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EARLY CONTRIBUTIONS TO THE GROUNDWATER HYDROLOGY OF ALBERTA

by

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EARLY CONTRIBUTIONS TO THE GROUNDWATER HYDROLOGY OF ALBERTA

ABSTRACT

Ten short papers are presented describing selected investigations of interest carried out by the Groundwater Division of the Research Council during its early years. Included are areal investigations and a microbasin study, reports on bedrock topography and bedrock channel distribution, results of an artificial recharge and of an induced infiltration project, and a discussion of geophysical prospecting for groundwater. Appendices list hydrologic units and equations in common use.

PURPOSE AND SCOPE OF GROUNDWATER INVESTIGATIONS IN ALBERTA

by
R. N. Farvolden

The Groundwater Division of the Research Council of Allberta was formally organized by the Government of Alberta in 1955. In its initial stage, an experienced groundwater geologist was appointed on a temporary basis to conduct a general survey of the water conditions of the province (Foster and Farvolden, 1958). The results of this general survey were then used to make recommendations to the Research Council Technical Advisory Committee concerning groundwater work required in the province. Personnel were added over a two-year period and the Groundwater Division has now been operating at full strength for five years. This report deals with some of the early studies of the Groundwater Division and discusses the approach used in trying to be of immediate service to the public while at the same time laying the foundations necessary to meet future problems.

There is always difficulty in reconciling the work of the scientific data-gathering agency with that of the government service agency. In the case of a groundwater survey, this difficulty is simplified somewhat because the practical and research problems are often the same or closely related. The groundwater program, in its initial stages, was designed to capitalize on this fact. All of the major projects reported in this bulletin have been undertaken with the intention of solving particular groundwater supply problems and at the same time continuing the program of mapping and understanding the groundwater resources of the province. Thus they conform to the basic aim of the Groundwater Division—the determination of the nature, extent and proper development of a valuable natural resource.

When designing a program for the groundwater survey, it is only natural that one should think of the many water supply problems that are before us today and of their possible solution. It is important to remember that the problems of today are mostly those of economics: that for most water supply problems encountered there are several alternative sources of water. The solution lies in assembling sufficient data so that a decision can be made as to which source is the most suitable. There can be no doubt that this work is worthwhile if it aids in the efficient development of private, municipal, or industrial water supplies. However, a wholly different problem may arise in the future, when the available supply of water may be insufficient to maintain modern standards. At that time it will be essential that the water resources of Alberta be outlined in such detail that complete development is possible. The work of the Groundwater Division will then be judged solely on the basis of its contribution to this complete development. Therefore, any service work or line of investigation that does not contribute to an understanding of the occurrence, movement, and development of groundwater is to be avoided, and any opportunity to gain knowledge is to be exploited.

For this reason some enquiries are investigated much more thoroughly than others. In one case, some special feature of a groundwater problem may mean that the solution of the problem will serve to increase the general knowledge of groundwater within the province. In another superficially similar case, no such feature will exist. In the latter case, there is no justification for extensive investigation by the Groundwater Division, and the response to such an inquiry will be based on information already available in the files of the Division and on the staff's general knowledge of the area in question. The Groundwater Division has a duty to the public, but this duty will best be fulfilled if the emphasis in the overall groundwater program remains as much as possible on the more general aspects, being shifted to the more particular only when certain unusual or special features warrant such a move.

The professional staff of the Groundwater Division presently consists of six geologists and two geophysicists. Each geologist has been assigned a specific area of the province, thus enabling him to become familiar with, and responsible for, the problems peculiar to his area. Most of the geologists also have been assigned a research or groundwater survey project. Although groundwater inquiries from within his assigned area may interfere to a greater or lesser extent with a geologist's research project, it must be stressed that a great deal of valuable information is obtained as a result of these requests.

The two geophysicists are responsible for the conduct of geophysical surveys required to supplement available geologic data. They and the one geologist who carries out a large number of the hydrologic well tests work in all inhabited areas of the province.

It is almost impossible to set a value on either the Division's field work or on the resource of groundwater, so that the economics of the survey will never be known for certain. However, by using an extremely rough estimate of the savings realized in successes such as the Driedmeat Lake recharge project and the Police Point (Medicine Hat) industrial water-supply, both of which are reported in this bulletin, it can be said with conviction that now, or in the near future, this Division will pay its way in terms of savings to the public.

OCCURRENCE AND MOVEMENT OF GROUNDWATER IN ALBERTA

by
W. A. Meneley

Introduction

All potable groundwater in Alberta is obtained from clastic sediments, ranging in age from late Cretaceous to Recent. The major aquifers may be subdivided into two broad categories: bedrock aquifers and surficial aquifers. Bedrock aquifers typically include fine-grained sandstone, sandy shale, shale and coal strata, which have intergranular and fracture permeability ranging from less than 1 gpd/ft^2 to about 50 gpd/ft^2 (gallons per day per square foot). These sediments, which are of late Cretaceous to Oligocene age are unconsolidated to completely indurated. The surficial aquifers, which are largely unconsolidated, include alluvial sand and gravel deposits—probably of late Tertiary age—Pleistocene alluvial sand and gravel deposits, outwash and stream-trench deposits, underlying or intercalated with till. The permeability of these aquifers ranges approximately from 1 gpd/ft^2 to 2,500 gpd/ft^2.

It is estimated that, although more than 75 per cent of the water wells in the province are completed in bedrock aquifers, more than half of the total groundwater production is obtained from wells completed in the surficial aquifers. The average sustained yields of most wells completed in bedrock aquifers are estimated to be less than 5 gpm (gallons per minute). Only relatively few bedrock wells, such as those completed in the Ribstone Creek and Birch Lake Sandstones, and in the Wapiti, Edmonton and Paskapoo Formations, are known to be capable of producing more than 60 gpm.

The most prolific surficial aquifers are the alluvial deposits of Tertiary, Pleistocene, and Recent rivers. The Tertiary and Pleistocene stream channels are completely or partially filled with younger glacial deposits and, as their precise ages cannot yet be determined, they are collectively termed *buried channels*. Where highly permeable surficial aquifers occur below the water level of an adjacent, perennial body of water, large quantities of water may be obtained by induced infiltration (Meyboom, 1963). If favorable geologic conditions occur, it may be possible to recharge surficial aquifers artificially (Farvolden, 1963c). Where water cannot be added to the aquifer by induced infiltration or by artificial recharge, the sustained yield of the aquifer cannot exceed the amount of natural recharge intercepted by the aquifer.

The maximum potential recharge in Alberta (Meyboom, 1961a; Farvolden, personal communication) is currently estimated to be approximately three per cent of the total precipitation, or about 0.04 feet per year. This means that any aquifer in Alberta dependent upon natural recharge may be subject to depletion if the mean annual withdrawal from the aquifer exceeds about seven million gallons per square mile. The area over which depletion may occur will extend away from the pumping centers if permeability is large, or will be restricted to their immediate vicinities if permeability is small. This phenomenon is related to the shapes of the cones of depression in the two cases. The cone of depression for a pumping well completed in a relatively impermeable aquifer will be steeper than that for a well pumping at the same rate in a more permeable aquifer, and the drawdown in the less permeable aquifer will be greater close to the well but less at great distances from it. Most bedrock aquifers are not sufficiently permeable to permit withdrawal of this limiting quantity of water per square mile; however, if there is an inordinately large number of wells, local depletion may occur. For example, in some subdivisions in the Calgary area, local depletion of aquifers in the Paskapoo Formation is evident in the immediate vicinity of the affected subdivisions, although there is no depletion of the Paskapoo as a whole (Meyboom, 1961b).

The amount of groundwater available for utilization is limited, among other things, by the climatologic environment of the prairies. Nevertheless, groundwater is now, and will continue to be, a vital source of water for the people of Alberta. It is essential that the regime of groundwater be fully understood in order that this resource may be developed to its maximum.

Other contributions to this bulletin discuss some of the problems that have been encountered while evaluating the groundwater resources of various areas in the province. The purpose of the following contribution is to give an outline of the scientific principles that underlie these hydrogeologic investigations.

Hydrology

Description of the Flow System

Hubbert (1940) described the mechanics of gravitational flow in a homogeneous isotropic medium. His concept of the flow system provides a framework within which the movement of groundwater in Alberta may be discussed.

Below the free-water surface, water moves under the influence of gravity from a region of higher fluid-potential toward regions of lower fluid-potential following a path that in the case of an isotropic medium is everywhere perpendicular to the equipotential surfaces. The fluid potential is the algebraic sum of the hydrostatic pressure potential and the gravity potential, and it may be determined from measured static or nonpumping water levels in wells.

It can be shown (*op. cit.*, p. 788-793) that flow will occur from a region having a lower hydrostatic pressure toward a region having a higher hydrostatic pressure only if the latter region has a lower fluid potential. In Alberta, maximum fluid potentials are found under topographically high areas (Jones, 1959; Farvolden, 1961b). It follows, therefore, that groundwater will flow away from topographically high areas toward topographically lower areas. If it is considered that the flow system is in dynamic equilibrium, then water must be continuously added in the upland areas to replace water that is naturally discharged in the lowland areas. Because the only available source of water in an upland area is precipitation, groundwater flow in Alberta is dependent upon precipitation for its sustenance.

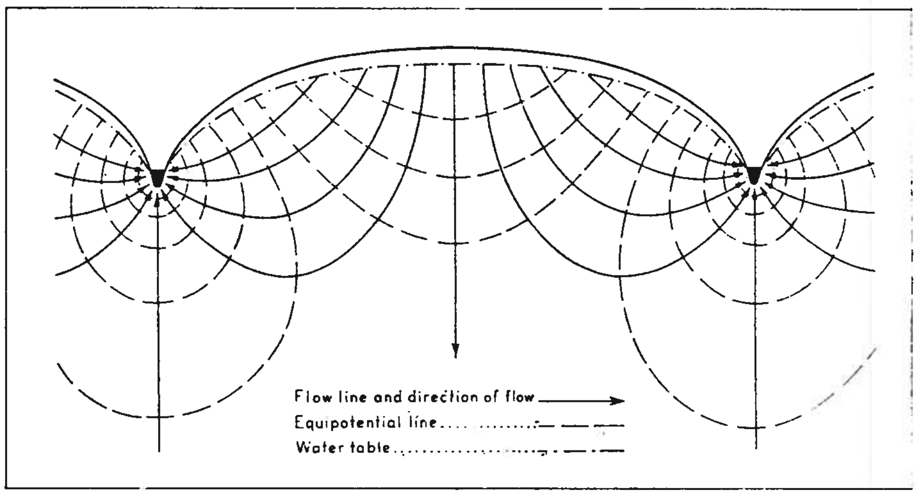

Figure 1. Potential distribution and flow pattern in a homogeneous, isotropic medium.

Figure 1 shows a cross section of the potential distribution and flow pattern in a homogeneous isotropic medium, constructed according to the assumptions: (a) that recharge takes place everywhere along the surface and (b) that discharge takes place only in the rivers and creeks or in their immediate vicinity. Here, water enters the flow system in the upland area, moves downward and radially outward, and ultimately discharges back to the ground surface in the lowland areas. In such a flow system, a series of piezometers, set at progressively greater depths in the upland area, will record a progressively decreasing fluid potential as the depth increases, while piezometers located in the lowland area will record increasing fluid potential as the depth increases. The potential along any streamline must continually decrease in the direction of movement, since flow requires a finite expenditure of energy to overcome frictional resistance. In a homogeneous isotropic medium, the maximum depth of groundwater circulation cannot be defined; however, the flow density or flux (the volume of flow per unit area of cross section per unit time) decreases rapidly with increasing depth (*op. cit.*, p. 930). It should be noted that the vertical scale in figure 1 is highly exaggerated. If vertical and horizontal scales were equal, the topographic surface would closely approximate a horizontal plane.

The depicted potential distribution may be applied equally well to the potential distribution below a single hummock in hummocky moraine terrain, to a drainage basin, or to the gravitational flow system of the entire Western Canada sedimentary basin.

In Alberta, the medium through which groundwater moves by gravitational flow includes a wide range of materials of widely different physical properties. The plains area of Alberta lies on the eastern flank of the Alberta syncline and the regional dip is westward, rarely exceeding 100 feet per mile. On the plains, therefore, the medium may be considered to be horizontally layered. Within this medium some strata will have sufficient permeability to yield water to a well and thus may be considered as aquifers in the sense defined by Meinzer (1923, p. 52). Over most of Alberta, glacial deposits overlie a sequence of sandstone and shale strata of late Cretaceous and Tertiary ages. The thickness and permeability of the glacial deposits are variable; highly permeable materials, however, constitute only a small part of their total volume. Similarly, almost all the sandstone strata are highly lenticular and are separated by sandy shale and shale.

Although permeability appears to be principally intergranular in nature, there are some reasons for inferring that fracture permeability is a factor in groundwater flow in the Stettler area. Fractures were observed in core samples from a test hole in the area and estimated permeabilities from pumping tests were too high to be typical of a material which is best described as a silty shale. The minimum value that can be assigned to the permeability is about 30 gpd/ft^2. Abnormally high circulation losses during drilling also attest to the possibility of fracture permeability.

In its gross aspect, the medium through which flow takes place in Alberta is relatively homogeneous and composed essentially of materials having very low permeability. It is probable that the ratio of the overall horizontal permeability (P_h) to the overall vertical permeability (P_v) is the same, or nearly the same, as that for the shale that constitutes most of the medium. The medium will therefore be anisotropic to some degree as yet undetermined, with respect to permeability. The effect of anisotropy (Hubbert, 1940, p. 902) is that the streamlines and the potential surfaces are no longer orthogonal. The magnitude of the angle of intersection will depend upon the direction of flow relative to the permeability axes of the medium.

Anisotropy of the type found in Alberta will tend to deflect the streamlines into the horizontal plane. This will tend to accentuate the decrease in flow density with increasing depth and to diminish the depth of effective groundwater circulation.

Consideration can now be made of the potential distribution in a typical segment of a flow system in Alberta, consisting of an entire drainage basin. Water will enter the flow system in the recharge area which, in accord with the original assumptions, will comprise almost the entire interstream area. Water enters the flow system at the free-water surface and initially moves downward at some angle greater than the slope of that surface (*op. cit.*, p. 930). However, because the medium is anisotropic, the streamlines will be distorted toward the horizontal. The free-water surface conforms to the surface topography and although the direction of groundwater movement in the recharge area is generally downward, some water will be discharged in local topographic depressions, creating minor discharge areas within the interstream recharge area. The effect is accentuated by the anisotropy of the medium. Thus, minor flow systems will be found in the vicinity of local topographic highs, and a portion of the groundwater will be discharged back to the ground surface to replenish sloughs or to be immediately evaporated. It should be emphasized that there will be no recharge to the regional flow system from lakes and sloughs in upland areas until the free-water surface adjacent to the surface water declines to an elevation lower than the lake water elevation.

Groundwater not discharged locally will continue to move downward and laterally toward the dicharge area. The flux is proportional to the amount of recharge, the permeability, and the available potential difference which is controlled principally by the topographic relief in the drainage basin. It is not possible at this time to define any depth below which there is no significant groundwater circulation; however, over most of the plains area of Alberta, water encountered below a depth of 500 feet rarely contains less than 2,000 ppm (parts per million) total solids. It is considered that the rate of groundwater circulation below this depth is very low. Within the recharge area, the rate of decrease of the fluid potential with increasing depth should rapidly lessen and should tend to approach zero at some finite depth.

The direction of groundwater movement in the discharge area will be upward and the fluid potential should increase with increasing depth, but at a rapidly decreasing rate which should tend to approach zero at some finite depth. Groundwater discharged to streams constitutes the baseflow component of stream discharge.

Piezometric Surface

A piezometric surface defines the fluid potential distribution on an arbitrary two-dimensional surface within a flow medium. It is usually constructed by contouring the elevations of the static water levels in wells completed in a common aquifer. Where the aquifer dips at less than three degrees the gradient on this piezometric surface is essentially equal to the potential gradient that causes tangential flow parallel to the bounding surfaces of the aquifer (Hubbert, 1940, p. 910). This, however, does not imply that there is no transverse flow either into or out of the aquifer. On the contrary, if there is flow in the aquifer, the water must both enter and leave the aquifer via the so-called "confining beds" (*op. cit.*, p. 911).

Over most of Alberta, no extensive continuous aquifers exist because of the general nature of the lithology. Nevertheless there is a continuous flow within this medium, and the observed fluid potential in a recharge area will depend to a great extent upon the depth of the well (Farvolden, 1961b; Meyboom, 1961b). In such a medium it is desirable to select, if possible, some arbitrary datum plane upon which to determine the piezometric surface. In practice this requirement cannot be met and instead only those wells within a limited elevation range are considered in constructing the piezometric surface map.

The piezometric surface will usually conform to the major features of the topographic surface and the latter will, therefore, indicate the general direction of flow in the medium.

Hydrologic Characteristics of Typical Alberta Aquifers

The practical study of groundwater is concentrated on the examination of the distribution and hydrologic properties of aquifers, that is, of those segments of the medium through which flow takes place that are capable of yielding water in economic quantities. Classically, there are two types of aquifers, unconfined and confined. An unconfined or water-table aquifer is one in which the upper bounding-surface of the contained water is the free-water surface, along which the pressure is atmospheric. A confined or artesian aquifer is one in which the upper bounding-surface of the contained water is the boundary between the aquifer and a much less permeable lithologic unit. The pressure along the upper bounding-surface in this case is greater than atmospheric. In Alberta typical confined aquifers are the Milk River and Bulwark Sandstones of Cretaceous age.

Flow of water into a well in an unconfined aquifer is characterized by both horizontal and vertical components, whereas flow into a well that completely penetrates a confined aquifer is distinguished by horizontal components only, provided that the permeability of the aquifer is very much greater than the permeabilities of the confining beds. If this does not hold true, some vertical flow-components appear as water leaks into the aquifer from the confining beds.

In Alberta the medium through which flow takes place commonly consists of thin discontinuous layers of materials having markedly different permeabilities. The average permeability along any path perpendicular to the layering will be less than the horizontal permeability within the more permeable layers. If a well completed across an appreciable thickness of such a medium is pumped, water will move fairly freely horizontally through the permeable layers into the well but vertical movement will be restricted by the low average vertical permeability. In this sense the thickness of the medium across which the well is completed behaves as a confined aquifer. The boundaries of the upper and lower confining beds are then taken to be planes parallel to the bedding and passing through the upper and lower limits of the producing section of the well.

If a free-water surface is to be observed when drilling a well, it must coincide with a permeable stratum. If the permeability is very small, however, water will drain into the well only very slowly and it may not be apparent that the free-water surface has been encountered. Nevertheless, if the well is left undisturbed for a sufficiently long period, the water level in the well will eventually coincide with the free-water surface. This may be observed, for example, when drilling is recommenced after a well has been undisturbed overnight.

In Alberta almost all aquifers appear to be artesian in that the water level almost always rises above the depth at which the aquifer is encountered. This artesian property may originate either in an artesian aquifer of the classical type or in an aquifer that owes its artesian properties to the layered nature of the strata. The height to which the water will rise above the top of the aquifer depends only on the position of the aquifer in the medium, and is equal to the hydrostatic pressure within the aquifer at the point at which the well intersects it. The available drawdown is defined as the vertical distance from the static water level to the top of the aquifer and it is, therefore, equal to the hydrostatic pressure. The flow system in Alberta, as was previously remarked, is essentially homogeneous in its gross aspect. Locally, however, it is heterogeneous and consists of interspersed permeable deposits in a medium of much smaller permeability. The available drawdown in a well that taps only the relatively impermeable matrix may appear to be anomalously small when it is compared to the available drawdowns for wells tapping nearby permeable deposits. If, however, the well is allowed to remain undisturbed long enough to allow the process of slow drainage into it to be completed, the anomaly will normally be removed.

When a well completed in an artesian aquifer is pumped, a region of pressure decrease called the cone of influence is developed around the well, within the area of which all water is moving toward the well. The water withdrawn from the well is derived from the following sources:

(1) storage within the aquifer,

(2) storage within the confining beds,

(3) recharge to the aquifer under naturally existing potential gradients that is intercepted by the cone of influence, and

(4) recharge to the confining beds, by precipitation or otherwise, over the area of the cone of influence.

When pumping starts, the aquifer will initially behave as an ideal artesian aquifer; essentially all the water produced will be derived from storage within the aquifer. As pumping progresses, however, the aquifer will depart to a greater or lesser extent from its initial behavior, depending upon the relative significance of the above factors.

The long-term behavior of the aquifer will be quite different; depending upon local conditions, one of the following conditions may result after pumping has continued for a long time:

(1) the aquifer may continue to behave as an ideal artesian aquifer from which essentially all water is derived from storage, such as the Milk River Sandstone and the Bulwark Sandstone,

(2) the aquifer may behave as a leaky artesian aquifer, reaching equilibrium at some time when the leakage and the up-gradient contribution are equal to the discharge, such as the Edmonton Formation at Stettler, or

(3) it may behave as an anisotropic unconfined aquifer, so that the overlying confining beds are gradually dewatered, and equilibrium, when it is reached, occurs when the discharge balances the natural recharge.

In any of the above situations, if the physical dimensions of the aquifer limit expansion of the cone of influence of a well or well field to some radius smaller than that required to attain equilibrium, then eventual total depletion of the aquifer will result if pumping is continued for a sufficiently long period of time.

The maximum rate at which a well can produce in a given aquifer depends primarily upon the transmissibility and storage coefficient of the aquifer. The safe yield of an aquifer over a long period of time is primarily dependent upon the physical dimensions of the aquifer and the average natural recharge to the aquifer. The aquifer coefficients can be determined from pumping tests; however, the safe yield of the aquifer can be evaluated only by a detailed study of the hydrologic and geologic environments.

GROUNDWATER GEOLOGY AND HYDROLOGY OF THE ANDREW AREA, ALBERTA

by
E. Gordon Le Breton

Introduction

Purpose and Scope of the Report

A survey carried out during the summer of 1958 was designed to determine the groundwater resources of the Andrew area. The area of investigation includes townships 55 to 57 within ranges 14 to 17*. This area is included on the Edmonton topographic sheet (Canada National Topographic Series Sheet 834, scale 1:250,000) and is outlined in figure 2.

The practical significance of this study was to inform the village of Andrew (population 650) of the prospects of obtaining groundwater for a proposed water works system. At present, the residents obtain water from many shallow wells completed either in bedrock or in glacial drift. There is no evidence that an assured municipal supply is available from the aquifers currently used, for wells in these aquifers seem only able to supply domestic requirements.

Preliminary work on the geology and groundwater resources of the area was followed by geophysical surveys, and drilling was later carried out to test the conclusion and findings of this work.

Description of the Area and Physiography

The climate of the area is continental, semiarid. Figures from the weather station at Ranfurly which is 40 miles southeast of Andrew are as follows: mean temperature for July, 63.2°F; mean temperature for January, 4.2°F; and mean annual temperature, 35.3°F. The average annual precipitation is 16.89 inches.

Apart from the east part of the area, the topography is relatively flat, with small valleys generally less than 50 feet deep incised into the surface. The relief is greatest in the east part of the area, where the land surface ranges in height from 1,850 feet above sea level in the North Saskatchewan River Valley (Tp. 57, R. 14), to over 2,450 feet above sea level on the top of the highest hill (Tp. 56, R. 14), 4 miles south of the river.

Crossing the center of the area from the northwest to the southeast is a broad depression 5 miles wide, which separates two areas of higher relief in the west and southwest, and in the east and northeast (Fig. 3).

* Unless otherwise stated, all locations are west of the Fourth Meridian (110° west longitude).

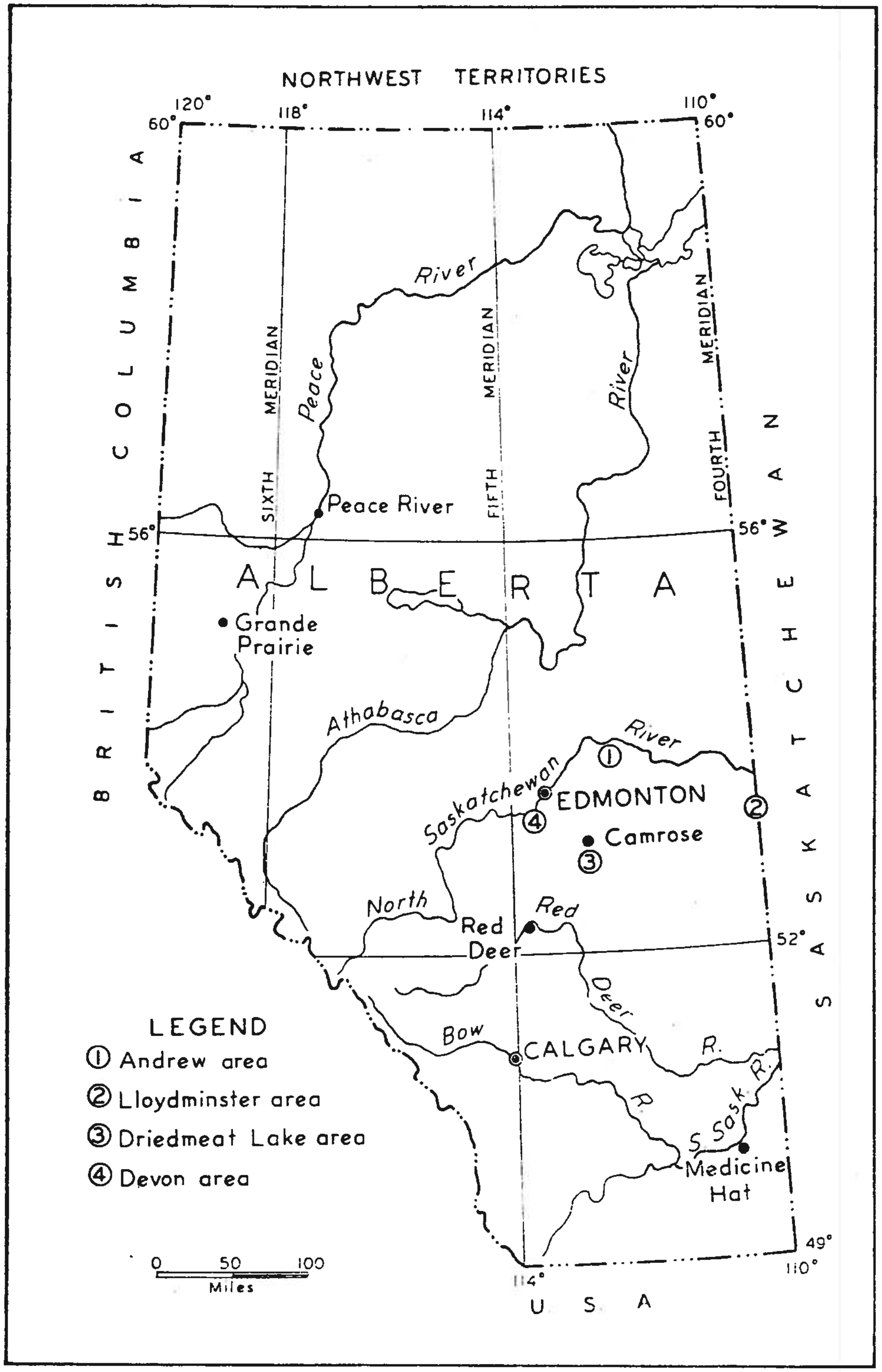

Figure 2. Index map.

Small temporary and permanent streams draining surface runoff from the higher areas flow to the southwest or to the northeast into the broad central depression. Surface water is then discharged away from the area toward the northwest into the North Saskatchewan River (Tp. 58, R. 17) and to the southeast (Tp. 55, R. 14).

Acknowledgments

The author wishes to express his thanks to local drillers and residents for information supplied during the field survey, and to Texaco Exploration Company for seismic shot-hole data.

Groundwater Geology

The geological map (Geol. Surv. Can. Map 505A, Tofield, Alberta) shows the distribution of the Grizzly Bear Formation, the Birch Lake Formation and the Pale and Variegated Beds. Shaw and Harding (1954) considered it impossible to distinguish these strata in subsurface correlation and include them as members of the Belly River Formation. In Shaw and Harding's discussion the status of the Belly River unit varies in different areas. On the southern plains of Alberta, the Belly River has group status, for in this region it is possible to make the distinction between the Foremost and Oldman Formations. In east-central Alberta, in the Vermilion area (Nauss, 1945), the Belly River is a formation which is divided into a number of members of interfingering marine shales and deltaic sandstones. West of a line from township 44, range 9 to township 55, range 14, the westward limit of marine incursion, the Belly River Formation is described as a series of grey, brownish-grey, greenish-grey, argillaceous, bentonitic sandstones, closely interbedded with brownish-grey to grey, carbonaceous shales and siltstones (Shaw and Harding, 1954).

Most of the area is covered by a deposit of ground moraine consisting chiefly of till with some lenses of sand and gravel. The ground moraine ranges in thickness from zero to 100 feet and averages about 40 feet. Where the drift cover has been eroded, bedrock outcrops are found. In the northwest part of the area in township 57, range 17, is a covering of dune sand estimated to have a maximum thickness of 40 feet. The covering of ground moraine is broken only by stream-trench systems and the granular deposits of these channels form one of the two important types of groundwater sources.

The map of the bedrock topography (Fig. 3) is based principally upon data obtained from seismic shot-hole logs, and upon some water-well logs. Shot-hole logs at spacings of one-quarter of a mile were available in road allowances from east to west across the area in township 55, and for two miles to the north into township 56. The bedrock surface shows very little relief except in the extreme east part of the area where a low hill rises to a height of about 300 feet above the general bedrock surface. The area is included in the Mundare Plain region (Farvolden, 1963a) described later in this bulletin. The bedrock channels that do appear on this map are the second important type of aquifer in the area. From a study of aerial photographs, it is believed that these bedrock valleys may have been cut mainly by glacial erosion. However, some interglacial or preglacial erosion may also have taken place, and the configuration of the bedrock topography is considered to be the result of a combination of these three phases of erosion. In what follows no attempt will be made to distinguish the origin of any of the bedrock valley features. The bedrock topography shows a very close resemblance to and exerts a strong influence on the present-day topography.

Seismic profiles were run along the north-south roads one and two miles west of Andrew, and also along east-west roads to the north and south of Andrew (Fig. 3). The purpose of the profiles was to obtain cross sections of the bedrock topography to aid in the search for buried glacial or preglacial drainage-channels in the area and to fill in the large gaps where no lithologic logs were available. The results of the work gave no indication of the presence of buried channels other than those previously known from surface expression and drill-hole data.

A resistivity survey was limited to testing short profiles and isolated points. The purpose of this was to gain some idea of the depth and extent of surface sand and gravel at the places tested. Analysis of the results confirmed previous indications that porous and permeable materials were likely to be shallow and confined laterally.

