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GEOLOGY and GROUND WATER RESOURCES of the KALISPELL VALLEY, NORTHWESTERN MONTANA

by
R. L. Konizeski, Alex Brietkrietz,
and R. G. McMurtrey
U. S. Geological Survey

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of the
KALISPELL VALLEY, NORTHWESTERN MONTANA

by
R. L. Konizeski, Alex Brietkrietz,
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ABSTRACT

The Kalispell Valley, a trench-like depression formed by down faulting in the Precambrian rocks during late Paleocene-Eocene time, was partly filled with material eroded from the nearby mountains during Tertiary time. In Pleistocene time the sediments of Tertiary age were partly eroded and the remnants were buried beneath ice-contact and glaciolacustrine deposits. As the ice from the last glacial stage melted, Lake Missoula expanded northward and the Kalispell Valley was inundated. Sand, silt, and clay (glacial flour) were deposited in Lake Missoula. Their thickness locally amounts to several hundred feet. While the lake was receding about 12,000 years ago, the Flathead River and its tributaries entrenched their courses about 100 feet into the unconsolidated valley-fill deposits. The flood plains were subsequently broadened and graded to an altitude of about 2,880 feet at which the lake had stabilized. Gravity data indicate a maximum depth of about 4,800 feet of valley-fill deposits of Tertiary and Quaternary age.

Several aquifers were defined during the study. The oldest is bedrock of Precambrian age. Water-bearing units of Pleistocene age include a deep artesian aquifer, two shallow artesian aquifers, and three perched aquifers of small areal extent. Water-bearing units of Recent age include flood-plain deposits of sand and of gravel.

The bedrock of Precambrian age yields adequate quantities of water to domestic and stock wells. It is recharged by precipitation and by seepage from overlying unconsolidated rocks.

The artesian aquifers of Pleistocene age supply many domestic and stock wells. In some places, wells that tap the deeper artesian aquifer flow as much as 225 gallons per minute and can yield as much as 1,500 gallons per minute when pumped. Recharge to the artesian aquifers is mainly along the mountain fronts. Discharge is to other aquifers near the center of the valley by upward percolation through overlying poorly permeable beds.

The three perched aquifers of Pleistocene age are (1) dune and lacustrine sand on the east and central terraces, (2) glacial outwash northwest of Kalispell, and (3) glacial drift in the pothole area. The dune and lacustrine sand is poorly permeable and yields less than 10 gallons per minute to wells. The glacial outwash yields as much as 200 gallons per minute to wells but its areal extent is small. The glacial drift in the pothole area has been tapped by few wells, therefore yield and permeability were not estimated. Recharge to the perched aquifers is by precipitation and seepage from streams. Discharge is by evaporation, transpiration, and seepage to springs, to streams, and to other aquifers.

The flood plain of the Flathead and Whitefish Rivers is underlain by an average of 28 feet of gravel north of Kalispell and about the same average thickness of sand south of Kalispell. The gravel aquifer yields as much as 2,000 gpm to wells, and the average coefficient of transmissibility is about 1,100,000 gallons per day per foot. The sand aquifer yields a few gallons per minute to domestic wells, and the average transmissibility is estimated as 7,500 gallons per day per foot. The gravel and sand aquifers are recharged mainly by precipitation, applied irrigation water, and seepage from the river or lake during relatively high stages. Discharge is by evaporation and transpiration and by seepage to the river or lake during relatively low stages.

MONTANA

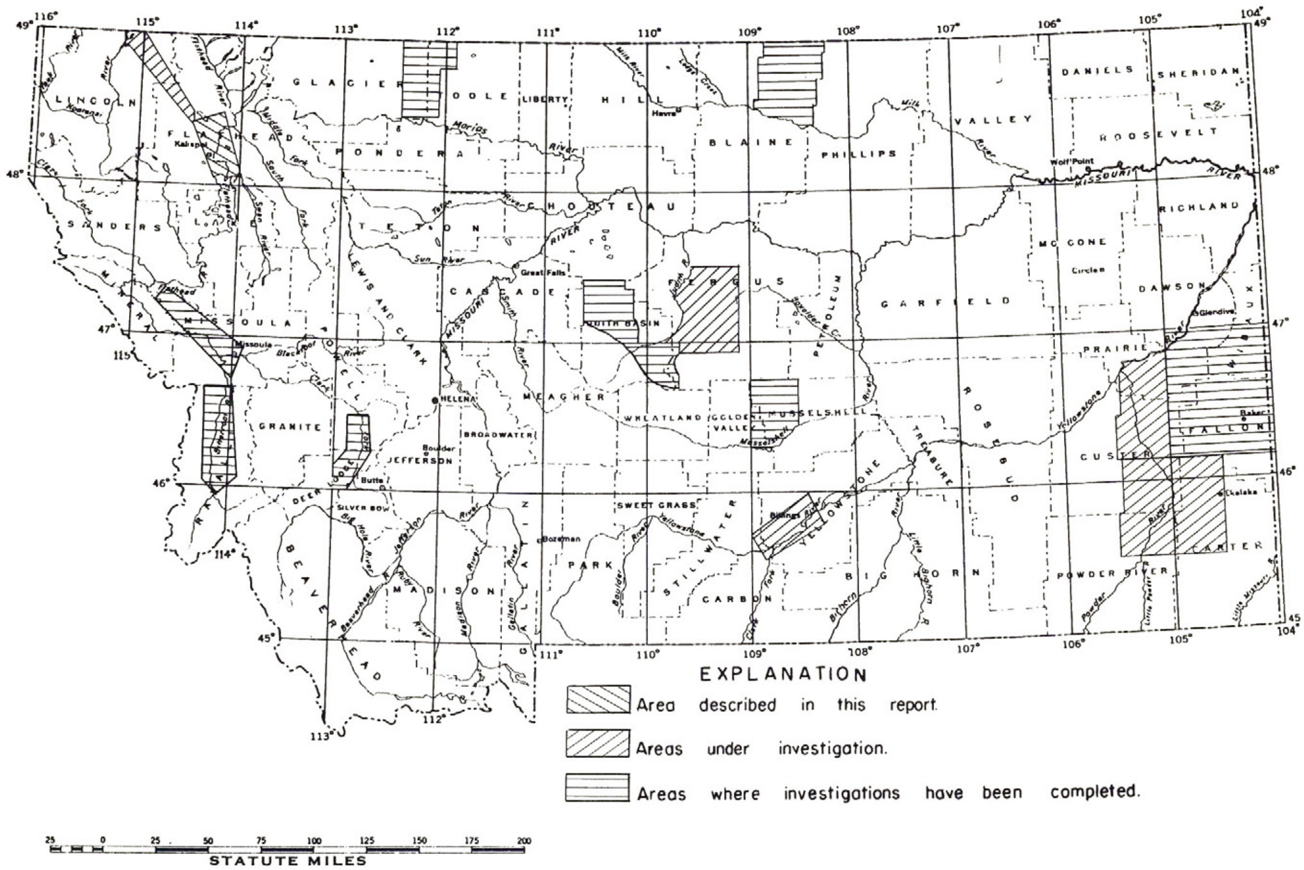


Figure 1.—Location of cooperative ground-water investigations, 1955-67.

INTRODUCTION

PURPOSE AND SCOPE

Water is Montana's most vital natural resource; it is essential to the economy of the state. Economic growth in the Kalispell Valley is directly dependent on optimal development of the water resources. The purpose of this investigation, begun in 1964 and completed in 1967, was to describe the principal aquifers and the occurrence, movement, quantity, and quality of the ground water. It is part of a State-Federal cooperative program begun in 1955 to appraise the ground-water resources and to provide data for planning their optimal use and development (fig. 1).

ACKNOWLEDGMENTS

The collection of data for this report was facilitated by the excellent cooperation and assistance from farmers and landowners throughout the valley. Special thanks are due those who permitted the installation of water-level recorders and who allowed water-level measurements to be made in their wells. Well drillers Homer McClarty, Gordon DeYoung, and Olsen and Justin provided logs of wells and other information. Thanks are due the Flathead County Clerk and Recorder's Office, Montana State Highway Commission, Bureau of Mines and Geology, Fish and Wildlife Service, and Kalispell Water Department for data supplied that contributed to the progress of the investigation.

The assistance of Don Peterson, Dean Kleinkopf, and Penelope Miller of the Branch of Regional Geophysics, Geologic Division, U. S. Geological Survey, in the reduction of gravity data and the computing of gravity cross sections is gratefully acknowledged.

PREVIOUS INVESTIGATIONS

Principal hydrologic papers are those of Cady (1941); McDonald (1946); and Erdmann (1947). Geologic references are those of Salisbury (1901); Pardee (1910); Daly (1912); Davis (1916, 1921); Clapp (1932); Alden (1953); Johns (1963); Richmond (1965); and Richmond and others (1965). Williams and Jackson (1960) prepared a soil survey, and The Pacific Power and Light Co. (1964) published an industrial survey of the area. Water-level measurements in wells between Flathead Lake and Kalispell were made from 1928-44 and from 1947-49 to determine the effects on the water table of the regulation of Flathead Lake. The water-level measurement data are in Water-Supply Papers 777, 817, 840, 845, 886, 910, 940, 990, 1020, 1100, 1130, and 1160. Basic data obtained during this investigation are in a report by Brietkrietz (1966).

WELL-NUMBERING SYSTEM

Wells and springs are assigned numbers in accordance with the U. S. Bureau of Land Management system of land subdivision. The "B" prefix indicates that the well is west of the Montana Principal Meridian and north of the base line. The first numeral indicates the township, the second the range, and the third the section in which the well or spring is located. Lowercase letters that follow the section number indicate the location of the well within the section. The first letter denotes the quarter section (160-acre tract) and the second the quarter-quarter section (40-acre tract). These subdivisions are designated a, b, c, and d, in a counterclockwise direction, beginning in the northeast corner. If more than one well is located in the same 40-acre tract, consecutive numbers beginning with 1 are added to the well number. Thus the second well inventoried in the SE¼ SE¼ sec. 21, T. 30 N., R. 21 W., would be numbered B30-21-21dd2 (fig. 2).

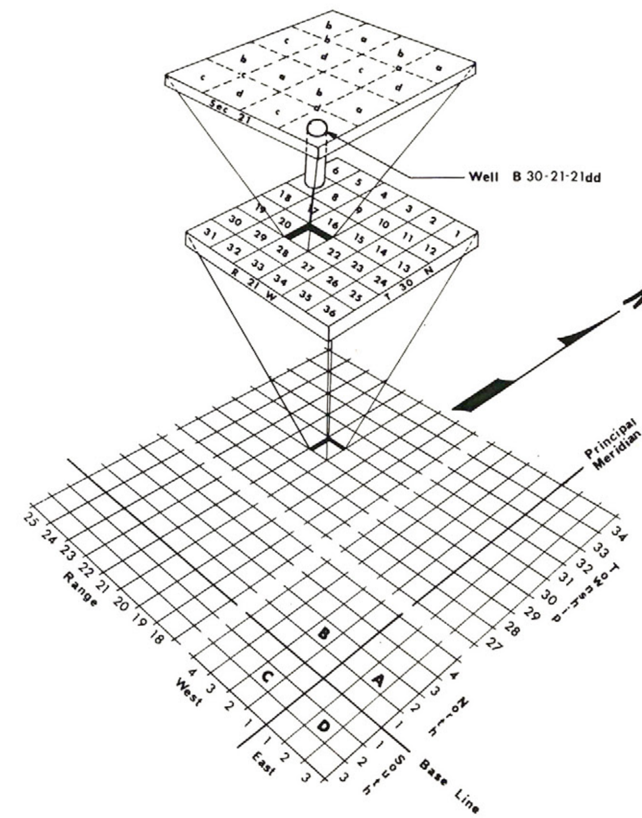


Figure 2.—System of numbering wells.

GEOGRAPHY

LOCATION AND EXTENT

The Kalispell Valley is mostly within Townships 27 to 31 North and Ranges 20 to 22 West, in Flathead County, northwestern Montana. It is a north-south-trending intermontane basin of about 325 square miles, 28 miles long by 7 miles wide at the south end and as much as 15 miles wide in the central and northern parts (fig. 3). The valley is in the Northern Rocky Mountain physiographic province (Fenneman, 1931, fig. 82).

TOPOGRAPHY

The Kalispell Valley is bounded by the Swan Range, Teakettle Mountain, Whitefish Range, Salish Mountains, and Flathead Lake. The Stillwater and Swan Valleys are expressions of the Rocky Mountain trench and join the Kalispell Valley in the northwest and in the southeast (pl. 1).

The uniform front of the Swan Range stands about 4,000 feet above the valley floor. The average altitude of the front is about 7,000 feet. The range is incised by numerous subparallel canyons that are separated by well-rounded ice-scoured ridges. In typical cross sections, the upper parts of the canyon walls are U-shaped, but V-shaped canyons have been cut into the flat bases of the U's. Between Lake Blaine and Bad Rock Canyon, the upper ends of the ridges slope valleyward at angles ranging from 20 to 30 degrees from the horizontal. Between 200 and 600 feet above the valley floor, the ridges are truncated to subtriangular faces. The upper parts of these faces slope toward the valley at angles that range from 50 degrees to nearly vertical. The steep slopes are generally on rough fractured rock, whereas the gentler slopes above them are on rock that has been scoured and beveled by ice. The lower parts of the subtriangular faces are mantled by talus.

The bare, rounded, ice-scoured crest of Teakettle Mountain is separated from the Swan Range by Bad Rock Canyon and stands 2,800 feet above the valley at a maximum altitude of 5,936 feet. Summit altitudes of the ice-scoured ridges of the south end of

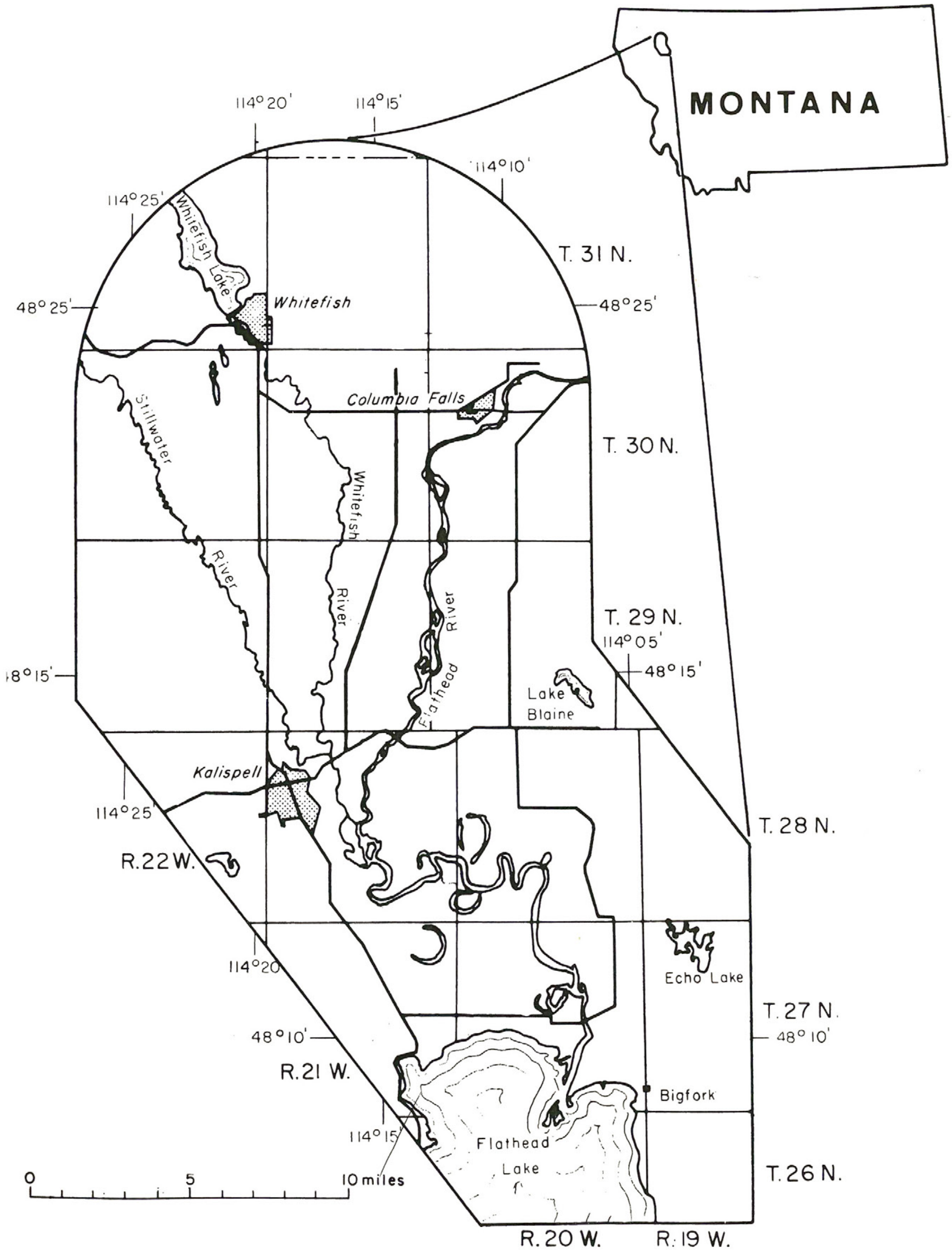


Figure 3.—Study area, principal towns, and drainage features.

the Whitefish Range are mostly between 5,500 and 6,500 feet, or 2,500 to 3,500 feet above the valley floor.

A 4-mile-wide area between Whitefish Lake and the Stillwater River is formed of southeast-trending, ice-scoured hills. Most summit altitudes are between 3,500 and 4,000 feet, but Lion Mountain rises to 4,387 feet, or about 1,300 feet above the bordering lowlands.

Maximum altitudes in the deeply glaciated Salish Mountains west of the valley are mostly less than 3,500 feet, or only about 500 feet above the valley floor.

The principal topographic features in the valley are (1) an east valley terrace, (2) a central valley terrace, (3) ice-scoured hills between Whitefish Lake and the Stillwater River, (4) glaciated terrain northwest of Kalispell, and (5) the flood plain of the Flathead and Whitefish Rivers. The east valley terrace, between the flood plain and the mountains, ranges from 3 to 4 miles in width, and stands about 80 to 150 feet above the flood plain. In some places, the western edge of the terrace is delineated by 50- to 100-foot scarps, and in other places by broad, lower terraces. Near Columbia Falls it is bordered by narrow, superposed terraces. Near Bigfork the floor of the Swan Valley merges with the terrace. A kame and kettle area (fig. 4) forms hummocky topography on the terrace between the mouth of the Swan Valley and Lake Blaine. North of Lake Blaine the terrace has south-trending swell-and-swale topography, partly masked by dune sand. Lake Blaine occupies the south end of a low swale that extends northward along the base of the Swan Range for 9 miles to the mouth of Bad Rock Canyon.

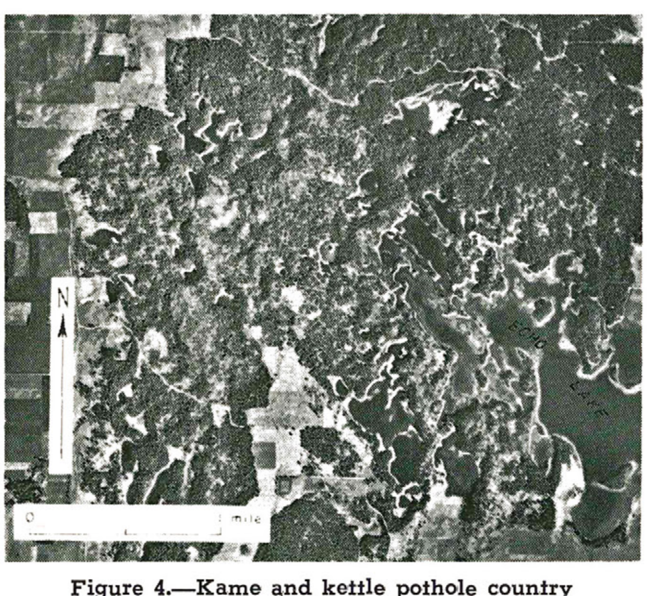


Figure 4.—Kame and kettle pothole country near Echo Lake.

The central valley terrace extends 5 miles north of Kalispell. This terrace is ½ mile to 2½ miles wide and stands 50 to 80 feet above remnants of lower terraces that are 20 to 30 feet above the flood plains along the Flathead and Whitefish Rivers. The northern part of the central valley terrace is marked by rolling sand hills and the southern part by swell-and-swale topography. From a point about 5 miles northwest of Kalispell almost to the mouth, the Stillwater River has cut a trench 50 to 80 feet below the terrace and has isolated two small terrace remnants in the northeastern outskirts of Kalispell in sec. 8, T. 28 N., R. 21 W.

The ice-scoured hills between Whitefish Lake and the Stillwater River are perhaps the dominant features in the valley; they extend from Lion Mountain to a point 5 miles north of Kalispell. The maximum altitude is about 3,900 feet, the average width is about 3 miles, and the land slopes gradually southeastward for about 8 miles to a minimum altitude of 3,030 feet at the head of the central valley terrace. The undulating surface of the hills is everywhere marked by narrow, southeast-trending ridges. The hill area is bordered on the west by the Stillwater flats, which average about 1½ miles in width.

Kame-and-kettle topography northwest of Kalispell, between the central valley terrace and the Salish Mountains, grades south into northwest-trending drumlin topography. West of Kalispell the drumlin topography grades south into west-trending recessional-morainal topography. Directly northwest of Kalispell the morainal topography merges with the central valley terrace.

The flood plain of the Flathead narrows from an average width of about 6 miles in the south end of the valley, where it is incised by multitudinous channel scars and stream meanders, to an average width of 1 mile in the north end of the valley. The shoreline of Flathead Lake extends in two gentle arcs for 7 miles across the south end of the flood plain. Altitudes of the flood plain range between 2,900 feet at the south end and 3,010 feet at the north end.

DRAINAGE

The Flathead River flows from Bad Rock Canyon into the northeastern corner of the valley, then southwest for 5 miles past Columbia Falls, then south for 12 miles to Kalispell. From Kalispell the river flows to Flathead Lake through a series of great meanders. The average gradient above Kalispell is about 6 feet per mile; below Kalispell it decreases abruptly to about 1 foot per mile during the period of minimum impoundment of Flathead Lake.

From 1928 to 1965, the Flathead River had an average annual discharge at Columbia Falls of 6,-975,000 acre-ft (acre-feet) or 9,634 cfs (cubic feet per second). Maximum discharges recorded June 1894, May 23, 1948, and June 9, 1964, were 142,000 cfs, 102,000 cfs, and 176,000 cfs, respectively. Minimum discharge recorded Dec. 8, 1929, was 798 cfs (U. S. Geological Survey, 1955, 1964). Water-level fluctuations in Flathead Lake range from a planned winter low of about 2,883 feet above mean sea level to a summer high of 2,893 feet.

