Geology and Ground-Water Resources of the Deer Lodge Valley Montana

GEOLOGICAL SURVEY WATER-SUPPLY PAPER 1862

Prepared in cooperation with the Montana Bureau of Mines and Geology, Butte, Montana

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R. L. KONIZESKI, R. G. McMURTREY, *and* ALEX BRIETKRIETZ

With a section on GRAVIMETRIC SURVEY
E. A. CREMER III

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By R. L. KONIZESKI, R. G. MCMURTREY, and ALEX BRIETKRIETZ

ABSTRACT

The Deer Lodge Valley is a basin trending north-south within Powell, Deer Lodge, and Silver Bow Counties in west-central Montana, near the center of the Northern Rocky Mountains physiographic province. It trends northward between a group of relatively low, rounded mountains to the east and the higher, more rugged Flint Creek Range to the west. The Clark Fork and its tributaries drain the valley in a northerly direction. The climate is semiarid and is characterized by long cold winters and short cool summers. Agriculture and ore refining are the principal industries. Both are dependent on large amounts of water.

The principal topographic features are a broad lowland, the Clark Fork flood plain, bordered by low fringing terraces that are in turn bordered by broad, high terraces, which slope gently upward to the mountains. The high terraces have been mostly obscured in the south end of the valley by erosion and by recent deposition of great coalescent fans radiating outward from the mouths of various tributary canyons.

The mountains east of the Deer Lodge Valley are formed mostly of Cretaceous sedimentary and volcanic rocks and a great core of Upper Cretaceous to lower Tertiary granitic rocks; those west of the valley are formed of Precambrian to Cretaceous sedimentary rocks and a core of lower Tertiary granitic rocks. Field relationships, gravimetric data, and seismic data indicate that the valley is a deep graben, which formed in early Tertiary time after emplacement of the Boulder and Philipsburg batholiths. During the Tertiary Period the valley was partly filled to a maximum depth of more than 5,500 feet with erosional detritus that came from the surrounding mountains and was interbedded with minor amounts of volcanic ejecta. This material accumulated in a great variety of local environments. Consequently the resultant deposits are of extremely variable lithology in lateral and vertical sequence. The deposits grade from unconsolidated to well-cemented and from clay to boulder-sized aggregates. Throughout most of the area the strata dip gently towards the valley axis, but along the western margins of the valley they dip steeply into the mountains.

In late Pliocene or early Pleistocene the Tertiary strata were eroded to a nearly regular, valleywide surface. In the western part of the valley the erosion surface was thinly mantled by glacial debris from the Flint Creek Range. Still later, probably during several interglacial intervals, the Clark Fork and its tributaries entrenched themselves in the Tertiary strata to an average depth of about 150 feet. The resultant erosional features were further modified by Wisconsin to Recent glaciofluvial deposition.

Three east-west cross sections and a corrected gravity map were drawn for the valley. They indicate a maximum depth of fill of more than 5,500 feet in the southern part. Depths decrease to the north to approximately 2,300 feet near the town of Deer Lodge.

The principal source of ground water in the Deer Lodge Valley is the upper few hundred feet of unconsolidated valley fill. Most of the wells tapping these deposits range in depth from a few feet to 250 feet. Water levels range from somewhat above land surface (in flowing wells) to about 150 feet below. Yields of the wells range from a few gallons per minute to 1,000 gallons per minute. Generally, wells having the highest yields are on the flood plain of the Clark Fork or the coalescent fans of Warm Springs and Mill Creeks.

Discharge of ground water by seepage into streams, by evapotranspiration, and by pumping from wells causes a gradual lowering of the water table. Each spring and early summer, seepage of water from irrigation and streams and infiltration of water from snowmelt and precipitation replenish the ground-water reservoir. Seasonal fluctuation of the water table generally is less than 10 feet. The small yearly water table fluctuation indicates that recharge about balances discharge from the ground-water reservoir.

A generalized representation of the water table in part of the valley indicates that ground water moves toward the Clark Fork from the east and west, but in detail the slope of the water table and the direction of movement of the water vary throughout the valley. In the southwestern part movement of ground water is generally northeast at a gradient of about 50 feet per mile with components of flow towards the tributary streams. In other parts ground water moves toward the Clark Fork under a gradient of 70 feet per mile or more. The rate of movement of ground water in the valley is generally less than 3 feet per day.

INTRODUCTION

PURPOSE AND SCOPE

The study of the geology and ground-water resources of the Deer Lodge Valley is part of a cooperative program with the Montana Bureau of Mines and Geology to evaluate the ground-water resources of Montana (fig. 1). The study was began in July 1957 and the fieldwork was completed in 1962. It was made in order to determine (1) the character, thickness, and extent of the water-bearing materials, (2) the source, occurrence, and direction of movement of the ground water, (3) the quantity and availability of the ground water, (4) the seasonal, annual, and long-term changes in ground-water storage, and (5) areas from which substantial supplies of underground water can be obtained.

ACKNOWLEDGMENTS

The writers wish to express their appreciation to all who aided this study. Thanks are extended to those individuals who willingly supplied information about their wells, consented to the use of their wells for water-level measurements, and allowed admittance to their land. Well drillers, especially C. F. Wroble, were most helpful in furnishing well logs and other useful information. The following organiza-

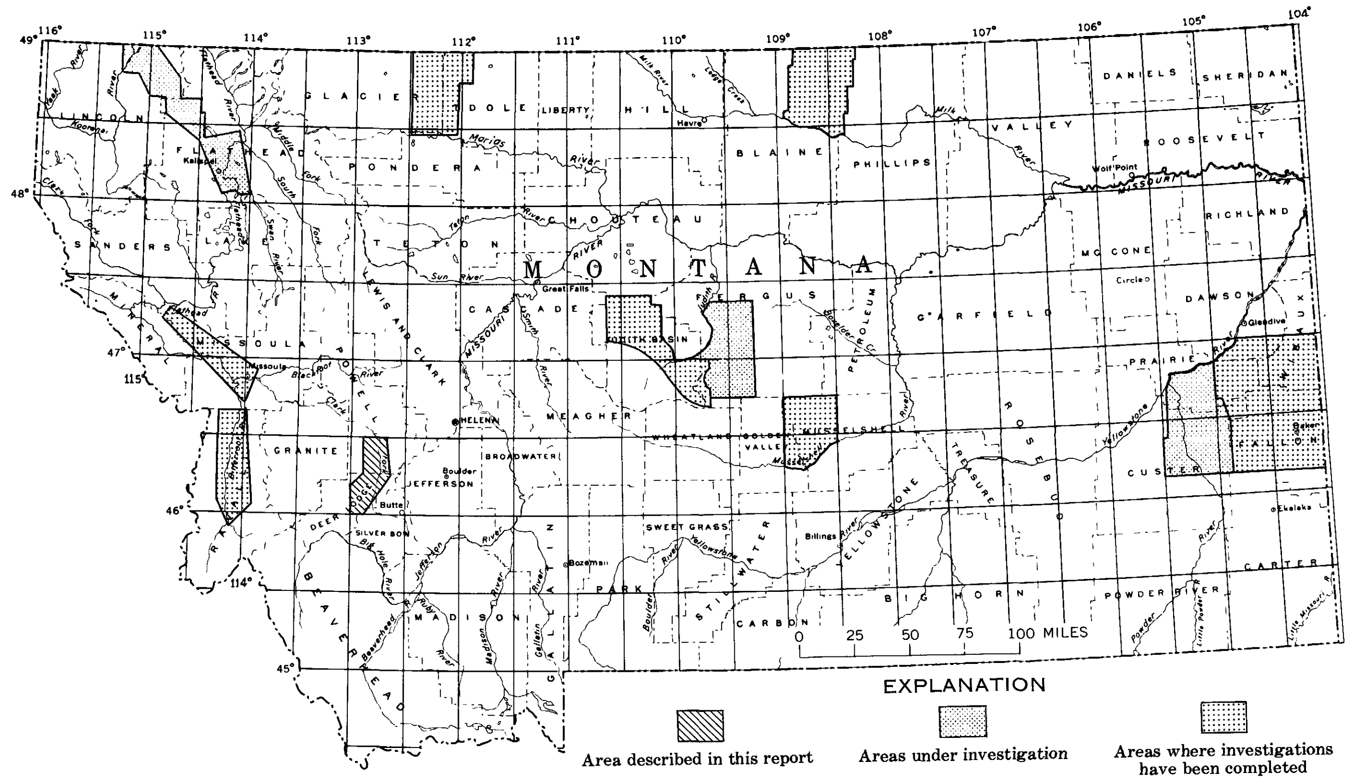

FIGURE 1.—Cooperative ground-water investigations in Montana, 1955–67.

tions contributed to the progress of the investigation: Anaconda Copper Mining Co., Mount Haggin Land and Livestock Co., Montana State Prison, Montana State Hospital, Montana Power Co., and the U.S. Soil Conservation Service.

PREVIOUS INVESTIGATIONS

During the summers of 1899–1900 Prof. Earl Douglass of Montana State University prospected for fossil vertebrates and studied the Cenozoic geology of the Deer Lodge Valley. The results of his work were published in 1901 and have formed the basis for most later descriptions of Cenozoic stratigraphy in the area. Subsequent geologic and hydrologic investigations include a discussion of the regional stratigraphy of Billingsley (1913), a description of the coal resources of the valley by Pardee (1913), a generalized description of the geology and topography of the area by Campbell (1915), and a discussion of the Cenozoic history of the valley by Alden (1927).

Perry (1933) included a brief description of the ground-water resources of the valley in a report on the possibilities of ground-water supply for certain towns and cities of Montana. Pardee (1951) attributed the origin of the valley to structural deformation. Alden (1953) discussed glacial deposits along the western margins of the valley. Konizeski (1957) described Pliocene ecological relationships within the valley. Mutch (1960) outlined the glacial chronology of the Flint Creek Range, and Csejetey (1962) presented a detailed discussion on the stratigraphic relationships of outcropping Tertiary strata at the base of the range near Anaconda. Smedes (1962) related certain volcanic rocks near Anaconda with the Tertiary Lowland Creek Volcanics near Butte. Preliminary reports on the water resources of the area have been published (Konizeski and others, 1961, 1962).

Numerous mine reports include references to the area.

METHODS

The geology was mapped on aerial photographs. The geologic map includes the valley and a border zone, which shows the relationship of the valley fill to the consolidated rocks of the surrounding mountains. A gravimetric survey was made during the 1961 field season. The data were used to interpret the approximate basement profile of the valley, structural relationships, and the approximate depth of valley fill.

From a well inventory made during the field seasons of 1957 and 1960, data were compiled for 270 wells and springs in the area, and a network of 84 observations wells was selected to determine water-table fluctuations. Data from the inventory and water-level measurements in the observation wells are on open file at the U.S. Geological Survey office in Helena, Mont. The altitude of the water surface in 129 wells, measured in September 1960, was used to make a map showing contours on the water table. The hydrologic properties of water-bearing materials were determined by 17 single-well and three multiple-well aquifer tests.

WELL-NUMBERING SYSTEM

The wells described in this report are assigned numbers that are based on their location within the system of land subdivision used by the U.S. Bureau of Land Management. The well number shows the location of the well by quadrant, township, range, section, and position within the section. This method of well numbering is shown in figure 2. The first letter of the well number gives the quadrant of the meridian and base-line system in which the well is located. The first numeral indicates the township, the second the range, and the third the section in which the well is located. Lowercase letters that follow the section number show the location of the well within the quarter section (160-acre tract) and the quarter-quarter section (40-acre tract). These subdivisions are designated "a," "b," "c," and "d" in a counterclockwise direction, beginning in the northeast quadrant. If two or more wells are within the same 40-acre tract, consecutive digits beginning with 1 are added to the well number. For example, well B4–10–36ab3 is in the NW¼NE¼ sec. 36, T. 4 N., R. 1 W. and is the third well inventoried in that tract. Springs are numbered in the same manner.

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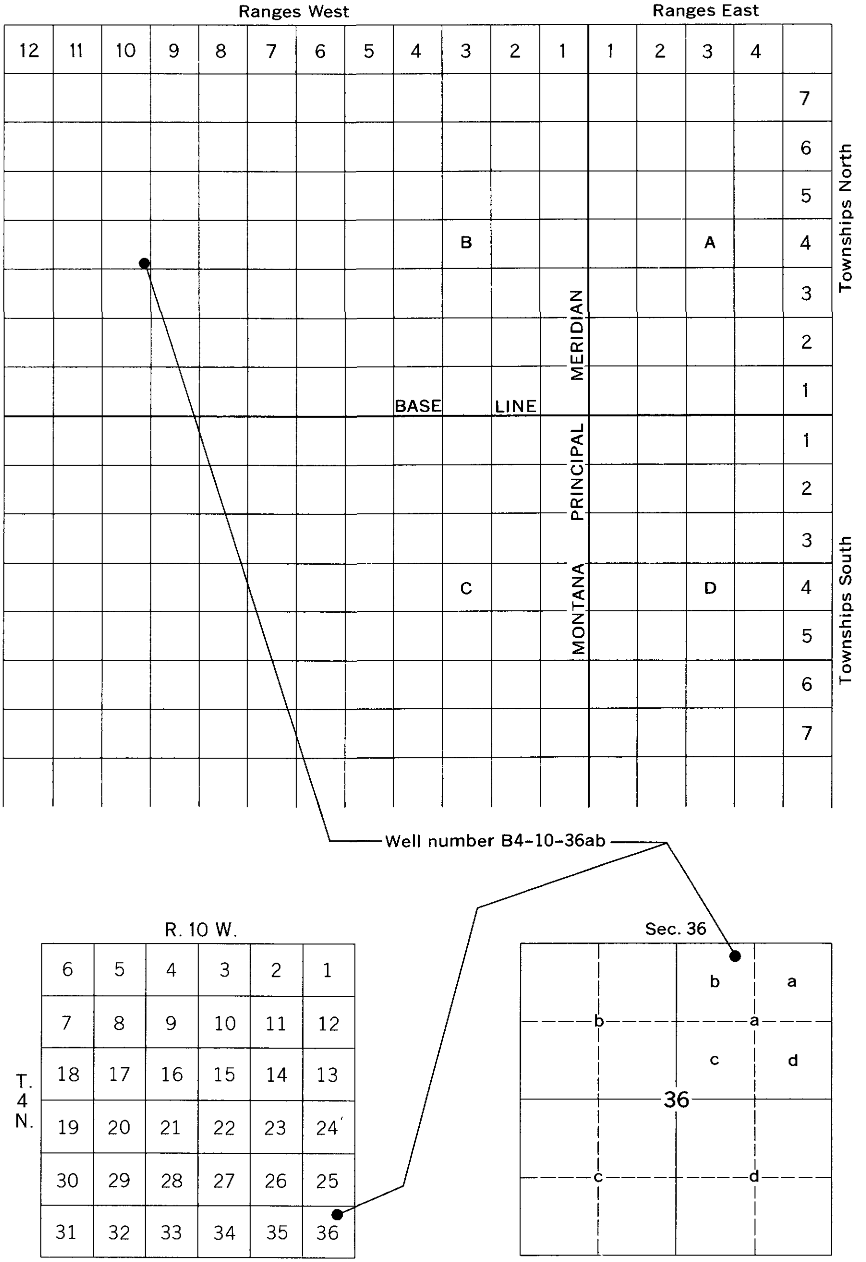

FIGURE 2.—Well-numbering system.

