along trends which, with one exception, show little direct relation to the structural trends of later Laramide deformation. If some parallelism did exist during certain periods, it was only temporary and in large part probably coincidental. Commonly the reverse is true; Paleozoic "highs" and "lows" correspond to Laramide "lows" and "highs," respectively. (Compare isopach maps with fig. 2.) The one exception is along the southeast margin of the basin, where an area coinciding roughly with the Laramie Mountains was relatively positive during much of Paleozoic time. This positive area, which widened south and southeast, is outlined by thickness lines for both the Cambrian and Mississippian strata (figs. 6, 14); Ordovician and Devonian rocks, if initially deposited in the area, were eroded before successively younger strata were laid down. One of the most pronounced periods of uplift was Middle Pennsylvanian (Atoka) time, when the positive area extended westward and included much of the region now occupied by the Granite Mountains. In terms of Pennsylvanian paleogeography, it has been referred to by Mallory (1963) as the Pathfinder uplift.

ECONOMIC GEOLOGY

OIL AND GAS

Paleozoic rocks are among the most important reservoirs for oil and gas in the Wind River Basin; they constitute the primary objectives for drilling in many areas. The Park City Formation and equivalents and Tensleep Sandstone are the most prolific, each being productive in about 20 fields (drill data as of Aug. 1, 1963). The Madison Limestone yields oil and gas in two fields, the Amsden Formation (Darwin Sandstone Member) in one. Wells drilled to Precambrian rocks in several of the producing fields have not found significant amounts of petroleum in Devonian or older rocks.

Nearly all production from Paleozoic rocks has been from anticlinal folds along the basin margins. Conditions favoring stratigraphic entrapment, such as textural and porosity changes within carbonate strata, wedging out of beds beneath unconformities, and changes in facies, exist in certain areas. Many of these potential sources, however, have not been adequately studied or explored. The Cottonwood Creek field in the southeastern part of the Bighorn Basin produces from a stratigraphic trap formed by the carbonate-to-shale facies change within the Park City and Goose Egg Formations (Biggs and Espach, 1960, p. 80), but attempts to locate similar traps in the south-central part of the Wind River Basin have as yet been unsuccessful. Owing to excessive drilling depths, the Paleozoic rocks remain virtually untested in the central part of the basin.

PHOSPHATE ROCK

Permian rocks have long been recognized as a potential source of phosphate in central and western Wyoming (Weeks and Ferrier, 1906; Weeks, 1907; Blackwelder, 1911; Condit, 1924). Owing to accessibility and proximity to a railroad terminal, the deposits along the east flank of the Wind River Range near Lander have been of particular economic interest. King (1947) studied the Lander deposits, and, more recently, Sheldon (1957) reported on the phosphate rocks at selected localities in the northwestern part of the range. Phosphate rock is also present in nearly all outcrops of the Park City Formation and equivalents in the southwestern and northern parts of the Wind River Basin, but these have not been extensively studied.

The phosphate-bearing rocks generally form two distinct zones within the Permian sequence along the east flank of the Wind River Range. These zones have been referred to as the "lower phosphate zone" and "upper phosphate zone" by Condit (1924) and King (1947) and as units B and D by Sheldon (1957). Each zone occurs as distinct beds of phosphorite, generally less than 1 foot thick and rarely more than 2 feet thick, and as grains or pellets disseminated through shale, sandstone, and limestone, with which the phosphorite is interbedded. According to King (1947, p. 12), the phosphate rock in the Lander area is classed as low to medium grade; the lower zone ranges from 21.5 to 29.7 percent P_2O_5 and the upper zone from 15.2 to 20.2 percent P_2O_5.

None of the phosphate deposits in the region has yet been exploited. The reserves, mining conditions, and power and transportation facilities are adequate, but the thinness and relative low grade of the phosphate beds have been the principal deterrents (King, 1947, p. 81). A determination that the Lander deposits are economic undoubtedly would provide an incentive to search for additional reserves in the adjacent areas.