The Stillwater and Whitefish Rivers are the two principal tributaries of the Flathead River in the Kalispell Valley. The Stillwater River meanders for 40 miles south from the northwestern corner of the valley, is joined by the Whitefish River, then enters the Flathead River a mile southeast of Kalispell. The average gradient is about 2 feet per mile. At the U. S. Geological Survey gaging station near Whitefish, the average annual discharge, from 1930 to 1950, was 246,200 acre-ft or 340 cfs. Maximum discharges recorded May 6, 1948, and June 9, 1964, were 4,330 and 1,480 cfs, respectively (U. S. Geological Survey, 1955; Boner and Stermitz, 1967).

The Whitefish River flows from Whitefish Lake, in the northwestern corner of the valley, for 15 miles to join the Stillwater River near Kalispell. Its average gradient is about 6 feet per mile. Average annual discharge, from 1929 to 1950, was 138,300 acre-ft or 191 cfs. Maximum discharges recorded on May 30, 1948, and June 9, 1964, were 1,290 and 1,400 cfs, respectively (U. S. Geological Survey, 1955; Boner and Stermitz, 1967).

Sixteen smaller streams flow into the valley from the mountains. Most of the flow is either diverted for irrigation or percolates into the unconsolidated valley fill. Typical are Big Lost Creek and Cedar Creek. Big Lost Creek flows into the valley about 7 miles northwest of Kalispell and is diverted into the Ashley Creek irrigation ditch. Cedar Creek flows into the northeast end of the valley and percolates into the alluvium near Columbia Falls.

A number of small spring-fed streams head within the valley. Among these are two perennial Spring Creeks; one heads in the Evergreen district northeast of Kalispell, the other a few miles northwest of Kalispell. Mill Creek rises from a large spring near the Creston National Fish Hatchery. Many smaller springs rise along the east-valley and central-valley terrace scarps and in various localities about the Stillwater flats.

The Kalispell Valley contains more than 40 lakes. The largest of these, Whitefish Lake, is impounded behind a moraine in the northwestern corner of the valley. It is 6 miles long by about 1 mile wide. Water flows into the lake from small inlets at the northern end.

Echo Lake (700 acres) and Lake Blaine (400 acres) are the largest of more than 30 pothole lakes in the kame-and-kettle area on the east valley terrace. Both lakes have surface inlets but receive most of their water from subsurface springs. The other pothole lakes range from somewhat less than an acre to as much as 200 acres. All are less than 30 feet deep and are fed from subsurface springs. The springs are fed by recharge from creeks that head in the mountains and percolate into the porous glacial deposits.

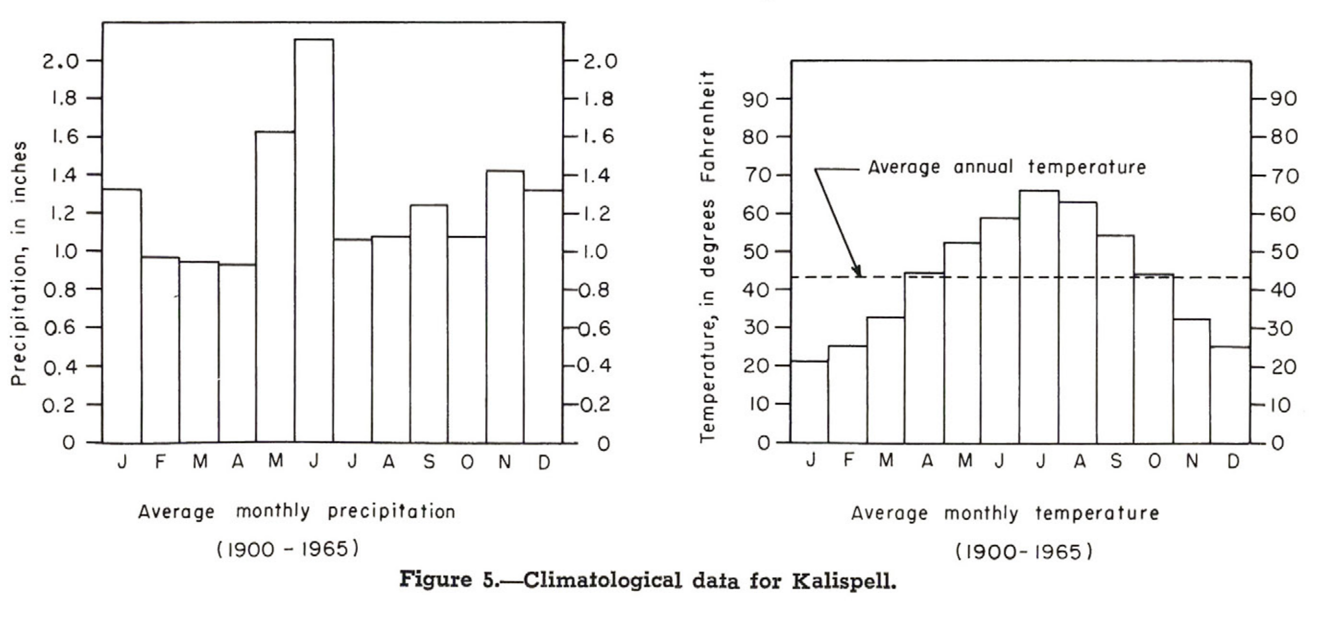


Figure 5.—Climatological data for Kalispell.

CLIMATE

At Kalispell, the average annual precipitation from 1900 to 1965 was 15.11 inches. About one-fourth of the precipitation occurs in May and June; the rest is fairly evenly distributed throughout the year (fig. 5).

The average annual temperature at Kalispell from 1900 to 1965 was 43.2°F. The average temperature in January, the coldest month, was 21.3°F, and in July, the warmest month, 65.4°F. The growing season averages 150 days at Kalispell as contrasted to only 99 days at Columbia Falls.

POPULATION AND ECONOMY

The population of the Kalispell Valley (about 30,000) is concentrated around Kalispell (10,151), Columbia Falls (2,132), Whitefish (2,965), Somers (800), and Bigfork (500). The valley's inhabitants are mostly employed by industries directly dependent upon stable water supplies. The Anaconda Aluminum Company's water-cooled reduction plant at Columbia Falls is the largest employer in the valley; the labor force exceeds 700. The Great Northern Railroad employs more than 500 people in its operations and maintenance center in Whitefish. Private, State, and National forest lands surrounding the valley provide annually about 300,000,000 board feet of timber, most of which is processed in 36 local mills. More than 1,200 persons were employed in the manufacture of lumber in the summer of 1964 (Pacific Power and Light Company, 1964).

More than 10 percent of all Christmas trees sold in the United States come from this region. Most are harvested from naturally forested lands, but many of the best trees are grown on irrigated farm plots. There are about 1,100 farms in the valley, and a total of 30,780 acres of irrigated cropland (U. S. Soil Conservation Service, written communication). Principal agricultural products are beef, dairy products, hay, grain, and sweet cherries. More than 10,000 head of beef cattle are fed annually. Most of the hay and grain is used locally to feed livestock. Sweet-cherry production exceeded 3,360,000 pounds in 1966, and the record production was 5,500,000 pounds in 1967, but production varies greatly according to weather conditions.

Kalispell Valley has spectacular scenery and many opportunities for water-oriented outdoor recreation. Because of its location near Glacier National Park, the area is becoming an important tourist center. Flathead Lake in particular is a tourist attraction of national as well as local importance. Many of the lakes in the valley are surrounded by numerous summer homes and resorts. Fall and winter sports such as hunting and skiing help maintain the economy on a year-round basis.

GEOLOGY

ROCKS OF PRECAMBRIAN AGE

The mountains near the Kalispell Valley consist of Precambrian sedimentary rocks (pl. 1). Gray to greenish-gray argillite and light-gray quartzite of the Ravalli Group of the lower part of the Belt Series crop out along the base of the Swan Range and Teakettle Mountain, the southwestern margins of the Whitefish Range, and the east slopes of the Salish Mountains for about 7 miles north from Ashley Creek. Gray, silty, stromatolite-bearing dolomite of the Piegan Group of the middle Belt Series overlies rocks of the Ravalli Group at higher altitudes in the Swan Range and on Teakettle Mountain. The Piegan Group crops out around most of the southern margins of the Whitefish Mountains and the Salish Mountains. The regional stratigraphy, structural relationships, and gravimetric data indicate that rocks of the Ravalli and Piegan Groups also underlie the floor of the valley beneath unconsolidated to consolidated deposits of Cenozoic fill.

ROCKS OF TERTIARY AGE

Unconsolidated to semiconsolidated Tertiary rocks occur in many northern Rocky Mountain intermontane basins of comparable size, but none are exposed in the Kalispell Valley. Late Eocene or early Oligocene rocks crop out in the Flathead Valley about 30 miles north of the project area (Russell, 1954). Alden (1953, p. 31, 47) described Tertiary(?) sediments in and about the south end of the Mission Valley near Moiese, Montana. The Rocky Mountain Trench contains exposures of coarse clastic material of Miocene age near Fort Steele, British Columbia (Rice, 1937); Garland and Kanasewich (1961, p. 2500) and Smith (Structure and stratigraphy of the Whitefish Range, northwest Montana, Princeton University Doctoral thesis, 1963) indicated that Tertiary rocks probably occur at depth within the trench. Probably most of the valley fill (as much as 4,800 feet) in the Kalispell Valley is Tertiary. The lack of outcrops is probably due to three circumstances: (1) during the Pleistocene, glaciation may have removed or reworked the upper part of the Tertiary fill; (2) during the Quaternary, subsidence of the valley floor and deposition of a thick mantle of glacial, stream, and lake

Figure 6.—Glaciolacustrine silt and sand north of Kalispell in sec. 7, T. 28 N., R. 21 W.

deposits may have covered the Tertiary rocks; and (3) Tertiary outcrops in this region are difficult to identify (Erdmann, 1944, p. 64-67; Ross, 1959, p. 68).

ROCKS OF QUATERNARY AGE

In the south end of the valley, well-bedded (in

Figure 7.—Boulder till in recessional moraine west of Kalispell in sec. 14, T. 28 N., R. 22 W.

places, varved) glaciolacustrine clayey silt and sand (fig. 6) more than 600 feet thick interfinger with and overlie gravel of unknown thickness. In the middle and north end of the valley, gravel is overlain by

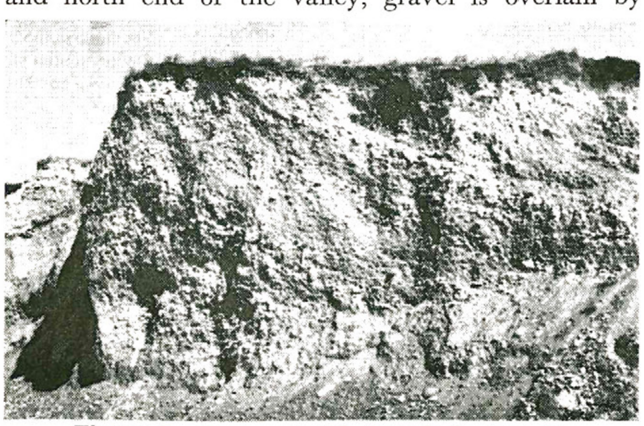


Figure 8.—Kame gravel deposit in sec. 31, T. 29 N., R. 21 W.

boulder till, kame deposits (fig. 7, 8), and outwash gravel 10 to 400 feet thick. The till, kame deposits, and outwash gravel interfinger with and are overlain by glaciolacustrine clayey silt and sand (fig. 9).

Figure 9.—Unconformity between kame and glaciolacustrine deposits exposed in roadcut in sec. 36, T. 29 N., R. 22 W.

Dune sand, locally as much as 75 feet thick, unconformably overlies the glacial and glaciolacustrine deposits.

Specimens of volcanic ash were collected from four localities (pl. 1 and fig. 10) at the contact be-

Figure 10.—Volcanic ash at unconformity between dune and lacustrine deposits exposed in roadcut in the SW¼ sec. 22, T. 30 N., R. 21 W. Volcanic ash exposed just below hammer.

tween drift and dune sand. On the basis of field studies by Roald Fryxell and petrographic examination by Miss Virginia C. Steen, Fryxell (written communication, 1965) stated that "it (the ash) is correlative . . . with ejecta from the eruption of Glacier Peak in the northern Cascade Range of Washington about 12,000-13,000 years ago . . ." Fryxell further stated ". . . the stratigraphic position . . . and the geographic location of this sample approximately 30 miles north of the terminal moraine at Polson, are of significance because identification of the ash as being of Glacier Peak origin provides:

1. A limiting date of more than 12,000 years for retreat of the Flathead Lobe from its maximum at Polson;

2. A limiting date of 12,000 years for draining of Glacial Lake Missoula and the waters in which the kame sand was deposited;
3. Evidence that retreat of the Flathead Lobe was approximately synchronous with recession of the Okanogan Lobe from the Columbia Plateau to the west, where ash from the Glacier Peak eruption also lies north of the terminal moraines (Fryxell, 1965); and
4. Support for previous correlation of the maximum advance of the Cordilleran Ice Sheet during Pinedale time with the middle stade of the Pinedale Glaciation (Richmond et al., 1965)."

Because of the above conclusions, and because of the absence of soil zones or regional unconformities within the drift, the exposed Kalispell Valley glacial and glaciolacustrine deposits are believed to be Wisconsin (Pinedale) in age and the overlying dune sand late Wisconsin to Recent (late Pinedale to Recent).

The north-central and western parts of the valley, except for the Stillwater flats, are underlain mostly by till that forms moraines and drumlins, and by glacial outwash that is overlain by a thin mantle of glaciolacustrine silt and sand. The till is a heterogeneous mixture of poorly sorted gravel, cobbles, and boulders, some more than 10 feet in diameter, in a clayey to sandy matrix (fig. 11). The glacial outwash differs

Figure 11.—Glacial till exposed in roadcut in sec. 33, T. 29 N., R. 22 W.

from the till in that it is better sorted, poorly bedded, and includes boulders 25 feet or more in diameter. The Stillwater flats are underlain by fine-grained, finely bedded (locally varved) glaciolacustrine deposits 1 foot to 240 feet thick.

The east and central valley terraces are underlain mostly by till and by kame deposits of well-rounded, well to poorly sorted, stratified gravel and cobbles. Kame deposits more than 100 feet thick are exposed in gravel pits on the east bank of the Flathead River in sec. 2 and 3, T. 28 N., R. 21 W., and within the city limits of Kalispell in sec. 8, T. 28 N., R. 21 W. (fig. 12). The till and kame deposits are

Figure 12.—Stratified kame deposit of gravel and sand in sec. 8, T. 28 N., R. 21 W.

overlain by a thin mantle of glaciolacustrine silt and sand. Dune sand mantles large areas of the northern part of the central valley terrace and the east valley terrace north of Creston. Except in fresh exposures, however, dune sand is generally indistinguishable from glaciolacustrine sand.

Two miles northeast of Bigfork, crevasse-filling deposits cover about 2 square miles. The deposits are coarse well-sorted sand that was deposited during Pinedale time in a lake marginal to the ice sheet.

The Recent flood-plain alluvium is mostly debris reworked from the drift and deposited in a wide variety of environments. In the north end of the valley, it consists almost entirely of well-sorted, interbedded gravel and sand, but in the south end of the valley it consists mostly of silt and sand.

STRUCTURE

The Kalispell Valley is underlain by a huge downdropped block of earth's crust. The valley's asymmetry about its north-south axis is the surface expression of normal faulting along the east side and may be related to a system of hidden faults along the west side and north end. Surface relationships and gravimetric data (p. 12), indicate that the underlying block of consolidated basement rocks is tilted to the east.

The crustal block underlying the valley is cut at an angle of about 25 degrees west of north by a system of normal faults associated with the southeast-trending Rocky Mountain Trench. The Lion Mountain area between Whitefish Lake and the Stillwater River is bounded by normal faults that probably extend southeast beneath unconsolidated fill into the mouth of the Swan Valley (Erdmann, 1947, p. 145; Johns, 1963, p. 44).

Truncated spur ends at the base of the Swan Range north of Lake Blaine are associated with a fault at the base of the range, although they have been elsewhere described as glacial features (Davis, 1921, p. 89). Evidence of their structural origin is as follows: 1. The tops of the spurs are ice scoured and beveled, but the truncated spur ends are not; 2. The valley-wall profile flattens at the tops of the truncated spurs; 3. The triangular spur-end facets are part of a regionally uniform joint system (fig. 13) along which movement is clearly indicated by slickensides; 4. Great talus accumulations at the base of the spurs indicate formation of the spurs after glaciation.

Figure 13.—Jointing on fault scarp east of Lake Blaine (bedding indicated by short dashed lines).

The combined evidence indicates that the spurs were formed by movement along the fault after retreat of the ice sheet.

A mile-wide swale along the base of the Swan Range between Lake Blaine and the mouth of Bad Rock Canyon may be an area of subsidence. Logs of wells drilled in the swale show 400 feet to 460 feet of lacustrine silt and sand, contrasted to only 70 to 155 feet in wells drilled about a mile west. The swale, therefore, is underlain by at least 300 feet more silt and sand than the area to the west. The greater thickness of lacustrine deposits may indicate subsidence relative to the bordering mountains along the mile-wide swale when the area was covered by the waters of Glacial Lake Missoula or ancestral Flathead Lake (late Pinedale to Recent).

Several east-trending faults were mapped by Johns (1963, pl. 3) along the Swan Range and the north end of the Mission Range. One east-trending fault is exposed in a new road cut in the SE¼ sec. 32, T. 27 N., R. 19 W.

No evidence was observed that would indicate that the Recent valley fill has been structurally deformed, but most of the exposures are in central parts of the valley, away from the mountain fronts.

GEOLOGIC HISTORY

It is not possible to date the origin of the Kalispell Valley from evidence within the project area, but the early history can be inferred by analogy with that of adjacent areas. The valley probably was formed during a period of normal faulting after a late Paleocene or Eocene episode of compression that produced folding and thrust faulting in the Whitefish Range and Glacier Park regions to the north and east (Ross, C. P., 1959, p. 95; Johns, 1963, p. 39-48). The intermittently subsiding basin was partly filled with erosional waste from the marginal highlands, concomitantly with successive structural movement during Tertiary time. Prior to Pleistocene time some of this material was probably stripped from the valley floor but most of it was eventually buried beneath glacial drift.

Alden (1953, p. 2, 127) supposed ". . . on a priority grounds, that there were as many distinct stages of glaciation in western Montana . . . as have been demonstrated as occurring in the central part of the upper Mississippi basin." And ". . . pre-Wisconsin ice overwhelmed the whole range." Pre-Wisconsin drift probably was deposited within the valley but it has not been definitely identified.

In middle Wisconsin (early to middle Pinedale) time a great lobe of the Cordilleran ice sheet in British Columbia advanced down the Rocky Mountain Trench and entered the Kalispell Valley at its northwestern corner near Whitefish. This lobe and coalescent alpine glaciers from the Whitefish Range were joined near Bad Rock Canyon by a great alpine glacier that originated in the highlands of Glacier National Park. As the resultant coalescent ice mass flowed southward, it was joined near Bigfork by the northwestwardly flowing glacier from Swan Valley (fig. 14). Except for a single small divergent lobe that advanced southwest up Ashley Creek, the main ice mass advanced southeast beyond the Kalispell Valley.

Glacial scour and erratics on the summit of Lion and Teakettle Mountains, the Salish Mountains, the northern end of the Mission Range (fig. 15), and near Somers (fig. 16) indicate that these topographic highs were completely overridden by ice. Near Columbia Falls and Whitefish the ice mass was at least 3,000 feet thick, in the middle of the valley near Kalispell about 2,500 feet thick, and in the south end of the valley about 2,200 feet thick (Alden, 1953, p. 118).

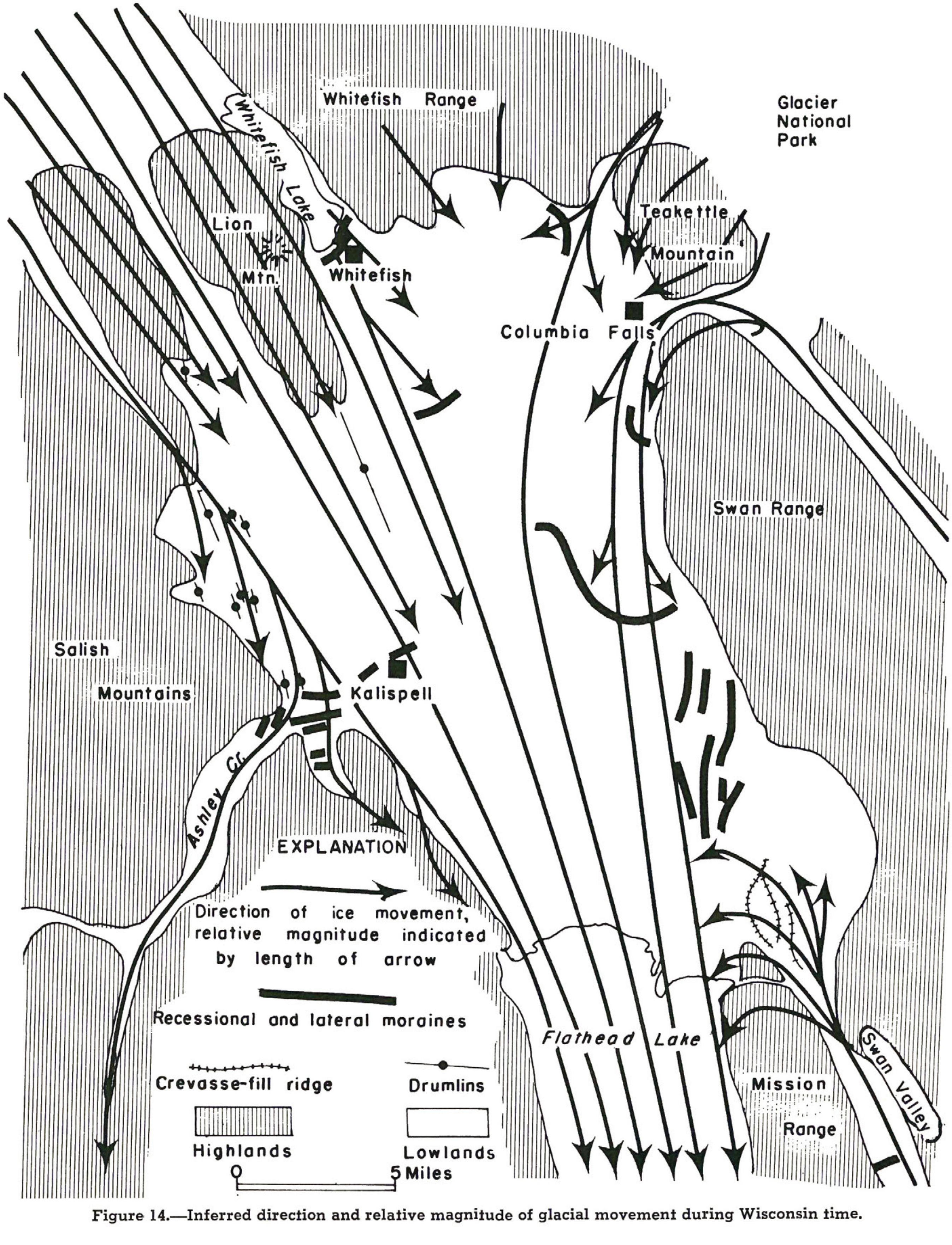


Figure 14.—Inferred direction and relative magnitude of glacial movement during Wisconsin time.