GEOGRAPHY

LOCATION AND EXTENT OF VALLEY

The Deer Lodge Valley includes about 300 square miles within Powell, Deer Lodge, and Silver Bow Counties. It extends northward from near Gregson to a relatively narrow place near Garrison (fig. 3). The term "valley" as used in this report includes the area of relatively low altitude and low relief that is bounded on the east by a group of low, rounded mountains and on the west by the rugged Flint Creek Range.

TOPOGRAPHY

The Deer Lodge Valley is a basin trending north-south near the center of the Northern Rocky Mountains physiographic province (Fenneman, 1931, p. 223). The Continental Divide is roughly parallel to and 5 to 15 miles east of the valley (fig. 3). The mountains in this area are locally known as the Deer Lodge Mountains and are mostly below 8,000 feet, but a few attain altitudes of more than 8,500 feet. The topography is generally low and rolling with tree-covered slopes and great open parks. However, a deeply glaciated area of high local relief exists east of Deer Lodge around Thunderbolt, Cliff, Negro, and Bison Mountains.

The Flint Creek Range is a deeply dissected dome about 28 miles long in the north-south direction by 20 miles wide. Much of the area is more than 8,000 feet in altitude, and several peaks are more than 9,000 feet. Mount Powell, the highest peak in the range, attains an altitude of 10,171 feet. Many deep gorges are separated by knife-edged divides and depths on the order of 2,000 feet or more are not unusual. U-shaped canyons, arêtes, cirques, ice scour, and other indications of glaciation are much in evidence.

The valley topography is dominated by great terraces, designated here as high terraces, that slope gently downward from the mountains and end in abrupt scarps above low terraces bordering the broad flood plain of the Clark Fork. The slope of the high terraces toward the Clark Fork ranges from 4° or 5° near the mountains to less than 1° near the low terraces. The high terraces range in altitude from 5,700 feet to 4,600 feet. In the northwestern part of the valley they are about 125 feet above the low terraces and have a maximum width of nearly 5 miles. The high terraces narrow and become progressively lower towards the south end of the valley. Throughout the valley their average width is about 4 miles and they have been deeply dissected. The flood plains of the tributary streams coalesce and merge with the low terrace along the Clark Fork.

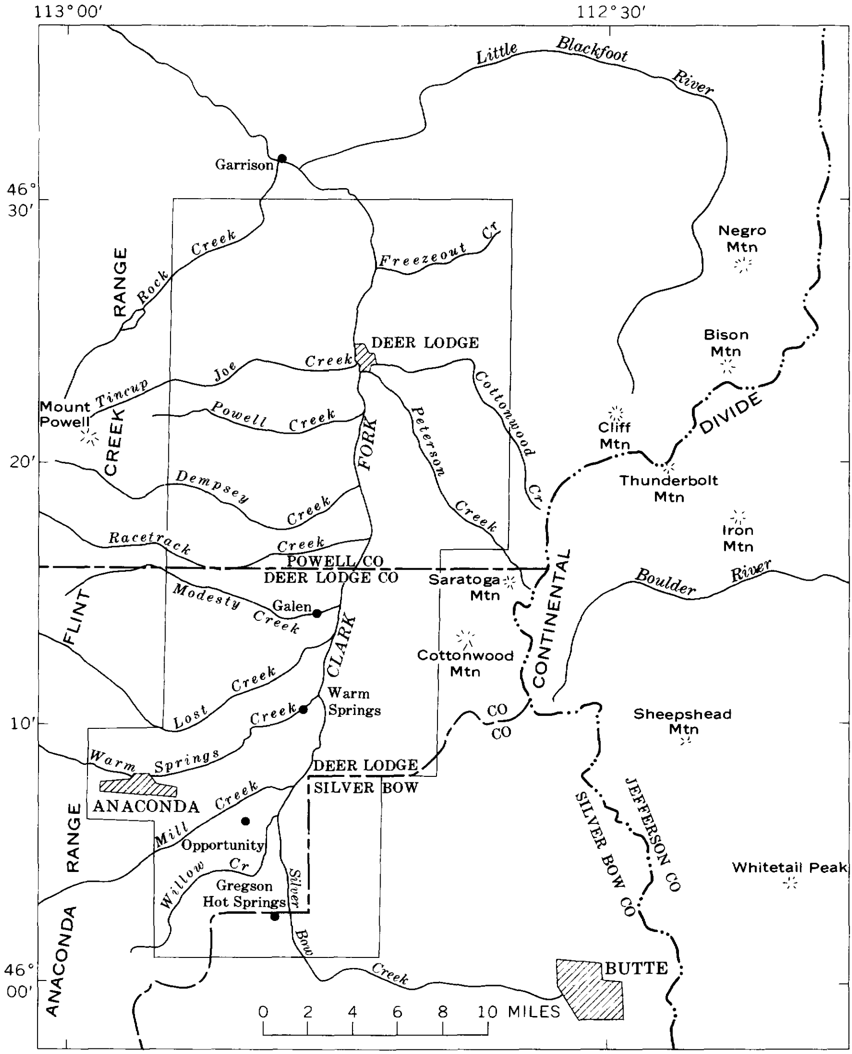

Figure 3.—Study area and principal topographic and drainage features.

In the southwestern part of the valley all these several topographic features merge and tend to lose their identities because of dissection of the terraces and emplacement of great coalescent fans. However, isolated remnants of the high terraces occur about 600 feet above the valley floor. Most of the major tributary valleys west of the Clark Fork are modified by great Wisconsin moraines.

DRAINAGE

Above its junction with the Little Blackfoot River the Clark Fork and its tributaries drain about 1,200 square miles. Because the area east of the valley is generally semiarid, streams from there are relatively small. Conversely because the Flint Creek Range is generally subhumid, the streams draining that area are relatively large. Two perennial tributaries enter the valley from the east; seven, from the west; and two, from the south. Numerous intermittent tributaries drain the lower mountain slopes on both sides of the valley. The Clark Fork turns sharply to the northwest near Garrison and flows for many miles through a series of deep gorges.

The Clark Fork enters the south end of the valley through a narrow gorge at an altitude of about 5,100 feet and leaves the valley at an altitude of about 4,400 feet. Between Gregson and Warm Springs the average gradient is about 26 feet per mile and between Warm Springs and Garrison the average gradient is about 8 feet per mile. The mean annual flow of the Clark Fork at Deer Lodge has been estimated (Frank Stermitz, oral commun., 1961) at about 250±75 cfs (cubic feet per second). In its upper (southern) reaches, the stream is held against the eastern wall of the valley by the great coalescent fans of Mill, Warm Springs, and Lost Creeks. In its middle and lower reaches its flood plain is about 1 mile wide and lies slightly east of the valley axis. The stream's meandering course is marked by numerous cutoff oxbows, sloughs, and marshes wherein grow a variety of hydrophytic and riparian plants.

Tributaries to the Clark Fork flow from consolidated rocks in the mountains onto unconsolidated fill in the valley where much of their flow is lost by seepage. Most of the remaining flow is diverted for irrigation and industrial use.

CLIMATE

The Deer Lodge Valley has a semiarid climate characterized by long cold winters, short cool summers, low precipitation, and moderate winds. The average monthly temperature and precipitation at Deer Lodge are shown in figure 4. The highest recorded temperature at Deer Lodge during the years 1931–66 was 100° F, and the lowest temperature was 39° F below zero. Changes from midday temperatures of more than 80° F to nighttime temperatures of less than 50° F are common in July. The length of the growing season is highly variable but averages about 95 days. September 5 is the average date of the first killing frost and June 5, the last killing frost.

The highest recorded yearly precipitation was 14.67 inches in 1938; the lowest was 5.91 inches in 1935. Approximately 50 percent of the

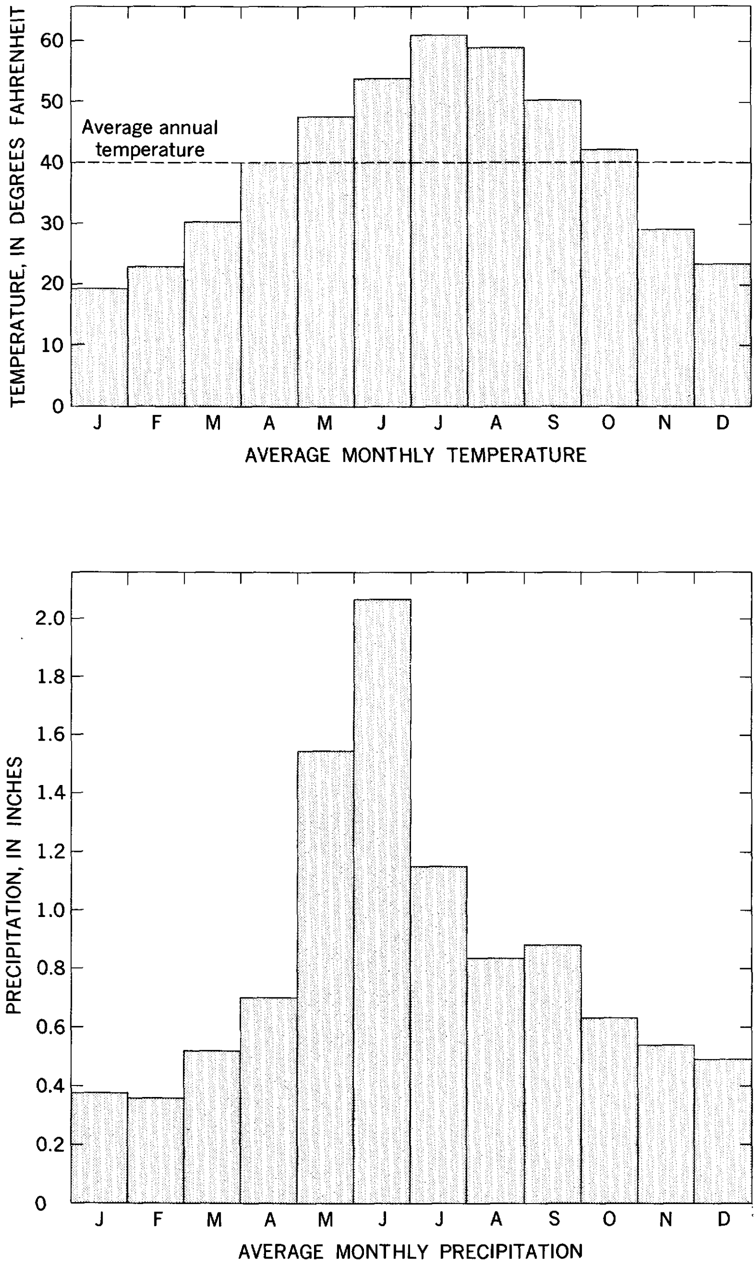

FIGURE 4.—Average monthly temperature and precipitation at Deer Lodge, 1931–66. (From U.S. Weather Bureau records.)

average annual precipitation is during May, June, and July; nearly 75 percent occurs from April through September. At Deer Lodge the periods 1936 through 1951 and 1954 through 1959 generally had above average precipitation; the other periods generally had below average precipitation.

HISTORY AND CULTURE

The Deer Lodge Valley derived its name from an Indian expression meaning white-tailed deer's lodge. It was so called because of large numbers of white-tailed deer in the valley and because of a cone-shaped mound at Warm Springs. This mound, about 40 feet high, was formed by a thermal spring. Steam issuing from the mound reminded the Indians of smoke rising from an Indian lodge.

One of the earliest settlers in the area was John Francis Grant, who in 1855 raised cattle in the valley. In the early 1860's successive groups of gold prospectors passed through on their way to the mining settlements. Many returned to the village of Cottonwood (Deer Lodge) and the surrounding area to become tradesmen and ranchers. A Federal penitentiary was built in 1871, and it became the State prison in 1873. In 1878 the first college in Montana was opened at Deer Lodge. In 1883, Marcus Daly, founder of the copper mining industry at Butte, located a smelter near Warm Springs Creek because of the favorable water supply. The smelter encouraged immigration and settlements sprang up around the site. Anaconda is one early settlement that remains.