OTHER COMMODITIES

The sandstone and carbonate rocks of the Paleozoic formations are readily available for a variety of construction uses in many areas. None of the strata has yet been used extensively as building stone. The Flathead Sandstone would probably be the best source, because of its color, resistance to weathering, and thin bedding. Other thick sandstone units, such as the Tensleep Sandstone and the Darwin Sandstone Member of the Amsden Formation, are more massive, friable, and porous.

Hematite-bearing red shale that occurs above the Darwin Sandstone Member of the Amsden Formation at many localities may prove to be a source of low-grade iron ore. Gruner and Smith (1955, p. 31) reported carnotite associated with barite and fluorite in rocks of Cambrian(?) age near Whiskey Mountain (sec. 12, T. 40 N., R. 107 W.), in the northern part of the Wind River Range. Localized radioactive deposits have also been reported from the Tensleep Sandstone in sec. 31, T. 31 N., R. 97 W., and from the Phosphoria Formation in secs. 7 and 10, T. 30 N., R. 97 W. (Wilson, 1960, p. 23). The extent and potential of the deposits are not known.

REFERENCES CITED

Agatston, R. S., 1954, Pennsylvanian and Lower Permian of northern and eastern Wyoming: Am. Assoc. Petroleum Geologists Bull., v. 38, no. 4, p. 508–583.

——— 1957, Pennsylvanian of the Wind River Basin, *in* Wyoming Geol. Assoc. Guidebook, 12th Ann. Field Conf., 1957: p. 29–33.

Amsden, T. W., and Miller, A. K., 1942, Ordovician conodonts from the Bighorn Mountains of Wyoming: Jour. Paleontology, v. 16, no. 3, p. 301-306.

Andrichuk, J. M., 1951, Regional stratigraphic analysis of Devonian System in Wyoming, Montana, southern Saskatchewan, and Alberta: Am. Assoc. Petroleum Geologists Bull., v. 35, no. 11, p. 2368–2408.

——— 1955, Mississippian Madison group stratigraphy and sedimentation in Wyoming and southern Montana: Am. Assoc. Petroleum Geologists Bull., v. 39, no. 11, p. 2170–2210.

Bearce, D. N., 1963, The Sacajawea Formation and the Darwin Sandstone in the southeast Wind River Mountains, Wyoming: Missouri Univ., unpub. masters thesis, 61 p.

Bell, W. G., 1955, The geology of the southeastern flank of the Wind River Mountains, Fremont County, Wyoming: Wyoming Univ., unpub. Ph.D. dissertation, 204 p.

Berg, R. R., 1961, Laramide tectonics of the Wind River Mountains, *in* Wyoming Geol. Assoc. Guidebook 16th Ann. Field Conf., 1961: p. 70–80.

Biggs, C. A., 1951, Stratigraphy of the Amsden formation of the Wind River Range and adjacent areas in northwestern Wyoming: Wyoming Univ., unpub. masters thesis, 121 p.

Biggs, Paul, and Espach, R. H., 1960, Petroleum and natural gas fields in Wyoming: U.S. Bur. Mines Bull. 582, 538 p.

Blackwelder, Eliot, 1911, A reconnaissance of the phosphate deposits in western Wyoming: U.S. Geol. Survey Bull. 470, p. 452–481.

——— 1913, Origin of the Bighorn dolomite of Wyoming: Geol. Soc. America Bull., v. 24, p. 607–624.

——— 1918, New geological formations in western Wyoming: Washington Acad. Sci. Jour., v. 8, no. 13, p. 417–426.

Bogrett, J. W., 1951, Geology of the northwestern end of the Rattlesnake Hills, Natrona County, Wyoming: Wyoming Univ., unpub. masters thesis, 70 p.

Boutwell, J. M., 1907, Stratigraphy and structure of the Park City mining district, Utah: Jour. Geology, v. 15, p. 434–458.

Branson, C. C., 1937, Stratigraphy and fauna of the Sacajawea formation, Mississippian, of Wyoming: Jour. Paleontology, v. 11, no. 8, p. 650–660.