Figure 15.—Aerial photograph showing glacial scour on north end of the Mission Range.

Because of their massiveness and areal extent, the glaciers were able to quarry and transport huge volumes of debris to the valley. Lithologic correlation indicates that some of the source areas were far to the north, but rounding and beveling of the marginal highlands shows that most of the debris was derived locally. Much clayey boulder till was deposited in the lee of the Whitefish Range, Lion Mountain, and the Salish Mountains. Well to poorly sorted and bedded sand, gravel, and boulders accumulated locally in cracks and fissures within the ice, especially along the east side of the valley. Several recessional boulder moraines were emplaced west of Kalispell.

As the ice finally melted in middle Pinedale time, the Flathead arm of Glacial Lake Missoula expanded northward and inundated the entire Kalispell Valley to an altitude of about 3,400 feet. Sand, silt, and clay (glacial flour) accumulated in the glacial lake to a thickness of several hundred feet and mantled the older ice-contact and glaciolacustrine deposits. The absence of shore-line features indicates that the lake remained at this altitude for a relatively short period. About 12,000 years ago it receded to its 1938 level of 2,883 feet above sea level.

Figure 16.—Glacial scour above roadcut south of Somers in sec. 26, T. 27 N., R. 21 W.

Immediately afterward the exposed part of the former lake bottom was mantled by a few inches of airborne volcanic ash, transported from the Glacier Peak region of Washington. Most of the ash was reworked into aeolian deposits derived from the underlying lacustrine silt and sand, but some of it was preserved in local topographic lows and covered by dune sand.

As the lake, which is base level for the area, receded, the Flathead River and its tributaries entrenched their courses into the unconsolidated drift. The flood plains were subsequently broadened and graded to the 2,883-foot altitude at which the lake stabilized.

GRAVIMETRIC SURVEY

A gravity survey was made in the Kalispell Valley to determine the approximate depth of valley fill and to help interpret subsurface features and basement rock configuration. The gravity data are referenced to the Kalispell airport station of Woollard (1958), which has a value of 980.5819 gals.

FIELD METHODS

Gravity readings were made at about 370 points on an approximate 1-mile grid by means of a portable temperature-compensated Worden gravity meter. The uncorrected readings are probably accurate to the nearest 0.05 milligal relative to one another. Nearly all stations were at bench marks or at road intersections, section corners, and other places where altitudes had been determined by instrumental leveling during field work before completion of 7½-minute topographic quadrangle maps. The altitudes at a few stations were obtained by instrumental leveling after gravity readings were made. Maximum error in altitude probably is less than ± 1 foot. Position control was obtained from topographic quadrangle maps and is accurate within about 0.1 minute.

To determine the gravity meter drift, readings were made at a selected base station at the beginning and end of each day's work and at a previously occupied location at intervals of about 2 hours. To obviate errors in reading the meter, two readings were taken at each station and then averaged.

REDUCTION OF DATA

The gravity observations were adjusted for meter drift and for latitude, free air, Bouguer, and terrain corrections. The drift correction was made in the manner described by Dobrin (1960, p. 222). Latitude, free air, and Bouguer corrections were calculated by the Regional Geophysics Branch of the U. S. Geological Survey on the Survey's computer in Denver, Colorado. Bouguer values were obtained for densities of 2.67 and 2.10 gm/cm^3 (grams per cubic centimeter). Terrain corrections from a circular inner radius of 2.615 out to 166.700 kilometers and based on a density of 2.67 gm/cc were calculated on the Geological Survey's computer in Washington, D. C. The inner zone terrain corrections from the gravity stations out to 2.615 km were calculated by hand, with the aid of the Hammer chart.

INTERPRETATION OF DATA

Dobrin (1960, p. 242) has pointed out the importance for all users of gravity data to realize that interpretation is subject to many limitations and cannot give a unique answer to a specific problem. As independent geologic control increases, the questions to be answered by gravity information become more restricted and answers usually become more definite. Gravity methods are fairly good for determining the form and depth of a basin filled with light material and surrounded by heavier bedrock. Geologic maps of the Kalispell Valley reduced the number of variables and the ambiguity in the interpretation.

BOUGER GRAVITY MAP

The gravity map (pl. 2) shows Bouguer anomalies computed for a density of 2.67 gm/cm^3. Tables of observed gravity, station location and elevation, terrain effect, and Bouguer anomalies computed for a density of 2.10 gm/cm^3 are on file at the U. S. Geological Survey office in Helena, Montana.

The main feature shown on the map is an elongate gravity low that parallels the Swan Range for about 15 miles south of the Whitefish Range. The southern end of the low is skewed to the southeast and trends toward a gravity low at the mouth of the Swan River valley northeast of Bigfork. Another prominent feature is a gravity high that extends northwest from Bigfork. Two other features of note are anomalies east of Whitefish and northwest of Kalispell.

All of the gravity anomalies correlate with mapped geologic features. High gravity occurs over exposures of basement rocks around the periphery of the valley, and relatively low gravity is associated with varying thicknesses of low-density Cenozoic deposits. The large gravity low covering the northeastern part of the area indicates a structural depression filled with low-density rocks. It bears out the structural relationships discussed on page 9. The steep gradient (7 milligals per mile) on the east side of the low is indicative of the fault or fault zone along the western front of the Swan Range. The asymmetry of the anomaly, the lowest gravity values being east of the topographic axis of the valley, suggests eastward tilting of the underlying block of basement rocks.

The anomalies shown in the southern and western part of the area are probably associated with the Rocky Mountain Trench. The gravity high northwest of Bigfork is probably caused by an extension of bedrock northwest from the Mission Range beneath the valley fill. The Mission Range presumably separates the south end of the Trench into two prongs; one trends southeast into the Swan Valley and the other trends south along the west side of the Mission Range. The gravity low northeast of Bigfork lies in the narrow southeast-trending prong. It indicates a deep basin in the Trench containing low-density valley fill deposits. The basin probably represents a downfaulted block along longitudinal and transverse faults. Ross (1959, pl. 2) mapped a fault longitudinal to the Trench along the west side of the Swan Range and a transverse fault cutting the longitudinal fault; Johns (1963, pl. 3) mapped two more transverse faults. The longitudinal fault and a projection of the northern transverse fault coincides with the northeast side and the northwest end of the gravity low. The structural low along the west side of the Mission Range may be a south-trending prong of the Rocky Mountain Trench. An alternate hypothesis is that the low is not directly related to the Trench, but may be an expression of a north-trending fault that crosses the Mission Range structural high and intersects the Trench at about 25°.

The anomaly east of Whitefish shown by the abrupt change in direction of the contours suggests a bedrock high extending into the valley. It is related to the mapped fault along the northeast side of Whitefish Lake, which is believed to extend across the Kalispell Valley into the mouth of the Swan Valley (page 9). The anomaly northwest of Kalispell is probably related to a projection of the fault shown by Johns (1963, pl. 3) in Ashley Creek.

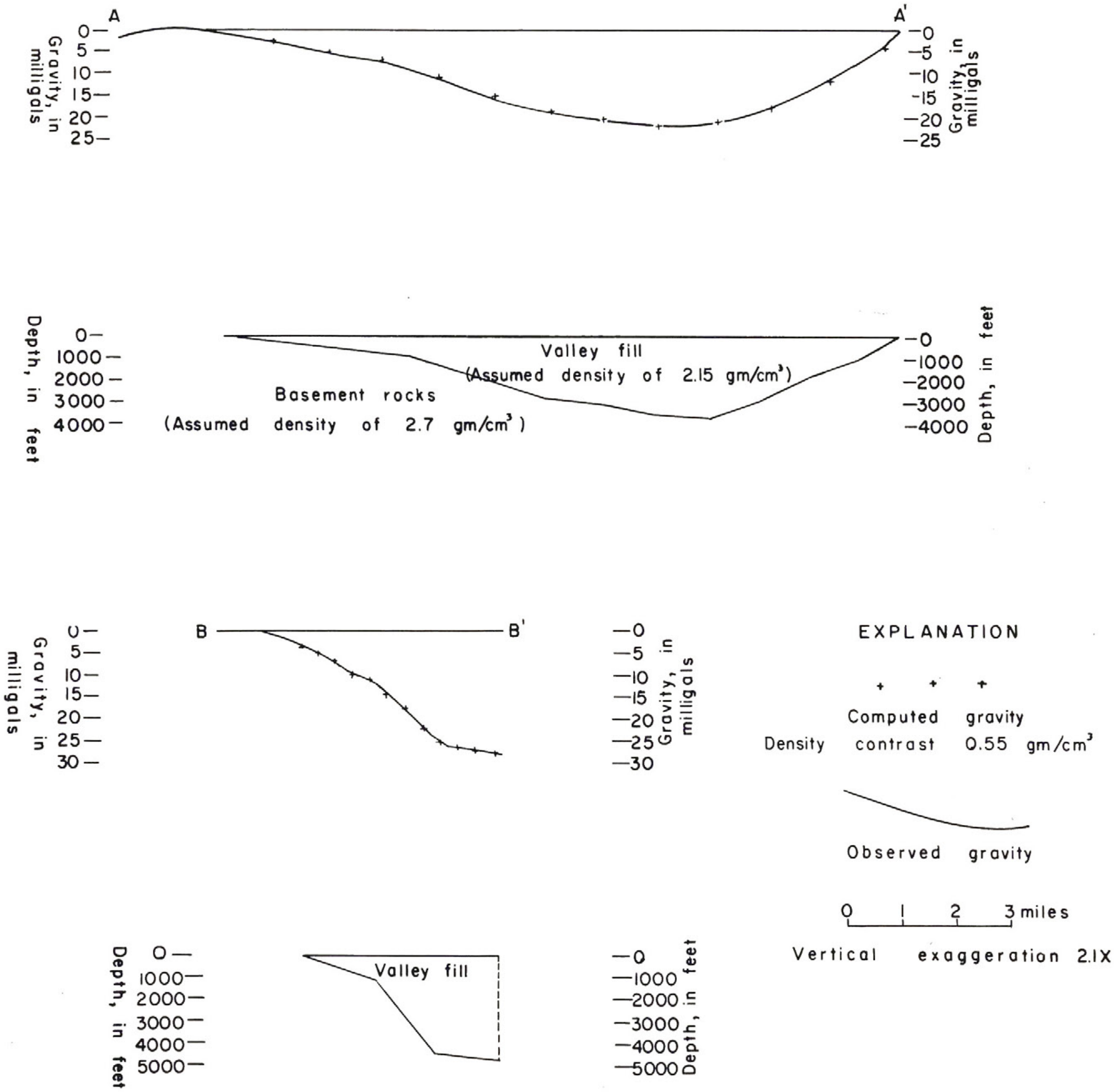


Figure 17.—Gravity profiles showing inferred basement rock configuration.

GRAVITY PROFILES

Two profiles (fig. 17) were calculated from the terrain-corrected gravity data by M. Dean Kleinkopf and Penelope Miller of the Regional Geophysics Branch of the U. S. Geological Survey. Calculations were made by use of the Talwani 2-dimensional modeling program. In profile A-A', an east-dipping regional gradient of about 0.5 milligals per mile was removed; in profile B-B', the regional gradient was assumed to be flat. A density contrast of 0.55 gm/cm³, based on an assumed density of 2.70 gm/cm³ for basement rocks and 2.15 gm/cm³ for valley fill, is believed to be accurate within ±0.1 gm/cm³. The depth of valley fill calculated is proportional to the density contrast; a lower density contrast would require a thicker section of valley fill.

Profile A-A' shows a residual anomaly, relative to outcrops of bedrock, of about –23 milligals just west of Columbia Falls. The calculated depth of valley fill in that area is 3,700 feet. The sediments thin gradually to the west and abruptly to the east.

Profile B-B′, which does not extend completely across the valley, shows a residual anomaly of about -28 milligals east of Echo Lake. The maximum calculated depth of valley fill is about 4,800 feet. The valley is narrow, and the sediments are probably in fault contact along the southwest side, as shown by a sharp break in profile and a steep gradient to the southwest.

GROUND WATER

This section of the report deals with the accumulation and movement of water in the various aquifers described during this study. A brief general description of the surface- and ground-water relations in the Kalispell Valley is followed by more detailed descriptions of recharge, movement, discharge, and quality of water in each aquifer.

Recharge to aquifers from precipitation and infiltration from streams is greatest from April through July because stream runoff from accumulated winter snow in the surrounding mountains is augmented by an average of 3.76 inches of rain in May and June. Along the flood plain of the Flathead and Whitefish Rivers, applied irrigation water and infiltration from the Flathead River during high stages are important sources of recharge to the alluvium during the spring and summer. Ground water moves slowly through the aquifers and eventually either reaches streams to become base flow or is discharged by wells and springs or by evapotranspiration.

An artesian aquifer is overlain by relatively impermeable beds and contains water under sufficient pressure to rise above the top of the aquifer if tapped by a well. Such a well is known as an artesian well. An artesian well flows when the pressure is sufficient to raise the water above the land surface (fig. 18).

A water-table aquifer is exposed at the surface and below a certain level contains a saturated zone. The surface at the top of the saturated zone, at which the pressure is atmospheric, is known as the water table. Water in a well casing will not rise above the water table. Ground water may be held above the main ground-water reservoir by relatively impermeable beds and is known as perched water.

Water-bearing zones in the Precambrian basement rocks are not distinctly separated so the rocks are treated as one aquifer, which is termed the Precambrian bedrock aquifer.

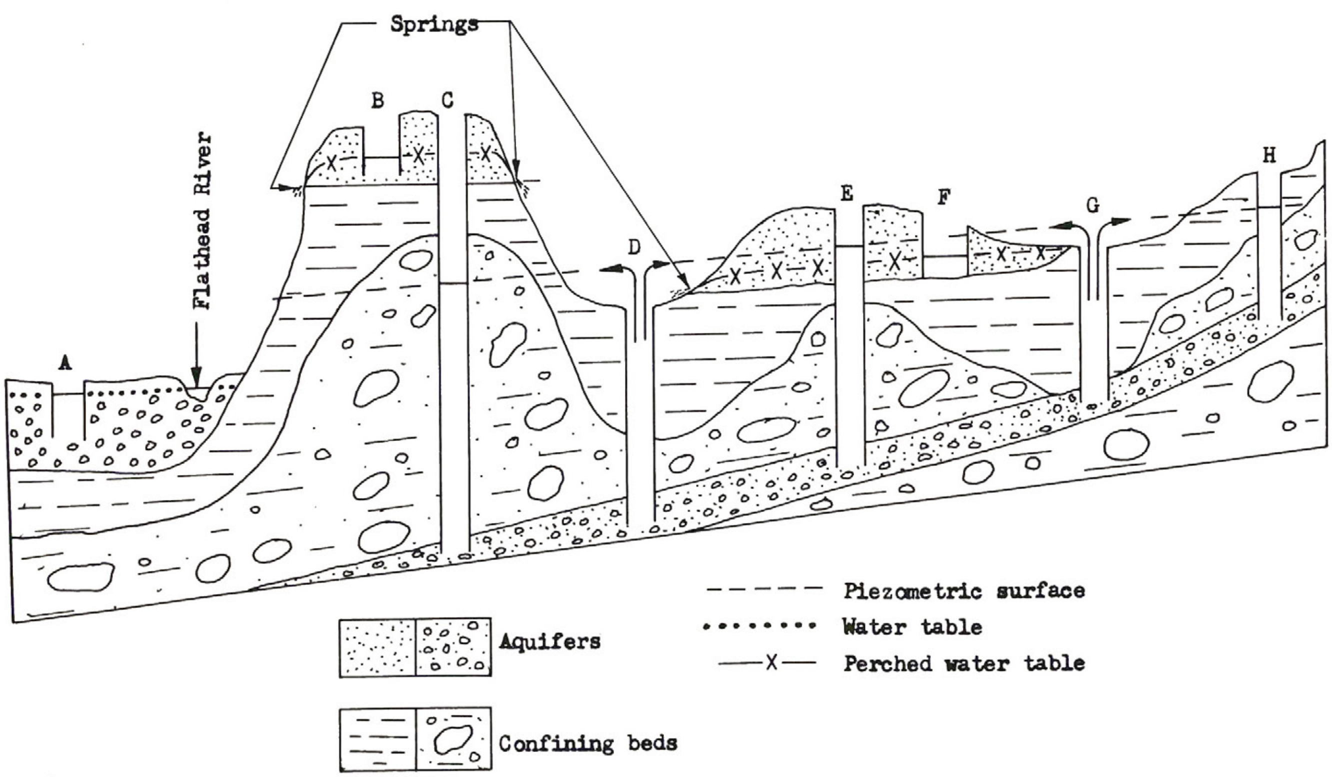


Figure 18.—Diagram showing various occurrences of ground water in the Kalispell Valley. A, water-table well; B and F, perched-water wells; C, E, and H, non-flowing artesian wells; D and G, flowing artesian wells.

The principal aquifers defined during the study are in the upper 600 feet of valley fill. These aquifers of Pleistocene and Recent age are glacier-related deposits of sand, gravel, and clay that may be intermixed as in deposits of drift, or they may be mostly sand and gravel as in beds of outwash.

Aquifers in the Pleistocene and Recent rocks may be broadly separated into artesian and water-table aquifers. For this report, the artesian aquifers are further separated on the basis of well depth and pressure-head differences into a Pleistocene deep artesian aquifer and Pleistocene shallow artesian aquifers. The water-table aquifers are separated on the basis of lithology, water-bearing characteristics, and dissimilar water-level fluctuations into Pleistocene perched aquifers, a Recent flood-plain gravel aquifer and a Recent flood-plain sand aquifer. Each of the catagorized aquifers is discussed in the following sections.

PRECAMBRIAN BEDROCK AQUIFER

WATER-BEARING PROPERTIES

In Precambrian bedrock, the primary openings or pore spaces between the grains are so minute that water moves through secondary openings such as joints or fractures. The joints or fractures have a small storage capacity and serve mainly as conduits. Water is continually supplied to the openings from precipitation and by leakage from adjacent deposits. Without replenishment, the water stored in joints and fractures would soon be depleted.

The Precambrian bedrock yields water to wells and springs where it crops out or where it is thinly covered by poorly permeable deposits. Wells have been drilled into bedrock along the east and west shores of Flathead Lake, at Bigfork and Somers, and southwest of Kalispell where the bedrock yields enough water for domestic use. It is not a source of large ground-water supplies.

Ground water in the Precambrian bedrock occurs under artesian conditions where the water is confined by overlying less-permeable deposits or the walls of joints in the bedrock. Some wells along the east and west shores of Flathead Lake have water levels many feet above the bedrock surface they penetrate. Well B26-20-12dd2, (190 feet deep) penetrated 82 feet of bedrock, and the static water level is about 78 feet above the bedrock surface. A few wells drilled into bedrock have sufficient pressure to flow but flows are generally less than 1 gpm (gallon per minute).

DEPTH AND YIELD OF WELLS

Depths of 26 wells producing from bedrock range from 35 to 467 feet. The wells were drilled 6 to 195 feet into bedrock, averaging about 77 feet. In general, if bedrock is near or at the surface, greater penetration of bedrock can be expected for a given yield than if the bedrock is buried by relatively impermeable unconsolidated sediments, but the well penetrating unconsolidated sediments and bedrock may have a greater total depth because of the thickness of overlying material. Well B27-21-35dd is 99 feet deep and was completed in 94 feet of bedrock, whereas well B27-20-13cd is 285 feet deep and was completed in only 33 feet of bedrock.

The yield of bedrock wells is variable and generally small. Reported yields of 24 wells range from 0.5 to 33 gpm and average about 9 gpm; 17 yield less than 10 gpm.

Reported yield is not an adequate measure of well success, as it does not take into account the drawdown required for a given yield. The productivity of a well and the permeability of the aquifer may be estimated from the specific capacity, which is the yield in gallons per minute divided by the drawdown in feet. For example, well B27-21-24cb, which is 467 feet deep and penetrates 185 feet of bedrock, was pumped at 33 gpm with 97 feet drawdown. The specific capacity, therefore, is about 0.3 gpm/ft (gallons per minute per foot of drawdown). The specific capacities of 24 wells range from 0.01 to 1.3 gpm/ft and average 0.2. The low specific capacity of bedrock wells indicates that large drawdowns are required for small yields.

PLEISTOCENE ARTESIAN AQUIFERS

The principal artesian aquifers in the valley are Pleistocene unconsolidated rocks, which were deposited by melt water and which underlie extensive deposits of till and silt. Two zones of sand and gravel are separated by a variable thickness of hard clayey till, known locally as hardpan, or by lacustrine silt. The lower zone occurs throughout most of the valley and is termed the deep artesian aquifer. The upper zone is delineated in a small area near Creston and in T. 29 N., R. 20 W. In these areas the water-bearing deposits are designated the shallow artesian aquifers.

The deep artesian aquifer is a series of beds of unconsolidated sand and gravel separated by discontinuous beds of fine-grained material. Because the beds of fine-grained material are discontinuous, the sand and gravel beds are hydraulically connected. The thickness of the deep artesian aquifer is generally unknown because the top has been penetrated only a few feet, except in well B29-20-27cb where the aquifer is at least 364 feet thick.

Overlying the deep artesian aquifer and hydraulically separating it from other aquifers are relatively impermeable beds of till and of lacustrine silt and clay. The till is generally poorly sorted, predominantly fine grained, and in some places well cemented. Its thickness ranges from 10 to 400 feet but probably averages about 100 feet. The lacustrine silt and clay is well bedded and well sorted. Its thickness ranges from 20 to 600 feet or more and averages about 200 feet.