In 1960 the population of Anaconda was 12,054 and that of Deer Lodge was 4,651. The patients and staff at hospitals in Warm Springs and Galen number more than 800 and 200 respectively. Opportunity is a small town of several hundred inhabitants.

AGRICULTURE

Seasonal shortage of water supply is a major consideration for successful farming in the valley. Most irrigated crops are raised on the Clark Fork flood plain, the low terraces, and the high terraces on the west side of the valley. The high terraces on the east side of the valley are mostly dry-farmed. Hay is the major irrigated crop and is used locally to support the livestock industry. Potatoes are the largest cash crop in the valley, the area being one of the largest producers of potato seed in Montana. Other crops are wheat, barley, and corn.

GEOLOGY

REGIONAL STRATIGRAPHY

The consolidated rocks marginal to and underlying the Deer Lodge Valley are, for convenience, referred to in this report as basement rocks (pl. 1). They are described briefly to implement later discussions of the geology of the valley fill and the ground-water regimen. The information is mostly from detailed descriptions by others, particularly Knopf (1953); Weeks and Klepper (1954); Chapman, Gottfried, and Waring (1955); Mutch (1960); Ruppel (1957, 1961); and Csejetey (1962).

The highlands marginal to the Deer Lodge Valley are formed on a great variety of rocks that range from Precambrian to early or middle Tertiary age. The Flint Creek Range is formed of a variety of structurally deformed Precambrian metamorphic rocks and Paleozoic and Mesozoic sedimentary rocks that have been intruded by granitic masses. One intrusive mass, the Philipsburg batholith, has been tentatively dated as middle Eocene(?). In general, the sedimentary rocks crop out in a narrow belt around the eastern flanks of the range. The oldest (Precambrian) rocks are farthest from the valley, and the youngest (Mesozoic) rocks are nearest to the valley. The Precambrian rocks consist of about 25,000 feet of fine-grained clastic rocks; the Paleozoic rocks consist of about 6,000 feet of limestone, dolomite, and some quartzite; and the Mesozoic rocks consist of about 15,000 feet of fine-grained clastic rocks.

More than 2,800 feet of welded tuff, described by Smedes (1962, p. 260) as a rapidly accumulated unit of the Lowland Creek Volcanics, crops out intermittently along the base of the range northward from the south side of Warm Springs Creek to Spring Gulch. Baadsgaard, Folinsbee, and Lispson (1961, p. 697) obtained a potassium-argon date of 49 million years (middle Eocene age) from biotite in a dike, which Smedes (1962, p. 264) describes as part of a system of dikes that cut the welded tuff unit west of Butte. Smedes assigned a late Oligocene age to the tuff because of (1) a small flora collected by Csejetey (1962) from associated sedimentary rocks in the Warm Springs Creek and Lost Creek areas and (2) the fact that the volcanic rocks are unconformably overlain by lower Miocene deposits west and southwest of Butte (Wood, 1936, p. 12–13). The Lowland Creek Volcanics was reassigned to the Eocene on the basis of a potassium-argon determination of 48 to 50 million years (Smedes and Thomas, 1965).

The marginal highlands to the east and south of the valley are formed on about 15,000 feet of Lower to middle Cretaceous sedimentary and volcanic rocks. These rocks have been intruded by Upper

Cretaceous to lower Tertiary (Boulder batholith) granitic rocks, which are in turn overlain by patches of lower to middle Tertiary volcanic rocks. The Cretaceous sedimentary and volcanic rocks consist of about 11,000 feet of fine-grained clastic rocks that are overlain by about 4,000 feet of andesitic pyroclastic rocks and flows. The lower to middle Tertiary volcanic rocks are mostly siliceous tuffs and flows. In general, the highlands northeast of the valley are formed on andesitic rocks; the highlands southeast of the valley are formed on granitic rocks; and the highlands near the southern end of the valley are formed on siliceous volcanic rocks and granitic rocks.

INTRAVALLEY STRATIGRAPHY

TERTIARY VOLCANIC ROCKS

An elongate belt of rocks near the axis of the valley has been described as the Garrison Vent (Mutch, 1960). It trends N. 30° W, from Mullan Creek past the northern boundary of the study area and is about 2 miles wide by several miles long. These volcanic rocks are relatively resistant to weathering and form an erosional remnant of the high terrace. Around the southeastern margin of this high terrace remnant they intrude metamorphosed Kootenai Formation and Colorado Shale. The volcanic rocks are bordered locally to the southwest by Miocene sediments and to the southeast by sediments of probably Miocene age. The contacts are poorly or not at all exposed, but some of the sediments of probably Miocene age contain locally derived volcanic detritus. The vent is believed to be of Tertiary, probably pre-Miocene age.

TERTIARY SEDIMENTARY DEPOSITS

The Deer Lodge Valley is partly filled with a great mass of material (valley fill) eroded from rocks in the surrounding mountains and with smaller amounts of volcanic ejecta. The valley fill consists mostly of unconsolidated to semiconsolidated Tertiary sedimentary deposits, but some consolidated Teritary sediments are exposed around the western margins of the valley, and others probably occur at depth. All are referred to here as Tertiary strata.

Tertiary strata underlie the high terraces and are elsewhere overlain by Quaternary alluvium. More than 2,000 feet of section is exposed over about 120 square miles or two-fifths of the valley. A series of gravimetric profiles (p. 29) indicate a maximum thickness of more than 5,500 feet of valley fill on a basement profile of moderate relief. An average of about 25 feet of this material is believed to be Quaternary alluvium; the remainder, Tertiary strata. A Montana Power Co.

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test well (State 1–1–22) about 5 miles southwest of Deer Lodge in sec. 22, T. 7 N., R. 10 W. penetrated 2,495 feet of Tertiary strata and bottomed in tuff(?) at 2,536 feet. The upper 300 feet of section is sand and gravel. The remainder, except for some pebble conglomerate between 450 and 860 feet and cobbles(?) between 820 and 860 feet, is interbedded limestone, shale, sandstone, and grit. The clastics are mostly quartz, calcareous detritus, and various percentages of bentonitic clay. About a tenth of the total section, or 235 feet, is made up of 18 beds of limestone and two of dolomite. This calcareous material was apparently derived from Paleozoic and Mesozoic limestone formations, which once may have mantled the Boulder batholith and still partly mantle the Philipsburg batholith and associated stocks in the Flint Creek Range. Granitic debris from either the Boulder or Philipsburg batholith is abundant between 820–860, 1,505–1,515, and 2,115–2,130 feet.

The upper 300 feet of sand and gravel penetrated in this well is middle Pliocene. Because of the ubiquitous presence of bentonite in the deeper deposits, the abrupt change of lithology at 300 feet, and the granitic detritus, the underlying strata are assumed to be older than Pliocene and younger than the middle Eocene(?) batholiths.

At least 1,600 feet of mostly coarse-grained Tertiary strata of extremely varied lithology crops out intermittently along the base of the Flint Creek Range from the vicinity of Warm Springs Creek northward to Robinson Creek (pl. 1). In some places the strata overlie Cretaceous rocks; in other places they apparently underlie Pliocene deposits. Some beds are almost entirely volcanic-rich erosional debris derived from associated Lowland Creek Volcanics; some beds are bentonitic; others contain various percentages of erosional detritus from Precambrian, Paleozoic and Mesozoic sedimentary rocks and middle Eocene(?) granitic rocks. The rocks range from fissile shale to poorly sorted boulder conglomerate with boulders as large as 4 feet in diameter. Some of the beds are so well indurated as to fracture across the grains; some are semiconsolidated and others are unconsolidated. Because the beds have been described elsewhere in considerable detail (Mutch, 1960, and Csejetey, 1962) they are discussed here only briefly.

The northernmost exposure of the 1,600 feet of Tertiary strata is a small patch of cemented colluvium formed of and overyling Cretaceous rocks at the mouth of Robinson Creek canyon in sec. 5, T. 7 N., R. 10 W. More than 200 feet of bentonitic conglomerate and arkose crops out farther to the south between Racetrack Creek and Dempsey Creek. From clay lenses in the conglomerate and arkose (SW¼ sec. 7, T. 6 N., R. 10 W.) were collected specimens of *Equisteum* sp. (Cretaceous to early Tertiary) and *Alnus microdentoides* (identification by Erling Dorf, Princeton Univ.). Both species occur in Oligocene(?) beds in the Missoula Valley about 75 miles to the northwest. *A. microdentoides* is known from only the two localities.

About 250 feet of lime-cemented conglomerate and intercalated lenses of sand crop out between Spring Gulch and Modesty Creek. The poorly exposed basal beds in sec. 6, T. 5 N., R. 10 W. are mostly erosional debris from the underlying volcanic rocks, but the upper beds are entirely erosional debris from Paleozoic rocks immediately to the west. These conglomerates have been tentatively described as Eocene (Mutch, 1960) and Miocene or Pliocene (Csejetey, 1962).

From volcanic-rich, granite-bearing sediments intercalated within the welded tuff south of Lost Creek in sec. 25, T. 5 N., R. 11 W., Csejetey (1962) has collected specimens of five species of fossil plants. All are common to the volcanic-rich beds in the Missoula Valley which were originally described by Douglass (1901, p. 4, 5) as "doubtfully Oligocene." On the basis of these relationships Erling Dorf (in Csejetey, 1962) interpreted the Anaconda flora as "probably of Late Oligocene age, or possibly slightly younger, i.e. early Miocene."

Douglass did not discuss the evidence on which he based his original tentative age designation of beds in the Missoula Valley. However, Jennings (1920, p. 388), after discussing the relationships of the small Missoula Valley flora of 15 species with the not yet satisfactorily dated Florissant flora of 115 species, concluded that "I can see no reason for not accepting Douglass' claim that the beds * * * are of Oligocene age." The much larger Florissant flora was not conclusively dated until many years later after numerous studies had been made and other stratigraphically definitive fossils had been found (MacGintie, 1953, p. 1–198).

If the questionable Missoula Valley Oligocene date is accepted as valid, two formidable weaknesses still remain in assigning a correlative age to the 1,600 feet of Tertiary strata in the Deer Lodge Valley. First, these beds contain a small flora that ranges from Cretaceous to middle or late Tertiary. None of the species are known to be restricted to the time interval in question. It could be argued, therefore, that the specific duplication of the Missoula and Deer Lodge floras is simply a reflection of similar environments rather than stratigraphic correlation. The second weakness is that the radiometric dates of Baadsgaard and others indicate that the Lowland Creek Volcanics and, therefore, the intercalated plant beds are Eocene.

The distribution and extremely varied lithology of the beds may relate to rapid deposition in a wide variety of unstable environments along the base of the ancestral Flint Creek Range. However, without stratigraphically definitive evidence it is impossible to know whether deposition of the 1,600 feet of strata took place almost contemporaneously during a relatively short orogenic interval or whether these beds interfinger with or grade into the coarse Pliocene and Miocene aggregates elsewhere in the valley. The latter would indicate deposition over a much longer interval. Because of the above considerations, it is apparent that their age remains an open question.

About 350 feet of well-bedded, well-consolidated to unconsolidated fluvial clay, silt, sand, and pebble conglomerate of early Miocene age have been measured in various outcrops on the west side of the Deer Lodge Valley north of Mullan Creek. However, because of local deformation and talus cover, definite correlation of most of the individual sections was not possible. The sediments are of variable composition but consist mostly of well-rounded grains of quartzite and granitic rocks. Some of them, near the Garrison Vent, include subangular fragments of locally derived volcanic debris. Metamorphism is not apparent in the Miocene sediments, although it is evident in the underlying Cretaceous rocks. Because of the above relationships the sediments are believed to be younger than the vent. An early Miocene vertebrate fauna (Konizeski, 1957; Wood and Konizeski, 1965) has been collected from these sediments.

North of Willow Creek, interbedded silt, sand, and conglomerate of early Miocene age is overlain by more than 100 feet of bentonitic-rich lacustrine clay and silt of middle or perhaps late Miocene age (Pardee, 1951, p. 81–82). A small patch of semi-indurated sand and conglomerate at the head of Johnson's Gulch on the east side of the valley has been mapped as Miocene(?) (Konizeski, 1957, p. 137). Two small patches of lacustrine deposits of probably Miocene age occur near the northeastern margins of the valley. One of them in the NE¼ sec. 8, T. 7 N., R. 8 W. is plastered on Cretaceous volcanic rocks at an altitude of 5,700 feet, or about the maximum altitude of the high terrace and 1,200 feet above the Clark Fork. The deposit consists of a basal, volcanic-rich conglomerate overlain by interbedded silt, sand, and some volcanic ash. These sediments are well bedded, strike about N. 10° E., and because of primary slumping, dip into the valley at about 20°.

More than 300 feet of unconsolidated to semiconsolidated middle Pliocene sediments underlies most of the high terraces on both sides of the valley. Along the eastern margins of the valley they overlap granitic and volcanic basement rocks; typical exposures occur near the head of Perkins Gulch in sec. 16, T. 5 N., R. 9 W. In the northwestern part of the valley they unconformably overlie lower Miocene sediments. Typical exposures occur in the high terrace scarp along Mullan Creek in sec. 25, T. 8 N., R. 10 W. In the southwestern part of the valley the Pliocene sediments are apparently faulted against and overlie the lower or middle Tertiary beds (fig. 5).

The Pliocene sediments are generally well bedded and usually retain their primary orientation of a few degrees towards the valley axis and to the north. They consist of several depositional types, which grade into and interfinger with each other. Included are colluvial deposits that grade into coarsely bedded piedmont fans of grit and coarse sand which, in turn, interfinger with lamellar flood-plain deposits of fine sand, silt, and clay. In the central parts of the valley the Pliocene deposits commonly consist of interbedded channel gravel and sand (fig. 5). The channel deposits extend towards the mouths of various tributary canyons; for example, from the vicinity of Deer Lodge towards the mouths of Peterson and Cottonwood Creeks.