Groundwater in the Bedrock

The succession in the undivided Belly River Formation typically is a sequence of alternating clay, poorly or completely cemented sandstone beds, ironstone bands, silty sandstone, sandy clay, carbonaceous shale and coal seams. There is an absence of lateral continuity of any particular stratum; and consequently the bedrock is without any distinctive extensive aquifers. This limits the occurrence of recoverable groundwater to individual sand lenses and small coal-seams, but these appear to behave as members of a larger single hydrologic unit. The balance of the unit is made up of the less permeable materials from which water cannot be removed economically.

The data available from pump tests, though very inadequate (Table 1), show that the wells completed in the aquifers tested are poor producers, and are able to supply only domestic requirements of water. Many wells in the area can often be pumped dry under normal conditions of usage, and yields exceeding 3 to 5 gpm (gallons per minute) must not be expected

Table 1. Bedrock Aquifer Test Data, Andrew Area

Well location ¼	Sec.	Tp.	R.	Rate of pumping or bailing (gpm)	Length of test (hours)	Available drawdown (feet)	Estimated safe yield (gpm)	Trans-missi-bility (gpd/ft)
NE	12	55	16	6	1	190	0.5	6
NE	27	55	16	15	1	240	5	53
NW	32	56	16	5	2	164	2	20

from bedrock sources. Despite the very limited data available, it is believed that the range of 6 to 53 gpd/ft (gallons per day per foot) for the coefficient of transmissibility is representative of the Belly River Formation.

Groundwater in the Glacial Drift

Wells in the glacial drift derive water from three sources: lenses of sand and gravel in the ground moraine, dune sand, stream-trench deposits, and interglacial channel deposits.

Wells finished in sand and gravel lenses in the ground moraine usually yield supplies of water sufficient only for domestic requirements.

In the northwest part of the area covering township 57, range 17 is an area of dune sand estimated to have a maximum thickness of 40 feet. This sand is saturated at very shallow depths, from 8 feet to 15 feet below the surface, and groundwater is obtained by means of bored or dug wells less than 20 feet deep. This aquifer is a more important source of water supply than either the sand and gravel lenses in the ground moraine or the shallow bedrock aquifers. Wells finished in this sand furnish ample supplies of water for domestic and stock purposes, and there are no reports of these wells having been pumped dry under normal conditions of usage.

The granular deposits in the interglacial channels and stream-trench systems constitute the best aquifers in the area. All the pump-test data available on glacial drift aquifers (Table 2) refer to aquifers within these deposits. These data reveal transmissibility figures ranging from at least 135 to 1,500 gpd/ft. It is apparent that there are prospects of developing wells to produce about 50 gpm in these deposits, and that they are the only possible sources for municipal and industrial supplies of groundwater.

Some sorted granular deposits occur within and near the village of Andrew. However, the very shallow depth of these deposits seriously limits their reliability for the development of municipal water supplies. Research Council test holes to the west, south and southeast of Andrew (Fig. 3) did

Table 2. Drift Aquifer Test Data, Andrew Area

Well location ¼	Sec.	Tp.	R.	Rate of pumping or bailing (gpm)	Length of test (hours)	Available drawdown (feet)	Estimated safe yield (gpm)	Transmissibility (gpd/ft)	Remarks
NE	11	56	15	11	48	70	5	135	Pumping rate began at 25 gpm, but dropped to 11 gpm
NE	11	56	15	26	24	130	90	1,500	
SE	14	56	15	7	52	200	—	—	
NW	5	56	16	60	0.5	45	—	—	Water flowed. Drawdown at 60 gpm reported as 5 feet in 29 min.
NE	16	56	16	16	0.5	22	—	—	Water flowed. Drawdown at 16 gpm reported as 17 feet in 24 min.
SE	32	56	16	7.7	11	10	2.5	524	

not encounter sand and gravel below 15 feet. As the depth to the non-pumping water level is 9 feet, and as there may be a fluctuation in the water level of about 2 or 3 feet, the very low head in this aquifer does not allow for a drop in the water level in dry years, nor for heavy demands that may be made upon the aquifer by the residents. The above considerations indicate that these deposits are marginal insofar as their suitability for the development of municipal supplies is concerned.

Hydrology

Piezometric Surface

The piezometric surface of the Andrew area (Fig. 4), is based on water levels in wells ranging in depth from 11 feet to 100 feet. Most of the measurements were made by the writer. From figure 4 it can be seen that the movement of groundwater is from local centres where the piezometric surface is high to areas where it is low. The discharge of groundwater coincides with that of the surface water, and takes place along interglacial channels, stream trenches and creeks in the area.

The important feature of the piezometric surface for the area is its close resemblance to the surface topography. This suggests that recharge to the aquifer takes place by local precipitation. The similarity of the piezometric surface to a water-table surface indicates the aquifers in the area are only partially confined. There is no significant difference between the water levels for wells completed into bedrock and drift aquifers, and it is not possible to contour separate piezometric surfaces for these two types of aquifers. Thus, both the bedrock and drift aquifers appear to belong to one hydrologic system.

It is believed that groundwater produced in the area is derived mostly from replenishment of the aquifers by local precipitation. This opinion is supported by reports of fairly rapid fluctuations of water levels after rainfall during the summer months and by some reports of the shortage of water during long dry spells, and the reliance on the occurrence of regular rainfall. Table 3 illustrates the general decline in water level during the dry summer of 1958. Because of the importance of regular recharge of the aquifers by rainfall and because there is no evidence to prove a long-term lowering of the piezometric surface throughout the area, it is believed that the storage capacity of the aquifers in the area is very limited and that only a small quantity of groundwater in the area is derived from storage.

Water-Level Fluctuations

After the initial survey of farm wells in June, 1958, nine wells, considered to be unused, were selected and measured at the end of August

Table 3. Water Level Decline, Andrew Area

Well location				Depth of well (feet)	Depth to water (feet)		Rise or fall (feet)
¼	Sec.	Tp.	R.		June	August	
NW	22	55	15	46.7	11.7	11.08	+0.62
SW	6	56	14	24.6	11.7	14.00	−2.30
SW	7	56	14	16.7	3.7	5.54	−1.84
SE	10	56	15	38.2	13.8	14.50	−0.70
NW	24	57	15	36.0	5.60	11.14	−5.54
NW	5	57	16	62.0	7.0	9.65	−2.65
SE	5	58	15	42.3	6.1	8.21	−2.11
SW	2	58	16	80.0	20.0	15.7	+4.30
SW	3	58	17	44.8	6.25	13.3	−7.05
		Total rise					4.92
		Total fall					22.19
		Net fall					17.27
		Average fall					1.92

to observe the effects of the dry summer in the area. Graphs for three observation wells in the area are shown in figure 5 with temperature and precipitation data recorded by the weather station at Ranfurly, 40 miles southeast of Andrew.

The figures in table 4 for temperature and precipitation are for a weather station at Vegreville which is in an adjacent area about 30 miles to the southeast of Andrew.

Table 4. Meteorological Data, Vegreville Weather Station

Month	Temperature (°F) Maximum	Minimum	Average	Precipitation (inches)
May	84	28	56.5	1.16
June	87	34	57.6	1.34
July	93	37	62.0	0.96
August	92	36	63.5	1.91
September	79	30	49.8	4.33
October	67	18	42.3	0.13

From the information presented in tables 3 and 4, and in figure 5, it can be observed that there is a possible correlation between water level fluctuation and rainfall. The decline in the water level observed in most wells in table 3 was expected, and the longer-term records shown on figure 5 also reflect the influence of precipitation.

Well Completion

Most bedrock wells are bored or dug and are completed as open holes. Many of these wells have 8 feet or more of cement cribbing near the surface to prevent surface seepage. There are very few drilled wells in the area and in some of these sand problems have occurred.

Well completion difficulties do occur in drift aquifers due to quicksand problems. Because of these problems many bored wells do not penetrate the aquifer, for the well driller usually terminates the well as soon as the sand and water are encountered.

When sand problems arise in wells completed in either bedrock or drift aquifers, adequate consideration should be given to well completion and development. Careful drilling and sampling is therefore important so that the drillers will have the information necessary to properly screen and gravel pack wells if this is required (Farvolden, 1961a).

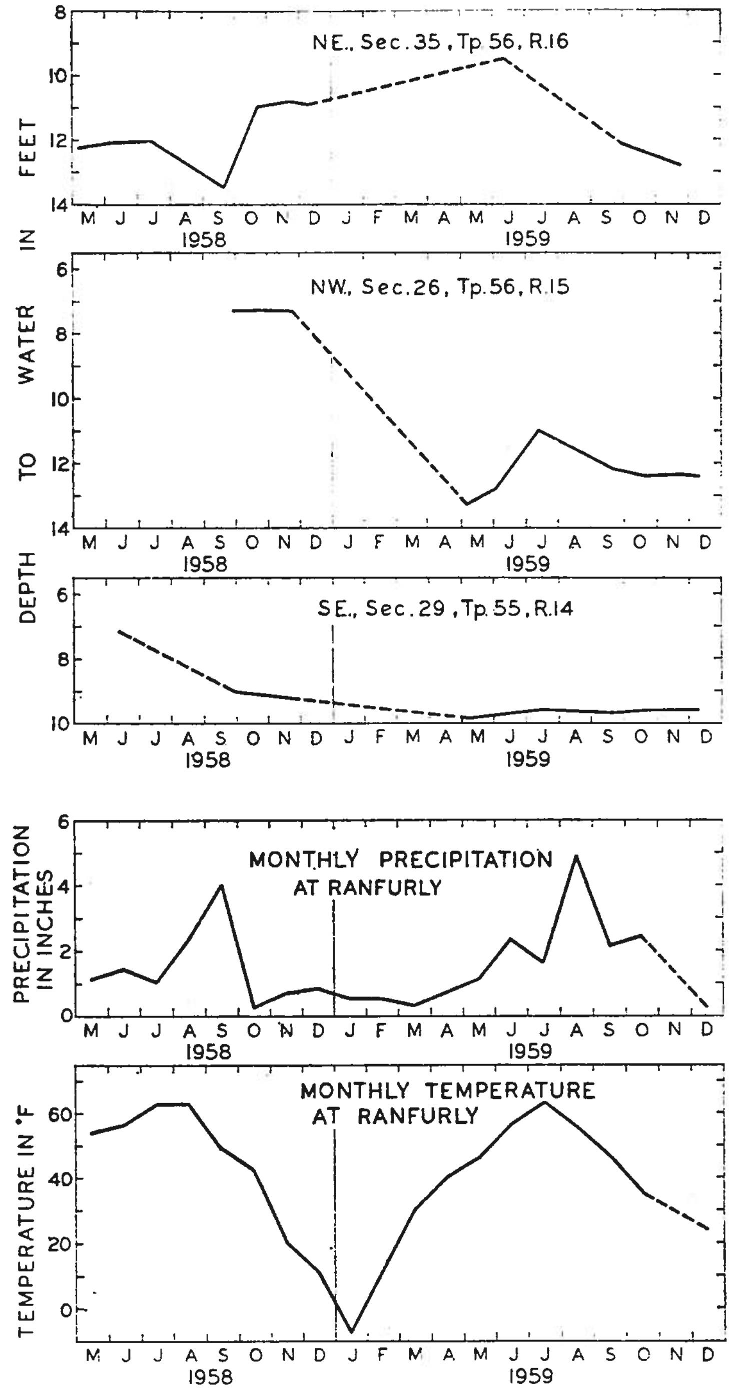

Figure 5. Well hydrographs and meteorological data, Andrew area.

Groundwater Quality

The total solids content of the bedrock waters ranges from 430 to 5,200 ppm (parts per million) and averages about 1,500 ppm.

The hardness of the bedrock waters, which is due to calcium and magnesium bicarbonates, ranges from 5 to 1,000 ppm, but is generally less than 300 ppm. However, because most of the bedrock wells sampled were shallow, about 50 feet deep, the hardness is commonly quite high for bedrock waters in Alberta. Where vertical seepage of groundwater has taken place to depths of 200 to 300 feet, permitting base exchange of calcium for sodium, softening of the water has occurred. This has been caused by its movement through a sequence of strata containing bentonitic sandstones. Water from deeper wells has a hardness of 5 to 65 ppm, which is more representative of bedrock supplies (Foster and Farvolden, 1958).

The following constituents are also found in bedrock waters: chlorides, ranging from zero to 125 ppm, and averaging 33 ppm; iron, ranging from 0.1 to 2.0 ppm, and averaging 0.5 ppm; and sulfates, reported to be as high as 2,400 ppm, but usually below 500 ppm, the suggested limit for human consumption. Figure 6 shows graphically the composition of groundwater in bedrock and drift aquifers.

Where contamination occurs, it is mostly caused by the nitrate content which has been reported to be as high as 140 ppm.

The chemical quality of water from the glacial drift cannot be discussed in any detail, for there are only two analyses available of water from this source. However, it can be expected that water from these deposits will probably be hard and high in iron.

Economics of Groundwater Supply

Generally, the development of groundwater supply in the area is only for domestic and stock requirements.

There is a very distinct preference for dug and bored wells which appears to have been influenced by three factors: the quantity of water which can be obtained from the well at any one particular time, the cost of the well, and the equipment possessed by local water-well drillers. Large-diameter wells are preferred to drilled wells because their storage capacity compensates, to some degree, for low yields. The costs of bored wells about 50 feet deep are much less than those for drilled wells of the same depth. The former may be about $100.00 per well and the latter, complete with pumping equipment, about $400.00 per well. Drilled wells average about 200 feet deep in the area and will cost nearer $1,000.00.

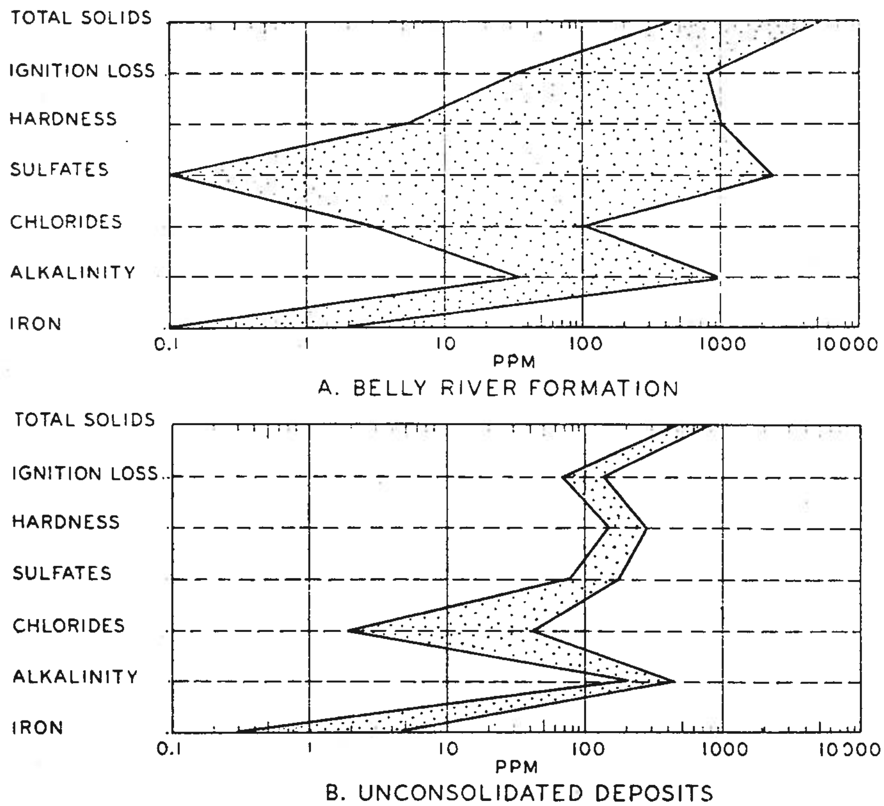

Figure 6. Ranges in chemical composition of groundwaters, Andrew area.

Municipal wells, for which capacities of over 20 gpm may be required, are likely to be more expensive. It may be necessary to screen and gravel pack the wells, and the costs may range from \$3,000.00 to \$5,000.00. However, the major item of expenditure for the village of Andrew will be determined by the proximity of sites—interglacial channels and stream-trench systems—at which adequate supplies of groundwater are available. The cost of laying pipeline may be about \$20,000.00 per mile.

The results of this study, based upon the available information gathered during the survey, show it is likely, though not definitely proven, that sufficient supplies of groundwater for the village of Andrew will be obtained only from interglacial channels or stream-trench systems. The location listed fourth in table 2—about 5 miles from the village—is suggested as the best site for further prospecting. Additional exploration and pump testing at this location may cost about \$2,000.00. If such exploration should prove successful, the total expenditure for the village of Andrew in bringing groundwater supplies from this site might be over \$100,000.00.

Conclusions

Bedrock and most drift aquifers in the area are poor producers, and are suited only for domestic and stock water supply requirements.

It is only in the interglacial channels and stream-trench systems that moderately large capacity wells, about 50 gpm, may be obtained. It is in these channels that future prospecting for groundwater supplies for municipal purposes should take place.

Well completion can often be an important consideration in the development of private and municipal wells in both the bedrock and drift aquifers. Appropriate well finishing and development are factors likely to increase well yields.

Groundwater from the bedrock and drift aquifers, though generally suitable for human consumption, is likely to be hard and high in iron.

The cost of obtaining groundwater supplies for municipal use is largely influenced by the location of villages in relation to interglacial channels and stream-trench systems.

GROUNDWATER GEOLOGY AND HYDROLOGY OF THE LLOYDMINSTER AREA, ALBERTA

by
E. Gordon Le Breton

Introduction

Purpose and Scope of the Report

The city of Lloydminster (population 5,800; location 53° 15′ north latitude and 110° west longitude, Fig. 2) has been supplied with water from wells from the time that the area was first settled in 1903. About 1955, it became obvious that additional supplies of water were required to meet the growing demands of the city. Thus, a survey was conducted in the summer of 1957 to study the groundwater resources of the area lying within townships 48 to 52, range 26, west of the Third Meridian to range 2 west of the Fourth Meridian.

This report is based on a study of well records supplied by water-well drillers, water supply papers published by the Geological Survey of Canada, information supplied by oil companies, data obtained from a survey of many farm wells, and reports on the chemical quality of ground-water from samples analyzed by the Provincial Analyst. In interpreting the geology, aerial photographs, geological reports and maps, lithologic logs from drillers, and electric logs supplied by oil companies were used to support the field studies. This work was followed by geophysical prospecting using seismic and earth-resistivity methods. Finally, three holes were drilled to test the results and conclusions of this work.

Description of the Area and Physiography

The climate of the area is continental, humid. Figures supplied by the Meteorological Branch, Department of Transport, Edmonton, for the Vermilion weather station, 45 miles west of Lloydminster, are as follows: mean temperature for July, 61.8°F; mean temperature for January, 2.5°F; the mean annual temperature, 33.6°F; and average annual precipitation, 16.41 inches.

The land surface is flat to gently rolling and the local relief seldom exceeds 20 feet except near valleys. The surface slopes at 20 feet per mile from a northwest-trending ridge flanking Blackfoot Creek in the southwest part of the area, toward Big Gully Creek in the northeast. This ridge forms a topographic divide in the area, and the surface runoff, mainly consisting of temporary streams, has its main drainage to the northeast into Big Gully Creek (Fig. 7).

Acknowledgments

The author wishes to express his thanks to local drillers and residents for information supplied during the field survey, and also to the personnel of Excelsior and Husky Oil Refineries for their assistance and co-operation.

Groundwater Geology and Hydrology

Glacial drift, forming an almost continuous cover over the whole area, overlies the Lea Park Formation, and the divided or the undivided Belly River Formation (Fig. 8); bedrock outcrops occur only along Big Gully Creek and west of Blackfoot Creek.

Within the area covered by this report, it is estimated that 80 to 90 per cent of the water wells are completed in the Ribstone Creek Sandstone, the only significant bedrock aquifer. A very few wells, however, are known to obtain water from the Victoria Sandstone which lies about 100 feet below the base of the Ribstone Creek Sandstone. This is the sandstone referred to by Hume and Hage (1947a, b) as forming "the upper 180 feet of the Lea Park shale". The remaining wells are completed in sand and gravel deposits in the glacial drift.

The area lying between Big Gully Creek and Blackfoot Creek is covered by a deposit of ground moraine, consisting chiefly of till and containing minor lenses of sand and gravel. The thickness of this moraine varies from 20 to 175 feet and averages about 50 feet. Numerous crevasse fillings are superimposed upon the ground moraine and are a prominent feature of the surface topography (Gravenor, 1956). These are long narrow ridges commonly up to 1 mile in length. The most important glacial features with respect to groundwater supplies in the area are the spillways of Big Gully Creek and Blackfoot Creek (Fig. 7). These spillways are steep-sided valleys cut by glacial meltwater (Gravenor, 1956). In places, as a result of changes in gradient or of slackening in the rate of flow of glacial meltwater in the final stages of deglaciation, they contain thick deposits of sand, silt and some gravel.

The Ribstone Creek Sandstone Aquifer

Most wells in the area obtain water from the Ribstone Creek Sandstone; consequently this is the only bedrock aquifer that will be discussed. The top of the Ribstone Creek Sandstone is at depths of from 50 to 150 feet below ground level; its thickness ranges from 50 to 150 feet and averages about 100 feet. This unit consists of a succession of interbedded sandstone, silty sandstone, and siltstone strata (Nauss, 1945; Shaw and Harding, 1954). In Research Council test hole No. 1 siltstone and silty sandstone strata, here considered to be the Ribstone Creek Member,

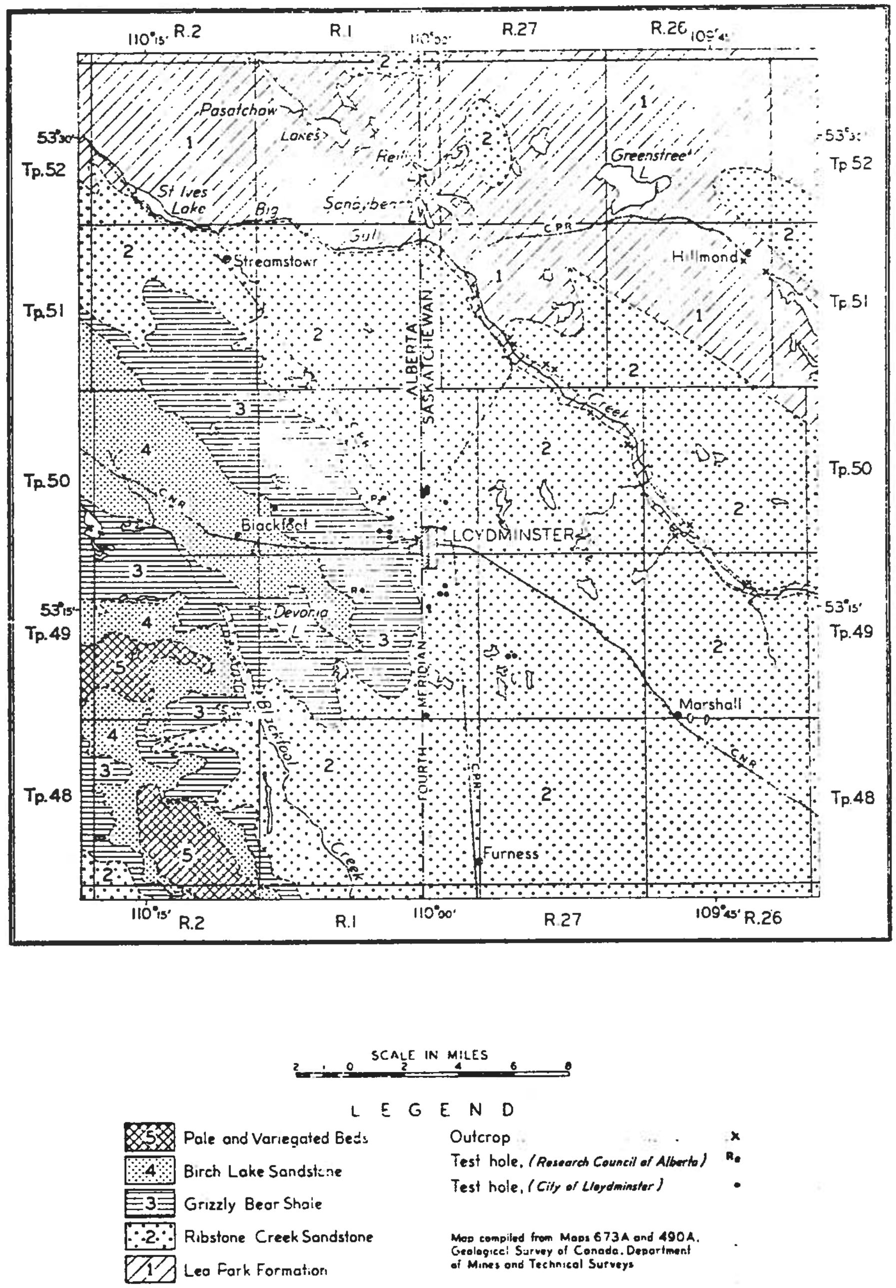

Figure 8. Geological map, Lloydminster area.

were encountered from 140 feet to 260 feet. In Research Council test hole No. 2 (Fig. 8) hard sandstone, 1 foot thick, was encountered at 116 feet, underlain by interbedded silty sandstone and thin sandstone beds to a depth of 215 feet. Samples of the Ribstone Creek Sandstone obtained from these test holes show it to be a grey, brown and green sandstone, consisting of approximately 60 per cent quartz and 40 per cent chert grains cemented with calcite. The degree of cementation ranges from poor to complete, and the degree of sorting is very variable within the sandstone beds, giving rise to wide and unpredictable variations in permeability.

Variation in development of the sandstone beds of the Ribstone Creek Member is best illustrated by electric logs, and three logs typical for the area have been selected to give a cross section A-B 12 miles long, trending from northwest to southeast (Fig. 9). The resistivity curves illustrate the irregular occurrence of siltstone and thin sandstone beds within the Ribstone Creek Member. The absence of lateral continuity of the more porous and permeable sandstone beds makes the search for groundwater very difficult. From the test drilling and electric logs, the thickness of the Ribstone Creek Sandstone within the area is shown to range from 99 to 150 feet.

Wells completed in the upper 50 feet of the Ribstone Creek Sandstone may produce from 5 to 15 gpm (gallons per minute) and they provide ample water for domestic and stock requirements. Wells completed in the lower 100 feet of the aquifer, which is more permeable, have capacities ranging from 15 to 120 gpm. Test holes located to the southeast and southwest of the city (Fig. 8) and completed in the lower part of the Ribstone Creek Sandstone indicated that the Ribstone Creek Sandstone has a lower permeability there and is suited to supply water only for domestic and stock purposes. In holes located to the north and west of Lloydminster, pump test results show that wells can be obtained with capacities ranging from 25 to 100 gpm. Existing wells 200 feet and 260 feet deep, at the Excelsior Oil Refinery (SE. ¼, Sec. 2, Tp. 50, R. 1*), and city wells Nos. 2 and 3, about 150 and 170 feet deep, have reported capacities ranging from 50 to 120 gpm. This shows the lower part of the Ribstone Creek Sandstone aquifer has varying permeability, the highest values being found within and to the northwest of the city. Examination of the data from test drilling and existing water wells indicates, therefore, that the best locality for prospecting for groundwater supplies lies to the north and west of Lloydminster. For the second, third and sixth holes listed in table 5 from which the most useful of the data have been obtained, the transmissibility ranges from at least 3,100 to 4,700 gpd/ft (gallons per day per foot), and the average transmissibility is 3,400 gpd/ft.

* Unless otherwise stated, all locations are west of the Fourth Meridian (110° west longitude).

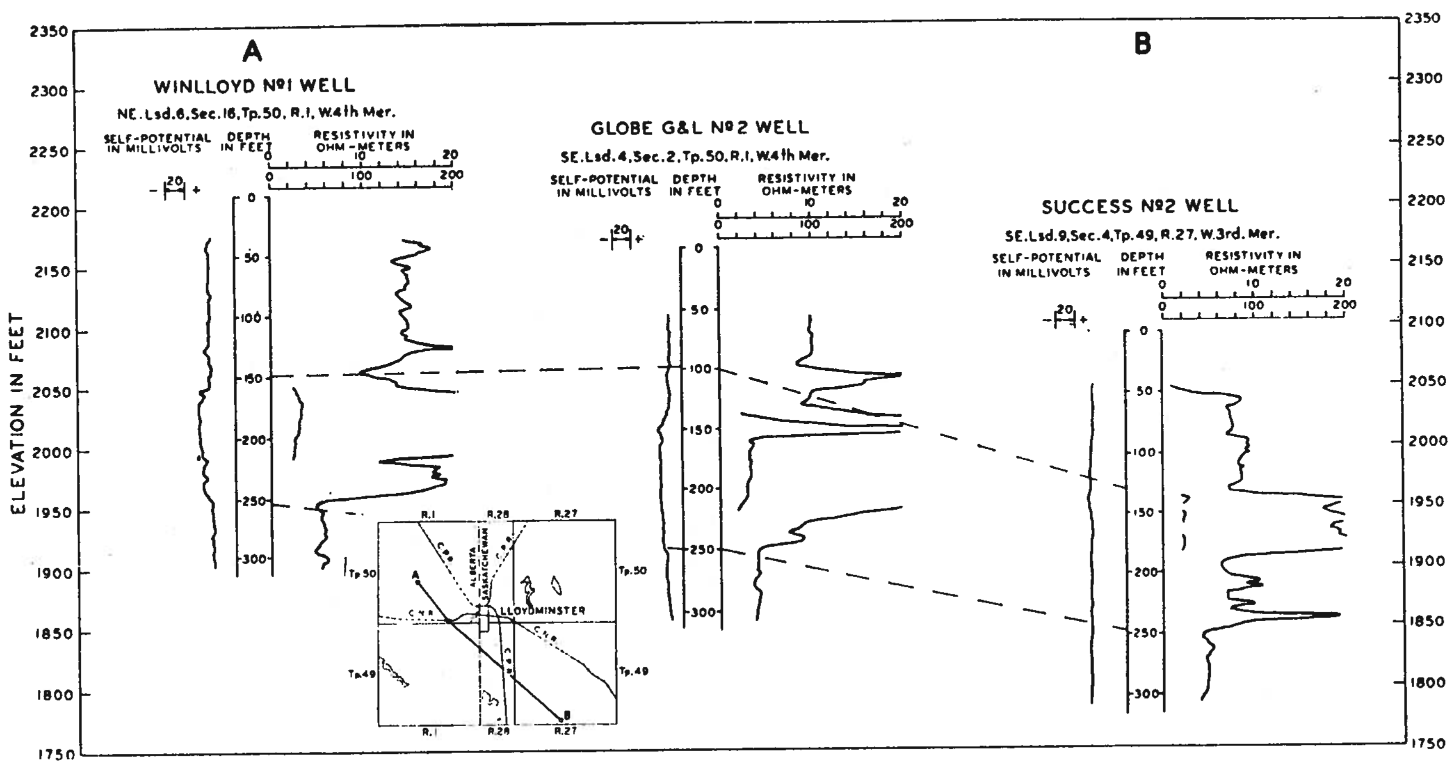

Figure 9. Electric log cross section, Lloydminster area.