The shallow artesian aquifer near Creston is a bed of outwash sand and gravel about 60 feet thick. Cemented gravel about 40 feet thick separates this aquifer from the deep artesian aquifer. Clay, cemented gravel, and boulder till about 75 feet thick overlie the aquifer.

The shallow artesian aquifer in T. 29 N., R. 20 W. is a group of lenses of sand and gravel in the till, or sand and gravel resting in shallow depressions on the till surface. It is generally overlain by silt and dune sand.

PLEISTOCENE DEEP ARTESIAN AQUIFER
HYDROLOGIC OPERATION

A description of the hydrologic operation of an aquifer includes discussion of movement, recharge, discharge, and changes in storage of water in the aquifer. The areas and sources of recharge and discharge may be determined in the field, but the direction and rate of movement are usually determined indirectly from piezometric maps. The horizontal direction of movement of water is downslope, nearly at right angles to the contour lines (lines of equal altitude of the piezometric surface). Piezometric maps for the Kalispell Valley were made from pressure-head measurements in wells, and the measurements were related to mean sea level by instrumental leveling.

The piezometric map of the deep artesian aquifer (pl. 3) was constructed from measurements made during August and September of 1965. The general slope of the piezometric surface is toward the flood plain of the Flathead and Whitefish Rivers. Water moves from the edges of the valley toward the rivers, where it is eventually discharged into the alluvium beneath the flood plain.

The steepening of gradient (greater than 50 feet per mile) shown by the narrow spacing of the contour lines south of Creston may be due to (1) ground-water discharge from many flowing wells, (2) a decrease in transmissibility of the aquifer, or (3) an area of high recharge a short distance to the east.

Near the swale along the base of the Swan Range (page 10), the hydraulic gradient steepens from 10 feet per mile on the west to 40 or 50 feet per mile on the east. This may be due to a ground-water barrier (pl. 3), as suggested by these facts: (1) the top of the deep artesian aquifer in well B29-20-4aa2, a quarter of a mile west of the depression, is 340 feet higher than the top of the deep artesian aquifer in well B29-20-3ba, half a mile to the east; (2) the water level in well B29-20-4aa2, which is west of the swale, fluctuated 2.68 feet during 1966 compared to 8.27 feet in well B30-20-34bd on the east; and (3) water in the aquifer east of the swale is somewhat more mineralized than water on the west. The ground-water barrier may be a fault that displaces the aquifer or it may be a zone of low transmissibility.

The quantity of water flowing downgradient through a cross section of the deep artesian aquifer was estimated by use of Darcy's equation, which may be expressed as

$$Q = TIL$$

where:

Q = rate of discharge in gallons per day
T = transmissibility in gallons per day per foot
I = hydraulic gradient in feet per mile
L = length of section in miles.

A 6-mile cross section was selected along the 2,940-foot piezometric contour line between Mill Creek and the Flathead River. The hydraulic gradient is about 12 feet per mile and the transmissibility is 3,400 gallons per day per foot. (The transmissibility was determined from a test on well B28-21-12a and is probably too low because the well did not fully penetrate the aquifer and because of excessive drawdown due to improper well construction.) By substituting these values in the equation, the flow is:

$$Q = TIL = 3{,}400 \times 12 \times 6 = 240{,}000 \text{ gpd or } 170 \text{ gpm.}$$

The rate of ground-water movement through this section is estimated to be less than 0.1 foot per day.

Water-level fluctuations in deep artesian wells may be placed in three categories: short-term, seasonal, and long-term. Short-term changes may be caused by atmospheric pressure changes or earthquakes and indicate aquifer characteristics. Seasonal fluctuations are the difference between the lowest and highest water level during the year and represent yearly changes in storage. Long-term changes show increases or decreases of water stored in an aquifer over a period of years.

The water level in artesian wells responds to changes in barometric pressure; a drop in barometric pressure causes water to rise in the casing, and a rise in barometric pressure causes it to decline (fig. 19).

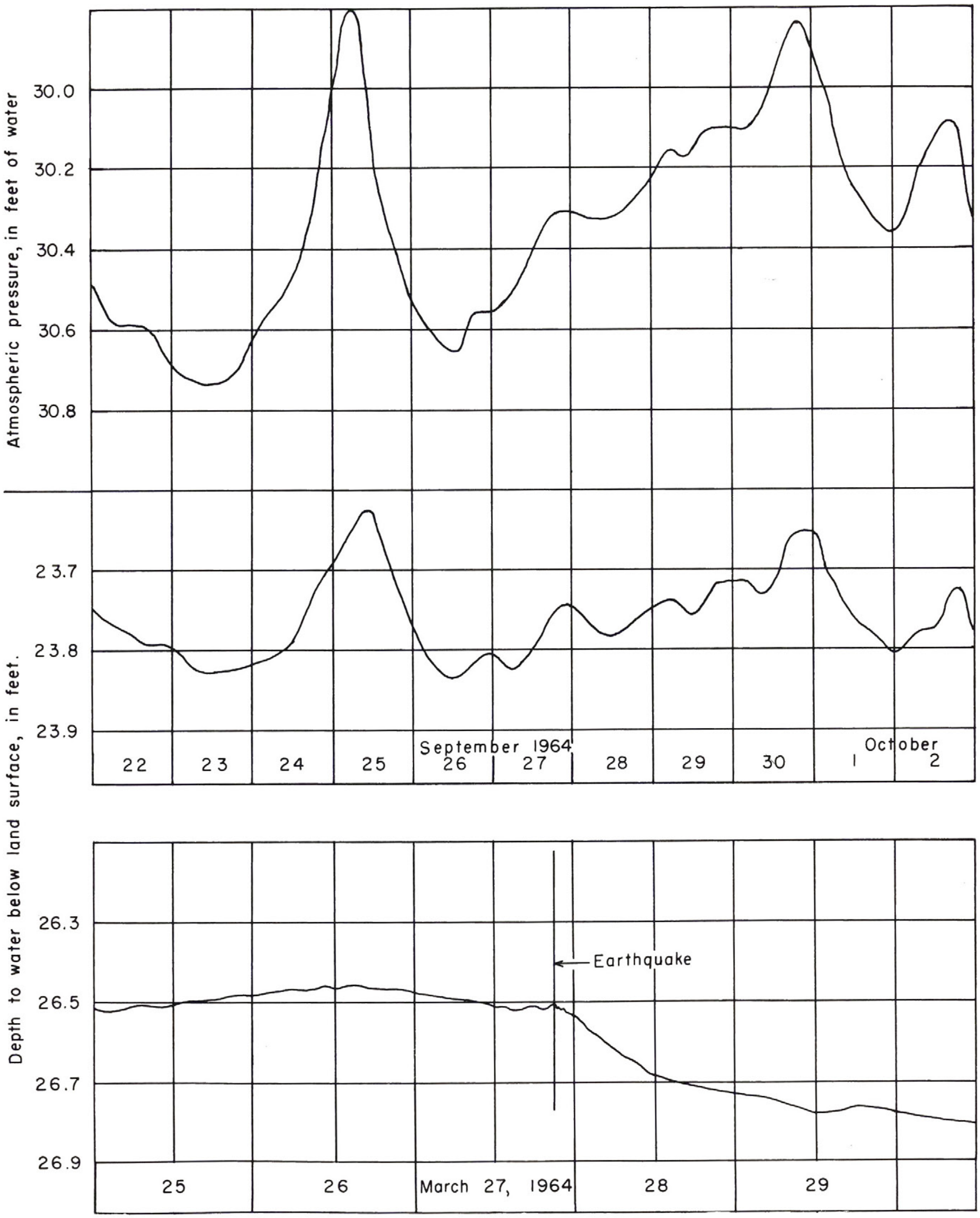


Figure 19.—Some examples of short-term water-level fluctuations in well B29-20-29bd due to barometric pressure changes and due to the Alaskan earthquake.

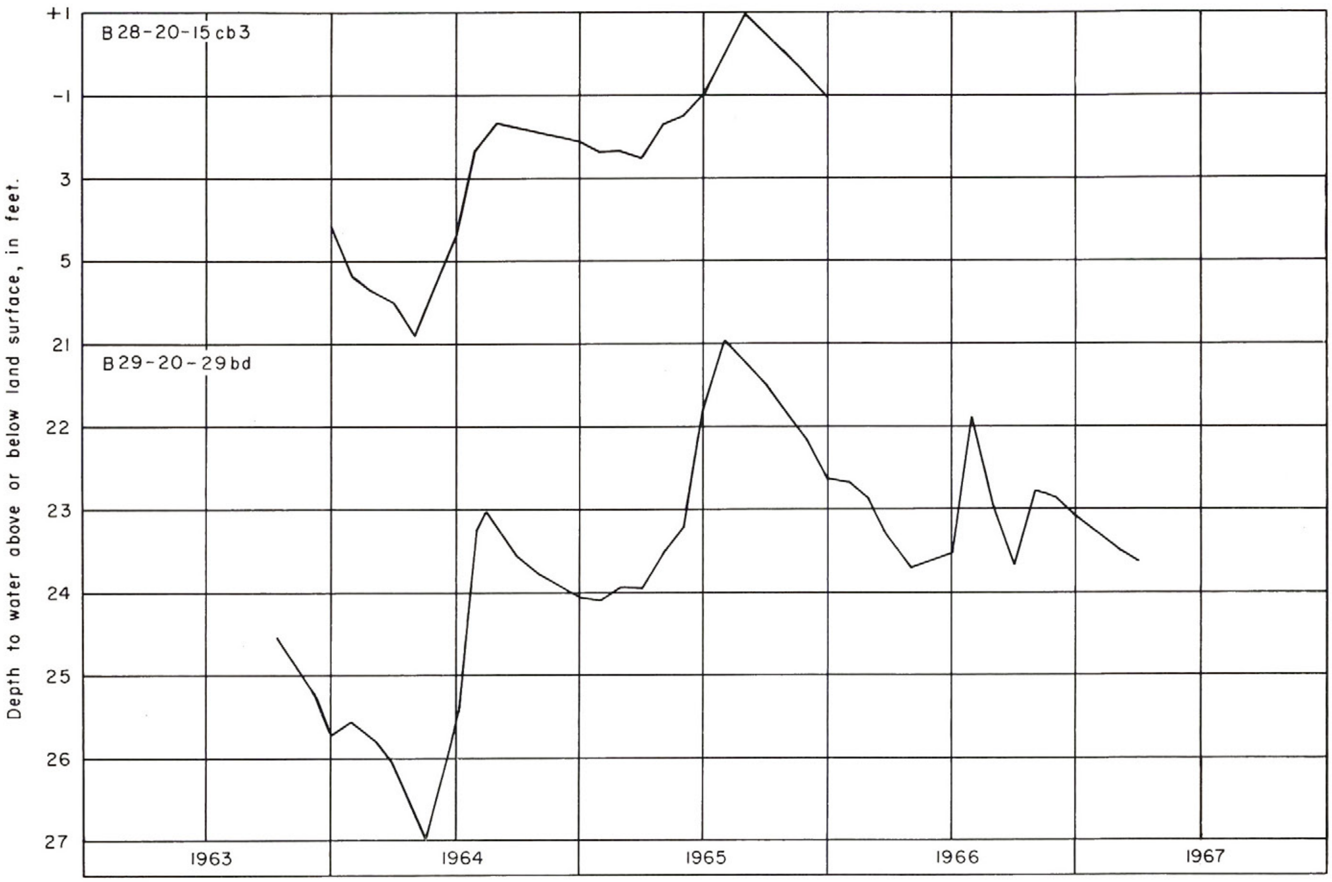


Figure 20.—Hydrographs of deep artesian wells.

The barometric efficiency of an aquifer may be determined by dividing the water-level change by the corresponding barometric pressure change (Ferris and others, 1962, p. 85) when both changes are expressed in the same units. For example, during September 25-26, 1964, the water level in well B29-20-29bd declined 0.21 foot and the barometric pressure rose 0.76 inches of mercury (0.86 foot of water). Therefore, the ratio 0.21/0.86 indicates a barometric efficiency of 24 percent.

The effects of an earthquake in Alaska on March 27, 1964, were recorded in well B29-20-29bd. These fluctuations are of very short duration and very sharp. The total fluctuation was about 0.65 foot (fig. 19).

The amount of ground water taken into and released from storage during the year varies from season to season. The water levels in wells generally are lowest in March or April and are highest in July. Greater than average precipitation in 1964 resulted in higher annual low-water levels the following year (fig. 20). Annual precipitation in 1964 was 22.36 inches compared with 15.22 inches in 1963. The lowest water level in well B29-20-29bd in 1964 was 27 feet below ground level and about 24 feet below ground level in 1965. In 1964, the lowest water level in well B29-20-15cb3 was about 7 feet below ground level whereas in 1965 it was only 2.5 feet below ground level.

The deep artesian aquifer is recharged by infiltration of snow-melt runoff from the mountains and by precipitation. Most recharge enters the aquifer where it is near the surface along the base of surrounding mountain ranges. Runoff in 1965 from the west-facing slope of the Swan Range between Echo Lake and Bad Rock Canyon was estimated to be about 70,000 acre-ft, on the assumption that unit runoff from the western side of the Swan Range is comparable to that from the west side of the Flathead Range. This runoff percolates into the ground at the base of the range. About 40,000 acre-ft of the runoff is discharged by contact springs near the base of the range, leaving no more than 30,000 acre-ft to recharge the deep artesian aquifer.

The deep artesian aquifer is discharged by flowing wells, seepage to other aquifers and into streams, evapotranspiration, and pumping. Measured well flows ranged from about 1 to 225 gpm. About 700,000 gpd (gallons per day) was discharged from flowing wells during 1965. Some wells near Creston did not flow in 1964 but were flowing when measured in 1965. Most of the flow from wells becomes streamflow, percolates downward to recharge shallow aquifers, or is evaporated.

Water may be discharged from an artesian aquifer by seeping upward through less permeable deposits to the land surface where it eventually is evaporated or transpired. Sloughs on the low-lying ground north of Flathead Lake may be supplied partly by ground water discharging from the deep artesian aquifer. The sloughs coincide with the trough in the piezometric surface north of the lake (pl. 3), where water from the aquifer migrates upward to the alluvium beneath the flood plain to be discharged either into the Flathead River or into the atmosphere by evapotranspiration from the sloughs.

Domestic and stock use from the deep artesian aquifer in 1966 was estimated to be 150,000 gpd or about 50 million gallons per year. This estimate is based on a per capita consumption of 50 gpd per person in rural areas, 20 gpd for milk cows, and 10 gpd for beef cattle and horses (MacKichan and Kammerer, 1961). Water pumped for irrigation during the peak of the irrigation season is estimated to be 1 million gpd. Kalispell has two deep wells, which supply supplemental municipal water. The total pumpage from these wells was 43 million gallons in 1964, 12 million gallons in 1965, and 60 million gallons in 1966. The Anaconda Aluminum Co. pumps about 1.4 billion gallons per year from three wells that apparently tap the deep artesian aquifer. A 300-foot well at the Creston National Fish Hatchery is used to help maintain fresh water of constant temperature in the fish tanks. The well flows 195 gpm during the summer and is pumped at 700 gpm in the winter.

WATER-BEARING PROPERTIES

Six aquifer tests were made to determine the coefficient of transmissibility (T) of the deep artesian aquifer. The coefficient of transmissibility is the rate of flow of water at the prevailing temperature, in gallons per day, through each vertical strip of the aquifer 1 foot wide extending the whole saturated thickness of the aquifer, under a hydraulic gradient of 100 percent. The data obtained from aquifer tests were analyzed according to the Theis recovery formula (Ferris and others, 1962, p. 100-102), which may be written

$$T = \frac{264Q}{\Delta s'}$$

in which

T = coefficient of transmissibility, in gallons per day per foot

Q = pumping rate, in gallons per minute

s' = residual drawdown of the water level, in feet

$\Delta s'$ = change in residual drawdown of the water level, in feet, per log cycle of time.

For example, when well B28-21-1ca was pumped at 6 gpm for 1 hour and then allowed to recover, the change in residual drawdown over one log cycle of time was 4.40 feet. Then

$$T = \frac{264Q}{\Delta s'} = \frac{264 \times 6}{4.40} = 360 \text{ gpd/ft}$$

The results of the six aquifer tests are in Table 1.

Table 1.—Results of aquifer tests of the deep artesian aquifer

Well number	Average discharge (gpm)	Drawdown after 1 hour of flow or pumping (feet)	Specific capacity (gpm/ft)	Coefficient of transmissibility (gpd/ft)
B28-20-16ad	5	6	0.8	2,800
B28-21-1ca	6	17	.4	360
B28-21-12a	325	167	1.9	3,400
B29-20-20ab	12	7	1.7	800
B29-20-20ca	50	38	1.3	3,800
B31-21-28cc	3	12	.3	550

The coefficient of transmissibility at flowing artesian well B31-21-28cc was obtained from a recovery test by use of a mercury manometer. The shut-in head was 21.1 feet above land surface. After the valve was opened, the well flowed an average of 3 gpm for 1 hour and the pressure head was lowered 12.3 feet. A recovery test was made by closing the valve and measuring recovery of pressure head. The computed value of T was 550 gpd/ft.

Most specific capacity estimates were from small-diameter domestic and stock wells (table 2). The average specific capacity is about 2 gpm/ft, and the average discharge is 31 gpm. The drawdown and discharge rate for each well were obtained from drillers' logs or short pumping tests. Many of the wells penetrate only a few feet of the aquifer and are open only at the end, which accounts for the low specific capacities. For example, wells B28-21-12a and B28-21-12ab are only about a quarter mile apart and the permeability of the aquifer is probably about the same at each well. Well B28-21-12a is 10 inches in diameter, penetrates 16 feet into the aquifer, and has six perforations per foot of casing. The specific capacity is 2 gpm/ft. Well B28-21-12ab is 7 inches in diameter, penetrates only 4 feet into the aquifer, and has no perforations in the casing. The specific capacity is only 0.4 gpm/ft. Partial penetration of the well and entrance losses to the casing increase the

Table 2.—Specific capacities of deep artesian wells

Well number	Drawdown (feet)	Discharge (gpm)	Specific capacity (gpm per foot)
B27-20- 3cd	3	2	0.7
10ab	2	1	.5
B27-21- 1ba	4	50	10
12ab	32	34	1.1
B28-20- 2bb	26	65	2.5
3dd2	25	40	1.6
10ad	48	25	.5
15bc	2	2	1
16ca	1	20	20
20dc	3	8	3
34cc	3	30	10
B28-21- 2ca	6	15	2
5cb1	5	15	3
6ba	145	10	.1
7dd	67	1,200	18
12a	166	325	1.9
12ab	83	30	.4
12cb	18	50	2.8
14dd	90	15	.2
19aa2	142	17	.1
20bb	47	1,500	32
20db1	55	25	.4
28bb	76	10	.1
B28-21-33bb	9	40	4
33cd1	102	20	.2
B28-22- 1dd	45	100	2.2
11ca	75	150	2.0
11cb	50	10	.2
12dc	165	100	.6
13ba	2	15	7
B29-20- 8cb	40	16	.4
9ac1	35	20	.6
10cb2	75	14	.2
18aa2	18	20	1.1
20ab	50	25	.5
27bb	100	12	.1
28ca	89	25	.3
29cd	28	30	1.1
33ca	43	35	.8
35dd2	14	15	1.1
B29-21- 5ba1	94	15	.2
19cb1	40	15	.4
20cc	42	15	.4
30cd	62	20	.3
B29-22-10dc1	105	12	.1
24dd	28	20	.7
25ad	43	12	.3
B29-22-27dc	13	10	.8
B30-20- 5cd	35	10	.3
6cb	13	8	.6
8ca	76	60	.8
18ba	6	50	8
20da	20	10	.5
27bd1	4	20	5
29ba	18	20	1.1
32dd	25	10	.4
B30-21- 5bb	40	20	.5
12dd	25	15	.6
13ad	35	10	.3
28ba2	66	8	.1
34da	96	13	.1
B30-22- 1db	32	20	.6
12ab	200	7	.04
13dd	30	6	.2
33ab	73	50	.7
34cd	31	50	1.6
B31-21-34ac2	45	20	.4
B31-22-24ca	38	60	1.6

drawdown. Thus the specific capacity is less than would be expected in a fully penetrating and fully screened well tapping the same aquifer (fig. 21). The average specific capacity of five irrigation wells is 6 gpm/ft. Their discharge ranges from 50 to 1,500 gpm and averages about 540 gpm. The average specific capacity of two municipal supply wells is 26 gpm/ft, and the average discharge is about 1,400 gpm.

The storage coefficient of the Pleistocene deep artesian aquifer may be estimated as described by Jacob (1940). If it is assumed that the hydrostatic pressure in the aquifer is effective over the entire plane of contact between the aquifer and overlying confining bed, the storage coefficient (S) is related to the barometric efficiency (BE) as follows:

$$S = (dpm/E)\,(1/BE)$$

where d is the specific weight of water, p the porosity of the aquifer, m the thickness of the aquifer, and E the bulk modulus of elasticity of water. The specific weight is 62.3 pounds per cubic foot, p assumed to be 0.40, m is 360 feet (p. 16), BE is 0.24 (p. 19), and E is approximately 300,000 pounds per square inch at the prevailing ground-water temperature. Substituting these values and a conversion factor into the equation results in

$$S = \frac{(62.3)\,(.40)\,(360)}{(144)\,(300{,}000)\,(.24)} = 8.7 \times 10^{-4}$$

This estimate is within the range that might be expected in an artesian aquifer and seems to be reasonable. If the aquifer thickness were better known and if the porosity were known, the estimate would be improved.

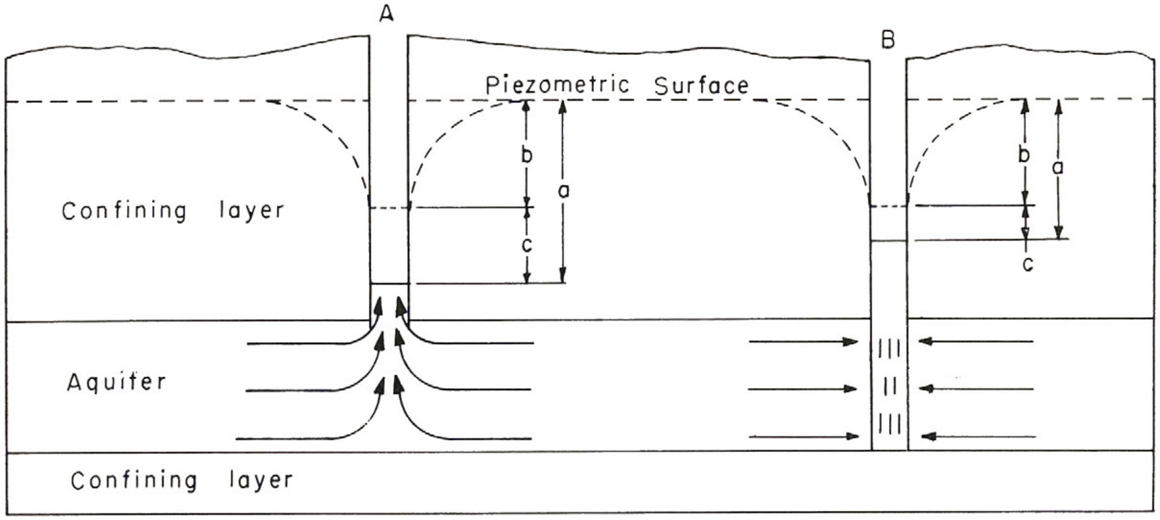


Figure 21.—Diagrammatic sketch comparing drawdown in a pumped well (A) with unperforated casing and (B) with perforated casing extending the full thickness of the aquifer. (a, total drawdown; b, drawdown owing to aquifer loss; c, drawdown owing to well loss; arrows indicate flow into casing.)