——— 1939, Pennsylvanian formations of central Wyoming: Geol. Soc. America Bull., v. 50, no. 8, p. 1199–1226.

Branson, E. B., and Branson, C. C., 1941, Geology of the Wind River Mountains, Wyoming: Am. Assoc. Petroleum Geologists Bull., v. 25, no. 1, p. 120–151.

Branson, E. B., and Greger, D. K., 1918, Amsden formation of the east slope of the Wind River Mountains of Wyoming and its fauna: Geol. Soc. America Bull., v. 29, p. 309–326.

Burk, C. A., 1954, Faunas and age of the Amsden formation in Wyoming: Jour. Paleontology, v. 28, no. 1, p. 1–16.

Burk, C. A., and Thomas, H. D., 1956, The Goose Egg formation (Permo-Triassic) of eastern Wyoming: Wyoming Geol. Survey Rept. Inv. 6, 11 p.

Burton, G. C., 1952, Geology of Squaw Creek area, Wind River Mountains, Fremont County, Wyoming: Missouri Univ., unpub. master thesis.

Chronic, John, and Ferris, C. S., Jr., 1961, Lower Paleozoic outlier in southeastern Wyoming [abs.]: Geol. Soc. America Spec. Paper 68, p. 86.

Cloud, P. E., Jr., 1955, Physical limits of glauconite formation: Am. Assoc. Petroleum Geologists Bull., v. 39, no. 4, p. 484-492.

Collier, A. J., and Cathcart, S. H., 1922, Possibility of finding oil in laccolithic domes south of the Little Rocky Mountains, Montana: U.S. Geol. Survey Bull. 736–F, p. 171–178.

Condit, D. D., 1916, Relations of Embar and Chugwater formations in central Wyoming: U.S. Geol. Survey Prof. Paper 98-O, p. 263–270.

——— 1924, Phosphate deposits in the Wind River Mountains, near Lander, Wyoming: U.S. Geol. Survey Bull. 764, 39 p.

Coryell, H. N., and Johnson, S. C., 1939, Ostracoda of the Clore limestone, Upper Mississippian, of Illinois: Jour. Paleontology, v. 13, no. 2, p. 214–224.

Croneis, C. G., and Funkhouser, H. J., 1938, New ostracodes from the Clore formation [Ill.]: Denison Univ. Sci. Lab. Jour., v. 33, p. 331–360.

Darton, N. H., 1901, Geology and water resources of the southern half of the Black Hills and adjoining regions in South Dakota and Wyoming: U.S. Geol. Survey 21st Ann. Rept., pt. 4, 489–599.

Darton, N. H., 1904, Comparison of the stratigraphy of the Black Hills, Bighorn Mountains, and Rocky Mountain front range: Geol. Soc. America Bull., v. 15, p. 379–448.

——— 1906a, Geology of the Owl Creek Mountains: U.S. 59th Cong., 1st sess., S. Doc. 219, 48 p.

——— 1906b, Geology of the Bighorn Mountains: U.S. Geol. Survey Prof. Paper 51, 129 p.

——— 1906c, Fish remains in Ordovician rocks in Bighorn Mountains, Wyoming, with a résumé of Ordovician geology of the Northwest: Geol. Soc. America Bull., v. 17, p. 541–566.

——— 1908, Paleozoic and Mesozoic of central Wyoming: Geol. Soc. America Bull., v. 19, p. 403–470.

Darton, N. H., and Paige, Sidney, 1925, Central Black Hills, South Dakota: U.S. Geol. Survey Geol. Atlas, Folio 219, 7 maps.

Deiss, Charles, 1938, Cambrian formations and sections in part of Cordilleran trough: Geol. Soc. America Bull., v. 49, no. 7, p. 1067–1168.

DeLand, C. R., 1954, Cambrian stratigraphy and Upper Cambrian trilobites of the southwestern flank of the Wind River Mountains, Wyoming: Wyoming Univ., unpub. masters thesis, 135 p.