The data presented in table 5 were obtained from short pump tests conducted during an exploration program. These pumping tests were intended to disclose only major variations in the possible well yields at different sites, but the data probably do provide fairly reliable values for the coefficient of transmissibility. No values were obtained for the coefficient of storage, but from knowledge of other bedrock aquifers in the province, it may be about 5×10^{-4}. The calculations for transmissibility were based on the modified nonequilibrium formula (Cooper and Jacob, 1946). The equations used are explained in Appendix B.

Because the dimensions of the aquifer are unknown, the total amount of groundwater in storage in the Ribstone Creek Sandstone cannot be determined. However, some observation of the effect of groundwater usage upon storage in the area can be made from the map of the piezometric surface (Fig. 7), which shows a major cone of depression produced by high groundwater extraction around Lloydminster. This cone of depression has considerably changed the local configuration of the piezometric surface, and it is believed that the piezometric surface has been lowered a maximum of 35 feet during the past 20 years. This indicates that the many closely spaced, high-capacity wells have derived much of their water from storage. The reported declines in production rates and the falling water levels in industrial, municipal and some private wells are caused by mutual interference of closely spaced, high-capacity wells and have resulted in a reduction in the available drawdown.

However, the declines in the yields of industrial and some private wells at Lloydminster may not be caused only by the drop in the water level. The optimum development of groundwater from the Ribstone Creek Sandstone is dependent upon good well-completion practices, maintenance and spacing. There is little cause for concern over well completion in the case of domestic and stock wells which need to be pumped at rates of only 3 to 5 gpm. The present methods of open-hole or slotted-casing completions appear to be adequate. However, a sand problem commonly occurs in industrial and municipal wells pumping at rates exceeding 15 gpm. Suitable well completion and development are necessary in such wells, and screening and possibly gravel packing are desirable to deal with this sand problem. With appropriate completion, the well efficiency should improve and thus higher production rates may be attained. Regardless of the type of completion, wells will require periodic servicing to maintain high yields and this will vary with the extent to which they are used, and the strata from which they obtain water. Well spacing is also an important factor in groundwater development in the area. Though no conclusive evidence is available, in order to obtain near-maximum yields from wells in the same zone, spacings of 1,500 to 2,000 feet may well be necessary.

Table 5. Aquifer Test Data, Lloydminster Area

Well location ¼	Sec.	Tp.	R.	Mer.	Rate of pumping or bailing (gpm)	Length of test (hours)	Available drawdown (feet)	Estimated safe yield (gpm)	Trans-missibility (gpd/ft)	Aquifer	Remarks
NW	2	50	28	3	4	9	29	40	3,500	Ribstone Creek Sandstone	Pumping of a nearby well affected the results of this test
SW	14	50	28	3	14	7	36	50	3,100	"	Test holes 250 feet apart
SW	14	50	28	3	4.5	13	16	35	4,700	"	
NE	36	49	1	4	35	2	45	20	900	"	Private well
NW	36	49	1	4	30	2	45	15	800	"	Private well
SE	14	50	1	4	10	7	80	120	3,200	"	
SW	1	52	28	3	—	—	—	350+	—	Glacial sand	New city wells at Sandybeach Lake

NOTE: Except for the two private wells, the measurements were taken in gauge holes about 4 feet from the pumping wells.

An automatic water well recorder was placed at the Husky Oil Refinery in the city of Lloydminster on an abandoned well completed in the Ribstone Creek Sandstone to record the effects on the aquifer of the pumping of several closely spaced wells. From the hydrograph shown in figure 10, it is clearly demonstrated that up to the end of 1959, the fluctuations in the water level have been caused by the seasonal demand for groundwater. There is a noticeable drop in the water level during the summer months when the demand is high, and the recovery takes place during the winter months when groundwater usage is lower.

Drilling to depths in excess of 250 to 300 feet, the approximate lower limit of the Ribstone Creek Sandstone, is not advised. There is little prospect of encountering any deeper water-bearing sands in the underlying Lea Park Shale, although the Victoria Sandstone is locally developed. Where the Ribstone Creek Sandstone is absent, supplies of water are unlikely to be obtained from the bedrock, and the glacial drift will have to be explored for water supply.

Groundwater in the Glacial Drift

There are two types of aquifer in the glacial drift. One of these, the sand and gravel lens, occurs in the ground moraine, and the other—the sorted granular deposits—is found in the spillways. Water-bearing sand and gravel lenses are occasionally found at depths from 20 to 50 feet, but are suitable only to supply domestic and stock requirements of water. The water is generally obtained by means of bored or dug wells, and quite often these can be pumped dry under normal conditions of domestic and stock usage.

The spillway deposits provide good groundwater reservoirs which are readily replenished by infiltration from precipitation. The water table in these deposits is about 10 feet below ground level, and the general direction of groundwater movement is the same as that of the surface drainage. Where the deposits are sufficiently thick and well sorted, they constitute the more prolific aquifers in the area. The most important spillway is the extension of Big Gully Creek from Sandybeach Lake northwest of Pasatchaw Lake (Fig. 8). Both these lakes are slightly north of the map area. Pump tests conducted near Sandybeach Lake indicated that induced infiltration from the lake could give rise to wells with very high yields. As a result of these tests, the city of Lloydminster chose to develop two wells in the vicinity of the lake. Each well yield exceeds 350 gpm. The total storage in the area between Pasatchaw Lake and Sandybeach Lake is estimated to be 11.8×10^9 gallons. Further work is essential to determine the percentage of this supply of groundwater that could be economically developed for industrial use. There are other spillways in the area, but the

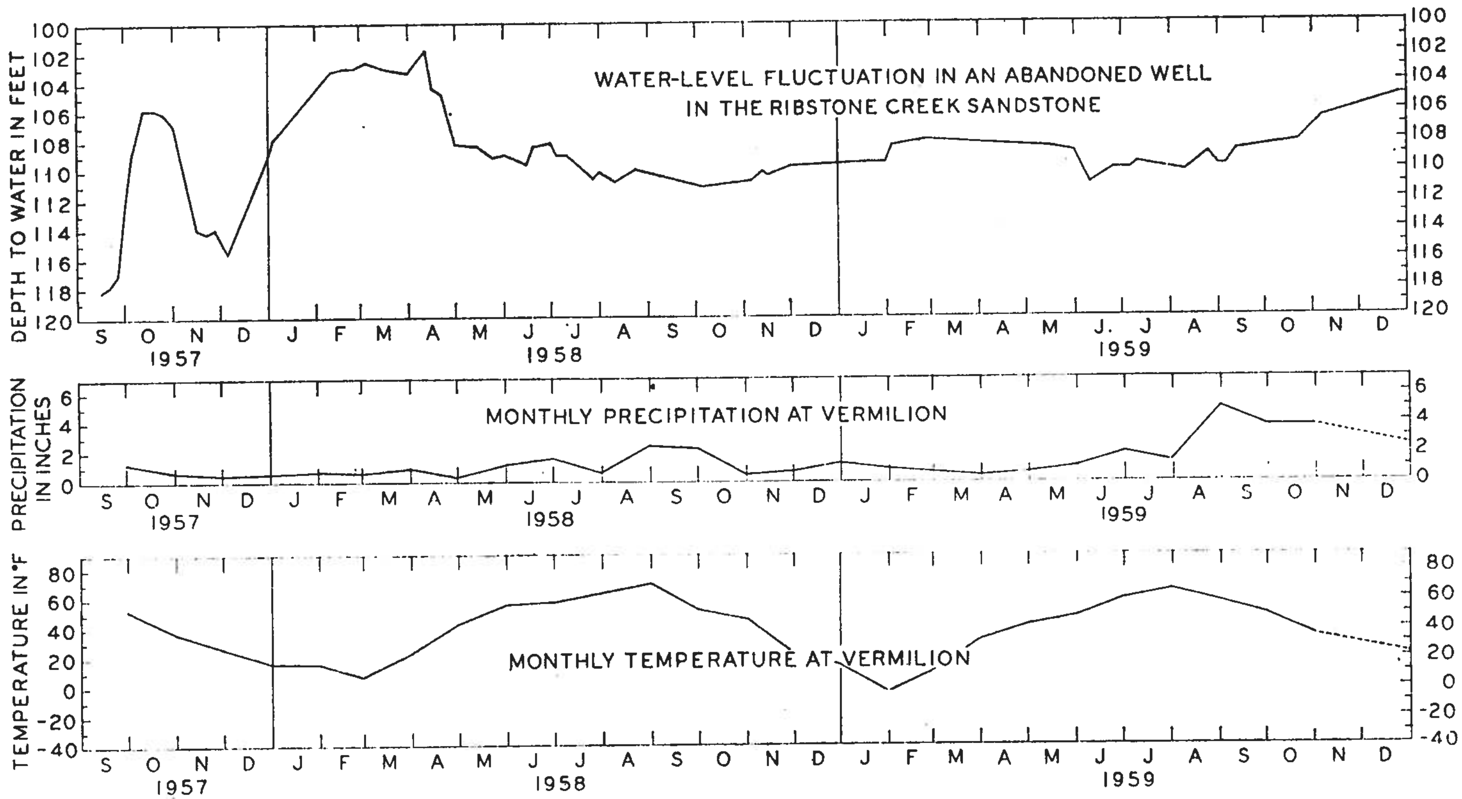

Figure 10. Well hydrograph and meteorological data, Lloydminster area.

potential groundwater supply will have to be determined very carefully; of these, the deposits in Blackfoot Creek are possibly the most important for groundwater supply. The results of earth-resistivity surveys at points along this creek have indicated porous and permeable deposits up to and over 100 feet in thickness. At present only shallow wells are sited in these deposits and, although ample supplies of water are obtained from them for domestic and stock requirements, test drilling is necessary to determine whether high-capacity wells could be developed. Maps outlining spillways and other glacial features over the area between 110° and 112° west longitude and 53° and 54° north latitude were constructed by Ellwood (1961).

Because the spillways are a very important feature of the glacial geology in relation to groundwater supplies, methods of locating these spillways and determining their potential productivity are worthy of some discussion.

Spillways can be located on aerial photographs and it is also possible to determine thereby which portions contain glaciofluvial deposits. Field investigations are required to determine the lithology and thickness of the deposits. Geophysical surveys using seismic and resistivity methods may be used to determine the depth of the spillways and the thickness of coarse-grained deposits. Drilling and pump testing are essential to determine the transmissibility and storage of the deposits, and the capacity and spacing of water wells.

Piezometric Surface

The piezometric surface (Fig. 7) for the Lloydminster area has been based on water levels in wells 100 to 260 feet deep drilled into the Ribstone Creek Sandstone, and also includes some wells reported to have been completed in sand and gravel. Figure 7 shows that most of the groundwater movement is toward the northeast. It also shows that the spillways of Big Gully Creek and Blackfoot Creek serve as discharge channels that help to drain the Ribstone Creek Sandstone. The most important observation derived from the examination of the piezometric map is the close resemblance between the piezometric surface and the topography. This suggests that recharge to the aquifer takes place by local precipitation.

Groundwater Quality

The Ribstone Creek Sandstone Aquifer

Groundwater in the Ribstone Creek Sandstone is very hard. Within a radius of 2 miles of the city of Lloydminster, the hardness ranges from 405 to 595 ppm (parts per million) calcium and magnesium carbonate.

Throughout the whole area, the hardness ranges from 50 to 940 ppm and averages 410 ppm. These high hardness figures for the Ribstone Creek Sandstone in this area are rather anomalous because the bedrock aquifers in Alberta are characterized by soft water (Foster and Farvolden, 1958). It is thought that in this area groundwater is very hard because of the absence of bentonite in the overlying materials through which the water percolates.

The total solids content ranges from 740 to 2,340 ppm, and averages 1,245 ppm. Although the sulfate content generally has a range from 230 to 500 ppm, averaging 376 ppm, a few cases are known where the sulfate exceeds 500 and reaches 860 ppm. Other constituents present in the water include the chloride ion, ranging from 13 to 78 ppm and averaging 35 ppm; sodium, ranging from 20 to 680 ppm; and iron which ranges from 0.2 to 5.2 ppm and averages 1.2 ppm. Figure 11 shows graphically the chemical composition of the groundwater in the Ribstone Creek Sandstone aquifer. The figures quoted above were from 35 analyses of well water.

Glacial Drift Aquifers

Groundwater obtained from drift aquifers in the area is commonly extremely hard, and a few samples have been collected that exceeded 1,000 ppm in hardness. The hardness ranges upward from 160 ppm and averages 507 ppm calcium and magnesium carbonate. The total solids content ranges from 220 to 4,138 ppm and average 1,069 ppm. The sulfate content ranges from zero to 1,603 ppm and averages 240 ppm. Other constituents present in the water include the chloride ion which ranges from 2 to 120 ppm and averages 24 ppm; nitrates, ranging from zero to 120 ppm and averaging 22 ppm; and iron, ranging from zero to 1.0 ppm and averaging 0.2 ppm. The figures for chemical constituents of groundwater from drift aquifers were obtained from eight samples of well water. The range in chemical composition of drift waters is shown graphically on figure 11.

Economics of Groundwater Supply

The most optimistic estimates of well capacities for bedrock wells near Lloydminster are about 100 gpm. To develop a groundwater supply of half a million gpd* (gallons per day) may well require an extensive drilling and pump-testing program.

Drilling may have to be conducted at three or four sites before one is located that is suitable for pump testing. The pump test should also include

* 1 gpm=1440 gpd

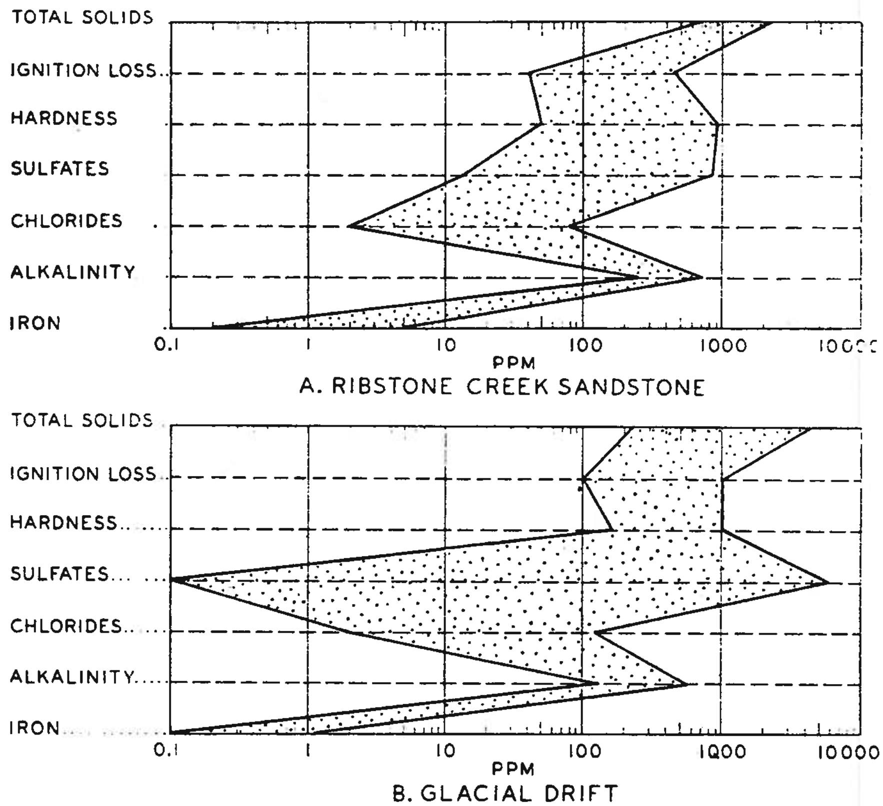

Figure 11. Ranges in chemical composition of groundwaters, Lloydminster area.

two observation wells at distances that will depend on geologic and hydrologic conditions. Four or five such sites may be required to locate an adequate supply. The results obtained may indicate that to develop a well field, well spacings of 1,500 feet to 2,000 feet are necessary. Preliminary test drilling and pump testing for each site located may cost $4,500.00. The cost of well connections and piping of water will be about $20,000.00 per mile. There are also the costs of well completion, well development, and of erecting pump houses—about $5,000.00 per well. Such a program will have to be considered for the development of adequate municipal or industrial groundwater supplies in the area.

However, if really large quantities of water are required for industrial purposes, then the economics of groundwater development from the spillways should be considered. Exploration of these will include costs of resistivity surveys, test drilling, pump testing, and completion of water wells. The total cost for all of this work may be about $20,000.00 to test each favorable location determined from aerial photographs.

Conclusions

Where the Ribstone Creek Sandstone aquifer is present in the area, groundwater supplies can be obtained. Locally, this aquifer is sufficiently porous and permeable for the development of municipal and industrial wells with capacities up to 100 gpm.

The glacial-drift aquifers may often supply domestic and stock requirements, but only in the spillway deposits can very high capacity wells with yields of about 350 gpm be obtained.

The shape of the piezometric surface in the area closely resembles the topography indicating recharge to the aquifer takes place by local precipitation.

Groundwater from the bedrock and drift aquifers, though generally suitable for human consumption, is very hard and has to be softened for use in boilers, homes, and laundries.

GEOPHYSICAL PROSPECTING FOR GROUNDWATER IN ALBERTA

by
D. H. Lennox

Introduction

Geophysical prospecting is the application of physical methods to the exploration for minerals and other subsurface deposits of value to man. Geophysical methods based on measurements at the earth's surface do not yield as complete information about the subsurface as test drilling, but their use is often justified when comparative costs are considered. Seismic refraction and earth resistivity, both surface-measurement methods, are being used in Alberta for groundwater exploration. Because the data they supply lack the detail that may be obtained from a test hole, they will never even under favorable circumstances replace test drilling, but they can reduce substantially the amount required and aid the geologist in a more intelligent selection of drilling sites.

Geophysical measurements may also be made beneath the earth's surface, using instruments lowered into wells and boreholes. These instruments and the associated instrumentation at the surface are known as well loggers. Loggers of many different types have been developed to measure many different physical quantities. The Research Council logger can be used to measure earth resistivity, earth potentials and gamma radiation. Advantages of a logger over the usual driller's log include the more precise location of formation boundaries and the possibility of using the data to calculate formation porosities and salinities of formation waters.

Acknowledgments

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Seismic Surveys

The seismograph is used in Alberta to determine drift thickness and its variation. From these data the locations of buried preglacial channels, and the courses that they follow, may be ascertained and promising drilling locations suggested. Bedrock over most of the province is shallow, necessitating the use of the seismic refraction method (Jakosky, 1957). Reverse shooting, described by Woollard and Hanson (1954), is the normal field technique. In this technique the detectors, or geophones, are equally spaced along a linear spread and explosive charges are fired successively at the two ends of the spread. The seismograph records for each geophone the elapsed time between the firing of one of the charges and the arrival of the resulting shock wave at the geophone. A seismogram is thus obtained for each of the two shots fired. The graphs of arrival times plotted against distances from the shot points determine the values of the seismic wave velocities in the drift and in the bedrock. Firing of a shot at each end of the spread allows for the effect on the observed velocities of dipping beds, if they are present. Woollard and Hanson (1954) describe the calculation of true velocities, apparent dips, and depths to bedrock at any point along the length of the spread. This calculation is unnecessarily complicated for the determination of bedrock depths in Alberta, and the simplified treatment of the same data adopted by Harris and Peabody (1946) has been substituted for it. If the dip of the bed is 10 degrees or less, the error introduced by the simplified calculation may be neglected.

During the first three summers that a seismic crew was in the field for the Research Council, about 250 miles of seismic line were covered with an average density of about one spread per mile and an average spread length of one-third of a mile. Most of the seismic prospecting was confined to the area included between the Fourth and Fifth Meridians, from townships 1 to 60, and the discussion in this report concerns only these results. About 30 miles of seismic line were covered in the Peace River area, just sufficient to indicate that the general conclusions derived from the southern Alberta data probably do not hold true in the Peace River country.

In southern Alberta the contrast between drift and bedrock velocities is small, varying between 1,000 and 2,500 fps (feet per second). Even the latter figure is somewhat lower than the minimum figures quoted by other authors (Johnson, 1954; Moore, 1952; Pakiser, 1957; Woollard and Hanson, 1954). Spread length is determined by the desired depth of penetration and by the velocity contrasts. Spread lengths in Alberta would be shorter and more manageable if the velocity contrasts existing in other areas were effective here.

Seismic Velocity Variations

Seismic wave velocity in a material is a function of the density of the material and its elastic constants (Nettleton, 1940) and the velocity differences observed on comparing lithologically dissimilar formations are attributable, at least in part, to their differences in composition. Differences are also observed, however, when similar formations are compared. The explanation lies in the relative amounts of compaction of the formations being examined. When a material is compressed the density and the elastic constants change, and the seismic wave velocity increases. A sedimentary unit will become consolidated with age alone, but the action will be accelerated if other units are deposited on top of it. There is, therefore, a relation existing between seismic velocity and geologic age and depth of burial. Faust (1951, 1953) has investigated the problem and has derived an empirical relation for an average sand and shale section. Acheson (1959) has published an empirical relation expressing velocity as a function of depth which he found to be applicable, with a small variation of the constant involved, over the Western Canada sedimentary basin. A statistical examination of the data collected by the Research Council has been made in an attempt to relate the velocities observed and their variations in a similar way to geologic age and depth of burial.

Drift Velocities

Table 6 lists the observed drift velocities in four discrete groups, significantly different from one another according to Student's t-test (Snedecor, 1956). The standard deviation for each group is also tabulated.

Table 6. Drift Velocities, Southern Alberta

Group	Velocity (fps)	Standard deviation (fps)
D1	1,800	200
D2	5,000	450
D3	5,850	400
D4	6,450	300

In addition to the velocities given in the table, a velocity appreciably less than 1,100 fps, the speed of sound in air, is observed in all areas. This velocity appears to be characteristic of a relatively thin surface layer, generally 15 feet or less in thickness. A similar low-velocity layer has been observed in other areas (Johnson, 1954; Moore, 1952) and a like phenomenon arises in the transmission of sound through water containing air bubbles (Wood, 1946). In the latter case the pressure variations act almost exclusively on the gas volume and a sound wave travelling through this medium is damped and its velocity of transmission lowered. Wood (1946) shows that for suitable proportions of air and water the velocity in the combined medium is lower than in either air or water alone. A similar phenomenon may be acting in the low-velocity surface layer.

Each area investigated by the seismic survey exhibited characteristic drift velocities which fitted into one or more of the groups summarized in table 6. The most common velocity was D2, occurring in 80 to 90 per cent of the spreads. The other three velocities have been found to date only in a 25-mile wide strip, lying immediately west of the Saskatchewan border and extending from Medicine Hat northwards to the Red Deer River. There is a good possibility that D1 and D3 are associated with the same sedimentary material in an unsaturated and a saturated condition, respectively. Comparable differences between elastic wave velocities in dry and water-saturated samples of the same porous medium have been observed in the laboratory (Wyllie, Gregory and Gardner, 1956) and similar field results in other areas (Kane and Pakiser, 1961) have been attributed to the transition across the water table from an aerated to a saturated zone.

There is some reason to believe that the D2 velocity group belongs to the same type of drift material as the D1 and D3 groups. The D2 group would be representative of velocities at the upper surface of an uppermost drift layer and the D3 of velocities at some depth within this layer, this depth being determined by the thickness of unsaturated D1 drift. The difference between the velocities in the two groups would, therefore, be a direct consequence of the differing depths of burial. Further support is given this hypothesis by the fact that the D2 group has not been observed in the Medicine Hat area, whereas D1 and D3 are observed only in this area.

The D4 velocity group may represent a drift of different origin or age. Johnson (1954) has correlated seismic drift velocity with age differences in an otherwise homogeneous drift in Illinois.

The boundaries between the two areas characterized by these distinct drift velocity conditions have not been defined with any certainty as yet. A large gap exists on all sides of the Medicine Hat strip in which no seismic surveying has been done by the Research Council. Future seismic results in the normal course of groundwater exploration should locate the two areas with more accuracy.

Bedrock Velocities

Table 7 lists the observed bedrock velocities which, like the drift velocities, may be separated into four discrete groups.

For the bedrock velocity groups there is a correlation with geologic age, but it is the reverse of that normally expected. B3 occurs mainly in Tertiary and Upper Cretaceous sediments, B2 in somewhat older sediments and B1 in the oldest Cretaceous formations. B4 occurs only in conjunction with B3 and is believed to represent a deeper horizon in the bedrock. Geographically, B3 and B4 have been found only in a narrow strip between

Table 7. Bedrock Velocities, Southern Alberta

Group	Velocity (fps)	Standard deviation (fps)
B1	6,650	350
B2	7,550	550
B3	8,300	450
B4	10,700	700

Calgary and Lethbridge, B2 in most of the other areas investigated and B1 within two small regions in that part of the province in which B2 is normally found.

To determine whether there is any relation existing between bedrock velocities and depths of burial, velocities were plotted against corresponding calculated amounts of drift cover for the members of each velocity group. The results for B1 and B2 are shown in figure 12. There is an obvious tendency for an increase in velocity to be associated with an increase in the amount of overburden. Correlation analysis (Snedecor, 1956) may be used to determine whether this seeming interdependence of the two quantities is statistically significant.

The correlation coefficient, calculated from the observed data, will be zero if there is absolutely no correlation between the two variables being tested. It will be unity if one is a function of the other, and only of the other. As the degree of correlation varies between these extremes, the correlation coefficient varies between zero and one. If there is a scattering of the observed points it will be necessary to test, using Student's t-test (Snedecor, 1956), whether the calculated value of the correlation coefficient is significantly different from zero. The correlation coefficient for the data shown in figure 12 is 0.679 and it is significantly different from zero. Stating the result in another way, bedrock velocity, at least for the B1 and B2 groups, does depend on the amount of drift cover, although this alone does not explain the variations observed.

It was previously remarked that the four bedrock velocity groups were discrete, meaning that the differences among them were statistically significant. This statement was based solely on a consideration of the observed velocities and not on the correlation between velocity and the amount of drift cover. A more reasonable interpretation of the data would now appear to be that the significant difference observed between the B1 and B2 groups was a consequence only of significant differences in the amounts of overburden in the B1 and B2 areas. If B1 and B2 are considered as members of a single velocity group, then there are two bedrock velocity areas outlined in southern Alberta to date: the relatively large B1-B2 area and the smaller area in which are observed B3 and the underlying B4. This interpretation has the virtue of removing the anomalous B1 islands from the B1-B2 areas.

Correlation coefficients were calculated for the B3 and B4 groups but were found to be not significantly different from zero. This result was most likely a consequence of the relative scarcity of data, rather than an indication that the velocity in these cases was not dependent on the depth of burial.

To return to the B1-B2 results, it is of interest to compare them with Faust's (1951, 1953) and Acheson's (1959) results for the variation of velocity with depth. Both authors have found an empirical relation, expressing velocity v as a function of depth z, of the form:

$$v = az^{1/n} \qquad (1)$$

It can be shown theoretically (Gassmann, 1951) that the magnitude of the constant a is determined by the elastic constants, density and porosity of the material and that n should equal six. Faust's results, based on the analysis of well-velocity surveys that covered extended areas in the United States and Canada, agree with the theoretical predictions. Acheson's work, on the other hand, suggests n is greater than 6 and usually in the neighborhood of 11 for similar surveys in the Western Canada sedimentary basin.

The bedrock velocities examined in this study are restricted to those essentially horizontal velocities occurring at or near the top of the bedrock surface. Those investigated by Faust and Acheson were vertical velocities and the ranges of depth extended over many thousands of feet. Despite these differences, it is of interest to determine a velocity-depth relation similar to equation (1) from the B1-B2 results, for Gassmann has shown that, except for a difference in the value of a, the law governing the variation of velocity with depth is the same for both horizontal and vertical velocities.

To determine the empirical velocity-depth relation in the form of equation (1), logarithms were taken of the velocities and of the corresponding amounts of drift cover. From the averages of the two, and from the slope of the regression line of log velocity on log depth, the required relation is derived. The calculated value of a is 4,000 fps and of n is 8 ± 1, and the resulting velocity-depth curve is shown in the scatter diagram of figure 12. This result, together with that of Acheson, suggests that a simple relation of the form of equation (1) with n equal to 6 does not hold for the Western Canada sedimentary basin. There are two possible alternatives that might be considered. Either n, because of some phenomenon not considered by Gassmann, is really greater than 6, or n is equal to 6 and the velocity-depth relation is of the form:

$$v = az^{1/6} + b \qquad (2)$$

There are good theoretical reasons for believing velocity to be dependent on depth as indicated by equation (2). According to Gassmann, equation (1) holds only if the pores in the material are interconnected and filled with vacuum or a gas. If the pores are liquid filled, the velocity-depth curve has a finite intercept on the velocity axis. It is interesting to note that a finite intercept has been observed in the laboratory by Paterson (1954) during an investigation of horizontal velocities in packings of glass spheres. Although the pores were air filled, the author felt that, due to the relatively high wave frequencies employed, the system possibly behaved like a liquid-filled system. If the Alberta results are used for the determination of an equation of the form of (2), a is found to be 2,550 fps and b to be 1,700 fps. The second version of the velocity-depth curve has not been plotted in figure 12 because it virtually overlaps the first curve. The maximum difference between the two curves is about 35 fps, occurring over the range of overburden thickness from 20 to 300 feet.

Earth-Resistivity Surveys

Earth-resistivity surveys have been carried out in Alberta since 1956, mainly during the summer months, using crews of high-school and university students. In contrast to the seismic refraction method, earth resistivity finds its greatest success, not in determining the elevation of the bedrock surface, but in prospecting directly for sands and gravels. These are the aquifers for which the geologist is looking and if resistivity could be relied on to locate them under all conditions, there would be little or no need for other geophysical techniques. The ability of the resistivity method to locate sand or gravel deposits depends on the thickness of the deposit relative to its depth of burial and on the resistivity contrast with the overlying material. The wrong combination of these factors can make a good aquifer undetectable. Hackett (1956) has described a location in Illinois where 25 feet of drift cover were sufficient to conceal the presence of 40 feet of sand and gravel.

