Geology of the Georgetown Canyon-Snowdrift Mountain Area, Southeastern Idaho

GEOLOGICAL SURVEY BULLETIN 1153

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By EARLE R. CRESSMAN

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A study of an area of high-grade phosphate rock deposits

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GEOLOGY OF THE GEORGETOWN CANYON-SNOWDRIFT MOUNTAIN AREA, SOUTHEASTERN IDAHO

By Earle R. Cressman

ABSTRACT

The Georgetown Canyon-Snowdrift Mountain area consists of 160 square miles in the central Peale Mountains of southeastern Idaho. It is bounded on the north by the 42°37′30″ parallel of latitude, on the east and west by the 111°07′30″ and 111°22′30″ meridians of longitude, and on the southwest and south by an irregular line extending from Georgetown, Idaho, around the south end of Meade Peak, the highest point in the Peale Mountains. The highest altitude is 9,960 feet and the lowest 6,200 feet.

Except for some basalt of Tertiary age, the rocks exposed are all sedimentary, ranging in age from Mississippian to Recent. The Mississippian rocks consist of the Madison and Brazer limestones, the Pennsylvanian of the lower part of the Wells formation, and the Permian of the upper part of the Wells formation, the Grandeur tongue of the Park City formation, and the Phosphoria formation. Triassic strata are the Dinwoody, Thaynes, and Ankareh formations. The Jurassic consists of the Nugget sandstone, the Twin Creek limestone, and the Preuss and Stump sandstones, and the Cretaceous system is represented by the Gannett group. The Tertiary system consists of red conglomerate of possible Eocene age, the Salt Lake formation of Pliocene age, and olivine basalt of possible Pliocene age. There are no glacial deposits, but high-level gravels and hill wash in the ranges and large dissected alluvial fans in the valleys may be partly Pleistocene in age. Recent deposits are terrace gravel, landslides, small active fans, and alluvium.

The main structural feature is a folded overthrust, originally named the Bannock by Richards and Mansfield (1912) but here renamed the Meade overthrust, that underlies all but the easternmost, southernmost, and southwesternmost parts of the area. The overthrust plate consists almost entirely of upper Paleozoic and Lower Triassic rocks; it is characterized by folds that strike north-northeast and range from open to tight and overturned, by several thrust and tear faults, and by many small transverse faults. Jurassic and Cretaceous rocks that comprise the lower plate are also folded where exposed, but apparently they are not markedly folded beneath the thrust plate.

The Meade overthrust is a bedding-plane fault on which movement was initiated in the early stages of compression, the rocks of the upper plate being folded during thrusting. The horizontal displacement was probably at least 18 to 20 miles. There is no evidence of more than one major period of compression, and the thrust was probably folded in the later stages of the same orogeny that produced the thrusting. This major orogeny, the Laramide, can be dated directly only as post-Lower Cretaceous and pre-Eocene. Normal faults that in places offset Pliocene beds disrupt the earlier structure on the west side of the ranges. Some Recent normal faulting has occurred.

The Late Cretaceous or Paleocene orogeny produced land of sufficient height to supply coarse clastics, but the maximum elevation was probably considerably

less than it is today. In Eocene time red conglomerate was deposited across the area on a relatively smooth surface that beveled folds in the older rock. In Oligocene or Miocene time the present ranges and valleys were defined by warping or faulting, most of the red conglomerate was stripped, and the underlying pre-Eocene surface was dissected. Deposition of the Salt Lake formation in the valleys began at least by the close of late Miocene time and continued through most of the Pliocene. Normal faulting in late Pliocene or Pleistocene time tilted beds of the Salt Lake formation, and a pediment was developed across them. The pediment has since been extensively dissected.

Thick extensive beds of phosphate rock occur in the Meade Peak member of the Phosphoria formation, but they are being mined only in Georgetown Canyon. The phosphatic beds are rarely exposed at the surface and must be sampled in bulldozer trenches and drill holes. Several outcrop belts of the Meade Peak member probably could be strip mined, but, except for those in Georgetown Canyon, they are distant from railroads and paved highways.

INTRODUCTION

LOCATION

The Georgetown Canyon-Snowdrift Mountain area (fig. 1) is in the Peale Mountains [1] of southeastern Idaho, about 6 miles north-northeast of Montpelier, Idaho, 11 miles east-southeast of Soda Springs, Idaho, and 2¾ miles west of the Idaho-Wyoming State line. It includes parts of Bear Lake and Caribou Counties, and most of the area is within Caribou National Forest. The area consists of two adjacent 7½-minute quadrangles and parts of two others.

PURPOSE OF THE WORK

Permian strata in the Peale Mountains contain some of the thickest and highest grade deposits of phosphate rock in the western phosphate field. To aid in the selection of possible sites for mining and to serve as a basis for calculating reserves, the U.S. Geological Survey is mapping seven 7½-minute quadrangles within the central Peale Mountains. Geologic maps of two of these quadrangles, the Dry Valley (Cressman and Gulbrandsen, 1955) and the Johnson Creek (Gulbrandsen and others, 1956), have already been published. A geologic map of the Snowdrift Mountain quadrangle, previously released in preliminary form (Cressman, 1957), is published in this report at a scale of 1:24,000 as plate 1, and structure sections through the same area are shown on plate 2.

The geology of southeastern Idaho has been described in a comprehensive report by George R. Mansfield (1927), who considered the northern two-thirds of the Peale Mountains to be part of the great Bannock overthrust sheet that had been recognized and named by Richards and Mansfield in 1912. Mapping of the Dry Valley and

[1] The Peale Mountains (fig. 2) were named by Mansfield (1927, p. 24) for A. C. Peale of the Hayden surveys. The term is not frequently used by residents of the area, but it is extremely useful in discussions of regional geography and geology.

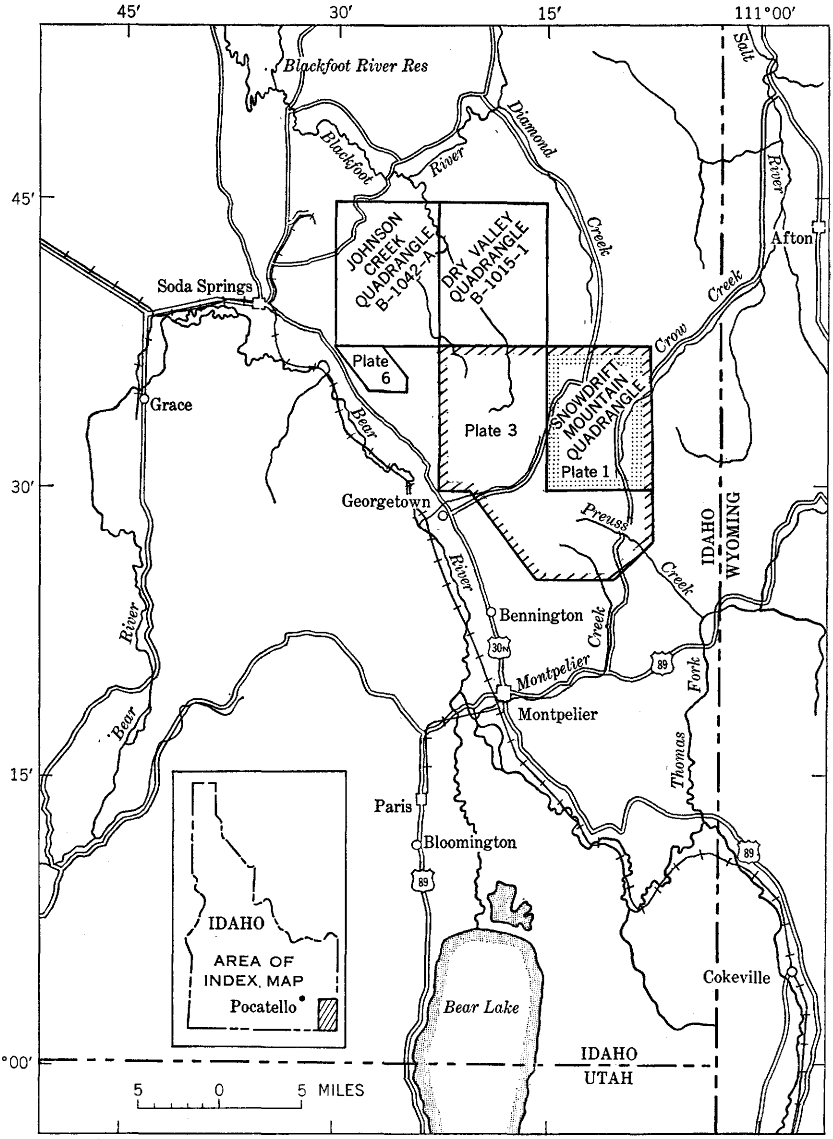

FIGURE 1.—Index map of southeastern Idaho showing location of area mapped for this report and areas covered by published 1:24,000 geologic maps.

Snowdrift Mountain quadrangles raised some questions about parts of the earlier structural interpretation (Cressman and Gulbrandsen, 1955, p. 265); therefore the region of relatively low-angle thrusting in Georgetown Canyon and around the south end of Snowdrift Mountain, critical to the concept of the Bannock overthrust, was remapped.

The resulting geologic map, combined with a somewhat generalized map of the Snowdrift Mountain quadrangle, is published here at a scale of 1:48,000 (pls. 3, 4).

FIELDWORK

The fieldwork on which this report is based was done during the summers of 1953 through 1957. Alvin F. Holzle assisted in the work during part of the summer of 1953 and William Glen during 1956. Charles H. Marshall measured several stratigraphic sections in June of 1955.

The Snowdrift Mountain quadrangle was mapped on a topographic base with a 20-foot contour interval. The 7½-minute quadrangle that adjoins the Snowdrift Mountain quadrangle on the west and the area south of the two quadrangles were mapped mostly on aerial photographs and transferred to a topographic base with a 100-foot contour interval by means of a Kail plotter. A small part near the west edge of the quadrangle was mapped directly on a topographic base (contour interval of 100 feet) by open-sight alidade.

Color names conform with the Rock-Color Chart distributed by the Geological Society of America (Goddard and others, 1948).

PREVIOUS WORK

Geologic studies in southeastern Idaho before 1927 have been summarized by Mansfield (1927, p. 5). This Professional Paper has been the one indispensable work on the geology of the region and will remain so for years to come. The Snowdrift Mountain quadrangle of this report is the southwest quarter of the Crow Creek 15-minute quadrangle of Mansfield's report, the 7½-minute quadrangle west of the Snowdrift Mountain quadrangle is the southeast quarter of the Slug Creek 15-minute quadrangle, and the area mapped south of these two 7½-minute quadrangles is in the northern part of the Montpelier 30-minute quadrangle. Reports dealing largely with phosphate deposits are cited on pages 90–92.

ACKNOWLEDGMENTS

I have benefited greatly from many discussions in both field and office with F. C. Armstrong and T. M. Cheney, who have been mapping adjacent areas, and with V. E. McKelvey, who organized and directed the investigation of the western phosphate field.

GEOGRAPHY

The northward-trending valley of Crow Creek, ½ to 1 mile wide along much of its course, divides the Georgetown Canyon-Snowdrift Mountain area into two parts. The Gannett Hills lie east of the valley and constitute about one-fifth of the area; parts of several ranges included in the Peale Mountains form most of the area, west of the valley (fig. 2). Bear River and Bear Lake valleys lie just west

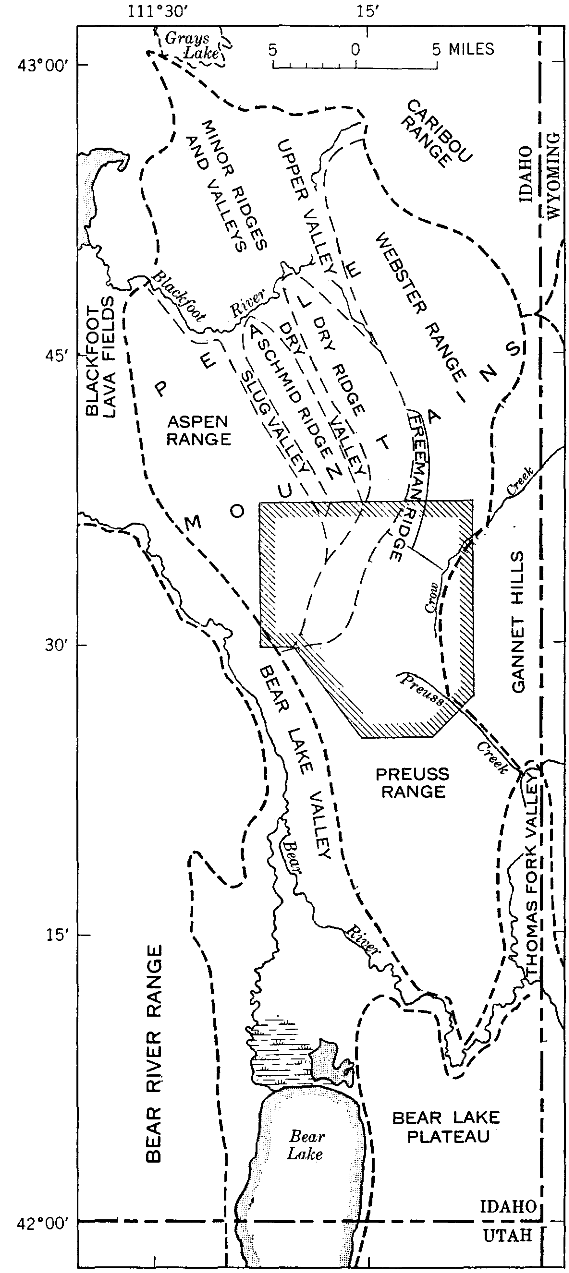

FIGURE 2.—Map of southeastern Idaho showing the principal physiographic features.

of the area. Red Mountain in the southern part of the Gannett Hills rises 8,840 feet above sea level and stands nearly 2,000 feet above Crow Creek valley; but elsewhere in the Gannett Hills, although hillsides are steep, the local relief generally does not exceed 1,100 or 1,200 feet. The mountains west of Crow Creek consist of northward-trending ridges that are separated by narrow valleys and canyons and are cut by several steep, narrow transverse canyons, such as the canyon of Deer Creek, Wells Canyon, and lower Georgetown Canyon. The highest point in the area is Meade Peak at an altitude of 9,960 feet; the lowest point is 6,200 feet near the mouth of Georgetown Canyon. In general, the altitude of the major divides in the Peale Mountains decreases from east to west; thus, the crest of Snowdrift Mountain is mostly more than 9,000 feet and that of Dry Ridge between 8,500 and 9,000 feet, whereas most crests in the Aspen Range are between 8,000 and 8,500 feet above sea level. Local relief is as much as 3,300 feet near Meade Peak and between 2,000 and 3,000 feet throughout much of the Preuss Range, but it is only about 1,600 feet in the Aspen Range. Slopes are steep throughout most of the area. Divides are rather sharp in the Preuss Range but they are somewhat more rounded in the Aspen Range.

The Great Basin-Columbia River drainage divide crosses the area from southeast to northwest. Crow and Slug Creeks drain into the Columbia by way of the Blackfoot and Snake Rivers; Preuss, Montpelier, and Twin Creeks, drain into Bear River which empties into the Great Salt Lake.

North- and east-facing hillsides are generally forested with aspen, Douglas-fir, and lodgepole pine; alpine fir and Engelmann spruce grow in some of the moist mountain valleys and on some of the higher slopes and limber pine is found on the highest divides. South- and west-facing slopes are not generally forested, but are covered with mixed grass and sagebrush or with mountain-lilac, serviceberry, chokecherry, and other mountain brush. Mountain mahogany grows on some relatively dry slopes along Lower Georgetown Canyon and near the west front of the Aspen Range. Sagebrush is found at all elevations.

The mapped area is crossed by several dirt roads from which even the most remote places can be reached by foot in 1 or 2 hours. U.S. Highway 30 N. and the Union Pacific Railroad pass through Bear Lake and Bear River valleys, and a railroad spur has recently been constructed up Georgetown Canyon to the Central Farmers Fertilizer Co.'s plant at the mouth of Phosphoria Gulch.

STRATIGRAPHY

The rocks exposed in the Georgetown Canyon-Snowdrift Mountain area (pl. 5) range in age from Early Mississippian to Recent, and except for several patches of basalt of late Tertiary age, they are all sedimentary. The rocks from Early Mississippian to Early Cretaceous age comprise an apparently conformable sequence about 24,000 feet thick. Presentage assignments indicate breaks in deposition within this sequence during the Late Pennsylvanian, Middle Triassic, and possibly during the Early Pennsylvanian, Late Permian, latest Jurassic and earliest Cretaceous. Strata of Early Mississippian through Jurassic age are mostly limestone, quartz sandstone, and shale deposited in a shallow marine environment. Lower Cretaceous rocks are continental red sandstone and conglomerate derived from an orogenic land a few tens of miles west of the area and deposited in a rapidly subsiding marginal basin. Post-Cretaceous formations are all postorogenic continental deposits—in part fluviatile, in part lacustrine, and in part the products of mass wasting.

The thickness of pre-Mississippian Paleozoic rocks that underlie the area is not known; rocks of this age in the Bear River Range to the west are 14,000 feet thick (Armstrong, 1953) but they are only 2,300 feet thick in the Bedford quadrangle, Wyoming, 15 miles northwest of the Snowdrift Mountain quadrangle (Rubey, 1958). The most reasonable interpolation yields a thickness of 5,000 to 10,000 feet for these strata in the Georgetown Canyon-Snowdrift Mountain area.

No systematic attempt was made to collect fossils, and except for the Wells formation, no new information has been obtained on the age of the stratigraphic units. Comprehensive discussions of the fauna and ages of the formations are given in Professional Paper 152 (Mansfield, 1927).

MISSISSIPPIAN SYSTEM

The oldest rocks exposed in the mapped area are of Mississippian age; moreover, no older strata are found anywhere within the Peale Mountains. The Mississippian beds comprise two formations, the Madison limestone and the Brazer limestone. Although the two formations differ in gross aspect, similar rock types are found in both, and it is difficult to map a consistent contact between them in areas of complex structure.

MADISON LIMESTONE

The type area of the Madison limestone is in the vicinity of Three Forks, Mont., where it was named by Peale in 1893. Collier and Cathcart (1922) raised the Madison to group rank in the Little Rocky Mountains of Montana and divided it into two formations—a lower thin-bedded unit named the Lodgepole limestone and an upper massive poorly bedded unit named the Mission Canyon limestone. This nomenclature is now generally followed throughout western Montana. The term Madison was applied to lower Mississippian rocks of the Randolph quadrangle of northern Utah by Richardson (1913) and to equivalent strata in southeastern Idaho by Mansfield (1927, p. 60).

The base of the Madison limestone is not known to be exposed in the Georgetown Canyon-Snowdrift Mountain area, although some poorly exposed grayish-yellow fine-grained calcareous sandstone on the south side of lower Georgetown Canyon may conceivably be the shoreward equivalent of the Leatham formation of Holland (1952), which is the basal Mississippian unit of the Bear River Range. The Madison either forms the cores of anticlines or is terminated downward by thrust faults. The formation is at least 1,000 feet thick; judging from structure section *F–F'* of plate 4, it may be more than 1,500 feet thick near Meade Peak, but this area is structurally complex and the thickness may not be reliable.

The Madison limestone typically consists of dark-gray to black finely crystalline to aphanitic limestone in beds 6 inches to 1 foot thick. Much of the limestone closely resembles basalt in both color and texture on fresh surfaces. North of Georgetown Canyon the formation contains several massive beds of light-gray medium-crystalline and coarsely crystalline crinoidal limestone, similar to much of the limestone in the overlying Brazer formation. The uppermost and most persistent of the massive crinoidal beds forms a conspicuous cliff on the hill between the two forks of Twin Creek and is easily seen from the highway just south of Georgetown; it is well shown by one of Mansfield's photographs (1927, pl. 38*A*). The crinoidal limestone beds apparently pinch out to the south, and none seem to be present at the south end of Snowdrift Mountain.

The uppermost massive crinoidal limestone of the Madison is overlain by an inch or so of nearly black coarsely pelletal phosphorite that has been mapped as a marker bed. Overlying the phosphorite is 150 or 200 feet of thin-bedded aphanitic limestone that is similar to that in the rest of the Madison, but it contains many irregular layers and nodules of black chert mottled brown and black on weathered surfaces. (Similar black chert occurs above some of the other crinoidal limestones, but not in such abundance nor throughout so great a thickness.) The contact between the Madison and the overlying Brazer formation has been placed at the base of another massive crinoidal limestone bed that overlies the aphanitic limestone-black chert unit.

The crinoidal limestone beds generally crop out conspicuously, but the rest of the Madison is seldom well exposed and commonly forms steep slopes littered with yellowish-gray-weathering limestone float. The thin-bedded limestone has been particularly susceptible to brecciation, and major thrusts that transect the formation are bordered by a zone from 10 to more than 100 feet thick of recemented yellowish-orange breccia. The Madison is also extensively brecciated adjacent to the normal faults that are common in the Aspen Range, and the conspicuous brecciated area in Big Canyon was derived mostly, if not entirely, from the Madison.

Holland (1952, p. 1731) noted the similarity between the Madison limestone of the Bear River Range and Lodgepole limestone of Montana, and Sando, Dutro, and Gere (1959) have introduced the name Lodgepole for the equivalent unit in the Crawford Mountains. The Madison of the Georgetown Canyon-Snowdrift Mountain area is also very similar in lithology to the type Lodgepole, but because the continuity of these beds with the Lodgepole of the type area has not as yet been demonstrated, the name Madison is retained in this report.

The Madison limestone is Early Mississippian in age (Sando and others, 1959).

BRAZER LIMESTONE

The Brazer limestone was named by Richardson (1913) for Brazer Canyon in the Crawford Mountains near Randolph, Rich County, Utah, and the name was shortly thereafter extended to equivalent rocks in southeastern Idaho by Mansfield and his associates. The Brazer has recently been restudied in its type area by Sando, Dutro, and Gere (1959), who describe the formation as being predominantly dolomite. They recommend that the name Brazer be restricted to the dolomite facies, but they do not propose a name for the limestone and sandstone facies of southeastern Idaho and adjacent areas where the name Brazer has long been used. The same formational name probably should not be applied to such diverse units as the dolomite of the type section, the predominantly light-colored massively bedded crinoidal limestone and sandstone of the Peale Mountains, and the dark-gray cherty and argillaceous limestone of Dry Lake near Logan, Utah (Williams, 1943, p. 596); but the facies relationships are not sufficiently known to warrant renaming the formation in the area described in this report.

The Brazer limestone as mapped in the Georgetown Canyon-Snowdrift Mountain area is 1,600 to 2,000 feet thick; most of the thicknesses measured from the geologic map are about 1,600 feet. A well drilled by the Standard Oil Co. of California in Dry Valley about 10 miles north of the area of this report passed through more than 7,000 feet of Brazer (Young, 1953), but no surface exposure anywhere in the vicinity suggests a thickness of even one-third of this figure.

The following section, measured by T. M. Cheney on the north side of Wells Canyon in the Snowdrift Mountain quadrangle, totals 1,665 feet. The base is not exposed, but the sequence of beds indicates that not more than 200 or 300 feet are missing.

Section of Brazer limestone

[Measured at mouth and on north side of Wells Canyon in the SW¼SW¼ sec. 11, SE¼ sec. 10, T. 10 S., R. 45 E., Caribou County, Idaho. Thickness-of-bedding terms are defined as follows: Massively bedded, greater than 1 ft; thick bedded, 0.6 to 1 ft; thin bedded, less than 0.6 ft. Crystal-size terms are coarsely crystalline, greater than 4 mm; medium crystalline, 1 to 4 mm; finely crystalline, less than 1 mm; dense crystals not visible to unaided eye. Grain-size terms of sandstone are those of Wentworth (1922)]

Unit	*Thickness (feet)*	*Cumulative thickness*	*Description*
			Wells formation, basal beds only
93	10. 0		Limestone and sandstone interbedded: sandy beds are laminated (laminae averaging 1/16 to ¼ in. thick) and crossbedded.
			Brazer limestone
92	8. 0	8. 0	Limestone: upper 5.0 ft dense to fine-grained and gray; contains more than 20 percent chert nodules and lenses; lowermost 3.0 ft medium crystalline. Unit thick-bedded to massively bedded.
91	1. 4	9. 4	Limestone: sandy, light-gray (weathers brown), thin-bedded. Contains laminae of well-rounded quartz grains and is locally crossbedded.
90	23. 0	32. 4	Limestone: dense to fine-grained, light-gray, thick-bedded to massively bedded; forms 10 to 16 ft cliff; brachiopods and horn corals in 6-ft zone 4 ft above base of unit.
89	26. 0	58. 4	Limestone: slightly sandy, very light yellow, thick-bedded to massively bedded, porous.
88	42. 4	100. 8	Limestone: dense to medium-crystalline, thick bedded to massively bedded; brachiopods and horn corals in 3-ft zone 12 ft above base of unit.
87	40. 3	141. 1	Limestone: medium-crystalline to coarsely crystalline, gray, thick-bedded; interval is 20 percent covered with silty limestone in float.
86	37. 5	178. 6	Limestone: uppermost 2 ft argillaceous(?) dolomitic(?); 27 to 29 ft medium to coarsely crystalline, thick bedded; lowermost 6–8 ft fine grained, gray, thick to massively bedded, cliff forming; blastoids, and horn corals, in 2-ft zone 3 ft above base of unit.
85	2. 1	180. 7	Dolomite(?): calcareous, tan to light-pink, thin- to thick-bedded, slightly argillaceous.
84	1. 2	181. 9	Limestone: very fine grained, pink, thin-bedded.
83	2. 2	184. 1	Dolomite: calcareous, very fine grained, light-brown and very light brown, massively bedded.
82	40. 1	224. 2	Limestone: fine- to medium-grained, gray, thick-bedded; forms 3- to 4-ft ledges; 0.5-ft dolomite(?) bed similar to unit 83, is 37.5 ft above base; 10 percent of unit is covered.
81	48. 0		Covered.
80	9. 5	281. 7	Limestone: dense to fine-grained, dark-gray (weathers light brown), thin- to thick-bedded; some horn corals are at base of unit.

Brazer limestone—Continued

Unit	Thickness (feet)	Cumulative thickness	Description
79	17.0	298.7	Limestone: slightly sandy, medium-crystalline to coarsely crystalline, light-brown on weathered surface, thin- to thick-bedded; coarse sandpaper-like texture on weathered surface; large horn corals concentrated in 0.5-ft zone 16.0 ft above base of unit.
78	14.5	313.2	Limestone: fine-grained, dark-gray (weathers light gray), thick-bedded to massively bedded.
77	19.0	332.2	Covered.
76	78.8	411.0	Limestone: dense to fine-grained, dark-gray (weathers gray), thick-bedded to massively bedded; coarsely crystalline bed (1.0 ft thick) 77.8 ft above base. Calcite joints ramify throughout unit, and ferruginous material in calcite on joints gives brown splotchy appearance to weathered surface.
75	2.0	413.0	Sandstone: calcareous, fine-grained, purple to light-red, thick-bedded; contains some siltstone in uppermost part, and contact with overlying limestone is gradational.
74	107.3	520.3	Similar to unit 76.
73	3.6	523.9	Siltstone: calcareous, reddish-brown (weathers light brown), thick-bedded; rust-colored specks are on both fresh and weathered surfaces; lowermost part of bed is slightly silty limestone.
72	1.4	525.3	Limestone: similar to unit 76.
71	1.8	527.1	Siltstone: calcareous, reddish-brown (weathers light brown); unit contains about 5 percent fine-grained quartz sand.
70	5.5	532.6	Limestone: very fine grained, gray, thick-bedded to massively bedded; 0.6-ft zone of oolites 3.0 ft above base of unit.
69	3.2	535.8	Siltstone: calcareous, brown; lower contact is gradational through 1.0 ft.
68	2.7	538.5	Limestone: medium-grained, gray.
67	1.8	540.3	Limestone: silty, gray (slight reddish-brown tinge), thin-bedded.
66	39.3	579.6	Limestone: fine-grained, gray, thin-bedded to massively bedded; 10 percent of unit covered, and silty limestone fragments in talus indicate that this may be interbedded with limestone.
65	14.4	594.0	Siltstone and sandstone: upper 5 ft tan, thin- to thick-bedded siltstone; 1.4 ft very fine grained, massively bedded sandstone; lower 7 ft calcareous fine-grained tan fissile to thin-bedded sandstone.
64	28.4	622.4	Limestone: fine-grained, gray, thick-bedded to massively bedded; oolitic limestone 0.5 ft thick 5.0 ft above base, 0.5 ft thick 13.0 ft above base and 0.4 ft thick 15.0 ft above base; oolites not seen on fresh surface but are easily distinguished on weathered surface.