DeLand, C. R., and Shaw, A. B., 1956, Upper Cambrian trilobites from western Wyoming: Jour. Paleontology, v. 30, no. 3, p. 542–562.

Denison, R. H., 1951, Late Devonian fresh-water fishes from the western United States: Fieldiana Geology, v. 11, no. 5, p. 221–261.

Denson, M. E., Jr., and Morrissey, N. S., 1952, The Madison group (Mississippian) of the Big Horn and Wind River Basins, Wyoming, *in* Wyoming Geol. Assoc. Guidebook, 7th Ann. Field Conf., 1952: p. 37–43.

Denson, N. M., 1942, Late Middle Cambrian trilobite faunas and stratigraphy of Alberta, Montana, Wyoming, and Utah: Princeton Univ., unpub. Ph.D. dissertation.

Duncan, Helen, 1956, Ordovician and Silurian coral faunas of western United States: U.S. Geol. Survey Bull. 1021–F, p. 209–236.

Easton, W. H., 1962, Carboniferous formations and faunas of central Montana: U.S. Geol. Survey Prof. Paper 348, 126 p.

Flanagan, P. E., 1955, Geology of the Mud Creek area, Hot Springs County, Wyoming: Wyoming Univ., unpub. masters thesis, 57 p.

——— 1962, The Mississippian-Madison rocks of the Wind River Basin, Wyoming [abs.], *in* Program for Rocky Mountain Section, Am. Assoc. Petroleum Geologists, 12th Ann. Mtg., Salt Lake City, Utah: p. 20.

Flower, R. H., 1946, Ordovician cephalopods of the Cincinnati region: Bull. Am. Paleontology, v. 29, no. 116, 656 p.

——— 1952, New Ordovician cephalopods from eastern North America: Jour. Paleontology, v. 26, no. 1, p. 24–59.

Foster, H. L., 1947, Paleozoic and Mesozoic stratigraphy of northern Gros Ventre Mountains and Mount Leidy Highlands, Teton County, Wyoming: Am. Assoc. Petroleum Geologists Bull., v. 31, no. 9, p. 1537–1593.

Gooldy, P. L., 1947, Geology of the Beaver Creek-South Sheep Mountain area, Fremont County, Wyoming: Wyoming Univ., unpub. masters thesis, 75 p.

Gorman, D. R., 1962, A study of the Amsden Formation: Illinois Univ., unpub. Ph.D. dissertation, 167 p.

Gruner, J. W., and Smith, D. K., Jr., 1955, Annual report for April 1, 1954 to March 31, 1955: U.S. Atomic Energy Comm. RME 3020, 37 p.

Harrison, J. W., 1938, Pennsylvanian stratigraphy of the Laramie Range of southeastern Wyoming: Wyoming Univ., unpub. masters thesis, 85 p.

Hembree, M. R., 1949, Geology of the Fossil Hill area, Wind River Mountains, Fremont County, Wyoming: Missouri Univ., unpub. masters thesis.

Henbest, L. G., 1954, Pennsylvanian Foraminifera in Amsden formation and Tensleep sandstone, Montana and Wyoming, *in* Billings Geol. Soc. Guidebook 5th Ann. Field Conf., 1954: p. 50–53.

——— 1956, Foraminifera and correlation of the Tensleep sandstone of Pennsylvanian age in Wyoming, *in* Wyoming Geol. Assoc. Guidebook 11th Ann. Field Conf., 1956: p. 58–63.

Hoare, R. D., and Burgess, J. D., 1960, Fauna from the Tensleep Sandstone in Wyoming: Jour. Paleontology, v. 34, p. 711–716.

Hose, R. K., 1955, Geology of the Crazy Woman Creek area, Johnson County, Wyoming: U.S. Geol. Survey Bull. 1027–B, p. 33–118.

Howell, B. F., and others, 1944, Correlation of the Cambrian formations of North America [chart 1]: Geol. Soc. America Bull., v. 55, no. 8, p. 993–1003.