In the field, resistivity measurements are made by putting a known current into the ground through two current electrodes and simultaneously measuring the voltage drop across two potential electrodes. From the current and voltage a quantity known as the apparent resistivity may be derived. The theory of the method is given by Jakosky (1957), who also lists a number of alternative ways of arranging the electrodes. That most usually used, and the one selected for resistivity surveying in Alberta, is

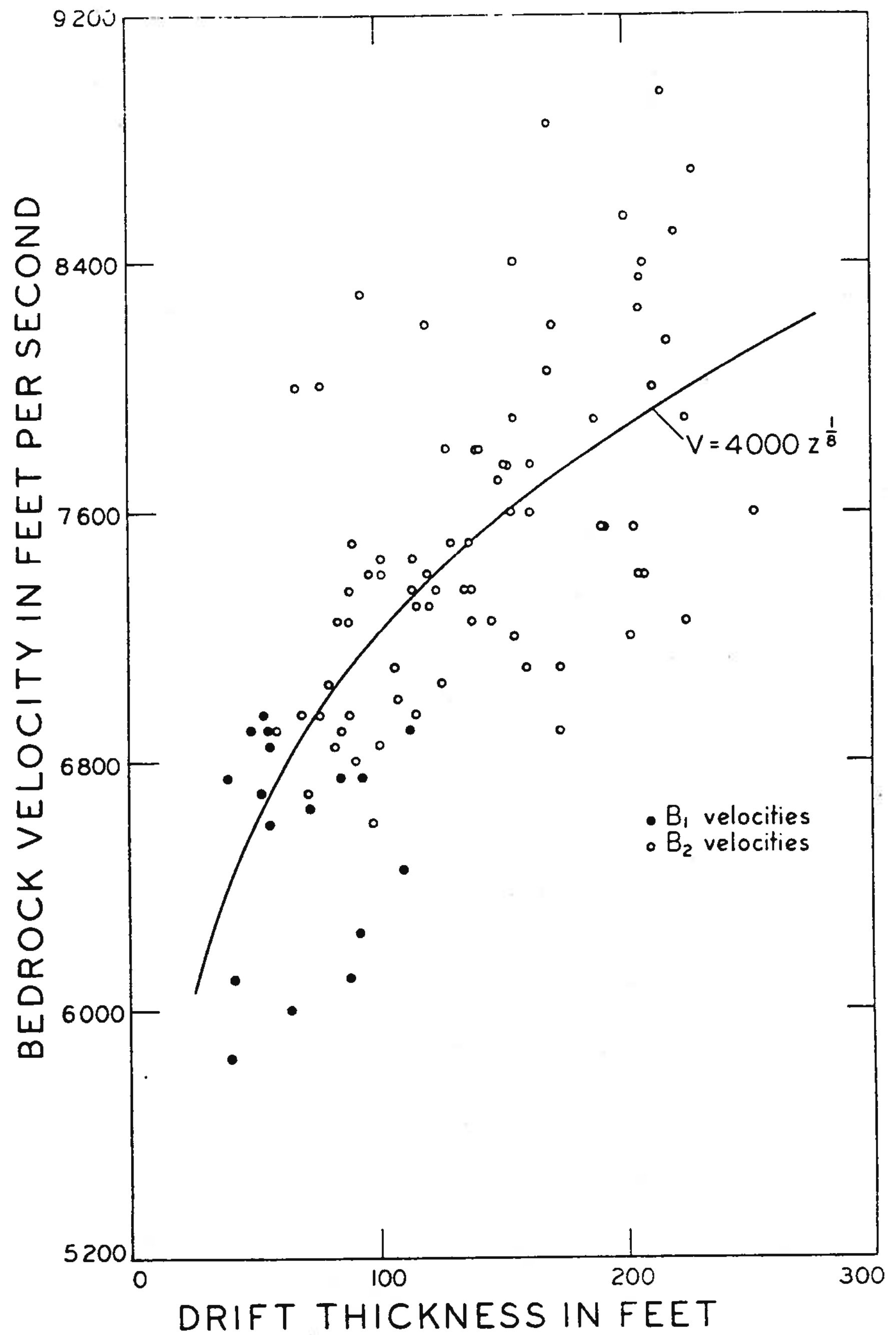


Figure 12. Relation between bedrock velocity and drift thickness.

the Wenner configuration in which the electrodes are evenly spaced along a line, the outer pair being the current electrodes and the inner the potential electrodes. The distance between two adjacent electrodes is called the electrode- or a-spacing and is considered to be roughly equivalent to the depth of penetration. Data may be obtained either by keeping the central point of the electrode array fixed and taking current and potential readings for a series of electrode spacings, or only one or two spacings may be used at any one location and a larger area covered. The former method, called depth-profiling or depth-sounding, is most suited for quantitative treatment of the data, while the latter, known as the step-traverse or lateral-profiling method, may be used to delineate the areas in which the anomalies occur. Depth-profiling has been the one most generally employed by the Research Council.

Resistivity Results

Resistivity data, obtained mainly in the area included between the Fourth and Fifth Meridians, from townships 1 to 60, can be correlated with known subsurface materials as shown in table 8.

Table 8. Resistivities of Subsurface Materials, Alberta

Material	Resistivity (ohm-meters)
Clayey tills	10- 20
Sandy tills	20- 40
Dry surface sands	70- 150
Near-surface gravels	500-2,000
Upper Cretaceous and Tertiary shales	1- 30

Typical Alberta depth-profiles are shown in figures 13A to 13D. The most usual case is represented by figure 13A, in which there is a gradual decrease in apparent resistivity from till to bedrock values as the electrode spacing is increased. Figure 13B represents an identical geologic situation in which till and bedrock resistivities are equal, or nearly so. Both these cases are ambiguous in that sand or gravel may actually be present, but in insufficient quantity or at too great a depth for detection. If, on the other hand, a depth-profile similar to that in figure 13C is obtained, showing an increase in apparent resistivity with electrode spacing for at least part of the profile, the presence of sand or gravel beneath the surface is definitely indicated. If the deposit is substantial, with only a superficial drift cover, a depth-profile such as that in figure 13D is obtained. A depth-

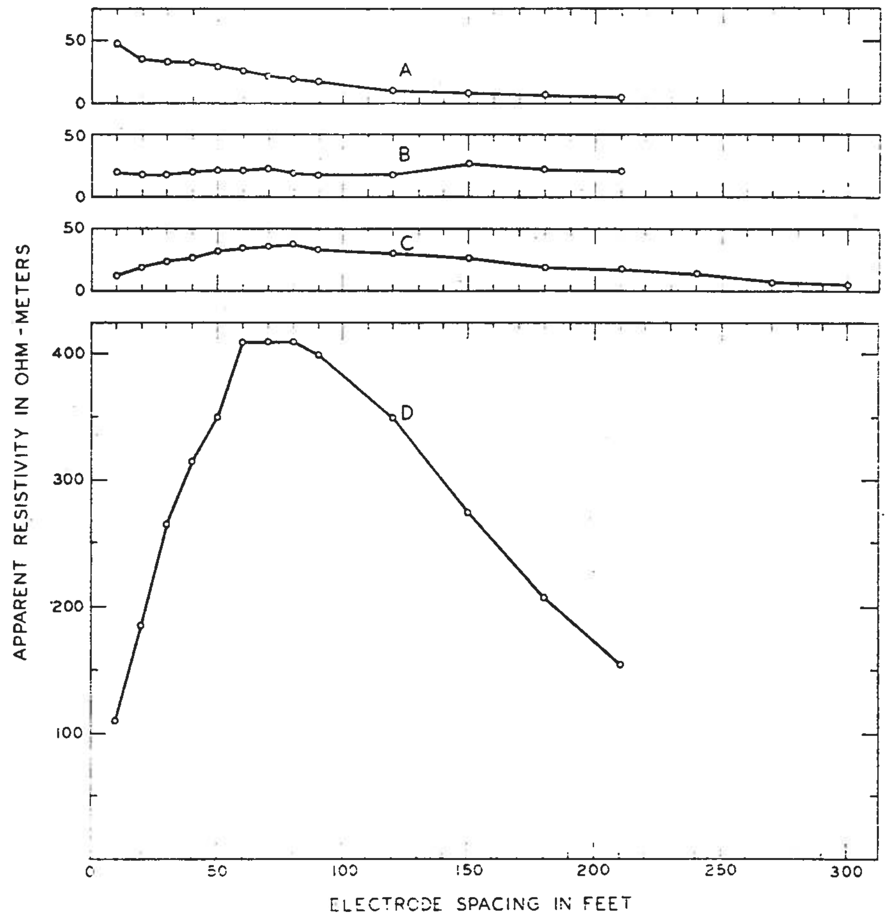


Figure 13. Resistivity depth-profiles.

profile of this type is, almost without exception, the result of surveying over recent alluvial gravel deposits.

Interpretation of resistivity results in Alberta has been almost exclusively of this qualitative nature. Various quantitative methods exist, many of which have been summarized by Roman (1952). Those which are the more sophisticated mathematically (Roman, 1959; Mooney and Wetzel, 1956) usually assume the simplest geologic conditions and consider only horizontal, homogeneous and isotropic layering. Alfano (1959), however, has recently applied rigorous mathematics to more complicated geologic situations. Attempts to analyze Alberta data quantitatively have met with very little success, but experiments are continuing. It is probable that the differences between the actual and the idealized geology are in large part responsible for the failure of the quantitative approach. In particular the assumption of homogeneous beds is probably invalid in many cases.

Efficiency of Earth-Resistivity Surveying

Table 8 shows that those Alberta earth materials that are potential aquifers have higher resistivities than those that are not. Resistivity surveying was originally employed by the Groundwater Division in the hope that this fundamental resistivity difference between aquifers and nonaquifers would provide a reliable method of detecting the former but, as has been remarked above, the masking effect of the overlying drift cover frequently renders an aquifer undetectable. This effect has already been noted for Hackett's results in Illinois (1956) and it has been possible to investigate it in a statistical manner in Alberta.

A number of resistivity surveys in Alberta have been run over buried preglacial channels containing appreciable amounts of sand and gravel. In some of these areas sufficient data from other sources have been obtained to allow the drawing of isopach maps both for the sand and gravel beds and for the overburden. With the aid of these maps estimates of the amount of overburden and the thickness of sand and gravel can be made for each resistivity station. If the former is denoted by d, and the latter by t, some statistics can be derived on the relation between success in detecting sand and gravel and the ratio $t/(t+d)$. This ratio will vary from unity for a sand or gravel bed at the surface to zero for a bed which is either infinitely thin or infinitely removed from the surface. Table 9 presents this statistical data for 89 stations where the subsurface conditions were known with reasonable accuracy. According to the table, Hackett's (1956) 40 feet of sand and gravel under 25 feet of drift have approximately a 50 per cent chance of being detected from the results of a single resistivity station. No attempt has been made to evaluate the results in terms of the resistivity contrast between drift and bedrock.

Table 9 will prove useful in a later section for evaluating the real costs of resistivity exploration.

Well-Logging

An excellent survey of logging techniques and their application is given by Jakosky (1957). Applications to groundwater exploration are described by Jones and Buford (1951) and Pryor (1956). The techniques utilized in the Research Council logger are resistivity, self-potential and gamma-ray measurements.

The basic principles of resistivity measurements are the same, whether these measurements are made on the surface or in a borehole, but the

Table 9. Success in Resistivity Exploration as a Function of Aquifer and Overburden Thickness

Ratio, aquifer thickness to total drift cover (t/t+d)	Total number of stations	Stations at which anomalies observed	Percentage success
0.00-0.19	22	5	23
0.20-0.39	36	9	25
0.40-0.59	14	7	50
0.60-0.79	12	6	50
0.80-1.00	5	4	80

practical details differ. A number of electrode arrangements are again possible. The electrodes may be either grounded at the earth's surface or located at points along a probe which is lowered into the borehole. The various electrode arrangements in common use differ both in electrode location and in spacing between the electrodes. The accuracy with which lithologic boundaries are located and their true resistivities measured is a function of electrode arrangement and of electrode spacing relative to bed thickness. Generally resistivity logs for a number of electrode arrangements are recorded simultaneously so that quantitative information can be derived for all the beds encountered.

As in the case of surface measurements, sands and gravels give resistivity highs and clays and shales give lows. Coal seams, which are normally undetectable in the Alberta plains by either seismic refraction or surface resistivity because they are usually only a few feet thick, also give resistivity highs. Coal seams make good aquifers if only small quantities of water are required and completion of a well in a coal seam is facilitated if the exact depth and thickness of the coal have been established by resistivity logging.

The self-potential measurement is a measurement of the naturally occurring voltage drop between a grounded electrode at the surface and an electrode on the probe. Usually the self-potential gives a nearly straight curve opposite shale beds with excursions toward more negative voltages when the probe passes permeable beds. This voltage change is caused mainly by an electrochemical phenomenon occurring at the contacts between the drilling mud and the connate water in the permeable beds and the adjacent shales. Differences in the concentrations of dissolved salts cause electromotive forces to arise and these are recorded by the self-potential log. Self-potential curves are usually combined on a single log with one or more resistivity curves and the combined information used to determine formation boundaries and porosities and the salinity of formation waters.

Gamma-ray logging shares with neutron logging, the other common type of radioactivity logging, the advantages of being able to produce usable logs in cased holes and in open holes containing no fluid. Casing acts as a short circuit for resistivity and self-potential logging. The gamma-ray probe consists of a scintillation counter to detect the weak natural radioactivity of the formations, and a photomultiplier tube and preamplifier to amplify the signal sufficiently for transmission to the surface. All formations are radioactive to a small degree and a change usually occurs at formation boundaries. Sandstones are usually low in radioactivity and clays and shales noticeably higher. A gamma-ray log gives a curve similar to the self-potential in that permeable beds are distinguished from impermeable.

Typical test-hole logs are shown in figure 9.

The resistivity electrode arrangement normally used in the Research Council logger is that commonly known as the single electrode, which is well suited for determining formation boundaries, but is inadequate for determining true resistivities. Because most of the logging done by the Groundwater Division has been for the purpose of accurately locating formation contacts, this electrode arrangement has, in many cases, provided a valuable supplement to the driller's log.

Economics of Geophysical Exploration for Groundwater

Apparent Costs

To the groundwater geologist, geophysical methods represent only an assortment of tools which he can employ in determining the nature of the subsurface. How these tools are used, and even whether they are used at all, is dependent on the amount of information they supply and the cost of obtaining it relative to the more conventional geologic methods of prospecting. In Alberta the only geologic method providing data comparable to that supplied by seismic refraction and earth resistivity is test drilling.

Shallow test-drilling in Alberta—to depths of 200 to 250 feet or less—can vary in cost from about \$0.50 to \$1.50 a foot. Costs per foot for deeper test holes increase with the footage. For \$0.50 a foot the geologist gets a minimum of information, often little more than the location of the drift-bedrock contact, whereas for \$1.50 a foot he gets a very detailed lithologic log. Such a detailed log, however, is probably only applicable in the immediate vicinity of the test hole so that an areal interpretation of the test-drilling data will, almost inevitably, lack the detail of the individual logs on which it is based. Nevertheless, such an interpretation will still result in a fairly comprehensive areal picture of the subsurface.

A single geophysical determination, in contrast to a single test hole, gives a picture of the average subsurface conditions in the vicinity of the measurement. In this picture, all fine detail is lost. In the seismic-refraction method, average velocities for drift and bedrock are obtained, along with an average depth to bedrock. In earth resistivity, results are of an even more qualitative nature. A seismic survey over an area gives somewhat less detailed information than a test-drilling program that covers the same area and penetrates to the same depth. This lack of detail is more pronounced in the case of earth resistivity.

Geophysical methods, however, do have one advantage over test drilling in that their costs do not depend, in the same direct way, on the depth of penetration. In the case of seismic refraction, the same physical arrangement of the shot holes, cable and geophones can be used to detect bedrock at depths up to 250 or 300 feet, so that the cost of a seismic depth-determination is independent of the depth of penetration until these depths are reached. Greater depths require special spreads which approximately double the cost per determination. Earth-resistivity costs per station depend to a certain extent on the required depth of penetration, but the main expense is in setting up the stations. Because geophysical costs are largely independent of depth, there is a depth, which depends upon the geophysical method and the circumstances under which it is used, beyond which it is cheaper to use the geophysical method than the drill. This depth, for convenience, is called herein the "critical depth". Above the critical depth it is cheaper to drill than to use the geophysical method in question. Below, it is cheaper to use geophysics.

It is apparent from the foregoing discussion of drilling costs that, if it can be safely assumed that the geologist will be satisfied with a bare minimum of subsurface information from his test-drilling program, the calculation of the critical depth should be based on drilling costs of $0.50 per foot. It is unlikely, however, that the geologist with a drilling rig at his disposal will be so easily satisfied and it is generally true, in Alberta at least, that the average test-drilling program will cost about $1.00 per foot. For this reason it is probably more realistic to base the critical-depth computation on drilling costs of $1.00 per foot. The critical depth will, of course, be doubled if the drilling costs are halved.

Table 10 shows the calculation of the critical depth for the seismic and resistivity methods. Three sets of calculations are shown for the seismic method. The difference in the critical-depth figure between the

Table 10. Apparent Costs of Geophysical Exploration

Type of Survey	Total expenses (dollars)	Number of depth determinations or resistivity profiles	Cost per determination (dollars)	Critical depth (feet)
Complete seismic season (1958)	17,210	155	111	111
Single seismic survey (1958)	3,144	41	77	77
Seismic optimum	1,145	25	46	46
Resistivity	3,300	198	17	17

first two columns shows the effect of doing a large number of surveys covering, in the main, fairly small areas and involving considerable travel between areas. The critical depth for the season is about 50 per cent greater than that for the single survey. The third column represents the sort of results that might be attained if there were no down time due to weather, or servicing and maintenance of cables and equipment, and if trips to pick up explosives could be reduced to a bare minimum. Because a number of the factors involved are beyond the geophysicist's control, it is improbable that this level of efficiency will be approached except over short periods, but the table does indicate that the factors that are controllable at the planning stage can be utilized to effect a considerable reduction in costs.

Real Costs

The data in table 10 on apparent costs of seismic exploration may be taken, without any serious error being introduced, to represent the real costs of seismic exploration. Failure to obtain an estimate of the drift thickness occurs only when subsurface layering deviates in some pronounced manner from the simple mathematical model on which computations are based. Such deviations are rare in Alberta, affecting perhaps one per cent of the seismic results.

To determine the true costs of resistivity exploration, however, the data in table 10 must be considered in conjunction with those in table 9 concerning the effect of geologic factors on success in resistivity prospecting. The success percentages recorded in table 9 may be regarded as the probabilities of discovering the corresponding aquifers from the results of a single resistivity station, assuming that the station is located over the aquifer.

Let p_1 = probability of discovering an aquifer from the results of a single resistivity station located over the aquifer.

p_n = probability of discovering an aquifer from the results of n resistivity stations located over the aquifer.

It may be shown that:

$$p_n = p_1 [1 + (1 - p_1) + (1 - p_1)^2 + \cdots + (1 - p_1)^{n-1}] \quad (3)$$

Making use of the algebraic expression for the sum of a geometric series and solving for n:

$$n = \log (1 - p_n)/\log (1 - p_1) \quad (4)$$

Tables 9 and 10 indicate that, although the cost of a single resistivity station is low, the probability of finding an aquifer from a single station is also low. If earth-resistivity exploration is to have a reasonable chance of success, the density of stations must be increased considerably as compared to the density of seismic spreads or test holes. The required density n is a function of the desired probability of success p_n and of the thickness and anticipated depth of burial, which determines the value assumed for p_1. The results in table 11 giving true resistivity costs are based on the data in tables 9 and 10. The table lists the calculated station densities, the costs of these stations and the corresponding critical depths, for a number of different geologic conditions and desired probabilities of success.

Table 11 can be used to estimate the advisability of employing resistivity prospecting in searching for sand and gravel deposits. A few examples will serve to illustrate. Let it be assumed that the average thickness of the deposits in a buried preglacial channel is known to be about 200 feet and that the channel contains sand and gravel beds which are some 40 feet thick. A choice is to be made between exploring the channel with a number of test holes or surveying it with earth resistivity. From table 11 it can be seen that if 11 resistivity stations are planned for each test hole thought necessary, there is only one chance in twenty that the sand and gravel beds will be missed. The cost of 11 stations will buy about 184 feet of drilling, which is just sufficient to reach the sand and gravel deposits. The real cost of resistivity prospecting will be virtually the same as that of test drilling in this case, and there will be a slight chance that the former will be unsuccessful.

In the case of another buried channel, also containing about 200 feet of deposits but with 30 feet or less of sand and gravel, the probability of success for a single station will be appreciably less than 0.24, and the required station density will be such that it will almost certainly be preferable to resort to a test-drilling program. If, on the other hand, drift thickness in the channel is only 100 feet and the sand and gravel deposits are about 20 feet thick, so that the value of t/(t+d) is the same as in the first case considered, the 184 feet of drilling which the cost of 11 resistivity stations will buy will be more than sufficient to reach and completely penetrate the sand and gravel beds.

In general, costs of resistivity surveying are found to increase as drift thickness increases and as sand or gravel thickness decreases, in agreement with previous qualitative predictions. The station densities required to give the method a reasonable efficiency cause the costs to rise to such an extent that, except under very favorable conditions, it can no longer be regarded as a relatively cheap method of geophysical exploration. Furthermore, the method is either uneconomical or inefficient, or both, if used in prospecting for thin aquifers.

If the subsurface geologic picture were as uncomplicated as the mathematical models on which quantitative analysis is usually based, a statistical approach to the economics of resistivity prospecting would be unwarranted. The subsurface would then consist, at least over areas with dimensions large compared to those of the electrode configurations, of a number of horizontal, homogeneous layers of constant thickness. If, in any given area, one of the layers should be an aquifer, it would be theoretically possible to determine unequivocally whether or not it could be detected in a resistivity survey. Detectability would depend upon the thickness and resistivity of the aquifer and the thicknesses and resistivities of the overlying layers as well as on the sensitivity of the instrument.

Because the actual geologic situation commonly departs radically from the idealized mathematical model with respect to any or all of the conditions mentioned above, and because resistivity contrasts can also vary between areas, though the mathematical model may be satisfied, an element of unpredictability is introduced into any summary of success in resistivity exploration, such as that given in table 9. The data given in table 9 are characteristic of resistivity surveys from a few selected areas in Alberta and would not necessarily be duplicated if a similar investigation were carried out in other areas of the province. However, it is not unreasonable to assume that table 9, and table 11 which is based on the data in table 9, are generally valid over the plains area of Alberta and, in any case, use of table 11 in assessing the costs of resistivity exploration is more enlightening than consideration of resistivity costs per station alone.

Conclusions

Of the three geophysical techniques employed to any great extent by the Research Council to date in the search for groundwater in the

Table 11. True Resistivity Costs, as a Function of Geologic Conditions and Probability of Success

Ratio, aquifer thickness to total drift cover (t/t+d)	Probability of success for single station (p_1)	Desired probability of success p_n=0.80			Desired probability of success p_n=0.95			Desired probability of success p_n=0.99		
		Station density (n)	Cost of n stations (dollars)	Critical depth (feet)	Station density (n)	Cost of n stations (dollars)	Critical depth (feet)	Station density (n)	Cost of n stations (dollars)	Critical depth (feet)
0.20	0.24	5.9	98	98	10.9	182	182	16.8	280	280
0.40	0.38	3.4	57	57	6.2	103	103	9.6	160	160
0.60	0.50	2.3	38	38	4.3	72	72	6.6	110	110
0.80	0.65	1.3	22	22	2.9	48	48	4.4	73	73
1.00	1.00	1.0	17	17	1.0	17	17	1.0	17	17

province, two — seismic and earth-resistivity prospecting — have been utilized to the extent that definite ideas can be formed of the advantages they offer and their economy of operation. Well-logging, on the other hand, has been in use for only a short period and an accurate evaluation of its potentialities cannot yet be made.

Seismic exploration has proven almost 100 per cent reliable in the determination of drift thickness and has located, or assisted in the location of, a number of buried preglacial channels. With efficient planning of the seismic survey, it is economically competitive with test-hole drilling if drift thickness exceeds about 80 feet. If, for some season, it is impossible to plan a survey in the most efficient manner, seismic exploration is still competitive for drift thicknesses of 110 feet and more.

Earth resistivity, by contrast, is probably uneconomic when compared to test drilling except when the aquifer sought is fairly thick or close to the surface. Variations in subsurface conditions introduce some uncertainty into the search for an aquifer by resistivity methods. To reduce the uncertainty the geophysicist must increase the areal density of his resistivity determinations as compared to the density required by other methods. Depending on geologic conditions, the required density may be so great as to render exploration costs for this method prohibitive.

BEDROCK TOPOGRAPHY EDMONTON—RED DEER MAP-AREA, ALBERTA

by
R. N. Farvolden

Introduction

The accompanying map (Fig. 14) was compiled from all information available pertaining to the bedrock surface in the Edmonton and Red Deer map-area. This area lies between 52° and 54° north latitude and between 112° and 114° west longitude and includes townships 35 to 58, ranges 15 to 28, west of the Fourth Meridian*. The main reason for undertaking this study was to determine the locations of bedrock channels, which are often important in groundwater studies. The sharp gullies that have been eroded into the bedrock surface during or after the retreat of the last glacier from the area are not important in groundwater studies in this area and will not be discussed further. The features of the bedrock surface that are discussed are thought to be preglacial in origin. Names have been assigned to the main features of the bedrock surface for convenience in describing the bedrock topography.

Acknowledgments

Almost all the subsurface data used in this report were contributed by water-well contractors and firms engaged in petroleum exploration in Alberta. The assistance of the Alberta Water Well Drilling Association and Texaco Exploration Company is especially appreciated.

Most of the ideas presented here are the result of discussions between the author and his colleagues at the Research Council of Alberta and the benefits derived from these discussions are gratefully acknowledged. Dr. A. MacS. Stalker of the Geological Survey of Canada very kindly made available a preliminary map of the surficial geology of the Red Deer-Stettler area on which the course of the Red Deer Bedrock Channel from Red Deer to Driedmeat Lake can be easily followed. This map has since been published (Stalker, 1960).

Bedrock Features

Upland Areas

The bedrock surface of the upland area of the Edmonton and Red Deer map-area may be divided into five main regions (Fig. 14). These are:

* Unless otherwise stated, all locations are west of the Fourth Meridian (110° west longitude).

(1) the Joffre Hills,

(2) the Edmonton Plain,

(3) the Mundare Plain,

(4) the Battle Plain, and

(5) the Endiang Plain.

The bedrock plains are separated from one another by low divides to which names have been given. Experience has shown that if the locations of these divides are known, the search for bedrock channels is simplified considerably. The physiographic boundary, unnamed, between the bedrock plains and the Joffre Hills is also indicated on figure 14, for convenience by the same symbol as that used for the divides.

The Joffre Hills comprise the southwest portion of the map-area and they coincide with that part of the map-area underlain by sandstone strata belonging to the basal portion of the Paskapoo Formation. These sandstones are massive and comparatively well indurated and they form the core of the Joffre Hills. The surface expression of the Joffre Hills forms an escarpment visible for some distance from the plains to the east. The escarpment rises from an elevation of 2,600 feet at its foot to over 3,400 feet at its highest point in township 37, ranges 25 and 26, southeast of the city of Red Deer. Local relief is nearly 700 feet and both major and minor divides are easy to distinguish. The interstream areas can be described as swell and swale topography in which the long, smooth slopes tend to become steep near the valleys.

The Edmonton Plain is bounded on the south by the Joffre Hills and the Kingman Divide and on the east by the Cooking Lake Divide. The north and west boundaries do not appear on this map. The Edmonton Plain is bisected by the North Saskatchewan Bedrock Channel. North of the channel the plain is relatively flat, and rises gradually from 2,000 feet in elevation in the northeast to over 2,400 feet in the hills in township 54, ranges 27 and 28. The gradual rise is broken by an escarpment in townships 55, 56, and 57, ranges 23 and 24, which is not too obvious on the map (Fig. 14) but which has a rather prominent surface expression. South of the North Saskatchewan Channel the surface slopes toward the channel from the Kingman and Cooking Lake Divides. The slopes are mostly gentle, being seldom steeper than 50 feet per mile. In the Cooking Lake Divide and the escarpment and hills mentioned above, the relief sometimes reaches 100 to 200 feet. For the most part, however, relief is of the order of 20 to 30 feet in the interstream areas.

The Mundare Plain is that portion of the map-area drained by the Vegreville Bedrock Channel and its tributaries. It is bounded on the west by the Cooking Lake Divide and on the south by the Kingman Divide. The north and east boundaries are outside of the area considered. The plain rises gradually from an elevation of 2,000 feet in the northeast to 2,300 feet just below the divide in the west, and there the gradients increase toward the crest of the divide area at an elevation of over 2,500 feet. Apart from the banks of the bedrock channels, the only steep gradients on the Mundare Plain are found at the juncture of the Cooking Lake and Kingman Divides, an area of low bedrock hills. For the most part, the Mundare Plain coincides closely with the present surface and is flat and featureless, but north of the Vegreville Channel (where borehole control is good) a number of sharp, narrow gullies are eroded into the surface of the plain. The origin of these gullies has not been established but some of them may be part of the bedrock-channel system and therefore pertinent to this discussion.

The Battle Plain lies east of the Joffre Hills between the Kingman Divide on the north and the Halkirk Divide on the south. The east boundary is beyond the map-area. The slope of the surface is largely controlled by the two major bedrock channels that cross the plain. At the eastern edge of the map-area the Battle Plain has an elevation of 2,200 feet and it rises to 2,500 feet in the northwest, 2,600 feet in the west, and 2,800 feet at the highest point in the Halkirk Divide. Local relief is seldom over 30 feet except near the channels. A system of gullies known to be of Pleistocene and Recent age has been eroded into the surface of the Battle Plain and is responsible for the irregular appearance of the surface.

Little is known of the Endiang Plain which lies to the south of the Halkirk Divide and east of the Joffre Hills, except that it is the northwest portion of a large bedrock plain drained by tributaries which join a major channel far to the south.

Bedrock Channels

Four major bedrock channels are present in the map-area. Two of these, the North Saskatchewan and Red Deer Channels, rise in the mountains to the west whereas the other two, the Vegreville and Buffalo Lake Channels, have their headwaters within the area considered.

The North Saskatchewan Channel is a consequent channel which crosses the northwest corner of the area. The valley is broad, the right bank rising gently to the south while the left bank is nearly cliff-like in places near the city of Edmonton. A large tributary, the Onoway Channel, enters the North Saskatchewan Channel from the north in township 53, range 25. Above the confluence with the Onoway Channel the North Saskatchewan Channel has a gradient of over 10 feet per mile, but below the confluence the gradient is about 5 feet per mile. There is evidence of a narrow gully being present at the bottom of the main channel. The channel might alternatively be shown to follow the present Sturgeon River valley as far as township 56, range 23, before turning southward into the North Saskatchewan Channel, but the available evidence favors the interpretation shown on the map.

The Onoway Channel drains a large area to the west of the map-area and is a major bedrock feature. Other tributaries to the North Saskatchewan Channel are not important but the present river meets and follows one of them for several miles just above the city of Edmonton.