The distribution of these depositional types of Pliocene sediments is not exposed in any single or continuous series of outcrops. The colluvium and channel deposits are mostly covered by Quaternary alluvium. However, Pliocene colluvium is exposed near the heads of several of the east valley draws, particularly in Perkins and Woodard Gulches. Fan and flood-plain deposits crop out along the high terrace scarps on both sides of the valley. Channel deposits are exposed in the Galen gravel pit (sec. 36, T. 6 N., R. 11 W.) (fig. 6), the Dempsey Creek gravel pit (sec. 20, T. 7 N., R. 9 W.) and the Powell County gravel pit (sec. 18, T. 6 N., R. 9 W.).

The colluvium and fan deposits are of variable lithology in accordance to the source rocks and travel distances to the sites of deposition. The flood-plain deposits are buff colored owing to chemical weathering of included iron-bearing rock and mineral fragments. The flood-plain deposits are mostly silt sized as shown in the following geologic section, but include local accumulations of pebbles, fine sand, and clay.

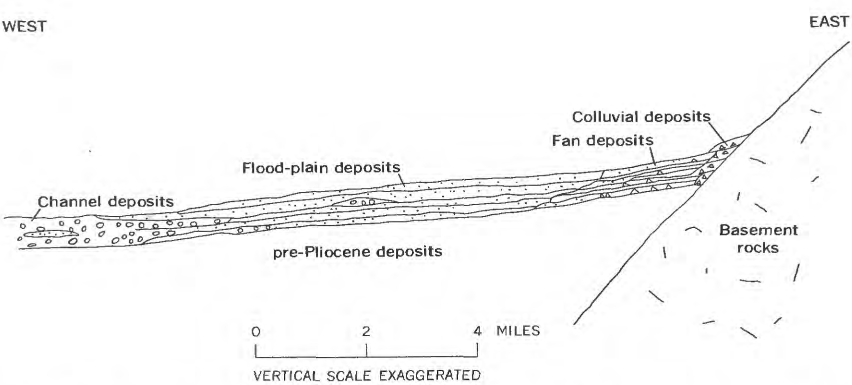

FIGURE 5.—Diagrammatic section across east side of Deer Lodge Valley showing environmental types of Pliocene sediments and their east-west distribution.

FIGURE 6.—Pliocene channel deposits at the Galen gravel pit, sec. 36, T. 6 N., R. 11 W.

Section of Pliocene flood-plain strata measured across the high terrace scarp, 4½ miles northeast of Galen in NE¼ sec. 16, T. 6 N., R. 9 W.

	Thickness (feet)
Silt, buff, finely bedded, semiresistant to weathering; contains as much as 10 percent grit, and pebble lenses 1–2 in. thick and 3–10 ft. long of subangular to subrounded fragments of granitic and volcanic rocks	30
Silt, buff, fine-bedded; contains as much as 5 percent grit	4
Silt, buff, lamellar	4
Silt, buff, lamellar, resistant to weathering; contains 10 percent volcanic ash	4
Silt, buff, lamellar	25
Silt, gray to buff, lamellar, resistant to weathering; contains 25 percent volcanic ash	2
Silt, buff, lamellar	25
Silt, buff, lamellar, resistant to weathering	0. 25
Silt, buff, lamellar	6
Sand and grit, light-buff, crossbedded; contains a few percent of granitic and volcanic pebbles as much as ½ in. in diameter, and fossil vertebrates	4
Covered	5
Silt, buff, finely bedded; contains 15 percent grit, 5 percent subangular to subrounded granitic and volcanic rock pebbles as much as 1½ in. in diameter	6
	115. 25

Lenses of finely bedded volcanic ash are sometimes intercalated between beds of the flood-plain silt. In most instances, for example, near the head of a dry gulch in the SE¼ sec. 29, T. 5 N., R. 9 W., ash overlies a few inches of lamellar clay which, in turn, overlies a few feet of well-rounded, coarse channel gravel and boulders. These relationships indicate a transition from stream channel to lacustrine, apparently cutoff oxbow, environments of deposition.

The channel deposits are well-rounded detritus from the marginal highlands. In some beds the material has an average grain size of more than 4 inches and includes boulders greater than 1 foot in diameter. The deposits are typically crossbedded, well sorted, and unconsolidated, but some, notably in secs. 20, 21, 29, and 32, T. 7 N., R. 9 W., are cemented by a manganese precipitate or by lime.

A small patch of lime-cemented paludal and lacustrine silt is exposed west of Deer Lodge in the NW¼ sec. 32, T. 8 N., R. 9 W. It interfingers laterally with flood-plain silt and overlies channel gravel. More than 100 feet is exposed from the middle of which was collected a single upper molar of *Neohipparion* sp. (Pliocene), USNM 22878. A 2-foot lens of dense lacustrine lime is exposed in gullies tributary to Johnson's Gulch in sec. 3, T. 5 N., R. 9 W. Fresh-water lime (probably of pre-Pliocene age) was penetrated in the Montana Power Co. test hole (State 1–1–22) at about 1,200 feet.

A pair of fragmented lower jaws of *Prosthennops crassigenis* (Pliocene), USNM 22877, was collected during the course of this study from flood-plain silt in the SW¼ sec. 28, T. 6 N., R. 10 W. (Specific identification by C. Lewis Gazin, U.S. Natl. Mus.) Konizeski (1957, p. 142–144) described 111 specimens of 13 species of Pliocene mammals collected mostly from channel, flood-plain, and fan deposits.

QUATERNARY MORAINE DEPOSITS

The glacial deposits of the Flint Creek Range have been described by Calkins and Emmons (1915), Pardee (1951), Alden (1953), Mutch (1960), and Csejetey (1962).

The oldest recognizable Quaternary deposits (older moraine deposits, pl. 1) in the Deer Lodge Valley are a few low mounds of glacial debris, great outwash fans, boulder trains, and occasional large erratics scattered about the high terraces. Low mounds of this material occur on the high terrace about 1 mile east of the mountain front in secs. 21 and 22, T. 8 N., R. 10 W. These unstratified mixtures of deeply weathered boulders and gravel are apparently relics of morainal topography. Extending eastward from them over the high terrace in the vicinity of Marsh Creek are geat glaciofluvial fans, and boulder and gravel trains. These deposits range in thickness from about 30 feet near the head of Marsh Creek to about 10 feet at the end of the terrace. A recent erosion surface cuts the glacial deposits at the base of the mountain front and its west margin is a fault scarp exposing the basement rocks, which are capped by older moraine (fig. 7). A large, deeply weathered granitic erratic, known locally as Sheep Rock, lies about 1 mile east of the mountain front in sec. 21, T. 8 N., R. 10 W. It

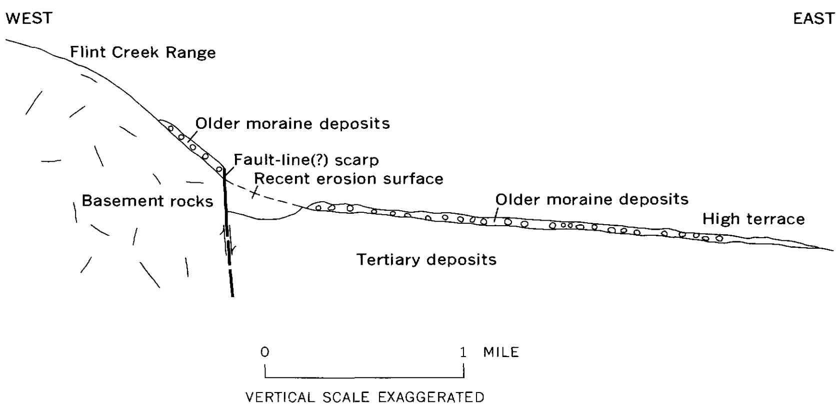

FIGURE 7.—Diagrammatic section showing topographic, age, and structural relationships of older moraine, high terrace, recent erosion surface, and fault-line(?) scarp south of Rock Creek.

is now split into three segments but was originally about 20 feet in diameter.

Pardee (1951, p. 61) described the older moraine as early Pleistocene. Alden (1953, p. 66) proposed a pre-Wisconsin, possibly early Pleistocene age. Mutch (1960) reviewed the age relationships of the drift in considerable detail. He concluded that it is of two ages: an early (pre-Wisconsin) drift similar to that mapped as Buffalo Glaciation in other parts of the Rocky Mountains, and an intermediate (early Wisconsin) drift similar to that mapped as Bull Lake(?) by Poulter (1957) in the Anaconda Range about 20 miles to the southwest.

The older moraine deposits have a more advanced stage of weathering and a more subdued topography than the Wisconsin drift in the area. This indicates that a considerable amount of time and erosion separated the two. Until more positive evidence is produced, the older moraine is perhaps best considered simply as of pre-Wisconsin age.

Terminal moraines of Wisconsin age extend onto the low terraces from the canyons of Rock Creek, Tincup Joe Creek, Dempsey Creek, and Racetrack Creek (pl. 1). The Racetrack moraine is typical. The Racetrack glacier, which was about 14 miles long, flowed eastward down a great gorge and terminated in a broad loop at the base of the mountains. It deposited lateral moraines along the lower reaches of the gorge and a terminal moraine at the mouth. The lateral moraines are about 1½ miles long and 200 feet high. The terminal moraine extends about 1 mile downvalley from the lateral moraines in a broad loop that is about 1½ miles across, 100 feet high, and 5,200 feet above sea level. It is composed mostly of unweathered granitic boulders, some of which are more than 20 feet in diameter. Glaciofluvial fans and gravel trains extend eastward down the broad Racetrack-Dempsey Creek alluvial flat.

No terminal moraines occur at the mouths of Lost Creek, Warm Springs Creek, and Mill Creek, although (1) those drainages were occupied by the greatest ice sheets in the area, (2) their mouths lie at about 5,200 feet above sea level, about the same altitude as the terminal lobe of the Racetrack moraine, and (3) recessional moraines occur in those drainages in areas upstream (and outside) the study area (Calkins and Emmons, 1915, p. 11). The huge coalescent fans of these three streams may be formed of reworked glacial debris derived from Wisconsin terminal moraines. This would explain (1) the great fans, unique in their size and location in the southwestern areas of the valley, (2) the very poor sorting of the fan material near the mouths of the canyons, and (3) the absence of Wisconsin moraines in those areas.

QUATERNARY ALLUVIUM

Almost three-fifths of the Deer Lodge Valley, or 180 square miles, is mantled by Quaternary alluvium. The material is chiefly flood-plain and fan deposits. It has a maximum thickness of about 100 feet at the heads of the Warm Springs Creek, Lost Creek, and Mill Creek fans but has an average thickness of about 20 feet on the Clark Fork flood plain. It generally overlies Pliocene flood-plain silt and clay near the edge of the valley and Pliocene interbedded channel sand and gravel in the central areas. In a few restricted localities, for example, the Willow Creek flat, the Quaternary alluvium has an average depth of less than 10 feet and lies on volcanic basement rocks. The Cottonwood Creek-Fred Burr Creek fan has an average thickness of about 10 feet and overlies Pliocene silt. Along the southeastern margins of the valley, the alluvium consists of a few feet of slope wash derived locally from underlying granitic basement rocks of the Boulder batholith.

The flood-plain alluvium is composed of mixtures and interbedded lenses of gravel, sand, silt, lignitic clay, and caliche capped by carbonaceous soil. These materials were derived primarily from the marginal fans and secondarily by reworking of the underlying and (or) bordering Tertiary sediments.

The fan deposits are of extremely variable composition in accordance with their source, location, and travel distance from source to site of deposition. The coalescent Lost Creek, Warm Springs Creek, and Mill Creek fans consist mostly of interbedded lenses of fine gravel, sand, silt, clay, and carbonaceous soil at the edge, but the grain size rapidly coarsens to predominant boulders and coarse gravel near the mouths of the three canyons. Variation of the lithology of the fan deposits is exemplified by the material in the Tri-City gravel pit near the head of the Mill Creek fan in sec. 8, T. 4 N., R. 10 W., as contrasted with the material in the Pioneer gravel pit near the head of the Clark Fork fan in sec. 36, 4 N., R. 10 W. (table 1).

The glaciofluvial fans below the Wisconsin moraines are composed of mixtures and lenses of sand, gravel, and boulders. Grain size generally diminishes from boulders and coarse gravel near the moraines to fine gravel and sand at the outer margins of the fans. The Cottonwood Creek-Fred Burr Creek fan contains no glacial debris and consists mostly of silt, sand, and gravel capped by a few feet of sandy soil.

Two unique Quaternary deposits occur on high terrace remnants near the mouth of Warm Springs Canyon. The first is north of the canyon where about 3 square miles of uneven erosional surfaces are formed on welded tuff and Tertiary beds. Scattered about the topographic highs and sometimes forming thin sheets of material about one boulder thick are well-rounded, water-worn boulders of Belt

TABLE 1.—*Report on samples of gravel from Deer Lodge Valley, Mont.*

[Submitted by U.S. Dept. of Commerce, Bur. of Public Roads, Washington, D.C., 1955. Values are in percent of size fraction]

Source	Lab. No.	Size fraction (inches)	Quartzite	Granite	Fine-grained volcanics, mostly dacite	Gneiss	Miscellaneous
Tri-City pit NW¼ NE¼ sec. 8, T. 4 N., R. 10 W	92951	2–1½	12	75			13
		1½–1	22	64	4	7	3
		1–¾	18	70	3	5	4
		¾–½	21	65	6	3	5
		½–⅜	21	70	4	4	1
		As received	20	66	4	6	4
Pioneer pit NE¼ SW¼ sec. 36, T. 4 N., R. 10 W	92954	2–1½	94				6
		1½–1	80	1	14		5
		1–¾	78	2	17		3
		¾–½	74		24		2
		½–⅜	58	5	34		3
		As received	78	1	17		4

quartzite. These boulders average about 8 inches in diameter, but some are more than 3 feet. Their source is about 20 miles to the west up Warm Springs Creek. Their occurrence here may be explained in three ways.