DEPTH AND TYPE OF WELLS

The depths of 122 wells that tap the deep artesian aquifer beneath the east and central valley terraces range from 109 to 405 feet and average 175 feet. The depths of 38 wells near Flathead Lake range from 125 to 480 feet and average about 230 feet; 48 wells west of the Stillwater River range from 111 to 340 feet deep and average about 195 feet. East of the ground-water barrier (pl. 3) on the east valley terrace the average depth of 5 wells is about 450 feet.

Nearly all recently drilled domestic and stock wells are constructed with 7-inch steel casing and are equipped with submersible pumps. The casing is normally not perforated, and water enters only from the bottom. Irrigation and municipal wells are constructed with 10- to 16-inch casing perforated opposite the water-bearing zone and are equipped with vertical turbine pumps.

The water levels may be considerably different in adjacent shallow and deep wells. For example, the depth to water in wells B30-21-28ba1 and B30-21-28ba2 in the same house basement was 97 feet greater in the 255-foot well than in the 26-foot well. Wells B29-20-18aa1 and B29-20-18aa2 are in the same well pit; one is 176 feet deep, and the water level is 62 feet below that in its companion well, which is 20 feet deep. In these two examples, the shallow wells tap unconfined aquifers and the companion wells tap the deep artesian aquifer. Other comparisons between depths to water in deep artesian wells and depths to water in shallow wells are in Table 3.

Table 3.—Comparison of depths to water in wells in deep artesian aquifer and overlying aquifers

Well location	Aquifer	Depth of well (feet)	Depth to water below land surface (feet)	Distance between wells (feet)	Remarks
B27-20-11ba	Perched sand and gravel	31	24.4	300	Land surface is 11 feet lower than 11bb
11bb	Deep artesian	220	23.8		
B27-21- 1ba	Deep artesian	346	5.5	2,100	Land surface is 6 feet higher than 36dd
B28-21-36dd	Flood-plain sand	19	12.1		
B27-21- 4ad1	Deep artesian	264	3.4	1,000	Land surface is 3.5 feet higher than 4ad2
4ad2	Flood-plain sand	14	13		
12ab	Deep artesian	480	+6.3	2,800	Land surface is 9 feet higher than 12ad; flowing well
12ad	Flood-plain sand	9	6.5		
B28-20- 7aa	Perched lacustrine sand	17	14.5	500	Land surface is 28 feet lower than 8bb
8bb	Deep artesian	165	30.1		
10dd1	Shallow artesian	105	+18.8	200	Flows 8 gpm
10dd2	Deep artesian	300	>+21.0		Land surface is 21 feet higher than 10dd1; flows 195 gpm
15cb1	Deep artesian	164	+4.6	25	Flowing well
15cb2	Perched lacustrine sand	70	15.1		

Table 3 (continued)

Well location	Aquifer	Depth of well (feet)	Depth to water below land surface (feet)	Distance between wells (feet)	Remarks
15db	Shallow artesian	103	+9.6		Flowing well; land surface is 29 feet lower than 15cb1 and 2; 15db is 2,600 feet from 15cb1 and 2
18bd1	Perched lacustrine sand	58	52.3	25	
18bd2	Deep artesian	405	22.7		
B28-21- 5cb1	Deep artesian	124	73.5	2	Both wells in same pit
5cb2	Perched lacustrine sand	87	26.1		
12a	Deep artesian	264	+11.7	1,800	Flowing well
12aa	Perched lacustrine sand	20	14.5		
20db1	Deep artesian	249	18.2	25	
20db2	Perched lacustrine sand	36	21.0		
27ab1	Deep artesian	264	+3.7	100	Land surface is 3 feet higher than 27ab2; flowing well
27ab2	Flood-plain sand	21	6.9		
33cd1	Deep artesian	184	9.5	10	
33cd2	Flood-plain sand	35	21.7		
B28-22- 1dd2	Perched lacustrine sand	30	26.4	200	
1dd3	Deep artesian	135	63.2		
11da1	Deep artesian	136	47.9	5	
11da2	Perched lacustrine sand	14	9.5		
B29-20- 4aa1	Perched dune sand	21	7.8	2	Both wells in same pit
4aa2	Deep artesian	170	70.5		
9ac1	Deep artesian	177	65.4	1,300	Land surface is 2 feet lower than 9dc1
9dc1	Shallow artesian	126	21.1		
5bb1	Perched dune sand	17	8.3	100	
5bb2	Deep artesian	151	80.3		
9ac1	Deep artesian	177	65.4	200	
9ac2	Perched dune sand	9	7.1		
33ca	Deep artesian	189	19.4	1,600	
33cd	Perched lacustrine sand	6	3		
B29-21- 5cd1	Deep artesian	131	82.9	50	
5cd2	Perched dune sand	17	8.2		
B29-21- 6aa1	Deep artesian	229	136.7	400	Land surface is 19 feet higher than 6aa2
6aa2	Perched lacustrine sand	11	8.6		
19ba1	Deep artesian	300	96.8	400	Land surface is 8 feet higher than 19ba2
19ba2	Perched dune sand	6	2.6		
19cb1	Deep artesian	286	92.9	200	
19cb2	Perched sand and gravel	100	46.5		
19dc	Perched dune sand	8	5.6	2,800	Land surface is 8 feet lower than 20cc
20cc	Deep artesian	278	100.8		
B29-22- 8ad	Deep artesian	211	173.9	1,600	Land surface is 13 feet higher than 8cc
8cc	Perched sand and gravel	72	55.6		
13dd1	Deep artesian	287	96.9	400	Land surface is 6 feet higher than 13dd2
13dd2	Perched dune sand	25	16.3		
27cc	Perched sand and gravel	11	6.9	300	Land surface is 12 feet lower than 28dd
28dd	Deep artesian	180	140.0		
B30-20-20dc1	Deep artesian	120	84.5	400	Land surface is 37 feet higher than 20dc2
20dc2	Perched dune sand	7	3.5		
27bd1	Deep artesian	445	49.5	200	Land surface is 7 feet higher than 27bd2
27bd2	Perched dune sand	24	19.7		
29cc1	Deep artesian	120	33.8	5	
29cc2	Perched dune sand	15	7.9		
32dc1	Deep artesian	160	73.5	2	Wells are in same pit
32dc2	Perched dune sand	10	5.3		
34bd	Deep artesian	425	11.7	800	
34ca	Perched dune sand	14	8.8		
B30-21- 9aa1	Deep artesian	120	17.6	10	
9aa2	Perched lacustrine sand	21	13.3		
10dd1	Deep artesian	164	26.7		Well inside of shallower well
10dd2	Perched lacustrine sand	23	18.2		
14ca1	Deep artesian	134	11.6	25	
14ca2	Flood-plain gravel	20	18.7		
B31-21-29dc1	Deep artesian	250	+36.4	25	Flowing well
29dc2	Perched lacustrine sand	36	13.7		
34ac1	Deep artesian	153	100.0	25	
34ac2	Perched sand and gravel	7	5.1		

CHEMICAL QUALITY OF WATER

Water from the deep artesian aquifer is primarily of the calcium and magnesium bicarbonate type and moderately hard. The concentration of dissolved solids in 24 samples ranged from 116 to 316 ppm (parts per million) and averaged about 200 ppm; one sample contained 848 ppm (table 4). The hardness is caused chiefly by calcium and magnesium, which reflect the calcareous composition of the bordering bedrock and of the material composing the aquifer. The average total hardness (as $CaCO_3$) of water from 25 wells was 182 ppm and ranged from 51 ppm to 245 ppm.

The chief chemical constituents in troublesome amounts are iron and fluoride. The concentration of iron in water from ten wells ranges from 0.5 to 3.8 ppm and averages 1.4 ppm as contrasted to a maximum of 0.3 ppm recommended by the U. S. Public Health Service (1962) for drinking water used on interstate carriers. Water from these ten wells stains plumbing fixtures, cooking utensils, and laundry. Commercial softeners are used to alleviate the problem. Water from well B30-20-27bd1 contains 4.4 ppm fluoride, and prolonged drinking of this water may cause dental fluorosis (U. S. Public Health Service, 1962). Water from this well tastes salty because it contains large amounts of sodium, potassium, and chloride.

Temperature of water from 25 wells that tap the deep artesian aquifer ranged from 48° to 66°F and averaged about 50°F, or about 7° above the mean annual air temperature. High water temperatures correspond to deep wells. For example, well B28-20-15cb3, which is 168 feet deep, yields water at 48°F, whereas well B27-21-12ab, which is 480 feet deep, yields water at 53°F. The warmest observed ground water (66°F) in the valley is from well B30-20-27bd1, which is 445 feet deep, and may obtain water from a fault zone.

PLEISTOCENE SHALLOW ARTESIAN AQUIFERS

The Pleistocene shallow artesian aquifer near Creston and another in T. 29 N., R. 20 W., are of minor importance in the valley, but are discussed because they are poorly connected to other aquifers. Water levels and well depths are considerably different between wells tapping the deep and shallow artesian aquifers (table 3).

HYDROLOGIC OPERATION

The piezometric surface of the shallow artesian aquifer near Creston (fig. 22) slopes west away from

Table 4.—Chemical analyses of water from selected wells and springs, Kalispell Valley, Montana

(Analyses made by Montana State Board of Health. Analytical results in milligrams per liter.)

Location	Date of Collection	Aquifer	Temperature (°F)	Iron (Fe)	Calcium (Ca)	Magnesium (Mg)	Sodium Potassium (Na+K)	Bicarbonate (HCO_3)	Carbonate (CO_3)	Sulfate (SO_4)	Chloride (Cl)	Fluoride (F)	Nitrate (NO_3)	Dissolved Solids	Total Hardness ($CaCO_3$)
B27-20- 3ab	11- 3-64	Deep artesian	49	0	31	15	1	159	0	6	2	0	0	146	138
B27-20- 8aa	9-29-65	Flood-plain sand	...	1.36	118	112	255	700	0	200	250	.6	208	1,480	755
B27-20-20ab	9-22-64	Flood-plain sand	...	14.12	108	20	20	406	0	15	31	0	.9	402	352
B27-20-26ab	9-29-65	Precambrian bedrock	...	.96	36	34	14	300	0	6	5	.2	.9	234	230
B27-21-12ab	9-22-64	Deep artesian	53	0	47	26	14	290	0	5	7	.1	0	220	224
B27-21-12dc	11- 4-64	Flood-plain sand	48	4.00	88	31	21	421	0	8	21	0	5.0	382	347
B28-20- 3bb	11- 3-64	Deep artesian	...	.20	43	18	6	223	0	7	2	0	0	180	180
B28-20-15cb3	11- 3-64	Deep artesian	48	.54	26	24	9	204	0	11	3	0	0	162	163
B28-20-18bd2	11- 4-64	Deep artesian	49	0	39	21	0	204	0	5	3	0	0	160	184
B28-20-20dc	9-22-64	Deep artesian	51	.54	45	13	9	214	0	3	5	.1	0	166	163
B28-20-22aa	9-22-64	Shallow artesian	50	.14	53	17	0	232	0	4	3	0	0	188	204
B28-21-1cb1	9-24-64	Deep artesian	...	0	33	12	12	177	0	5	5	0	0	140	133
B28-21- 2dd	11- 4-64	Perched lacustrine sand	...	.60	43	19	33	247	0	28	9	0	13.3	246	184

B28-21- 6da	4-20-54	Flood-plain gravel	....	0	47	9	4	183	0	2	5	.1	0	175	155
B28-21-7dd	2-10-55	Deep artesian	....	0	50	26	2	244	0	8	3	0	0	225	232
B28-21-15aa	9-30-65	Flood-plain gravel	....	0	38	11	12	190	0	4	4	0	1.1	150	140
B28-21-19bc2	2-29-65	Precambrian bedrock	....	.40	74	16	23	238	0	11	23	0	68	360	250
B28-21-20bb	6- 4-64	Deep artesian	....	.10	70	12	5	275	0	3	2	0	.5	238	225
B28-21-20db1	9-29-65	Deep artesian	....	1.46	58	22	2	280	0	3	3	0	0	220	235
B28-21-23ca	11- 4-64	Flood-plain sand	....	.34	57	14	24	198	0	20	22	0	46.7	288	199
B28-21-33cd2	9-29-65	Flood-plain sand	....	3.74	68	20	22	335	0	12	7	.1	0	296	250
B28-21-35cb	9-29-65	Deep artesian	51	0	42	26	19	293	0	4	5	.1	0	212	210
B29-20- 3ba	9-22-64	Deep artesian	....	.76	35	17	18	223	0	5	6	.7	0	180	158
B29-20- 9ac1	10- 1-58	Deep artesian	....	.10	40	51	6	351	0	10	14	.1	2	316	321
B29-20- 9cc	9-22-64	Perched dune sand	49	.1	72	9	2	247	0	6	3	0	7.5	234	214
B29-20- 9dc1	9-28-65	Shallow artesian	....	0	86	52	10	525	0	10	5	.2	0	430	430
B29-20-18aa2	9-22-64	Deep artesian	50	.60	59	19	3	271	0	5	2	.2	0	214	224
B29-20-33ca	11- 2-64	Deep artesian	....	3.80	45	26	12	290	0	3	1	0	.5	228	219
B29-21- 2aa	9-29-65	Flood-plain gravel	....	0	38	10	9	177	0	4	3	.2	0	132	135
B29-21- 7dd	9-22-64	Perched dune sand	....	.1	122	40	28	244	0	18	52	.1	292	788	469
B29-21-15cb	11- 2-64	Flood-plain gravel	....	.14	51	14	2	216	0	3	2	0	1.4	180	184
B29-21-19cb1	11- 3-64	Deep artesian	49	2.10	53	2	23	201	0	18	3	0	0	184	143
B29-21-20bc	11- 3-64	Perched dune sand	51	0	53	20	11	235	0	14	11	0	21.2	250	214
B29-21-21bd	9-23-64	Flood-plain gravel	....	0	57	12	0	223	0	2	2	0	1.8	194	194
B29-21-34cc2	9-29-65	Flood-plain gravel	....	0	52	17	3	238	0	5	4	.1	0	210	200
B29-22- 3cd	11- 3-64	Perched sand and gravel	....	0	43	26	0	235	0	5	0	0	8.0	194	214
B29-22-10dc1	11- 3-64	Deep artesian	....	0	37	4	12	134	0	20	2	0	0	116	107
B29-22-21bb	9-29-65	Perched sand and gravel	....	0	58	22	19	268	0	10	8	0	40	280	235
B29-22-27dd	9-22-64	Deep artesian	....	0	46	38	4	320	0	8	3	0	5.3	246	230
B29-22-34dc	11- 3-64	Perched sand and gravel	....	0	59	12	0	229	0	4	1	0	1.1	180	199
B29-22-35dd2	11- 3-64	Deep artesian	....	0	61	16	12	281	0	7	1	0	2.6	218	219
B30-20-19dd	9-22-64	Perched sand and gravel	47	.14	40	15	47	284	0	10	5	.1	14.3	280	260
B30-20-21da	9-28-65	Perched sand and gravel	....	0	80	18	15	342	0	9	5	.2	8.0	288	275
B30-20-27bd1	11- 2-64	Deep artesian	66	0	12	5	335	827	12	0	52	4.4	2.7	848	51
B30-20-32cb	11- 2-64	Perched dune sand	....	0	65	50	68	403	0	27	29	.2	135	600	367
B30-20-33bc	11- 2-64	Deep artesian	....	0	41	32	9	287	0	10	3	.2	.5	222	235
B30-20-34ca	11- 2-64	Perched dune sand	....	.14	126	11	39	381	0	29	12	.2	99	546	362
B30-21- 9dd	9-29-65	Perched lacustrine sand	....	0	78	25	0	330	0	11	5	.2	5.3	300	295
B30-21-21dd	9-29-65	Perched dune sand	51	.12	124	30	39	387	0	35	44	.1	122	630	435
B30-21-26ba	9-23-64	Flood-plain gravel	....	0	61	13	5	244	0	9	3	.2	3.2	210	204
B30-21-28ba1	9-22-64	Deep artesian	50	2.66	57	13	9	244	0	9	2	.1	2.1	204	194
B30-22-12ab	9-29-65	Deep artesian	....	1.16	50	17	30	310	0	0	4	0	0	228	195
B30-22-25aa	9-22-64	Deep artesian	....	.50	69	18	19	320	0	15	5	.1	0	258	245
B31-21-28cc	11- 3-64	Deep artesian	....	0	35	15	3	183	0	1	3	0	0	142	148

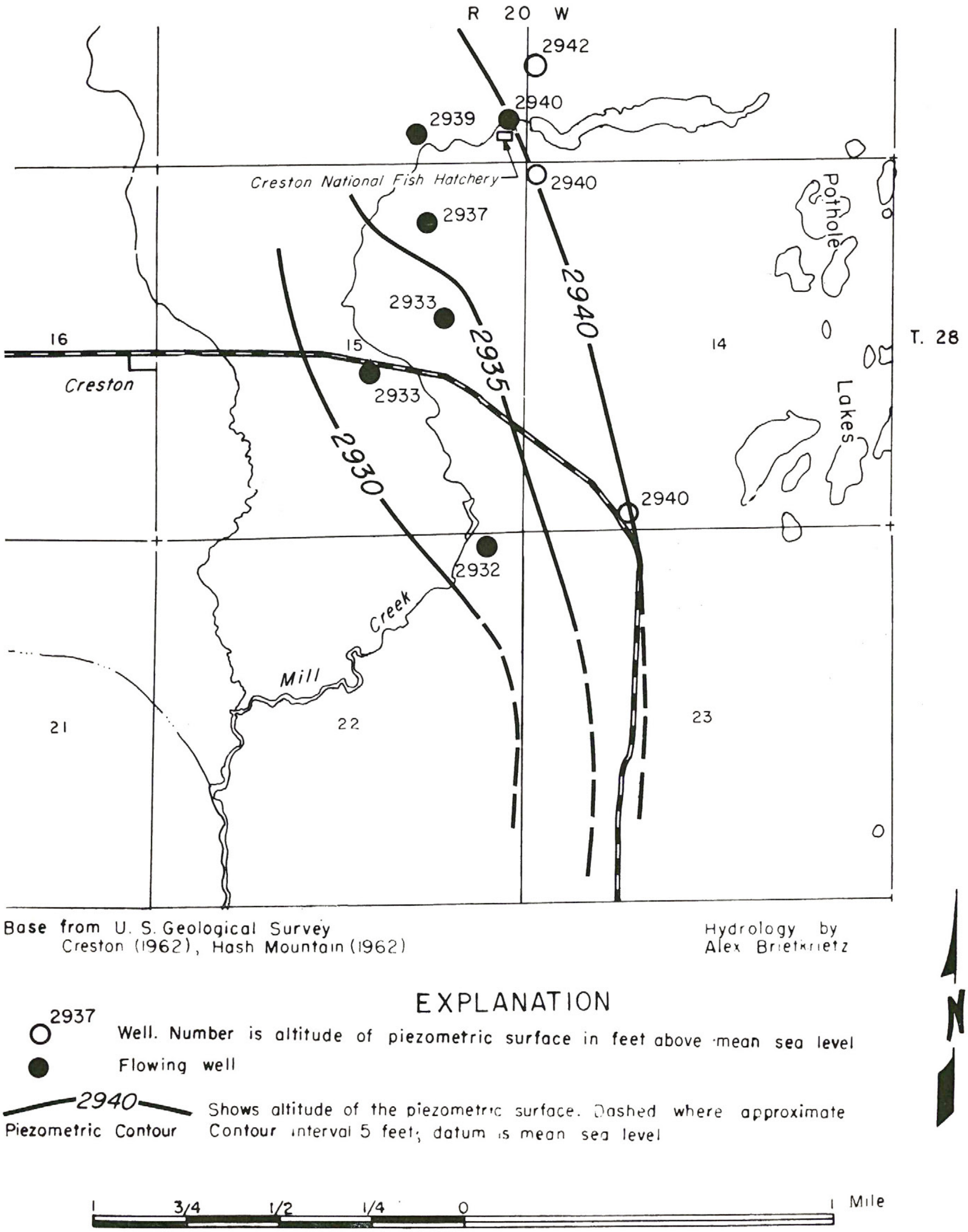
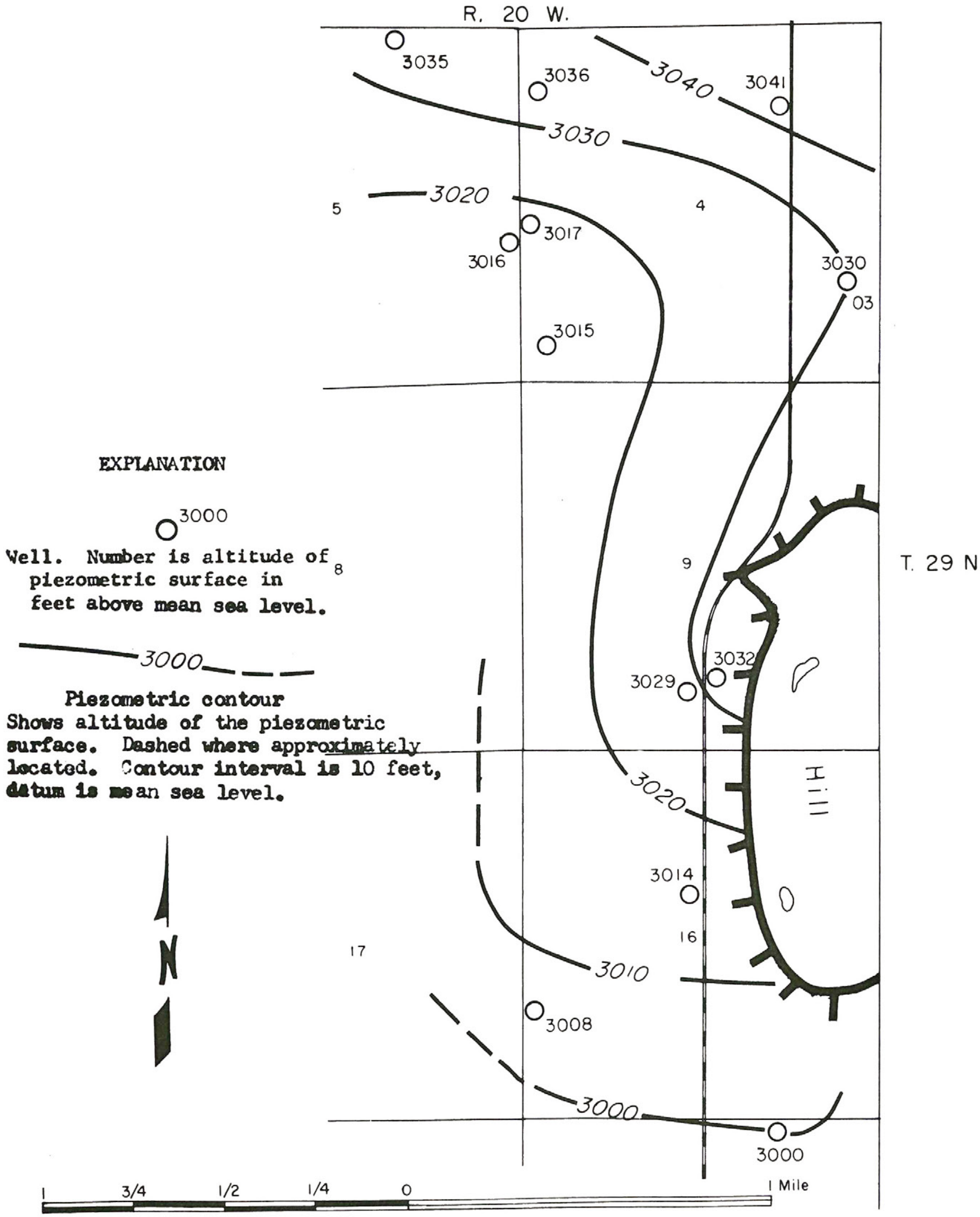


Figure 22.—Contours on the piezometric surface of the shallow artesian aquifer in the Creston area.