The Vegreville Channel rises near the juncture of the Cooking Lake and Kingman Divides, extends nearly due east, and then near the town of Vegreville turns northeast to follow the regional consequent slope. The channel is narrow and has steep banks throughout its extent shown on this map. The two tributaries that enter the channel from the south have broad valleys with low, gently sloping sides but the remainder of the tributaries have valleys much like that of the main channel and are difficult to define. The gradient is high in the headwaters and in some of the tributaries but it is about 6 to 7 feet per mile for the main portion of the channel. Above Beaverhill Lake and below Vegreville there is evidence of the presence of the bedrock channel on the present land surface but between these two points there is no such evidence.

The Red Deer Bedrock Channel enters the map-area west of the town of Innisfail (Tp. 35, R. 28). It cuts through the Joffre Hills in a wide channel which has nearly a consequent direction. South of Wetaskiwin (Tp. 46, R. 24) the channel makes a wide turn to the east and follows a course slightly south of east across the Battle Plain. All subsurface information suggests that the floor of the valley is flat. The gradient is from 8 to 10 feet per mile, except for the portion east of range 18, where the gradient is less than 5 feet per mile. From Menaik (Tp. 43, R. 25) to Driedmeat Lake (Tp. 45, R. 20) the presence of the channel is not obvious from the present topography. This is also the case along the lower part of the valley in the vicinity of Strome (Tp. 44, R. 20). Throughout the remainder of its course the valley is coincident with topographic lows in the present land surface. Driedmeat Hill, on the west side of township 45, range 19, is a bedrock high capped by Saskatchewan sands and gravels. This feature must have been an island in the middle of the valley. A similar feature may be present in township 45, ranges 22 and 23, but in this case there is no closure of contours because the contour interval is too large.

The Buffalo Lake Channel rises in the Joffre Hills and before it reaches the Battle Plain it is a major feature in the bedrock topography. This rapid development in size is caused by the fact that the tributaries that rise in the Joffre Hills are well developed and have steep gradients.

Where the channel enters the Battle Plain it widens, and the form of the contours suggests that the broad, relatively flat valley floor has a narrow gully incised into it. The gradient is about 6 feet per mile. Southwest of Buffalo Lake, the channel follows a depression in the present surface, but east of Buffalo Lake its course is not evident from the present topography. The Buffalo Lake Bedrock Channel is tributary to the Red Deer Bedrock Channel, the confluence being about 20 miles east of the map-area.

South of the Halkirk Divide the headwaters of several bedrock channels are present. As in the headwaters portions of the other bedrock channels, the present surface is nearly coincident with the bedrock surface. In the vicinity of Sullivan Lake in townships 35 and 36, range 15, bedrock is exposed over a wide area.

Groundwater Resources

With minor exceptions, the only possible source of large supplies of groundwater in the Edmonton-Red Deer area is in the sand and gravel deposits that occur in the bedrock channels. In many places these deposits are well sorted and permeable and make excellent aquifers. In others the amount and nature of the matrix material render the deposits impermeable, and in still other portions of the channels, sorted, granular sediments are absent altogether.

The North Saskatchewan Channel contains large deposits of gravel below the confluence with the Onoway Channel but only sand has been reported in drill holes above this point. The gravel deposits are up to 40 feet in thickness and are reportedly quite permeable in some places. The Onoway Channel also carries considerable gravel but bedrock is close to the surface along the lower reaches and the gravel may not be saturated in some places. Almost anywhere along the North Saskatchewan Channel or the tributaries mentioned, it is reasonable to explore for groundwater where quantities of up to one million gpd (gallons per day) are required.

There is no obvious source of gravels for the Vegreville Channel other than Tertiary gravels that may have been present at one time on the highlands near the channel or on the hills in the headwaters area. However, gravel has been reported in many places and successful wells have been completed in the gravel. Near Tofield, in township 51, range 19, the narrow channel contains thick gravel beds and northwest of Vegreville in township 52, range 15, a well with a capacity of 100 gpm (gallons per minute) was used to supply the town for some years. It is quite likely that near these and other localities along this channel, well fields capable of producing one-half million gpd can be developed. Sand and gravel deposits have been reported along the tributaries but they are not expected to be capable of producing large quantities of water.

The Red Deer Channel apparently contains sand and gravel deposits of varying permeability throughout its course on the map-area. Exploration anywhere along the channel for water supplies of up to one million gpd appears justified. Little is known of the tributary that enters the channel from the west in township 45, range 21, but in the lower end at least, no gravels are expected to occur.

The sediments of the Buffalo Lake Channel are much finer than those of the Red Deer Channel. This is not surprising for, as in the case of the Vegreville Channel, there is no source of coarse material available except Tertiary gravel deposits of an earlier erosion cycle. Reports of gravel in the channel are not frequent and test drilling at the east end of Buffalo Lake in township 41, range 20, encountered no sorted granular material other than fine sand. Reports of sand in the channel are prevalent, and where supplies of water of up to one-half million gpd are required, exploration of this channel is justified. The tributaries that rise in the Joffre Hills carry thick sand and gravel deposits in places and should not be ignored in the search for water in the area.

The channels draining south from the Halkirk Divide do not contain granular water-bearing sediments because they have never been occupied by aggrading streams, nor is there any source of coarse granular material.

BEDROCK CHANNELS OF SOUTHERN ALBERTA

by
R. N. Farvolden

Introduction

Over much of the glaciated area of North America the stream valleys that existed prior to glaciation are now partially or completely buried by glacial drift. The channels of these earlier drainage systems commonly contain sand and gravel deposits which originated in the same manner as sand and gravel bars along present-day rivers. These deposits are for the most part exposed only where recent erosion has removed the drift cover. Where extensive sand and gravel deposits are well sorted and occur below the water table, they are excellent aquifers. In an area such as southern Alberta, underlain by shale and impermeable sandstone, the only aquifers capable of yielding large supplies of groundwater, apart from gravel bars along present-day rivers, are the sand and gravel deposits associated with bedrock channels. In present interstream areas these aquifers may be the only possible source of large supplies of water. In semiarid, southern Alberta these aquifers have a special significance for surface supplies of water are often difficult to develop because all but the main trunks of the larger rivers are intermittent streams.

The locations of some of the bedrock channels are known from surface and subsurface data and this report is an attempt to show the pattern of the bedrock channel systems.

Purpose and Scope of the Report

This report presents the results of observations carried out during four years of field work in Alberta. The logs of approximately 8,000 shallow borings have been tabulated and used in compiling the map of bedrock channels. Unfortunately, about 75 per cent of the logs are from borings in the Edmonton and Red Deer areas (Farvolden, 1963a) and only 25 per cent are from the rest of the area. Nevertheless, it is believed that sufficient information has been collected to allow mapping of the locations, pattern and gradients of the major bedrock channels. In some places the exact positions of the channels are known, but in others it is likely that errors of two or three miles, or more, will be revealed by future work.

The map accompanying this report (Fig. 15) has been compiled for use in the exploration and development of groundwater resources in Alberta. It is hoped that by mapping, in a general way, the divide areas and the bedrock channels the development of this type of aquifer will be encouraged.

The pattern of the bedrock channels and the relation of the bedrock surface to the present-day surface is used as evidence to explain the erosional history of the Alberta plains.

Description of the Area and Physiography

The area included in this report is that part of the plains region of Alberta lying east of 115° west longitude and south of 54° north latitude. The Rocky Mountains and the foothills belt form the southwestern boundary of the area but farther north the western edge of the map-area is well within the plains region because of the northwesterly strike of the mountain front.

The surface slopes from an elevation of about 4,000 feet in the plains region of southwestern Alberta to about 1,900 feet at Frog Lake in township 57, range 3, west of the Fourth Meridian. The surface is rather flat and featureless except for several distinctive upland areas, some of which rise hundreds of feet above the surrounding plains, and for the valleys of present-day streams which are deeply incised into the plains surface. The area is almost completely covered by a mantle of glacial drift which is 30 to 100 feet thick over most of the area. Morainal deposits and abandoned meltwater channels relieve the otherwise monotonous plains topography.

The annual precipitation for the area ranges from less than 14 inches in the south to over 18 inches in the north. The January mean temperature ranges from over 15°F in the south to 0°F in the north and the July mean temperature is about 65°F.

Acknowledgments

As in the case of the preceding report (Farvolden, 1963a), water-well contractors and the Alberta Water Well Drilling Association were responsible for important contributions to the available data. Petroleum exploration companies, in particular the Texaco Exploration Company, also assisted materially. The suggestions and comments of colleagues at the Research Council are again gratefully acknowledged.

The Bedrock Surface

The present-day surface is nearly coincident with the bedrock surface over most of the southern half of Alberta. The major upland areas are underlain by bedrock and the major bedrock channels are, for the most part, coincident with broad depressions in the present land surface. A large number of narrow, glacial meltwater channels have been incised into the surficial deposits and in many cases into the underlying bedrock. The majority of these channels are now abandoned except for intermittent streams, but some of them still carry permanent streams. The map of the

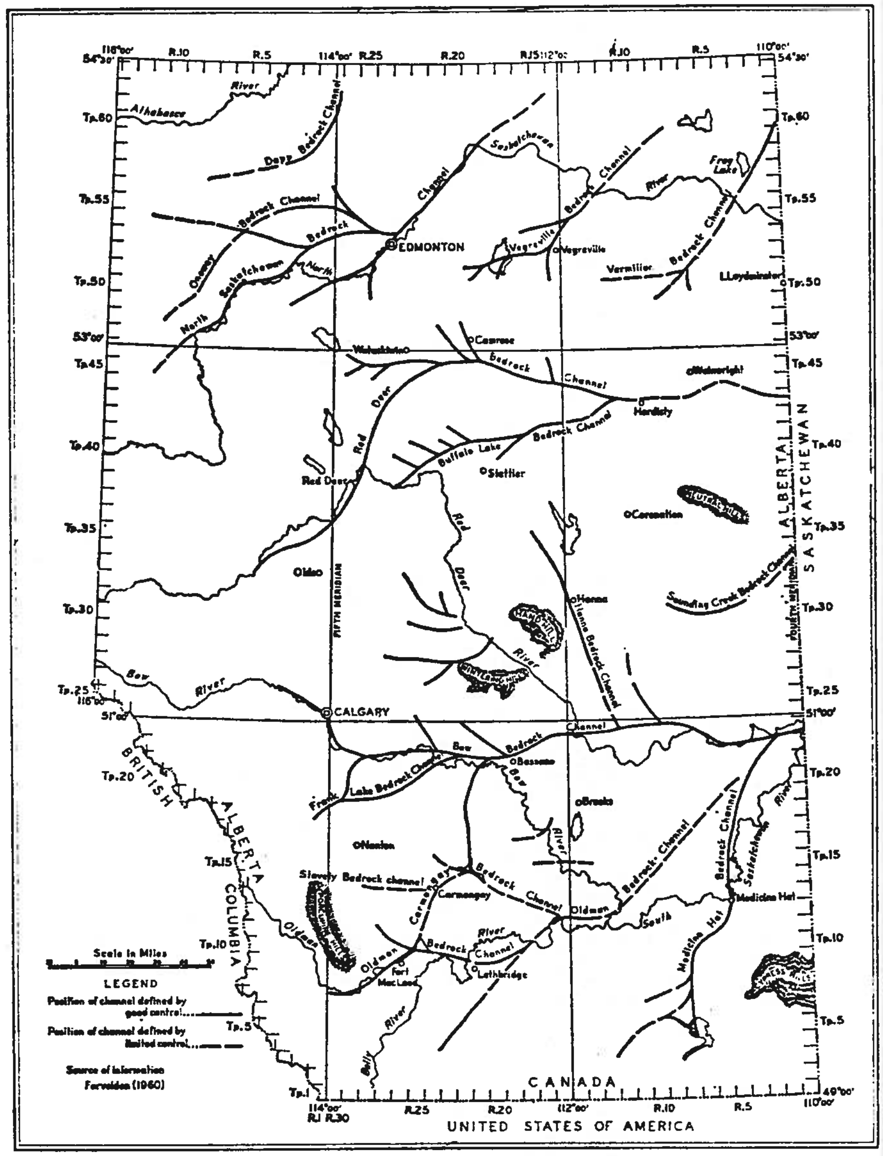


Figure 15. Bedrock channels of southern Alberta.

bedrock channels (Fig. 15) does not show these features for it is presented as a map of the preglacial drainage system, or perhaps more correctly, as a map of the drainage system that existed prior to the last glaciation of the area.

Development of the Bedrock Surface

Since the last uplift of the Rocky Mountains, probably in Oligocene time, the history of the western plains has been one of degradation although, during two intervals at least, degrading was halted and aggrading on a large scale took place. It seems likely that in each case aggraded material was deposited on a newly formed peneplain. The older of these two surfaces is the Cypress Plain, and the younger is the Flaxville Plain (Warren, 1939). Both of these surfaces have been nearly removed by erosion and only remnants of them rise above the present prairie level.

Erosion has been more or less continuous since about the end of Miocene time and, because there is no evidence of major post-Miocene tectonic movement in Alberta, the individual valleys must have had the opportunity to approach maturity or possibly old age. The tendency of the bedrock channels to occupy broad lowland areas between widely separated, low and rounded upland areas lends support to the idea of a mature plain drained by mature streams. There seems to be no reason to suspect that the drainage systems were not completely integrated and dendritic, as suggested by Lobeck (1939, p. 484), for the underlying rocks are flat-lying shales and soft sandstones.

The gradients from the interstream uplands toward the bedrock channels appear to be relatively even. If erosion took place by scarp retreat, it progressed to a degree that the valleys were several miles wide and the hills in interstream areas became quite rounded. The net effect is the same as that to be expected from sheet erosion.

Most major rivers in southern Alberta occupy ancient bedrock channels along some portion of their course, and here it is common to find the stream channel to be alternately incised into bedrock and then underlain by alluvium for short distances (Fig. 16). The outcrop pattern and the occurrence of alluvium below river level is obviously controlled by the relation between the location of the present channel and that of the bedrock channel. In some instances 50 feet of alluvium have been found below river level but this likely means that the channel had a different course and a lower local base level at one time than it has now. An example of this is found in the Red Deer River valley at Red Deer. It is suggested that the long period of uplift and erosion that began in middle Tertiary time has not changed significantly, and the regional base level of the streams of southern Alberta is as low or lower than it has ever been since the uplift of the Rocky Mountains.

Several of the bedrock channels have been mapped in detail where sufficient subsurface control is available. Where this has been possible, the gradient of the bedrock channel is approximately that of a similarly located

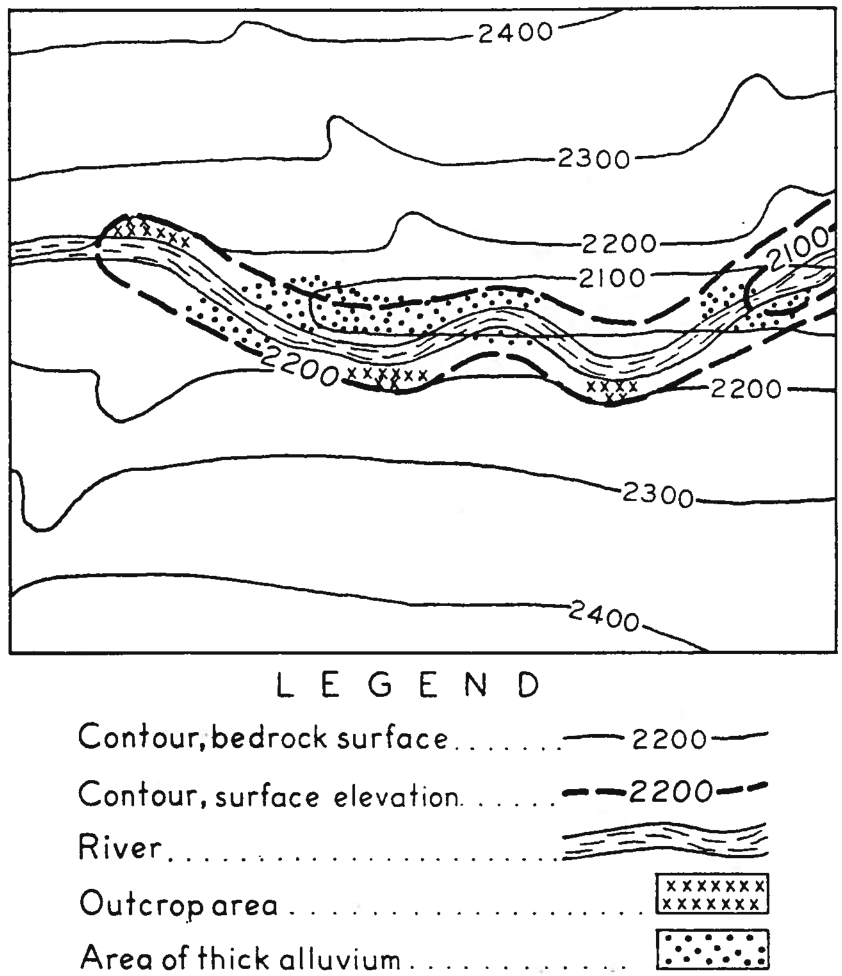


Figure 16. Present-day river in a bedrock channel.

valley today. This condition is postulated to hold throughout the area and it is used to help determine possible connections between segments of bedrock channels. The uniformity of the gradients of the bedrock channels is strong evidence that there has been little or no tilting of the land mass since preglacial time.

The block diagrams (Figs. 17A to 17G) are an attempt to illustrate a possible sequence of erosion and uplift that can be used as a working hypothesis to explain the features of the present and past plains surfaces and their drainage systems. Figure 17A represents the Cypress Plain. Remnants of this plain are capped with gravel strata for which there is indirect paleontological evidence to indicate an Oligocene age (Warren, 1939). Deposition of the gravels may have been contemporaneous with the last major uplift of the Rocky Mountains. Figures 17B, 17C, and 17D illustrate that, although the land mass continued to rise, the drainage systems and land surface developed to maturity and possibly old age. This cycle must have been repeated several times for the remnants of different erosion surfaces are found throughout Alberta (Warren, 1939). Each time the land was uplifted and another cycle began the streams would continue to occupy the same channel, because that would remain the lowest ground. Thus, the locations of the main features of the present bedrock topography, particularly the main divide or upland areas, and the stream channels, were determined at an early stage in the history of the plains.

Some evidence indicates that, in places, the divides have been breached by headward erosion and that stream piracy occurred. This has caused major changes in the pattern of the bedrock channels but not in the overall bedrock topography. As the individual drainage systems were developing, the controls on local base level, such as lakes and waterfalls, were removed. The gradients of the headwater streams would thereby be alternately decreased and increased, in turn causing changes in the rates of headward erosion in the divide regions. Then, if the relations of elevations and gradients between the streams on either side of the divide were favorable, stream piracy could result. The whole pattern of the drainage system might thus be changed. Stream piracy is a common phenomenon in nature and there is some evidence to indicate it was a factor in determining the pattern of bedrock channels in Alberta. If it is assumed that the bedrock topography is of preglacial origin, stream piracy is the most likely explanation for the apparent downstream bifurcation of several bedrock channels.

In several bedrock channels for which there is good subsurface control, there appears to be a deep, narrow gully at the bottom of the channel. This is illustrated in figure 17E. It is interpreted as evidence of a relatively rapid uplift of the land mass similar to minor uplifts that had previously started new cycles of erosion, as illustrated in figure 17B. In this instance, however, the cycle was interrupted by continental glaciation. Figure 17F shows that, although the drift is not nearly thick enough to obscure the major features of the bedrock topography, segments of some bedrock channels are completely buried by the drift. Most bedrock channels are still coincident with topographic depressions and are now occupied by streams or lakes. In some cases, however, the drainage was diverted by the glacier and the bedrock channels were buried in drift and abandoned (Fig. 17G).

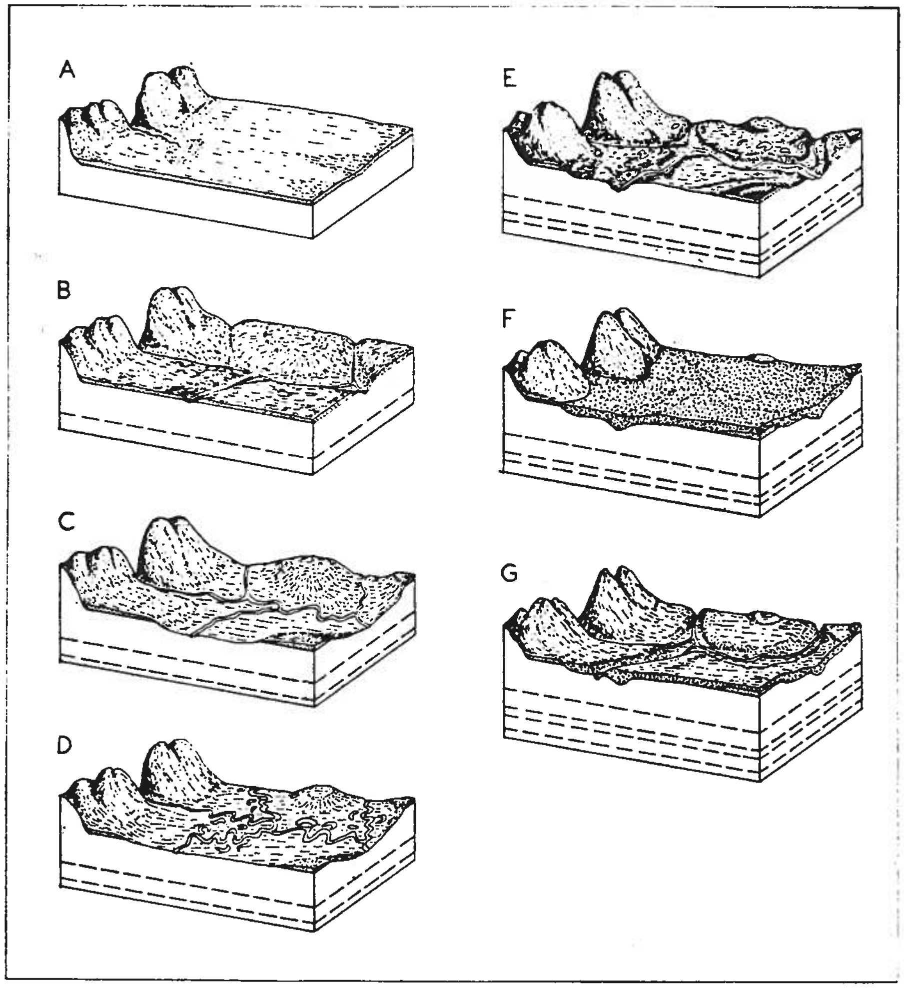


Figure 17. Evolution of a system of bedrock channels.

A. Deposition surface in Oligocene time.

B. Streams dissect the plain as the land is uplifted.

C. Uplift continues but not fast enough to prevent maturing of the drainage system and land surface.

D. Uplift is very slow and rivers approach old age.

E. Relatively rapid uplift causes rejuvenation of the drainage systems. This sequence may have been repeated several times.

F. The surface is overrun by continental glaciers leaving a mantle of drift over all but the highest hills.

G. The drainage system developing since the retreat of the glaciers follows the preglacial system in some areas, but deviates from it in others. A slight uplift has left the land surface higher than at any time since the Oligocene.

The Bedrock Channel Systems

The North Saskatchewan Bedrock Channel System

The North Saskatchewan Bedrock Channel crosses the northwest corner of the map-area following a northeast direction. The channel rises in the mountains but is not shown above Tp. 47, R. 9, W. 5th Mer. on this map. From this point downstream to Tp. 51, R. 3, W. 5th Mer., the present North Saskatchewan River follows the ancient bedrock channel. Then the river leaves the bedrock channel which forms a broad arc that curves to the north through Big Lake (Tp. 53, R. 26, W. 4th Mer.) to the city of Edmonton. Here the present river reoccupies the bedrock channel and continues to do so downstream to Tp. 58, R. 19, W. 4th Mer. The abandoned portion of the bedrock channel is, in places, partly obscured by a thick drift cover and in other places is easily detectable from the present topography. East of Tp. 58, R. 19, W. 4th Mer. the river likely leaves the bedrock channel once more but there is no proof of this supposition at present.

A tributary bedrock channel that rises in the bedrock uplands area around Tp. 49, R. 3, W. 5th Mer. is occupied by the present North Saskatchewan River from the Fifth Meridian to the city of Edmonton. A major channel joins the North Saskatchewan Bedrock Channel in Tp. 52, R. 2, W. 5th Mer. This channel has been mapped mainly on the basis of present topography and its genetic relation to the main channel is not known.

The Onoway Bedrock Channel swings in a wide arc from Tp. 50, R. 8, W. 5th Mer. around Wabamun Lake (Tp. 53, R. 5, W. 5th Mer.) and joins the North Saskatchewan Bedrock Channel at the east end of Big Lake. The upper portion of this channel as shown on figure 15 is occupied by the Pembina River, the valley of which changes markedly where it leaves the bedrock channel. The lower portion of the Onoway Bedrock Channel is well known from subsurface data and the middle reaches coincide with a depression on the present land surface. The relations between this channel and the North Saskatchewan Bedrock Channel system suggest that the Onoway Bedrock Channel may be an early stream course, abandoned due to stream piracy. It is also possible that this bedrock channel has been formed through some process associated with glaciation and should therefore have been omitted from this map.

The Pembina River in Tp. 60, R. 1, W. 5th Mer. occupies the Dapp Bedrock Channel, the tentative location of which is shown, although adequate subsurface data to substantiate this interpretation are not available.

The Vegreville Bedrock Channel System

The Vegreville Bedrock Channel rises in the plains area, in the upland west of Beaverhill Lake (Tp. 51, R. 18, W. 4th Mer.) and trends northeast. The portion of the channel west of Tp. 54, R. 12, W. 4th Mer. is well established by subsurface data and is a deep, narrow gully in the bedrock surface, in places completely filled with glacial drift. The northern half of the bedrock channel has been located on the basis of the present topography and scattered well logs.

The Vermilion Bedrock Channel is not well established by subsurface data, except in the headwaters region near Birch Lake (Tp. 50, R. 11, W. 4th Mer.). However, scattered well logs indicate a headwater tributary pattern as shown on the map, and below the town of Vermilion (Tp. 50, R. 6, W. 4th Mer.) the present Vermilion River follows the old valley for a short distance. Near Frog Lake (Tp. 57, R. 3, W. 4th Mer.) the presence of the channel has been confirmed by subsurface data, and there is likely a junction with the Vegreville Bedrock Channel in the vicinity of Cold Lake (Tp. 63, R. 1, W. 4th Mer.), just outside the map-area.

The Red Deer Bedrock Channel System

The Red Deer Bedrock Channel completely traverses the central Alberta plains. The upper portion of the channel nearly coincides with the valley of the present Red Deer River but in Tp. 39, R. 27, W. 4th Mer. the river leaves the bedrock channel which, east of this locality, is partially buried by glacial drift. The upper Red Deer Bedrock Channel follows a fairly constant northeast direction for about 90 miles and then swings east in a wide arc. It is possible that at an earlier stage of development than that shown on the map, the floor of the upper Red Deer Bedrock Channel was slightly higher than it is now and the channel did not swing east as noted above, but maintained its northeasterly direction to join the North Saskatchewan Bedrock Channel near Edmonton. Stream piracy is a mechanism that may have caused the abandoning of this course. Subsequent erosion has nearly obliterated the former channel. A similar argument could be advanced in favor of an early connection between this bedrock channel and the Vegreville Bedrock Channel but the intervening bedrock topography is not as favorable, although the channel pattern seems suggestive.

West of Hardisty (Tp. 43, R. 9, W. 4th Mer.) the Red Deer Bedrock Channel is well defined by both surface and subsurface information but there is only scattered subsurface control with which to fix the position of the lower end of the channel. However, logs of wells in Tp. 43, R. 1, W. 4th Mer., the nature of the topographic low that runs through Reflex Lake and Manito Lake just east of the Alberta-Saskatchewan boundary, and the fact that the elevation of the floor of the bedrock at the boundary (1,825 feet) is close to that required by the gradient (3.5 feet per mile), indicate the location for the channel that is shown on the map is the best choice at present.

The Buffalo Lake Bedrock Channel rises in the highland area east of Red Deer (Tp. 38, R. 27, W. 4th Mer.) and rapidly becomes a major channel because it is joined by several short but large tributaries. It is a major tributary to the Red Deer Bedrock Channel, the confluence likely being as shown on the map. Throughout the upper half of its course the channel is coincident with present-day valleys but the lower half is not readily apparent from surface features. It seems possible that, had headward erosion of this channel not been interrupted, stream piracy would have occurred, causing diversion of the ancestral Red Deer River through the Buffalo Lake Bedrock Channel.

The Bow River Bedrock Channel System

The Bow River Bedrock Channel rises in the Rocky Mountains and, after entering the plains region, follows an easterly course across southern Alberta. Above Bassano (Tp. 21, R. 18, W. 4th Mer.) the present Bow River occupies the bedrock channel or is diverted only slightly from it. Thirty miles below Bassano the present Red Deer River has a similar relation to the lower portion of the same bedrock channel, following it very closely for the most part. The intervening portion of the channel is mapped on the basis of several water-well logs which indicate the presence of a depression on the bedrock surface between the upper and lower channel segments, with the correct relations with regard to elevation and direction.

The Bow River Bedrock Channel is joined by several tributaries from the north and at least two tributaries from the south. Those that enter from the north all drain a broad upland area that forms the divide between the Bow River and Red Deer Bedrock Channels. West of the present Red Deer River the streams that previously drained south and had a base level controlled by the Bow River Bedrock Channel now have a base level controlled by the canyon of the Red Deer River. Thus the present-day streams, where the drainage has become sufficiently well integrated, have eroded through and in some cases removed all the coarse alluvium along their channels, and bedrock outcrops along these channels are common. As in the region of the headwaters of the Buffalo Lake Bedrock Channel, these tributaries are shown on the map in greater detail than other minor tributaries because they are easily recognized from the present topography.

The upper portion of the Hanna Bedrock Channel is well established from subsurface data and topographic features but the lower half of the channel is not. Information on the Sounding Creek Bedrock Channel is scant, but it may be a tributary of the Red Deer Bedrock Channel.

The present Highwood River, above the town of High River (Tp. 19, R. 29, W. 4th Mer.) follows an ancient bedrock channel. At this point the river makes a turn to the north but the bedrock channel continues to follow an easterly direction crossing the upper portion of the valley of the Little Bow River above Frank Lake (Tp. 18, R. 27, W. 4th Mer.), before bending northeast to cut through a bedrock upland area in a broad, deep bedrock valley. This feature has been named the Frank Lake Bedrock Channel. It is thought that the ancient Highwood River once occupied the Frank Lake Bedrock Channel but was diverted from this course to its present course by stream piracy.