1. It has been suggested (F. A. Swenson, oral commun., 1961) that the boulders were transported to and emplaced in their present locale by early-day mill workers to prevent erosion of the old smelter foundations. Lending credence to this hypothesis is the fact that in one area the boulder fields stop abruptly at the boundary between relatively easily erodable sediments, which are the plant beds of Csejetey (1962), and the welded tuff. However, it is difficult to visualize early-day horse-powered transportation of boulders 2 or 3 feet in diameter for distances of 20 or more miles. Furthermore, the material occurs in lesser amounts in many areas where there is no evidence of early mill workings.
2. It may be that the boulder fields are erosional lag deposits. In large areas much well-rounded Belt debris is incorporated within the underlying Tertiary deposits. However, its distribution in the Tertiary deposits does not seem to satisfy the requirements for its distribution in the boulder field.
3. Perhaps the most probable explanation is that the Belt debris had a glacial origin comparable with that of the older moraine in the

Marsh Creek area and was originally deposited on the high terraces about the mouth of Warm Springs Canyon. Interglacial erosion removed most of the other material and left the topographic highs capped by boulders.

The second unique deposit is on the high terrace remnants south of the mouth of Warm Springs Creek canyon. A great bed of travertine slopes gently valleyward and covers about 1 square mile. Outlying patches extend westward about 2 miles above the mouth of the canyon. The travertine averages less than 10 feet in thickness but ranges up to 20 feet and overlies about 10 feet of poorly cemented boulder conglomerate. Numerous impressions of plant stems are encrusted in the travertine, which was formed by springs along the base of the mounttains. The Anaconda Hot Springs still flow and lime is being added to the extensive travertine deposits.

Silt and sludge from the settling basins of the Washoe Smelter (pl. 1) cover about 10 square miles of the Mill Creek, Warm Springs Creek, and Lost Creek alluvial fans. This material has an average thickness of about 20 feet and a maximum thickness of about 75 feet.

REGIONAL STRUCTURE

Most of the sedimentary basement rocks in the mountains around the Deer Lodge Valley have been folded and faulted. Along the front of the Flint Creek Range they are generally folded into a series of anticlines and synclines that trend northeast-southwest. The Range has been described by Pardee (1951, p. 402) as a domal structure that was probably formed by folding rather than faulting. Mutch (1960) agrees with Pardee that the northern part of the Flint Creek Range appears to be arched but describes the Mount Powell fault as evidence to show that the change from broad gentle slopes around the margins of the northern part of the range to abrupt valley-wall relief south of Rock Creek is probably due to recent faulting. The remarkably straight erosion scarp between Rock Creek and Robinson Creek cuts uniformly across basement rocks and unconsolidated drift along its 2½-mile extent (pl. 1, fig. 7) and may be a recent faultline scarp.

INTRAVALLEY STRUCTURE

As indicated by gravity data (p. 29), the minimum altitude of the basement rocks is east of Anaconda and is about 700 feet below sea level. A water well drilled by the Northern Pacific Railroad Co. in Cenozoic fill at Garrison bottomed in basement rocks at about 4,334 feet above sea level; therefore the valley is a closed structural basin.

Most of the exposed lower or middle Tertiary beds and the associated welded tuff unit of the Lowland Creek Volcanics have been greatly deformed by folding, faulting, or slumping. Some are completely overturned. Near the mouth of Warm Springs Canyon they are variously oriented; but between Spring Gulch and Dempsey Creek, there is generally a westward trend in dip (pl. 1).

Much of the Miocene fill is also deformed. Across the north end of the valley, in the area between Marsh Creek and Rock Creek, the dip trends south-southeast (pl. 1). The few patches of exposed Miocene sediments on the east side of the valley are undeformed.

The dip of the exposed Pliocene strata is about parallel to the slope of the high terrace and is probably a function of the original environment of deposition. However, some of the Pliocene strata between Modesty Creek and Spring Gulch are down faulted to the west and dip into the Flint Creek Range (pl. 1). The down faulting and the westward dip of the lower or middle Tertiary strata are probably due to local subsidence along the base of the Flint Creek Range.

Seismic data, recorded by the Montana Power Co., indicate that the Mount Powell and associated faults extend into the valley and cut at least the oldest Tertiary strata. The seismic data also indicate about 300 feet of vertical displacement along the main (normal) fault near the mountain front, but only about 200 feet at a distance of 3 miles into the valley. Only about 100 feet of displacement was recorded along the southern (thrust) fault. As there are no good exposures along the faults, it cannot be determined if the Pliocene strata have been cut. Because there are no surface expressions of the faults within the valley, most movement must have been before the high terraces were formed, although recent movement is indicated at the base of the mountain front.

The Miocene lacustrine deposits plastered onto the lower mountain slopes above intravalley Pliocene fill might be ascribed to post-Miocene, pre-Pliocene erosional sequence, but their position is more logically explained by faulting. The northward lateral migration of cross-terrace ravines (Konizeski, 1957, p. 138) indicates that moderate uplift in the marginal eastern areas of the valley occurred after the deposition of the older drift.

TERTIARY HISTORY

According to Weeks and Klepper (1954, p. 1320–1321), major folding and local thrust faulting in the northern Boulder batholith region east of the Deer Lodge Valley culminated in the very Late Cretaceous or very early Tertiary emplacement of the batholith. "Uplift and erosion during Paleocene and Eocene time resulted in a mountainous terrane and partial deroofing of the batholith. Oligocene gravel and rhyolitic sediments accumulated in subsiding intermontane basins;

tilting of Oligocene basin deposits, partly associated with movement on the range front faults, preceded deposition of stream gravels of Miocene and Pliocene age. Pediments formed in late Tertiary to Pleistocene(?) time."

Ruppel, in a summary of the evolution of landforms in the Basin quadrangle (1957, p. 94, 1963, p. 83), suggests a mature mountainous Oligocene erosion surface of perhaps 3,000 feet of maximum relief and a broad, deeply weathered late Miocene(?) surface that was nearly flat but above which a few rounded mountains rose perhaps 500–1,000 feet. He postulates further that this surface was then uplifted and that Pliocene and Pleistocene erosion carved a landscape similar to that of today.

Smedes (1962, p. 259) depicts deposits of the basal unit of the Lowland Creek Volcanics as accumulating on the Oligocene erosion surface described by Ruppel. If the radiometric date of Baadsgaard is valid, this surface would be early Eocene or older, and erosion of the country rocks and breaching of the batholith progressed at a much faster rate than has formerly been supposed.

The Deer Lodge Valley probably was a major topographic low during deposition of the Lowland Creek Volcanics (early Eocene). The volcanics filled various canyons about the base of the Flint Creek Range and the abrupt relief and orientation of these canyons suggests a major topographic low to the east. The volcanics crop out on the east, west, and south sides of the valley (Smedes, 1962, fig. 1) and probably accumulated in the valley as indicated by a test well (State 1-1-22), which bottoms in tuff(?) at 2,536 feet. Because this well generally penetrated fine-grained sedimentary deposits and some limestone beds, the topographic relief in the valley was probably relatively low during the early to middle Tertiary.

During Miocene time the ancestral Clark Fork may have flowed southward as postulated by Perry (1934, p. 6–7). In late Miocene or early Pliocene the north end of the Deer Lodge Valley was down faulted at least several hundred feet and the Tertiary deposits were faulted and tilted.

In Pliocene time the lithology of the accumulating valley fill indicates regional topographic relief comparable to that now in existence (Konizeski, 1957, p. 147). Coarse gravel and sand were deposited along the channel of the ancestral Clark Fork in the center of the valley; silt and minor amounts of fresh water lime were deposited in the bordering flood plain; and colluvium and slope wash were deposited along the margins of the valley. Intermittent falls of volcanic ash were incorporated in the accumulating erosion detritus. Some of the Pliocene and older deposits were faulted and tilted in the late Pliocene.

QUATERNARY HISTORY

The high terraces are remnants of a valleywide, erosional surface formed during either late Pliocene or early Pleistocene before deposition of the older moraine. This is evidenced by local beveling of middle Pliocene strata and by the capping outwash trains from the older moraine. After deposition of the older moraine but before Wisconsin time, the Clark Fork and its tributaries entrenched themselves about 150 feet into the Tertiary strata and into the basement rocks at the outlet of the valley. The entrenchment was during erosional cycles that perhaps were related to glacial-interglacial periods and (or) intermittent tectonic activity north of the study area. The low terraces were cut during the next to last cycle; the surface beneath the Clark Fork flood-plain alluvium was cut during the post-Wisconsin cycle. Between Mill Creek and Racetrack Creek the high terraces have been mostly beveled to an intermediate level, but some remnants still remain along the mountain front. Great Wisconsin moraines were subsequently deposited on the low terraces at the mouths of the principal tributary canyons north from Racetrack Creek and perhaps at the mouths of Mill, Lost, and Warm Springs Creeks. Large boulder trains have also been deposited on the low terraces below the still fresh Racetrack, Dempsey, Tincup Joe, and Rock Creeks Wisconsin moraines. About 20 feet of Recent alluvium, derived from a variety of sources around the margins of the valley, has accumulated on the Clark Fork flood plain.

Uplift, which occurred after deposition of the older moraine and as late as Recent time, affected the southeastern areas of the valley. The uplift was accompanied and followed by northward migration of tributary streams on the east side of the valley. The recently formed scarp trending north-south at the base of the mountain slopes between Robinson Creek and Mullan Creek may be a faultline feature.

GRAVIMETRIC SURVEY

By E. A. Cremer III

A gravity survey was made in the southern part of the Deer Lodge Valley, Mont., in 1960 to determine approximate depths of valley fill and, if possible, to interpret subsurface features and bedrock configuration. The investigation was made under the auspices of the University of Montana, and the work was supervised by faculty members R. M. Weidman and John Hower. The gravity meter was rented by the U.S. Geological Survey and the University of Montana.

METHODS

Gravity readings were made at more than 300 stations (pl. 2) with a portable, temperature-compensated World-Wide gravity meter. The uncorrected readings are believed to be accurate to the nearest 0.05 mgal (milligal) relative to each other. The observed gravity values and corrections are available at the U.S. Geological Survey office in Helena, Mont. About one-half of the stations were on three traverses across the valley and three short trunklines in the southwest part of the area. Stations on the cross-valley traverses were about one-fourth mile apart; those on the short lines were about one-eighth mile apart. The other one-half of the stations were randomly scattered throughout the area where altitude data were available. Altitudes at the stations were determined by instrumental leveling and are generally accurate to the nearest foot. To determine the instrumental drift, an additional reading was made at a station where a reading was taken one or two hours before.

REDUCTION OF DATA

The observed gravity values were corrected for drift, then the standard free-air, Bouguer, and latitude corrections were made as described by Nettleton (1960, p. 41–58). An assumed density of 2.5 was used for the Bouguer correction.

Terrain corrections using tables published by Hammer (1939) were made for 28 selected stations throughout the area and interpolated values were applied to all other stations.

By comparing gravity values of stations on basement rock, a regional gravitational gradient was found to be representable as a flat surface sloping to the southwest. This surface was mapped and superimposed on the overall gravity map. The appropriate value was then subtracted from each station thus leaving a residual gravity map showing the anomaly caused by varying depths of fill (pl. 2). The maximum regional correction for the entire area was 29.9 mgal; however, the maximum correction for the area covered by the residual gravity map was only 13.7 mgal. This correction limits the precision of the final values but is probably sufficiently accurate to justify the 2-mgal interval.

RESULTS

Without other information, there is no unique interpretation of gravity data. In this gravimetric survey, surface and subsurface geology aided the interpretation of the gravity data.

The residual gravity map (pl. 2) shows a rapid decrease in gravity along the sides and southern end of the valley. The values in the central part of the valley are comparatively constant. A gravity low in the southern part of the area has about 20 mgal of residual gravity relief relative to bedrock areas. The low basement rock values in the western part of T. 5 N. could be caused by low-density volcanic tuffs (R. L. Konizeski, written commun., 1960).

Three east-west sections (pl. 2) were calculated by using a graticule as described by Nettleton (1940, p. 115) to convert corrected gravity readings to depth to basement rock. An assumed density contrast of 0.6 gram per cubic centimeter between the valley fill and basement rocks was used. The sections were prepared to determine the approximate thickness of valley fill and to help interpret structural relationships.

The outstanding feature of the three sections (pl. 2) is a sudden, large break in slope at the eastern end of each. This almost certainly represents a major fault along the foot of the mountains on the east side of the valley. This fault is probably a continuation of a known fault southeast of the valley (R. L. Konizeski, written commun., 1960). There is also definite indication of a smaller fault at the western end of section A–A′ and, to a lesser degree, of section B–B′. The magnitude of the break, the sharpness of the subsurface features, and the depth of valley probably preclude the possibility of the anomaly being caused by a terrace. Section C–C′ did not extend far enough west to cross a northern projection of the known Mount Powell fault that extends into the valley between sections B–B′ and C–C′ and bends northward along the foot of the Flint Creek Range. Because the valley fill is thicker at the western end of the section, which is near the mountains, than it is towards the center of the valley, the Mount Powell fault probably extends at least as far north as section C–C′. The above geophysical evidence indicates the valley was formed by faulting.