Base from U.S. Geological Survey
Columbia Falls South, 1962

Hydrology by
Alex Brietkrietz

Figure 23.—Contours on the piezometric surface of the shallow artesian aquifer in T. 29 N., R. 20 W.

the pothole lake area, indicating recharge to the aquifer from the pothole lake area. The piezometric surface averages about 20 feet lower than the piezometric surface of the deep artesian aquifer.

Ground water in the shallow artesian aquifer in T. 29 N., R. 20 W., (fig. 23) flows southwestward, except in sec. 9 and 16, where it flows west around a drumlin-like deposit of relatively impermeable drift. The gradient of the piezometric surface is about 20 feet per mile, and the surface averages about 50 feet higher than the piezometric surface of the deep artesian aquifer.

When the pothole lakes are full, seasonal flowing wells start to flow, and perennial flowing wells yield more water. Most discharge is by flowing wells, which flow 3 to 10 gpm and discharged about 46,000 gpd during 1965. Measured flows were somewhat less in 1966 than in 1965. In 1965, wells B28-20-10dc and B28-20-10dd1 flowed 10 and 5 gpm, respectively, and in 1966 flowed 8 and 2 gpm, respectively. Pumpage from domestic and stock wells is roughly 10,000 gpd.

WATER-BEARING PROPERTIES

An aquifer test was attempted on a 105-foot well (B28-20-10dd1) flowing 8 gpm at the Creston National Fish Hatchery. The well was shut in and the head recovered to 18.8 feet above land surface within a few minutes. The well was allowed to flow and the head abruptly dropped to 2 feet above land surface. Transmissibility was not determined but the specific capacity was 0.5 gpm per foot of drawdown. The specific capacity of flowing well B28-20-10dc was 0.3 gpm per foot of drawdown.

PLEISTOCENE PERCHED AQUIFERS

Three perched aquifers are separated from the underlying artesian aquifers by relatively impermeable clay, till, or cemented gravel. The perched aquifers are:

1. Dune and lacustrine sand of low permeability, which crops out on the terraces. As wells in this aquifer may go dry during years of below-average precipitation, many owners have drilled wells to the underlying deep artesian aquifer.

2. Outwash sand and gravel, which fills depressions between drumlins of clayey boulder till northwest of Kalispell. The thickness is unknown but some wells have penetrated as much as 70 feet of the sand and gravel.

3. Glacial drift of relatively well sorted sand and gravel in the pothole lake area. The drift is relatively permeable, but because surface water is abundant, few wells have been drilled into it.

HYDROLOGIC OPERATION

In April 1965, depths to water were measured in wells tapping the perched aquifers, and a perched water-table map was constructed (pl. 4). The map shows that ground water moves in several directions

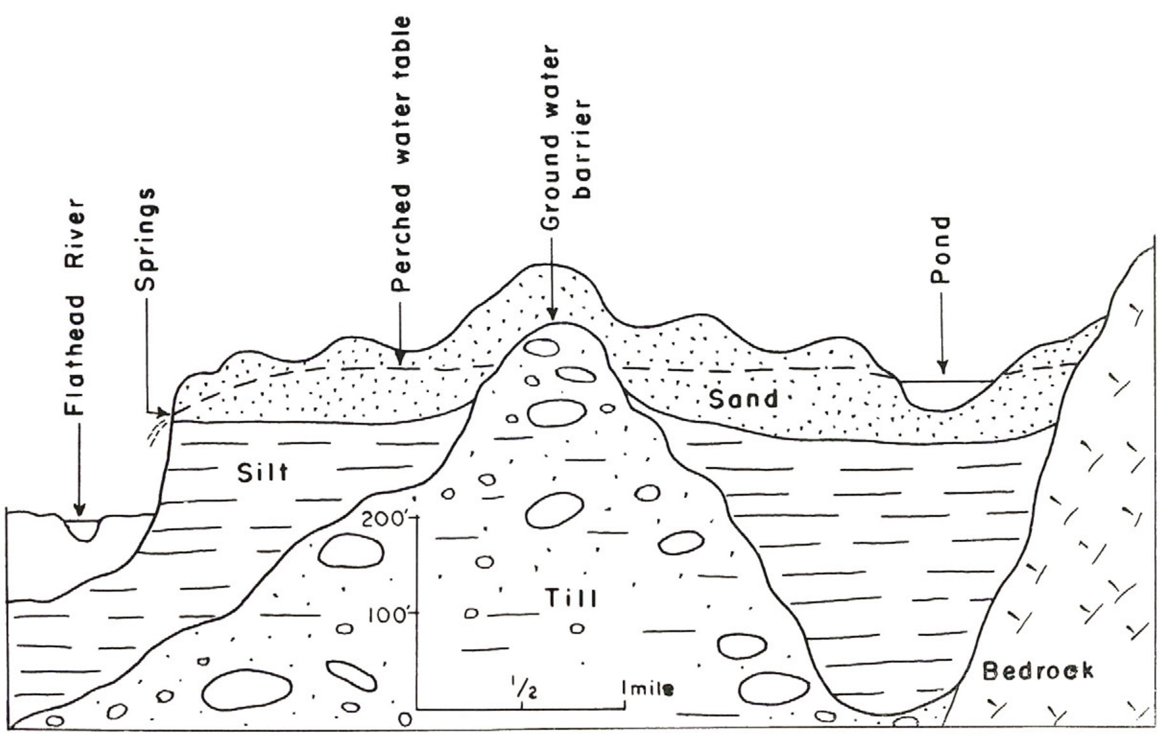


Figure 24.—Sketch showing ground-water barrier.

beneath the valley terraces because of well-defined ground-water barriers of relatively impermeable material (fig. 24). The ground-water barriers coincide with prominent drumlin-like features of relatively impermeable drift. Several miles from the barriers, ground water moves roughly parallel to the slope of the land surface.

Most recharge to the dune and lacustrine sand is from precipitation. The amount of recharge depends upon the amount and seasonal distribution of

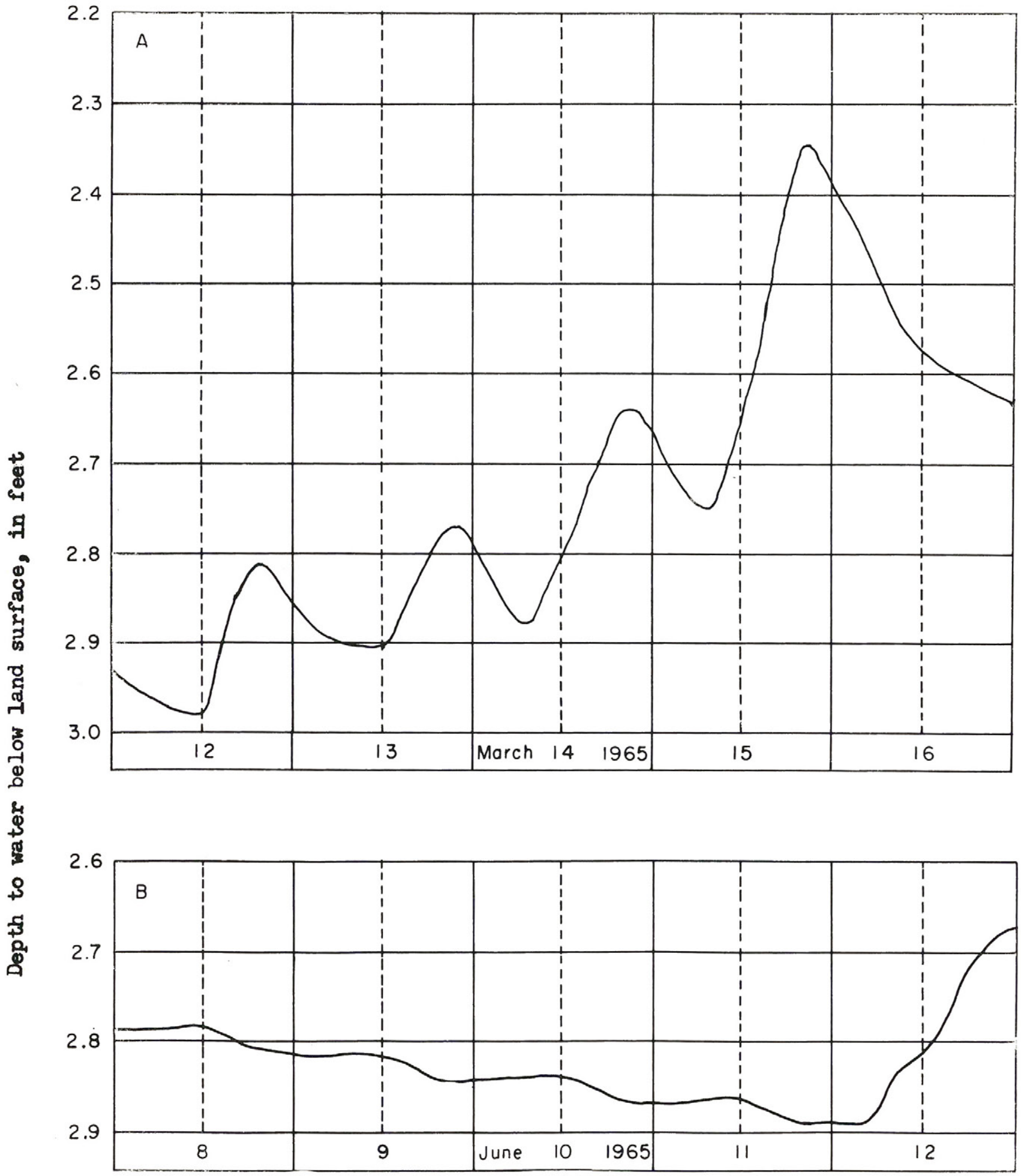


Figure 25.—Hydrographs of well B29-21-17ccl showing (A) diurnal fluctuations caused by snowmelt, and (B) diurnal fluctuations caused by evapotranspiration and by response to precipitation.

precipitation, vegetative cover, and the permeability of the soil. The dune sand forms a hummocky topography sparsely covered with grass. Owing to the high infiltration capacity of the soil, much of the precipitation recharges the aquifer.

Two excellent examples of the correlation between fluctuations of the perched water table and precipitation are shown in Figure 25. Snowmelt may recharge the sand when the daily air temperature exceeds 32°F and the ground is not frozen. Figure 25 shows diurnal water-level changes in well B29-21-17cc1 caused by daily temperature variations. On March 15, 1965, the water level rose 0.40 foot in response to a maximum temperature of 50°F. When the air temperature dropped below 32°F, the water level dropped. On March 16 the maximum daily temperature was below freezing, and there was no recharge from snowmelt. On the basis of diurnal fluctuations in well B29-21-17cc1 and an assumed specific yield of 0.10, the average recharge rate from snowmelt was calculated to be 0.05 foot per day. Therefore, from March 12 to 14, about 45 acre-ft of water recharged the 300-acre area near the well where the depth to water and permeability are relatively constant. On June 12-13, 1965, 1.37 inches of rain caused the water level in well B29-21-17cc1 to rise 0.22 foot.

Recharge to the outwash sand and gravel northwest of Kalispell is mainly seepage from Lost, Big Lost, and O'Neil Creeks. The combined flow of these three creeks was about 30 cfs on May 9, 1967, where they first cross the outwash sand and gravel; almost the entire flow disappears within 2 miles. Cedar Creek north of Columbia Falls also loses much water by seepage. On April 27, 1966, Frank Stermitz estimated that seepage from the creek was about 34 cfs in 1.25 miles of stream channel (written communication, 1966).

Discharge from the dune and lacustrine sand aquifer is by: (1) transpiration (2) evapotranspiration from ponds (3) flow from springs and seeps, and (4) pumping from wells. Wherever the water table in the dune and lacustrine sand aquifer is near land surface, water is discharged to the atmosphere during the summer as shown by diurnal water-level fluctuations caused by daily changes in temperature (fig. 25). Highest daily water levels occur in the morning when the air is relatively cool. Water levels start to decline about noon, when the air warms and evapotranspiration rates increase. About 10 p.m., the water level stops declining because the air has cooled and evapotranspiration rates have decreased. Calculations based on diurnal fluctuations in well B29-21-17cc1 and an assumed specific yield of 0.10 indicate that about 0.03 foot of water per day was discharged. Therefore, about 45 acre-ft of water was evaporated or transpired between June 7 and 11 from the 300-acre area near the well where the depth of water and permeability are relatively constant. The maximum daily air temperatures during this period ranged between 75°F and 90°F.

Some of the ground water in the dune sand becomes trapped in depressions in the surface of the underlying till when the perched water table falls below the lip of the depressions (fig. 26). In a small area of trapped water in sec. 1 and 12, T. 29 N., R. 22 W., the water table is never higher than the divide, because evapotranspiration is equal to or greater than recharge.

Water from some permanent shallow perched water-table ponds on the terraces, for example, Morning Slough in sec. 3, T. 29 N., R. 20 W., is transpired by cattails or other plants, is evaporated, or flows overland to Lake Blaine. Water from a pond in sec. 28 and 29, T. 29 N., R. 20 W., is pumped for irrigation.

Contact springs and seeps flowing from the

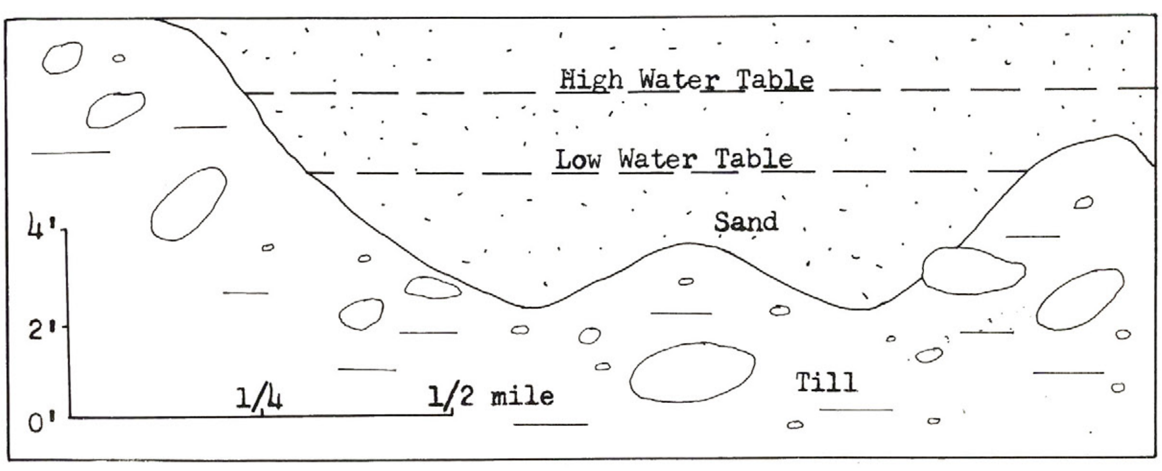


Figure 26.—Sketch showing the occurrence of trapped and free ground water in the dune sand.

perched water are common along the scarps of the terraces at the contact between the dune sand and the lacustrine silt or till (fig. 24). Most of the springs are a few feet to about 60 feet above the Flathead and Whitefish Rivers. Measured yields of these springs range from 1 to 95 gpm; about 300,000 gpd is discharged from them. Some are used for stock water and domestic supplies and one is used for irrigation.

Much water is discharged from numerous contact springs along the western edge of the pothole area east of Creston and from contact springs northwest of Kalispell. The springs east of Creston discharge water from the pothole lakes and issue along the edge of permeable gravelly drift overlying a relatively impermeable, hard, thick-bedded, lime-cemented gravel deposit. The springs northwest of Kalispell issue at the contact of very permeable sand and gravel overlying less permeable clayey till. Discharge measurements of selected contact springs throughout the valley are in Table 5.

Table 5.—Discharge of selected contact springs

Spring number	Date of measurement	Discharge (gpm)
B27-20-11ac	9- 2-66	6,500
B28-20-11dc	9- 1-66	15,300
B28-20-23cc	9- 2-66	200
B28-22-10aa	8-31-66	800
B29-22- 3cd	9-14-66	900
B28-22-34cd	8-31-66	1,300
B29-22-34dc	8-31-66	450

In the last 10 years, many wells penetrating the dune and lacustrine sand have been replaced by wells drilled to the deep artesian aquifer. The estimated domestic pumping rate from the dune and lacustrine sand aquifer has decreased 10,000 gpd in the last 10 years; in 1966 about 53,000 gpd was pumped.

Most of the pumpage from the outwash sand and gravel aquifer northwest of Kalispell is for irrigation. About 300 acre-ft is used during the irrigation season, and 10 acre-ft is pumped each year

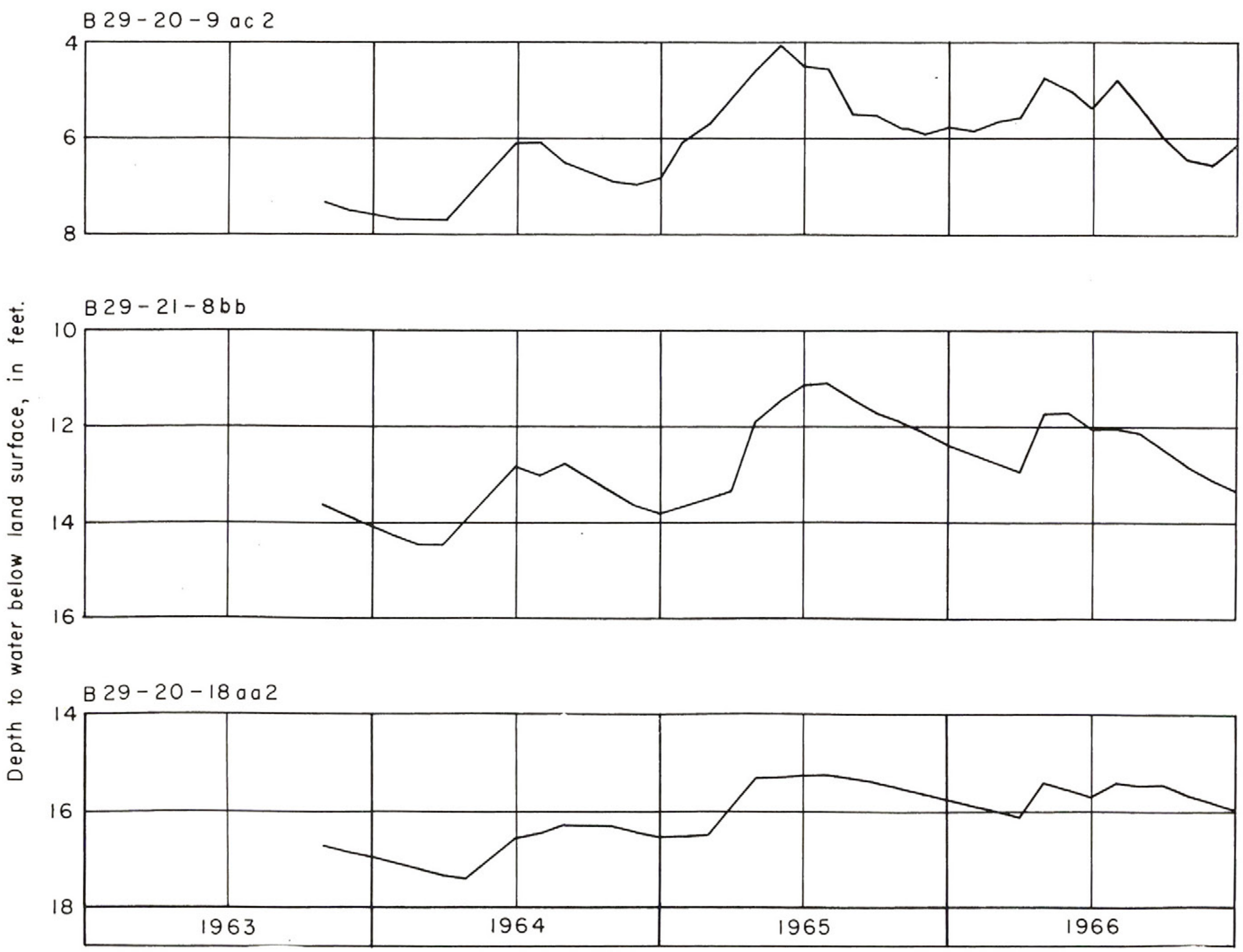


Figure 27.—Hydrographs of wells tapping dune sand.

for domestic uses. Of the total pumped for irrigation, about 22 acre-ft is from spring B29-22-3cd.