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Brazer limestone—Continued

Unit	*Thickness (feet)*	*Cumulative thickness*	*Description*
63	9. 5	631. 9	Sandstone: calcareous, fine-grained, light-brown (weathers tan), thick-bedded.
62	1. 0	632. 9	Limestone: dense, dark-gray (weathers light gray), massively bedded.
61	6. 2	639. 1	Sandstone: calcareous; similar to unit 63; contact with underlying unit gradational through 0.8 ft.
60	2. 0	641. 1	Limestone: similar to unit 62.
59	10. 2	651. 3	Sandstone: similar to unit 63.
58	4. 2	655. 5	Limestone: similar to unit 62.
57	4. 0	659. 5	Sandstone: similar to unit 63; contact with underlying unit gradational through 0.5 ft.
56	5. 8	665. 3	Limestone: dense; light-gray on weathered surface, dark-gray (weathers light gray); thick-bedded to massively bedded.
55	3. 0	668. 3	Sandstone: similar to unit 63.
54	0. 8	669. 1	Limestone: similar to unit 62.
53	2. 2	671. 3	Sandstone: similar to unit 63; contact with underlying unit gradational through 0.6 ft.
52	3. 8	675. 1	Limestone: massively bedded, similar to unit 62.
51	10. 5	685. 6	Covered: calcareous sandstone makes up 90 percent of talus fragments.
50	25. 4	711. 0	Limestone: fine- to medium-grained, gray, massively bedded; forms cliff or small ledges.
49	4. 5	715. 5	Limestone: fine-grained, grayish-brown, thick-bedded.
48	7. 6	723. 1	Sandstone: calcareous, fine-grained, grayish-brown (weathers dark tan), thin-bedded to massively bedded; cross laminated.
47	3. 5	726. 6	Limestone: silty, laminated (laminae composed of quartz grains); 0.5 ft of oolitic limestone at base of unit.
46	5. 8	732. 4	Limestone: fine-grained, gray, massively bedded; upper 1.5 ft dense light-brown limestone.
45	15. 2	747. 6	Sandstone: similar to unit 48; 80 percent covered, but float indicates that most of unit is calcareous sandstone.
44	2. 0	749. 6	Limestone: dense to fine-grained, gray, massively bedded.
43	22. 0	771. 6	Covered: except for a 0.7-ft bed of fine-grained gray limestone 12.0 ft above base of unit; float indicates unit is comprised of 80 percent calcareous sandstone and 20 percent limestone.
42	32. 5	804. 1	Limestone: fine- to coarse-grained, gray, thick-bedded to massively bedded limestone outcrops form small ledges, intervals between ledges containing fine-grained calcareous sandstone float; 20 percent covered.
41	6. 0	810. 1	Sandstone: very fine grained, very light brown, slightly calcareous, slightly dolomitic(?).

Brazer limestone—Continued

Unit	Thickness (feet)	Cumulative thickness	Description
40	8. 0	818. 1	Limestone: medium-grained in upper ⅓ of unit, dense to fine-grained in lower ⅔ of unit; thick-bedded, slightly dolomitic(?); forms cliff.
39	22. 5	840. 6	Covered: upper 15 ft contains float of very fine grained sandstone similar to unit 41; lower 6 ft is comprised for the most part of coarse-grained limestone chips.
38	6. 3	846. 9	Limestone: coarsely crystalline, dark-gray, thick-bedded; smells slightly petroliferous; contains 5 percent small chert nodules.
37	17. 5	864. 4	Limestone: sandy, fine-grained, gray, thin- to thick-bedded; smells slightly petroliferous; shaly bed 2 ft thick in upper part of unit; brachiopods and bryozoans in 2-ft zone 14.0 ft above base; brachiopods, bryozoans, and horn corals in 5-ft zone at base of unit.
36	17. 0	881. 4	Limestone: fine- to medium-grained, massively bedded; oolitic limestone 0.6-ft thick 8 ft above base of unit.
35	8. 0	889. 4	Limestone: fine- to medium-grained, gray, massively bedded; contains thin discontinuous sandstone laminae; forms 6- to 8-ft cliff.
34	5. 5	894. 9	Limestone: dense to fine-grained, gray, massively bedded; upper 2 ft of unit contains scattered coarsely crystalline white calcite grains.
33	45. 0	939. 9	Sandstone: medium-coarse-grained, soft, tan and red, cross laminated, thick-bedded to massively bedded, slightly calcareous sandstone and calcareous fine-grained hard dark-brown thick-bedded to massively bedded sandstone; unit is 50 percent covered, and float indicates some fine-grained limestone.
32	20. 0	959. 9	Sandstone: calcareous, very fine grained; upper ⅔ of unit hard thick bedded to massively bedded; lower ⅓ of unit thin-bedded; smells petroliferous.
31	21. 0	980. 9	Limestone: medium-crystalline to coarsely crystalline, brown and black (weathers dark gray and tan), thick-bedded to massively bedded; contains small horn corals; bryozoans in uppermost 5 ft.
30	6. 0	986. 9	Sandstone: calcareous, medium-coarse- to coarse-grained, light-gray, massively bedded, locally crossbedded; rock breaks through quartz grains giving quartzitic appearance on fresh surface.
29	8. 5	995. 4	Limestone: dense to fine-grained, light-gray to light-brown, thick-bedded; contains a few poorly preserved horn corals.
28	31. 5	1, 026. 9	Limestone: coarsely crystalline, light-gray to black (weathers light gray), massively bedded.

Brazer limestone—Continued

Unit	*Thickness (feet)*	*Cumulative thickness*	*Description*
27	13. 5	1, 040. 4	Limestone: dense to fine-grained, light-gray and brown, thick-bedded; some coarse fragments of crystalline calcite throughout unit; contains poorly preserved gastropods, brachiopod fragments, medium-sized horn corals, and crinoid stems; 20 percent of unit covered.
26	13. 0	1, 053. 4	Limestone: medium- to coarse-grained, light-gray to pale-brown, thick-bedded to massively bedded; contains some chert nodules.
25	40. 0	1, 093. 4	Limestone: dense to medium-grained, light-gray, thin-bedded to massively bedded limestone; contains some chert nodules and lenses.
24	32. 0	1, 125. 4	Limestone: medium- to coarse-grained, dark-gray on weathered surface, thick-bedded to massively bedded; contains a few poorly preserved horn corals.
23	31. 0	1, 156. 4	Limestone and chert: dense to medium-grained, dark-gray (on weathered surface), thick-bedded to massively bedded limestone containing chert nodules and lenses that comprise more than 20 percent of unit; horn corals in 4-ft zone 2.6 ft above base of unit; 30 percent of unit is covered.
22	5. 0	1, 161. 4	Limestone: similar to unit 25; 0.5 ft at base is laminated.
21	62. 0	1, 223. 4	Limestone: similar to unit 24; contains a few coarsely crystalline beds less than 1 ft thick; 10 percent of unit covered; brachiopods and horn corals in 5-ft zone 10 ft above base of unit.
20	23. 0	1, 246. 4	Covered: limestone and chert float similar to unit 23.
19	35. 0	1, 281. 4	Limestone: dense to medium-grained, pale-brown (weathers light gray), massively bedded; medium-grained material restricted to lower 4 ft of unit; a 2.0- to 4.0-ft zone of oolitic limestone is 15.0 ft above base; cliff forming in places.
18	15. 5	1, 296. 9	Limestone and chert: fine-grained, light-brown (weathers light gray), thick-bedded to massively bedded limestone with chert nodules and lenses constituting more than 20 percent of unit; many white calcite veinlets; 40 percent covered.
17	6. 0	1, 302. 9	Limestone: fine-grained, dark-gray (weathers pale brownish gray), thick-bedded to massively bedded; contains some small specks of red ferruginous material.
16	7. 5	1, 310. 4	Sandstone: calcareous, fine- to medium-coarse-grained; light brown on weathered surface; massively bedded; upper 0.5 ft is calcareous siltstone; 50 percent of unit covered.
15	20. 5	1, 330. 9	Limestone: dense to fine-grained, gray to light-brown (weathers light gray), thick-bedded to massively bedded; contains 0.6-ft oolitic limestone 8.5 ft above base of unit; forms small ledges.

Brazer limestone—Continued

Unit	*Thickness (feet)*	*Cumulative thickness*	*Description*
14	18. 8	1, 349. 7	Oolitic limestone and limestone interbedded: dark-gray thick-bedded to massively bedded oolitic limestone interbedded with fine- to medium-grained orange thick-bedded to massively bedded limestone; oolitic limestone constitutes about 70 percent of unit; 10 percent of unit is covered with float containing some sandy limestone underlying orange limestone units.
13	32. 5	1, 382. 2	Sandstone: calcareous, light-brown to yellow, thin- to massive-bedded, very fine to fine-grained; locally laminated; 1-ft bed of oolitic limestone is 14.0 ft above base of unit; quartz grains grade from ¼ mm at base of unit to 1/16 mm at top; 60 percent of unit is covered with talus containing about 20 percent of limestone fragments.
12	37. 0	1, 419. 2	Limestone: coarsely crystalline limestone (20 percent) interbedded with fine- to medium-grained limestone (60 percent); both are gray to brown (weather very pale brown), thick-bedded to massively bedded; coarse-grained units are thickest near base; chert nodules and lenses occur in fine- to medium-grained limestone.
11	49. 0	1, 468. 2	Limestone: fine- to medium-grained; dark gray and grayish brown on fresh surface; massively bedded; forms series of small ledges.
10	38. 0	1, 506. 2	Sandstone: calcareous, fine-grained; light brown on fresh surfaces; massively bedded and laminated; quartz grains subrounded; 60 percent of unit covered, 40 percent of talus being limestone fragments.
9	4. 0	1, 510. 2	Limestone: dense to fine-grained, light-brownish-gray, massively bedded; oolitic limestone makes up 10 to 70 percent of unit and is concentrated in 1- to 2-ft lenticular zones.
8	22. 5	1, 532. 7	Limestone: sandy, brown (weathers light brown), thin-bedded to massively bedded; smells petroliferous; medium-sized horn corals in upper 5 ft of unit, brachiopods and bryozoans in thin-bedded lower 6 ft of unit.
7	5. 7	1, 538. 4	Limestone: coarsely crystalline; dark gray on weathered surface; thick bedded; contains horn corals.
6	17. 0	1, 555. 4	Limestone: fine- to coarse-grained limestone interbedded with fine-grained sandy limestone; chert nodules and lenses are throughout unit; contains horn corals.
5	11. 0	1, 566. 4	Limestone: fine- to medium-grained; dark gray on fresh surface; thick bedded; contains scattered chert nodules.

Brazer limestone—Continued

Unit	Thickness (feet)	Cumulative thickness	Description
4	22. 0	1, 588. 4	Limestone: sandy, fine- to coarse-grained, thick-bedded to massively bedded; quartz grains are in laminae; 25 percent of unit covered with float that indicates some fine-grained limestone in upper half of unit.
3	3. 0	1, 591. 4	Limestone: coarsely crystalline; dark gray on fresh surface; thick bedded.
2	38. 5	1, 629. 9	Limestone and chert: limestone is dense to fine grained, dark gray on fresh surface, thick to massively bedded; chert nodules and lenses constitute more than 20 percent of unit; coarsely crystalline limestone in thin beds or lenses make up less than 10 percent of unit; horn corals in 0.3- to 1.0-ft zone 27.5 ft above base of unit and in 0.3-ft zone 6.0 ft above base of unit.
1	35. 0	1, 664. 9	Limestone: coarsely crystalline; light to dark gray on fresh surface; massively bedded. Base of bed is approximate axis of Boulder Creek anticline.

Characteristic rock types of the Brazer are (1) gray to light-gray massively bedded medium-crystalline and coarsely crystalline crinoidal limestone that locally contains many horn corals (most of the medium-crystalline and coarsely crystalline limestone in the Wells Canyon section is crinoidal); (2) dark- and medium-gray lithographic limestone; (3) pelletal limestone that grades laterally into lithographic limestone like that described above (most of the "oolites" described in the Wells Canyon section are not concentrically laminated); (4) tan medium-grained well-sorted quartz sandstone, much of which is friable; (5) light-brown to reddish-brown calcareous siltstone. Of these, the medium-crystalline and coarsely crystalline limestone is the most common and characteristic. The various rock types are arranged in a sequence recognizable throughout the area.

The basal 200 to 400 feet of the Brazer consists of light-gray finely to coarsely crystalline crinoidal limestone that contains some intercolated beds of lithographic limestone and some of dark aphanitic limestone similar to that in the Madison. Very irregular nodules of chert are scattered throughout. The unit generally crops out in cliffs and ledges, and the contact with the underlying finer grained and less resistant Madison limestone could generally be located in mapping to within the error of plotting at the compilation scale of 1:40,000. In the Wells Canyon section beds 9 and below probably belong to this interval.

Overlying the basal unit is 600 to 900 feet of interbedded sandstone, crinoidal limestone, and lithographic limestone; these are beds 11 through 65 in the Wells Canyon section. The sandstone is concentrated at the top and the base of this unit at Wells Canyon, but in the more western exposures sandstone is interbedded throughout the interval. Some chert nodules occur in the lower part. In the Aspen Range the base of the sandy interval has been mapped where it could be located with confidence, but no attempt has been made to project the contact through areas of poor exposure.

A few beds of reddish-brown siltstone are intercalated in limestone at the top of the sandy sequence, but otherwise nearly all the uppermost 600 feet of the Brazer consists of limestone. From about 600 to about 350 feet below the top of the formation the limestone is the dark lithographic type. These beds are capped by 50 to 100 feet of limestone, mostly crinoidal, that commonly forms ledges, ridges, and cliffs and locally contains many large well-preserved horn corals. A well-defined swale that contains distinctive black and white laminated siliceous shale float generally is developed on the 50 feet or so of beds that overlie the crinoidal limestone. The uppermost 150 to 250 feet of beds, those above the swale, are light-colored massively bedded resistant limestone, some of which is crinoidal but most of which is aphanitic and dense and has a faint suggestion of bioclastic textures on fresh wet surfaces. Irregular chert nodules occur in the uppermost part. The uppermost two resistant limestone units separated by the swale containing the laminated shale form a distinctive double ridge or cliff, conspicuous on aerial photographs, that is recognizable over the entire area.

All the units described above could be mapped separately if exposures were good and the structure simple, but the rock types are so similar from the top to the base of the formation that the Brazer could not be divided into members suitable for mapping in the complexly folded and faulted areas.

The massive limestones of the Brazer are resistant to weathering and underlie the highest peaks and ridges, such as Meade Peak, Snowdrift Mountain, Dry Ridge, and Harrington Peak.

In north-central Utah a black phosphatic shale member is present at the base of the Brazer limestone; the nearest exposure of the member to the area of this report is at Laketown Canyon at the south end of Bear Lake (Cheney, 1957, p. 12). The thin phosphatic bed that is about 200 feet below the Brazer-Madison contact as mapped in the Aspen Range and the overlying cherty limestone in the uppermost part of the Madison may be the lateral equivalent of the phosphatic shale.

There is no evidence within the area of any major and widespread unconformity within the Brazer, between the Brazer and Madison, or in the upper part of the Madison. According to Williams and Yolton (1945) and Parks (1951), a major disconformity separates the Brazer limestone and the overlying Wells formation in the Bear River Range. In the Peale Mountains, however, the widespread continutiy of the uppermost two ledge-forming limestone units and the intervening laminated shale of the Brazer strongly argues against a disconformity at this horizon.

The age of the Brazer limestone has generally been considered to be Late Mississippian, but Sando, Dutro, and Gere (1959) report that in the type area only the upper third is of Late Mississippian age.

PENNSYLVANIAN AND PERMIAN SYSTEMS

WELLS FORMATION

The Wells formation was named by Richards and Mansfield (1912, p. 689–693) for Wells Canyon in the Snowdrift Mountain quadrangle. The type section measured by them on the north side of Wells Canyon totaled 2,400 feet thick and was informally divided into an upper siliceous limestone 50 feet thick, a middle sandy series 1,600 feet thick, and a lower sandy and cherty limestone series 750 feet thick.[2] The Wells is from 1,500 to 2,000 feet thick in other parts of the area, and the considerably greater thickness of 2,400 feet in the type section most probably results from either a gentle flexure, such as that shown in structure section *C-C′* of plate 2, or from repetition by normal faults.

The following section was measured by A. F. Holzle where Deer Creek crosses the east limb of Snowdrift anticline and is about 2 miles from the type section. The exposures are far from complete, but this is the best exposed continuous section of the Wells formation within the area.

Section of Wells formation

[Measured on north side of Deer Creek, SE¼ sec. 33, T. 9 S., R. 45 E., Caribou County, Idaho. Thickness-of-bedding terms are defined as follows: Massively bedded, greater than 1 ft; thick bedded, 0.6 to 1 ft; thin bedded less than 0.6 ft. Crystal-size terms are coarsely crystalline, greater than 4 mm; medium crystalline, 1 to 4 mm; finely crystalline, less than 1 mm; dense crystals not visible to unaided eye. Grain-size terms of sandstone are those of Wentworth (1922)]

Grandeur tongue of Park City formation

[Contact with Phosphoria formation not exposed]

Unit	*Thickness (feet)*	*Cumulative thickness*	*Description*
71	15	15	Dolomite: hard, massive, light-brownish-gray (near top) to light-gray (near base), finely crystalline; contains dark-blue to black chert layers and nodules that weather white.
70	2	17	Sandstone: white, very fine grained to medium-grained.

[2] The Wells has recently been redefined to exclude the upper siliceous limestone, now termed the "Grandeur tongue of the Park City formation" (McKelvey and others, 1956, p. 2842, and 1959, p. 15). In this report the Grandeur is described separately, but it was mapped with the upper member of the Wells formation.

Grandeur tongue of Park City formation—Continued

Unit	*Thickness (feet)*	*Cumulative thickness*	*Description*
69	8	25	Limestone: silty, hard, massive, light-brown, cliff forming.
68	15	40	Limestone: thin- to thick-bedded, very fine grained; upper 5 ft light brown, rest light gray.
			Wells formation, upper member
67	5	45	Limestone: thin- to thick-bedded, reddish-brown, coarsely crystalline.
66	12	57	Limestone: thin-bedded, dark-brown, dense; contains chert layers near base.
65	130	187	Covered.
64	12	199	Sandstone: massive, white, very fine grained to medium-grained, poorly sorted; grades from bed below.
63	5	204	Sandstone: calcareous, hard, massive, light-reddish-brown, very fine grained.
62	10	214	Sandstone: calcareous, massive, yellowish-brown to white, very fine grained to medium-grained, poorly sorted.
61	3	217	Sandstone: calcareous, hard, massive, weak reddish-white, very fine grained to fine-grained, poorly sorted.
60	20	237	Sandstone: massive, bright-reddish-brown, very fine grained to fine-grained, poorly sorted; weathered outcrops are rounded and pockmarked; forms prominent ledge.
59	1.5	238.5	Sandstone: white, very fine grained to fine-grained, poorly sorted.
58	4	242.5	Sandstone: thick-bedded, yellowish-brown, very fine grained to fine-grained, poorly sorted.
57	3.5	246	Sandstone: crossbedded, white (weathers reddish brown in part), very fine grained to fine-grained, poorly sorted.
56	3	249	Limestone: silty(?), massive, light-brown, finely crystalline.
55	4	253	Sandstone: massive, reddish- and yellowish-gray, fine-grained.
54	8	261	Sandstone: massive, light-gray, fine-grained.
53	30	291	Covered.
52	7	298	Sandstone: massive, pale-yellowish-white, very fine grained to fine-grained, poorly sorted; lower 1 to 2 ft has light-reddish cast and weathers deep red; forms conspicuous ledge.
51	5	303	Sandstone: thick-bedded, yellowish-white, very fine grained to fine-grained, poorly sorted.
50	5	308	Sandstone: massive, white, fine-grained.
49	10	318	Sandstone: massive, yellowish-white to buff, very fine grained to fine-grained, poorly sorted.
48	40	358	Covered.

Wells formation, upper member—Continued

Unit	Thickness (feet)	Cumulative thickness	Description
47	10	368	Sandstone: massive, light-brown to buff, fine-grained.
46	8	376	Sandstone: thick-bedded, white, very fine grained.
45	3	379	Limestone: sandy, hard,thin-bedded, light-yellowish-gray, dense; reddish-gray near top.
44	24	403	Sandstone and limestone interbedded; thin-bedded buff fine-grained sandstone (40 percent) interbedded with sandy limestone (10 percent).
43	65	468	Covered.
42	14	482	Sandstone: massive, very pale yellowish white, fine-grained; contains alternating laminae of hard and soft sandstone; cliff former.
41	10	492	Sandstone: thick-bedded, very fine grained.
40	12	504	Sandstone: massive, light-yellowish-brown (weathers light brown), very fine grained to fine-grained, poorly sorted.
39	55	559	Covered.
38	18	577	Sandstone: massive, yellowish-brown (weathers light-brown), fine-grained.
37	110	687	Covered.
36	10	697	Sandstone: hard, thick-bedded, light-reddish-gray to yellowish-gray (weathers light gray and light reddish gray), very fine grained.
35	4	701	Sandstone: calcareous, hard, massive, light-reddish-yellow, very fine grained.
34	21	722	Sandstone: hard, massive, reddish-brown, very fine grained.
33	160	882	Covered.
32	15	897	Sandstone: thick-bedded, reddish-gray, very fine grained to fine-grained, poorly sorted.
31	85	982	Covered.

Wells formation, lower member

Unit	Thickness (feet)	Cumulative thickness	Description
30	25	1, 007	Limestone: slightly sandy, hard, thick-bedded, dark-reddish-gray; lower 3 ft are dark brownish-gray.
29	4	1, 011	Limestone: sandy, massive, light-brown.
28	8	1, 019	Limestone: massive, dark-gray, coarsely crystalline.
27	40	1, 059	Limestone: fossiliferous, light-brownish-gray (weathers light brown); sandy in upper part; contains large chert nodules throughout unit and beds of chert in 20-ft zone in middle of unit; fossils stand out boldly on weathered surfaces.
26	12	1, 071	Limestone: sandy, hard, massive, light-brownish-gray.
25	5	1, 076	Sandstone: calcareous, thin-bedded, light-gray.
24	11	1, 087	Limestone: thin- and thick-bedded, light-brownish-gray; sandy in part; chert nodules in upper part, chert layers in basal 3 ft.
23	2	1, 089	Limestone: argillaceous, thin-bedded, light-brown, dense to very finely crystalline.

Wells formation, lower member—Continued

Unit	*Thickness (feet)*	*Cumulative thickness*	*Description*
22	6	1, 095	Limestone and chert interbedded: hard thick-bedded light-brownish-gray slightly sandy limestone (80 percent) interbedded with irregularly bedded dark-blue chert; chert is nodular in basal 2 ft.
21	3	1, 098	Limestone: hard, massive, light-brown, coarsely crystalline; contains numerous small chert concretions.
20	6	1, 104	Sandstone: slightly calcareous, hard, massive, medium-brown, very fine grained.
19	12	1, 116	Covered.
18	5	1, 121	Limestone: massive, light-brown to gray, finely crystalline; contains numerous chert nodules in upper 3 ft; lower 2 ft is dark gray, coarsely crystalline, and sandy; brachipods in upper part.
17	2	1, 123	Sandstone: hard, massive, reddish-brown (weathers light brown), fine-grained.
16	5	1, 128	Sandstone: hard, massive, light-gray (weathers light brown), very fine grained to fine-grained, poorly sorted.
15	20	1, 148	Limestone: sandy, thick-bedded, light-gray.
14	10	1, 158	Limestone: similar to unit 15 but contains many chert nodules.
13	5	1, 163	Limestone: similar to unit 15.
12	5	1, 168	Limestone: hard, dark-gray.
11	8	1, 176	Limestone: similar to unit 15.
10	2	1, 178	Sandstone: light-reddish-brown, fine-grained.
9	4	1, 182	Limestone: massive, light-gray; contains a few chert nodules.
8	6	1, 188	Sandstone: thick-bedded, yellowish-brown to buff, fine-grained.
7	10	1, 198	Limestone: grades from dark gray and coarsely crystalline at top to medium gray and very finely crystalline at base.
6	9	1, 207	Sandstone: calcareous, massive, medium-gray (weathers light tan).
5	30	1, 237	Covered.
4	6	1, 243	Limestone: slightly sandy, massive, light-brownish-gray (weathers light gray), dense.
3	10	1, 253	Limestone: sandy, massive, brownish-gray (weathers light gray), finely crystalline.
2	18	1, 271	Limestone: massive, light-gray, dense; sandy in upper 8 to 10 ft, chert nodules in lower 4 ft.
1	300	1, 571	Mostly covered, although some sandstone ribs are exposed. Base of covered interval is top of the uppermost ledge of the Brazer limestone.

LOWER MEMBER

The lower member of the Wells formation thins from 800 or 900 feet in the Aspen Range to 500 or 600 feet in the eastern part of the area. A thickness of 1,500 feet was noted a few miles north of the Snowdrift Mountain quadrangle (Cressman and Gulbrandsen, 1955, p. 260), and Richards and Mansfield (1912, p. 692) record a thickness of only 100 feet 2 miles north of the type section; the greater figure, however, may be the result of repetition by faulting, and the lesser thickness of 100 feet could not be confirmed.

The lower member consists of limestone and sandstone in beds several feet thick that contrast with the massive ledges of the uppermost Brazer even when seen from some distance. The member is fairly well exposed and typically crops out as a series of low ribs on grass and sagebrush slopes. The contact with the Brazer can be located closely in many areas by the abrupt change of bedding characteristics and by the first appearance of Wells-type sandstone.

The lower part of the lower member (the covered interval of 300 ft at the base of the formation in the Deer Creek section) is dominantly brown-weathering, gray very finegrained to fine-grained calcareous quartz sandstone with some interbedded sandy limestone, but the upper part is mostly gray limestone and sandy limestone with some intercalated sandstone. Chert nodules and layers are present throughout, but they are particularly abundant in the uppermost 100 feet; chert nodules 0.5 to 0.8 foot in diameter in limestone 50 to 100 feet below top of the member form a horizon recognizable throughout much of the area. Some limestone is composed largely of flattened oolites and is diagnostic of the member. Crinoid plates are present in small amounts in many beds, and brachiopods are abundant in others; fusulinids and horn corals have been found, but they are not common. The sandy limestone is commonly laminated on weathered surfaces, but crossbedding is rare.

Phosphatic chert has been noted at the very top of the member at one locality. Phosphatic rocks occur at the top of the member in the Stewart Flat quadrangle, which is just north of the Snowdrift Mountain quadrangle (T. M. Cheney, oral communication, 1957) and in some areas north of the Blackfoot River (V. E. McKelvey, oral communication, 1957); so the phosphatic horizon is probably more widespread in the area of this report than the single observation would suggest. A black shale 3 feet thick is exposed in the uppermost part of the member in a roadcut in Georgetown Canyon, but it has not been seen in natural exposures.

In the Aspen Range the lower member contains more limestone than it does in the eastern part of the area and is thicker bedded, so that the member there is more like the upper part of the Brazer in general aspect.

UPPER MEMBER

The upper member of the Wells formation (including the overlying Grandeur tongue of the Park City formation) is 1,000 or 1,100 feet thick in the Georgetown Canyon-Snowdrift Mountain area, although thicknesses of 1,300 to 1,500 feet have been noted in the Dry Valley quadrangle to the north (Cressman and Gulbrandsen, 1955, p. 261). The member is predominantly rather nonresistant sandstone and commonly underlies the steep wooded flanks of ridges that are held up by the much more resistant Brazer limestone. The contact between the upper and lower members of the Wells is marked in most places by an abrupt change from limestone outcrops to sandstone float, but where the contact is well exposed is seen to be gradational through a zone of interbedded sandstone and limestone.

The upper member of the Wells may be divided into three parts, each constituting about one-third of the member; these are helpful in determining position within the member, but they are not sufficiently distinct to be practicable as mapping units in areas of poor exposure or complex structure. Only the upper zone is apparent in the section measured at Deer Creek, but even there the units may be detected in the field by float and by small outcrops not included in the measured section.

The lowest third of the member consists of very fine grained and fine-grained quartz sandstone. It does not commonly crop out but generally forms talus in which the fragments are blocky and brown or reddish brown. The weathered fragments are light in weight, porous, and noncalcareous. Several beds of light-gray finely crystalline or dense limestone, similar in appearance to much of the limestone in the lower member of the Wells, occur in the middle of the unit on the west side of Snowdrift Mountain and on the east side of Harrington Peak, but they have not been observed in the easternmost exposures of the formation.

The middle third of the upper member is mostly light-gray or light-brownish-gray finely crystalline to dense dolomite and very fine grained dolomitic and calcareous sandstone. The dolomite beds are poorly exposed along Deer Creek, but on the west side of Georgetown Canyon about 1¼ miles south of Phosphoria Gulch they form prominent dip slopes and are well exposed in gullies. Some dolomite beds contain lenses and nodules of black chert and closely resemble the Grandeur tongue of the Park City formation. At the divide at the extreme south end of Dry Valley chert is present in beds so thick that on superficial inspection the float resembles that of the Rex chert member of the Phosphoria formation.

The uppermost third of the upper member consists mostly of nonresistant grayish-brown and reddish-brown calcareous sandstone. The sandstone is typically very fine grained to fine-grained and contains well-rounded and frosted grains of medium sand scattered throughout; it appears bimodal and rather poorly sorted compared with other sandstone in the Wells, but the one sample for which a size analysis was made is unimodal and has a Trask sorting coefficient of only 1.25. Much of the sandstone has been brecciated and rehealed and is commonly porous and vuggy in surface exposures. Many of the breccia fragments are laminated, and it is apparent in many exposures that the breccia is not sedimentary but has resulted from fracturing of the rock, slight rotation of the fragments, and recementation. An interval about 100 feet thick at the top of the member contains red sandstone and limestone. These red beds generally form a brush- or tree-covered swale, but the uppermost part is exposed in the Deer Creek section and parts are exposed in bulldozer roads of the Central Farmers Fertilizer Co. on the east side of upper Georgetown Canyon. Pink sandstone and anhydrite have been reported in the upper part of the Wells formation from the Sheep Creek test well on Bear Lake Plateau (Zeni, 1953); they are probably the equivalents of the upper unit of the member.