An ambiguous feature shown by the three sections is the bedrock high in the middle of the valley. This may be attributed to faulting of the valley floor or erosion but is more likely caused by lava flows down the center of the valley. A flow of lava or a series of interbedded flows of less thickness than the indicated bedrock high, but closer to the land surface, could result in a calculated valley floor similar to that shown. Andesite flows at the north and south ends of the valley give credence to the explanation. The tight lines in the vicinity of sec. 21, T. 4 N., R. 10 W. (pl. 2) are caused by an andesite flow.

The decreasing thickness of the valley fill from more than 5,500 feet in the south to about 2,300 feet near Deer Lodge indicates that at one time drainage from the area may have been to the south as postulated by Perry (1934, p. 6–7).

SUMMARY

There is a large fault along the eastern side of the Deer Lodge Valley and probably a smaller one along the western side. These faults indicate a tectonic origin of the Deer Lodge Valley. The maximum depth of fill is more than 5,500 feet east of Anaconda. The depth of fill decreases to the north.

GROUND WATER

Scientific studies have shown that (1) practically all ground water is derived from precipitation, (2) most usable ground water is an important component of the hydrologic cycle, (3) ground water obeys natural laws, and (4) the occurrence of ground water is intimately associated with the geology of the area.

DEFINITION OF SELECTED HYDROLOGIC TERMS

The following definitions are based largely on those given by Meinzer (1923b). A few terms not included in the following list are defined where they are introduced in the text.

Aquifer, a formation, group of formations, or part of a formation that will yield ground water in useful quantities.

Artesian aquifer, a confined *aquifer* in which ground water rises in a well above the point at which it is found in the aquifer.

Confining bed, a bed which overlies an *aquifer* and which, because of its low permeability relative to the aquifer, prevents or impedes upward loss of water and pressure; a similar bed beneath an aquifer that prevents or impedes downward loss of water and pressure.

Drawdown, lowering of the water level in a well as the result of pumping.

Effluent flow, flow of water from the ground-water reservoir to surface water.

Evapotranspiration, the combined discharge of water to the air by direct evaporation and plant transpiration.

Flowing well, an artesian well through which water is forced above the land surface by pressure in the *aquifer.*

Influent flow, flow of water into the ground-water reservoir from surface water.

Permeability, a measure of the capacity of an *aquifer* to transmit water.

Permeability, field coefficient of, the rate of flow of water, in gallons per day under prevailing conditions, through a cross section of *aquifer* 1 foot high and 1 mile wide, under a hydraulic gradient of 1 foot per mile.

Piezometric surface, an imaginary surface that everywhere coincides with the static head of water in an *aquifer.*

Porosity, the ratio of the volume of the openings in a rock to the total volume of the rock.

Recovery, the residual *drawdown* after pumping has stopped.

Specific capacity, a measure of the productivity of a well; the amount of water, in gallons per minute, that is yielded per foot of *drawdown.*

Storage, coefficient of, a measure of the capacity of an *aquifer* to store and release water; the volume of water released from or taken into storage per unit surface area of the aquifer per unit change in the component of head normal to that surface.

Transmissibility, coefficient of, the rate of flow of water, in gallons per day under prevailing conditions, across each mile strip extending the saturated thickness of the *aquifer*, under a hydraulic gradient of 1 foot per mile. It is equal to the *field coefficient of permeability* multiplied by the saturated thickness of the aquifer, in feet.

Water table, the surface within the *zone of saturation* where the pressure is atmospheric.

Water-table aquifer, an *aquifer* which is not confined above. In this type of aquifer the water level in a well indicates the *water table.*

Zone of aeration, the zone in which the open spaces in the rocks are filled with air and water.

Zone of saturation, the zone in which the open spaces in the rocks are completely filled with water.

PRINCIPLES OF OCCURRENCE

In the Deer Lodge Valley most of the ground water occurs in the pore spaces between the grains of the Quaternary and Tertiary sediments; some occurs in the joints and fractures of the indurated volcanic rocks. Part of the water from precipitation, from irrigation, and from influent streams seeps into the soil and percolates downward through the zone of aeration to the zone of saturation. In the zone of saturation the open spaces, which are generally interconnected, act as conduits through which ground water moves. The rate of movement of ground water is measured in feet per day or feet per year, whereas that of surface water is measured in feet per second. Stored ground water seeps into stream channels to maintain streamflow throughout the year.

Ground water is discharged by springs, wells, and effluent streams, or by evaporation and transpiration (evapotranspiration). Under natural conditions the discharge from the reservoir was equal to the recharge. Irrigation of agricultural land and industrial development in the valley changed the natural conditions. However, the change was far enough in the past that a new balance between recharge and discharge has been fairly well established.

The geology determines whether the water is artesian (confined) or water table (unconfined). In the Deer Lodge Valley, water in the Tertiary rocks is generally confined, but water in the Quaternary rocks is generally unconfined. The geology also determines to a great extent the depth to water in the underground reservoir. In some places in the valley, as in seeps and swamps, the water is at the land surface. In the flood-plain alluvium of the Clark Fork and some of its tributaries, the water table is within 5 to 10 feet of the land surface. In some of the alluvial fans and under most of the terraces the water table may be from 10 to more than 150 feet below the surface. Depths of wells in the valley range from a few feet to 250 feet.

HYDROLOGIC PROPERTIES OF WATER-BEARING MATERIALS

The hydrologic properties of an aquifer include the ability to store and to transmit water, as measured by the porosity and by the coefficients of storage (S), permeability, and transmissibility (T). These properties control (1) the ability of the aquifer to take water into storage and release it from storage and (2) the movement of water through the aquifer from the area of intake to the area of discharge.

The coefficient of transmissibility of the Quaternary alluvium and Tertiary sediments was obtained in several places in the Deer Lodge Valley by aquifer tests. Pumping water from a well develops a cone of depression in the water table around the pumped well. The base of the cone is the static water level and the height is equal to the drawdown in the well. The drawdown in the pumped well depends upon the pumping rate, the time since pumping began, the transmissibility and storage properties of the aquifer, and the well construction. The area affected by pumping (area of influence) depends upon the pumping rate, the time since pumping began, and the transmissibility and storage properties of the aquifer. The cone of depression around a pumped well increases in depth and area until it intercepts enough rejected recharge or reduces natural discharge to equal the amount of water being pumped from the well. After pumping is stopped, the water table surrounding the well will eventually return almost to its original position. By pumping a well at a known constant rate for a measured time and by recording drawdown during pumping and recovery after pumping, enough data can be collected to use in formulas developed by Theis (1935, p. 519–524) and by Jacob (1947, p. 1047–1070) for computing the coefficients of transmissibility and storage. The coefficients of transmissibility and storage can be used with other data to (1) estimate yield and drawdown for proposed wells, (2) determine the amount of ground water flowing through an aquifier, and (3) determine the rate at which ground water is moving.

MULTIPLE-WELL TESTS

An aquifer test using the pumped well and one or more observation wells is termed a multiple-well test. The coefficient of transmissibility of the aquifer at three sites was determined from this type of test (table 2). At one test site, the observation well was a domestic well; at the other two sites, temporary observation wells were installed by driving a ¾-inch pipe into the ground. This type of test setup is not ideal for testing an aquifer because the aquifer thickness at the observation well is not known and generally the test well only partly penetrates the aquifer. However, data obtained from the tests are worth-

TABLE 2.—*Aquifer test data*

[Geologic source: A, Quaternary alluvium; T, Tertiary sediments]

Well	Geologic source	Depth of well (feet)	Pumping rate (gpm)	Drawdown in pumped well (feet)	Length of test (minutes)	Specific capacity (gpm per ft)	Coefficient of transmissibility (gpd per ft)	Remarks
B4-10- 5ab	A, T	160	61	----------	30	----------	95,000	Casing too wet to measure water level while pumping.
15aa	A	23.1	11	10.6	470	1	20,000	Large entrance loss into casing. Observation well 40 ft away.
B5-10-17dc	A(?)	71.1	270	29.5	1,800	9	77,000	Large entrance loss. Observation well 550 ft away.
23cb	A, T	150	580	38.2	450	15	20,000	
24aa	T	200	975	102.3	220	10	15,000	Screened and gravel packed.
B6- 9- 4ab	A	18.1	28	4.1	300	7	80,000	Large entrance loss into casing.
4bb	A	14.3	27	.8	200	34	40,000	
4cc	A	8.0	10	1.8	180	6	25,000	
7ad2	A	35	92	3.8	200	24	35,000	
7dc	A	40	110	2.2	250	50	175,000	Observation well 50 ft away.
B6-10-14bc	A, T	126	85	3.0	300	28	80,000	
15aa	A	12.6	30	2.5	300	12	35,000	
B7- 9-17ad	T		135	7.2	30	19	38,000	
32da	A	34	22	2.7	150	8	60,000	
B8- 9-15ab	T(?)	19.1	9	4.4	200	2	800	
15ba1	A	23.6	55	2.0	350	28	50,000	
21da1	A	14.4	25	.7	240	36	50,000	
21da2	A	12.5	52	2.6	330	20	35,000	
27ac	T(?)	11.2	5	1.6	120	3	1,000	
B9- 9-34cc	T	17.4	3	2.0	120	1	600	

while when used with the knowledge of the conditions under which the tests were made. Well B6–9–7dc fully penetrates the aquifer and is perforated and developed; the saturated thickness did not appreciably decrease during the test, and the observation well was about twice the saturated thickness (m) from the pumped well. Therefore, results are believed to be fairly good. Neither well B4–10–15aa nor B5–10–17dc fully penetrates the aquifer. The drawdown in the observation well at test site B4–10–15aa was affected by the partial penetration of the pumped well, but no correction could be made because the saturated thickness of the aquifer was not known. Because the observation well at test site B5–10–17dc was not affected by partial penetration, no correction was required. The maximum drawdown during the test on wells B4–10–15aa and B5–10–17dc was 10.6 feet and 29.5 feet respectively; the theoretical drawdown in a well with an effective radius of 1 foot (based on the calculated values of T and an estimated value of S of 0.05) was less than 1 foot and 5 feet, respectively. Therefore, most of the measured drawdown in the pumped wells was due to entrance loss, and the saturated thickness probably did not decrease sufficiently to necessitate correction of drawdowns. Water from a stream about 200 feet south of well B4–10–15aa apparently decreased the rate of drawdown in the observation well.

In the development of Theis's formulas, it was assumed that the coefficient of storage is constant and that water is instantaneously released from storage as the head declines. In tests made under water-table conditions the water is derived from storage by gravity drainage, which is generally slow. Thus, the coefficient of storage as calculated from the aquifer test data appears to increase as pumping time increases and drainage is more complete. Because complete drainage of the rocks in the cone of depression might require several weeks or months of pumping, the coefficient of storage was not determined.

SINGLE-WELL TESTS

An aquifier test using only the pumped well is termed a single-well test. Theis (1935, p. 522) stated, "If a well is pumped for a known period and then left to recover, the residual drawdown at any instant will be the same as if pumping of the the well had been continued but a recharge well with the same flow had been introduced at the same point at the instant pumping stopped." Therefore, the coefficient of transmissibility can be determined from the rate of recovery (as well as the rate of drawdown) of the water level in a well. Todd (1959, p. 97, footnote) stated, "An advantage of a recovery analysis of a pumped well is that * * * it implies a constant discharge Q, which often is difficult to control accurately in the field." The coefficient of transmissibility of the aquifer near 17 wells was determined from single-well tests (table 2), page 33. On five of the tests, computations were made from the rate of drawdown; on another five, computations were made from the rate of recovery; and on the remaining seven, computations were made from both the rate of drawdown and the rate of recovery (values obtained on each well by both methods agreed reasonably well). Testing aquifers by this method is not ideal, but much useful information can be obtained.

Generally the storage coefficient cannot be determined from rate of drawdown or recovery in the well being pumped, because the effective radius of the well is not known. The coefficient of storage was not determined from any of these tests.

SPECIFIC CAPACITY

The specific capacity of a well is computed by dividing the discharge from a well by the drawdown. The specific capacity is a function of the coefficients of transmissibility and storage and, therefore, indicates the hydrologic properties of the aquifer. Because drawdown is not only dependent on the water-yielding properties of the aquifer but also on the well construction and development, the relation between specific capacity and coefficient of transmissibility may differ from well to well (table 2). Actual drawdown is greater than the theoretical drawdown in the best designed and constructed wells and may be several times as great in poorly constructed and poorly developed wells. The specific capacity decreases with increasing discharge and with increasing time. Therefore, using the specific capacity determined from a short pumping test to predict drawdown after a long period of pumping and (or) after pumping at a different rate could cause some large errors.

Theis and others (1963, p. 331–340) outlined a method for estimating the transmissibility of a water-table aquifer near a well of known specific capacity. They noted many limitations of the method but also indicated its usefulness where no other data are available. This method was used in the Deer Lodge Valley to check 19 of the values of the coefficient of transmissibility obtained from aquifer tests and to estimate the coefficient of transmissibility in the vicinity of 11 other wells (table 3). Of the 19 values checked, four were the same as

TABLE 3.—*Coefficient of transmissibility values estimated from specific capacities*

[Remarks: T, Tertiary sediments; A, Quaternary alluvium]

Well	Specific capacity (gpm per ft)	Coefficient of transmissibility (gpd per ft)	Remarks
B4- 9-31cb	0.6	1,000	T.
10-10ac	3	5,000	T(?).
17bd	13	20,000	A, T(?). Perforated and developed.
11-1cb	1	1,000	T.
B5-11-33ca1	50	85,000	A. Perforated and developed.
B6-10- 5aa1	26	40,000	A, T. Perforated and developed.
B7- 9- 4ba	40	70,000	T. Screened and developed.
20cc	13	20,000	T.
10- 2cd	2	2,000	T.
B8- 9-33cc	30	50,000	T. Screened and developed.
10-23db	.5	800	T.

those obtained from pumping tests, six were reasonably close, and nine were much lower. The nine were low because they were from specific capacities of small-diameter wells with unperforated casings that have large entrance losses. The estimated values (table 3) for perforated or screened and developed wells are probably correct to within 25 percent; those values of 1,000 or less are probably correct to within an order of magnitude; and the others are probably somewhat low.