A net gain in ground-water storage in the dune sand between March 1964 and March 1966 is shown by three hydrographs (fig. 27). The water level in well B29-20-9ac2 rose 2.1 feet; in well B29-21-8bb, 1.5 feet; and in well B29-20-18aa2, 1.2 feet. Precipitation during 1964 was above average and caused the increase of ground-water storage. The increase of storage in the dune sand between 1964 and 1966 was estimated as 10,000 acre-ft. This estimate was made by multiplying the surface area of the aquifer (about 67,000 acres) by the average change in water level between 1964 and 1965 (1.5 feet) and by an assumed storage coefficient of 0.1. The total amount of water in storage at the end of 1965 was calculated as 180,000 acre-ft in the estimated average saturated thickness of 27 feet.

Changes in ground-water storage in the pothole-lake area are indicated by observations of water-level fluctuations in Echo, Cabin, and Plummers Lakes (fig. 28), which seem to be representative of the pothole lakes. During 1965 Echo Lake was highest in July, Cabin Lake in September, and Plummers Lake in August. The high water level in Cabin Lake occurs latest because the lake has no surface inlet and is farthest from streams that recharge the area. Echo Lake has a surface inlet, which accounts for its peak in July. Echo Lake rose 4.3 feet between May and July 1965 for a gain in lake storage of 3,000 acre-ft. From July 1965 to May 1966, the lake level dropped about 5 feet for a loss of 3,500 acre-ft. The net loss between May 1965 and May 1966 was 500 acre-ft. The other pothole lakes including Lake Blaine gained roughly 6,000 acre-ft in the summer of 1965 and the losses were about 6,000 acre-ft during the winter of 1965-66. Based on the average lake fluctuation of 4.3 feet between May and July of 1965, it is estimated that the gain in ground-water storage in the drift in the pothole-lake area was about 11,000 acre-ft.

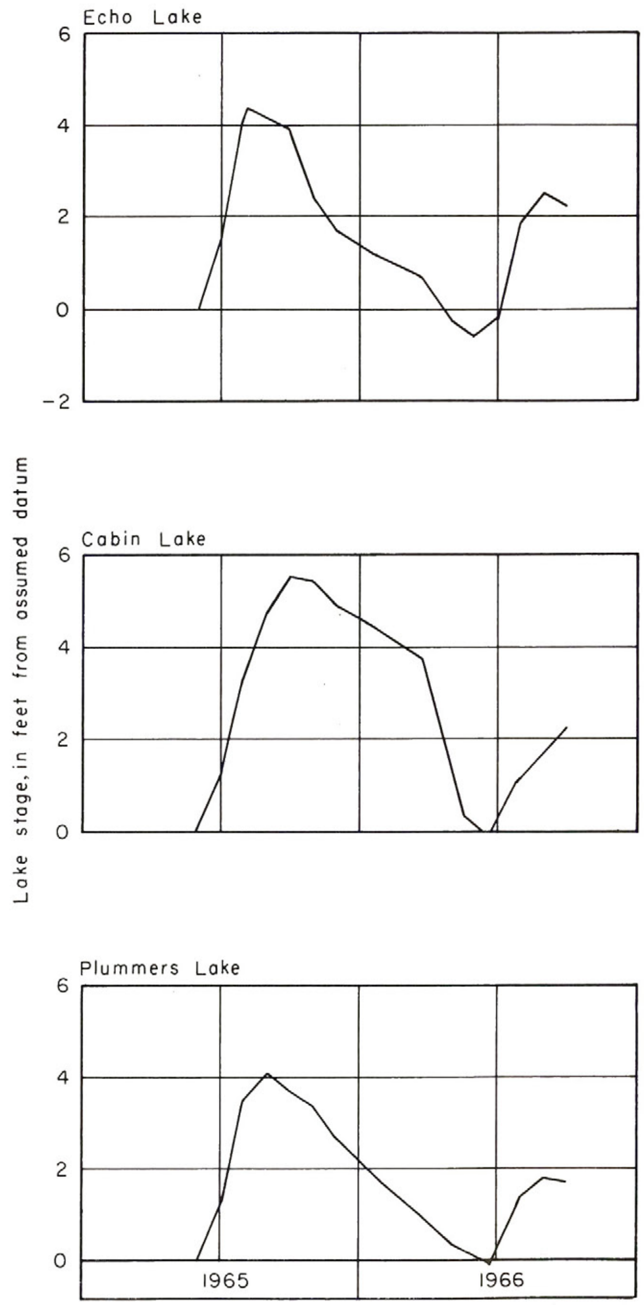


Figure 28.—Hydrographs of three pothole lakes.

WATER-BEARING PROPERTIES

Aquifer tests were made for two wells that tap the outwash sand and gravel northwest of Kalispell. Well B29-22-17dd was pumped at 183 gpm. The maximum drawdown was 2 feet, indicating a specific capacity of 92 gpm/ft. A coefficient of transmissibility of 570,000 gpd/ft was determined from the test data. Well B29-22-9cc was pumped at 12 gpm. The test was not satisfactory, however, because after 2 minutes of pumping, the maximum drawdown of 1.88 feet was reached. The specific capacity of this well is about 6 gpm/ft.

DEPTH AND YIELD OF WELLS

The depths of 95 wells that tap the dune and lacustrine sand on terraces range from 6 to 40 feet and average 20 feet. Most are 3 to 6 feet in diameter, dug by hand, cribbed with concrete casing, and completed only a few feet below the water table. A few wells are driven sand points. Sustained yields are small, generally less than 10 gpm, but the large-diameter dug wells provide storage, and initial pumping rates can be more. If dug wells are pumped too heavily, the sand "heaves" into the casing.

The depths of 14 wells that tap outwash sand and gravel northwest of Kalispell range from 22 to 94 feet and average 53 feet. Drilled wells are constructed with 7-inch steel casing, which extends 10 to 20 feet below the perched water table. The yield is more than adequate for domestic and stock use. Dug wells are cribbed with large-diameter concrete casing and are completed only a few feet below the water table. Irrigation wells are constructed with large-diameter (mainly 48 inches), galvanized-iron

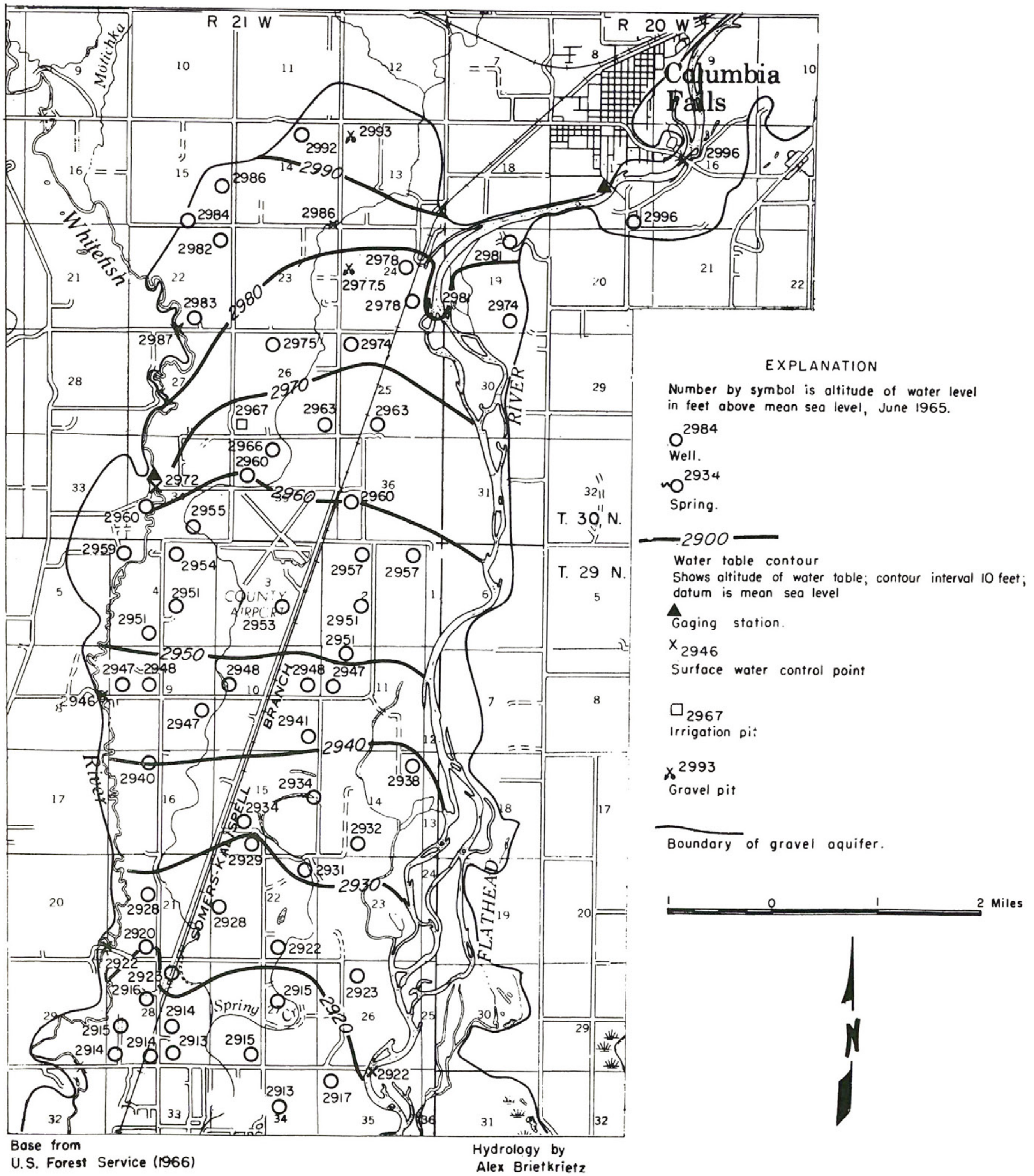
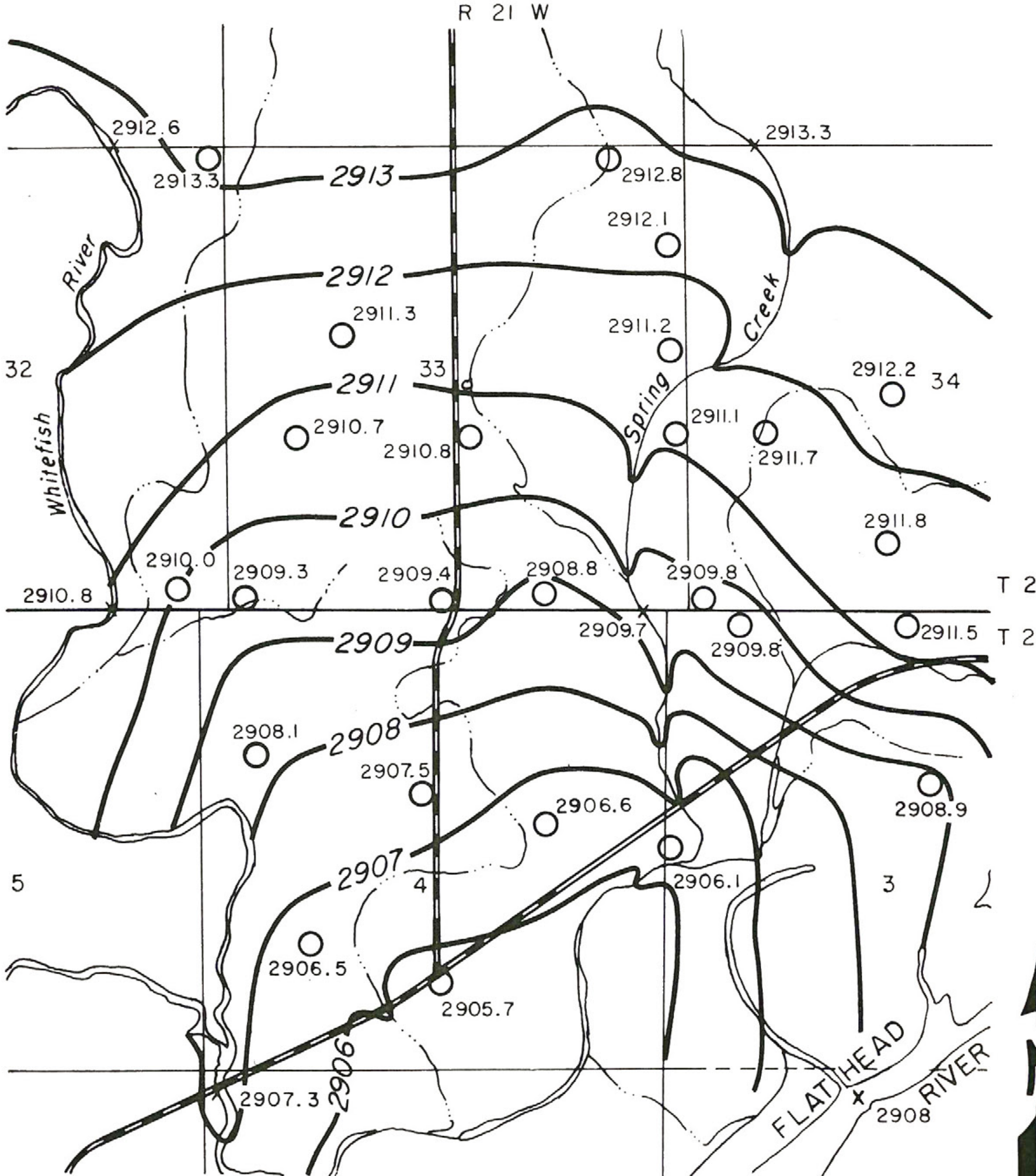


Figure 29.—Contours on the water table in the gravel aquifer between Columbia Falls and the Evergreen area.



Base from U. S. Geological Survey
Kalispell (1962)

Hydrology by
Alex Brietkrietz

EXPLANATION

O 2908.1 — Well completed in gravel aquifer. Number is altitude of water level, in feet above mean sea level, June, 1965.

2910 — Water-table contour
Shows altitude of water table; contour interval 1 foot; datum is mean sea level.

x 2908 — Surface water control point. Number is altitude of water surface, in feet above mean sea level.

1 1/4 1/2 3/4 0 1

Scale 1: 24 000

Figure 30.—Contours on the water table in the gravel aquifer, Evergreen area.

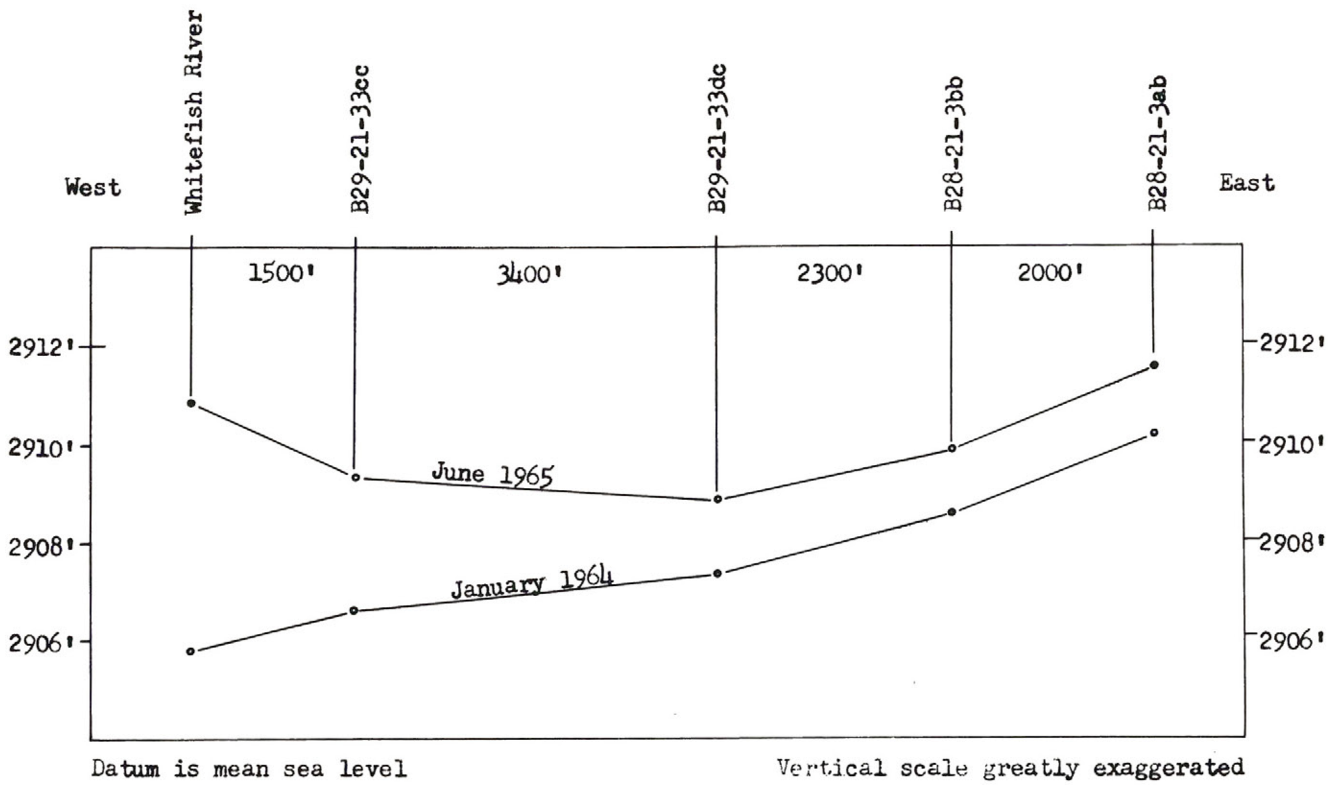


Figure 31.—Water-table profiles, Evergreen area.

casing. Most of these wells are dug with a backhoe and completed a few feet below the water table; their yields range from 70 to 200 gpm. This type irrigation well can be dug deepest when the perched water table is lowest, generally in March or April.

CHEMICAL QUALITY OF WATER

Water samples from the dune and lacustrine sand were collected from 13 sites (table 4); the water is a calcium bicarbonate type and generally contains the greatest concentration of dissolved solids of any water sampled throughout the valley. The dissolved solids range from 234 to 788 ppm and average 449 ppm. The water is very hard; total hardness ranges from 184 to 469 ppm and averages 318 ppm. The hardness may reflect a large percentage of limestone particles in the sand. Nitrate concentration ranged from 5.3 to 292 ppm; water from four domestic wells contained more than the recommended limit of 45 ppm. Significant concentration of nitrate in ground water may be caused by contamination from barnyards or domestic sewage, by oxidation of organic matter in the soil, or by commercial fertilizers applied to fields. The trapping and relative stagnation of water (p. 30) probably also contributes to the nitrate concentration. Excessive concentration of nitrates in drinking water may cause serious blood changes in infants (Comly, 1945).

The temperature of water from several shallow wells ranged from 50 to 52°F in July and from 47 to 50°F in December. Water from several contact springs was 46 to 48°F in July.

RECENT FLOOD-PLAIN AQUIFERS

The alluvium underlying the flood plain of the Flathead and Whitefish Rivers north of Kalispell (north of sec. 21, 22, 23, T. 28 N., R. 21 W.) is gravel containing a few lenses of sand and silt—the gravel aquifer. The gravel aquifer averages about 28 feet thick and partly fills a trough cut into relatively permeable till or silt. The aquifer is the most permeable in the Kalispell Valley and yields water to domestic, stock, industrial, and irrigation wells. Many wells were contaminated by floor waters of June 1964, however, and the aquifer is subject to man-caused pollution because of its permeability. South of Kalispell, the gravel abruptly grades downstream into sand and silt—the sand aquifer. The sand aquifer is only slightly permeable and yields poor water to domestic and stock wells.

THE GRAVEL AQUIFER

HYDROLOGIC OPERATION

Water-level measurements were made during June 1965 in shallow wells, open holes, gravel pits, and streams in order to construct a water-table map of the gravel aquifer (fig. 29). The water table slopes south at an average of about 8 feet per mile. At the time the wells were measured, the Flathead and Whitefish Rivers were at relatively high stages and were recharging the aquifer, as indicated by a downstream bending of the water-table contours near the rivers. When the river stage is relatively low, the flow may reverse direction so that the ground water discharges to the rivers.

The water-table map of the Evergreen area (fig. 30) shows that the Flathead and Whitefish Rivers are recharging the aquifer except in the northwesternmost part, where the 2,913-foot contour swings upstream instead of downstream. Profiles of the water table in the Evergreen area were determined from water-level measurements in four wells in a line perpendicular to the Flathead and Whitefish Rivers. The profiles (fig. 31) show that water was moving from the direction of the Flathead River during January 1964 while the aquifer was discharging into the Whitefish River. During June 1965, both rivers were at high stages and water was moving toward the center of the aquifer, as indicated by the dish-shaped profiles.

The rate of ground-water flow through the gravel aquifer (fig. 29) was determined from the slope (12 feet per mile) of the water table across a 3-mile section along the 2,950-foot contour and the coefficient of transmissibility (1,300,000 gpd/ft at well B29-21-11b). These values were substituted in Darcy's equation:

$Q = TIL = 1{,}300{,}000 \times 12 \times 3 = 47{,}000{,}000$ gpd or 73 cfs. The estimated velocity of ground-water flow through the section is 50 feet per day.

The gravel aquifer is recharged by applied irrigation water, by precipitation, and by infiltration from the Flathead and Whitefish Rivers during relatively high river stages. Discharge is by evapotranspiration, pumping, and seepage into the Flathead and Whitefish Rivers during relatively low river stages. The rivers' effect on the water levels in the aquifer is shown by variations in river flow and corresponding changes in ground-water level (fig. 32). Well B29-21-34cc1 in the Evergreen area is about 10 miles downstream from the gaging station on the Flathead River at Columbia Falls; the well is about 5,700 feet from the river. The elapsed time between a peak in discharge and river stage at Columbia Falls and the corresponding water-level peak in the well is about two days. During January 1964, the water level in well B29-21-34cc1 and the flow of the river were nearly constant; in February the flow of the river decreased from about 11,000 cfs to about 3,000 cfs and the water level in the well lowered 0.40 foot. As the river flow decreased, water from the aquifer began to drain into the river causing a lowering of head in the aquifer. On April 17 the flow of the river was 11,500 cfs after a flow of about 1,500 cfs on April 1. The water level in the well rose 0.22 foot on April 19 because the river stage was higher than water levels in the aquifer. During May, the river gradually increased in flow and reached a peak of 42,500 cfs on May 21. The water level in the well gradually rose 1.44 feet to a peak on May 23. A decline in water level in the well of about 0.22 foot between May 24 and 28 corresponds to the decrease in river flow between May 22 and 26.