AGE AND CORRELATION

Girty (*in* Mansfield, 1927, p. 73) dated the Wells formation as Pennsylvanian on the basis of megafossils, mostly brachiopods. However, most of Girty's collections were from the lower member, and recent evidence indicates that most of the upper member is Permian. In 1953, J. Stewart Williams (p. 39), referring to unpublished work by oil company geologists, assigned the upper member of the Wells as used in this report to the Permian and the lower member to the Pennsylvanian. Fusulinids collected by K. P. McLaughlin from the upper member of the Wells in the Johnson Creek quadrangle were studied by L. G. Henbest who considered them to indicate an early Wolfcamp age (Henbest, written communication, 1954; McKelvey and others, 1959, p. 36).

More recent fusulinid collections from the Peale Mountains, collected mostly by V. E. McKelvey and studied by R. C. Douglass, of the U.S. Geological Survey, now permit the lower half of the Wells to be dated with considerable confidence. The results were summarized by McKelvey and others (1959, p. 36) and are presented in more detail below. Most of these collections were taken from a section measured in reconnaissance by McKelvey on the north side of the Blackfoot River about 20 miles north-northwest of the area described in this report. The measured section is followed by Douglass' fossil identifications and age assignments. Collections from elsewhere in the area that can be keyed to McKelvey's section are also included.

Section of the Wells formation in the northern part of the Wooley Range, Caribou County, Idaho

[Measured by V. E. McKelvey. Most of upper member described from exposures in NE¼ sec. 24, T. 6 S., R. 42 E., and NW¼ sec. 30, T. 6 S., R. 43 E.; lower 375 feet and all the lower member measured in SW¼ sec. 36, T. 6 S., R. 42 E]

	Thickness (feet)
Grandeur tongue.	
Park City formation:	
Dolomite, hard, light-gray. Contains a few cherty layers in upper and middle part and a 5-ft bed of yellowish-brown fine-grained sandstone about 8 ft above the base	85
Wells formation, upper member:	
P. Covered interval. Probably yellowish-brown, very fine grained sandstone	50
O. Sandstone, medium-hard, fine- to medium-grained, poorly sorted, massively bedded; forms rounded outcrops. Contains a few thin white or light-gray dolomite beds (including one a few feet above the base) and a minor amount of reddish-brown sandstone	420
N. Sandstone, calcareous, hard, fine-grained, poorly sorted, crossbedded, light-yellowish-gray (dark reddish brown or brownish gray weathered). Lower part forms prominent scarp	60
M. Dolomite and sandstone, incompletely exposed. Light-gray sandy dolomite in lower part; three 5- to 10-ft beds of yellowish-brown massive fine-grained calcareous sandstone crop out in upper part; intervening covered beds are probably soft sandstone	190
L. Sandstone, calcareous, hard, fine-grained, light-gray (dark greenish brown weathered). Forms low scarp	45
K. Limestone, hard, thin-bedded to massive; contains numerous masses of black chert. A 1-ft light-gray spicular chert layer occurs about 50 ft above the base. A few thin yellowish-brown sandstone layers are present in this same zone and in another zone 100–150 ft above the base	200
Covered interval. A dolomitic sandstone forms low scarp about 30 ft above the base. Remainder seems to be mostly soft sandstone	125
J. Sandstone, slightly calcareous, fine-grained, coarsely laminated, locally crossbedded, light-yellowish-gray (reddish brown weathered). Forms low scarp	10
I. Dolomite, hard, massive, vuggy, light-gray. Covered interval. Appears to be interbedded yellowish-brown sandstone and gray limestone	95
Covered interval. Apparently gray, fossiliferous limestone similar to underlying unit	20
H. Limestone, hard, gray; contains abundant bryozoans and much crinoid debris. Upper 5 ft contains 2 irregular 4-in. dark-gray (reddish brown weathered) chert layers, and forms low scarp	65

	Thickness (feet)
Wells formation, upper member—Continued	
G. Sandstone, finely laminated (much of it is crossbedded), reddish-brown; forms poor outcrops here and elsewhere but yields much coarse blocky float. Contains conspicuous 5- to 15-ft layers of gray limestone, calcareous sandstone, and dolomite 10, 70, and 240 ft above the base	275
F. Sandstone, yellowish-brown and light-gray; lower 10 ft contains much reddish-brown chert. Two 5-ft limestone beds—one 10, the other 45 ft above base. Silicified fusulinids (VEM 32–57) occur in sandstone about 50 ft above the base	100
Total thickness of upper member	1, 655
Wells formation, lower member:	
E. Interbedded gray bioclastic limestone, sandy limestone, and yellowish-brown sandstone (latter constitutes about 20 percent of unit). Much of limestone is laminated and contains large septarian chert nodules. Horn corals are abundant about 50 ft from top (VEM–2–58). Fusulinids occur about 20 ft below top (VEM–31–57)	425
D. Interbedded bioclastic limestone, sandy limestone, and sandstone (latter makes up about 10–15 percent of unit). Nodules of chert scattered throughout. Some of limestone is well laminated. Bench at base is formed by 10-ft bed of sandstone	160
C. Limestone, bioclastic, gray; about 5–10 percent of unit is made up of cherty beds and another 5–10 percent is composed of thin sandstones. Some of the limestones are finely laminated, and some contain flattened oolites	130
B. Interbedded bioclastic limestone, cherty limestone, and fine-grained calcareous sandstone (latter makes up about 20 percent of unit). Some of the limestone is oolitic	50
A. Sandstone, fine-grained, reddish- and yellowish-brown. Contains a 3-ft sandy limestone about 20 ft above base and another about 75 ft above base	130
Total thickness of lower member (rounded)	895
Total thickness of Wells formation	2, 550

Age, stratigraphic position, and location of fusulinid collections from the Wells formation in southeastern Idaho

[Identification and comments by R. C. Douglass. See measured section above for stratigraphic position of lithologic zones referred to by letter]

Lower Permian, late Wolfcamp age:

f 12480, VEM–37–57. Zone H? 300–400 ft above base of the upper member of the Wells formation, T. 8 S., R. 44 E. (north side of Mabie Canyon). Contains *Millerella* sp., *Pseudofusulinella* sp., and *Schwagerina* sp. aff. *S. elkoensis* Thompson and Hansen, 1954. The forms in this sample are characteristic of rocks of late Wolfcamp or Hueco age.

Middle Pennsylvanian, Des Moines age:

f 12475, VEM–32–57. Top of zone F. Sec. 36, T. 6 S., R. 42 E., (Wooley Valley divide). *Fusulina* sp. The specimens are altered chemically and abraded mechanically. They occur in a sandy matrix and appear to have been roughly size sorted. The specimens themselves are of Des Moines (Middle Pennsylvanian) age, but the age of the rock cannot definitely be determined.

f 12484, VEM–41–57. Zone F, 15–20 ft above the base, sec. 27, T. 8 S., R. 42 E. (north side of Trail Canyon). Contains ?*Staffella* sp. and *Wedekindellina* sp.

f 12483, VEM–40–57. Top of Zone E, sec. 27, T. 8 S., R. 42 E. (north side of Trail Canyon). Contains *Millerella* sp., ?*Staffella* sp., and *Fusulinella* sp.

f 12479, VEM–36–57. Zone E, 2–5 ft below top; T. 8 S., R. 44 E. (north side Mabie Canyon). Contains *Millerella* sp., ?*Staffella* sp., *Fusulinella* sp., and *Fusulina* sp.

f 12478, VEM 35–57, Zone E, 5–10 ft below top, sec. 33, T. 6 S., R. 43 E. (east slope of Wooley Range). Contains *Millerella* sp., ?*Staffella* sp., *Fusulinella* sp. and *Fusulina* sp. This association appears to represent a horizon near the Atoka-Des Moines boundary of the midcontinent region.

f 12474, VEM–31–57, Zone E, 20 ft below top, sec. 36, T. 6 S., R. 42 E. (Wooley Valley divide). *Fusulina* sp. aff. *F. leei* Skinner, 1931. This Middle Pennsylvanian form represents an early stage in the development of *Fusulina* and has been recognized in rocks of early Des Moines age from several areas.

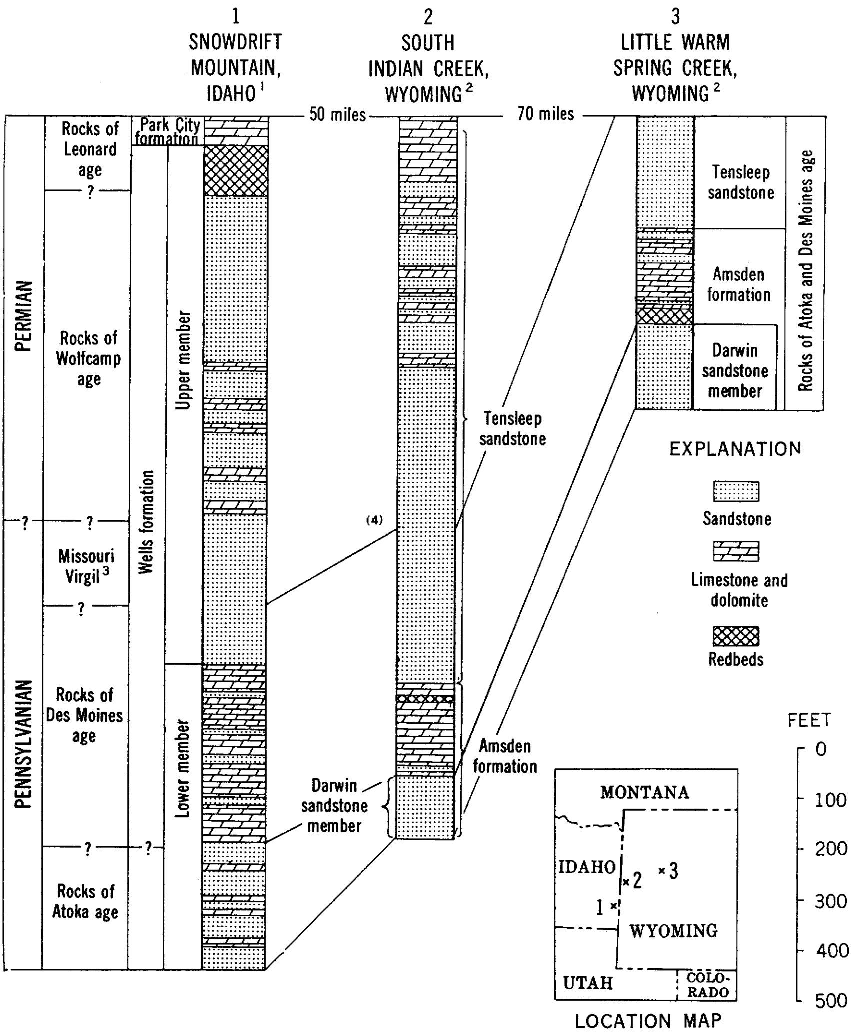
f 12470, ERC–28–15. Zone E, about 300 ft above base of Wells formation. NE¼ sec. 30, T. 10 S., R. 45 E. (Snowdrift Mountain). Contains *Endothyra* sp., *Millerella* sp., *Staffella* sp., *Fusulina* sp., *Wedekindellina* sp.

Middle Pennsylvanian, Atoka age:

f 12481, VEM–38–57. Zone B?, about 100 ft above base of Wells formation. Sec. 27, T. 8 S., R. 42 E. (north side of Trail Canyon). Contains *Millerella* sp., ?*Staffella* sp., and *Fusulinella* sp.

The collections indicate that the lower member of the Wells formation is of Middle Pennsylvanian age and includes equivalents of both the Atoka and Des Moines of the midcontinent section. The upper member of the Wells contains some beds of Middle Pennsylvanian age at the base, but at least the upper two-thirds are Permian. It is not known whether all that part of the Wells that is Permian is of Wolfcamp age or whether Leonard equivalents are present at the top. Beds of Late Pennsylvanian age have not been identified, but they might be present in the lower third of the upper member immediately overlying the highest beds known to be of Des Moines age.

The Pennsylvanian part of the Amsden formation of western Wyoming is of Atoka and possibly early Des Moines age (Henbest, 1956, p. 59) and is thus at least partly equivalent to the lower member of the Wells (fig. 3). If any equivalents of the Mississipian part of the Amsden are present in southeastern Idaho, they must be represented by the lowermost part of the Wells or the uppermost part of the Brazer. Comparison with sections of the Tensleep and Amsden



[1] Generalized composite section for Snowdrift Mountain-Georgetown area.
[2] Sections, formational and age assignments, and correlation within Wyoming by Love (1954).
[3] Rocks of Virgil or Missouri age are not known to be present in the Snowdrift Mountain area, but they might be present in this interval.
[4] Fusulinids of Missouri or Virgil age reported from this horizon of South Indian Creek (Love, 1954).

FIGURE 3.—Tentative correlation of the Wells formation of southeastern Idaho with the Tensleep and Amsden formations of western Wyoming.

formations of Wyoming in the correlation diagram of Love (1954) suggests that the lower sandy part of the lower member of the Wells of southeastern Idaho may be the equivalent of the Darwin sandstone member of the Amsden in the Jackson area of Wyoming (fig. 3).

The new data also suggest that most of the part of Love's South Indian Creek section that is shown by Love as of Missouri, Virgil, and possibly Wolfcamp age is probably of Wolfcamp age.

PERMIAN SYSTEM

GRANDEUR TONGUE OF THE PARK CITY FORMATION

The Grandeur tongue of the Park City formation was named by Cheney and others (*in* McKelvey and others, 1959, p. 12) for Grandeur Mountain in the central Wasatch Mountains of Utah. The beds assigned to the member in southeastern Idaho were previously included in the Wells formation (the upper siliceous limestone of Richards and Mansfield, 1912, p. 691).

The Grandeur tongue consists of about 75 feet of dense light-gray carbonate rock that is mostly dolomite but includes limestone near the base. The upper third or half of the tongue contains beds, lenses, and nodules of black and dark-gray chert that weathers light bluish gray. Some of the dolomite in the upper part contains abundant small silicified brachiopods. Locally, on the north side of Deer Creek on the east limb of Webster syncline, several beds of phosphorite about an inch or so thick occur 5 or 10 feet below the top of the tongue. A well-defined contact between the Grandeur tongue and the underlying Wells formation is exposed in only a few places, the Deer Creek section being one such locality. Because of the thinness of the tongue and the absence of a sharp well-exposed mappable basal contact, the Grandeur was not mapped separately but was grouped with the upper member of the Wells formation.

The Grandeur tongue is an excellent guide to the overlying Meade Peak phosphatic shale member of the Phosphoria. Caution must be used in its identification, however, for in the absence of the silicified brachiopods or of abundant phosphatic float, the Grandeur can be confused with cherty dolomite in the middle of the Wells formation.

The Grandeur tongue in southeastern Idaho is of Leonard age (Williams, *in* McKelvey and others, 1959, p. 36).

PHOSPHORIA FORMATION

Richards and Mansfield (1912, p. 684) named the Phosphoria formation from Phosphoria Gulch in the area of this report. At the type locality the formation consists of three members; these are, from youngest to oldest, the cherty shale, the Rex chert, and the Meade Peak phosphatic shale. The following section, adapted from McKelvey and others (1959, p. 20, 21), was measured on the west slope of Snowdrift Mountain 2 to 3 miles north of Phosphoria Gulch.

Section of the Phosphoria formation on Snowdrift Mountain

[Meade Peak phosphatic shale member measured in a bulldozer trench in the NW1xNW1x sec. 8, T. 10 S., R. 45 E., by T. M. Cheney, J A. Peterson, R. G. Waring, R. A. Smart, and E. R. Cressman (Smart and others, 1954, p. 15). Lower part of Rex chert member measured by V. E. McKelvey in artificial and natural exposures ad'acent to opencut mine of the Central Farmers Fertilizer Co. in the NW1xSW1x sec. 8 upper part of Rex and the cherty shale member measured by McKelvey in natural and artificial exposures in the NE1xNE1x sec. 18]

Thickness (feet)	Cumulative thickness	
		Dinwoody formation (lower beds only):
		Limestone, argillaceous, hard (forms natural outcrop), brownish-gray to pale-brown, thin-bedded; lower 0.7 ft consists of grayish-brown mudstone containing scattered black phosphatic pellets
		Phosphoria formation:
		Cherty shale member (includes a tongue of the Retort [phosphatic shale member of Phosphoria fm.] indicated by "Rt" in front of the description):
1. 0	1. 0	Rt: Phosphorite, cherty(?), nodular, grayish-brown, hard; contains casts of sponge spicules(?).
1. 3	2. 3	Rt: Mudstone, soft, brownish-black, fissile.
1. 3	3. 6	Rt: Phosphorite, cherty, hard, black, pelletal.
3. 0	6. 6	Rt: Mudstone, soft, brownish-gray, fissile.
1. 7	8. 3	Rt: Dolomite, hard (forms natural outcrop), brownish-gray, massive. Weathered surface is pale brown and deeply etched.
22. 0	30. 3	Rt: Mudstone, soft, black to grayish-brown, fissile.
3. 5	33. 8	Chert, argillaceous, hard (forms natural outcrop), black, thick-bedded.
13. 0	46. 8	Cherty mudstone and mudstone; hard black and brownish-gray thin-bedded cherty mudstone (80 percent) and soft brownish-gray mudstone (20 percent).
6. 3	53. 1	Mudstone, slightly dolomitic, medium-hard, dark-gray, fissile.
10. 0	63. 1	Mudstone, cherty, medium-hard, brownish-gray, thin-bedded.
1. 0	64. 1	Dolomite, hard, dark-gray, massive; contains chert nodules in upper 0.2 ft. Weathered surface is pale brown and deeply etched.
24. 2	88. 3	Mudstone, cherty, locally dolomitic, medium-hard, brownish-gray, thin-bedded. Some fracture surfaces are stained reddish brown and moderate orange.
4. 0	92. 3	Dolomite, argillaceous, hard, brownish-gray; contains deeply weathered parts that are soft and pale reddish brown.
4. 4	96. 7	Mudstone, locally dolomitic or cherty, medium-hard; brownish gray, fresh and pale brown to moderate yellowish orange weathered; thin-bedded.
96. 7		Total thickness of cherty mudstone member.

Phosphoria formation—Continued

Rex chert member (includes a tongue of the cherty shale member, designated by "CS" in front of the description):

Thickness (feet)	*Cumulative thickness*	
21. 6	118. 3	Chert and mudstone: hard (forms natural outcrop) dark-gray nodular thin-bedded chert (75 percent) and interbedded thin zones of cherty soft to medium-hard brownish-gray (stained pale brown to moderate yellowish orange on fractures) fissile and thin-bedded mudstone (25 percent).
11. 3	129. 6	Chert, hard (forms conspicuous natural outcrop), black and dark-gray (weathers reddish gray), thick-bedded; contains irregular chert pebbles and nodules.
21. 2	150. 8	Chert, hard (forms conspicuous natural outcrop), dark-gray (weathered surface has conspicuous light-gray bands), thick-bedded.
10. 2	161. 0	Chert, hard (forms natural outcrop); dark-gray (reddish gray and moderate orange weathered), thin- and thick-bedded; contains abundant irregular cylindrical concretions as much as 0.3 ft. in diameter and 1.0 ft long, mostly inclined in relation to bedding planes; unit not observed in nearby sections.
15. 9	176. 9	Cs: Chert and cherty mudstone: cherty medium-hard (poorly exposed) brownish-gray (moderate yellowish-brown and reddish-brown coatings on weathered surfaces and joints), thin-bedded mudstone (35 percent) interbedded with hard thin-bedded nodular chert (65 percent).
9. 2	186. 1	Chert, hard (forms upper part of prominent cliff) dark-gray (light gray and reddish gray weathered), thick- and massive-bedded, nodular.
26. 5	212. 6	Chert, hard (forms prominent cliff), dark-gray (light gray weathered), thick- and massive-bedded; beds pinch and swell.
23. 7	236. 3	Chert, hard, black (moderate orange to reddish gray on weathered surfaces and joints), thick-bedded; contains abundant spicules.
16. 0	252. 3	Chert and limestone: hard black thin- and thick-bedded finely laminated chert (70 percent) interbedded with and irregularly replacing hard black (pale brown and yellowish brown weathered) thin- and thick-bedded finely laminated limestone (30 percent). In nearby sections this unit is as much as 45 ft thick.
155. 6		Total thickness of Rex chert member.

Phosphoria formation—Continued

Thickness (feet)	Cumulative thickness	
		Meade Peak phosphatic shale member:
25. 3	277. 6	Mudstone, medium-hard, grayish-brown, commonly spheroidally weathered; nodular phosphorite, about 0.5 ft thick at top of unit; another bed approximately 1 ft thick about 3 ft above base.
15. 5	293. 1	Upper phosphate zone: coarsely oolitic medium-hard brownish-gray thin- and thick-bedded phosphorite and interbedded thin layers of soft yellowish-brown slightly phosphatic mudstone; pisolitic phosphorite at top, hackly fractured oolitic phosphorite at base.
9. 4	302. 5	Phosphorite, argillaceous, soft, grayish-brown; increasingly vanadiferous toward top. In fresh sections, rocks of this zone are only moderately phosphatic and contain abundant carbonate lenses as much as 1 ft thick.
16. 4	318. 9	Upper nodular zone: medium-hard pale-brown (grayish black with greenish-yellow stains in fresh sections) thick-bedded mudstone and interbedded thin layers of nodular grayish-brown phosphorite; hard massive limestone 2 ft thick ordinarily present at top but here weathered to mudstone.
7. 3	326. 2	Mudstone, soft brownish-gray, thin-bedded; contains some phosphatic layers.
22. 5	348. 7	Lower nodular zone: soft to medium-hard brownish-gray (hard and grayish black in fresh sections) thick-bedded mudstone and interbedded thin layers of nodular grayish brown phosphorite; greenish yellow stains abundant.
28. 4	377. 1	Mudstone, soft, brownish-black, thin-bedded, phosphatic in upper half; chert bed about 0.5 ft thick 10 ft above base. Fresh sections contains abundant lenses of carbonate rock.
26. 6	403. 7	Mudstone, phosphatic, soft, grayish-brown, thin-bedded.
7. 5	411. 2	Mudstone, soft, grayish-brown; phosphatic in upper two-thirds. In fresh sections upper and lower 2 ft are carbonate beds; the lower one is known as the "false cap."
22. 0	433. 2	Phosphorite, argillaceous, brownish-gray, thin-bedded. Rock from this zone is used in electric furnace manufacture of phosphorus.
1. 4	434. 6	Mudstone, phosphatic, soft, grayish-brown. Where unweathered, this bed is phosphatic argillaceous limestone, known as the "cap lime."
7. 0	441. 6	Lower phosphate zone: medium-hard thin-bedded brownish-black phosphorite. Rock from this zone is used to make superphosphate and triple-superphosphate fertilizer.

Thickness (feet)	Cumulative thickness	
		Phosphoria formation—Continued
		Meade Park phosphatic shale member—Continued
10.6	452.2	Carbonate rock, argillaceous, medium-hard, dark-gray, thick-bedded; bed of bioclastic phosphorite 0.7 ft thick at base contains abundant fish scales.
199.9		Total thickness of Meade Peak member.
		Park City formation (upper bed only of Grandeur tongue):
		Dolomite, siliceous, hard, light brownish-gray thick-bedded.

MEADE PEAKE PHOSPHATIC SHALE MEMBER

McKelvey (McKelvey and others, 1956, p. 2845) has renamed the phosphatic shale member of Richards and Mansfield (1912, p. 683) the Meade Peak phosphatic shale member, and describes it as follows (McKelvey and others, 1959, p. 22):

the Meade Peak phosphatic shale member in its type area is composed mainly of dark carbonaceous, phosphatic, and argillaceous rocks. Mudstone and phosphorite are the chief end-member rock types; dark dolomite and limestone are subordinate types. Common mixed-rock types and distinctive textural varieties include both slightly and highly carbonaceous mudstone (the latter pyritic in fresh exposures); finely pelletal phosphatic mudstone; dense, structureless phosphorite, fine-grained pelletal phosphorite, oolitic phosphorite, pisolitic phosphorite, nodular phosphorite, and bioclastic phosphorite generally composed of fish scales or brachiopod shells; and finely crystalline argillaceous and phosphatic (pelletal) dolomite or limestone. The vertical sequence of these rocks in the type area is symmetrical—upper and lower halves are almost mirror images of each other (McKelvey, 1949, fig. 10). Thus, the sequence from the base upward consists of bioclastic phosphorite, mudstone, high-grade phosphorite becoming progressively more argillaceous upward, and slightly phosphatic mudstone; then the reverse sequence to the top of the member.

The Meade Peak phosphatic shale member is seldom exposed, but its position usually is marked by a swale between resistant rocks of the Grandeur tongue of the Park City and the Rex chert member of the Phosphoria. Near-surface sections generally show considerable effects of weathering: the rocks are soft and brown, or lighter, in color; carbonate rock is leached to mudstone; and nearly all the rocks are enriched in phosphate. Rocks in unweathered sections are uniformly dark gray or black, hard, and more calcareous and less phosphatic. The Meade Peak in southeastern Idaho ranges from 125 to 225 feet in thickness. Much of this variation may be the effect of deformation and weathering.

The thickness of the Meade Peak member in the Georgetown Canyon-Snowdrift Mountain area ranges from 200 feet in the Georgetown syncline to about 150 feet on the east limb of Webster syncline, judging from both artificial exposures and the width of outcrop (or more properly, the width of the swale between exposures of the Grandeur tongue of the Park City formation and the Rex chert member of the Phosphoria formation). Outcrop widths indicating thicknesses of much less than 150 feet, such as those on the west limb of the Georgetown syncline, undoubtedly result from faulting.

The base of the Meade Peak member in this area is marked by a thin phosphorite bed that contains fish scales and rests directly on light-colored dolomite of the Grandeur tongue; float from this bed often is found immediately overlying the Grandeur. Where the Grandeur tongue forms dip slopes, the basal contact of the Meade Peak has been placed at the topographically highest occurrence of phosphorite. Although this procedure probably closely locates the contact in most places, it is possible that in some localities the highest phosphorite float may be from thin phosphorite seams within the upper part of the Grandeur.

REX CHERT AND CHERTY SHALE MEMBERS

The Rex chert member of the Phosphoria formation was named by Gale, but the name was first published by Richards and Mansfield (1912, p. 684). Although the name is derived from Rex Peak in the Crawford Mountains of Rich County, Utah, McKelvey (*in* McKelvey and others, 1956, p. 2847) has concluded that Richards and Mansfield intended the type area to be Phosphoria Gulch. As originally defined, the Rex chert member consisted of about 150 feet of massively bedded chert, containing some limestone at the base and some cherty mudstone in the middle, overlain by about 100 feet of cherty mudstone. McKelvey (*in* McKelvey and others, 1956, p. 2847–2848) redefined the Rex to exclude the 100 feet of cherty mudstone at the top which was named the cherty shale member. In this report the cherty shale member has been mapped separately only on the gently dipping east limb of the Webster syncline where the outcrop belt is wide; elsewhere, it has been grouped with the Rex member as a single map unit.

The combined thickness of the Rex chert and cherty shale members is about 240 to 300 feet. Thicknesses obtained from the map are not everywhere reliable because the upper contact was generally located by float or topography.

McKelvey (*in* McKelvey and others, 1959, p. 26) has described the Rex chert member as follows:

> In the type area and at most other localities where the Rex is exposed it includes a wide variety of cherty rocks that range from almost pure vitreous translucent chert to impure dull opaque chert to mudstone, phosphorite and carbonate rock that contain discrete masses of chert. Major impurities in the chert are argillaceous and carbonaceous matter, quartz sand, calcite, dolomite, or apatite; minor impurities include pyrite or glauconite. The rocks of the Rex range in color value (lightness) from black to nearly white; intermediate shades are mostly low in chroma (color saturation). The bedding of the cherts ranges from even to wavy or lenticular and from an inch to as much as 2 feet in thickness. Some beds are structureless (massive) but many are composed of lenses, elliptical

nodules, cylindrical masses, or highly irregular masses. The cylindrical masses are either parallel to each other and steeply inclined with respect to the bedding planes, or nonparallel and flat lying; the latter type gives bedding surfaces a fucoidal appearance. Some of the chert beds contain sparse to abundant sponge spicules (Keller, 1941, p. 1292; Cressman, 1955, p. 11) which are composed of chalcedony or microcrystalline quartz and which range in preservation from poor to good; they can be seen with a hand lens or even with the naked eye. Many other beds have spicules that are visible under high magnification or dark-field illumination; still other beds contain no recognizable organic remains. Stylolites, some with a vertical relief of 4 inches, are present in many beds, particularly in relatively pure spicular cherts.

Three sections of the Rex and upper shale in the Phosphoria Gulch area have been measured recently; the units * * * [described in the preceding measured section] are recognized in all of them. The cherty mudstone zone, 10 to 15 feet thick, in the middle of the Rex is probably an eastern tongue of the cherty shale member; it is not found in adjacent areas, where the Rex consists of an uninterrupted sequence of chert. The basal zone containing lenses and layers of carbonate is recognizable at most localities in southeastern Idaho, though it ranges in thickness from 5 to 65 feet.

Extensive lenses of coarse light-gray bioclastic limestone occur near the middle of the Rex at several localities in southeastern Idaho, notably Wood Canyon, North Stewart Canyon, Timber Creek, Deer Creek, Sage Creek, South Canyon, and the ridge north of Hot Springs. These lenses are as much as a mile in outcrop length and range from 0 to as much as 50 feet in thickness; northwest of Timber Creek they seem to constitute nearly the whole of the Rex interval. Their lithology is highly distinctive and closely resembles that of the Franson member of the Park City in western Wyoming; no observations suggest they were ever physically continuous with the Franson, but they are outliers of the same facies.