The Oldman Bedrock Channel System

The main branch of the Oldman Bedrock Channel drained part of the Rocky Mountains and plains of southern Alberta as does the present-day Oldman River, which follows, more or less, the ancient bedrock channel. However, at an early stage of development of this bedrock channel, the river followed a northerly course below Fort Macleod (Tp. 9, R. 26, W. 4th Mer.), the channel passing west of Blackspring Ridge (Tps. 12 and 13, R. 22, W. 4th Mer.). This segment of the channel is given the name Carmangay Bedrock Channel in this report. At Carmangay (Tp. 13, R. 23, W. 4th Mer.) the channel makes a wide turn to the east, and its course around the north end of Blackspring Ridge is marked by Travers Reservoir (Tp. 14, R. 22, W. 4th Mer.). East of Travers Reservoir there are four bedrock channels or segments of channels, any one of which may have carried the main stream flow from the southern Alberta Rocky Mountains at one time.

The main branch of the Oldman Bedrock Channel follows the same general topographic low as the present river from Fort Macleod to its confluence with the Bow River (Tp. 11, R. 13, W. 4th Mer.). Bedrock crops out in places where the present river crosses the bedrock channel (Horberg, 1952); this suggests that the present river may have a base level that is lower than that of the ancient river. The location of the lower half of the Oldman Bedrock Channel is not well established and is drawn on the basis of meager surface and subsurface control.

The Medicine Hat Bedrock Channel formerly drained the southeast corner of Alberta. It is a large tributary of the Oldman Bedrock Channel but the location of the confluence shown on the map is based on conjecture. The gradients at various points in the Medicine Hat Bedrock Channel indicate nearby base level control, for, in the upper half of the channel the gradient is steep (10 feet per mile) but below the city of Medicine Hat the gradient is gentle (less than 3 feet per mile), as required for an accordant junction with the Oldman Bedrock Channel. The location of the Medicine Hat Bedrock Channel is well established by subsurface data above the city of Medicine Hat and is fairly well established below that point (Meyboom, 1963).

The Groundwater Resources of the Bedrock Channels

The permanent streams of southern Alberta are widely separated and, the country being semiarid, large water supplies are often difficult or impossible to develop economically at interstream localities. The bedrock formations that underlie the major part of the area are not sufficiently permeable to yield large quantities of groundwater to wells. The bedrock channels, on the other hand, commonly contain sand and gravel strata that are excellent aquifers, and the mapping of the bedrock channels is therefore a necessary step in determining the groundwater resources of the province. Unfortunately, the bedrock channels do not contain sand and gravel at all places, and exploration thus far has indicated that along some portions of every channel only fine-grained alluvium is present and good aquifers do not exist. In many places the bedrock channels coincide with the valleys on the present surface and erosion has removed most of the alluvium that may have been deposited in the channel at one time. In such a channel there is little hope of encountering coarse alluvium below the water table.

The bedrock channels shown on the map vary with respect to the accuracy with which their locations are known. For instance, in the Edmonton-Red Deer area (Fig. 14) the courses of the bedrock channels are quite well defined, whereas knowledge of the pattern and location of parts of the Oldman Bedrock Channel leaves much to be desired.

The distribution of gravel along the course of any stream channel may depend upon the availability of coarse clastic material and the regimen of the stream. The rocks of the area weather to fine-grained material and were not the source of the gravel found in the bedrock channels. However, the detritus from the more indurated rocks of the mountains has been carried into the plains region and deposited in stream channels since early Tertiary time. The only source of coarse sediments for bedrock channels that rise in the plains is the gravel that was deposited during an earlier stage of the development of the present surface (Figs. 17A to 17G) and that has since undergone another cycle of erosion and deposition. Thus, of the bedrock channels that arise in the plains, those that pass near upland areas capped by gravel should be the most likely to contain coarse alluvium. All the streams in figures 17A to 17G are surrounded by gravel-capped hills. All bedrock channels that rise in the mountains may be expected to contain coarse alluvium along some parts of their courses.

The regimen of a stream controls the sedimentation in the stream bed and one of the most important factors in this regard is the gradient of the stream. Coarse, well-sorted sediments are most likely to occur where the gradient decreases suddenly, which makes knowledge of the gradients of the bedrock channels of great importance. Deposition can also be expected if the load of a stream is increased suddenly as, for example, at the point where a tributary with a higher gradient joins the main channel.

At the intersections of present-day stream valleys with the ancient bedrock channels the coarse sediments of both the modern and ancient streams may be present and have sometimes been resorted and redeposited as thick permeable strata by the younger stream. At such sites groundwater recharge of these permeable strata by induced infiltration may be possible if the bedrock channel floor is lower than the present stream level (Meyboom, 1963). This is an ideal situation for the production of large amounts of groundwater and such locations are worthy of close examination where large supplies of water are required.

There is evidence that the aquifers of the Alberta plains region are being recharged by local precipitation, although the rate of recharge is certainly low. The determination of this rate of recharge is a necessary step in the proper evaluation of the groundwater resources of the bedrock channels.

It is obvious that the map presented in this study can serve only to indicate the areas most favorable for the occurrence of groundwater in abundance. It is also obvious that one is not justified in depending on supplies of groundwater from aquifers in bedrock channels until a thorough testing program has been conducted.

THE DRIEDMEAT LAKE ARTIFICIAL RECHARGE PROJECT

by
R. N. Farvolden

Introduction

Purpose and Scope of the Report

In many places where groundwater is an important and exploited resource, the groundwater reservoirs are being depleted because the rate at which water is being withdrawn from them exceeds the rate at which they are being recharged by natural processes. In some cases, artificial methods have been devised to increase the rate of recharge. A successful project of this sort has been established on a rock-defended gravel terrace at Driedmeat Lake and the water that is recovered now supplies the city of Camrose. This paper reports on the investigations into the geology and hydrology of the aquifer involved, and on the operation of the recharge system.

Description of the Area and Physiography

Driedmeat Lake is located in the plains region of Alberta, about 8 miles south of the city of Camrose and 60 miles southeast of Edmonton (Fig. 2). The lake is a natural, shallow reservoir on the Battle River, about 12 miles long and about one-half mile wide. The valley of the river has steep banks 150 feet high, and a flat bottom up to 1 mile wide. Bedrock exposures are common along the lower portion of the valley wall where the slope is steep enough to be bare of vegetation.

Tributary valleys become deeply incised as they approach the main valley. All local tributaries are intermittent and, during very dry summers or after a series of dry years, the Battle River has been reported dry.

Camrose is a city with a population of 6,000 and it is the business center of a large and rich mixed-farming district. The artificial recharge project at Driedmeat Lake was established to provide this growing community with an adequate water supply, for alternative sources involve either expensive treatment, or prohibitively long pipelines, or both.

Acknowledgments

This project was undertaken and financed by Calgary Power Limited of Calgary, Alberta, with the Research Council of Alberta acting in an advisory capacity. Field work was done under the direction of Mr. Percy Maggs of Calgary Power and the author.

Most of the information on the operation of the system was provided by Mr. Maggs and Mr. F. E. Balshaw, also of Calgary Power. Without the enthusiasm and determination of these two gentlemen, the project could not have been successful.

Mr. A. Beasley, the owner of the land on which the work was carried out, was thoughtful and co-operative throughout the long period of testing and experimentation.

For the help received from these and other people the author is extremely grateful.

Geology

The surface strata of the Camrose-Driedmeat Lake area belong to the Edmonton Formation of late Cretaceous age. This formation is a sequence of sandy shales and fine-grained, dirty sandstone beds and coal seams.

Between the time of deposition of the Edmonton Formation and the advent of the first period of glaciation, extensive erosion took place over the plains of Alberta. The Rocky Mountains were being uplifted and rivers, rising in this highland area, flowed across the prairies, much as they do today. These preglacial rivers cut valleys in the surface of the bedrock, the Edmonton Formation in this case. One of these rivers flowed from west to east in a bedrock channel that cuts through the present Battle River valley at about the center of Driedmeat Lake. Several tributaries entered the main valley from the north. This bedrock channel is the valley of the preglacial Red Deer River (Stalker, 1960) and further downstream it contains thick gravel aquifers.

During the Pleistocene Epoch, the area was overrun, at least once, by a continental ice sheet. A mantle of till, from 10 to 150 feet thick, covers the whole area except where it has been removed by recent erosion. Sorted granular deposits are found associated with the till, and aeolian and lacustrine deposits are also known in the area. Glacial drift has nearly filled in the preglacial valleys. The main channel of the preglacial Red Deer River is now merely a broad, shallow depression which is not easily discernible in all places along its course.

Origin of Driedmeat Lake Terrace

During the melting stages of the ice, when huge quantities of water were released, large lakes formed at the snout of the glacier. One of these lakes formed in a broad depression now occupied by the city of Edmonton (Gravenor and Bayrock, 1956; Hughes, 1958). The lake level was raised until the water finally spilled over a low divide, near the present site of the town of Leduc, into the lower ground beyond. The water ran southeast along the ice front and eroded a valley that curves in a broad arc across east-central Alberta. This valley, called the Gwynne Outlet (Gravenor and Bayrock, 1956), is about one mile wide, 150 feet deep, and 150 miles long. The Battle River, a misfit stream now occupies all but the upper 25 miles of it. Driedmeat Lake is a much widened portion of the Battle River.

Several miles above Driedmeat Lake, the Gwynne Channel crosses a tributary channel of the preglacial Red Deer River. This tributary channel contains large quantities of sand and gravel, and these granular materials were picked up by the eroding torrent and redeposited in bars downstream. At a later stage in development, the base level was lowered so that the stream began cutting down into these newly deposited sand and gravel bars, leaving only remnants of them high above the present valley floor. The terrace on which the recharge experiments were conducted is one of these high-level, sand and gravel bars.

Groundwater Hydrology

The sandstone strata and coal seams of the Edmonton Formation commonly carry fresh water that is characteristically soft and moderate to high in soda and iron. Wells completed in these beds furnish almost all the water produced in the area except for that used in the Camrose municipal system. Good wells for domestic and stock-watering supplies are common and the wells originally used to supply Camrose were capable of producing up to 20,000 gpd (gallons per day). However, there is little chance that better wells can be obtained and, since no fresh water is found below the Edmonton Formation in this area, it was decided that testing of the bedrock strata was not warranted in this particular program.

A test hole was drilled into the preglacial valley of the Red Deer River about 9 miles southeast of Camrose. The results of this and other nearby borings into the channel deposits indicate that these deposits are not suitable aquifers along this portion of the channel.

The till is usually very impermeable and commonly even dug wells are dry unless a lens of sand or gravel is encountered in the hole. Where porous, granular strata are encountered in a drill hole a good domestic well can be obtained, but the aquifers are mostly very limited in areal extent. The Pleistocene deposits are not important as aquifers in this area.

Hydrology of Driedmeat Lake Terrace

Near the upper end of Driedmeat Lake there are two terraces, one on either side of the valley. The terrace on which the recharge project is being conducted is on the north side of the lake (Fig. 18). It is about 3 miles long and 1,500 feet wide. For the most part, the surface of the terrace is flat, but a few old meander scars are present. The surface slopes from an elevation of 2,350 feet at the northwest end to less than 2,300 feet at the southeast end. The water level in the lake is at an elevation of 2,246 feet.

The terraces are rock defended, that is, they are bedrock terraces with a layer of granular material at the surface. The contact between the granular materials and the top of the bedrock is 20 feet or more above the lake level. Therefore, the gravels of the terrace cannot be recharged directly by induced infiltration from Driedmeat Lake.

A shallow depression along the north edge of the terrace carries spring runoff, and at the lower end of the terrace this depression becomes a gully which joins a small creek near the lake shore (Fig. 18). The south edge of the terrace nearest the lake is thus higher than the north edge. Test drilling and examination of outcrops indicate that the surface of the bedrock conforms to the surface of the terrace and also slopes from a high along the south edge toward the draw along the north edge of the terrace (Fig. 18). It is this feature that makes the gravel a groundwater reservoir for, without the bedrock dam along the south edge of the terrace, the groundwater would leak out to the valley and form a spring line along the north bank.

The surface of the terrace rises from the southeast to the northwest. Again, the bedrock surface conforms to the present-day surface and, since the water table is nearly horizontal, the saturated thickness of the gravels decreases toward the northwest. Between wells Nos. 11 and 10 (Fig. 18), the bedrock rises above the water table to form one boundary of the reservoir.

The sand and silt content of the gravels varies considerably both horizontally and vertically. All test holes within the limits of the reservoir encountered porous, saturated gravel overlying the bedrock, although in almost every case dirty and relatively impermeable sand and gravel comprised a portion of the section. These strata have the same effect as impermeable confining beds if they are encountered below the water table, because when the drilling tools penetrate the impermeable strata and enter the saturated, porous strata the water will rise in the hole, suggesting artesian conditions.

Natural recharge of the aquifer occurs in two ways. Precipitation and runoff onto the surface of the terrace result in recharge by influent seepage when the ground is not frozen. Precipitation, totalling 3.13 inches, in the second week of August, 1957 caused a rise in the water table of about 3 inches (Fig. 19). During spring runoff, however, the surface of the ground is frozen and impermeable and there is no evidence of recharge by snow meltwater. Natural recharge also occurs by percolation into the gravels from the drift and bedrock of the adjacent valley wall. The Battle River acts as a groundwater drainage ditch and the gravels of the terrace intercept the water moving toward it. The reservoir will thus show an increase in storage during the winter months after being depleted by pumping during the late fall. The average natural recharge is likely represented by the 50 gpm (gallons per minute) flow of the springs at the lower end of the terrace.

Analysis of pumping-test data confirms the geologic evidence concerning permeability variations in the gravels and the locations of hydrologic boundaries. For a homogeneous aquifer of infinite extent, provided sufficient time has elapsed since pumping began, the drawdown curves for the pumped well and the observation wells should be straight lines when drawdowns are plotted against the logarithm of the time (Cooper and Jacob, 1946). During a three-day pumping test from August 7 to 10, 1956, conducted on well No. 8, the drawdown curves for observation wells Nos. 4 and 7 (Fig. 20) showed that the water levels were dropping at a rate greater than would be anticipated if the aquifer were an extensive one. The 21-day test in October, 1956, showed similar downward-trending curves.

In subsequent pump tests it was found that observation wells Nos. 9 and 11 did not draw down as quickly as anticipated. This is likely due to a decrease in permeability in the gravel between these wells and the pumping well. This change in permeability is also suggested by the increased gradient in the water table near the springs—a gradient that could not be present in highly permeable gravels without much greater flows. The variations in flow of the springs correlate very well with the changes in water level in well No. 9, so the springs and well No. 9 must all be on the same side of the permeability barrier. Well No. 10 is apparently isolated from this hydrologic system altogether.

To estimate the amount of water in storage in the reservoir it is assumed that the reservoir is 4,000 feet long and 700 feet wide and that it has a saturated thickness of eight feet and a specific yield of 0.20. The total volume of water in storage is then:

$$4{,}000 \times 700 \times 8 \times 0.20 = 4.48 \times 10^6 \text{ ft}^3$$
$$= 2.8 \times 10^7 \text{ imperial gallons}$$

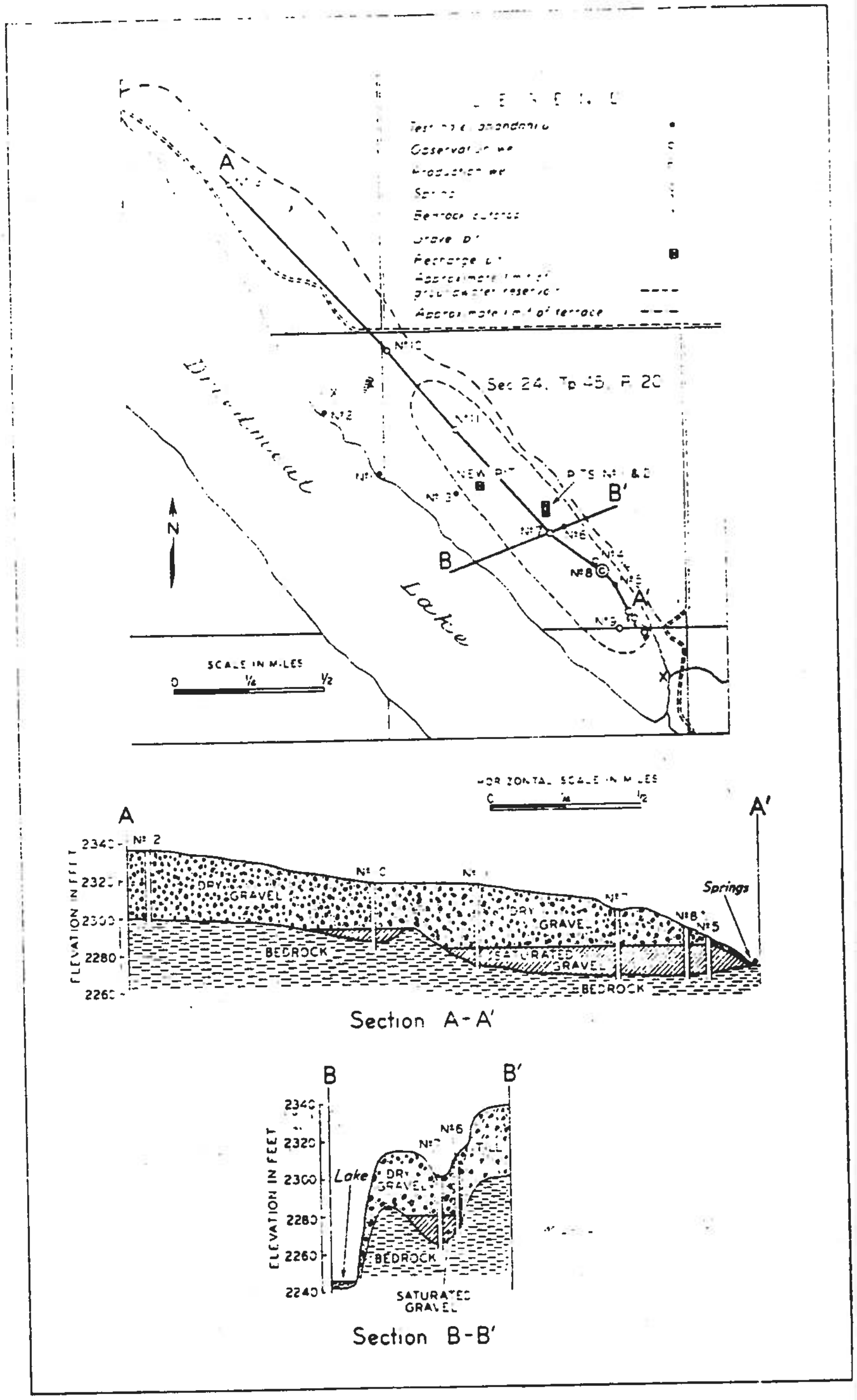


Figure 18. Map and cross sections, Driedmeat Lake terrace.

The drawdown in well No. 8 after it had been pumped 72 hours at 500 gpm was only 2 feet. This indicates a high permeability for the aquifer and very low entrance losses at the well. The gravels are so permeable that the water table remains nearly flat during pumping and, therefore, the reservoir can be totally depleted by one production well.

Artificial Recharge

Three problems had to be solved before artificial recharge could be considered feasible for this reservoir. First, the water had to be introduced into the gravels in such a way that relatively economic experimentation could continue. If experimentation proved successful, then the cost of more elaborate works might be borne. Secondly, a high proportion of the water recharged to the gravels had to reach the aquifer and be recoverable from the production well. Finally, the "high color" of the raw lake-water—40-50 ppm (parts per million)—had to be reduced sufficiently during the process of artificial recharge and subsequent recovery to ensure that no expensive treatment of the final product would be necessary. Only then would the process be economically successful.

Water-Spreading Experiments

Water spreading over large areas of native grass is a common technique for inducing influent seepage in artificial recharge projects (Todd, 1959a). During the first ten days of the pump test of No. 8 well in October, 1956, the water was discharged onto the ground near the well. It is obvious from the hydrograph for No. 4 observation well (Fig. 19) that some of this water was infiltrating down to the water table. When a pipeline was installed to carry the discharge water 1,000 feet away to the end of the terrace, the water level in the aquifer dropped.

Experiments in infiltration through the soil cover on Driedmeat Lake terrace showed the infiltration rate to be fairly large at first, but the rate began to decline within 24 hours. Water spreading over grass would be almost impossible in this area during the winter and, with these two difficulties immediately apparent, experiments in water spreading were abandoned.

Shallow Recharge-Pit Experiments

Two pits, each 50 feet long, 10 feet wide, and 2 to 4 feet deep, were excavated 1,000 feet from the production well and approximately 4 feet apart. The gravels in the pits varied as expected from very dirty gravel containing abundant silt and fine sand, to clean, well-rounded half-inch gravel. However, relatively dirty gravel was predominant and only in one corner of No. 2 pit was clean gravel uncovered. On October 15, 1957,

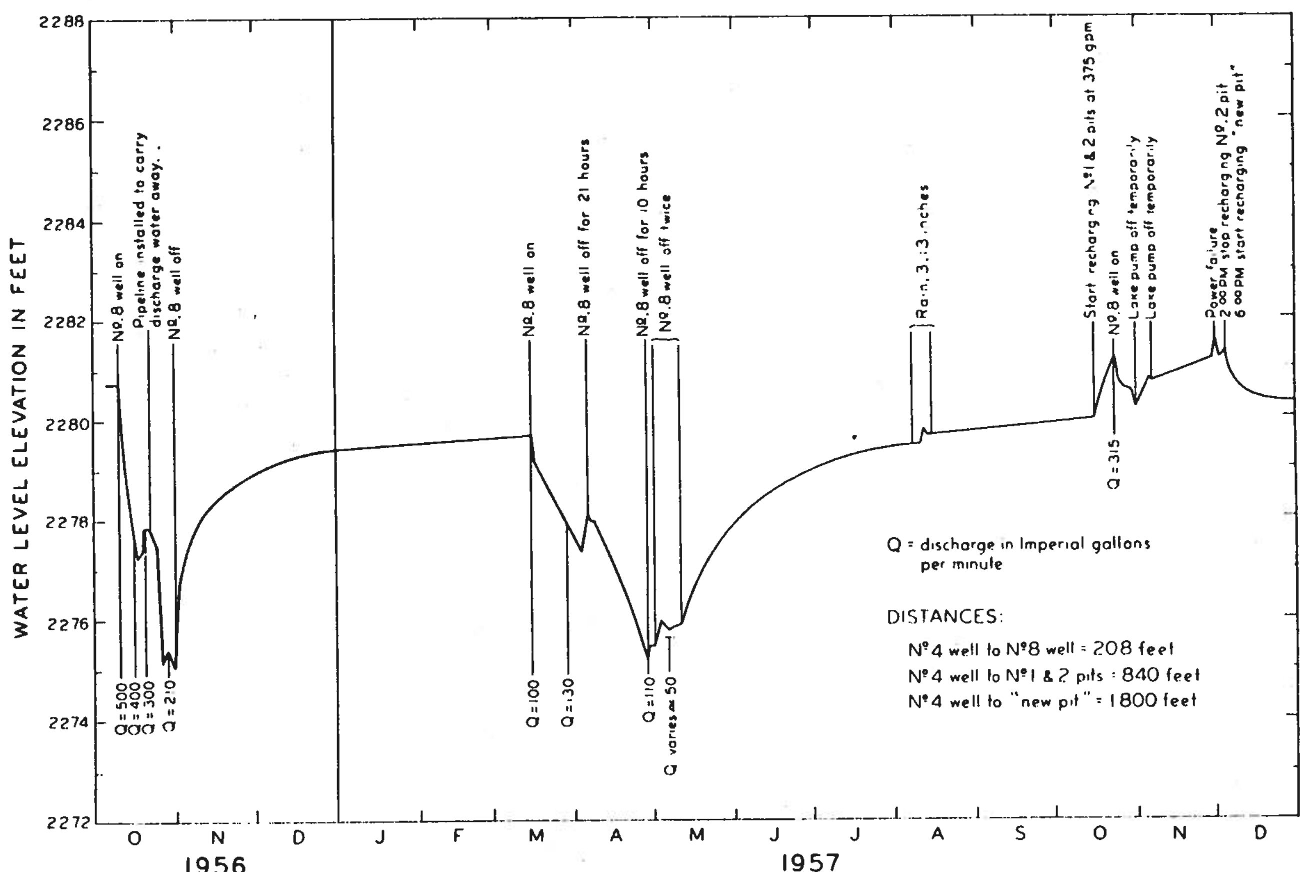


Figure 19. Hydrograph of No. 4 observation well, Driedmeat Lake terrace.

water was pumped up from the lake through a 6-inch temporary transite line and delivered to No. 1 pit at 375 gpm. This pit filled quickly and only small quantities of water were lost through seepage. The flow was diverted to pit No. 2. Most of the bottom of pit No. 2 had about the same hydrologic characteristics as pit No. 1. However, when the water reached the open gravel in one corner of pit No. 2, it was drained away immediately. The whole flow was directed to this corner and the water that had accumulated in the pit drained out. The area of clean gravels exposed in the pit and taking the water was about 3 feet square.

The water levels of all wells were observed as the experiment proceeded. The influence of the recharge was soon evident in the three nearest wells (Fig. 20). The other wells responded much more slowly, partly because of the greater distances, but mostly because of the decrease in permeability at either end of the reservoir. It was apparent that the water pumped into the pit was being added to the storage in the reservoir, where it could be recovered by pumping No. 8 well.

The one remaining problem was that of color removal. Six days after artificial recharge into No. 2 pit had commenced, the production well 1,000 feet away was pumped at 315 gpm, or 84 per cent of the rate of recharge. After 30 days of pumping, the color of water produced from this well had risen from 3 ppm to 17 ppm. It was decided that the recharge pits were too close to the producing well. A new pit was excavated 2,000 feet away from the producing well. At this location, very clean gravels were found after a little exploration with a backhoe excavator. The raw water was diverted from the old to the new pit and the hydrologic properties of the two pits appeared to be identical. The recharge test was continued throughout the winter and by April 3, 1958, the color of water from the production well appeared to be stable at 7 ppm. This was considered adequate proof that the process of recharge and infiltration through the gravel reduced the color of the raw water to an acceptable level, and the test was stopped.

During the 170 days of this test, 89 million imperial gallons of raw water were pumped into the gravels and 70 million imperial gallons were recovered from No. 8 well. During the same time the storage had been increased so, though exact figures are not known, a very high percentage of recovery is indicated.

Operation

During the summer of 1958, a pipeline was built from the terrace to Camrose and permanent equipment and structures, including a lake pumping station and a standby well, were installed at Driedmeat Lake. Since November 28, 1958, this system has supplied Camrose with water.

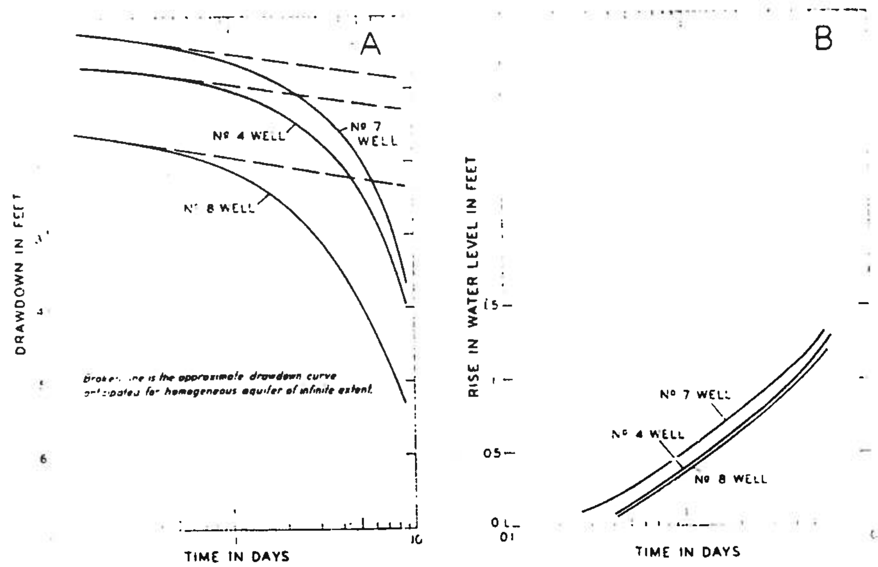


Figure 20. Pump and recharge test results, Driedmeat Lake terrace.

The recharge pit used in the tests has been replaced by a new pit in the same location. The pit has bottom dimensions of 8 feet by 24 feet but when water is being introduced at 300 gpm the bottom of the pit is not completely flooded (Fig. 21), and a staff gauge installed to show the depth of water in the pit has not yet been wet. Cold weather has not affected the recharge operation. The lake pumping station was shut down for the first two weeks of January, 1959, during which time the temperature varied between 0°F and a minimum of —31°F. No trouble was experienced in resuming operations despite these extreme conditions.

Color reduction between the pit and the production well is from 40-50 ppm to 7-8 ppm. Bacteriological analyses of the raw lake water sometimes give positive and sometimes negative results. Samples from the springs also showed occasional contamination even before recharge experiments had started, but this contamination is thought to be very local. So far all samples taken from the production well have given negative results.