SUMMARY OF HYDROLOGIC PROPERTIES

The principal source of ground water in the Deer Lodge Valley is the upper few hundred feet of unconsolidated and semiconsolidated valley fill of Quaternary and Tertiary age. The consolidated rocks of the surrounding mountains are barriers to ground-water movement and form boundaries along the sides of the ground-water basin.

The alluvium beneath the flood plain of the Clark Fork consists chiefly of medium- to well-sorted unconsolidated gravel. At seven aquifer test sites (wells B6–9–4ab, B6–9–4bb, B6–9–4cc, B7–9–32da, B8–9–15ba1, B8–9–21da1, and B8–9–21da2) in the alluvium the coefficient of transmissibility ranged from 25,000 to 80,000 gpd per ft (gallons per day per foot). Specific capacities ranged from 6 to 36 gpm per ft (gallons per minute per foot) of drawdown. The flood-plain alluvium is relatively thin and is not extensive laterally; therefore, the yield of shallow alluvial wells is limited to a few hundred gallons per minute even in areas of high transmissibility.

No wells were found in the moraine deposits, so that their hydrologic properties are not known. However, the lithology and texture of the moraine deposits indicate that the transmissibility would generally be much less than that of the flood-plain alluvium.

The terrace alluvium on the west side of the valley north of Dempsey Creek is fairly thin and most of the wells obtain water from the underlying Tertiary sediments. However, a few shallow dug wells tapping the alluvium along draws yields adequate water for stock and domestic use.

The alluvium south of Dempsey Creek and west of the Clark Fork flood plain is mostly fan material and generally much thicker than the alluvium underlying the low terrace. Adequate water for domestic and stock needs can be obtained from these deposits. Sufficient water for industrial or irrigation needs can be obtained locally. Transmissibility ranged from 20,000 to 175,000 gpd per ft for the alluvium in this area as determined near five wells (B4–10–15aa1, B5–10–17dc, B6–9–7ad2, B6–9–7dc, and B6–10–15aa) and estimated from the specific capacity of two wells (B4–10–11bd and B5–11–33ca1). The variation is due to differences in thickness and in lithology. Transmissibility is 95,000, 20,000, and 80,000 gpd per ft near three wells tapping the Quaternary alluvium and Tertiary sediments. Data indicate that much of the water comes from the alluvium. A transmissibility of 40,000 gpd per ft was estimated from a reported specific capacity of 26 gpm per ft of drawdown for an irrigation well (B6–10–5aa1) at the foot of the Dempsey Creek moraine. The owner's report indicates that the well is probably open to both alluvium and Tertiary sediments, but most of the water comes from the alluvium. Two other wells drawing water from both the Quaternary and Tertiary aquifers between Warm Springs Creek and Mill Creek reportedly yielded 800 and 1,200 gpm.

The relatively small deposit of alluvium east of the Clark Fork flood plain consists mostly of fan material and of detrital material, which was formed in place or transported a short distance. No aquifer tests were made in this deposit. Some water is obtained from shallow dug wells in the fan material of Cottonwood Creek and a few of the other east-side tributaries. Yields of these wells are reported to be small and many of them go dry in early spring. No wells were found in the detrital material (alluvial slope wash) in the southeast part of the valley. As recharge to this area is only from precipitation, it is doubtful that any large-yield wells could be obtained. Adequate water probably could be obtained locally for stock or domestic use.

The Tertiary sediments that yield water to wells are mostly unconsolidated to semiconsolidated fluvial silt and interfingering beds of unconsolidated sand and gravel. These sediments are generally less permeable than the alluvium but have a much greater thickness. Transmissibility ranged from 600 to 70,000 gpd per ft as determined near five wells (B5–10–24aa, B7–9–17ad, B8–9–15ab, B8–9–27ac, and B9–9–34cc) and estimated from specific capacities for 8 wells (B4–9–31cb, B4–10–10ac, B4–11–1cb, B7–9–4ba, B7–9–20cc, B7–10–2cd, B8–9–33cc, and B8–10–23db). Specific capacities ranged from less than 1 to 40 gpm per ft of drawdown. Eight of the 13 specific capacities were less than 10 gpm per ft of drawdown; only two were more than 20 gpm per ft and both of these were reported specific capacities on the Deer Lodge municipal wells. Geologic conditions in the vicinity of Deer Lodge and the well construction indicate that a high specific capacity is possible. One of the wells reportedly was tested at 1,300 gpm with 32 feet of drawdown after 24 hours of pumping.

WATER TABLE AND MOVEMENT OF GROUND WATER

The water table is an irregular sloping surface that conforms roughly to the topography and rises and falls as the aquifer is recharged and discharged. The water table has many irregularities that are caused by local differences in permeability and local and seasonal differences in withdrawals from, or additions to, the reservoir. For a given rate of ground-water flow, the water-table slope will be relatively gentle if the transmitting material is coarse and permeable, such as a clean sand and gravel; the slope will be steeper if the material is fine grained and less permeable, such as a fine sand or silty sand and gravel.

The slope of the water table and the direction of movement of ground water were determined from a water-table contour map made from measurements in June 1960 of the water levels in 159 wells. Most of the wells in the northern part of the area are in a narrow belt along the Clark Fork; in the southern part they are in a wider belt but are mostly on the west side of the Clark Fork. Therefore, the water-table contours could only be drawn for part of the area. With such scant data many of the details of the shape and slope of the water table could not be shown even in the area contoured (pl. 1) Ground water moves downslope perpendicularly to the contours. The water-table contour map shows that ground water moves generally toward the Clark Fork, but in detail the direction of movement varies. In the southwest part of the valley, ground water moves generally northeastward toward the Clark Fork; but it has a component of flow toward Warm Springs, Dutchman, Mill, and Willow Creeks. The slope of the water table in this part of the area averages about 50 feet per mile. Between Racetrack and Lost Creeks, the slope of the water table is generally eastward and averages about 70 feet per mile. Data are insufficient to contour the water table on the east side of the river or in the northwestern section except for a narrow strip along the Clark Fork in the northern part. The topography and the dip of the Tertiary beds indicate that ground water east of the river moves northwestward and ground water in the northwestern section moves eastward.

The quantity of water that flows through a cross section of the aquifer can be estimated by Darcy's equation written in the form (Ferris and others, 1962, p. 73):

$$Q = TIL \qquad (1)$$

where

Q=discharge in gallons per day,
T=coefficient of transmissibility in gallons per day per foot,
I=hydraulic gradient in feet per foot, and
L=width in feet of the cross section through which the discharge occurs.

From a cross section along the 5,050-foot contour line on the water table (pl. 1), the following data were obtained. The length of the section across the coalescent fans between Warm Springs and Mill Creeks is 14,000 feet and the slope of the water table at the section is 0.01 foot per foot. The coefficient of transmissibility as determined from an aquifer test on well B4–10–5ab is about 95,000 gpd per ft. Substituting these values in equation 1:

$$Q = 95{,}000 \times 0.01 \times 14{,}000$$
$$= 13{,}000{,}000 \text{ gpd.}$$

The velocity at which ground water moves can be determined by use of the equation:

$$v=\frac{Q}{7.48\,\theta\,A}$$

where

v=velocity in feet per day,
Q=flow through the cross section in gallons per day,
θ=porosity expressed as a decimal, and
A=cross-sectional area in square feet.

Assuming a thickness m of 100 feet, the area of the cross section is 1,400,000 (Lm=14,000×100) square feet. The estimated porosity is 40 percent. Substituting these values in equation 2, the velocity

$$v=\frac{13{,}000{,}000}{7.48\times 0.4\times 1{,}400{,}000}$$
$$=3 \text{ feet per day.}$$

Lower values of transmissibility more than offset steeper gradients in other parts of the study area; therefore, the ground-water velocity is less than 3 feet per day.

FLUCTUATIONS OF WATER LEVEL

The water table is a dynamic surface. Many factors cause it to rise or fall. The most important factor causing water-level fluctuations in the Deer Lodge Valley is change in volume of water in the ground-water reservoir. The amount and rate of fluctuation depends on the rate of loss or rate of replenishment of water in storage. Ground water is discharged by seepage into streams, by evaporation and transpiration (generally along streams), and by pumping from wells. This discharge gradually lowers the water table except when exceeded by recharge to the underground reservoir from snowmelt, precipitation, and seepage from streams and irrigation.

A record of water-level fluctuations furnishes valuable information about discharge, recharge, and storage. Monthly fluctuations of the water table in the Deer Lodge Valley were measured in about 40 wells in the northern part of the valley starting in the fall of 1957 and in about 45 wells in the southern part of the valley starting in the summer of 1960. Monthly measurements were continued in most of these wells until the fall of 1961, when two measurements per year (March and September) were made on about 70 of the wells. These measurements will be continued for a period of years to monitor long-term fluctuations.

Water-level fluctuations in the valley may be placed in two general groups—seasonal and long term. Seasonal fluctuations are an index of change in quantity of water in storage during the year and long-term fluctuations indicate trends over a period of years.

Seasonal fluctuations indicate the amount of water taken into or released from storage. In general, the water level in the valley declines in the winter and early spring, then rises in late spring and summer. Hydrographs indicate that location, depth to water, geologic setting, and climatic variations affect the seasonal fluctuations (fig. 8). The time of the annual peak varies in different parts of the valley and sometimes varies for different years in the same part of the valley.

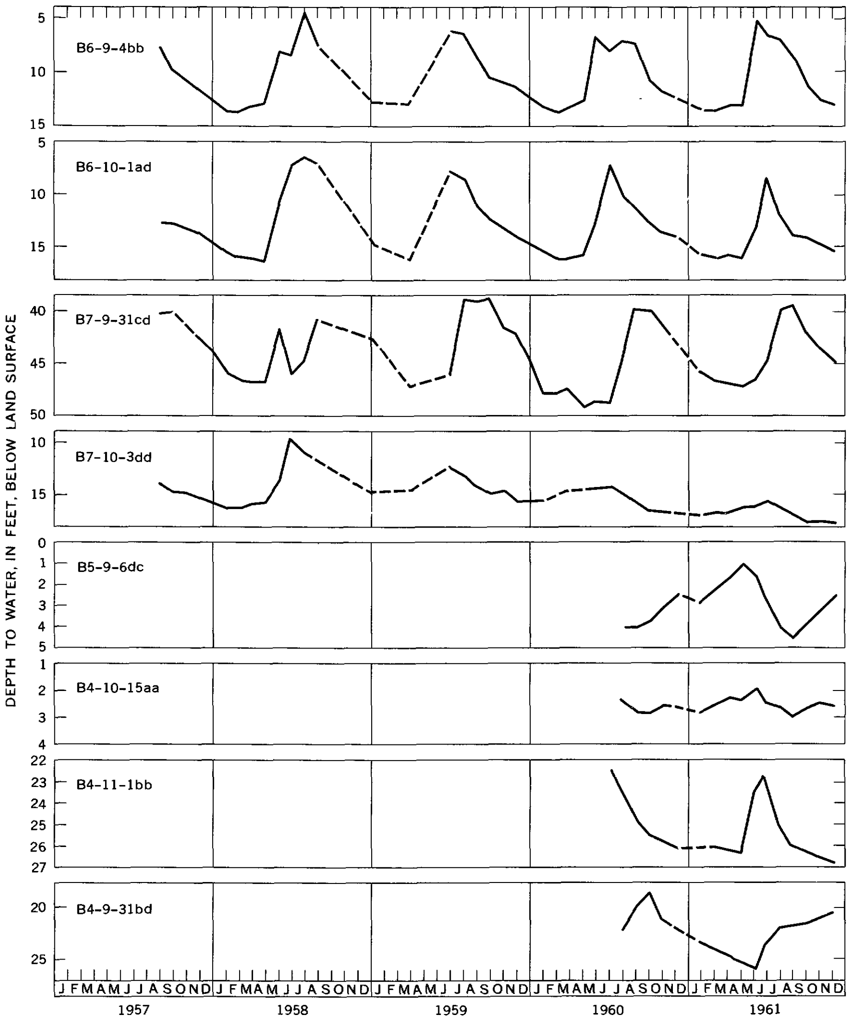

FIGURE 8.—Effect of location, depth to water, and geologic setting on seasonal fluctuations of the water level.

During the period of record, a few wells reached their annual peak in April or May; most of the wells reached the annual peak in June or July; several, in August or September; and a few, in October or November.

Well B6–9–4bb (fig. 8) is in the alluvium, and wells B6–10–1ad and B7–9–31db are in Tertiary sediments on the west side of the valley. The difference in the time of the annual peak is caused mostly by (1) the difference in depth to water in the wells, (2) the difference in sediment size and bedding of the material between the source of recharge and the water table, and (3) different amounts of local irrigation. Generally, water levels in the Tertiary sediments peak later than water levels in the alluvium. The hydrograph for well B5–9–6dc shows the water level rising from September through April and declining from May through September. This is fairly typical of a few shallow observation wells in areas of high water table and is due to evapotranspiration. At this particular well, alkali also indicates considerable evaporation. Water-level flunctuations in well B4–10–15aa also show the effect of evapotranspiration. In addition, they show that recharge during May was more than enough to offset the discharge by evapotranspiration and underflow. Well B4–11–1bb is near Warm Springs Creek, and recharge from spring runoff causes a rapid rise in the water level in May and a continued rise in June. Upstream storage and utilization of almost the entire low flow of Warm Springs Creek causes a rapid decline in July and August. Recharge is about equal to discharge from November through April. Well B4–9–31bd is artesian and reflects the local fluctuation of the piezometric surface of the basement rocks. The fairly constant rate of decline from October through May is not general throughout the valley.