The lens of light-gray coarsely crystalline bioclastic limestone at Deer Creek mentioned by McKelvey is in the trough of the Webster syncline where it occurs near the top of the Rex and extends southward for about half a mile from the creek. The width of the lens is not known, but it exceeds a quarter of a mile. The lens terminates abruptly; it is present on the south side of Deer Creek but not on the north. The limestone in South Canyon occurs throughout the Rex. The lenses do not thicken the Rex interval, but they seem to take the place of an equivalent thickness of chert.

The contact between the Rex chert member and the Meade Peak phosphatic shale member has not been seen in natural exposures, the contact generally having been placed at the base of the lowest rib of massive chert. The zone of interbedded chert and limestone at the base of the Rex is locally a useful guide to the contact, particularly in areas of moderate or low dip.

McKelvey (*in* McKelvey and others, 1959, p. 28) describes the cherty shale member as follows:

The lithology of the cherty shale is not as well known as that of the other units because of poor exposures in the area of its major development. The mudstone, which makes up a large part of the unit, is harder, more siliceous, and less carbonaceous than that of the Meade Peak. The chert is mostly thin-bedded and argillaceous. The cherty mudstone is thin-bedded and hard; its siliceous nature is due to disseminated chert and to quartz silt, which is much more prominent than microcrystalline quartz and chalcedony in some thin sections. Joint surfaces of the cherty mudstone are often etched in a manner resembling stylolites of low relief; no bedding-plane stylolites, however, have been observed.

The cherty shale member does not generally crop out but typically forms bare slopes littered with hard black platy mudstone float.

AGE

The Phosphoria formation is probably of Leonard and Guadalupe age. The fauna, age, and correlation are discussed in detail by Williams (*in* McKelvey and others, 1959, p. 38).

TRIASSIC SYSTEM

Strata in the area that are assigned to the Triassic system total about 6,000 feet. About 5,000 feet, comprising the Dinwoody and Thaynes formations, is well dated as Early Triassic; the remainder is unfossiliferous and is dated as Triassic by its position between beds of known Early Triassic and Middle Jurassic ages and by lithologic correlation with rocks known to be Triassic in other areas. The most recent study of the Triassic rocks of southeastern Idaho is that of Kummel (1954). His formational nomenclature is generally followed in this report but with some modification, particularly of formational contacts, to meet local mapping conditions.

DINWOODY FORMATION

The Dinwoody formation was named by Blackwelder (1918 and *in* Condit, 1916, p. 263) for Dinwoody Canyon on the northeast side of the Wind River Range of Wyoming. As originally defined, the formation at the type area consists of a brown-weathering sequence of greenish-gray shale with thin plates of argillaceous dolomite or limestone containing obscure pelecypod shells. It is underlain at the type area by the Ervay member of the Park City formation (McKelvey and others, 1956, fig. 4) and overlain by red shale and siltstone of the Chugwater formation.

The 1,600 to 1,800 feet of Triassic beds beneath the *Meekoceras* zone in the Georgetown Canyon-Snowdrift Mountain area was assigned by Mansfield (1927, p. 86) to the Woodside shale. Kummel (1953, 1954), however, has applied the name Dinwoody to the formation because of its lithologic similarity to the type Dinwoody and its dissimilarity to the dark red shale of the type Woodside. Certainly the lower part of the Dinwoody, as so defined in southeastern Idaho, closely resembles the Dinwoody of the Wind River Range. The upper part in southeastern Idaho contains rock types very similar to those in the type section of the formation, but the gross aspect is considerably different because of differences in their relative abundance and their bedding characteristics.

The Dinwoody formation was mapped as two members of nearly equal thickness.

LOWER MEMBER

The lower member of the Dinwoody formation consists mostly of thin-bedded to fissile light-grayish-brown to olive-brown shale and calcareous siltstone that grades upward into thick-bedded calcareous siltstone and silty limestone. The member averages about 700 feet thick in the Georgetown syncline, but it may be 800 or 900 feet thick in the Webster syncline. It is incompetent and poorly exposed, and thicknesses may range from 500 to 900 feet within a mile or so along strike without any other surface evidence of faulting. Thicknesses as much as 1,000 feet have been described in the Dry Valley quadrangle (Cressman and Gulbrandsen, 1955, p. 262) and as much as 1,400 feet in the Johnson Creek quadrangle (Gulbrandsen and others, 1956, p. 12).

The following section was measured by J. A. Peterson on the east limb of Georgetown syncline where it is crossed by upper Deer Creek.

Section of lower member of the Dinwoody formation

[Measured in a roadcut and gully on west side of road and on west side of the north fork of Deer Creek, N½ sec. 32, T. 9 S., R. 45 E., Caribou County, Idaho]

Unit	Thickness (feet)	Cumulative thickness	Description
			Dinwoody formation, upper member (basal part only)
30	25	--------	Limestone: medium- to thick-bedded; gray (weathers light gray).
29	150	--------	Covered: reddish-brown shale float in lower 50 ft.
			Dinwoody formation, lower member
28	75	75	Limestone: silty, medium-bedded, black-weathering; contains some interbedded brown and reddish-brown shale; upper 10 to 15 ft is gray limestone.
27	15	90	Shale.
26	2	92	Limestone: massive, pink-weathering.
25	39	131	Limestone: silty, brown- and black-weathering; contains a few shale beds.
24	62	193	Shale and limestone interbedded.
23	15	208	Limestone: silty, brown- and black-weathering.
22	70	278	Shale: contains a few beds of platy to flaggy limestone.
21	17	295	Limestone: silty, thin-bedded, black-weathering.
20	28	323	Shale (70 percent) and limestone (30 percent) interbedded.
19	45	368	Limestone and shale interbedded; contains bed of black-weathering silty limestone 4 ft thick in middle.

Dinwoody formation, lower member—Continued

Unit	*Thickness (feet)*	*Cumulative thickness*	*Description*
18	30	398	Shale: contains interbedded limestone in upper third.
17	15	413	Limestone: contains a few beds of shale.
16	42	455	Shale.
15	15	470	Limestone (60 percent) and shale (40 percent) interbedded.
14	15	485	Shale.
13	7	492	Limestone (80 percent) and shale (20 percent) interbedded.
12	26	518	Shale.
11	15	533	Limestone: flaggy.
10	32	565	Shale (80 percent) and limestone (20 percent) interbedded.
9	15	580	Covered: shale float.
8	10	590	Shale and limestone interbedded: limestone is flaggy.
7	20	610	Mostly covered: some shale and flaggy siltstone outcrops.
6	7	617	Shale.
5	15	632	Limestone (60 percent) and shale (40 percent) interbedded: limestone is flaggy.
4	13	645	Shale (80 percent) and limestone (20 percent) interbedded.
3	13	658	Limestone (60 percent) and shale (40 percent) interbedded: limestone is flaggy.
2	10	668	Shale: contains a few thin beds of limestone.
1	40	708	Mostly covered: a few limestone and shale outcrops; basal bed is tan silt stone; underlain by cherty shale member of Phosphoria formation.

Although there is some clay shale in the base of the member, most of the material described as shale in the measured section is thin-bedded olive-gray, olive-brown, and dusky-yellow calcareous siltstone. The limestone interbedded with the shale is gray and finely crystalline and is generally in beds 1 to 6 inches thick. Some limestone beds contain many vague impressions of pelecypod valves, and much of the limestone weathers a distinctive grayish brown.

The medium- and thick-bedded silty limestone in the upper 130 to 200 feet of the member is thicker bedded and more resistant than the underlying shale, and its position along nearly the entire length of the east limb of the Georgetown syncline is marked by low ridges and alined knobs. The silty limestone contains about equal amounts of silt and calcite and could as validly be termed "calcareous siltstone." It is typically dark brown and shiny black on weathered surfaces, and it breaks into blocky fragments. Truncated rolls are common in the upper part. Throughout the area the black-weathering silty limestone zone is capped by a gray limestone bed, about 15 feet thick, that in many places forms conspicuous dip slopes.

A swale is commonly developed on the uppermost part of the cherty shale member of the Phosphoria formation and on the basal shale of the Dinwoody, and the contact between the two formations is generally covered. Where the contact is exposed, such as at Black Dugway, it is sharp and conformable and is marked by a thin nodular phosphorite bed at the very top of the Phosphoria.

UPPER MEMBER

The upper member of the Dinwoody formation is about 900 feet thick in the mapped area, judging by outcrop widths. It contains the same rock types as the lower member—gray limestone, grayish-brown and olive-brown shale, and brown- and black-weathering calcareous siltstone—but the shale is much less abundant and the gray limestone is thicker bedded and much more conspicuous. The limestone beds crop out in ledges and ribs, and the member commonly underlies rounded knobs and hills of intermediate height. No section of the upper member has been measured in the Georgetown Canyon-Snowdrift Mountain area, but it is very similar in lithology, sequence, and thickness to the member in Kummel's Dry Ridge section (1954, pl. 36) which was measured on the west limb of the Georgetown syncline about 5 miles north of the Snowdrift Mountain quadrangle.

The dip slope developed on the uppermost bed of the lower member of the Dinwoody generally terminates in a swale about 100 feet wide. Although the swale is commonly covered by a shick tangle of brush, red siltstone float can often be found, and the covered interval is with little doubt composed of a tongue of the Woodside shale. The poor exposures would not permit the red beds to be mapped separately; they have therefore been included in the upper member of the Dinwoody.

The beds above the swale, by far the greater part of the member, may be divided into three rather indistinct parts. In the lower part the gray limestone beds are thick and are the dominant rock type; in the middle part the limestone beds are thin, and calcareous siltstone, black-weathering in part, is the dominant rock type; in the upper part limestone is again dominant. Much of the limestone of the upper member is bioclastic, and crossbedding is common. Some beds contain abundant poorly preserved pelecypod shells.

The uppermost limestone bed of the member as defined in this report contains the *Meekoceras* zone about 15 or 20 feet below the top. The zone, characterized by many well-preserved ammonites, is well developed throughout the area and is very useful as a marker bed. Both Mansfield (1927, p. 86) and Kummel (1954, p. 167) placed the upper contact of the Dinwoody beneath the *Meekoceras*-bearing limestone, but the most marked lithologic change is between the limestone and the overlying black shale, and the formational contact has been placed accordingly.

THAYNES FORMATION

The Thaynes formation was named by Boutwell (1907) for Thaynes Canyon in the Park City district of Utah. The formation there consists of limestone, sandstone, and shale and contains a red shale member in the middle; it is underlain by red beds of the Woodside formation and overlain by red beds of the Ankareh formation. In southeastern Idaho those beds from the base of *Meekoceras*-bearing limestone to the base of the Timothy sandstone were termed Thaynes group by Mansfield (1916; 1927, p. 87) and were divided into three formations—the Ross Fork limestone, the Fort Hall formation, and the Portneuf limestone. In the most recent study, Kummel (1954, p. 172) reduced Thaynes to formational rank, classed a red bed unit within the Portneuf limestone as the Lanes tongue of the Ankareh formation, included the Timothy sandstone in the formation as the uppermost member, but he did not use the names Fort Hall and Ross Fork. In the central Peale Mountains, Kummel (1954, p. 172) divided the Thaynes into seven lithologic units; these are, from oldest to youngest, the lower limestone (the *Meekoceras*-bearing limestone), the lower black limestone, the upper black limestone, a sandstone and limestone unit, the Portneuf limestone member, and the Timothy sandstone member. In this report the lower limestone of Kummel has been placed in the Dinwoody formation and the base of the Thaynes has been mapped at the base of Kummel's "lower black limestone." The formation as thus defined totals 2,500 to 3,200 feet thick, excluding the Lanes tongue of the Ankareh formation. It has been divided into six mapping units of which the lowermost corresponds to Kummel's "lower black limestone" and the upper three to Kummel's Timothy member and upper and lower parts of the Portneuf member; the other two do not correspond with Kummel's divisions.

Mansfield (1927, p. 86) noted the lithologic similarity between the Dinwoody and Thaynes formations and the somewhat arbitrary nature of the contact between them. Indeed, the similarity in both rock types and sequences is so striking that I doubt whether the Lower Triassic of the area would have been divided into two formations if it were not for the conspicuous widespread *Meekoceras* zone; it would either have been treated as one formation or have been divided into more than two.

No outcrops of the Thaynes formation sufficiently well exposed to measure were found in the area, but the Sheep Creek section measured by Kummel (1954, pl. 38) north of the Blackfoot River differs only in detail from the Thaynes of the Georgetown Canyon-Dry Ridge area.

LOWER BLACK SHALE MEMBER

The lower black shale member, the same unit as Kummel's lower black limestone, is 700 to 800 feet thick. In the southern part of Georgetown syncline it is mostly black and dark-gray shale, but it also contains two limestone units, each probably not more than 100 feet thick. The lower limestone is brittle, nearly lithographic, and black on fresh surfaces but bluish gray on weathered surfaces; it laminated, and breaks into small angular blocks. The upper limestone is medium gray to brownish gray and silty; it weathers to less regular and less angular fragments than the lower limestone. The black shale is rarely exposed, but the limestones, particularly the lower one, form distinctive outcrops.

In the vicinity of upper Crow Creek and Preuss Creek the black shale and the lower black limestone are present, but the upper limestone is replaced by 150 or 200 feet of thick-bedded gray limestone interbedded with some tan calcareous siltstone. The limestone contains many pelecypod and brachiopod shells, those of the brachiopod genus *Pugnoides* being particularly abundant. This resistant limestone unit is underlain and overlain by black shale and is a conspicuous easily recognizable marker, both in the field and on aerial photographs. It has been mapped as a separate unit but is included in the lower black shale member for nomenclatural purposes, although it is probably a tongue of the middle limestone member described by Kummel (1954, p. 175) in the area south and east of Montpelier.

PLATY SILTSTONE MEMBER

Overlying the lower black shale member are about 600 feet of tan-weathering brownish-gray silty limestone that is typically thin bedded and evenly bedded and weathers into large flat plates. The upper and lower parts of the member are generally somewhat thicker bedded than the central part and form chunky or blocky rather than platy fragments.

IRREGULARLY BEDDED SILTSTONE MEMBER

This member consists mostly of about 600 feet of poorly exposed thin-bedded and irregularly bedded brownish-gray calcareous siltstone that contains many small nodules of dense gray limestone. The thin-bedded siltstone grades upward into medium-bedded calcareous siltstone and very fine grained calcareous sandstone that constitutes the uppermost 50 to 100 feet of the member. Both the rock types and the general sequence are similar to those of the lower member of the Dinwoody formation; the principal distinguishing features are the irregular bedding of the shale, the nodular character of the limestone, and the coarser grain size of the uppermost beds. Some gray and black shale occurs within the member in the Dry Valley quadrangle

(Cressman and Gulbrandsen, 1955, p. 263), but none is exposed in this area.

PORTNEUF LIMESTONE MEMBER

The Portneuf limestone member is divided into two parts by the Lanes tongue of the Ankareh formation, a red-bed unit. Between upper Crow Creek and Preuss Creek the upper part of the Portneuf member is about 180 feet thick, but the thickness of the lower part and of the Lanes tongue of the Ankareh cannot be determined accurately because of faulting. Best estimates are 600 or 700 feet for the lower part of the Portneuf and 400 or 500 feet for the Lanes tongue. On the west side of the Left Fork of Twin Creek, limestone that is probably the upper part of the Portneuf is present but is too thin to map.

The Portneuf limestone member consists of interbedded limestone, siltstone, and sandstone, of which the limestone is the most abundant and conspicuous. The member is similar in appearance to the upper member of the Dinwoody formation, but the limestone beds of the Portneuf are thicker and more massive. The limestone is gray and generally very fossiliferous. Chert nodules and silicified shells occur in the upper part of the Portneuf and in the upper half of the lower part.

TIMOTHY SANDSTONE MEMBER

Mansfield (1920, p. 50) named the Timothy sandstone for Timothy Creek in the northern Peale Mountains. Kummel (1954, p. 172) reduced the status of the Timothy to that of a member of the Thaynes formation because it is "lithologically similar to the Thaynes formation and appears to be gradational to the Thaynes * * *." The lithologic similarity is not apparent to me, but Kummel's nomenclature is followed nevertheless.

The Timothy sandstone member as exposed in this area consists of massive beds of olive-gray fine- to coarse-grained sandstone that appears well sorted and contains many black grains that are probably chert. The sanstone is soft and friable and weathers into softly rounded outcrops. The exact thickness of the member is not known because the contacts are not well exposed, but it is between 250 and 400 feet.

ANKAREH FORMATION

The Ankareh formation was named by Boutwell (1907) for Ankareh Ridge in the Park City district of Utah. As originally defined, the Ankareh consisted mostly of red shale but included beds now assigned to the Nugget sandstone. Boutwell (1912, p. 59) subsequently redefined the formation to exclude the Nugget. Gale and Richards (1910) introduced the name Ankareh into southeastern Idaho, apparently for the beds now termed the Lanes tongue of the Ankareh, but placed all the beds between the base of the Timothy sandstone and the base of the Twin Creek limestone in the Nugget. Mansfield (1927, p. 84) abandoned the name Ankareh in southeastern Idaho, included the Lanes tongue of current nomenclature in the Thaynes formation, restricted Nugget to the rather uniform sandstone unit underlying the Twin Creek limestone, and divided the strata between the Thaynes and the Nugget into four new formations—the Timothy sandstone at the base, the Higham grit, the Deadman limestone, and the Wood shale at the top. In addition to placing the Timothy in the Thaynes, Kummel (1954) named the red beds in the Thaynes the Lanes tongue of the Ankareh formation, retained the Higham grit and the Deadman limestone as formations, and considered the Wood shale a tongue of the Ankareh. Kummel's nomenclature is followed in this report in that both the Wood shale and the red beds within the Portneuf are considered tongues of the Ankareh. However, the Deadman limestone has not been mapped separately; it is included in the Wood shale tongue because in most exposures it is too thin, because its contacts are too poorly defined to make it a practical mapping unit, and because it is both overlain and underlain by red shale.

The Higham grit, which overlies the Timothy member of the Thaynes throughout most of southeastern Idaho, is not present in either of the two exposures of the Upper Triassic rocks in the Georgetown Canyon-Snowdrift Mountain area. Its absence might have resulted from faulting, but there is no evidence for such faulting other than for the absence of the Higham, and it seems more probable that either the formation was not deposited or else was subsequently eroded. Rubey (1958) states that grit-sized conglomerate equivalent to the Higham occurs as lentils rather than as a continuous bed in the Bedford quadrangle of westernmost Wyoming.

The tongues of the Ankareh formation are described below.

LANES TONGUE

As noted in the section on the Portneuf member of the Thaynes, the Lanes tongue of the Ankareh formation is probably about 400 or 500 feet thick and is separated from the Wood shale tongue by nearly 600 feet of strata assigned to the Thaynes. The tongue consists of dark-red or maroon shale and very fine grained calcareous siltstone. It is poorly exposed, the best exposure being on Preuss Creek.

WOOD SHALE TONGUE

The following section of the Wood shale tongue of the Ankareh formation and the underlying beds of the uppermost Thaynes formation were measured on the north side of Preuss Creek by C. H. Marshall.

Section of the Wood shale tongue of the Ankareh formation and the upper part of the Thaynes formation

[Measured on a ridge on the north side of Preuss Creek, sec. 15, T. 11 S., R. 45 E., Caribou County, Idaho]

Unit	*Thickness (feet)*	*Cumulative thickness*	*Description*
			Wood shale tongue of Ankareh formation
21	42	42	Covered: red shale float; overlain by Nugget sandstone.
20	3	45	Shale: irregularly bedded, moderate-red to moderate-reddish-brown, nonresistant.
19	5	50	Covered: red shale float.
18	2	52	Limestone: hard, thin- to medium-bedded, light-gray to white, dense.
17	21	73	Covered: red shale float.
16	5	78	Sandstone: medium-grained.
15	32	110	Covered: red sandy shale float.
14	1	111	Dolomite: silty, hard, thin-bedded, pale-red.
13	27	138	Covered: red shale float.
12	14	152	Breccia: intimately mixed grayish-pink limestone and grayish-red shale; rock is mottled; 0.1 ft of dolomite at base.
11	42	194	Covered: red shaly sandstone float.
	21	215	Covered: gray limestone float.
10	16	231	Limestone: hard, medium light-gray to grayish-red.
9	100	331	Covered slope to gully bottom.
			Thaynes formation, Timothy sandstone member [387 ft, total thickness of Timothy member]
8	101	432	Covered slope up to first sandstone exposure.
7	11	443	Sandstone: soft, massive, yellowish-green, medium- to coarse-grained.
6	164	607	Covered: sandstone float.
5	111	718	Sandstone: soft, massive, olive-gray, fine- to medium-grained; contains many black grains; crossbedded in part.
			Thaynes formation, upper part of Portneuf limestone member
4	122	840	Covered: limestone float.
3	32	872	Limestone: thick-bedded to massive, dark-gray, fossiliferous; contains gray chert nodules.
2	4	876	Limestone: hard, massive, medium-gray, dense.
1	21	897	Limestone: fossiliferous, dark-gray (weathers light gray); contains gray to light-tan chert nodules and lenses as much as 5 in. thick and 2 ft long. Underlying beds are covered.

Units 10 and 14 are rock types typical of Mansfield's Deadman limestone. South of Dunns Canyon on the west side of the Pruess Range these limestone beds are much thicker and crop out in a conspicuous comb. In most exposures red shale crops out in unit 9.

The thickness of the Wood shale tongue is somewhat variable; the thickness of 331 feet obtained in the above section compares with a thickness of about 200 feet in the valley of Preuss Creek only 2,000 feet away, and outcrop widths in the same general area indicate thicknesses as much as 400 feet. The variation in thickness is interpreted to be original, but it might be the result of faulting.

The color of the Wood shale tongue, which gives the impression of being brilliant red—although it is only moderate red when compared with the Munsell color chart—is different from that of any other unit. The hard dense grayish-red and pale-red dolomite of Mansfield's Deadman limestone, which seems lavender until compared with the color chart, is also distinctive.

JURASSIC SYSTEM

NUGGET SANDSTONE

Veatch (1907) applied the name Nugget to a sequence of yellow, pink, and red sandstone beds bounded by gray fossiliferous limestone of the Thaynes below and by shale and limestone of the Twin Creek limestone above; the name was derived from Nugget Station in southwestern Wyoming. Gale and Richards (1910) were the first to use the name in southeastern Idaho. Mansfield (1920) restricted Nugget to the sandstone between the beds now termed the Wood shale tongue of the Ankareh formation and the Twin Creek, but apparently included in the top of the formation the red-bed unit that was subsequently placed in the Twin Creek limestone by Imlay (1950, p. 37). Nugget is used in this report as restricted by Imlay.

The Nugget sandstone is about 900 feet thick between upper Crow Creek and Preuss Creek, but is 1,700 feet thick where it crosses Dunns Canyon. These thicknesses must be taken with some caution; the Nugget does not crop out in well-exposed beds but forms float-covered ridges and knobs, and structure cannot be mapped within the formation.

The Nugget consists of massive orange-pink and reddish-orange fine-grained quartz sandstone. It contains some dark grains (presumably chert) but very much less than does the salt-and-pepper sandstone of the Timothy member of the Thaynes formation. Cross-bedding has been noted, but it is much less common than planar bedding lamination. The Nugget is resistant, but it does not generally crop out in bedded exposures; typically, it upholds high brushy ridges and knobs that are littered with a jumble of angular blocks that weather pinkish gray or pale red. On steep hillsides these angular blocks form talus that is difficult and dangerous to negotiate.

The contact between the Nugget sandstone and the underlying Wood shale tongue of the Ankareh formation is gradational through several tens of feet.

No fossils have been found in the Nugget sandstone, but Mansfield (1927, p. 96) considered the formation to be of Jurassic age because of several possible unconformities in the interval between the Nugget and the fossiliferous rocks of the Thaynes formation and because the Nugget was considered the equivalent of beds that elsewhere contain Jurassic fossils. More recently, Gudim (1956, p. 72), among others, has placed the Nugget of Wyoming in the Triassic because of an unconformity between the Nugget and the overlying Gypsum Spring formation and because of interfingering between the Nugget and the Chugwater formation reported by Hubbell.[3] The evidence cited by Gudim does not seem sufficient to change the customary Jurassic age assignment of the Nugget in southeastern Idaho.

TWIN CREEK LIMESTONE

The Twin Creek limestone was named for a stream of that name in southwestern Wyoming (Veatch, 1907, p. 56). Imlay (1950) has recently studied the Twin Creek in considerable detail and has divided it into seven members, which in nearly all the area of this report are lithologically distinct and mappable. A section of the formation on the north side of Preuss Creek has been published by Imlay (1953, p. 60). The following partial section was measured by C. H. Marshall along Books Creek in the Gannett Hills in the east-central part of the Snowdrift Mountain quadrangle.

Partial section of the Twin Creek limestone

[Measured on the north side of Books Creek, sec. 12, T. 10 S., R. 45 E., Caribou County, Idaho]

TWIN CREEK LIMESTONE

Member G

[87 ft., total thickness of member]

Unit	*Thickness (feet)*	*Cumulative thickness*	*Description*
42	20	20	Covered: overlain by Preuss sandstone.
41	27	47	Sandstone: calcareous, glauconitic, light-olive-gray (weathers yellowish gray), very fine grained.
40	5	52	Sandstone: calcareous, glauconitic, thin-bedded, light-olive-gray (weathers greenish gray), fine- to medium-grained.
39	24	76	Sandstone: calcareous, glauconitic, thin-bedded, greenish-gray (weathers olive gray), fine-grained, crossbedded in part.
38	11	87	Sandstone: calcareous, glauconitic, thin-bedded, light-olive-gray.

[3] Hubbell, R. G., 1954, Stratigraphy of the Jelm, Nugget, and Sundance formations of northern Carbon County, Wyoming: Unpublished Masters thesis, Wyoming Univ.

TWIN CREEK LIMESTONE—Continued

Member F

[332, thickness of member F. Section was measured across a thrust fault, and the true stratigraphic thickness is probably about 1,500 ft]

Unit	Thickness (feet)	Cumulative thickness	Description
37	42	129	Covered: shaly limestone float.
36	11	140	Limestone: fissile, light-olive-gray (weathers grayish yellow), lithographic.
35	100	240	Covered: shaly limestone float.
34	2	242	Limestone: similar to unit 36.
33	53	295	Covered: shaly limestone float.
32	4	299	Limestone: similar to unit 36.
31	27	326	Covered: shaly limestone float.
30	48	374	Limestone: similar to unit 36.
29	43	417	Covered: shaly limestone float.
28	2	419	Limestone: similar to unit 36.
			Members D and E undifferentiated [208, total thickness of members D and E, undifferentiated]
27	85	504	Limestone: thin-bedded, light-olive-gray, lithographic with conchoidal fracture, veined with calcite; ridge former.
26	26	530	Limestone, fissile.
25	11	541	Limestone: similar to unit 27 but does not form ridge.
24	11	552	Covered: shaly limestone float.
23	32	584	Limestone: similar to unit 27.
22	16	600	Limestone and sandstone: limestone is massive, light pinkish gray, and oolitic; sandstone is calcareous, massive, and light olive-gray, and very fine grained; forms rib.
21	27	627	Limestone: hard, thin-bedded, dark-gray.
			Member C [260, total thickness of member C]
20	213	840	Covered: shaly limestone float.
19	24	864	Limestone: thin-bedded irregularly bedded, light-olive-gray, dense; contains 25 to 50 percent small shell fragments.
18	12	876	Covered.
17	11	887	Limestone: thin-bedded and irregularly bedded, pale-yellowish-brown, dense.
			Member B [252 ft. exposed]
16	25	912	Limestone: hard, thin-bedded, light-gray.
15	4	916	Tuff: siliceous, hard, light-green (weathers into gray and very light green angular blocky fragments).
14	16	932	Limestone: glauconitic, hard, thick-bedded, light-olive-gray, slightly pelletal.
13	16	948	Limestone: sandy, glauconitic, thin-bedded, yellowish-gray, very finely pelletal.
12	5	953	Limestone: glauconitic, hard, thick-bedded, light-olive-gray, finely colotic; contains some coquina layers.
11	16	969	Covered.

TWIN CREEK LIMESTONE—Continued

Member B—Continued

Unit	*Thickness (feet)*	*Cumulative thickness*	*Description*
10	5	974	Limestone: sandy, glauconitic, hard, thin-bedded, yellowish-gray; contains many pellets, oolites, and fossil fragments of medium-sand size.
9	8	982	Covered.
8	5	987	Limestone: sandy, thick-bedded, light light-gray (weathers yellowish gray), finely oolitic.
7	5	992	Covered.
6	11	1, 003	Limestone: hard, thin-bedded, gray, fossiliferous.
5	58	1, 061	Limestone: sandy, thin- and thick-bedded, light-olive-gray (weathering pinkish and yellowish gray), very finely pelletal, fossiliferous and crossbedded in part.
4	17	1, 078	Limestone: hard, massive, medium-gray, fossiliferous.
3	21	1, 099	Limestone: similar to bed above but more resistant.
2	20	1, 119	Covered.
1	20	1, 139	Limestone: hard, shaly, gray to yellowish-gray. Reddish soil below probably marks contact with member A.

The Twin Creek limestone thickens from about 3,000 feet in the Gannett Hills to about 5,000 feet on the south and east sides of Dunns Canyon. These thicknesses are only approximate, because the formation contains two very incompetent members that make reliable thicknesses difficult to obtain.