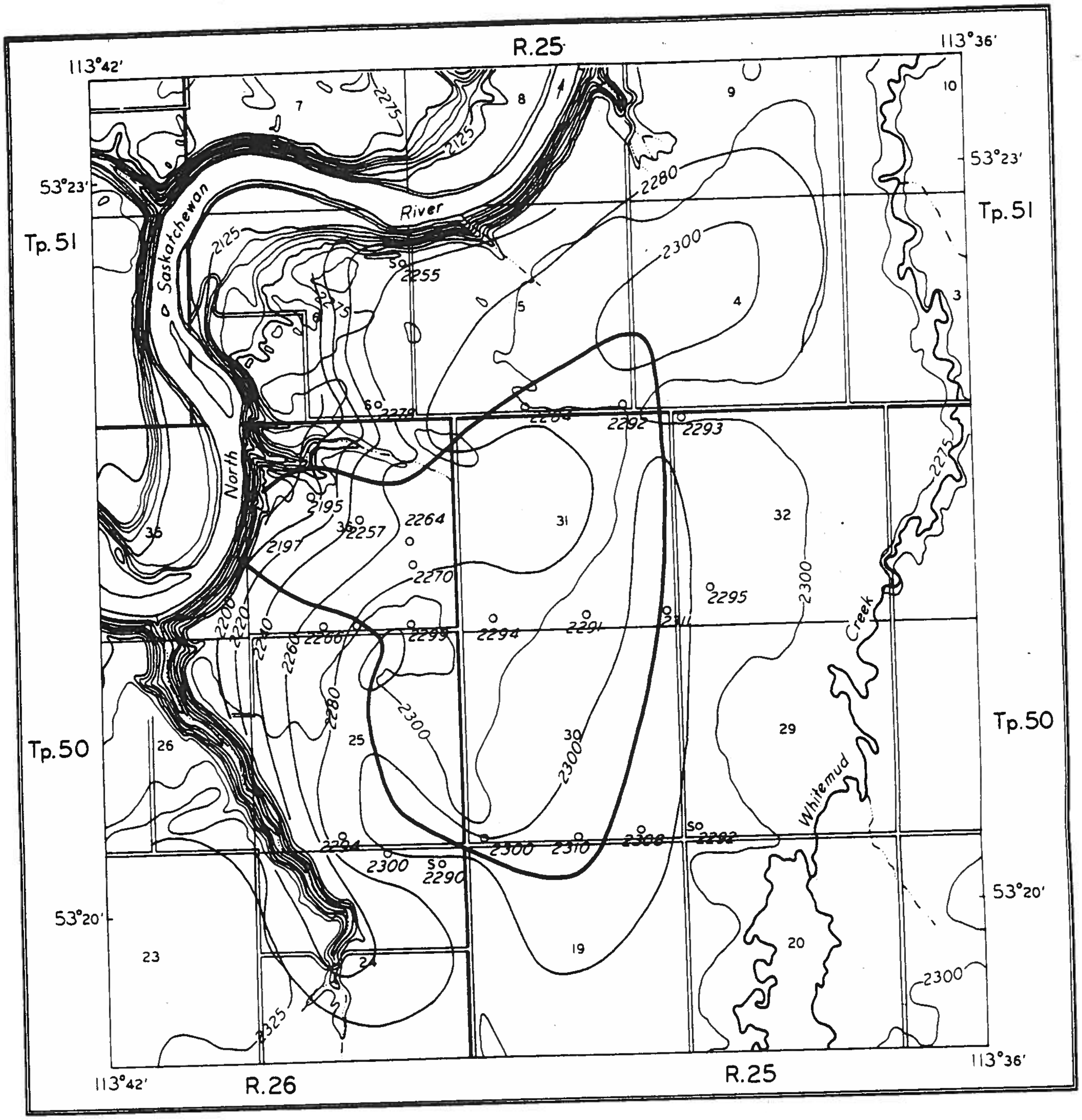
Chlorine is added to the raw water in quantities of about 3 ppm. Before the water enters the storage tanks in Camrose, chlorine and ammonium sulfate are added to give a total residual of 1.75 ppm. No other treatment is required but a micro-strainer has been installed at the lake pumping station to cut down on the amount of organic material entering the gravels where it might become a problem in the future.

The temperature of the water at the production well has varied from 38°F in December, 1958, to 44°F in July, 1959.

Figure 21. Pumping into recharge pit, Driedmeat Lake terrace.

Economics of the Project

The purpose in using this system instead of conventional treatment methods is solely economic for, though there are other benefits such as constant temperature and chemical quality, these alone would not justify use of the artificial-recharge system. It is estimated by the operators that construction of a treatment plant or enlargement of the old plant would have required $150,000.00 more in capital investment than the recharge system. The cost of chemicals for this type of raw water had risen to 11.5 cents per thousand gallons at the Camrose treatment plant, while chemicals for the recharge supply are 0.75 cents per thousand gallons. At a rate of production of 100 million gallons per year, the annual savings realized by the consumers in Camrose are:



SCALE IN MILES

1 ¾ ½ ¼ 0 1

CONTOUR INTERVALS

Surface elevation 25 feet

Piezometric surface 20 feet

Elevations in feet above mean sea level

LEGEND

Observation well, with water table elevation 2299

Shallow well .... so

Spring ....

Contour, surface elevation 2125

Contour, piezometric surface 2300

Groundwater divide ....

Road, loose surface ....

Stream, intermittent ....

Figure 29. Topography and piezometric surface near Devon.

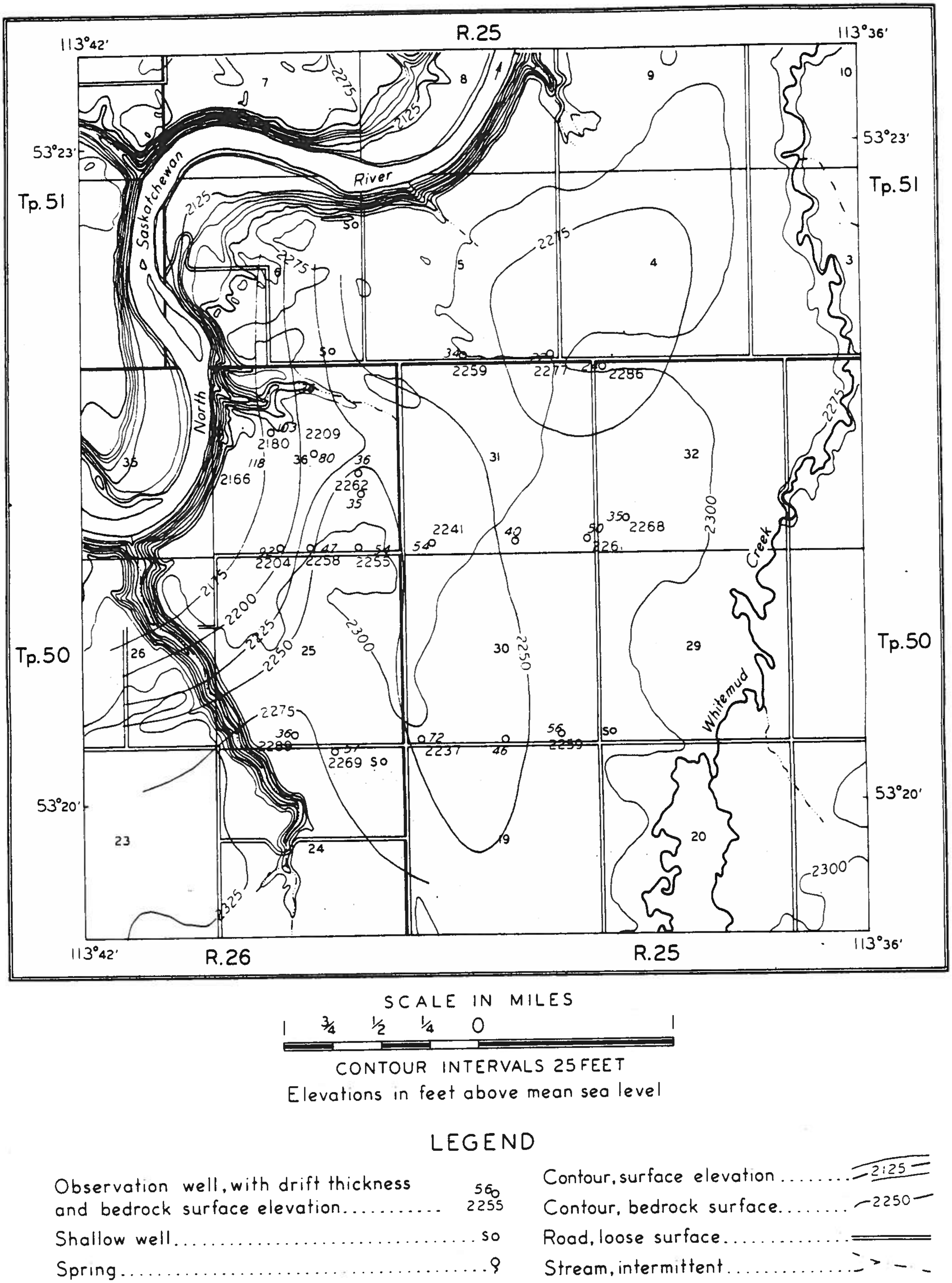


Figure 30. Bedrock topography and drift thickness near Devon.

$$\$150{,}000 \times 0.07 = \$10{,}500 \text{ (interest at 7 per cent on savings on capital expenses)}$$

$$\frac{100{,}000{,}000}{1{,}000} \times \$0.1075 = 10{,}750 \text{ (savings on chemical costs)}$$

$$\frac{100{,}000{,}000}{1{,}000} \times \$0.10 = \underline{10{,}000} \text{ (savings on labor costs)}$$

$$\$31{,}250$$

Conclusions

Where there is an abundant supply of surface water that is difficult or expensive to treat, and the geologic conditions are favorable, it may be worthwhile considering artificial recharge when planning new facilities. However, an ideal geologic setting, such as found at Driedmeat Lake, is likely quite rare. There is no possibility of putting a system like this into operation without considerable expenditure of time and money on the essential geologic and hydrologic testing.

INDUCED INFILTRATION, MEDICINE HAT, ALBERTA

by
P. Meyboom

Introduction

Purpose and Scope of the Report

This report deals with the results of the search for a large industrial groundwater supply, carried out by the Research Council of Alberta in close co-operation with the city of Medicine Hat.

Because the bedrock in this part of the province is not considered to be favorable for high-capacity wells, the city of Medicine Hat (Fig. 2) was advised to test the groundwater occurrences at Police Point, a meander terrace along the South Saskatchewan River, one-half mile east of the

Figure 22. Air photograph of Medicine Hat area showing meander terrace at Police Point.

city (Fig. 22). This terrace was expected to consist of fairly coarse gravel which, if extending far enough below river level, might yield large quantities of water by means of induced infiltration.

Location and Geology of Police Point

Police Point is a meander terrace approximately one-half square mile in area along the northern bank of the deeply incised South Saskatchewan River (Fig. 22). The surface of the terrace rises only 20 feet above river level and it lies approximately 100 feet below the adjacent upland. The presence of a luxuriant vegetation is the most striking difference between the appearance of Police Point and that of the dry and barren prairie that surrounds it.

Test drilling revealed that the terrace is covered with 10 feet of medium-fine to fine sand, below which lies a thick gravel deposit resting on shale of the Foremost Formation. The gravel is medium sized at the top and becomes gradually coarser toward the base. The thickness of this gravel bed varies from 39 feet in test hole No. 1 (Fig. 23) to 19 feet in test hole No. 4. In test holes Nos. 1 and 2, bedrock was reached at 55 and 58 feet below ground surface, respectively; in test hole No. 4, however, bedrock was reached at only 37 feet below ground surface. On the basis of test drilling farther away from the river (test hole No. 7) there appears to be a rise in bedrock and a lateral change from coarse gravel to fine sand. A schematic cross section is given in figure 23.

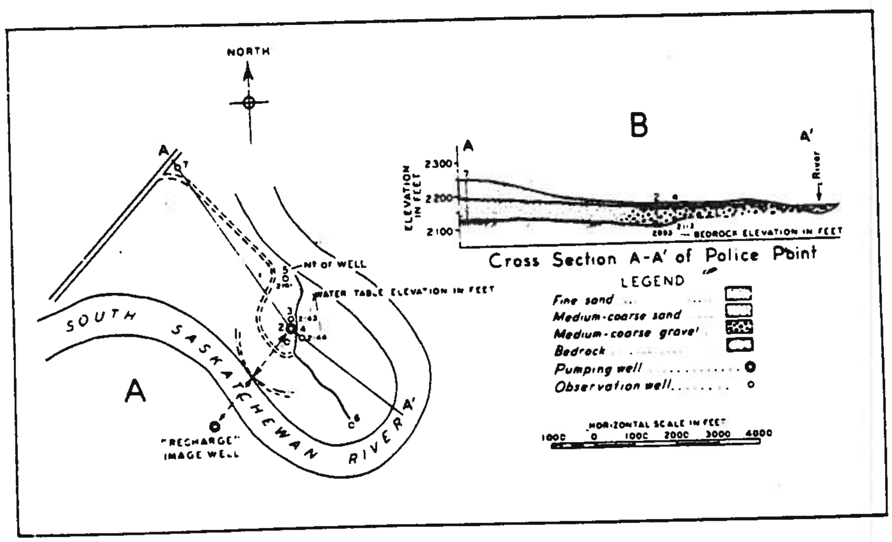


Figure 23. Map and cross section, Police Point, Medicine Hat.

These results suggest the presence of a channel-like depression in the bedrock surface, and it is believed from this and other evidence that the center line of this depression extends in a northerly direction, approximately parallel to the present South Saskatchewan River but a few miles west of it (Fig. 24). The existence of this channel was first suspected when drillers' logs from the Cavendish, Buffalo and Bindloss areas (Fig. 24) reported excessive drift thicknesses. By means of a seismic survey and earth-resistivity observations, this channel was traced southward to Medicine Hat and, during a geological field survey in 1958, part of its continuation south of Medicine Hat was also reconstructed (Common, 1958, personal communication). This channel is believed to be of preglacial origin. It is of the same magnitude as the present valley of the South Saskatchewan River, and its northward gradient of approximately 3 feet per mile is the same as the northward gradient of 3 to 4 feet per mile of the present river. Farvolden (1963b) called this channel the Medicine Hat Bedrock Channel.

As the base level of postglacial erosion in eastern and central Alberta is generally below the preglacial or glacial erosion base level, most present-day rivers have cut deep valleys, commonly into bedrock, thus exposing the glacial and preglacial deposits that occur in their valleys. Cases where preglacial or glacial deposits still lie below the present river valley bottom are, therefore, relatively rare in Alberta. For example, the thickness of most gravel beds along the major streams in the province seldom exceeds a few feet, and the fact that the gravels at Police Point extend at least 38 feet below the present South Saskatchewan River is exceptional.

Groundwater Hydrology

The static groundwater-levels in the test holes at Police Point were established and a groundwater gradient from southwest toward northeast was determined, indicating that part of the river flow travels through the gravels of the terrace, rather than around the meander. Any well obtaining its water from this reservoir can easily supply all desired quantities of water, provided that the aquifer can transmit them.

Hydrologic Boundaries and Aquifer Coefficients

For test purposes, well No. 2 was equipped with a suction pump, powered by a tractor. The well was pumped for 46 hours at 275 gpm (gallons per minute). Test holes Nos. 1, 3, and 4 were used for observation wells.

Semilogarithmic plots of the drawdowns in wells 1 and 4 versus the elapsed time since pumping began (Fig. 25) clearly show the existence of one positive or recharge boundary. The plot for well No. 4 possibly indicates the existence of a negative or discharge boundary. The positive

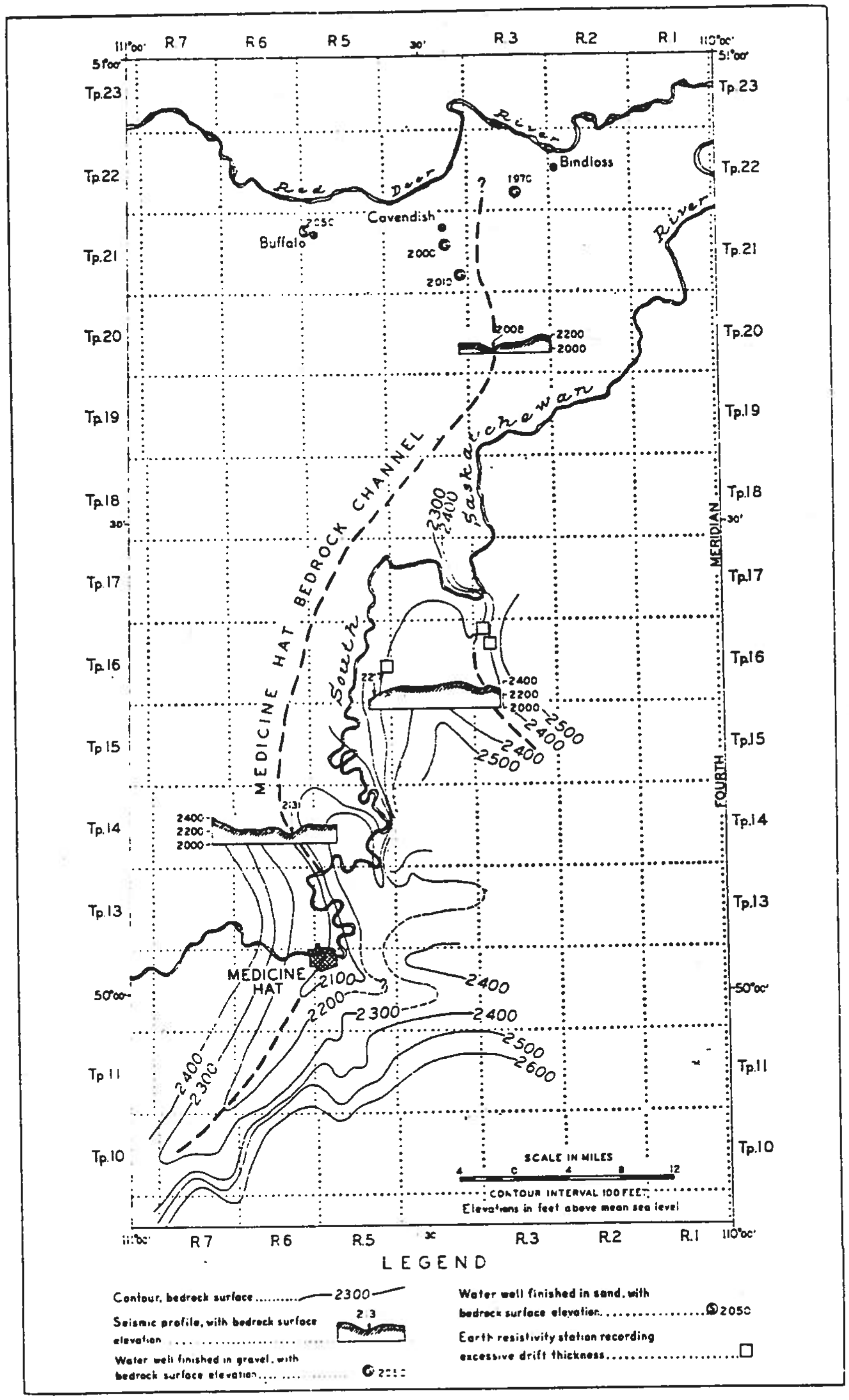


Figure 24. Medicine Hat Bedrock Channel.

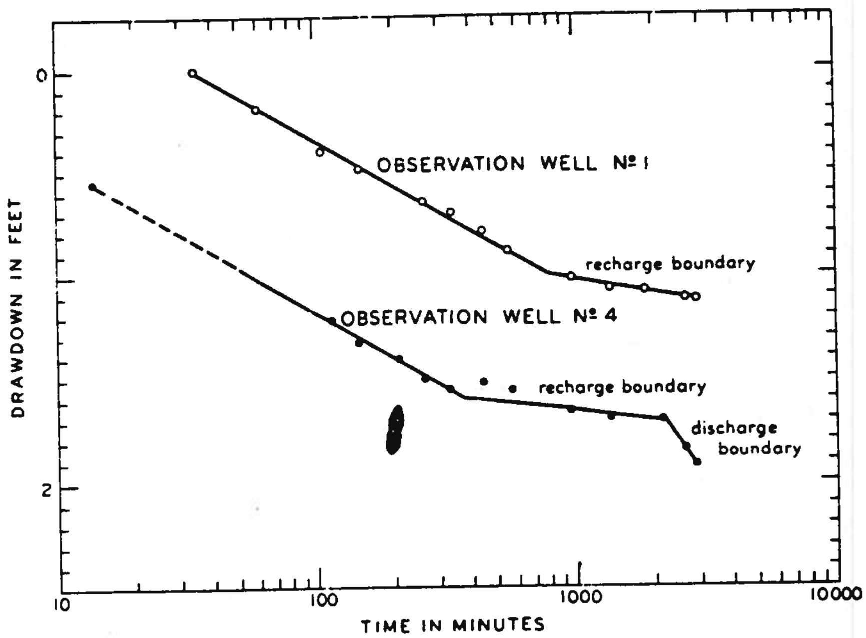


Figure 25. Pump test results, Police Point, Medicine Hat.

boundary is the South Saskatchewan River and along this boundary no drawdown occurs. The negative boundary, if it exists, will be found where there is a reduction in permeability. Although it is impossible to give an exact location for the negative boundary, it is believed that the lateral change in lithology may account for it (Fig. 23).

Because there can be no drawdown at the positive boundary, the cone of depression becomes distorted and the hydraulic gradient between the well and the river becomes steep compared to the gradient in other directions from the pumping well. As a result, more water moves toward the pumping well from the river than from other directions and, as time goes by, a greater and greater percentage of the water discharged by the well is drawn from the river and a smaller and smaller percentage from storage within the aquifer (Glover and Balmer, 1954). This process is known as *induced infiltration.*

In order to determine the coefficient of transmissibility, both the nonequilibrium formula and the Theis recovery formula (Theis, 1935) were used. The results of both methods agreed very closely and the average value for the coefficient of transmissibility was calculated to be 120,000 gpd/ft (gallons per day per foot).

The coefficient of storage is extremely difficult to determine from a short pumping-test in an aquifer of this type. The Theis nonequilibrium formula is based on the assumption that water is released from storage instantly with decline in head, but this is not often the case in a water-table aquifer. In the initial stages of pumping the volume of the cone of depression is substantially larger than would be anticipated from the volume of water that is pumped out (Ubell, 1956). This phenomenon is called *delayed yield* and causes the coefficient of storage obtained from the nonequilibrium formula to be too low. Table 12 shows several stages of the pumping test, each one characterized by a different apparent coefficient of storage.

The high value of 120,000 gpd/ft for the transmissibility and the unusually large thickness of saturated gravel resulting from the fortunate location of the preglacial channel directly under Police Point combine to make this site a most favorable one for a major groundwater development, at least insofar as water quantity is concerned.

Table 12. Effect of Delayed Yield on Coefficient of Storage, Medicine Hat

Time since pumping started (minutes)	Rate of pumping (gpm)	Volume withdrawn (cubic feet)	Volume of cone of depression (cubic feet)	Apparent coefficient of storage
100	275	4,400	504,381	0.008
992	275	43,648	629,641	0.069
2,565	275	112,860	656,638	0.17

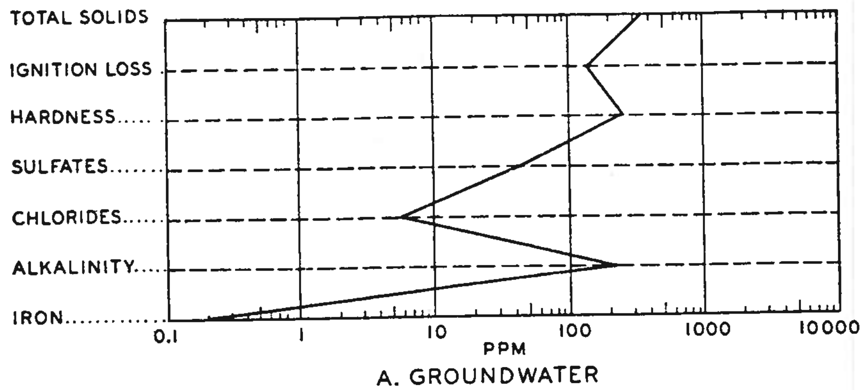
The actual coefficient of storage (specific yield) of the aquifer is estimated to be 0.20.

Chemical Characteristics of Infiltrated Water

The chemical compositions of infiltrated water and river water do not differ to a great extent. The results of chemical analyses are shown graphically in figure 26; the corresponding data are presented in table 13.

As can be seen from figure 27, the amount of total dissolved solids in the river water reaches a maximum during the November to January period of minimum runoff. During May, June and July the dissolved solids are less concentrated as a result of dilution by large quantities of direct runoff.

As has been pointed out earlier, some of the river water flows through the gravel terrace, moving at a much slower rate than the water in the river. The water that was pumped out during the test on Police Point probably



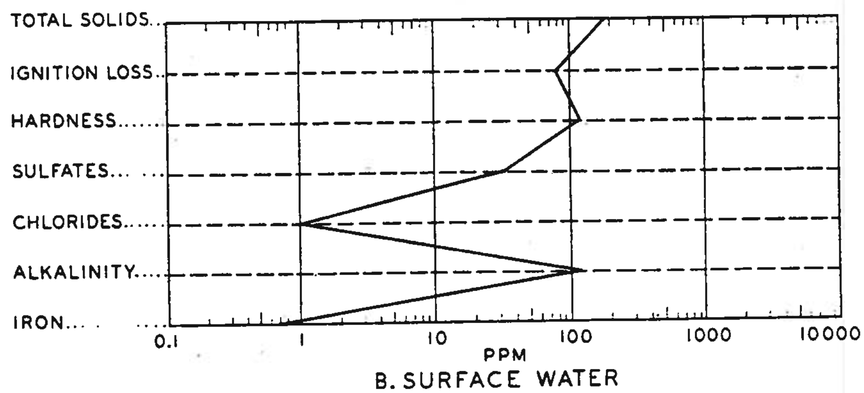


Figure 26. Chemical composition of groundwater and surface water, Police Point, Medicine Hat.

infiltrated into the aquifer during a period of higher total solid concentration in the river. Under conditions of continuous pumping, the difference in chemical quality between infiltrated water and river water should be less and the chemical composition of the infiltrated water should approach the quality of the river water more closely.

Temperature of Infiltrated Water

The relation between surface and infiltrated water temperature is very complex. Kazmann (1948) mentions several factors that influence the temperature of infiltrated water. These are:

(1) the mixing of groundwater with infiltrated water,

(2) the admixture of river waters of different temperatures while en route to the well,

Table 13. Comparison of Chemical Compositions of Infiltrated Water and South Saskatchewan River Water at Medicine Hat

(Analyses by Provincial Analyst, Edmonton)

Constituents	Groundwater*	Surface Water*
Total solids	336	196
Ignition loss	132	78
Hardness	240	125
Sulfates	42	32
Chlorides	6	1
Alkalinity	215	105
Nature of alkalinity	Bicarbonate of lime and magnesia	
Iron	0.2	0.8

* Constituents are expressed in ppm.

(3) the heat storage of the aquifer and the underlying rocks, and

(4) the conduction of heat upward and laterally within the aquifer, due to temperature gradients.

As a net result of these processes the aquifer produces water that has a temperature intermediate to the extreme temperatures of the river water.

For comparison, a graph is given in figure 28 showing temperatures for the Elbow River in Calgary (a 10-year average) and for infiltrated water obtained from the same river. In the same illustration, the temperature of the South Saskatchewan River in Medicine Hat is given for 1951-1952 (Thomas, 1956).

Measurements taken during the pump test gave a temperature of 42°F. It seems reasonable to expect the groundwater temperature to range from about 55°F during December to about 40°F during May and June at Police Point.

As a result of this temperature difference, the coefficient of viscosity will vary from about 1.55 centipoises during summer to about 1.21 centipoises during winter. The resulting increase in flow of water toward the well during the winter months may cause the winter drawdown to be slightly less than calculated.

Effects of Changing River Level upon the Groundwater Level

The aquifer of Police Point is a water-table aquifer and any change in river level will, therefore, be reflected in the groundwater level. Although no recent gauge recordings from the South Saskatchewan were

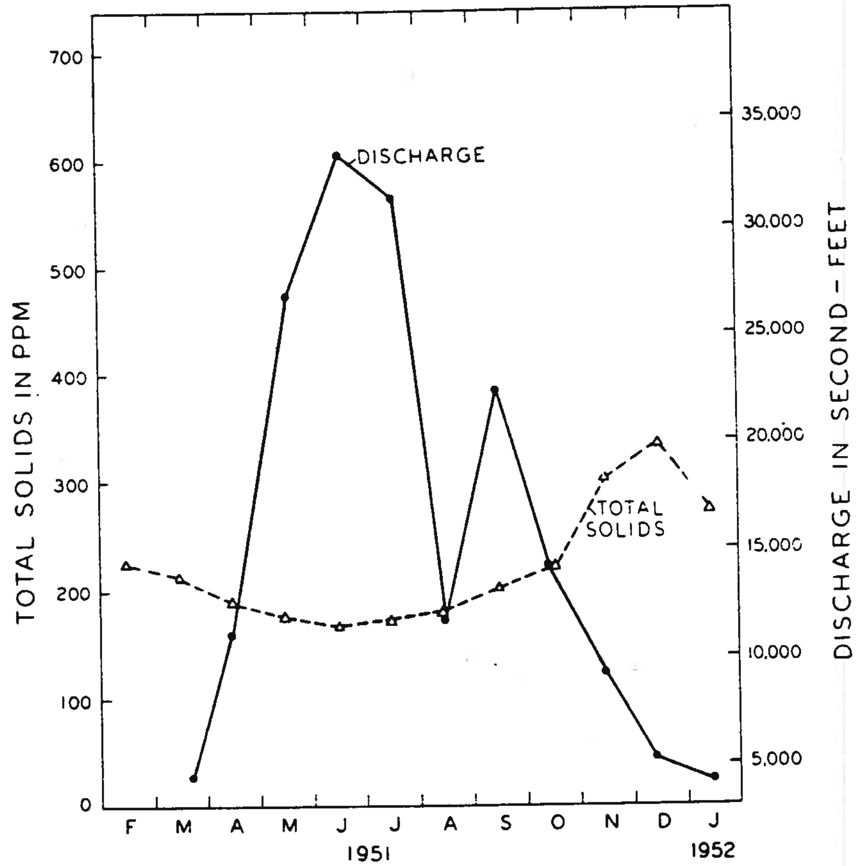


Figure 27. Variations in total solid content and specific discharge of the South Saskatchewan River at Medicine Hat.

readily available, it is believed that the average annual river fluctuation is less than 10 feet, mainly because of the artificial flow regulation associated with upstream irrigation works. The maximum river stage occurs during May and early June (Fig. 27); the minimum stage during December and January. The range between the highest and lowest recorded river stages is 30 feet (the difference between the water levels during the 1905 flood and the 1942 drought), but the saturated portion of the aquifer is sufficiently thick to deal even with emergency situations such as this. The groundwater level during August probably represents the average situation. During the winter months, the groundwater level may be 3 to 5 feet lower. However, if the production from Police Point aquifer is less than two million gpd (gallons per day), the entire quantity will be withdrawn from the upper portion of the gravel deposit, leaving the lower portion as an emergency reservoir.

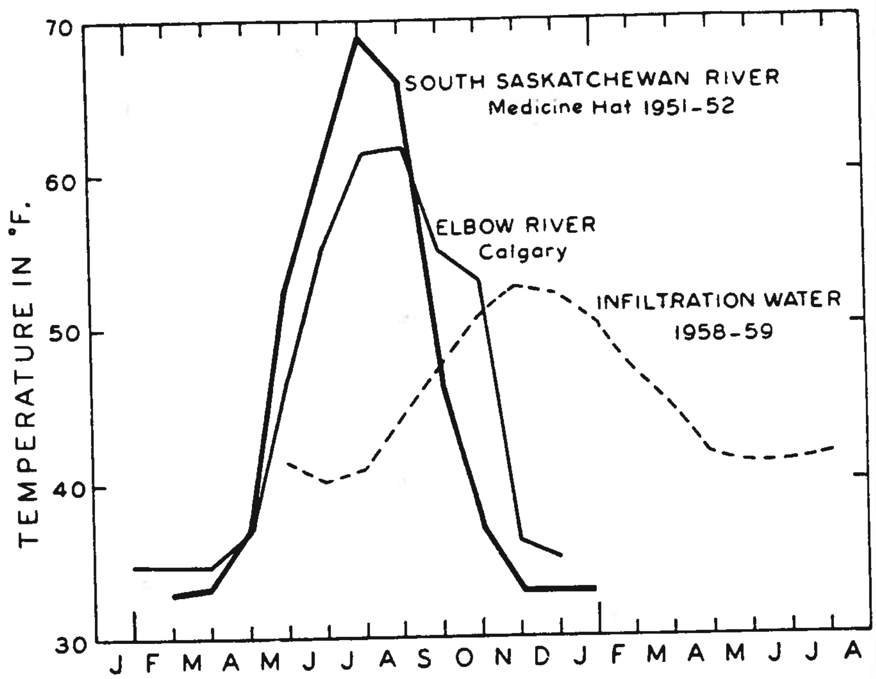


Figure 28. Temperature variations in groundwater and river water.