Jenkins, C. E., 1950, Geology of the Bates Creek-Corral Creek area, Natrona County, Wyoming: Wyoming Univ., unpub. masters thesis, 80 p.

Keefer, W. R., 1957, Geology of the Du Noir area, Fremont County, Wyoming: U.S. Geol. Survey Prof. Paper 294–E, p. 155–221.

——— 1965, Stratigraphy and geologic history of the uppermost Cretaceous, Paleocene, and lower Eocene rocks in the Wind River Basin, central Wyoming: U.S. Geol. Survey Prof. Paper 495–A, 77 p.

King, R. H., 1947, Phosphate deposits near Lander, Wyoming: Wyoming Geol. Survey Bull. 39, 84 p.

——— 1957, Stratigraphy of the Phosphoria formation in the Wind River Mountains, *in* Wyoming Geol. Assoc. Guidebook 12th Ann. Field Conf., 1957: p. 35–38.

Kirk, Edwin, 1930, The Harding sandstone of Colorado: Am. Jour. Sci., 5th ser., v. 20, p. 456–466.

Klapper, Gilbert, 1958, An Upper Devonian conodont fauna from the Darby formation of the Wind River Mountains, Wyoming: Jour. Paleontology, v. 32, no. 6, p. 1082–1093.

Lee, W. T., 1927, Correlation of geologic formations between east-central Colorado, central Wyoming, and southern Montana: U.S. Geol. Survey Prof. Paper 149, 80 p.

Lochman, Christina, 1950, Status of the Dry Creek Shale of central Montana: Am. Assoc. Petroleum Geologists Bull., v. 34, no. 11, p. 2200–2222.

Lochman-Balk, Christina, 1956, The evolution of some Upper Cambrian and Lower Ordovician trilobite families: Jour. Paleontology, v. 30, no. 3, p. 445–462.

——— 1957, Paleoecology of the Cambrian in Montana and Wyoming, chap. 8 *in* Ladd, H. S., ed., Paleoecology, v. 2 *of* Treatise on marine ecology and paleoecology: Geol. Soc. America Mem. 67, p. 117–162.

Lochman-Balk, Christina, and Wilson, J. L., 1958, Cambrian biostratigraphy in North America: Jour. Paleontology, v. 32, no. 2, p. 312–350.

Love, J. D., 1939, Geology along the southern margin of the Absaroka Range, Wyoming: Geol. Soc. America Spec. Paper 20, 134 p.

——— 1954, Tentative diagrammatic correlation of Tensleep, Amsden, Casper, and Hartville Formations in Wyoming, *in* Wyoming Geol. Assoc. Guidebook 9th Ann. Field Conf., 1954 [chart in pocket].

Love, J. D., Keefer, W. R., Duncan, D. C., Bergquist, H. R., and Hose, R. K., 1951, Geologic map of the Spread Creek-Gros Ventre River area, Teton County, Wyoming: U.S. Geol. Survey Oil and Gas Inv. Map OM–118.

Love, J. D., Tourtelot, H. A., Johnson, C. O., Thompson, R. M., Sharkey, H. H. R., and Zapp, A. D., 1947, Stratigraphic sections of Mesozoic rocks in central Wyoming: Wyoming Geol. Survey Bull. 38, 59 p.

Love, J. D., Weitz, J. L., and Hose, R. K., 1955, Geologic map of Wyoming: U.S. Geol. Survey.

Lyon, D. L., 1956, Geology of the north flank, eastern Bridger Range, Hot Springs County, Wyoming: Wyoming Univ., unpub. masters thesis, 95 p.

McCue, J. J., 1953, Facies changes within the Phosphoria Formation in the southeast portion of the Big Horn Basin, Wyoming: Wyoming Univ., unpub. masters thesis, 88 p.

McKee, E. D., 1945, Stratigraphy and ecology of the Grand Canyon Cambrian, Pt. 1 *of* Cambrian history of the Grand Canyon region: Carnegie Inst. Washington Pub. 563, p. 1–168.