Long-term fluctuations in well B7–10–3dd, which taps Tertiary sediments, show a continued and substantial decline in the annual peak. The decline is probably caused mostly by a decrease in the amount of irrigation in the vicinity because less than 20 acre-feet of water per year (operator's estimate) are pumped from the only well in the vicinity.

Water-level fluctuations in 73 wells were used to estimate the monthly change in ground-water storage for the period June through September 1961. The valley fill was divided by Thiessen polygons (Thiessen, 1911, 1082–1084) so that each polygon contained only one well and any point in the polygon was nearer to the enclosed well than any other well. The areas of the polygons were determined by a planimeter. The measured monthly change in water level in each well was multiplied by the area of the polygon encompassing the well, and the product was considered to be the bulk volume of material saturated or drained within the polygon during the month. The sum of the volumes for all polygonal areas in the valley was considered to be the total bulk volume of material saturated or drained. Monthly changes in the volume of ground water in storage were computed by multiplying the monthly changes in volume of saturated material by an estimated specific yield (0.1) of the material. These changes in storage are discussed under the section on discharge.

RECHARGE

The Deer Lodge Valley ground-water reservoir is recharged by infiltration of water from irrigation, precipitation and snowmelt runoff, and influent streams. The Clark Fork is normally an effluent stream, but for a short period during the spring runoff it rises high enough so that some water recharges the ground-water reservoir. Much of the water, which is diverted from the Clark Fork to irrigate 5,000 acres of land in the valley, percolates into the ground-water reservoir. Most of the major tributaries to the Clark Fork in the valley are perennial in their upper reaches, but during the summer the water is all diverted, leaving the streams dry for much of their course across the valley. Some of the streams flow in their lower reaches. From June to September 1961 about 16,000 acre-feet of water was diverted from Warm Springs Creek for industrial use. Total flow in the creek for the same period was about 24,000 acre-feet. Total flow in the other tributary streams was about 50,000 acre-feet, most of which was diverted for the irrigation of 30,000 acres of land. In addition, about 13,000 acre-feet of water was imported to the area from Rock Creek to irrigate about 4,000 acres of land. Some of the water diverted for industrial use and much of the water diverted for irrrigation percolates to the ground-water reservoir. Most of the tributary streams are influent for at least part of their course across the valley; therefore, when all the water is not being diverted, they become effective sources of recharge to the underlying and adjacent sediments.

Recharge from precipitation and snowmelt is governed by amount, distribution and intensity of precipitation; topography; permeability and moisture-holding capacity of the surficial deposits; consumptive use through evapotranspiration; and the capacity of the ground-water reservoir to store additional water. Probably very little of the precipitation and only a small amount of the snowmelt is recharged in the nonirrigated part of the valley; most of the available moisture in that area is transpired by plants or evaporated. The average annual precipitation is about 10 or 11 inches. Almost one-third of the total normally is in May and June, which are also the months of high recharge from irrigation, so that it is impossible to tell from the hydrographs how much of the recharge is from precipitation and snowmelt. It is probably small in comparison to the amount of recharge from irrigation. Estimates made from U.S Weather Bureau records indicate that precipitation adds about 150,000 acre-feet of water per year to the valley. Estimates made from water-level fluctuations (p. 41, 44), indicate a net increase in ground-water storage of about 14,000 acre-feet in June 1961 and about 1,500 acre-feet in July 1961.

DISCHARGE

Ground water is discharged from the Deer Lodge Valley by effluent seepage into streams, drains, springs, and seeps; by evaporation and transpiration; and by pumping water from wells. At places where the water table intersects the land surface, ground water is discharged by effluent seepage. During most of the year the Clark Fork is an effluent stream throughout the valley. Some of the tributary streams (notably Lost, Dutchman, Warm Springs, and Willow Creeks) are effluent in their lower reaches. Streamflow measurements made April 25, 1961, showed inflow to the area was about 190 cfs and outflow from the area was about 330 cfs. Ground-water discharge to the streams accounted for more than 40 percent of the total outflow from the area on that day. Measurements of the flow in Lost, Dutchman, Warm Springs, and Willow Creek during late April and early May indicated that more than 70 cfs of ground water was being discharged into these streams during that period.

Much of the ground-water discharge into Willow Creek (20 cfs on May 19, 1961) was from an 8-mile tile drainage system in the vicinity of Opportunity. The system was installed about 1913 and much of it is no longer effective. Local residents reported that many of the 12-inch main lines carried about 100 inches (2.5 cfs) of water for many years after the system was installed. One outlet measured May 19, 1961, was discharging 1.7 cfs. Most of the other outlets discharge directly into Willow Creek or into short tributaries of Willow Creek that probably flow because of water surfacing as a result of a plugged drain.

Farmers and ranchers in the valley have dug about 25 drainage ditches. The drains have a total length of about 9 miles and discharge an estimated 28 cfs of water into the Clark Fork through tributary stream channels (H. N. Smets, U.S. Soil Conserv. Service, written commun., Oct. 31 and November 14, 1961). A 3-mile open drain installed by the Anaconda Copper Mining Co., north of some newly constructed settling ponds, was discharging about 7 cfs on May 5, 1961. A similar drain south of the settling ponds was reported to have an average discharge of 8 cfs (John Grant, Anaconda Copper Mining Co., oral commun., 1961). It also discharges tailings water from the Washoe Smelter.

Discharge of ground water from springs and seeps is small in comparison to discharge by effluent streams and drains. Three thermal springs—Gregson Hot Springs, Warm Springs, and the Anaconda Hot Springs—discharge several hundred gallons per minute and are the largest springs in the valley.

A substantial amount of water is discharged from the valley by evapotranspiration from swamps, from areas where the water table is within a few feet of land surface during much of the year, and from many borrow pits that are partly filled with ground water. A large quantity of water is also discharged from the area by evaporation from the several thousand acres of settling ponds at the Washoe Smelter and from the open streams.

An estimate of water "use" (evapotranspiration and change in soil-moisture storage) from June to September 1961 (table 4) was made from inflow and outflow records, precipitation records, and change in ground-water storage estimates. Inflow measurements consisted of miscellaneous measurements of eight streams and two irrigation canals (table 5) and continuous streamflow records on German Gulch, Racetrack Creek, and Warm Springs Creek. Streamflow records of Warm Springs Creek were furnished by the Anaconda Copper Mining Co. The outflow was determined from daily gage heights and monthly discharge measurements of the Clark Fork near Garrison. From June to September 1961, the total "use" (125,000 acre-feet) was more than three times as much as the total discharge from the area as streamflow (41,00 acre-feet).

The amount of ground water discharged by pumping from wells is relatively small in comparison to the amount discharged by effluent streams and evapotranspiration. About 7,000 acre-feet of water per year is pumped from the ground-water reservoir.

TABLE 4.—*Monthly water supply of the Deer Lodge Valley, in acre-feet, for the period June–September 1961*

Month	Surface-water inflow	Surface-water outflow	Net loss in streamflow	Precipitation	Ground-water discharge	Ground-water recharge	Evapotranspiration and change in soil-moisture storage
June	51,100	19,600	31,500	18,900		13,700	36,700
July	20,500	6,000	14,500	17,100		1,300	30,300
Aug	14,300	5,100	9,200	6,700	8,700		24,600
Sept	12,100	10,300	1,800	28,700	3,200		33,700
Total	98,000	41,000	57,000	71,400	11,900	15,000	125,300

TABLE 5.—*Miscellaneous streamflow measurements, Deer Lodge Valley, 1961*

[Values are in cubic feet per second]

Date	Clark Fork (near Ramsey)	Willow Creek	Mill Creek	Lost Creek	Modesty Creek	Dempsey Creek	Peterson Creek	Cottonwood Creek	Tincup Creek	Tavener Ditch	Pauly Ditch
May 29	38.81	40.79	177.68	46.72	4.95	55.36	11.31	161.77	33.08	33.99	45.59
June 6	27.42	20.73	192.97	40.32	7.20	65.23	6.44	87.82	34.34	66.43	Dry
June 12	31.14	18.91	181.66	44.93	6.58	57.86	4.08	61.20	32.33	56.61	49.59
June 19	25.88	9.54	91.88	21.79	6.54	47.56	1.79	31.19	26.47	38.98	33.40
June 26	20.81	6.92	67.14	18.13	5.92	42.74	.58	19.33	20.34	33.39	31.46
July 5	30.75	5.31	43.30	16.44	----	28.81	<.5	12.27	16.62	34.38	29.87
July 10	29.43	3.68	45.90	14.02	5.70	22.93	<.5	8.37	8.68	31.34	29.56
July 18	27.47	2.34	38.39	12.89	4.11	19.29	<.5	4.76	8.85	35.64	26.97
July 31	25.86	----	19.17	10.90	4.76	11.18	<.5	3.75	8.65	33.44	52.59
Aug. 8	22.39	.84	15.96	10.48	3.37	7.03	<.5	2.11	5.81	----	----
Aug. 16	26.45	----	14.42	7.27	3.59	9.44	<.5	2.70	5.55	25.54	36.32
Aug. 28	25.23	1.13	13.05	7.56	3.20	10.65	<.5	2.81	4.04	----	----
Sept. 6	25.07	1.53	17.05	8.48	3.73	10.70	<.5	1.51	2.85	12.10	5.23
Sept. 15	25.73	1.50	18.46	7.72	3.61	7.34	<.5	4.93	4.97	----	----
Sept. 25	24.97	1.78	19.83	6.89	3.16	8.93	<.5	5.89	5.41	10.48	4.62

PRESENT DEVELOPMENT

In the Deer Lodge Valley, ground water is the principal source of water for municipal and domestic use. Only two wells are presently (1967) being used for irrigation, but three others have been drilled that will probably be used for irrigation in the future.

About 7,000 acre-feet per year of ground water is pumped from six municipal, four State institution, two irrigation, and several hundred domestic and stock wells. Almost 50 percent of the 7,000 acre-feet is pumped from three municipal wells in Anaconda; another 30 percent is pumped from three municipal wells in Deer Lodge. The wells in Anaconda tap alluvium along Warm Springs Creek. Two of the Deer Lodge wells tap Tertiary sediments; the other, a shallow well used mostly for standby, taps alluvium along the Clark Fork. Total pumpage from four State institution wells is about 1,000 acre-feet per year; most of the water is from Tertiary sediments. Probably less than 300 acre-feet per year is withdrawn from stock and domestic wells tapping alluvium and Tertiary sediments. About 200 acre-feet per year is pumped from two irrigation wells. The three unused irrigation wells are capable of yielding almost 500 acre-feet each irrigation season. Most of the water from the irrigation wells in use is from Tertiary sediments, but water from the other three will be mostly from Quaternary sediments.

POTENTIAL DEVELOPMENT

Because the entire flow of all the tributary streams rising in the mountains and flowing into the Deer Lodge Valley is appropriated, any future developments that require substantial amounts of water will have to use ground water or purchase surface-water rights.

Recharge to the valley fill is sufficient that additional withdrawals of ground water could be made without excessively lowering the water table. Because of the great variation of water-yielding properties of the sediments in different parts of the valley, test holes should be drilled before installation of industrial or irrigation wells. The most likely areas for the best wells are in the flood plain of the Clark Fork and coalescing fans of Mill and Warm Springs Creeks. In these areas, properly constructed and developed wells will yield as much as 300 gpm if they penetrate alluvium with a large saturated thickness. Properly constructed and developed wells tapping both the Quaternary and Tertiary sediments will yield 1,000 gpm or more. Wells with the greatest yield from the Tertiary sediments are gravel packed. Gravel packing is generally not necessary in the Quaternary sediments.

Wells that will yield 300 gpm or more can be developed in the Racetrack and Dempsey Creek fans. Because of the variation in saturated thickness and sediment size, test holes should be drilled before the drilling of production wells. The Tertiary sediments underlying these fans are predominently fine grained and do not yield sufficient water to wells for irrigation.

In the rest of the valley, the upper few hundred feet of sediments will generally yield water sufficient only for stock and domestic use. However, there are some aquifers at greater depths. A test hole (B7-10-22dd) drilled by the Montana Power Co. reportedly flowed 140 gpm from a depth of about 1,200 feet. The sodium content is great enough that the water is probably not suitable for irrigation of most soils (U.S. Soil Conservation Service, Bozeman, Mont., written commun. March 21, 1962). The water would probably be suitable for certain industrial uses or for public supply.

SUMMARY AND CONCLUSIONS

The principal aquifers in the Deer Lodge Valley are in the upper few hundred feet of valley fill, which includes sediments of both Quaternary and Tertiary age. The Quaternary alluvium is relatively thin; but domestic and stock wells obtain sufficient water except along minor tributary streams. Locally, water for irrigation of small areas can be obtained from the alluvium in the vicinity of the Clark Fork and its major tributaries. The Tertiary sediments are thick; their permeability is variable but generally low. Most wells in the Tertiary sediments yield less than 10 gpm per ft. of drawdown even when properly constructed and developed. However, the thickness of these sediments is great enough that yields of 1,000 gpm or more can be obtained locally. The most likely areas for obtaining the higher yielding wells are the flood plain of the Clark Fork and the coalescing fans of Mill and Warm Springs Creeks. Recharge in the valley is sufficient that additional withdrawals of water could be made without excessively lowering the water table.

The depth to water in the flood-plain area is generally less than 10 feet, but on the fans and terraces it ranges from 10 to 150 feet. Depth of wells ranges from a few feet to 250 feet. The general direction of movement of ground water in the valley is toward the flood-plain from the east and west. However, in the southwestern part of the valley, movement of ground water is generally northeastward with a component of flow toward the tributary streams.

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