Water pumped from the gravel aquifer for irrigation is estimated to be about 600 million gallons during June, July, and August. Water pumped for domestic use is about 230,000 gpd or about 80 million gallons per year. Of this 80 million gallons, about 150,000 gpd or 50 million gallons per year is pumped from wells in the Evergreen area. The daily pumpage from two industrial wells at the Anaconda Aluminum Co. plant is estimated to be 6 million gallons per day or 2.2 billion gallons per year. The municipal supply for Kalispell is primarily from a spring issuing from the gravel aquifer. The water pumped from this spring in 1964 was 548 million gallons; in 1965, it was 542 million gallons.

The volume of water in the gravel aquifer north of Kalispell was estimated by multiplication of the average saturated thickness of the aquifer by its area and by its storage coefficient. In June 1965, the average saturated thickness was 25 feet, the area 34,800 acres, and the storage coefficient is assumed to be about 0.20. Therefore, 170,000 acre-ft or about 55 billion gallons was in storage in June.

For the years 1964 through 1966, average fluctuation in the gravel aquifer from the high in May or June to the low in February or March was about 3 feet. This range indicates that the volume of water in temporary storage changes by about 21,000 acre-ft.

WATER-BEARING PROPERTIES

Well data and aquifer tests show that the gravel aquifer is very permeable and is capable of yielding more than 1,500 gpm to wells. Most of the large irrigation and industrial wells in the valley obtain

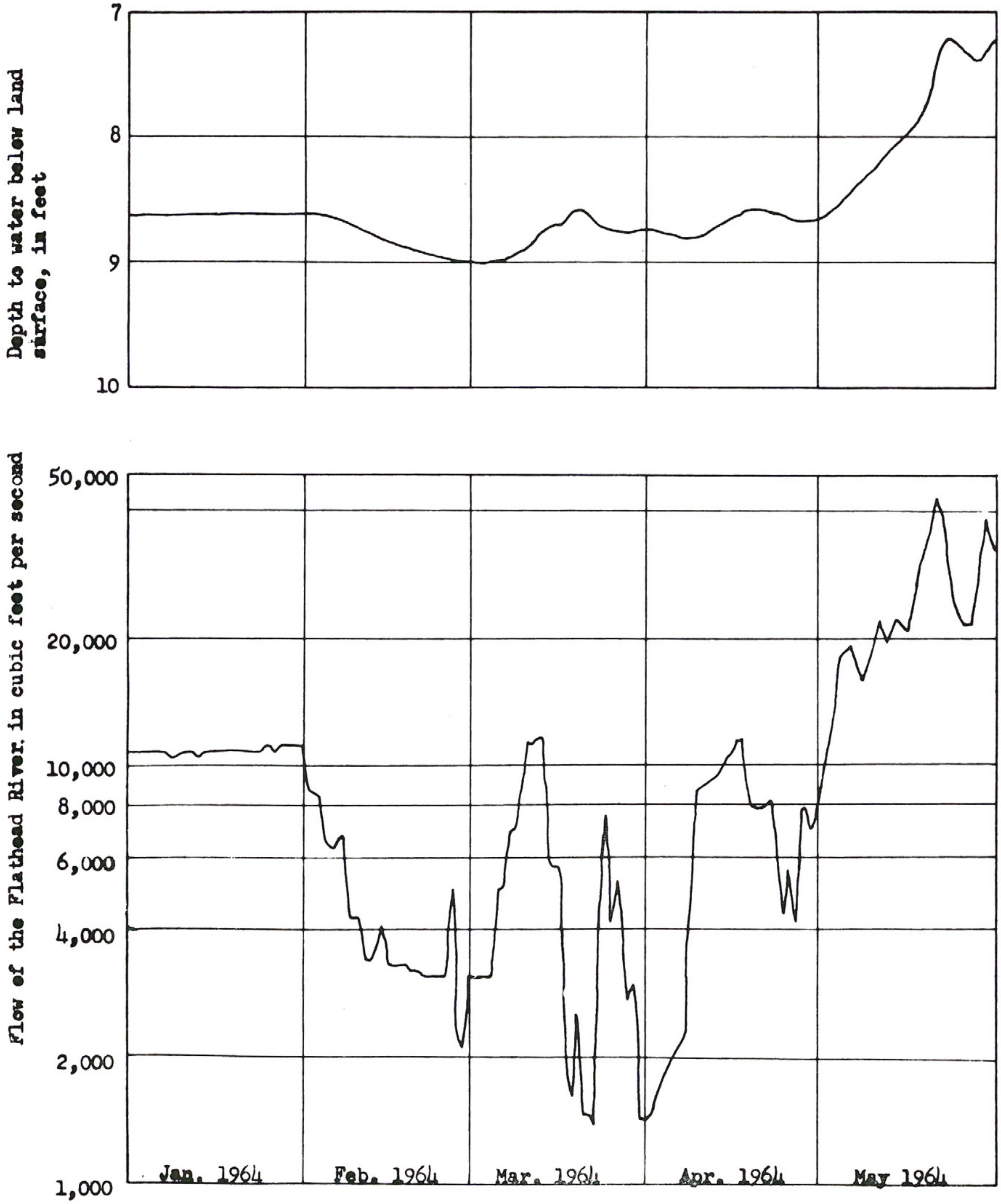


Figure 32.—Fluctuations of water level in well B29-21-34ccl caused by changes in flow of the Flathead River.

water from this aquifer. Two wells (B30-20-3dd1 and B30-20-3dd2) owned by the Anaconda Aluminum Co. indicate the water-bearing potential of the aquifer. One well was pumped at 1,400 gpm for 96 hours and was drawn down 2.2 feet; the specific capacity is 640 gpm/ft. The other well was pumped at 1,000 gpm for 96 hours and was drawn down 1.9 feet; the specific capacity is 530 gpm/ft.

The transmissibility and permeability were determined at three sites. Well B29-21-11b (fig. 29) was pumped at 250 gpm for three hours and the drawdown was 0.68 foot, indicating a specific capacity of 370 gpm/ft. From the Theis recovery formula the transmissibility was calculated to be 1,300,000 gpd/ft. The average coefficient of permeability, which is the quotient obtained by dividing the coefficient of transmissibility by the saturated thickness, is 87,000 gpd/sq ft. An arrigation well at the west quarter corner of sec. 22, T. 29 N., R. 21 W., was pumped for three hours at 480 gpm. The drawdown was 1.86 feet, which indicates a specific capacity of about 260 gpm/ft. The transmissibility is 800,000 gpd/ft, and the average permeability is 30,000 gpd/sq ft. Well B29-21-4db was pumped at 250 gpm for 30 minutes and the drawdown was only 0.83 foot, indicating a specific capacity of about 300 gpm/ft. The transmissibility is 1,300,000 gpd/ft, and the average permeability is 100,000 gpd/sq ft.

The aquifer tests indicate an average transmissibility of 1,100,000 gpd/ft and an average permeability of 75,000 gpd/sq ft. The tests also indicate that the water-bearing characteristics of the aquifer are almost uniform from place to place.

CHEMICAL QUALITY OF WATER

The hardness of water from seven wells tapping the gravel aquifer averaged 173 ppm and ranged from 135 to 204 ppm. The range in hardness probably reflects the mixing of relatively soft river water and ground water. Water from well B29-21-2aa, which is about 2,600 feet from the Flathead River, had a hardness of only 135 ppm. Water from well B29-21-34cc2, which is about 5,700 feet from the Flathead River, had a hardness of 200 ppm. Water from the gravel aquifer is of the calcium bicarbonate type and generally contains less iron and dissolved solids than water from other aquifers in the valley. Six of seven samples contained no iron in solution at the time of analysis, and the dissolved solids ranged from 132 to 210 ppm and averaged 179 ppm.

Some well owners in the Evergreen area noticed a medicinal taste and brown tinge to their water. Water samples sent to the State Board of Health were found to contain phenols. An investigation by personnel of the State Board of Health revealed that waste glue from an industrial plant was being placed in a pit dug below the water table, and phenol compounds dissolved from the glue were assumed to have migrated southward in the ground water. Dilution of the ground water by water from the Flathead River helped to keep the problem from becoming more widespread.

THE SAND AQUIFER

HYDROLOGIC OPERATION

The water-table contour map (pl. 5) was based on water levels measured between June 5 and 10, 1965, in wells, sloughs, and the Flathead River. The stage of the Flathead River was at an altitude of 2,891.6 feet at Foys Bend and 2,890.9 feet about 13 river miles downstream. The mean stage of Flathead Lake on June 7 was 2,890.8 feet. Major fluctuations of lake stage and of river stage downstream from about Foys Bend are nearly the same because of backwater from the lake. From October 1964 through April 1965, the lake stage steadily declined 7 feet from an altitude of 2,893 to 2,886. From May 1, 1965, until June 5, 1965, when the wells were measured, the lake stage rose steadily about 2 feet.

The configuration of the water table in the sand aquifer shows the effects of changes in the lake and river stages. Recharge by precipitation and discharge by evapotranspiration tend to modify the water table configuration, but the major features probably are caused by the lake and river. The mounds in the water table east of Church Slough and northeast Somers may be remnants of the high water table when the lake stage was at an altitude of 2,893 feet the previous October. As shown by the arrows in Plate 5, water is draining from the mounds to the river and lake, and to the trough in the water table between Foys Bend and Fennon Slough. The trough may be a remnant of the low water table when the lake stage was at an altitude of 2,886 feet the previous April. Water is draining into the trough from the two mounds, from the river near Foys Bend and near Fennon Slough, and from Flathead Lake about 2 miles east of Somers.

The yearly cycle of recharge and discharge in the sand aquifer is controlled mainly by the lake and river stages. The time of highest water level in the aquifer and the magnitude of the yearly rise vary throughout the aquifer, but the yearly fluctuations in water level are sinusoidal because the yearly fluctuation of lake stage is sinusoidal (fig. 33). In general, the lake and river recharge the aquifer when the lake stage is rising (April through June) and when

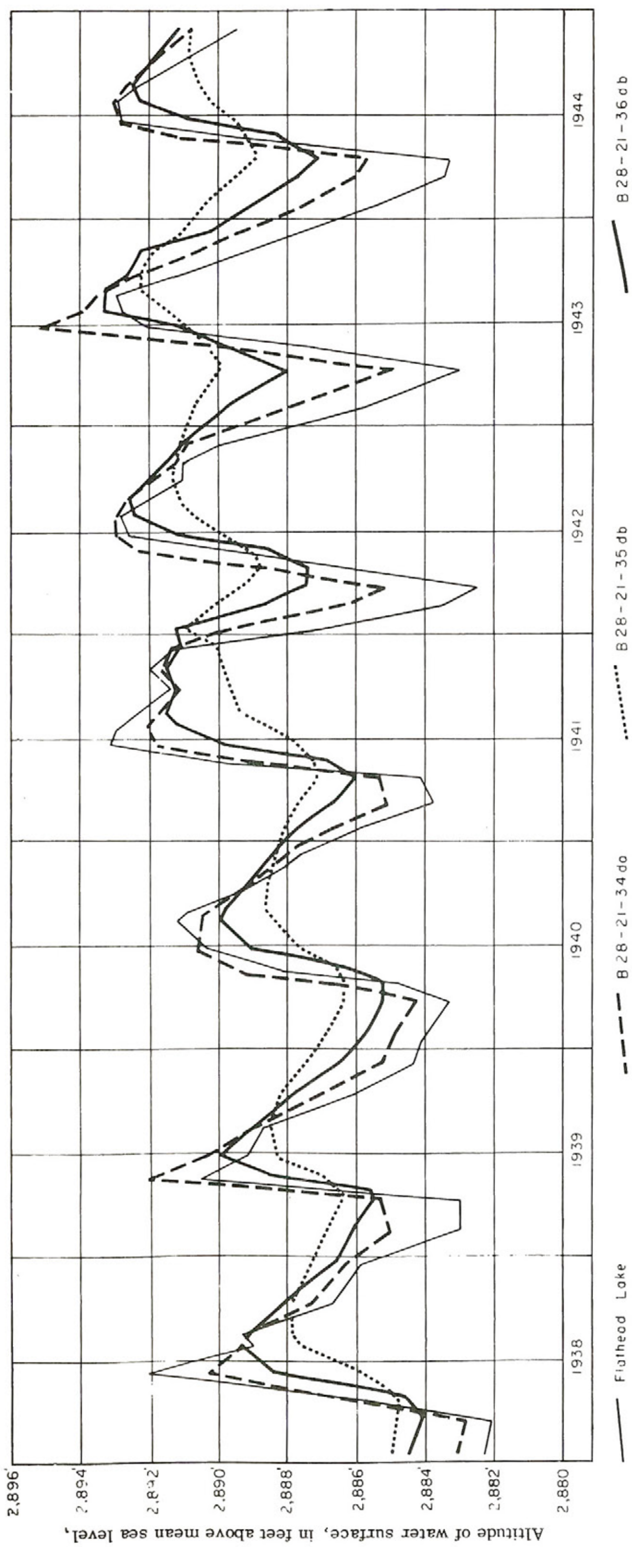


Figure 33.—Hydrographs of water-level fluctuations in Flathead Lake and in three wells tapping the sand aquifer.

the lake stage is highest (July and August). At most places the rate of rise of water level in the aquifer is less than in the lake, and the peak lags behind the lake peak. Water levels in the aquifer close to the river or lake rise at about the same rate as the lake, and the lag is measured only in days. The lake stage begins to fall in August, and water in the aquifer begins to discharge to the lake and river.

The magnitude of water-level fluctuations in the aquifer is progressively less with greater distance from the river and lake. The relation of lake- and river-stage change to water-level changes at various places in the aquifer is shown in Table 6, which is summarized from Cady (1941) and McDonald (1946).

Table 6.—Average annual water-level changes (1938-44) in the sand aquifer and in Flathead Lake

Well number	Distance from Flathead River (feet)	Average annual water-level change (feet)
B28-21-34da	250	7.44
B28-21-36db	800	4.97
B28-21-35db	1,500	2.36
B27-20- 8da	3,000	.35
Flathead Lake	--------	11.35

Recharge from precipitation causes water-level rises in the aquifer, especially noticeable at points a mile or more from the lake or river. As the soil is relatively permeable, a large percentage of the precipitation infiltrates the aquifer. A good example of recharge from precipitation is the 4.6-foot rise of the water level in well B27-20-8ba between March 9 and 10, 1966, in response to melting snow (fig. 34).

Evapotranspiration directly from the aquifer is mainly from sloughs or marshy areas. The 1,700 acres of swamps and sloughs south of the river discharge about 3,400 acre-ft of water per year by evapotranspiration if the yearly rate is assumed to be 2 feet of water per acre. Soil moisture that is evaporated or transpired is replenished by precipitation.

Pumpage from domestic wells is about 40,000 gpd. No ground water is pumped from the sand aquifer for irrigation. The town of Somers is supplied from Flathead Lake.

The total volume of ground-water storage in the sand aquifer during June 1965 was about 50,000 acre-ft. This figure is based on an area of 18,000 acres, as assumed storage coefficient of 0.10, and an average saturated thickness of 28 feet as calculated from well logs. The average annual change in water levels throughout the aquifer for the years 1964 through 1966 was about 2 feet, which indicates a yearly average change in temporary storage of 4,000 acre-ft.

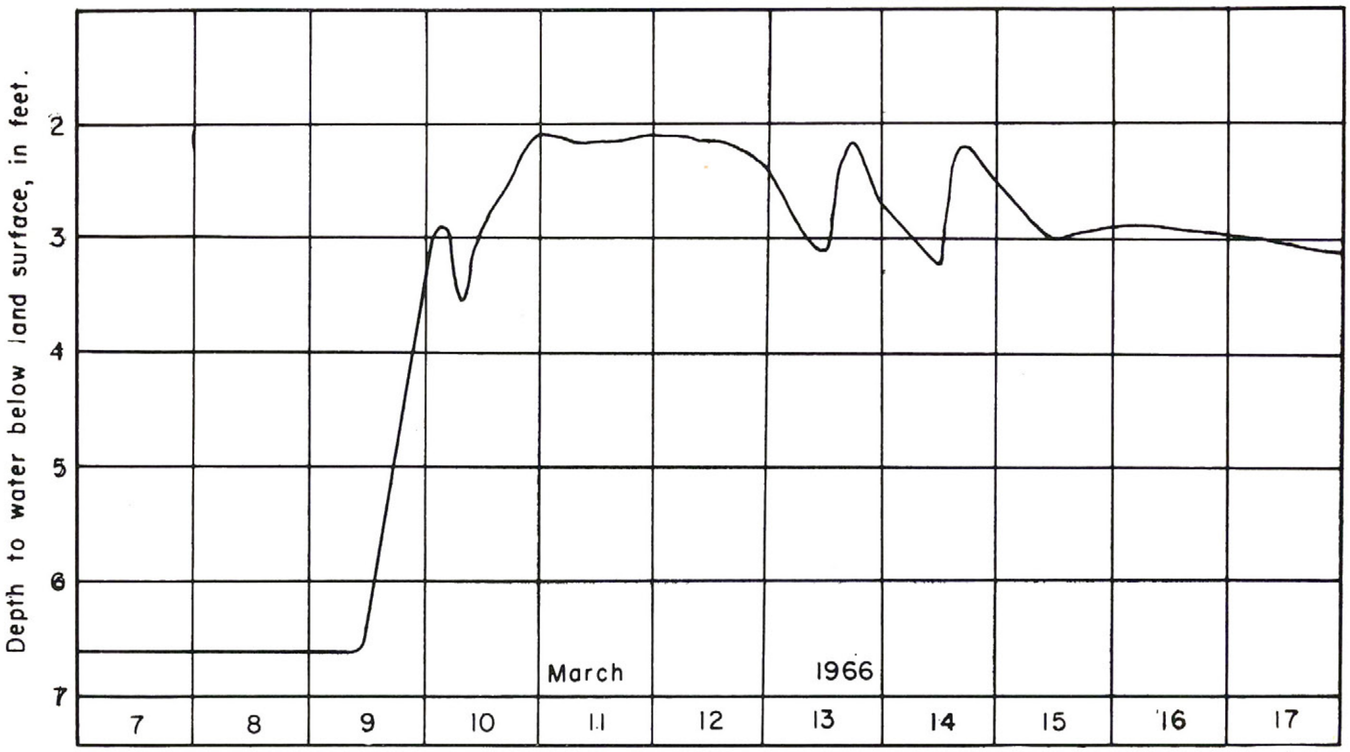


Figure 34.—Hydrograph of well B27-20-8ba showing recharge from snowmelt.

A long-term gain in ground-water storage during the years 1938-49 is due in part to maintenance of the water level of Flathead Lake at relatively high levels for about three months each year. The level is regulated by a dam constructed in 1938 across the outlet. The water levels in wells rose an average of 4.4 feet in an area of about 18,000 acres. The average net water-level rise multiplied by the area and by an assumed storage coefficient of 0.1 indicates that the total gain in storage between 1938 and 1949 was 8,000 acre-ft.

WATER-BEARING PROPERTIES

The ratio (T/S) of the coefficients of transmissibility (T) and storage (S) of the sand aquifer was calculated from fluctuations of the river and lake stages and corresponding water-level fluctuations in wells. The method, which was described by Ferris (1950), assumes, together with the usual simplifying assumptions, that sinusoidal fluctuations of a river or lake cause sinusoidal fluctuations of the water level in an aquifer that is hydraulically connected with the river or lake. Because of artificial regulation, the stages of Flathead Lake and of the river for about 13 miles above its mouth may be approximated as a sinusoidal curve having a period of one year. River-induced fluctuations of the water level in the aquifer may also be approximated as a sinusoidal curve that lags slightly behind the lake and river sinusoid and is of smaller amplitude (fig. 33). By comparison of the average range of ground-water fluctuations at various distances from the river with the average range of river stage (table 6), the ratio T/S was calculated to be 75,000 gpd/ft. If S is assumed to be 0.10, then T must equal 7,500 gpd/ft.

The transmissibility calculated by this method is probably correct within an order of magnitude and is reasonable for an aquifer composed of sand and silt. Transmissibility calculated from pump-test data would probably be smaller because of head losses near the pumped well and partial penetration of wells. Transmissibility calculated from cyclic fluctuations depends upon the assumed value of storage coefficient and is based on the assumptions of complete hydraulic connection between the aquifer and lake or river and of one-dimensional flow. These assumptions and dependency are not involved in calculation of the transmissibility from pump-test data.

CHEMICAL QUALITY

Excessive iron content is characteristic of the water in the sand aquifer. At the time of analysis, the average iron in solution in water from four wells was 2.36 ppm, the greatest in the valley. The maximum iron content was 14.12 ppm in well B27-20-20ab (table 4). Reddish precipitated iron can be seen along marshy areas and along the north shore of Flathead Lake where ground water discharges during low lake stages. The sides of many white houses are stained yellowish brown as a result of lawn sprinkling, and numerous bands of iron-stained rocks are exposed in excavations.

The water is of the calcium bicarbonate type, and the dissolved solids of four samples range from 288 to 402 ppm, averaging 342 ppm. The relatively large dissolved-solids content (1,480 ppm) of water from well B27-20-8aa is probably attributable to concentration of the dissolved solids by evapotranspiration. Soil near this well has been classified as of the saline-alkaline type (Williams and Jackson, 1960, p. 49-50, sheet 28).

SUMMARY

Areas that contain large quantities of good ground water have been delineated. The flood-plain gravel aquifer has the greatest potential for additional large increases in pumpage. This aquifer stores about 170,000 acre-ft of water and annually releases about 21,000 acre-ft of water to streams. About 1,100 acre-ft of water is pumped from the aquifer each year, and wells yield as much as 3,000 gpm. The aquifer is subject to pollution from disposal of industrial wastes and from sewage.

Development of water supplies from the deep artesian aquifer is accelerating, because the aquifer is a dependable source of ground water and is little affected by drought. Properly constructed and developed wells will yield as much as 1,500 gpm. Many more large-capacity wells can be developed without permanently lowering the piezometric surface. As the water contains some iron, treatment may be required.

The amount of water pumped from the dune sand is diminishing because wells may go dry in some years, although storage amounts to about 180,000 acre-ft. Low permeability results in small yields to wells, and the water may contain excessive nitrate.

The outwash sand and gravel aquifer northwest of Kalispell is of small areal extent. Even though this aquifer is relatively permeable, it has small irrigation potential compared to the flood-plain gravel and deep artesian aquifers.

Water from the flood-plain sand aquifer north of Flathead Lake contain much iron. Because of the iron content and the relatively low permeability of this aquifer, many wells have been drilled to the more favorable deep artesian aquifer below it.

The basement rocks supply water to wells in areas where there are no other aquifers. Wells yield a few gallons per minute from joints and fractures, but large drawdowns are required for small quantities of water.

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