MEMBER A

The reddish-orange sandstone of the Nugget is overlain by a red shale unit, 130 to 250 feet thick, that was included in the Nugget by Mansfield but has been placed in the Twin Creek as member A by Imlay (1950, p. 37). It consists mostly of soft red siltstone but contains one or two beds of yellowish-gray brecciated limestone. Mansfield (1927, p. 96) reported gypsum in the member where it is exposed in the banks of Crow Creek near Halfway House.

Imlay (1950, p. 39) correlates member A with the Gypsum Spring formation of central Wyoming and considers it of Bajocian (early Middle Jurassic) age.

MEMBER B

The thickness of member B ranges from 200 or 250 feet on the east side of the area to as much as 600 feet on the west side. The basal contact is generally covered and the upper contact is gradational, although through only a few tens of feet.

The member consists mostly of tan-weathering sandy glauconitic calcarenite, a few gray well-indurated shell beds, and some dense limestone. The calcarenites are mostly mixtures of calcite oolites, pellets, and shell fragments. Glauconite occurs as well-rounded grains and as infiltrations in fossil fragments, especially in echinoid spines. The admixed sand is mostly quartz, generally well-rounded, but subangular feldspar makes up 10 or 20 percent of the terrigenous material in some samples, and chert grains are common. Much of the sandy limestone is well bedded and breaks into flat regular slabs. Many beds are cross laminated. The member is rather resistant and commonly forms strike ridges of intermediate height.

The upper third of member B contains an unusual altered tuff bed, 4 to 8 feet thick, that consists almost entirely of quartz and analcime, analcime comprising as much as 40 percent of the rock (Gulbrandsen and Cressman, 1960). The bed is siliceous, brittle, and apple green.

Imlay (1953, p. 55) dates member B as Bajocian.

MEMBER C

Overlying the resistant sandy limestone of member B are 250 to 600 feet of light-gray and light olive-gray shaly limestone that comprise member C. The unit thins from west to east and, to a lesser degree, from south to north.

The limestone of member C is very dense and lithographic on fresh surfaces. The fissility suggests that the rocks are very argillaceous, but clay is not apparent either in hand specimen or thin section. The beds weather into very light gray flaky and splintery fragments that thickly cover slopes and swales developed on the member.

Imlay (1950, p. 39) states that the upper half of the member is not older than Bathonian (late Middle Jurassic).

MEMBER D

The soft shaly limestone of member C is overlain by a rib of limestone and sandstone that is, in turn, overlain by a soft silty zone. The silty zone is not exposed but forms a band of reddish soil. The rib and the overlying red unit comprise member D. Like member C, member D thins both eastward and northward, the eastward thinning being more pronounced. It ranges in thickness from 40 feet in parts of the Gannett Hills to nearly 600 feet on the ridge west of upper Montpelier Creek. North of Warm Creek in the Gannett Hills the member is too thin to map separately, and it has been grouped with member E. In the measured section on Books Creek, units 21 and 22 are parts of the lower rib of the member, but the overlying red unit is either not present or is covered.

In most of the area the basal part of the member consists of thin- to thick-bedded dense gray limestone similar to that of member E. The main part of the resistant rib is brownish-gray oolitic limestone and very fine grained glauconitic sandy limestone and calcareous sandstone. The brownish-gray oolitic limestone is characteristic of the member. The sandstone, like that of member B, is feldspathic.

The topographic expression of the lower resistant limestone and sandstone and the upper nonresistant red siltstone is so distinct that in parts of the area the member can be located accurately by topography alone.

Member D is Bathonian (Middle Jurassic) in age (Imlay, 1950, p. 40).

MEMBER E

Member E thickens from about 120 feet in the Gannett Hills to nearly 900 feet on the west side of Montpelier Creek. In its westernmost outcrop belt it thins slightly northward, being about 700 feet thick near Hawks Roost. It is a prominent ridge former.

In the Gannett Hills member E consists mostly of hard thin- to thick-bedded dense, lithographic light-olive-gray limestone. Jointing is well developed in nearly all exposures, and great care is necessary in distinguishing bedding from jointing. Silt laminae have been noted, but they are not common. Westward in the area sand appears in the base of the member, and on the south side of Whiskey Flat the basal 300 to 400 feet consists of interbedded gray lithographic limestone and cross-laminated grayish-brown fine-grained sandstone. The sandstone thins northward as well as eastward; it is present in significant amounts on the divide between South Canyon and Dunns Canyon, but at Hawks Roost sandstone is only a very minor component.

Member E grades into the overlying member, but the zone of gradation is thin and the contact can be closely located.

The member is of earliest Callovian (early Late Jurassic) age (Imlay, 1950, p. 40).

MEMBER F

Member F consists of shaly limestone that is generally poorly exposed and weathers to form light-colored bare slopes covered with splintery float. Exposures are closely jointed, and fracture cleavage has formed locally; bedding can seldom be determined with confidence and folds apparent in the underlying and overlying beds generally cannot be detected in member F; at places thrust faults occur in the member, but they are not marked by breccia zones. For these reasons, the thickness of member F is not known. It is about 1,600 feet thick in Thomas Fork Canyon about 7 miles southeast of the area, and the unfaulted thickness may be about the same in the vicinity of Preuss and Crow Creeks.

Lithologically the limestone of the member is very similar to that of member C. It is light olive gray and lithographic on fresh surfaces and breaks into grayish-yellow and white flakes, chips, and pencilshaped slivers. The shaly fracture suggests that the limestone is argillaceous, and scattered silt and clay laminae are present. Near the top of the member are several thick beds of bioclastic limestone that locally are useful markers.

The member is early Callovian in age (Imlay, 1950, p. 38).

MEMBER G

The uppermost member of the Twin Creek limestone is about 75 feet thick. However, both the upper and lower contacts are gradational, and map thicknesses are therefore somewhat variable.

Member G consists of thin-bedded olive-gray and greenish-gray very fine to medium-grained glauconitic calcareous sandstone; the sand is similar in composition to that of members B and D. The greenish-gray color on weathered surfaces is unmistakable, and the member is a valuable marker. Crossbedding is common.

Like members E and F, member G is early Callovian in age (Imlay, 1950, p. 41).

PREUSS SANDSTONE

Mansfield and Roundy (1916, p. 8) named the Preuss sandstone for Preuss Creek in the area of this report.

The thickness of the Preuss in the mapped area is not known. It is too poorly exposed and too uniform in character from base to top to enable the mapping of structure within the formation, and inasmuch as it crops out in a belt known to contain thrust faults and tight folds, the thickness cannot be determined from outcrop widths. An assumed thickness of 1,700 feet results in the simplest structural interpretation. Mansfield (1927, p. 99) states that the Preuss is 1,300 feet thick at the head of Thomas Fork Valley, and considering the rapid westward thickening of the rest of the Jurassic in the area, 1,700 feet is a reasonable thickness for the formation near Crow Creek.

The Preuss consists mostly of reddish-brown, pale-red, and grayish-red calcareous sandstone in beds 1 to 6 inches thick. The terrigenous material is similar in composition to that of sand in the Twin Creek limestone—quartz grains are the most abundant, but chert, glauconite, and feldspar are common. The average grain size of the sandstone ranges from very fine to very coarse, though fine-grained sand is the most common. There are some beds of granule conglomerate, and some sandstone contains shale chips. Most sandstone beds in the Preuss are less than 6 inches thick, although beds as much as 2 and 3 feet thick are not too uncommon. Crossbedding is rare. Roadcuts on the east side of Whiskey Flat expose several rhythmic sequences of interbedded sandstone and shale in which the sandstone beds are progressively thinner from base to top within each sequence.

A well drilled in the Preuss sandstone in Tygee Valley near the Idaho-Wyoming line about 9 miles north of the Snowdrift Mountain quadrangle was reported by Mansfield (1927, p. 340) to have penetrated 6 beds of salt totaling 96 feet thick. Another bed of salt, at least 12 feet thick and presumably in the Preuss, is present on the east side of Crow Creek 3 miles downstream from the east border of the Snowdrift Mountain quadrangle (Mansfield, 1927, p. 339). No exposures of evaporites in the Preuss have been found in the area of this report, but several small saline springs issuing from alluvium on the east side of Crow Creek near areas underlain by the Preuss suggest that salt beds may be present in that formation below the zone of weathering.

A typical exposure of Preuss sandstone is a rounded hill of moderate height, sparsely vegetated with grass and sagebrush, with pale-reddish-brown soil containing somewhat slabby sandstone float and with a few widely separated outcropping ribs of sandstone.

The Preusss is not fossiliferous but is underlain and overlain by beds of early Late Jurassic age.

STUMP SANDSTONE

The Stump sandstone was named by Mansfield and Roundy (1916, p. 81) for Stump Creek in the Freedom quadrangle of southeastern Idaho. In the Snowdrift Mountain quadrangle the formation as mapped is 300 to 500 feet thick. The mapped thickness is probably more variable than the true thickness inasmuch as both the upper and lower contacts are nearly everywhere located by the change in color of the soil and by float. The lower contact can generally be located with considerable confidence by the float, but in many places the contact with the overlying Gannett group cannot be located to within several hundred feet.

The Stump consists mostly of greenish-gray and olive-gray calcareous and glauconitic sandstone; the sand itself is mostly quartz but, like that of the other Jurassic formations, contains considerable feldspar. Much of the sandstone is crossbedded. Even though the Stump is thin bedded, it generally crops out in a ridge that on one side rises above slopes developed on the upper part of the Preuss sandstone and on the other side slopes down to a swale that separates the Stump from the Gannett group.

The following section was measured by C. H. Marshall in 1955.

Section of Stump sandstone at head of Books Creek, NE¼ sec. 7, T. 10 S., R. 46 E., Caribou County, Idaho

Unit	Thickness (feet)	Cumulative thickness	Description
18	53	53	Covered: float is typical of Stump. Contact with overlying Gannett group located at appearance of red float and soil.
17	11	64	Sandstone: calcareous, glauconitic, medium-hard, thin-bedded and cross-bedded, greenish-gray, fine-grained.
16	37	101	Covered.
15	16	117	Sandstone: calcareous, glauconitic, medium-hard, thin- and cross-bedded, greenish-gray, fine-grained.
14	11	128	Covered.
13	27	155	Sandstone: calcareous, glauconitic, medium-hard, thin- and cross-bedded, light olive-gray, fine-grained.
12	37	192	Covered.
11	32	224	Sandstone: calcareous, glauconitic, medium-hard, thin- and cross-bedded, greenish-gray, fine-grained.
10	27	251	Covered.
9	5	256	Sandstone: calcareous, glauconitic, medium-hard, thin- and cross-bedded, greenish-gray, fine-grained.
8	27	283	Covered.
7	53	336	Sandstone: calcareous, glauconitic, medium-hard, thin- and cross-bedded, greenish-gray, fine-grained.
6	4	340	Sandstone: calcareous, glauconitic, hard, thin- and cross-bedded, dark-grayish-green (weathers light yellowish gray), fine-grained; contains some ripple marks and shell casts.
5	8	348	Sandstone: calcareous, glauconitic, thin-bedded and irregularly bedded, light-grayish-green, fine-grained.
4	5	353	Sandstone: calcareous, glauconitic, hard, thin-bedded, dark-grayish-green (weathers light yellowish gray), fine-grained; contains some ripple marks and shell casts.
3	4	357	Sandstone: calcareous, glauconitic, thin-bedded and irregularly bedded, light-grayish-green, fine-grained.
2	5	362	Limestone: sandy, glauconitic, hard, thin-bedded, olive-gray (weathers light yellowish gray), finely oolitic; contains some ripple marks and shell casts.
1	40	402	Covered: contact with underlying Preuss sandstone placed at change of float from olive gray to reddish brown at base of unit.

According to Imlay (1950, p. 41) the Stump sandstone is of Oxfordian age (early Late Jurassic).

CRETACEOUS SYSTEM

GANNETT GROUP

The Gannett group was named by Mansfield and Roundy (1916, p. 82) from the Gannett Hills. In westernmost Wyoming the group divided into the following formations from base to top: the Ephraim conglomerate, the Peterson limestone, the Bechler conglomerate, the Draney limestone, and an unnamed red-bed unit (Moritz, 1953, p. 63). In the area of this report the Draney limestone and the overlying red-bed unit are not exposed, and the Peterson limestone has not been positively identified; the formational breakdown, therefore, cannot be used.

About 5,000 feet of beds is assigned with confidence to the Gannett group. About 2,000 feet above the Stump sandstone is a zone of light-gray calcareous conglomerate that contains, in addition to pebbles of older rocks, fragments similar in character to the Peterson limestone.[4] The zone is not a practicable mapping unit. If this light-colored zone is the equivalent of the Peterson, then the Gannett group as mapped includes both the Ephraim and Bechler conglomerates. Another possibility is that the rocks mapped as Gannett group are all Ephraim equivalents and that the overlying 1,000 feet of conglomerate discussed in the section below may be part of the Bechler conglomerate. However, the Ephraim is only about 800 feet thick along U.S. Highway 89 in Wyoming 8 miles southeast of Red Mountain, and the rate of thickening required if all the beds mapped as Gannett group are Ephraim conglomerate seems excessive.

The Gannett group crops out as massive ledges of conglomerate separated by intervals of reddish-brown soil that are probably underlain by sandstone and shale. Except for the calcareous unit in the middle, the entire group is reddish. Although little of the Gannett group is exposed other than conglomerate and some minor interbedded sandstone, actual outcrops of conglomerate make up only 5 to 10 percent of the total thickness of the interval on the slopes of Red Mountain, suggesting that the group is mostly sandstone and shale. Because of the interbedding of massive conglomerate ledges and nonresistant sandstone and shale, the Gannett group is especially susceptible to landsliding. One large slide has been mapped, but slides too small and indistinct to map have formed on a number of steep slopes eroded on the Gannett group.

The outcropping ledges of the Gannett group are composed of massive light-red pebble and cobble conglomerate containing mostly fragments of chert and hard sandstone in a sandy matrix. Limestone

[4] These light-colored beds were mapped as the Salt Lake formation of Tertiary age by Mansfield (1927, pl. 9), but they are conformable with the Gannett group both above and below and do not contain the tuffaceous material typical of the Salt Lake formation.

pebbles and cobbles are absent in the basal beds but are common in the rest of the unit. The average pebble diameter is about 1 or 2 inches, but it ranges considerably from bed to bed; the maximum size is about 10 inches. Many of the limestone and sandstone fragments have been derived from the Wells and Brazer formations. A very few quartzite pebbles derived from the Swan Peak formation of Ordovician age are present in the basal beds (F. C. Armstrong, oral communication, 1956), but they are extremely rare, and except for these few quartzite pebbles, no fragments older than Mississippian have been recognized.

The ledges also contain layers and lenses of light-red poorly sorted quartz and chert sandstone, much of which is pebbly. Crossbedding is common.

The Gannett group is considered to be of Early Cretaceous age on the basis of faunas from the Peterson and Draney limestones; but the age of the Ephraim conglomerate is not known, and it may possibly contain Jurassic beds in the lower part (Cobban and Reeside, 1952, p. 1030).

GANNETT(?) GROUP

At an altitude of about 8,300 feet on the west flank of Red Mountain the character of the conglomerate changes somewhat. Although also mostly light red in color, the conglomerate ledges above 8,300 feet are thicker, constitute more of the interval, and contain larger boulders, the maximum size being about 1½ feet. In contrast with the underlying unit, rounded boulders and cobbles of Swan Peak formation (Ordovician) are conspicuous even though they comprise a very small part of the conglomerate. Some purple quartzite fragments that may have been derived from the Brigham quartzite of Cambrian age were found in the uppermost beds. The unit is about 1,000 feet thick.

This conglomerate was mapped by Mansfield (1927, pl. 9) as Ephraim, but Reeside and Rubey (W. W. Rubey, oral communication) have suggested that it may be part of the Wasatch formation of Paleocene and Eocene age. No evidence of the age of these beds has been obtained in this study, but it seems unlikely that they could be part of the Ephraim inasmuch as that formation would then be at least 6,000 feet thick; they may be part of the Bechler conglomerate, but here, too, the thickness of beds assigned to the Ephraim would seem too great. On the other hand, the two units differ only in pebble size, the relative amounts of conglomerate, and in the relative amounts of Swan Peak boulders, and no clear indications of unconformity between the units has been found. The age of this conglomerate may thus be either Cretaceous or Tertiary.

TERTIARY SYSTEM

WASATCH FORMATION

Patches of red conglomerate are exposed on the west face of the Aspen Range north of Pine Canyon and in several small patches in The Hole near the head of Pine Canyon. At both places the conglomerate rests with profound angular unconformity on the underlying Mississippian rocks, although the relationship is somewhat obscured in The Hole by a younger gravel deposit. The conglomerate is very poorly sorted; boulders are as large as 3 feet in diameter, although the average size is about 3 or 4 inches. The fragments are mostly subrounded, and many have been brecciated after deposition. Many fragments can be recognized as having been derived from lower Paleozoic formations. Among these are boulders of Swan Peak formation, pebbles of dark dolomite from either the Jefferson dolomite (Devonian) or Fish Haven dolomite (Ordovician), and fragments from the Worm Creek quartzite member of the St. Charles limestone (Cambrian) and Brigham quartzite (Cambrian). The conglomerate does not crop out boldly but forms low rounded discontinuous ledges; areas of red soil containing lower Paleozoic boulders must be searched carefully to ascertain whether they are actually underlain by red conglomerate or whether the red bouldery material is merely a gravel capping on other formations.

The red conglomerate is younger than most of the folding and older than most of the range-front normal faulting, dating it very imprecisely as somewhere between Late Cretaceous and latest Tertiary in age; not even its age relations to the Salt Lake formation can be determined directly, for the two units are in fault contact. Red sedimentary rocks are not common in the post-Eocene of the central Rocky Mountains, probably because of the progressive regional climatic desiccation demonstrated by Axelrod (1950) to have started in about middle Eocene time, and the red conglomerates of the Aspen Range are therefore probably not younger than Eocene.

The red conglomerate was mapped as the Wasatch formation (Eocene age) by Mansfield (1927, p. 109, pl. 6). The exposures in the area of this report are small, isolated, and unfossiliferous; so their relation to the Wasatch of the type area is not known. However, the red conglomerate is similar in character to the Wasatch formation mapped by Mansfield on the Bear Lake Plateau, and the Wasatch of that area appears to be continuous, at least in part, with the fossil-bearing Wasatch of southwestern Wyoming (Mansfield, 1937, p. 109).

SALT LAKE FORMATION

The white and light-gray tuff, marl, sandstone, and conglomerate exposed along the west front of the Aspen and Preuss Ranges were assigned by Mansfield (1927, p. 110, pl. 6) to the Salt Lake formation, and his terminology is followed herein. Similar tuffaceous and conglomeratic rocks exposed in upper Slug Valley near Summit View and in upper Crow Creek valley are also included in the formation. Calcareous pebble conglomerate that crops out on the west side of Crow Creek where it leaves the Snowdrift Mountain quadrangle is also assigned to the Salt Lake, although the rock is not tuffaceous and may actually be younger than other beds assigned to the Salt Lake.

The Salt Lake formation either overlies the older rocks with marked angular unconformity or is in fault contact with them.

The Salt Lake formation consists of a lower tuffaceous member and an upper conglomeratic member that are not differentiated on the map. The lower tuffaceous member is mostly light-gray to white fine-grained tuff, dense limestone, oolitic limestone, and calcareous and tuffaceous sandstone and conglomerate. Except near the base and top of the member, conglomerate is not very common; wherever it is present, the pebbles are locally derived, and lower Paleozoic formations are not represented. The member is soft and is easily eroded. It underlies most of the low hills that stretch across to the Bear River Range and separate Bear River valley from Bear Lake Valley and forms conspicuous white outcrops where U.S. Highway 30 N. crosses the hills just north of Georgetown. The thickness of the tuffaceous member is not known, but Armstrong (1953) estimates about 4,000 feet near Soda Springs and a well drilled in the southern part of Bear River valley (just west of the area of this report) penetrated at least 2,000 feet and perhaps as much as 2,900 feet. Most of the beds are lacustrine as shown by the oolitic limestone and by the molluscan fauna (Yen, 1946, p. 485).

The upper member consists of cobble conglomerate. The cobbles are subangular and subrounded, and the cement is calcareous. The fragments have been derived locally; those from the lower Paleozoic formations are uncommon or absent. In the area of the map (pl. 3), the conglomerate beds are well exposed in a deep gully on the north side of lower Georgetown Canyon. The total thickness is not known, but it is more than 1,000 feet. The Salt Lake formation between Georgetown Canyon and the Left Fork of Twin Creek is composed of the conglomerate member, as is most of the Salt Lake in the mapped area south of Georgetown Canyon.

The conglomerate member may rest with slight angular unconformity on the tuffaceous member; the conglomerate beds on the east side of the Left Fork of Twin Creek dip 15° to 25° to the east, whereas the tuffaceous member on the east side of the creek dips as much as to 55° south. However, the contact between the members is not exposed, and the unconformable relations are by no means demonstrated.

Mansfield (1927, p. 110–112) provisionally dated the Salt Lake formation as Miocene. Southwest of Soda Springs it has been assigned provisionally to the "Pliocene-Pleistocene" on the basis of gastropods (Teng-Chien Yen, written communication, 1951) and to the Pliocene(?) on the basis of diatoms (K. E. Lohman, written communication, 1951). Fresh-water molluscans from the Salt Lake formation 9 miles northwest of Montpelier were identified by Yen (1946, p. 487) as upper Miocene. All the faunas on which these dates were based were collected from the lower tuffaceous member.

In recent years the Salt Lake in adjacent areas has been divided into several formations, each system of formational nomenclature generally being restricted to one basin (Slentz, 1955; Adamson and others, 1955; Merritt, 1956). Typically, the sedimentary rocks in these areas consist of a tuffaceous lacustrine formation, generally of Pliocene age but at places perhaps as old as Oligocene, overlain by a conglomerate that cannot be closely dated. The gross similarity between the Salt Lake formation in the area of this report and the late Tertiary basin sediments in the adjacent areas is evident, and the Salt Lake in the area of this report could certainly be divided into two formations. However, before any such formal subdivision is attempted, the extensive exposures in Bear Lake Valley should be mapped and collected in detail.

OLIVINE BASALT

A patch of vesicular olivine basalt about a mile in diameter is exposed in upper Slug Valley, and several small patches of very similar basalt have been found in the Aspen Range and near the south end of Schmid Ridge. Where the basalt rests on the lower tuffaceous member of the Salt Lake formation, it seems conformable with it, but it laps onto the adjacent Paleozoic rocks. The basalt thus seems younger than the lower tuffaceous member of the Salt Lake and is so shown on the map explanation (pl. 3). However, in the Soda Springs area Armstrong (1953) shows a basalt layer occurring within the Salt Lake formation, and the well drilled in Bear River valley just west of the mapped area passed through several layers of basalt within the Salt Lake. The relation of the basalt to the conglomeratic member of the Salt Lake is not known.

The source of the basalt is not known, but considering that most of the more recent basalt in the Soda Springs area issued from fissures and craters in Bear Lake Valley, the source of the olivine basalt was probably also within the valley.

Because of the conformable relations with the Salt Lake formation, the olivine basalt is believed to be no younger than Pliocene.

TERTIARY OR QUATERNARY SYSTEMS

HIGH-LEVEL PEBBLE AND BOULDER DEPOSITS

Most of The Hole, which is between 7,400 and 7,700 feet in altitude, is mantled by sandy soil that contains rounded cobbles and boulders of quartzite derived from the Swan Peak formation and small angular fragments of sandstone but very few limestone fragments. Many of the quartzite boulders have been stream worn so that their surface is smooth and highly polished, but much of the polish has subsequently been removed by weathering. The scarcity of limestone fragments indicates that the deposit is not simply the result of recent weathering of the underlying red conglomerate; rather, it is thought to be a pediment gravel formed where a reentrant of a late Tertiary or early Pleistocene pediment beveled the east-dipping red conglomerate.

A very similar deposit of sandy soil containing boulders of lower Paleozoic quartzite has been mapped at an altitude of about 7,400 feet near Summit View, where it may have been deposited in a pediment pass. The source of the boulders was presumably the Wasatch formation, which must have been much more widely distributed in the past than it is now.

Rounded fragments of lower Paleozoic quartzite rest on the patch of basalt that caps the divide on the west side of the Left Fork of Twin Creek. Similar material caps the Salt Lake formation above about 7,000 feet on the east side of the Left Fork of Twin Creek; the gravel capping differs from conglomerate in the Salt Lake formation in that the gravel contains many fragments of lower Paleozoic rocks, suggesting that it, too, is a pediment deposit. Neither of these latter occurrences is sufficiently thick or distinct to map.

QUATERNARY SYSTEM

OLDER COLLUVIUM AND ALLUVIUM

Several types of unconsolidated deposits that for the most part are not in equilibrium with the present drainage and topography are here grouped as "older colluvium and alluvium." However, the differences between the unconsolidated deposits are not so distinct as between the older formations, and there is doubtless some inconsistency in their classification and mapping; for example, some of the material mapped as older hill wash may well be still accumulating. Nevertheless, the general grouping is believed to be valid.

The relative ages of the units included in this group are uncertain.

OLDER GRAVEL

Patches of rounded pebbles and cobbles that have been almost entirely derived from the Brazer limestone rest on top of two small hills, one on the north side and one on the south side of the mouth of White Dugway Creek. The gravel is from 100 to 150 feet above the

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valley floor. A gravel deposit consisting mostly of debris from the Wells formation but containing rounded cobbles and boulders of Rex chert member of the Phosphoria formation, Brazer limestone, and quartzite from the Swan Peak formation occurs from 50 to 200 feet above the valley on the first knob south of Wells Canyon on the east side of Crow Creek.

These gravel deposits are higher than the reconstructed surface of the dissected fan that fills much of Crow Creek valley and are therefore older than the old alluvial fan deposits described below. On plate 3 these older gravels are grouped with the younger terrace gravels as a single unit.

OLDER ALLUVIAL FANS

Much of Crow Creek valley is covered by unconsolidated sand and silt that contains unrounded pebbles, cobbles, and small boulders. The deposits on the east side of the valley are red and calcareous, and the pebbles are derived from the formations exposed in the Gannett Hills. The red color has evidently been derived from the Preuss sandstone and the Gannett group, which comprise much of the drainage area. The deposits on the west side are gray and brown, and the pebbles have come from rocks exposed in the Deer Creek, Wells Canyon, and Clear Creek drainages. These deposits are now being dissected, but remnants of an original surface indicate that the sediments on the east side of Crow Creek valley were originally part of alluvial fans that spread out from Warm Creek and White Dugway Creek. Although no remnants of an original surface are preserved on the west side of the valley, the deposits there probably are also dissected fan remnants.

An area of older alluvium in upper Slug Valley is also included in the unit.

TERRACE GRAVELS

Stream-worn gravel, which is similar to that in the present stream channel and consists of all the resistant rock types in the drainage area, coats a strath terrace that is about 10 to 20 feet above the flood plain of Crow Creek. The gravel averages 8 or 10 inches in diameter, but some boulders are several feet across. There is much silt and sand matrix, and the terrace supports a growth of sagebrush. North of the mouth of Deer Creek stream-worn gravel has been mapped as high as 60 feet above the flood plain, and patches of rounded boulders too small and sparse to map have been found as high as 120 feet above the flood plain. More detailed work probably would demonstrate that the material here mapped as one unit occurs on several different terrace levels.

In the Black Dugway area, scattered water-worn boulders rest on shoulders on the west side of the road about 80 feet above the present stream level, but they are not sufficiently numerous or extensive to map.

In Georgetown Canyon, terrace gravels cap older rock at an altitude of about 100 feet above the stream; others seem to be remnants of an earlier valley fill, weakly cemented by caliche. Detailed work in Georgetown Canyon together with a study of terraces in Bear Lake Valley undoubtedly would reveal an interesting alluvial history.

YOUNGER COLLUVIUM AND ALLUVIUM

ALLUVIUM

Most of the material mapped as alluvium consists of silt, sand, and gravel deposits along stream courses and in valley bottoms. It also includes some Recent hill wash, particularly near Goodhart Spring in the northwest corner of the Snowdrift Mountain quadrangle, on the margins of Dry Fork Basin, and on the west shoulder of upper Georgetown Canyon.

ALLUVIAL FANS

It is characteristic of the area that actively accumulating alluvial fans are small. Where large enough, they have been mapped separately; otherwise they have not been distinguished from alluvium.

LANDSLIDES

Landslides are not so common as one might expect in an area of such steep relief. Only four slides have been mapped, although other small ones have been noted. The largest slide is in the Gannett Hills in the southeast corner of the Snowdrift Mountain quadrangle where a landslide of material from the Gannett group has blocked the north end of Ephraim Valley. Although not reflected by the contours on the topographic map, the surface of the slide is irregular and hummocky. Two small rocks slides were mapped in Deer Creek, one where the canyon crosses the Brazer limestone in the core of the Snowdrift anticline and the other where the canyon cuts the Brazer in the center of the Boulder Creek anticline. The rock slides are probably composite, for in parts limestone blocks support lichen, whereas in other parts the blocks are lichen free. High on the east side of South Canyon blocks of Brazer limestone have slid down a steep slope developed on the Dinwoody formation. As seen from across South Canyon, the topography of the slide surface is hummocky but subdued.

TALUS

Small accumulations of slide-rock are common, particularly where the Brazer limestone and the lower member of the Wells formation crop out; but only those accumulations large enough to obscure significant parts of the bedrock geology have been mapped. These are all derived from the Nugget sandstone.