McKelvey, V. E., Williams, J. S., Sheldon, R. P., Cressman, E. R., Cheney, T. M., and Swanson, R. W., 1956, Summary description of Phosphoria, Park City, and Shedhorn Formations in western phosphate field: Am. Assoc. Petroleum Geologists Bull., v. 40, no. 12, p. 2826–2863.

McKelvey, V. E., Cressman, E. R., Sheldon, R. P., Cheney, T. M., Swanson, R. W., and Williams, J. S., 1959, The Phosphoria, Park City, and Shedhorn formations in western phosphate field: U.S. Geol. Survey Prof. Paper 313–A, p. 1–46.

Mallory, W. W., 1963, Pathfinder uplift of Pennsylvanian age in southern Wyoming, *in* Short papers in geology, hydrology, and topography: U. S. Geol. Survey Prof. Paper 450–E, p. E57–E60.

Mansfield, G. R., 1927, Geography, geology, and mineral resources of part of southeastern Idaho, with descriptions of Carboniferous and Triassic fossils, by G. H. Girty: U.S. Geol. Survey Prof. Paper 152, 453 p.

Maughan, E. K., 1963, Mississippian rocks in the Laramie Range, Wyoming, and adjacent areas, *in* Short papers in geology and hydrology: U.S. Geol. Survey Prof. Paper 475–C, p. C23–C27.

——— 1964, The Goose Egg Formation in the Laramie Range and adjacent parts of southeastern Wyoming, *in* Geological Survey Research 1964: U.S. Geol. Survey Prof. Paper 501–B, p. B53–B60.

Miller, A. K., 1930, The age and correlation of the Bighorn formation of northwestern United States: Am. Jour. Sci., 5th ser., v. 20, p. 195–213.

——— 1932, The cephalopods of the Bighorn formation of the Wind River Mountains of Wyoming: Connecticut Acad. Arts and Sci. Trans., v. 31, p. 193–297.

Miller, B. M., 1936, Cambrian stratigraphy of northwestern Wyoming: Jour. Geology, v. 44, no. 2, pt. 1, p. 113–144.

Milton, A. A., 1942, Geology of the Circle Ridge-Spring Mountain area, Owl Creek Mountains, Wyoming: Iowa State Univ., unpub. masters thesis.

Moore, R. C., chm., and others, 1944, Correlation of Pennsylvanian formations of North America: Geol. Soc. America Bull., v. 55, no. 6, p. 657–706.

Morey, P. S., 1935, Ostracoda from the Amsden formation of Wyoming: Jour. Paleontology, v. 9, no. 6, p. 474–482.

Murphy, J. F., Privrasky, N. C., and Moerlein, G. A., 1956, Geology of the Sheldon-Little Dome area, Fremont County, Wyoming: U.S. Geol. Survey Oil and Gas Inv. Map OM–181.

Murphy, J. F., and Richmond, G. M., 1965, Geology of the Bull Lake West quadrangle, Wyoming: U.S. Geol. Survey Geol. Quad. Map GQ–432.

Newell, N. D., and Kummel, Bernhard, Jr., 1942, Lower Eo-Triassic stratigraphy, western Wyoming and southeast Idaho: Geol. Soc. America Bull., v. 53, no. 6, p. 937–996.

Peale, A. C., 1893, The Paleozoic section in the vicinity of Three Forks, Mont.,: U.S. Geol. survey Bull. 110, 56 p.

Phillips, D. P., 1958, Geology of the Sheep Ridge area, Hot Springs and Fremont Counties, Wyoming: Wyoming Univ., unpub. masters thesis, 135 p.

Powell, J. D., 1957, The geology of the Blackrock Ridge area, Hot Springs and Fremont Counties, Wyoming: Wyoming Univ., unpub. masters thesis, 118 p.

Rhodes, F. H. T., 1963, Conodonts from the topmost Tensleep Sandstone of the eastern Big Horn Mountains, Wyoming: Jour. Paleontology, v. 37, no. 2, p. 401–408.