STRUCTURE

GENERAL FEATURES

The Georgetown Canyon-Snowdrift Mountain area may be divided into two parts on the basis of the age of the outcropping pre-Tertiary rocks: The Gannett Hills, Whiskey Flat, and the area west of South Canyon and Montpelier Creek consist of Jurassic and Lower Cretaceous rocks, whereas the other and far larger part consists nearly entirely of upper Paleozoic and Lower Triassic rocks. These two areas are separated by a fault, or in places by a fault zone, that was shown by Richards and Mansfield (1912) to be the trace of a large folded overthrust, here termed the Meade overthrust.[5] The older rocks are thus parts of a large overthrust sheet. Erosion of parts of the upper plate has exposed the younger rocks of the underlying lower plate. Strata in both the upper and lower plates have been compressed into north-striking folds, and the west side of the area is broken by a number of normal faults that offset both the folds and the thrust.

STRUCTURE OF THE UPPER PLATE OF THE MEADE OVERTHRUST

Of the rocks exposed on the upper plate of the Meade overthrust (exclusive of the upper Triassic and Jurassic formations near Preuss Creek), the Brazer limestone, the lower part of the Wells formation, the Rex chert member of the Phosphoria, and the upper member of the Dinwoody formation are competent, whereas the Meade Peak phosphatic shale member of the Phosphoria, the lower member of the Dinwoody, and the lower black shale member of the Thaynes formation are incompetent. The Madison limestone and the upper part of the Wells formation seem intermediate in competency. The folds in the upper plate are concentric, as shown by the relatively uniform thickness of the mapping units regardless of their position in the folds and by the absence of cleavage, even in the most incompetent beds.

Evidence discussed in following paragraphs indicates that the thrust plane beneath the upper plate is about at the base of the Madison limestone and that most of the folding of the upper plate occurred during thrusting. When folding was initiated, the upper plate consisted of a sequence about 24,000 feet thick. The strata exposed on the upper plate (again exclusive of the Preuss Creek area) are a little

[5] Richards and Mansfield (1912) named the overthrust the Bannock and postulated its extension throughout much of southeastern Idaho. In particular, they believed Bear River valley to be a window in the thrust and thought the large overthrust on the west side of the valley to be part of the same fault as that in the Peale Mountains. Reappraisal of the structure of southeastern Idaho (Armstrong and Cressman, 1963) has shown that the Bear Lake window does not exist and that the thrusts on the two sides of the valley are different faults. The name Bannock is therefore replaced in the area of this report by the name Meade overthrust, the name being derived from Meade Peak. The thrust on the west side of the valley has been renamed the Paris thrust (Armstrong and Cressman, 1963).

less than 7,000 feet thick from the base of the Madison limestone. The upper plate as now exposed therefore represents only the lowermost part of the original folded sequence.

The axial surfaces of most of the folds dip to the west; the principal exception is the Snowdrift anticline, in which the axial plane is nearly vertical. The east limbs of several anticlines are overturned and cut by thrusts that for the most part occur within or between the incompetent lower member of the Dinwoody formation, the Meade Peak member of the Phosphoria formation, or the upper member of the Wells formation.

In the following discussion the individual folds and their related faults in the overthrust plate are discussed in sequence from west to east. Normal faults that displace beds in both plates and that formed after folding and thrusting are discussed on pages 80–83. Except for the Summit View anticline, the names of the folds are those first applied by Mansfield (1927).

DAIRY SYNCLINE

The Dairy syncline is the westernmost large fold in the older rocks. Where the syncline is best developed, it is tightly folded, markedly overturned to the east, and nearly recumbent (pl. 4, section *B-B'*). This is apparent only near the axial part, for erosion has removed most of the overturned limb. The south end of the fold is cut by a transverse fault and dropped downward to the west by a large normal fault. It disappears beneath the Salt Lake formation about a mile north of the juncture of the forks of Twin Creek. North of Harrington Peak the fold has been tipped eastward, apparently by normal faulting, and beds corresponding to the east limb of the fold actually dip gently to the east (pl. 4, section *A-A'*). Northward the fold gradually tips westward again. Like most other folds in the overthrust plate, the Dairy syncline plunges gently northward at an average of 2° or 3°. The west side of the syncline is bordered nearly its entire length by normal faults, along which the overturned beds of the west limb are in contact with right-side-up and gently dipping beds that underlie the western part of the Aspen Range. The anticline that must have been adjacent to the Dairy syncline on the west has been faulted out.

SUMMIT VIEW ANTICLINE

The Summit View anticline extends from the edge of the thrust plate south of Summit View northward for nearly 3 miles to the head of Slug Valley, where it abuts against a normal fault that marks the west side of Schmid Ridge. It plunges about 5° or 6° northward. An abrupt change in the strike of the trace of the axial surface just north of the Summit View road is thought to be the result of gentle flexing after the principal folding along an axis striking N. 15° W.

SCHMID SYNCLINE

The axial plane of the Schmid syncline strikes N. 10° W. for 1¼ miles north of where it terminates at the edge of the thrust sheet, but it then strikes north-northeast. Near the sharp bend in the strike, the fold is overturned to the west (pl. 4, section *B–B′*), in contrast to most folds in the area which have westward-dipping axial planes. It is tightly folded and overturned in its southern part but broadens northward into a relatively open fold; its axis plunges an average of 3° to the north. The west side of the Schmid syncline is cut by at least two west-dipping longitudinal normal faults, the westernmost one forming the west face of Schmid Ridge.

DRY VALLEY ANTICLINE

The trace of the axial plane of the Dry Valley anticline extends from near the south end of Dry Ridge, where it is truncated by a fault, north-northeastward along the west side of Dry Ridge to Dry Valley. It is mostly tightly folded, nearly isoclinal, and generally overturned slightly to the east; the segment just east of Summit View, however, is overturned to the west.

The east limb of the anticline is cut by a thrust zone. The largest thrust has a stratigraphic displacement of about 1,600 feet, and the other thrusts in the zone appear to be relatively small and discontinuous. Thrusts cut out part of the Brazer limestone, thin the Wells formation, thin or cut out the Meade Peak member of the Phosphoria formation, and thin the Rex chert member of the Phosphoria and the lower member of the Dinwoody. The thrusts die out northward before reaching the edge of the area. Southward they cross Georgetown Canyon at its bend, continue southeastward, and terminate against tear faults in South Canyon. The dips of most of the thrusts are not known, but they generally seem to be approximately parallel to the bedding and therefore probably dip 60° or 70° to the west. Several normal faults occur in the same fault zone as the thrusts and at one place repeat the Rex chert member three times.

GEORGETOWN SYNCLINE

The Georgetown syncline is a large complex fold, the west limb of which is overturned. It is as much as 3 miles across and extends in some form from the south end of the thrust sheet to 30 or 40 miles beyond the north end of the area. It plunges gently but irregularly northward.

In gross aspect the Georgetown syncline is broad and overturned to the east and has a minor fold developed on the normal limb (pl. 2).

The Lower Triassic rocks in the center of the syncline have been compressed into a number of small discontinuous folds that die out laterally and therefore probably vertically before reaching the Phosphoria formation. The overturned limb is thinned by the thrust zone described in the previous sections.

Where the surface trace of the Phosphoria formation swings around the plunging nose of the eastern minor syncline, it bends very sharply and is offset by several transverse faults. The Meade Peak member is noticeably thinned for at least 1,000 yards along strike on the east side of the fold and for about 1,000 feet on the west side. As shown in structure sections *C–C′* of plate 4 and *E–E′* of plate 2, both observed attitudes and required unit thicknesses indicate that this sharp flexure does not extend far down the plunge of the syncline but that the fold broadens rapidly to the north. The sharp bend and the transverse fault in the Phosphoria and the thinning of the Meade Peak member may be related in some way to the northeast-striking longitudinal fault within the Wells formation on the east limb of the syncline. This fault is shown as being nearly vertical in the sections and on the map, but it can be closely located in only two or three places, and it may well be a thrust that dips to the northwest.

South of Phosphoria Gulch details of the structure of the Georgetown syncline are obscured by poor exposures and complicated by faulting. Probably the faulted synclinal segment of the Wells, Phosphoria, and Dinwoody formations in South Canyon is part of the Georgetown syncline that has been displaced about a mile southeastward along the large tear fault that offsets the main ridge of Snowdrift Mountain.

SNOWDRIFT ANTICLINE

Throughout most of its length the Snowdrift anticline is narrow, tightly folded, and nearly isoclinal. The resistant Brazer limestone is brought to the surface along the axis and supports the high ridge of Snowdrift Mountain, the trace of the axial plane being either on or near the crest of the ridge. The southermost 2½ miles of the anticline has been displaced nearly a mile to the southeast along a tear fault; most of the segment of the fold south of the fault strikes a little west of north, contrasting with the strike of N. 20° E. of the rest of the anticline. The southern part of Snowdrift anticline plunges gently northward, but the fold axis throughout most of its length is about horizontal. On the east side of the anticline, the large tear that offsets the southern end of the fold turns southward into a thrust along which the Wells formation is thrust eastward onto the uppermost part of the Dinwoody formation. Near Greyhound Pass Spring, a high-angle fault that crosses the anticlinal axis at an oblique angle repeats the axis; in the northern part of the area a small thrust has developed between the Phospheria and Dinwoody formations of the east flank.

WEBSTER SYNCLINE

The west limb of the Webster syncline is steep, the dip ranging from 65° to vertical. The east limb is gently dipping; therefore the fold is rather open. Within the Phosphoria and Dinwoody formations the limbs of the folds are relatively straight, the dips on the flanks being nearly constant almost to the axis, and the fold is chevron like. The Phosphoria formation is the oldest unit exposed in the axial parts of the syncline; so the shape of the axial part of the fold in the older rocks is not known. In constructing the sections, however, it has been assumed that the bend is less abrupt in the beds beneath the Phosphoria. North of Deer Creek the syncline plunges 2° or 3° northward, the youngest beds exposed being the *Meekoceras* limestone at the north edge of the quadrangle, but south of Deer Creek no plunge is perceptible.

The thickness of the upper member of the Wells formation on the east limb, as determined from the outcrop width, is considerably greater than the thickness of the member elsewhere in the area, but the exposures are too poor to determine why. In the structure sections the excessive width is shown as the result of a terracelike flexure in the east limb, but it could be explained equally well by repetition by normal faults, such as the two that repeat part of the Phosphoria formation in the northern part of this area. Such faults are common in the east limb in the Stewart Flat quadrangle to the north (T. M. Cheney, oral communication, 1956).

The Webster syncline is truncated on the south by the transverse Sand Wash fault. South of this fault the Snowdrift anticline is bordered on the east by a broad wrinkled structural sag (pl. 2, section *F–F′*) that passes southward into a wide terrace (pl. 4 sections. *E–E′* and *F–F′*). The northern part of the Webster syncline is offset slightly by the Deer Creek fault.

BOULDER CREEK ANTICLINE

The Boulder Creek anticline is the easternmost fold of the upper plate; it has been thrust eastward along a complex thrust zone onto Jurassic rocks of the west limb of the Red Mountain syncline. The Brazer limestone is exposed in the center of the anticline for most of its length, but at the north and south ends the Wells formation is the oldest unit exposed.

The Boulder Creek anticline is divided into four segments by the transverse Deer Creek, Wells Canyon, and Sand Wash faults. Just north of Deer Creek the anticline is simple and overturned to the east, the overturned limb dipping about 70° to the west. About half a mile north of the fault a terrace, which broadens northward, is developed in the east limb between the axis of the fold and the overturned beds (pl. 2, section *A–A′*).

Between the Deer Creek and Wells Canyon faults most of the east limb of the anticline is covered, but exposures at the north end of the segment show that the east limb is overturned.

Most of the east limb is also covered in the segment between the Wells Canyon and Sand Wash faults, but exposures at the north end suggest that the anticline there is simple. Exposures on the north side of Clear Creek show that the fold in the southern part of the segment is complex, consisting of a broad flexure on the west, a gentle syncline, and a narrow overturned anticline on the east. A transverse anticline subparallel to the transverse faults is present at both the northern and southern ends of the segment.

South of the Sand Wash fault, the anticline extends for perhaps 2 miles as a gentle fold; but as the structural sag that separates it from the Snowdrift anticline disappears, the Boulder Creek anticline loses its identity.

TRANSVERSE FAULTS

The direction and amount of movement on the three transverse faults are difficult to determine because of the different shape of the folds on the two sides of each of the faults. Although not definitive, a graphic reconstruction of movement on the Deer Creek fault suggested that the vertical component was as large as, and probably larger than, the horizontal component; movement on the Sand Wash fault seems to have been largely vertical.

STRUCTURE OF THE LOWER PLATE OF THE MEADE OVERTHRUST

Of the formations exposed in the lower plate, the Nugget sandstone, most of the lower half of the Twin Creek limestone, the Preuss sandstone, and the Gannett group consist of competent rocks, and folds in these beds are concentric. However, the shaly limestone of member F of the Twin Creek limestone, 1,500 feet thick, is incompetent and locally has developed cleavage. Other incompetent units in the lower plate are the Wood shale tongue of the Ankareh formation and members A and C of the Twin Creek formation, but these are much thinner than member F of the Twin Creek and are intercalated with competent beds.

The Red Mountain syncline strikes a little east of north, paralleling the east edge of the thrust plate, but the other folds of the lower plate trend northwesterly. A few miles south of the area the strike changes to north-south, and this direction is maintained at least to the Utah border (Mansfield, 1927, pl. 9). The folds are more closely spaced than in the upper plate. Both the wave length and the amplitude of the folds decrease from west to east, perhaps reflecting the eastward thinning of the Jurassic section (pl. 4, section *H–H'*).

The westernmost fold in the mapped area is the Home Canyon anticline, a broad fold that brings the Nugget sandstone to the surface. According to Mansfield's map (1927, pl. 9), it is bordered on the west by the Bald Mountain syncline, which is west of the area of plate 3 of this report. Just south of Dunns Canyon the north end of the anticline is fragmented by normal faults, and north of the canyon it is concealed by the Salt Lake formation. The structure of the Jurassic and Upper Triassic beds on the Left Fork of Twin Creek seems to be that of a faulted overturned anticline (pl. 4, section *C–C'*) that may be the extension of the Home Canyon anticline. Northwest of Whiskey Flat, the uppermost Twin Creek limestone and the lowermost Preuss sandstone on the east limb of the Home Canyon anticline are in fault contact with Mississippian rocks of the overthrust plate; the thrust fault dips about 45° to the east and is approximately parallel to the dip of the Jurassic beds of the lower plate.

The next large fold exposed east of the Home Canyon anticline is a broad anticline that plunges about 10° to the northwest. It is separated from the Home Canyon anticline by several tight folds and two westward-dipping thrusts. The broad anticline is terminated on the northwest by a tear fault that runs into the thrusts on one end and abuts against the Meade overthrust on the other.

East of this broad anticline, Whiskey Flat is underlain by several small moderately tight folds in the Jurassic rocks (pl. 4, section *H–H'*). These folds terminate against the Meade overthrust and with one exception plunge northward.

The Red Mountain syncline strikes northward through the Gannett Hills nearly the entire length of the area. It contains rocks of the Gannett group in its axial portion, and they are the youngest beds known to be involved in the folding in the whole area. In the northern part of the area, the syncline is fairly tight and overturned eastward (pl. 2, section *C–C'*), but it broadens southward into a wide gentle fold. The axial plane, the trace of which is well defined in the Snowdrift Mountain quadrangle, dies out southward on the west side of Red Mountain; there the folding takes place about an axial plane just east of the area.

MEADE OVERTHRUST

The available evidence indicates that the Meade overthrust originated as a bedding plane shear within or at the base of the Madison limestone while the beds were still relatively flat and only slightly deformed (fig. 6*A*). The incipient thrust extended eastward at least 17 miles at this horizon, cut upward diagonally across strata of Mississippian to Middle Jurassic age, followed bedding at the top of the Twin Creek limestone eastward nearly 14 miles, and rose diagonally upward through the Jurassic and Cretaceous section. As the upper plate moved eastward, the beds comprising it were folded, and at the end of thrusting folded Mississippian beds rested on the comparatively undeformed Jurassic. The thrust surface itself was broadly folded, probably during the later stages of the same orogeny that produced the thrusting. The horizontal displacement of the eastern edge of the thrust as now exposed in the area is thought to have been about 18 or 20 miles.

DESCRIPTION OF THE SURFACE TRACE

The outcrop pattern of the fault trace indicates that, as shown by Richards and Mansfield (1912), the Meade overthrust is folded with a north-trending anticlinal flexure just east of the Left Fork of Twin Creek and a parallel broad synclinal flexure beneath Snowdrift Mountain. Both of these flexures plunge gently northward, as do the folds in the rocks of the upper plate. The trace is marked by a yellowish and reddish rehealed breccia zone several tens of feet wide wherever Mississippian rocks form the hanging wall. From its easternmost exposure to the south end of Snowdrift Mountain—that is, in all but the east limb of the synclinal flexure—the thrust separates Jurassic beds below from Mississippian beds above, and through most of this distance it separates beds near the Twin Creek-Preuss contact from the Madison limestone. The stratigraphic displacement is thus about 12,000 to 15,000 feet. East of the south end of Snowdrift Mountain, the stratigraphic displacement decreases rapidly by the occurrence of progressively younger beds in the upper plate adjacent to the thrust.

In the canyon of Preuss Creek, no stratigraphic displacement is apparent, but the following evidence indicates the presence of the fault. Overturned beds in the uppermost part of the Twin Creek limestone and the Preuss sandstone in the canyon of Preuss Creek have been folded into an isoclinal anticline with the youngest beds in the core, and the axial surface of this anticline has itself been flexed anticlinally. This peculiar fold, illustrated in figure 4, cannot be connected in any logical manner to the main body of the Preuss sandstone on the east or the Twin Creek limestone on the west, but must be separated from both by faults. This entire structural feature seems to be a slice of the thrust sheet that has been crumpled between two branches of the overthrust.

On the east side of Crow Creek the Meade overthrust is within Jurassic rocks, and the stratigraphic displacement is about 1,500 feet.

On the east limb of the synclinal flexure in the overthrust surface, several small thrusts that are laterally discontinuous splay upward

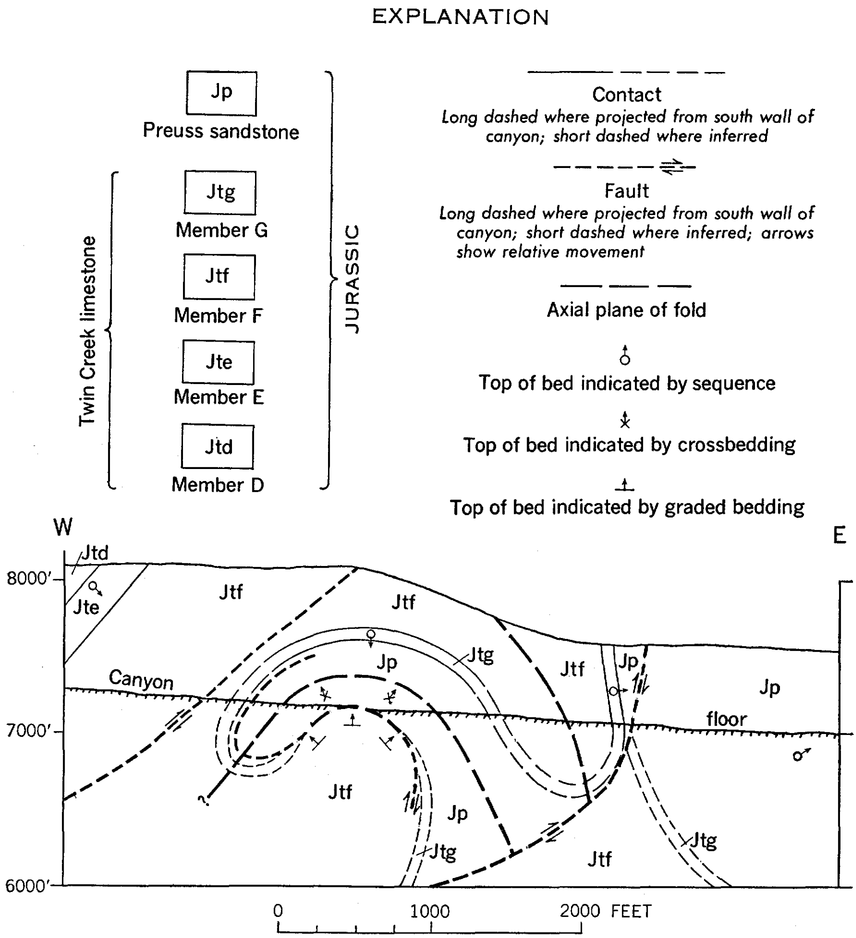

Figure 4.—Diagrammatic structure section of Upper Jurassic beds exposed on the north side of Preuss Creek.

from the Meade overthrust. These thrusts together with the sole thrust comprise the thrust zone in the valley of Crow Creek.

The overthrust surface truncates a complexly faulted overturned anticline in Jurassic rocks of the lower plate on the Left Fork of Twin Creek (pl. 3; pl. 4, sections *B-B'* and *C-C'*) and transects several small folds in the Twin Creek limestone and Preuss sandstone on the north side of Whiskey Flat (pl. 3), but elsewhere the beds beneath the thrust are about parallel to the thrust surface. On the west side of the Left Fork of Twin Creek the thrust dips about 20° NE., and the recorded dips in the beds of the lower plate are 10° to 35° in the same direction; at South Canyon, on the opposite side of the anticlinal flexure, the fault and the underlying beds dip about 45° E.; at the south end of Snowdrift Mountain both the fault and the underlying Jurassic beds dip moderately northward; and in the valley of Crow Creek the thrust and the adjacent beds dip steeply to the west, the beds being overturned.

Closer inspection of the anticline in the overthrust surface in the vicinity of lower Georgetown Canyon and the Left Fork of Twin Creek reveals some interesting relations, although they cannot all be interpreted with certainty because of the complex structure and the lack of exposures in some of the critical areas. The crest of the anticlinal flexure is fairly well exposed about three-fourths mile south of Summit View and one-half mile east of the Left Fork of Twin Creek. Here the attitude of the thrust changes abruptly from N. 50° E. 20° NW., on the west side, to N. 30° W. and nearly vertical on the east. Although a slight change in strike in the overlying Brazer limestone indicates some slight folding in the upper plate above the sharp bend in the overthrust surface, there is no fold in the upper plate commensurate with that in the fault surface. Three hypotheses have been considered to explain this relation.

First, it was taken as evidence that the anticlinal flexure in the overthrust surface was not the result of compressional folding; rather, the steeply dipping fault on the east side of the flexure was thought to have been originally a tear fault. This hypothesis was discarded because the supposed tear fault parallels beds in the lower plate rather than truncating them and because the fault also parallels the Home Canyon anticline.

The second hypothesis considered was that the steeply dipping fault was a younger normal fault that had dropped the overthrust plane downward on the east. A flexure in the overthrust surface would still be demanded, but it would be below the upper plate farther east, perhaps beneath the Dry Valley anticline. However, if the steeply dipping fault segment were a younger fault, it would have to extend both northward and southward into rocks of the upper plate, and no evidence of any such extension could be found. Exposures are such that small faults would be overlooked, but the continuity of the Summit View anticline west of Summit View is strong evidence that no fault of any considerable displacement extends northward into the upper plate.

The third hypothesis, now considered the most probable and illustrated in both map and sections (pls. 3, 4), is that the structural relations south of Summit View resulted from the development of the bend in the thrust surface at the site of an earlier small prethrust syncline in the Mississippian rocks. The possible stages in the development of the structure are illustrated in figure 5.

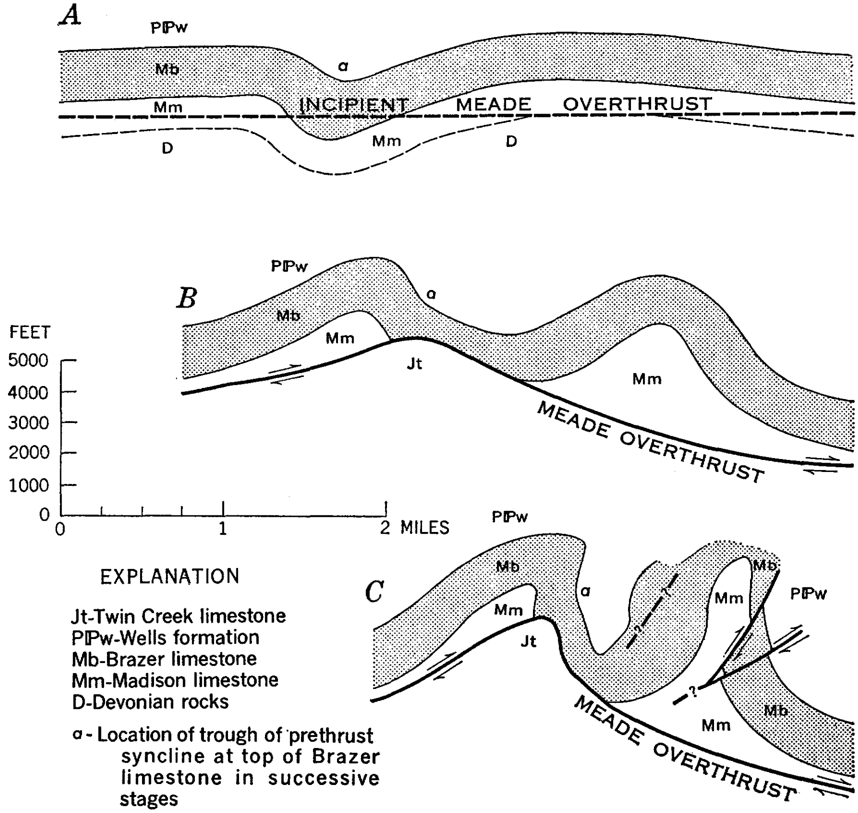

FIGURE 5.—Possible stages in the development of structure of the Meade overthrust plate on the Left Fork of Twin Creek. *A*, Initial stage; incipient thrust plane nearly parallel to bedding, but cuts small prethrust syncline. *B*, Thrusting nearly complete; upper plate has been thrust eastward onto Jurassic rocks; anticline develops in thrust surface at location of prethrust syncline in upper plste. *C*, Final structure.

The plate of Madison limestone on the divide between lower Georgetown Canyon and the Left Fork of Twin Creek is an erosional remnant of the overthrust sheet. Even though the general outline of the breccia zone that separates it on its north, west, and south sides from the Jurassic strata clearly shows the Madison to be in nearly horizontal contact with the younger rocks below, the margins of the breccia on the north side of the plate appear steeply dipping, perhaps as a result of some later movement. The outcrop of the breccia zone itself is as much as 1,200 feet wide, largely because of the very low dip of the fault. The anticline in the overthrust surface is partly exposed on the north margin of the overthrust plate where the dip of the fault changes eastward from nearly horizontal to about 35° E. before disappearing beneath alluvium. The flexure is thus much more gentle than it is to the north.

The narrow east-trending strip of Madison limestone on the south side of Georgetown Canyon is evidently part of the upper plate that was bent downward along an east-west axis, broken on the south by the east-strking high-angle Southside fault, and breached by the erosion of Georgetown Canyon. Judging by the difference in elevation of the Meade overthrust trace on the north and south sides of Georgetown Canyon, the vertical displacement of the Southside fault is a little more than 1,000 feet. If the fault extends eastward into the Mississippian rocks of Snowdrift Mountain, it must die out rapidly, for there are no mappable offsets in the northwest-trending contacts and faults to the east.

The nature of the north-striking and west-dipping fault that transects the west end of the Southside fault is not definitely known. Mansfield (1927, pl. 9) mapped it as part of the Bannock overthrust, and the similarity of the breccia zone along the fault to that which elsewhere marks the trace of the Meade overthrust, the anticlinal structure of the Madison limestone west of the fault, and indication of drag that may be seen in the Madison near the fault on the north side of Georgetown Canyon all suggest that the fault is indeed a thrust. However, the relative position of the phosphatic horizon in the slice of Madison west of the fault and in the overthrust plate between the forks of Twin Creek suggest that the fault is normal and that the Mississippian rocks west of the fault are part of the overthrust plate that has been dropped downward. In either event, the absence of a transverse fault in the Madison limestone near the mouth of the canyon indicates that the Southside fault developed contemporaneously with the fault that transects its western end.

STRUCTURE BENEATH THE OVERTHRUST PLATE

The trace of the Meade overthrust separates the Madison limestone from either the uppermost part of the Twin Creek limestone or the lowermost part of the Preuss sandstone nearly the whole distance from the west side of the Left Fork of Twin Creek to the south end of Snowdrift Mountain. In this distance the trace crosses structures in the upper plate from the Dairy syncline to the east limb of Snowdrift anticline. These relations suggest that the Meade overthrust separates the same two horizons—the Madison limestone and the uppermost Twin Creek limestone as far eastward as the east limb of Snowdrift anticline.

Evidence suggests that the Twin Creek limestone is nearly horizontal beneath the overthrust sheet. In section *H-H′* of plate 4, which is just south of the overthrust plate, the length of the contact between members C and D of the Twin Creek limestone from its outcrop on the west to a point along strike of the east edge of the thrust plate is

5.7 miles; in section *E-E′* of the same figure the length of the same contact from its western outcrop to beneath the thrust zone, assuming it to be nearly horizontal and parallel to the thrust as shown in the section, is about 6.9 miles. By deepening the westernmost syncline in section *H-H′* a permissible amount, the two lengths can be made comparable. This suggests that the length of the contact as shown in section *E-E′* is about correct and thus that the Jurassic rocks beneath the upper plate are not highly folded.

RELATIONS OF THRUSTING TO FOLDING

The relations, discussed above, that indicate that the thrust surface is between the Madison limestone and the top of the Twin Creek limestone beneath most of the upper plate, and the evidence that the Twin Creek limestone is nearly horizontal and not highly deformed beneath the thrust, suggest that the fault was initiated as a bedding thrust. More precisely, it probably originated as two bedding thrusts, one at the base of the Madison limestone and a more easterly one at the top of the Twin Creek limestone, the two thrusts being joined by a fault that crossed the intervening beds (fig. 6A). If so, the overthrust was initiated before most of the folding, and the folds of the upper plate and their associated thrusts and tear faults must have developed largely during the thrusting. Confirming evidence may be obtained by calculating the depth of folded rock in the upper plate (that is, the depth to the plane of basal detachment) from the amplitude and wave length of the folds (de Sitter, 1956, p. 190). The only fold in the upper plate sufficiently well exposed for such a calculation is the Snowdrift anticline in section *C-C′* of plate 4. Taking the base of the Wells formation to delineate the fold and the limits of the anticline as being the axis of the Webster syncline and the axis of the eastern minor syncline in the complex Georgetown syncline, the cross sectional area of Snowdrift anticline is 1.17×10^6 square feet, the length in the fold of the delineating horizon is 1.17×10^4 feet, and the distance across the base of the fold is 6.5×10^3 feet. From these data, the depth to the plane of basal detachment is determined as 2,250 feet below the base of the fold as defined by the base of the Wells, or 2,800 feet above sea level. If the Snowdrift anticline is taken as extending from the axis of the Webster syncline to the axis of the western minor syncline in Georgetown syncline, the depth of folding is calculated as 2,200 feet above sea level. These figures are only several hundred feet below the position of the thrust as placed in the structure section on the assumption that the fault must be somewhere within the Madison limestone. The calculations thus afford some supporting evidence that the thrust surface beneath the upper plate is within, or at least not far below, the Madison limestone; that the position of the thrust determined the amplitude and wave length of the folds of the upper plate; and thus that the folds developed during and not before the thrusting.

The Jurassic rocks of Whiskey Flat were apparently folded after most of the thrusting and during or after folding of the overthrust surface. This is indicated by the manner in which the fold axes in the lower plate swing around both sides of the south end of the thrust sheet as if deflected by it and by evidence discussed previously that the Twin Creek limestone is not folded beneath the thrust. The small folds on the east side of Whiskey Flat that abut against the fault probably resulted from crumpling of the Jurassic beds in front of the moving overthrust sheet. A small anticline must have existed before thrusting in the Jurassic rocks on the Left Fork of Twin Creek where the thrust truncates the crest of an overturned anticline, and the explanation offered in figure 5 for the structural relations in the upper plate in the same vicinity presupposes a small syncline in the Mississippian strata before thrusting.

DIRECTION OF MOVEMENT

With respect to its present orientation, the relative movement of the overthrust within the area was from the west-northwest. The principal evidence for this is the strike of the folds of the upper plate, but the following facies relations within the Twin Creek limestone also indicate that the movement must have been from well north of west. Members D and E of the Twin Creek limestone on the overthrust sheet near Preuss Creek are considerably thinner and contain much less sandstone than the same members as exposed in Whiskey Flat and in the southern part of the Home Canyon anticline. In the westernmost outcrop belt of the Twin Creek both members thin and become less sandy to the north so that as exposed near Hawks Roost they are more similar to the members as exposed at Preuss Creek.

The Meade overthrust is one of a number of faults that comprise a large arcuate zone of overthrusts in eastern Wyoming, southeastern Idaho, and adjacent States (Rubey, 1951). At the latitude of the Georgetown Canyon-Snowdrift Mountain area, the overthrust belt and most of the individual thrusts within it trend north-south, suggesting that the relative movement of the upper plates of most of the thrusts was from west to east. The movement in the Meade overthrust may also originally have been from west to east, the upper plate having been rotated clockwise to its present position during the later stages of compression. Such rotation might have resulted from crowding within the arc of the overthrust belt.

MAGNITUDE OF THRUSTING

Between Montpelier and Bennington (southwest of the mapped area), a west-dipping thrust plate of Madison limestone rests on beds as old as the Phosphoria formation that are part of the west limb of the Bald Mountain syncline which parallels the Home Canyon anticline (Mansfield, 1927, pl. 9). Inasmuch as this fault is approximately along strike with, and involves the same strata in, the upper plate as the Meade overthrust in its westernmost exposures, it is most probably an extension of the Meade overthrust. Therefore, rocks of the upper plate could not have been derived from areas east of the Bald Mountain syncline. The fact that the thrust cuts beds as old as Permian in the Bald Mountain syncline suggests that the thrust sheet could have been derived a short distance to the west. From Mansfield's (1927, pl. 12) section *W–W′*, it can be seen that the horizontal component of displacement in an east-west direction need only have been 7 miles.

Consideration of the relations shown in figure 6, however, suggests that the movement in the area of this report was more than twice this amount. Figure 6*B* represents the structure, both across the area and for some distance to the west, after thrusting but before normal faulting. The structure shown for most of the Preuss Range has been taken from section *C–C′* of plate 4, but because of the disruptions of the folds by range-front faults in the area of this report, the structure in the westernmost part of the range has been deduced from that in the Johnson Creek quadrangle (Gulbrandsen and others, 1956, pl. 1). The following folds are shown in the overthrust plate west of the Left Fork of Twin Creek: the Dairy syncline, an overturned anticline that was later faulted out, the Trail syncline projected southward from the Johnson Creek quadrangle, and an overturned anticline that is now disrupted by range-front faults. Evidence for the existence of the westernmost anticline is given by overturned beds along the west front of the Aspen Range, shown in both the Johnson Creek quadrangle and plate 6 of this report. The west limb of this anticline has been extended across Bear River valley in the most direct way permitted by exposures in the valley and in the Bear River Range (Mansfield, 1927, pl. 6). As measured from figure 6*B*, the horizontal offset of the Madison limestone is about 16 miles. Corrections for curvature of the fault plane indicate that the Mississippian limestone in the easternmost part of the upper plate must have moved from at least 18 or 20 miles west of its present position, and compensation for folding in the upper plate indicates that the western part of the thrust sheet moved about 24 miles. If the thrusting was from the northwest, the displacements would be somewhat greater.

In figure 6*B*, the Paris thrust on the west side of Bear River valley is shown as passing at depth into a bedding thrust between the Brigham quartzite and the Precambrian. The incipient Meade over-

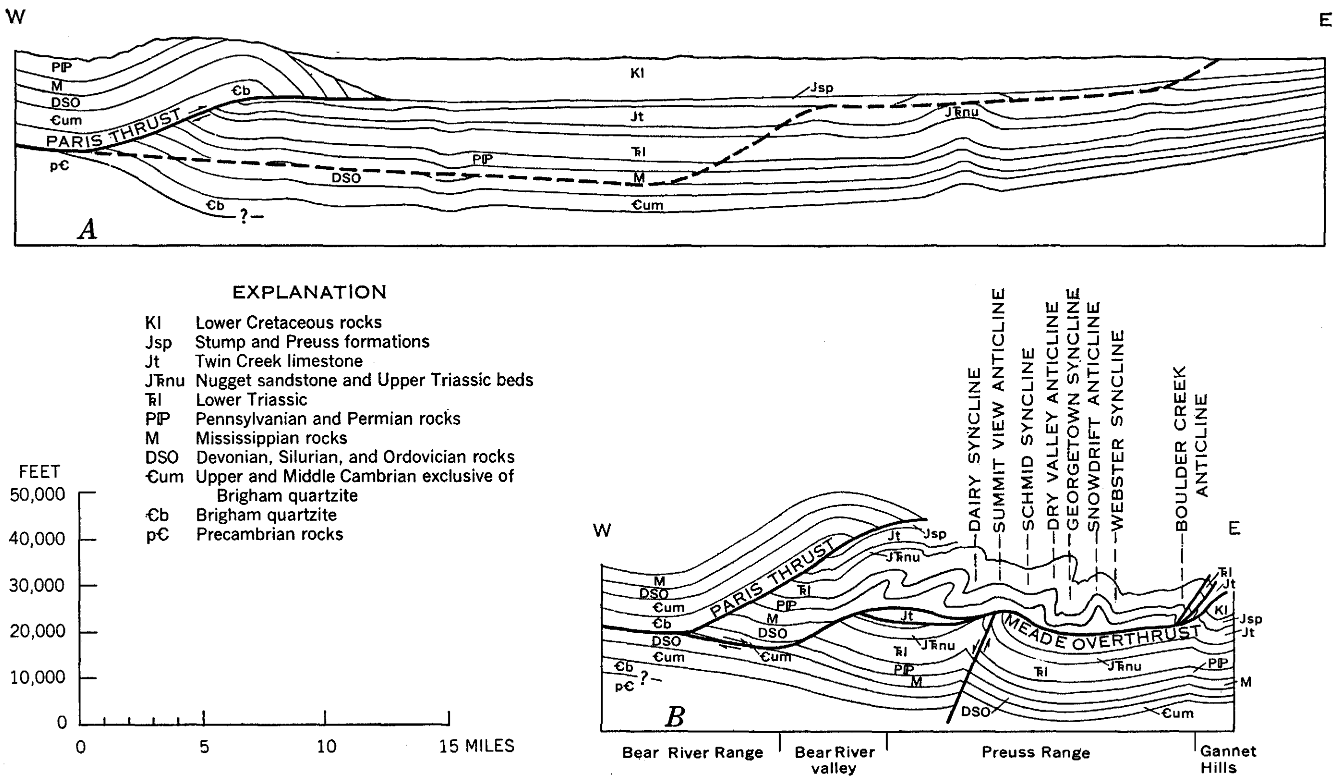

FIGURE 6.—Hypothetical structure sections across southeastern Idaho just before and just after development of the Meade overthrust. Sections at latitude of line *C–C′*, plate 3. *A*, Just before development of Meade overthrust; dashed line is initial position of thrust. *B*, Just after development of Meade overthrust and before block faulting; topographic surface is not indicated, and beds younger than the Twin Creek limestone are shown only beneath the Paris thrust.

thrust is shown as merging into this bedding thrust; that is, movement is assumed to have recurred on a sole thrust beneath the Bear River Range during Meade Peak thrusting. Relations illustrated at this depth are obviously interpretive. This particular interpretation has been chosen because the base of the Cambrian is known to be a favorable horizon for thrusting; however, the Meade overthrust may well offset the thrust on the east side of the valley and pass into a younger horizon, perhaps one of the Cambrian shales, or may extend well below the older fault. The Paris thrust is shown as older than the Meade overthrust on the basis of evidence discussed by Armstrong and Cressman (1963).

TIME OF THRUSTING

From relations within the Georgetown Canyon-Snowdrift Mountain quadrangle the Meade overthrust can be dated only very broadly as post-Gannett group (Early Cretaceous) and pre-Salt Lake formation (Pliocene). From adjacent areas the thrusting and folding can be dated more closely as post-Wayan (latest Early Cretaceous) and pre-Wasatch (Eocene). Rubey and Hubbert (1959, p. 190–192) conclude that in the western Wyoming-southeastern Idaho belt the thrusts developed first on the west and were followed successively by those on the east, and they imply that the Meade overthrust (termed by them the "east branch of the Bannock") was initiated at the end of the Early Cretaceous time.

DEFORMATION OF THE THRUST SURFACE

Mansfield (1927, p. 203) thought that the thrust surface was probably folded in Pliocene time, although he recognized the possibility that it might have occurred much earlier. Any Pliocene compression sufficiently strong to fold the thrust surface and to fold the Jurassic beds in Whiskey Flat should be evidenced by deformation of the Salt Lake formation, but at least in this area the Salt Lake is neither thrusted nor folded. The steep dips in the Salt Lake have resulted from the rotation of normal fault blocks. Thus, although neither the thrusting nor the deformation of the fault surface can be dated closely, there is no evidence of more than one major compressive period. Therefore, the fault surface probably was folded in the later stages of the same orogeny that produced the thrust and the folds.

GRAVITATIONAL SLIDING AS THE MECHANISM OF THRUSTING

Rubey and Hubbert (1959) have discussed in an important paper the possible mechanics of thrusting in the overthrust belt of southeastern Idaho and western Wyoming with special emphasis on the role of abnormally high fluid pressures. Most of their discussion is of that part of the belt east of the Georgetown Canyon-Snowdrift Mountain area, and they mention the Meade thrust (the East Branch of the Bannock in their paper) only incidentally. However, the Meade overthrust shares many of the attributes of the thrusts that Rubey and Hubbert discuss in more detail: It parallels bedding for considerable distances and probably originated in large part as a bedding thrust; making no allowance for folding, the displacement is probably at least 18 to 20 miles (compared with estimates of 10 to 15 miles for the more easterly thrusts); the original thickness of strata in the overthrust sheet was probably about 25,000 feet; the rocks are unmetamorphosed; and the breccia zone is thin. As in the area to the east, conditions were favorable for the development of abnormal fluid pressures; the total sedimentary section is thick and contains both highly compactible and less compactible beds, and the rate of sedimentation was rapid in the later stages of the depositional history.

In discussing the motive force of the thrusting, Rubey and Hubbert (1959, p. 196) state that if the fluid pressure-overburden ratio was 0.91 to 0.92, not improbable values, gravitational sliding could occur on slopes of only 2½° to 3°. Figure 6*A* was reconstructed from figure 6*B* to determine whether or not a 2½- to 3-degree eastward slope could have existed at the base of the Mississippian strata before thrusting. The lengths of the strata in both upper and lower plates were measured, and the formations then restored to their assumed positions before Meade thrusting. According to Rubey (*in* Moritz, 1953, p. 66), the rate of increase in grain size in the Gannett group indicates that the source of the sediments must have been very close to the type area in the Gannett Hills, perhaps 25 miles or less to the west. Armstrong and Cressman (1963) have concluded that the uplift from which these sediments were derived probably resulted from thrusting and that the Paris thrust on the west side of Bear River valley developed in the Late Jurassic or Early Cretaceous; this thrust cuts the Lower Triassic, the youngest beds exposed in its lower plate, and is assumed to have originally continued upward across the section to the Late Jurassic surface. Pebbles of Swan Peak formation, present in small amounts in the Gannett group, indicate that the headwaters of streams depositing the Gannett had at least locally dissected the source as deep as the Ordovician. The probable gradient of the streams depositing the Lower Cretaceous sediments has been taken as similar to many modern streams heading in mountainous areas; changing the gradient, reasonable amounts would have only a small effect on the diagram. The position of the earlier thrust, the depth of erosion of the source area, and the stream gradient together determine, rather imprecisely, the thickness of the Lower Cretaceous in the western part of the figure. The thickness of rocks of Mississippian to Early Cretaceous age in the western part of the Meade thrust sheet is established fairly closely by the thickness of the section in the Aspen Range; the thickness of the unexposed part of the section in the Gannett Hills has been interpolated by comparing thicknesses in the western part of the mapped area and, for the lower Paleozoic in the Bear River Range, in the Bedford quadrangle in westernmost Wyoming (Rubey, 1958). The remaining unknown that has not been estimated is the thickness of Upper Triassic and Jurassic rocks in the western part of the section. If the Upper Triassic and Jurassic thin to the west of the Aspen Range, as shown in figure 6*A*, then the 3-degree eastward slope required for gravitational sliding with the assumed fluid pressure-overburden ratio could have existed at the base of the Mississippian. This does not, of course, establish that gravitational sliding was the mechanism, but merely indicates that data within the area are not inconsistent with the hypothesis. No evidence as to whether or not the Upper Triassic and Jurassic rocks thin in the required manner has been found.

As stated by Rubey and Hubbert (1959, p. 197), thrusts caused solely by gravitational sliding would leave wide gaps in their rear. They discuss as one possibility that the broad intermontane valleys may have originated as gaps where the thrust plates pulled away. That Bear River valley could not be such a gap is shown first by the Lower Triassic rocks of simple structure that are exposed in parts of the valley (Mansfield, 1927, pls. 6, 9) and second by the fact that the valley is only half as wide as that part of the thrust sheet exposed in the Preuss Range. Rubey and Hubbert also very tentatively suggest that thrusts may have slid eastward off the Bear River Range. This possibility cannot be evaluated properly until more is known of the geology of the Bear River Range itself, but the hypothesis would seem to place the source of the Lower Cretaceous sediments too far west.

In summary, the available facts indicate that the Meade overthrust could have resulted from gravitational sliding, but the ranges west of the area of the report must be mapped to determine whether or not the gaps behind the plates demanded by the gravitational sliding hypothesis actually exist.

NORMAL FAULTS

The Aspen Range, the Preuss Range, and Schmid Ridge are all fault-block mountains uplifted along normal faults that mark their western margins, and most of the normal faults in the Georgetown Canyon-Snowdrift Mountain area are associated with this range-front faulting. The west front of the Aspen Range in particular has been broken into many small blocks by a complex system of normal faults. Most of these faults parallel the strike of the beds and are downthrown on the west, but a few are downthrown on the east and some are transverse to the strike of the beds. The faults are both large and small; of the faults intersected by section *C–C′* of plate 4, three have dip-slip displacements of 1,000 feet or more, yet the displacement of other mapped faults in the range-front zone is only 100 feet or so. Where thin, distinctive mapping units are involved, numerous small faults have been detected, suggesting that in thicker units, such as the Madison and Brazer limestones, the faulting is more complex than shown. Some of the faults undoubtedly die out laterally or terminate against transverse faults, but faults that trend into the Salt Lake formation are probably more continuous than shown. The normal fault at the mouth of Dunns Canyon may well be the same fault as that which separates the Brazer limestone from the Salt Lake formation just east of lower Pine Canyon, but even though these faults are both post-Salt Lake in age, they cannot be traced through the soft, poorly exposed, and uniform beds of that formation. The dips of only a few of the faults can be determined; they range from 50° to 68°. The observable dip-slip displacement across the normal fault zone in section *C–C′* of plate 4 is 7,000 feet; assuming a dip of 50° on the faults, the total throw across the zone is about 5,500 feet. A mile to the south (pl. 4, section *D–D′*), the total dip-slip displacement may be as much as 12,000 feet with a throw of 9,000 feet, but the relations are not so clear as those in section *C–C′*. Between the forks of Twin Creek, where the conglomerate unit of the Salt Lake formation has been faulted against the older rocks, the conglomerate beds dip eastward 15° or 20° toward the fault, and west of the area in the hills north of Georgetown the tuffaceous member of the Salt Lake dips as much as 60° to the east.

The zone of normal faults fronting the Aspen Range about 7 miles northwest of the mouth of Georgetown Canyon is illustrated on plate 6. Here, too, most faults are downthrown on the west, although a few are downthrown on the east. Those fault planes that can be measured dip 48° to 55°. The Salt Lake formation that is faulted against the Wells formation consists of the upper conglomerate unit and dips about 20° eastward into the range-front fault. Several deposits of calcareous tufa and sulfur-depositing springs are on or near the faults.

In post-Salt Lake time the movement on the westernmost fault on plate 6 was clearly normal, west side down. However, north of Swan Lake Gulch the same fault separates the Brazer limestone and the lower member of the Wells formation on the west from the upper member of the Wells on the east. These relations suggest that movement on the fault in pre-Salt Lake time may have been the reverse of that in post-Salt Lake time; that is, normal movement may have occurred along an earlier thrust fault. However, the observed relations could have resulted entirely from normal faulting. Before disruption of the Laramide structure by normal faults, an overturned anticline existed along the front of what is now the Aspen Range. This is evidenced by overturned beds along the front of the range in the Johnson Creek quadrangle (Gulbrandsen and others, 1956, pl. 1). The wedge of rocks just north of Swan Lake Gulch that are overturned 180° and in which the Meade Peak member of the Phosphoria formation is less than 100 feet thick is probably part of the attenuated overturned limb of the anticline that has been faulted downward. The Brazer and lower part of the Wells on the west side of the normal fault along the range front might have been derived from a higher part of the overturned fold and thus have been emplaced wholly by normal faulting.

Normal faults that border Slug Valley extend into the area but die out before reaching Summit View. The normal fault on the west side of Dry Ridge is an extension of the normal fault that has been mapped along the base of Dry Ridge in the Dry Valley quadrangle (Cressman and Gulbrandsen, 1955, p. 265, pl. 27).

Those normal faults that offset the Salt Lake formation can be dated as Pliocene or younger. These faults may have been active before, as well as after, deposition of the Salt Lake, but any such earlier movement could not be detected. Normal faults within the range can be dated directly only as postfolding, but they seem to be part of the same fault system as those faults that can be dated, and they, too, were probably formed mostly in later Tertiary or early Quaternary time. The Salt Lake formation now occurs only in structural valleys, and judging from the occurrence of the Salt Lake on upper Slug Creek and upper Crow Creek, at least some of these valleys were in existence at the beginning of Salt Lake deposition. That the source of pebbles was local also indicates that the valleys were formed early. No evidence that these valleys were originally defined by warping rather than by faulting has been found; the normal-fault systems, therefore, probably were initiated at least by Miocene time and perhaps as early as the Oligocene.

Fault relations in upper Slug Valley rather equivocally suggest several periods of normal faulting. The fault on the east side of the valley is clearly later than the Salt Lake formation and probably later than the basalt. The fault on the west side of the large basalt mass, however, seems to be prebasalt in age. Faults north of the basalt are shown as being older than the flow, but exposures there are so poor that the actual relations cannot be determined.

The normal fault that separates the Nugget and Salt Lake formations at the mouth of Dunns Canyon is marked by a recent fault scarp. A fault scarp can be seen along the front of the Preuss Range most of the way from Georgetown to Montpelier and is pictured by Mansfield (1927, pl. 40*A*). As shown by Mansfield's photograph, the scarp is accompanied by a depression that near Montpelier is a convenient location for the town dump. There is no historic record of movement along the range-front fault, but the scarp is comparable in freshness to some formed in Nevada early in this century and is probably not older than a few hundred years.

PHYSIOGRAPHY

GENERAL FEATURES

The Peale Mountains are generally considered part of the Middle Rocky Mountain physiographic province (Fenneman, 1931). Most of the component ranges are tilted fault blocks similar to those of the Basin and Range province. But the general aspect of the country differs from that of the Basin and Range province in that the valleys are more narrow and dissected, both ranges and valleys are less dry and more vegetated, and ranges comprise a larger proportion of the area. Although the major ranges and valleys are outlined by normal faults, most of the local relief is erosional. Other than for the small Recent scarps, fault scarps have been so eroded as to be unrecognizable. The Georgetown Canyon-Snowdrift Mountain area has been dissected to the stage of early maturity. Trunk streams are at grade, at least in their lower parts, divides are sharp, and the drainage is well adjusted to structure.

It is characteristic of the area that relief is closely adjusted to structure (Richards and Mansfield, 1914, p. 12). Crow Creek valley is located in a thrust zone where rocks have been shattered by faulting; Georgetown Canyon has been excavated partly in soft Lower Triassic rocks and partly in the nonresistant Wells formation in the center of the Georgetown syncline; Snowdrift Mountain and Dry Ridge are held up by the resistant Brazer limestone exposed along the area of anticline; and Whiskey Flat is developed on the very soft shaly limestone of member F of the Twin Creek formation. On a different scale, the uppermost beds of the Brazer, the Rex chert member of the Phosphoria, and the contact between the lower and upper member of the Dinwoody formation can all be traced along strike for miles in the Snowdrift Mountain quadrangle by their distinctive appearance on the topographic map. This dependence of topography on structure and the mature dissection of the area make the detection of older surfaces of erosion difficult and highly subjective.

EROSION SURFACES

Mansfield (1927, p. 11–19) discussed the physiographic development of southeastern Idaho in considerable detail and distinguished a number of erosion surfaces. These were, from oldest to youngest (1) a pre-Wasatch surface, (2) the Snowdrift peneplain, (3) the Tygee (pre-Salt Lake) surface, (4) the Gannett surface, (5) the Elk Valley surface, and (6) the Dry Fork surface. Only the Snowdrift surface was described as a peneplain; the others were either surfaces of considerable relief or were rock terraces along the major canyons and valleys. Remnants of all these surfaces were identified in the area of this report (Mansfield, 1927, pls. 8, 16, fig. 6). Mansfield's principal analytical method was the construction of profiles along divides both parallel to, and at right angles to, the trend of the main ridges, and to a large extent his conclusions were based on the evenness and accordance of divides.

I can add little to Mansfield's exhaustive discussion of surfaces within the ranges other than to caution that the identification of surfaces by even and accordant divides is hazardous in an area such as southeastern Idaho that is maturely dissected and consists of rocks of widely differing resistance. Remnants of cyclic surfaces probably exist in the region, but their identification and correlation are by no means obvious. The problems in interpretation may be illustrated by consideration of the Snowdrift peneplain. The principal evidence within the area for the peneplain is the relatively even crestline of Snowdrift Mountain that stands at altitudes of 9,200 to nearly 10,000 feet. The even crestline is indeed suggestive of peneplanation when seen from a distance (Mansfield, 1927, fig. 6); yet, the divide is sharp and is developed on the most resistant formation in the section, the Brazer limestone, that crops out in the core of the Snowdrift anticline. There seems no compelling reason to interpret the ridge crest as a peneplain remnant rather than merely the result of structurally controlled differential erosion.

The oldest and highest surface in the area that I believe can be identified with confidence is in The Hole at an altitude of 7,500 and 7,700 feet. This surface is mantled with gravel and is thought to be a reentrant of a pediment. Gravel deposits at an altitude of 7,000–7,300 feet on top of the beveled Salt Lake formation between the forks of Twin Creek and between Dunns Canyon and Georgetown Canyon are also thought to be on pediment remnants. The evidence that these surfaces are pediment remnants is principally their position along the range front. Standing on the Salt Lake formation on the divide between the forks of Twin Creek and looking westward, it takes only a little imagination to reconstruct from the summits of hills of the Salt Lake formation in Bear River valley a broad surface sloping westward to the center of the valley and then rising to merge with an erosion level in the Bear River Range. The pediment, if such it was, was developed nearly entirely on the Salt Lake formation, and only locally did it bevel the older rock of the ranges. Shoulders on the sides of both forks of Twin Creek suggest that the surface continued for a distance into the range up these two canyons, that on the Left Fork possibly extending to Summit View; but most of the shoulders are developed on hard stratigraphic units.

The progressive dissection of the pediment is recorded by gravels that cap hill summits at many different altitudes in Bear River valley. These gravels are similar in content to those on the highest levels, and, unlike the fragments in the Salt Lake formation, many are of lower Paleozoic quartzites. The gravels are probably the remnants of deposits on a series of pediments developed at successively different levels, and subsequently so dissected that their reconstruction is not possible without more detailed work than was possible in this study.

The lowest and youngest erosion surfaces in the area are strath terraces along the major streams that have been previously mentioned (p. 60).

GEOLOGIC HISTORY

GEOSYNCLINAL PHASE

The Georgetown Canyon-Snowdrift Mountain area, which is a short distance west of the Wasatch line of Kay (1951, p. 14), was the site of geosynclinal deposition from the Cambrian through Early Cretaceous. From comparison with neighboring areas, it can be assumed confidently that geosynclinal subsidence in the area from the Early Cambrian to the Mississippian was from 5,000 to 10,000 feet. Rocks within the area record subsidence of nearly 25,000 feet from earliest Mississippian to the end of the geosynclinal phase in middle Cretaceous time. In figure 7 the depth of burial of the base of the Madison limestone, the oldest horizon exposed in the area, has been plotted against geologic time in the same manner as was done by Rubey and Hubbert (1959, p. 192) for various parts of the overthrust belt. The diagram illustrates for this area relations similar to those shown by Rubey and Hubbert; the rate of burial was somewhat irregular, but it generally increased in the later stages of the geosyncline.

The geosynclinal phase can be divided into two parts. From the Cambrian through the Early Triassic, the area was on the east flank of the Cordilleran geosyncline and received most of its terrigenous material from the craton to the east. As in other miogeosynclines, most sediments were deposited in marine waters but above wave base. Beginning in Late Triassic or Early Jurassic time, terrigenous material was derived in large part from an uplift in the geosyncline west of the area; thus, the area formed the western part of an exogeosyncline (Kay, 1951, p. 17). The exogeosynclinal character was more marked in the Early Cretaceous, when the source land was but a few tens of miles to the west and the sediments continental, than it was in the Jurassic, when the source was more distant, the clastics finer grained, and the environment marine.

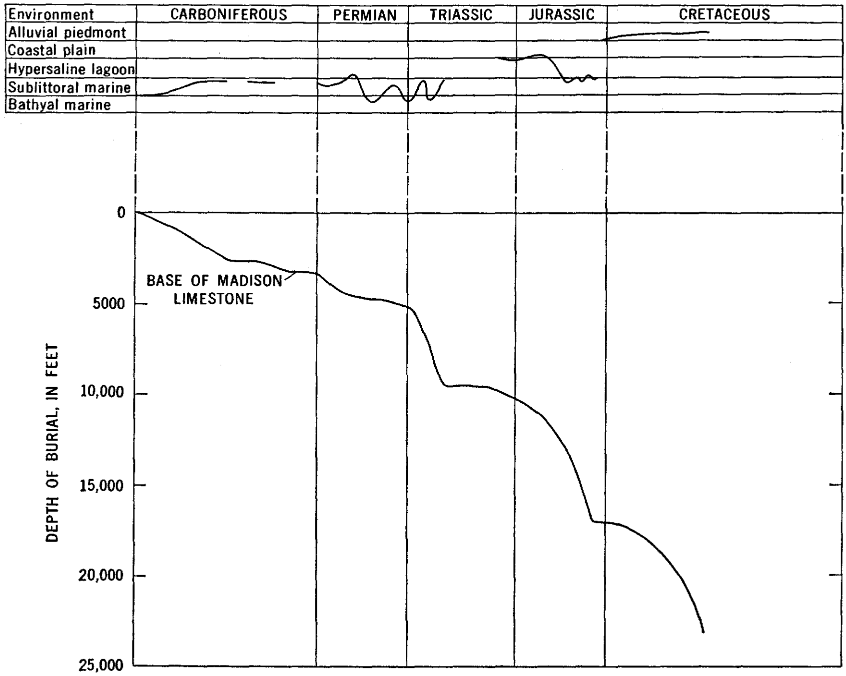

FIGURE 7.—Depositional environment and subsidence during the geosynclinal phase. Lengths of geologic periods are shown as proportional to their duration according to the Holmes "B" time scale.