Richards, R. W., and Mansfield, G. R., 1912, The Bannock overthrust, a major fault in southeastern Idaho and northeastern Utah: Jour. Geology, v. 20, p. 681–709.

Richardson, G. B., 1913, The Paleozoic section in northern Utah: Am. Jour. Sci., 4th ser., v. 36, p. 406–416.

Robinson, G. D., 1963, Geology of the Three Forks quadrangle, Montana: U.S. Geol. Survey Prof. Paper 370, 143 p.

Ross, R. J., Jr., 1957a, Origin of Ordovician units exposed near the Wind River basin, *in* Wyoming Geol. Assoc. Guidebook 12th Ann. Field Conf., 1957: p. 17–19.

——— 1957b, Ordovician fossils from wells in the Williston basin, eastern Montana: U.S. Geol. Survey Bull. 1021–M, p. 439–510.

Sandberg, C. A., 1963, Dark shale unit of Devonian and Mississippian age in northern Wyoming and southern Montana, *in* Short papers in geology and hydrology: U.S. Geol. Survey Prof. Paper 475–C, p. C17–C20.

Sandberg, C. A., and McMannis, W. J., 1964, Occurrence and paleogeographic significance of the Maywood Formation of Late Devonian age in the Gallatin Range, southwestern Montana, *in* Geological Survey Research 1964: U.S. Geol. Survey Prof. Paper 501–C, p. C50–C54.

Sando, W. J., and Dutro, J. T., Jr., 1960, Stratigraphy and coral zonation of the Madison Group and Brazer Dolomite in northeastern Utah, western Wyoming, and southwestern Montana, *in* Wyoming Geol. Assoc. Guidebook 15th Ann. Field Conf., 1960: p. 117–126.

Shaw, A. B., 1954, Correlation of the Paleozoic formations of Wyoming, *in* Wyoming Geol. Assoc. Guidebook 9th Ann. Field Conf., 1954 [chart in pocket].

——— 1955, The Amsden formation in southwestern and south-central Wyoming, *in* Wyoming Geol. Assoc. Guidebook 10th Ann. Field Conf., 1955: p. 60–63.

——— 1957, Cambrian of the southwestern Wind River Basin, Wyoming, *in* Wyoming Geol. Assoc. Guidebook 12th Ann. Field Conf., 1957: p. 9–16.

Shaw, A. B., and Bell, W. G., 1955, Age of the Amsden formation, Cherry Creek, Wind River Mountains, Wyoming: Am. Assoc. Petroleum Geologists Bull., v. 39, no. 3, p. 333–337.

Shaw, A. B., and DeLand, C. R., 1955, Cambrian of southwestern Wyoming, *in* Wyoming Geol. Assoc. Guidebook 10th Ann. Field Conf., 1955: p. 38–42.

Sheldon, R. P., 1957, Physical stratigraphy of the Phosphoria formation in northwestern Wyoming: U.S. Geol. Survey Bull. 1042–E, p. 105–185.

Sims, F. C., 1948, Geology of the west end of the Laramie Range, Natrona County, Wyoming: Wyoming Univ., unpub. masters thesis, 52 p.

Sloss, L. L., 1952, Introduction to the Mississippian of the Williston Basin, *in* Billings Geol. Soc. Guidebook 3d Ann. Field Conf., 1952: p. 65–69.

Stokes, W. L., 1953, Summary of Paleozoic and Mesozoic stratigraphy [Utah], *in* Intermountain Assoc. Petroleum Geologists Guidebook 4th Ann. Field Conf., 1953: p. 14–18.

Strickland, J. W., 1957, Summary of Mississippian and Devonian stratigraphy, Wind River Basin, Wyoming, *in* Wyoming Geol. Assoc. Guidebook 12th Ann. Field Conf., 1957: p. 20–28.