The earliest event recorded within the Peale Mountains is the deposition of the Madison limestone of Mississippian age. The dark color, fine grain size, and thin and even bedding of most of the Madison indicates that it was deposited mostly in tranquil water, but the light-colored crinoidal limestone indicates that the water was periodically agitated and aerated. The formation is therefore shown in figure 7 as having accumulated mostly in the outer sublittoral marine environment. The overlying Brazer limestone was deposited mostly above wave base in the inner sublittoral environment as clearly shown by the abundance of light-colored crinoidal limestone and well-sorted fine- and medium-grained sandstone. A similar depositional environment is indicated for the lower member of the Wells formation by beds of oolitic and highly fossiliferous limestone.

The hiatus in the Wells formation between the Middle Pennsylvanian and the Permian is not marked by an unconformity, by any evidence of subaerial exposure, or even by a distinct change in rock type. Perhaps during this lapse in deposition the area was part of a marine profile of equilibrium, such as that described by Keulegan and Krumbein (1949), developed during a pause in subsidence.

Renewed subsidence in Early Permian time resulted in shallow marine deposition of the sandstone and carbonate rock of the upper member of the Wells formation. Red beds in the uppermost part of the Wells were evidently laid down in a hypersaline lagoon, as attested by anhydrite cement observed in the Standard Oil Co. of California's Sheep Creek well 1 on Bear Lake Plateau (Zeni, 1953).

Much of the black shale of the Meade Peak phosphatic shale member of the Phosphoria formation was clearly deposited below wave base in the bathyal marine environment. McKelvey (*in* McKelvey and others, 1959, p. 23–25) has discussed the origin of this unit and concludes that the nondeposition or very slow deposition of diluting carbonates and clastics is a prerequisite to the formation of a phosphorite unit. This conclusion suggests that the deeper marine waters of Meade Peak time may have resulted as much from decreased rate of deposition as from an increased rate of subsidence.

The Rex chert member of the Phosphoria resulted from the accumulation of siliceous skeletal remains and the partial diagenetic reorganization of the silica (McKelvey and others, 1959, p. 27). Spicules from the sponge class Demospongia are abundant in the Rex. Sponges of this class are mostly marine and thrive at depths of less than a few tens of meters (M. W. de Laubenfels, oral communication, 1957). Depostion in the sublittoral environment is indicated not only by the Demospongia fauna but also by lenses of crinoidal limestone within the member. The cherty shale member probably formed at a water depth intermediate between that of Meade Peak time and Rex time.

Kummel (1957) has discussed the origin of the Lower Triassic of southeastern Idaho in considerable detail. It suffices here to note that the light-colored bioclastic limestone members are clearly shelf deposits, the shale members accumulated below wave base in the bathyal environment, and slump structure in the black-weathering silty limestone and calcareous siltstone suggest deposition on an intermediate slope. In its gross aspect the Lower Triassic records two major regressions and an intermediate transgression. The Timothy sandstone at the top of the Thaynes is a near-shore deposit that marks the culmination of the last regression (Kummel, 1957, p. 466).

Because the ages of the units assigned in this paper to the Upper Triassic are not known accurately, it cannot be determined whether or not there was a significant break in deposition within the sequence between the Thaynes formation and the Twin Creek limestone. However, the strata were either deposited subaerially or in very shallow water, and there may well have been periods of nondeposition or even erosion.

The environments of deposition of the Jurassic formations have been discussed by Imlay (1957). Briefly, the Jurassic of southeastern Idaho records three marine transgressions. The earliest transgression culminated in the deposition of member C of the Twin Creek in Bajocian time, the second in deposition of member F in early Callovian time, and the third in deposition of the Stump sandstone in the Oxfordian. The Oxfordian transgression was less marked than the earlier two. The regressive stages are recorded by the aeolian (and in this area perhaps partly littoral) Nugget sandstone and the lagoonal member A of the Twin Creek, by the red beds of member D, and by the lagoonal red beds and salines of the Preuss sandstone. According to Imlay (1957, p. 483–485), the Jurassic limestones were deposited in very shallow water as shown by an abundant molluscan fauna, the frequent occurrence of *Gryphaeu* and *Ostrea*, and by oolitic beds. The lithographic limestone of members C, E, and F were probably deposited in deeper water than the other Jurassic sediments as shown both by their texture and position in the sedimentary cycle, but even these are apparently sublittoral deposits (Imlay, 1957, p. 484). Tuffaceous material within the Jurassic rocks may reflect increased volcanism, but it may also have been the result of the disruption of prevailing northeasterly winds by the Jurassic orogenic land rising west of the area.

The record of the geosynclinal phase ends with the deposition of the Gannett group. These beds were deposited on an alluvial plain spread eastward before a source land that had risen a few tens of miles west of the area (Rubey *in* Moritz, 1953, p. 66).

OROGENIC PHASE

The Twin Creek limestone and the Gannett group record in their sediments progressive stages in the eastward migration of the orogenic belt. The orogenic phase finally involved the area of the Peale Mountains sometime between the Early Cretaceous and the Eocene, but most probably at the close of the Early Cretaceous. At that time, rocks of Madison and post-Madison age lying nearly horizontally in what is now Bear River valley were detached from the underlying pre-Madison beds and thrust eastward onto the top of the nearly horizontal Twin Creek limestone. At the east side of Crow Creek valley the thrust plane rose in the section to cross the uppermost Jurassic and Lower Cretaceous formations. During thrusting, the beds of the upper plate were folded and faulted. Continued compression folded the thrust surface into a broad anticline and syncline and also folded the Jurassic strata in Whiskey Flat.

The land that emerged as a result of the thrusting and associated folding was of sufficient elevation to supply pebbles to the Late

Cretaceous and earliest Tertiary sediments of western Wyoming; however, the warm, humid climate that continued to prevail in western Wyoming (Dorr, 1958, p. 1239; Axelrod, 1950) indicates that there was no lofty range of mountains and thus that erosion nearly kept pace with uplift. This, in turn, suggests that the orogeny was of considerable duration.

POSTOROGENIC PHASE

During and after the Laramide folding and thrusting and before deposition of the Wasatch formation of Eocene age, the area was deeply eroded. In the Bear Lake Plateau the erosion surface on which the Wasatch formation was deposited was relatively smooth, although the total relief was as much as 1,350 feet (Mansfield, 1927, p. 13). Eocene conglomerate may have originally been deposited across all or nearly all the area, although the only remnants are the red conglomerate on the west side of the Aspen Range and the conglomerate on top of Red Mountain. The base of the Wasatch in The Hole is now at an altitude of about 7,500 feet, but it has been downfaulted at least 2,000 feet and probably more. Therefore, before normal faulting it was at least level with, and perhaps considerably higher than, the rocks now at the crest of Snowdrift Mountain. The occurrence of lower Paleozoic pebbles near Summit View and on hill crests on the Salt Lake formation also suggest that the conglomerate was widespread, and scattered pebbles of lower Paleozoic quartzite have been found at an altitude of 9,200 feet on the east side of the head of South Canyon. It therefore seems plausible that the relatively smooth surface that underlies the Wasatch in the Bear Lake Plateau was also developed across the southern and central Peale Mountains in Late Cretaceous and early Tertiary time, and that a blanket of red Eocene conglomerate was deposited across the area.

In middle Tertiary time, the present ranges and basins of southeastern Idaho were defined either by warping or, more probably, by normal faulting. The drainage was disrupted by these crustal movements, and lakes were formed in which the Salt Lake formation was subsequently deposited. The regional deposition of the Salt Lake formation must have been preceded by considerable erosion during which most of the Wasatch formation was removed, for lower Paleozoic boulders are extremely uncommon in conglomerate of the Salt Lake formation, and the Salt Lake formation in the well drilled north of Georgetown rests directly on Mesozoic rocks. The present relief between the depositional base of the Salt Lake formation near Summit View and the top of Meade Peak is about 2,600 feet and was surely greater in Salt Lake time. The small exposure of Salt Lake on upper Crow Creek indicates also that that valley was as deeply eroded at the beginning of Salt Lake time as it is today.

Deposition of the Salt Lake formation may have continued through the Pliocene and perhaps into the Pleistocene. Toward the end of Salt Lake time, extensive boulder beds were deposited, probably as fans, perhaps as a result of rapid uplift of the range along the range-front fault or perhaps because of a change of climate. Normal faulting continued after deposition of the Salt Lake formation had ceased, and the beds of the Salt Lake were tilted, some steeply so.

After deposition of the Salt Lake formation and after its deformation by normal faulting, a broad pediment was eroded across the Salt Lake formation in Bear Lake Valley. At places the pediment may have extended fingerlike up Georgetown Canyon and up the Left Fork of Twin Creek to Summit View. The altitude of the pediment and its extreme dissection indicate that it was formed before the Recent, and inasmuch as the age of youngest fauna from the Salt Lake formation is "Pliocene-Pleistocene" (Teng-Chien Yen, written communication, 1951), the pediment may have developed in the early Pleistocene. As Bear River entrenched itself in the valley, the pediment was dissected by tributaries flowing westward from the range.

While the pediment on the west side of the area was being dissected, Slug Valley and Crow Creek valley were being reexcavated. After Crow Creek valley had reached nearly its present depth, several large fans refilled the floor of the valley. The most likely cause of the fan building is climatic change; the fans, therefore, are dated as Pleistocene, although an early Recent date is not impossible. More recently the fans have been dissected.

The postorogenic events that led to the development of the present landscape are summarized in figure 8; both the time and the extent of the events are of necessity shown more precisely than the actual evidence allows. The proposed history is similar to that advanced by Eardley (1955, fig. 9) for several areas in north-central Utah. In the areas discussed by Eardley, however, the period of erosion that preceded deposition of the equivalents of the Salt Lake formation apparently resulted in the formation of a widespread surface.

ECONOMIC GEOLOGY

PHOSPHATE DEPOSITS

PREVIOUS WORK

Phosphatic rocks were known to exist in the Georgetown Canyon area at least as early as 1906 (Weeks and Ferrier, 1907), and by December 9, 1908, when most public lands in the region were withdrawn from entry, placer claims had already been located on nearly all the Phosphoria outcrop on both sides of Georgetown Canyon and on the western outcrop in South Canyon. The U.S. Geological Survey began an intensive study of the phosphate deposits of southeastern

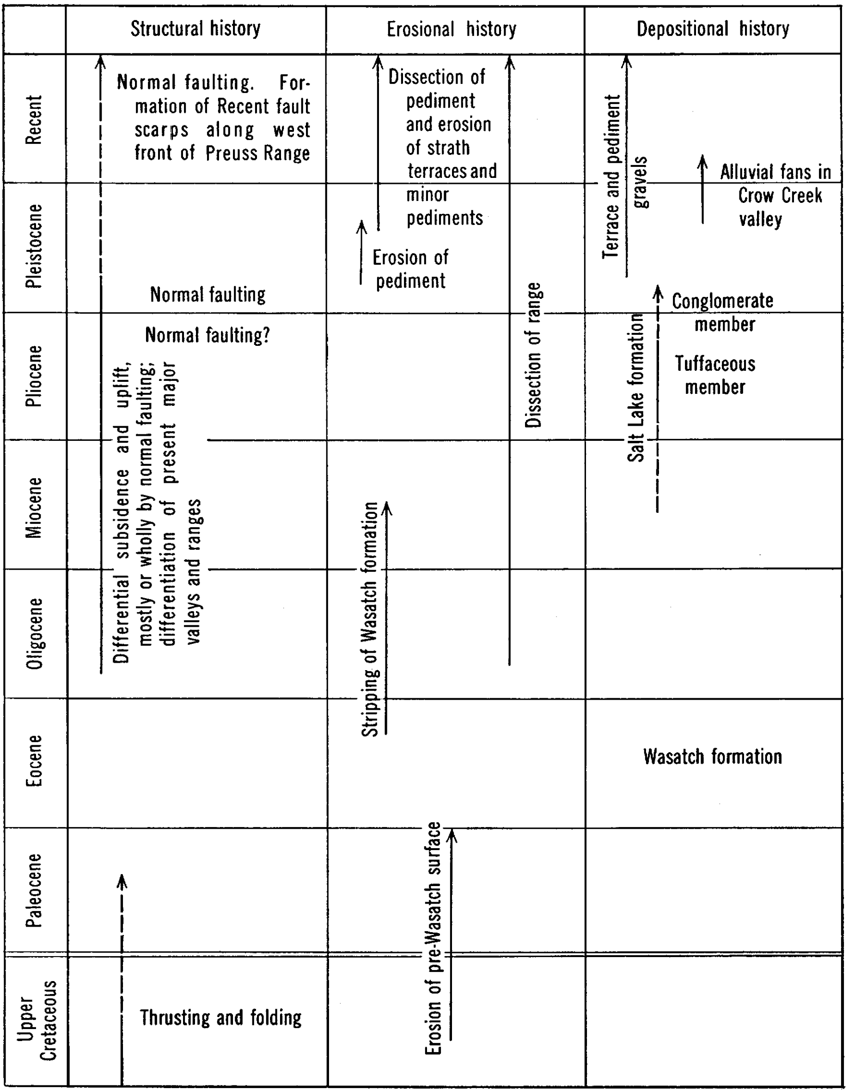

FIGURE 8.—Chart showing Cenozoic history of the central Peale Mountains and adjacent parts of southeastern Idaho.

Idaho and adjacent areas in 1908, and a map of the Phosphoria in Georgetown Canyon and a measured section with P_2O_5 analyses of the shale member were published by Gale and Richards in 1910. Geologic maps of most of the area were published by Richards and Mansfield in 1914; these were somewhat simplified versions of the maps later included in Mansfield's comprehensive paper on the region (Mansfield, 1927). During World War II, sections of the Meade

Peak phosphatic shale member were measured and sampled by the Geological Survey to evaluate the vanadium content (McKelvey and others, 1959, p. iii). Also during the war Deiss (1949) mapped the Phosphoria outcrop in the Webster syncline south of Deer Creek and in the Georgetown syncline from the north edge of the area to a line about the latitude of the summit of Georgetown Canyon. Deiss' report contains detailed stratigraphic sections of the Meade Peak member of the Phosphoria formation and several hundred partial chemical analyses. Lowell (1952) studied the petrography of the phosphatic rocks from the sections described by Deiss.

As part of the Geological Survey's current investigations of the western phosphate field, five sections have been measured in the Georgetown Canyon-Snowdrift Mountain area. These are shown graphically on plate 7.

CHARACTER AND OCCURRENCE

The Meade Peak phosphatic shale member of the Phosphoria formation contains great amounts of phosphorous in the mineral carbonate-fluorapatite (Altschuler and others, 1952). The member averages 11 to 15 percent P_2O_5 but the apatite is unevenly distributed, some beds containing more than 32 percent P_2O_5 and others containing less than 1 percent. The phosphate is concentrated in two zones, one near the base and one near the top of the member; the thickest and highest quality beds are at the base of the lower zone and the top of the upper (McKelvey, 1949, p. 274). Beds within the Meade Peak member are remarkably continuous and uniform in character throughout the area (pl. 4).

Most of the following brief summary of the petrography has been taken from Lowell (1952).

The carbonate-fluorapatite occurs for the most part in cryptocrystalline and seemingly amorphous pellets of sand size, although oolites and pisolites are common in the upper phosphatic zone and nodules are present near the middle of the member. Common accessory minerals in the phosphorite are quartz (mostly of silt size), muscovite, and calcite. The mudstone consists of angular and subangular quartz silt, muscovite, and clay minerals, and many beds contain much admixed calcite. Both the phosphate rock and mudstone contain much organic matter that colors the rock black or nearly black. Limestone beds are fine grained (0.016 to 0.061 mm in diameter) and impure; angular quartz silt grains are the most abundant contaminant, but clay minerals are also present and apatite pellets are abundant in some beds. In some surface exposures most of the carbonate has been removed by weathering, and the position of limestone beds is marked by a thin residue of mud.

In the Georgetown Canyon-Snowdrift Mountain area the Phosphoria formation is restricted to the overthrust sheet, where it underlies at least parts of the major synclines but has been eroded from the major anticlines. Because of the gentle northward plunge of the folds, the formation is present in only the northernmost parts of the Dairy and Schmid synclines and is absent from the southern part of the Georgetown syncline. The Meade Peak member is gently dipping, although mostly covered, on the east limb of the Dairy syncline, and it dips only 5° to 30° on the east limb of the Webster syncline; elsewhere, it is either steeply dipping or overturned. The member has been thinned or faulted out by thrusts along much of the overturned west limb of the Georgetown syncline and is commonly offset by many transverse faults, most of which are of too small displacement to show at the scale of the maps.

The depth of the Meade Peak member within the synclines can be estimated rather closely from the thickness of the overlying units, but some errors in depicting the member at depth doubtless result from the incompetency of the Lower Triassic formations. For example, it seems likely that the small folds in the Triassic of the Georgetown syncline die out downward within the Dinwoody formation, but the evidence is indirect (p. 65). The configuration of the Phosphoria in the more tightly folded parts of the Georgetown syncline is also somewhat conjectural, for the fold may be either more tight or more open than is shown in the structure sections.

PROSPECTING

Although the beds within the Meade Peak phosphatic shale member of the Phosphoria formation are remarkably continuous, local variations in thickness and grade and the possibility of small faults not detectable in surface exposures require that closely spaced sections be measured and analyzed in any locality considered for mining (Gulbrandsen and others, 1956, p. 18; Cressman and Gulbrandsen, 1955, p. 268). The Meade Peak member is seldom exposed naturally; its presence is deduced by float, by its position above the resistant Grandeur tongue of the Park City formation and below the Rex chert member of the Phosphoria, and by its topographic expression as a bench or swale. Sections must therefore be exposed artificially or sampled by drilling.

Core drilling has not been very successful in the Meade Peak member because of poor core recovery and difficulties in maintaining circulation. In an experimental drilling program of the U.S. Bureau of Mines, Long (1949) reported recoveries of 30 to 60 percent, and even though subsequent work by phosphate mining companies has resulted in somewhat improved recovery, most logging and sampling is now done from cuttings. Data from drill holes are invaluable, particularly in those areas where trenching is impossible; but drill holes cannot yield the detailed information on stratigraphy and structure that is obtainable from a good bulldozer trench or adit.

Bulldozer trenches must be located carefully and dug deeply to expose beds below the zone of creep (McKelvey and others, 1953, p. 3). Many trenches in the mapped area have failed to yield adequate information because they were poorly located or because they were not dug sufficiently deep. Some trenches located on crests and divides, particularly those on the west limbs of the Webster and Georgetown synclines, did not penetrate through the colluvial cover of float from the Wells formation, which at places is as much as 3 or 4 feet thick; others were located near the base of slopes where the colluvium was too thick to penetrate. Many trenches, even those in areas where the colluvium is thin, have not been dug sufficiently deep, and misleading thicknesses have been obtained by measuring sections in the zone of creep. Because of the inadequacy of much of the trenching, I doubt that the potentiality of the Phosphoria in much of the area has yet been properly evaluated, and it is unlikely that information from drill holes can be interpreted correctly without the control of excellent detailed surface sections.

Weathering of the Meade Peak member removes carbonates and organic matter so that surface exposures of the Meade Peak member are somewhat enriched in phosphate (McKelvey and Carswell, 1956, p. 384). Mansfield (1927, p. 213) noted that at Waterloo Hill near Montpelier a bed containing 38 percent P_2O_5 at the surface contains 34.8 and 33.7 percent P_2O_5 at two localities where it was sampled underground. Surface enrichment of phosphate must therefore be considered in evaluating data from trenches. Where the Meade Peak member dips steeply, the removal of carbonates and organic matter may cause the settling of adjacent beds with resultant thickening. The lower phosphate zone, overlain by a limestone bed and underlain by calcareous mudstone, may thus be several feet thicker in weathered surface exposures than at depth.

The uniformity of the Meade Peak member is such that in interpreting data from either surface sections or drill holes, faulting should seriously be considered to explain any large deviation from the stratigraphic sequence shown in the Snowdrift Mountain, Deer Creek, and Clear Creek sections.

ECONOMIC CONSIDERATIONS

The two important basic methods used in the west for treating phosphate rock are the sulfuric acid or wet process and the thermal process, which most commonly employs the electric furnace. The wet process requires rock with a minimum P_2O_5 content of about 31 percent, whereas the electric furnace requires rock with a minimum P_2O_5 content of 24 percent. In the Snowdrift Mountain section (pl. 7; Smart and others, 1954, p. 15) the lower phosphate bed is 7 feet thick and averages 33.3 percent P_2O_5 and is overlain by 14.6 feet of beds averaging 24.5 percent P_2O_5. The upper phosphate zone averages 28 percent P_2O_5 through a thickness of 22.9 feet. By selective mining a greater thickness of furnace-grade rock can be obtained, and by beneficiation, lower grade rock can be raised to meet the minimum requirements for both processes. The member contains nearly as much minable phosphate everywhere in the area where it has not been thinned by faulting, and it is apparent that there are very large reserves of rock suitable for either the wet or the furnace process.

The average value of phosphate rock at the mine in Idaho during 1957 was $4.65 per long ton (Ruhlman and Tucker, 1958). The value is dependent on the P_2O_5 content; so acid-grade rock is higher priced and furnace-grade lower priced. Because of the low value of the rock, and the wide extent of the deposits, the development of any one deposit is largely dependent on accessibility, the availability of acid or electricity, and the adaptability of the deposit to low-cost mining methods.

In southeastern Idaho, under present conditions, stripping seems to be the only economical mining method; even at Conda near Soda Springs underground mining that had been in progress for many years was stopped in 1956 in favor of stripping. Until recently, strip mining has been practiced only in areas of gentle dip or on dip slopes on the Meade Peak member, but it has recently been shown that it is feasible to mine by open-pit methods steeply dipping and vertical beds, especially on steep slopes.

DEVELOPMENT

Development of phosphate deposits within the area before 1927 has been discussed by Mansfield (1927, p. 261–266). Briefly, many hand trenches had been dug, and several adits had been driven into the Meade Peak member in Phosphoria Gulch, in the western outcrop in South Canyon, and in the outcrop on the northwest slope of Meade Peak; but there had been no production. A tramway was built from the bend in Georgetown Canyon to the outcrop on the northwest slope of Meade Peak. Little additional work was done in the area until after World War II, and most of the original workings are caved. No trace remains of the tramway.

The patented claims that include nearly all the Phosphoria outcrop on both sides of Georgetown Canyon are now owned by the Central

Farmers Fertilizer Co. In 1957 construction was started on a processing plant in Georgetown Canyon at the mouth of Phosphoria Gulch, and a railroad spur, an all-weather road, and a powerline were built up the canyon to the plant site. Open-pit mining was begun on the southern part of the east limb of Georgetown syncline in 1958, and the plant was completed in 1959. The mining is highly selective; 3 grades of rock averaging 18, 26, and 32 percent P_2O_5 are separated. The rock is trucked from the pit to a feeder and from there to the plant site by an endless conveyer belt. Present plans call for the mining of 900,000 tons of ore in a 4- to 5-month mining season.

At the plant, low-grade rock is washed and calcined to raise it to the grades required by the several steps in the manufacturing process. Elemental phosphorus is produced in the electric furnace and converted to phosphoric acid, containing 78 percent P_2O_5, which, in turn, is used to treat phosphate rock. The final product is water-soluble monocalcium phosphate containing 53 to 55 percent P_2O_5. The plant will produce annually in excess of 200,000 tons of phosphate fertilizer.

The Central Farmers Fertilizer Co. has found many more small faults in their open pit than had been expected from their prospecting trenches (David Hand, oral communication, 1959). I have not examined the pit and do not know the nature of the faults, but inasmuch as the major folds of the area are concentric, concentric shears should be expected on the fold limbs; the tighter the fold and the more steeply dipping the limbs, the greater should be the concentric shearing. The Meade Peak member is incompetent and might be particularly affected by such shearing. An analysis of structure in trenches and in existing strip operations with respect to the nature of the larger fold and the position in the fold might enable better predictions of structural complications that could be expected in a particular section of outcrop.

Other than in the areas of patented claims on the Georgetown Canyon outcrops, on the west outcrop in South Canyon, and on the small outcrop on the northwest flank of Meade Peak, nearly all the Phosphoria formation in the area is part of the western phosphate reserve. Leases are held on the Webster syncline south of Deer Creek and on the west limb of the Georgetown syncline north of Georgetown summit; both of these areas have been prospected by bulldozer, and in the Webster syncline by drillng, but no phosphate rock has been mined.

The steeply dipping Phosphoria on the east limb of Snowdrift anticline and the west flank of the Dairy syncline might be as successfully stripped as that on the east limb of the Georgetown syncline. The Meade Peak member is thinned drastically or cut out by thrusts along most of its outcrop belt on the west limb of Georgetown syncline, but the northernmost mile and the southernmost 2 miles might be sufficiently free of faults to be mined.

WATER

Until recently the water resources of the area had been developed only to the extent necessary for watering stock and for small-scale irrigation of pasture and hay meadows in Crow Creek Valley. However, the newly constructed plant of the Central Farmers Fertilizer Co. has demanded a larger water supply than can currently be supplied by the drainage area above the plant site, and water supply will probably be a critical factor for any other plant that might be located within the area. The water resources have not been studied expressly, but the following observations may be of some assistance to anyone interested in the hydrology of the area.

Crow Creek and the lower parts of Twin and Preuss Creeks are perennial. A gage on Twin Creek, just upstream from the junction with the Right Fork, has recorded an average discharge of 31.1 cubic feet per second or 22,520 acre-feet per year over a period of 16 years (U.S. Geological Survey, 1956). The drainage area supplying the flow is 22.2 square miles. The discharge of the other two streams has not been measured, but that of Crow Creek is probably somewhat smaller and that of Preuss Creek considerably smaller. Many other streams contain parts, some of considerable length, that flow perennially: but other parts of the same streams sink into the alluvial valley floor in the summer months.

Springs are numerous, but most are of small size. The more predominant types of spring occurrences are listed below.

1. At the contact of the Phosphoria and Dinwoody formations. Many springs issue from this horizon on both flanks of the Georgetown and Webster synclines. However, the springs are all very small, many being little more than small marshy seeps.

2. Along faults. The only clear-cut example of this type is on the north side of upper Deer Creek where several springs of considerable flow rise along the Deer Creek fault where it crosses the lower member of the Dinwoody formation. Elsewhere, the absence of any relation between faults and springs is striking.

3. From valley alluvium. The largest springs in the area rise from alluvium, generally where bedrock ribs crossing the valley floor block ground water moving downstream through alluvial gravels. Two examples are in South Canyon where the Nugget sandstone crosses the valley and at the bend in Georgetown Canyon where a rib of Carboniferous limestone constricts the canyon.

4. From colluvium. These springs are all insignificant seeps.

5. From fractured bedrock. Two examples of this type are in Dunns Canyon where small springs rise from the lower part of the Twin Creek limestone, and a half mile southwest of the summit of Georgetown Canyon where a tufa-depositing spring flows from a hillside underlain by the upper member of the Dinwoody formation. In neither example is there any obvious structural or stratigraphic reason for the location of the spring.

Except for the Gannett group, the Paleozoic and Mesozoic formations are not very porous. Even sandstone of the Wells formation that is porous in weathered outcrops is tightly cemented at depth (Zeni, 1953). This, together with the occurrence of springs, suggests that the Dinwoody formation, where highly fractured, is the most likely bedrock formation to carry ground water. Considering the generally small flow of the springs rising from this formation, the amount of water obtainable is probably small.

Two other sources of ground water remain. Considerable quantities could probably be obtained from alluvium in the canyon floors, but the withdrawal of water would undoubtedly result in diminished flow in the surface streams in the lower parts of the canyons. A second possibility is that water might be obtained from wells in the Salt Lake formation along the west front of the range (V. E. McKelvey, oral communication, 1950). The Salt Lake formation contains many very permeable beds, and much of the formation dips steeply toward the range front, suggesting that a considerable reservoir might exist within that formation.

OTHER MINERAL DEPOSITS

Other than phosphorite and water, the only mineral resources exploited within the area have been rock waste from the Nugget sandstone and the Twin Creek limestone used locally for road metal and sandstone from the Nugget used for flux in the extraction of elemental phosphorus by the electric furnace process. Sulfur and salt deposits have been developed on a small scale in past years in adjacent areas (Mansfield, 1927, p. 338, 341), but they are not being worked at present. The small sulfur deposits along the front of the Aspen Range in the area of this report are much too small to be considered for mining; and salt beds, if present in the Preuss sandstone, are at depth.

If population and industry develop in southeastern Idaho to the extent that the manufacture of Portland cement is economically feasible, an adequate source of nearly pure limestone can probably be found in the Madison and Brazer formations, and some of the Twin Creek limestone probably meets the specification of cement rock (Mansfield, 1927, p. 331, 334.)

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