Thomas, H. D., 1934, Phosphoria and Dinwoody tongues in lower Chugwater of central and southeastern Wyoming: Am. Assoc. Petroleum Geologists Bull., v. 18, no. 12, p. 1655–1697.

——— 1948, Summary of Paleozoic stratigraphy of the Wind River Basin, Wyoming, *in* Wyoming Geol. Assoc. Guidebook 3d Ann. Field Conf., 1948: p. 79–95.

——— 1952, Cambrian and Ordovician stratigraphy around the southern Bighorn Basin, Wyoming, *in* Wyoming Geol. Assoc. Guidebook 7th Ann. Field Conf., 1952: p. 32–36.

——— 1957, Geologic history and structure of Wyoming: Wyoming Geol. Survey repr. 18, 11 p.

Thomas, H. D., Thompson, M. L., and Harrison, J. W., 1953, Fusulinids of the Casper formation of Wyoming: Wyoming Geol. Survey Bull. 46, 56 p.

Tourtelot, H. A., 1952, Marine and evaporite facies of Permian and Triassic strata in the southern part of the Bighorn Basin and adjacent areas, central Wyoming, *in* Wyoming Geol. Assoc. Guidebook 7th Ann. Field Conf., 1952: p. 49–52.

——— 1953, Geology of the Badwater area, central Wyoming: U.S. Geol. Survey Oil and Gas Inv. Map OM–124.

Tourtelot, H. A., and Thompson, R. M., 1948, Geology of the Boysen area, central Wyoming: U.S. Geol. Survey Oil and Gas Inv. Map 91.

Van Houten, F. B., and Weitz, J. L., 1956, Geologic map of the eastern Beaver Divide-Gas Hills area, Fremont and Natrona Counties, Wyoming: U.S. Geol. Survey Oil and Gas Inv. Map OM–180.

Verville, G. J., 1957, Wolfcampian fusulinids from the Tensleep sandstone in the Bighorn Mountains, Wyoming: Jour. Paleontology, v. 31, no. 2, p. 349–352.

Walcott, C. D., 1908, Cambrian Brachiopoda-descriptions of new genera and species, No. 3 *of* Cambrian geology and paleontology: Smithsonian Misc. Colln., v. 53, no. 3, p. 53–137.

Wanless, H. R., Belknap, R. L., and Foster, Helen, 1955, Paleozoic and Mesozoic rocks of Gros Ventre, Teton, Hoback, and Snake River ranges, Wyoming: Geol. Soc. America Mem. 63, 90 p.

Weart, R. C., 1948, Geology of the northern flank of the Wind River Mountains, Fremont County, Wyoming: Univ. Syracuse, unpub. masters thesis.

Weed, W. H., 1900, Geology of the Little Belt Mountains, Montana: U.S. Geol. Survey, 20th Ann. Rept., pt. 3, p. 257–461.

Weeks, F. B., 1907, Phosphate deposits in western United States: U.S. Geol. Survey Bull. 340–K, p. 441–447.

Weeks, F. B., and Ferrier, W. F., 1906, Phosphate deposits in western United States: U.S. Geol. Survey Bull. 315–P, p. 449–462.

Weichman, B. E., 1958, Stratigraphy of the Phosphoria Formation in the southern portion of the Wind River Basin, Wyoming: Wyoming Univ., unpub. masters thesis, 72 p.

Wilson, W. H., 1960, Radioactive mineral deposits of Wyoming: Wyoming Geol. Survey Rept. Inv. 7, 41 p.

Woodward, T. C., 1957, Geology of the Deadman Butte area, Natrona County, Wyoming: Am. Assoc. Petroleum Geologists Bull., v. 41, no. 2, p. 212–262.

Yochelson, E. L., 1963, Paleoecology of the Permian Phosphoria Formation and related rocks, *in* Short papers in geology and hydrology: U.S. Geol. Survey Prof. Paper 475–B, p. B123–B124.

INDEX

[Italic page numbers indicate major references]

U.S. GOVERNMENT PRINTING OFFICE: 1966 O—205-540