Economics of Induced Infiltration

Where there are two possible sources of water and either supply could be used, preference should be given to the one that is most suitable and can be developed most economically.

In the case of Medicine Hat, a nationally known rubber industry showed that its choice of a new plant site would be strongly influenced by the presence of a suitable water supply because large quantities of water are desirable for cooling purposes in the manufacture of rubber products. About 43 million gallons per month were required, the temperature at the source never to exceed 72°F. Furthermore, the water had to be clear, pure, and not excessively hard. In many instances in Alberta, raw surface water does not meet these requirements and needs expensive treatment to remove these impediments to its use.

The economic advantages of induced infiltration are obvious. Water obtained by induced infiltration into Police Point does not require treatment, has a maximum temperature well below the permissible limit, is available in large quantities, and can be supplied at low cost.

RATE OF GROUNDWATER RECHARGE NEAR DEVON, ALBERTA

by
R. N. Farvolden

Introduction

Knowledge of the rate of natural recharge to the groundwater reservoir is one of the prime requirements in the assessment of the groundwater resources of any area. In Alberta, the form of the piezometric surface and the behavior of shallow aquifers after long periods of pumping have led to the belief that significant recharge does take place, but no quantitative studies have been conducted previously.

The area discussed in this report was selected for study because the hydrologic and geologic conditions indicated that the necessary field work and test drilling could be done economically, and because the local landowners were extremely co-operative with regard to right-of-entry privileges. The soil cover, vegetation, topography, and drainage are typical for this region of Alberta. The results might therefore be considered as indicative of the rate of groundwater recharge over a much wider area than that actually studied.

Purpose and Scope of the Report

The purpose of this report is to indicate the results of a study of the rate of recharge to the groundwater reservoir in an area near Devon, Alberta. The calculations are based on the assumption that the flow of a spring, which appears to be relatively uniform throughout the year, is in equilibrium with the total recharge within the groundwater basin drained by the spring. The area of the groundwater basin and the rate of flow of the spring were determined and these figures allowed quantitative evaluation of the annual recharge.

Description of the Area and Physiography

The area involved in this study comprises a few square miles adjacent to the North Saskatchewan River about 10 miles southwest of Edmonton, Alberta (Fig. 2), and includes parts of townships 50 and 51, ranges 25 and 26, west of the Fourth Meridian (Fig. 29). The surface is flat to gently rolling, the local relief on the uplands area being about 25 feet. The drainage is fairly good although several shallow depressions are swampy and are flooded during wet seasons. The deep ditches of the present road system certainly have improved the drainage and perhaps the low ground was previously wet throughout the year.

The valley of the North Saskatchewan River forms the north and west boundaries of the area (Fig. 29). The valley is 200 feet deep and, for the most part, the banks are precipitous. The uplands area is drained by several short, intermittent streams which flow into the river. These streams have accordant junctions with the river and, near the river valley, flow in deep, narrow ravines with steep gradients. One such stream forms the southwest boundary of the area under study. The area is bounded on the east by Whitemud Creek which flows north and enters the North Saskatchewan River near the city of Edmonton.

A low ridge forms a drainage divide between the North Saskatchewan River and Whitemud Creek. Another low ridge runs along the east bank of the valley in the southwest corner of the area (Secs. 25 and 26, Tp. 50, R. 26) and a series of low hills along the north side of Sec. 36, Tp. 50, R. 26, and Sec. 31, Tp. 50, R. 25, completes the enclosure of a small drainage basin which is drained into the North Saskatchewan River by a short, deep ravine in the north half of Sec. 36, Tp. 50, R. 26.

The area is part of a rich mixed-farming district. Most of the land is cultivated and sown to grain or feed crops. The poorly drained depressions are utilized for pasture. In the uplands portion of the area there is likely less than one quarter-section of land still covered by native poplar trees and willow brush. The area might be described as typical of the parkland region of Alberta.

The area has a humid, continental, microthermal climate with warm summers and cold winters. The mean annual precipitation is 17 inches, of which about 5 inches fall in the form of snow. The coldest month is January, with an average mean daily temperature of 8°F and the warmest month is July with an average mean daily temperature of 60°F (Canada, 1957).

Frost penetration during the winter months averages 6 feet, so the surface of the ground is impermeable during the winter and early spring. Most of the spring runoff occurs while the ground is still frozen and therefore very little of this water infiltrates into the ground. The period during which recharge can occur is likely limited to that between the beginning of May and the end of October.

Map 83H/5 east half, of the Canada National Topographic Series, provides the best map coverage of the area.

Acknowledgments

This study could not have been conducted without the co-operation of the local residents who allowed the establishing of observation wells on private property. This co-operation is gratefully acknowledged.

Groundwater Geology

The Edmonton Formation

Strata of the lower member of the Edmonton Formation of late Cretaceous age underlie the area of this report (Ower, 1958) and crop out along the banks and valley walls of the North Saskatchewan River and the ravines near the river. The strata are composed mainly of bentonitic and silty shales of nonmarine origin and are over 800 feet thick. Coal seams are common in the lower Edmonton Formation in this region. In many places elsewhere in Alberta where it has been encountered in drilling operations, the Edmonton Formation contains sandstone strata which commonly occur as lenticular bodies within the thick shale section. The sandstone strata are commonly good aquifers although they are never continuous over a wide area. In the small area of this report, the thin coal seams are not sufficiently fractured to be suitable aquifers and sandstone strata are apparently absent.

All the farms in the area obtain water from wells and almost all of them have experienced some difficulty in this regard. Very few drill holes encounter either coal seams or sandstone strata of sufficient thickness and permeability to be suitable aquifers. Wells 400 to 500 feet deep commonly yield only one quart of water per minute if pumped continuously. The water from strata over 200 feet deep has a high alkalinity due to the presence of sodium bicarbonate. Because the alkalinity increases with depth, there is no possibility of obtaining potable water at depths of over 500 feet.

The regional structure map (Alberta, 1957) indicates a strike of north 45 degrees west and a dip of 30 feet per mile southwest, although a Devonian reef underlies the area and local reversals in dip have been established by oil-well logs. Experience has shown that neither the coal seams nor the sandstone strata are dependable marker beds and local structure is always difficult to determine in this region by surface surveys. Since it is of no importance in the present study, no effort was made to establish the local structure.

Surficial Deposits

Surficial deposits of Pleistocene age cover the entire upland portion of the area. Drift thicknesses of 22 to 118 feet were recorded in the 26 test holes that were drilled (Fig. 30). Most of the drill holes encountered a layer of soft, brown or grey clay about 20 feet thick, underlain by a compact, grey till. Thin beds of well-sorted, fine to very fine sand were encountered in many of the holes, in some places between the upper soft clay and the till, and in some places within the till portion of the section. The upper clay is likely lacustrine in origin and may be part of the deposits of glacial

Lake Edmonton (Hughes, 1958). It is soft, unctuous, and plastic, and has a rich brown color near the surface but is grey below about 20 feet, probably because that is the depth of oxidation. The till is commonly sufficiently well compacted to be mistaken for shale on the basis of drilling rate. No boulders and few pebbles larger than pea-size were encountered in the till in any of the borings.

Where sand was encountered in test drilling, it was mostly very fine grained and of the sort known to drillers as "quicksand." The mechanical analyses of five typical samples are shown graphically in figure 31. The sand strata are lenticular and range up to 10 feet in thickness.

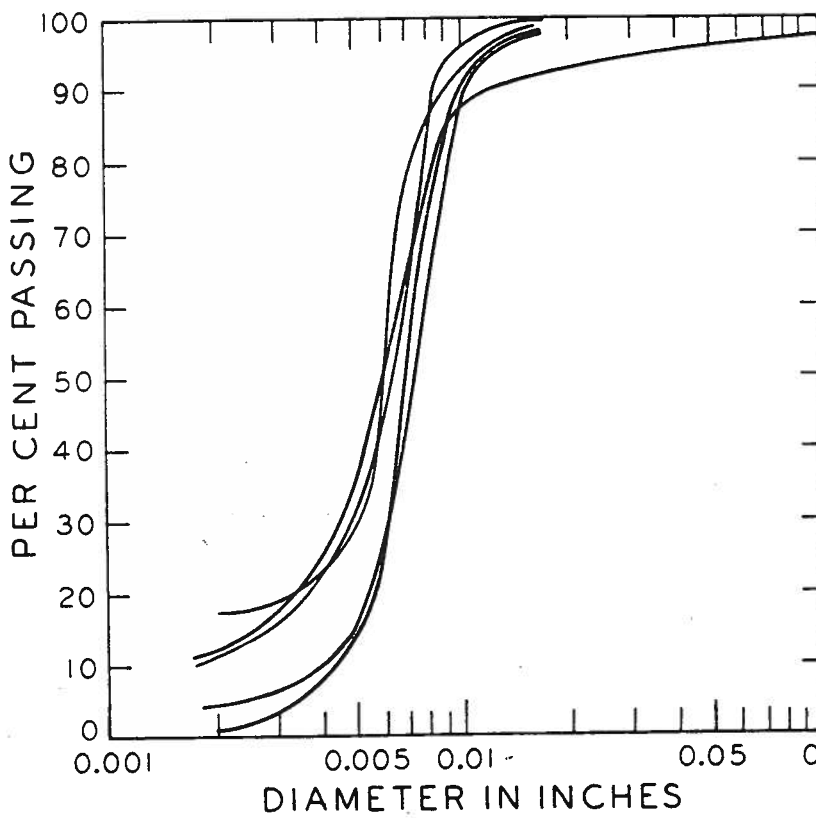


Figure 31. Mechanical analyses of sand samples from an area near Devon.

Locally, the search for water often leads to the construction of a shallow, large-diameter well to develop the fine sand beds of the surficial deposits. These attempts are usually only partially successful for, although there is abundant water in the sand to meet the requirements, improper design allows fine sand to enter the well and the result is either sand in the produced water or a nearly useless, sand-plugged well, or both. However, with proper techniques this type of unconsolidated sand can be developed as an aquifer (Farvolden, 1961a).

Groundwater Hydrology

Throughout most of Alberta the near-surface strata, considered on a regional basis, behave as a single hydrologic unit and the piezometric surface is a subdued replica of the surface topography. It follows that the surface-drainage basins and divides are also groundwater-drainage basins and divides. If the area and the discharge of any groundwater basin are known and if the discharge is in equilibrium with the recharge, then the rate of recharge to the groundwater reservoir can be determined, by equating recharge and discharge.

Attention was first focused on the area of this study by the presence of a spring that issues from the contact between the surficial deposits and the Edmonton Formation. Examination of the topography of the uplands in the immediate vicinity of the spring revealed the presence of the small surface-drainage basin already described, which drains into the North Saskatchewan River. The ravine that drains this small basin joins the river near the springs.

It was obvious that if the relation between the topographic and piezometric surfaces was the same as elsewhere in Alberta, there was an excellent opportunity to obtain quantitative data on the rate of groundwater recharge.

Groundwater Regimen

The piezometric surface was established (Fig. 29) using as control 22 observation wells and 4 shallow domestic wells. All of the observation wells were fitted with one-inch tubing, the bottom 2 feet of the tubing being perforated and filled with very coarse sand. In each observation well the perforated section of the tubing was placed across the sand strata that seemed to have the best characteristics as an aquifer. In test holes in which sand was not encountered, the perforations were set at the bottom of the drift section. The wells were measured periodically throughout the first five months of 1960. Each well showed both a rise and drop in the water level, which suggested that the wells were sufficiently sensitive for the purpose of this program.

The piezometric surface within this area corresponds closely to the topographic surface. The location of a groundwater divide depends upon the configuration of the piezometric surface. It is obvious that the contours on the piezometric surface could be drawn somewhat differently than shown on figure 29. This would cause a relocation of the groundwater divides and therefore a different figure for the area of the basin. However, with reasonable interpretation of the data, it is not possible to alter the area of the groundwater basin sufficiently to invalidate the final results.

From the form of the piezometric surface it is evident that recharge takes place throughout the area. A certain amount of the water that falls on the land surface as precipitation must pass through the soil horizon and zone of aeration and reach the water table.

The upper 500 or 1,000 feet of strata can be considered to be a single hydrologic unit, in which the horizontal permeability greatly exceeds the vertical permeability. In the recharge areas the hydrostatic pressure decreases with increasing depth and the movement of water below the water table is downward whereas, in the discharge area, water at depth will tend to move upward, toward the spring. As the water moves across the bedding planes, the flow lines are refracted at each stratigraphic interface (Hubbert, 1940, p. 846). The horizontal component of flow is greater than the vertical component, because the horizontal permeability exceeds the vertical permeability. A small but finite amount of water from the ground surface must enter and pass through the shale of the Edmonton Formation to recharge the poor aquifers occasionally found in that formation. In this way one can explain the close relation between the hydrostatic pressures in the surficial deposits and those in the Edmonton Formation, and the relation between the piezometric surface and the topography. The underlying Edmonton shale has extremely low permeability. Therefore, almost all the water that reaches the shale surface moves laterally along that surface in the direction of decreasing hydrostatic pressure, that is, toward the spring.

The spring marks the lowest point on the piezometric surface, and is the point of discharge. Water issues from several places along the bottom of a short, deep box canyon that is hardly noticeable on the topographic map. The bottom of this canyon is heavily wooded and a thick covering of moss and other vegetation obscures the exact location of the spring. It is most likely that the water issues from the contact between the surficial deposits and the Edmonton Formation.

The flow of the spring can be easily and accurately measured during the spring, summer and fall by means of measuring pails. During the winter months it was impossible to measure the flow with accuracy because of problems caused by ice in the channel. All evidence indicates that from May, 1959 to May, 1960, the spring flowed at a relatively steady rate of 28 $\pm$ 2 gpm (gallons per minute). The nature of the box canyon in which the spring is found is such that there is no possible error due to surface runoff for this is zero except immediately after a rain, and during the spring runoff.

Quantitative Evaluation of Recharge

The piezometric surface (Fig. 29) allows the determination of the area of the groundwater basin (the area receiving recharge) and the focus of the discharge (the spring). The only assumptions involved in evaluating the recharge are that the hydrologic system is in equilibrium and that the sum of the discharge of the spring and of pumpage by farm wells is equal to the recharge to the groundwater reservoir.

The total discharge of 30 gpm, made up of 28 gpm from the spring and 2 gpm pumpage by the farm wells, is equivalent to 2.52×10^6 cubic feet per year. The area enclosed by the groundwater divides is 2.9 square miles or 8.2×10^7 square feet by measurement of the basin on figure 29. The rate of recharge to the groundwater reservoir may then be calculated to be 0.031 feet per year, which is 2 per cent of the total annual precipitation.

Harder and Drescher (1954, p. 28) reported that, in Langlade County, Wisconsin, the average annual recharge over an area of 8.56 square miles was 0.45 feet per year. This is over 18 per cent of the total annual precipitation and compares reasonably well with the estimate of 10 to 12 per cent for the Chicago area given by Suter and co-workers (1959, p. 14). The recharge in Goshen County, Wyoming, was estimated to be 5 per cent of the annual precipitation of 14 inches or 0.058 feet per year (Rapp, Visher, and Littleton, 1957, p. 55). Babcock and Bjorklund (1956, p. 26) referred to an estimate by Morgan (1946) of 6 per cent of the annual precipitation or 0.07 feet per year for the recharge in the Cheyenne area of Wyoming. Keech and Dreeszen (1959, p. 47) computed an average recharge of 0.13 feet per year or about 6.5 per cent of the annual precipitation for a small area in Clay County, Nebraska, and stated that this was typical for the county. In another part of Nebraska, Newport (1959, p. 297) estimated that the recharge varies from 0.42 feet per year in an area of sand dunes to less than 1 inch per year for areas underlain by relatively impermeable rock.

Near Olds, Alberta, about 100 miles south of the area of this report, the annual recharge has been calculated to be about 0.04 feet or 3 per cent of the annual precipitation (Meyboom, 1960, personal communication). This figure was obtained using data from a flow net and a pumping-test determination of transmissibility.

In comparison, the rate of recharge calculated for the area of this report seems rather low. However, the result is thought to be valid and the low rate of recharge is attributed to:

(1) a low average annual precipitation,

(2) a long period during the winter in which no recharge can occur,

(3) slow melting of the ground frost in spring which prevents recharge during spring runoff,

(4) high evapotranspiration rates during the summer when recharge can take place, and

(5) relatively impermeable surface materials.

Conclusions and Recommendations

The annual rate of recharge to the groundwater reservoir over a small basin near Devon, Alberta, is 0.031 feet per year or about 2 per cent of the total annual precipitation. This rate may be representative of the rate of recharge over a much larger area of the Alberta "parkland" region, where similar conditions of climate, topography, geology, soil cover, and plant cover prevail.

In the area of study there is an excellent opportunity to use other methods to determine the recharge rate, and thereby perhaps develop and test suitable techniques for measuring the recharge rates in different areas. The rates in different areas may easily be determined where hydrologic conditions similar to those discussed in this report are found.

If the surface runoff and total precipitation can also be determined for the basins studied by this method, then reliable figures for evapotranspiration can be obtained. Investigations of this sort would lead to a better understanding of the groundwater regimen throughout Alberta.

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APPENDIX A. HYDROLOGIC TERMINOLOGY

by
W. A. Meneley

Introduction

The flow of liquids through porous media is governed by fundamental physical laws. Various expressions of these laws have been developed to suit the requirements of various disciplines and industries. These different expressions which, in essence, arise from differences in basic hydrologic units, lead to confusion which tends to inhibit the interchange of information. It is the purpose of this appendix to remove some of the confusion, firstly by giving explicit definitions of the fundamental hydrologic quantities and, secondly by stating the relations which hold among the units employed for these quantities.

Definitions

Permeability

The flow of liquids through a homogeneous porous medium may be expressed by Darcy's law, which in its differential form (Hubbert, 1940, p. 816) is:

$$q = -K \frac{\partial h}{\partial l} \tag{5}$$

where

q = specific discharge in the direction of flow

K = constant of proportionality which includes the coefficient of permeability

$\frac{\partial h}{\partial l}$ = potential gradient in the direction of flow

Hubbert (*op. cit.*, p. 816) demonstrated that the coefficient K can be expressed in terms of four components.

$$K = \frac{\rho g}{\eta} k \tag{6}$$

where

ρ = density of the fluid

η = viscosity of the fluid

g = acceleration due to gravity

k = a property of the medium alone pertaining to its ability to transmit water and defined by Hubbert as the coefficient of permeability. It has the dimensions of length squared (L^2)

The general expression for Darcy's law now becomes

$$q = \frac{-\rho g}{\eta} k \frac{\partial h}{\partial l} \tag{7}$$

Groundwater flow involves the movement of a single fluid, namely water, and it has been the accepted usage to include the properties of the fluid, as well as those of the medium, within the expression for permeability.

The "coefficient of permeability" employed in groundwater literature is denoted by P, and Darcy's law governing one-dimensional flow may be written:

$$Q = PIA \tag{8}$$

where

Q = discharge per unit time across the cross-sectional area A

P = permeability, dependent not only upon the coefficient of permeability k, but also upon the density and viscosity of the fluid. It has the dimensions of length divided by time (LT^{-1})

I = the potential gradient

A = cross-sectional area perpendicular to the direction of flow

The values of P obtained using equation (8) may be caused to vary radically simply by varying the temperature at which the experiment is performed, without altering any other factor. This is occasioned by the variation in fluid properties, chiefly the viscosity, with temperature. To take this into account, it is necessary to rewrite equation (8):

$$Q = \frac{\rho}{\rho_0} \frac{\eta_0}{\eta} \frac{g}{g_0} PIA \tag{9}$$

where ρ_0, η_0 and g_0 are the values of ρ, η and g under certain specified standard conditions.

When dealing with the flow of water under the usual range of physical conditions, the variations in ρ and g may be neglected, so that:

$$Q = \frac{\eta_0}{\eta} PIA \tag{10}$$

Some indication of the significance of the variation of the viscosity with temperature is given by the graph in figure 32. The graph may also be used for correcting permeability values obtained at conditions other than those specified to be standard.

The unit of permeability employed by the United States Geological Survey is defined by Meinzer as "the rate of flow (of water) in gallons a day, through a cross-sectional area of one square foot, under a hydraulic gradient of 100 per cent, at a temperature of 60°F (Stearns, 1928, p. 148)". Fishel (1946, p. 259) proposed that this coefficient of permeability be designated the *meinzer*.

The meinzer is equivalent to one U.S. gallon per day per square foot, and is, therefore, not the same thing as the fundamental unit of permeability in countries of the British Commonwealth, which is one imperial gallon per day per square foot.

European groundwater practice is also to consider the permeability as a function of the fluid as well as of the medium. It is defined as the volume of water at 15.4°C (59.72°F) that flows across a unit cross section of the medium under a unit hydraulic gradient (Prinz, 1923, p. 129, p. 167). It generally has the units in meters per day or centimeters per second.

The unit of permeability used in the petroleum industry is the darcy, defined as the flow in cubic centimeters per second of a fluid of unit viscosity through a cross section of one square centimeter under a pressure gradient of one atmosphere per centimeter (Muskat, 1946, p. 76). It is a quantity dependent solely on the properties of the medium.

Table 14 lists the factors to be used in converting permeability in any one of the units described above into permeability in any other system of units.

Field Permeability

The field permeability is the permeability of the medium, determined and applied under the prevailing field conditions. No correction is required for fluid viscosity variations in this case. If, however, it is necessary to convert to permeability units in a different system of measurement or if field permeability values obtained at different temperatures are to be compared, viscosity must then be taken into account. Wenzel (1942, p. 62) proposed the following temperature correction, which is derived from equation (10).

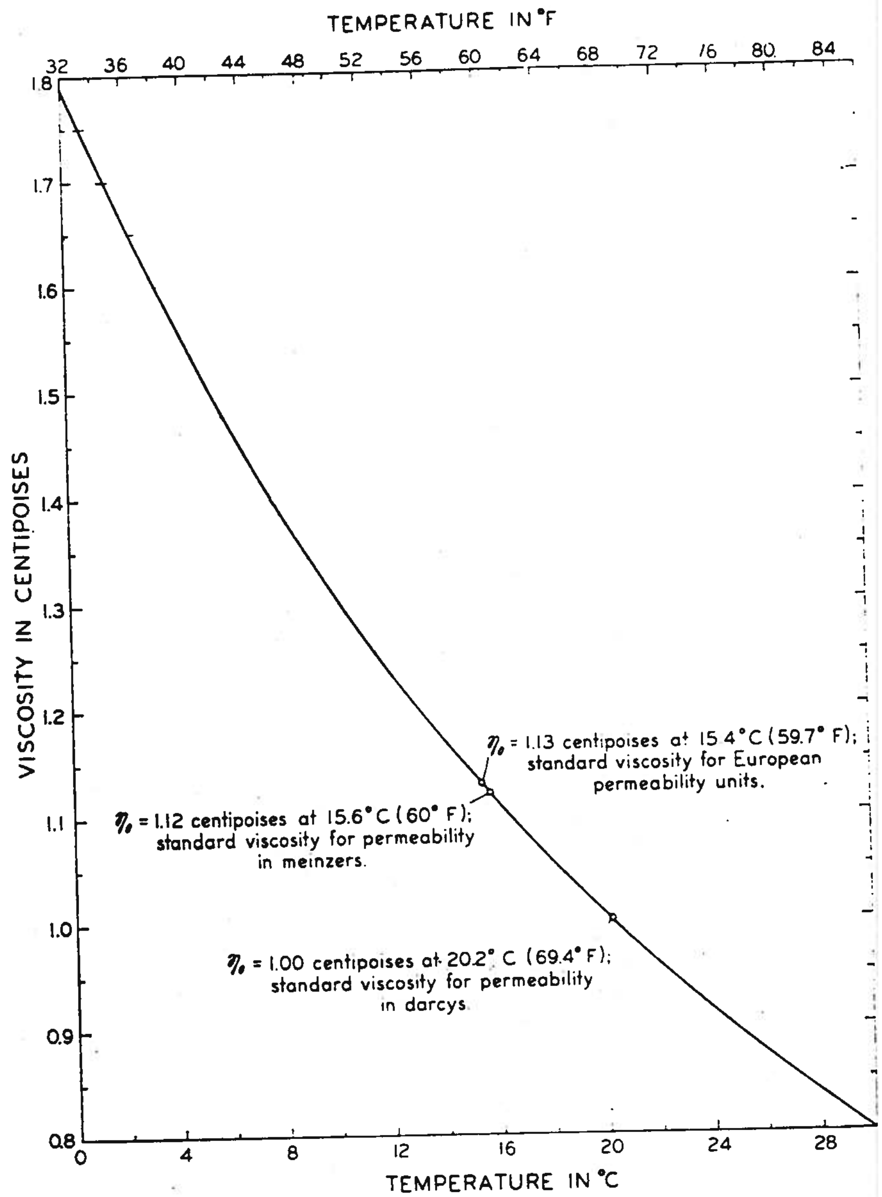


Figure 32. Variation in the viscosity of water with temperature.

$$P = \frac{\eta}{\eta_0} P_f \tag{11}$$

where

P_f = permeability at the prevailing field conditions

In practice, reported values of permeability represent the field permeabilities, unless otherwise indicated; the subscript is generally omitted.

Transmissibility

The coefficient of transmissibility or, simply, the ***transmissibility***, was defined by Theis (1935) as the rate of flow of water, at the prevailing groundwater temperature, in gallons a day, through a vertical strip of aquifer one foot wide, extending the full vertical height of the aquifer, under a unit hydraulic gradient. It is, therefore, the product of the field permeability and the thickness of the aquifer,

$$T = P_f m \tag{12}$$

where

T = transmissibility

m = thickness of the aquifer

Transmissibility has the dimensions L^2T^{-1}. It is analogous to similar terms in European groundwater and petroleum industry usage, expressing the product of the permeability and a given transmitting thickness. In Canadian groundwater studies, transmissibility is expressed as imperial gallons per day per foot, whereas in European terminology it is generally expressed as square meters per day.

Storage Coefficient

The storage coefficient S is defined as the volume of water released from, or taken into, storage from a prism of the aquifer having a base of unit dimensions and a height equal to the thickness of the aquifer, per unit change in the fluid potential normal to the surface of the aquifer (Theis, 1938). It was shown by Jacob (1940) that the storage coefficient is the sum of three components:

$$S = \rho g \theta m \left[\frac{1}{E_w} + \frac{b}{E_s} + \frac{c}{E_c} \right] \tag{13}$$

where

θ = porosity

Table 14. Conversion Table for Permeability Units

Unit = $\dfrac{\text{Volume}}{\text{area} \times \text{time} \times \text{potential grad.}}$	$\dfrac{ft^3}{ft^2 \times day \times (ft/ft)}$ at 60°F	$\dfrac{\text{imperial gallons}}{ft^2 \times day \times (ft/ft)}$ at 60°F	$\dfrac{\text{U.S. gallons}}{ft^2 \times day \times (ft/ft)}$ at 60°F (meinzer)	$\dfrac{m^3}{m^2 \times day \times (m/m)}$ at 15.4°C	$\dfrac{cm^3}{cm^2 \times sec \times (cm/cm)}$ at 20.2°C	$\dfrac{cm^3}{cm^2 \times sec \times (atm/cm)}$ at 20.2°C (darcy)
$\dfrac{ft^3}{ft^2 \times day \times (ft/ft)}$ at 60°F	1.00	6.23	7.48	3.04×10^{-1}	3.96×10^{-4}	4.10×10^{-1}
$\dfrac{\text{imperial gallons}}{ft^2 \times day \times (ft/ft)}$ at 60°F	1.60×10^{-1}	1.00	1.20	4.89×10^{-2}	6.36×10^{-5}	6.59×10^{-2}
$\dfrac{\text{U.S. gallons}}{ft^2 \times day \times (ft/ft)}$ at 60°F	1.34×10^{-1}	8.33×10^{-1}	1.00	4.07×10^{-2}	5.30×10^{-5}	5.49×10^{-2}
$\dfrac{m^3}{m^2 \times day \times (m/m)}$ at 15.4°C	3.28	2.04×10	2.46×10	1.00	1.31×10^{-3}	1.35
$\dfrac{cm^3}{cm^2 \times sec \times (cm/cm)}$ at 20.2°C	2.51×10^{3}	1.56×10^{4}	1.88×10^{4}	7.66×10^{2}	1.00	1.03×10^{3}
$\dfrac{cm^3}{cm^2 \times sec \times (atm/cm)}$ at 20.2°C	2.44	1.52×10	1.82×10	7.42×10^{-1}	9.67×10^{-4}	1.00

E_w = bulk modulus of elasticity of water

E_s = bulk modulus of the aquifer skeleton

E_c = bulk modulus of the intercalated clay strata

b = the negative ration of a given change in compressive stress to the change in hydrostatic pressure producing it. It ranges in magnitude from the porosity of the medium to unity, depending on the nature of the aquifer, particularly its degree of cementation

c = a dimensionless quantity that depends upon the thickness, configuration, and distribution of the intercalated clay beds

and the other units are as previously defined.

It is apparent that if a significant portion of the water derived from storage in an artesian aquifer is obtained by consolidation of intercalated shale strata, then the storage coefficient will vary with time, depending upon the rate of consolidation. For practical purposes, however, the rate of change of the storage coefficient decreases rapidly with time, and a reasonably representative value for the coefficient may be determined after pumping has continued for a sufficient time. This time will depend upon the permeability of the aquifer and the location of the observation wells.

Porosity

The porosity of a rock is the ratio of the volume of the void spaces to the total volume of the rock. It may be expressed as a decimal fraction or as a percentage.

Specific Yield and Specific Retention

The specific yield is the volume of water obtained from a unit volume of the medium by complete gravity drainage, expressed as a decimal fraction, or as a percentage. The specific retention is the volume of water retained in the rock by capillary attraction and molecular adhesion after gravity drainage has been completed. The sum of the specific yield and the specific retention is equal to the porosity.

Conversion between Systems of Units

The differences among the fundamental permeability units in the various systems of units in common use have been previously discussed and a table (Table 14) provided giving the conversion factors. Of the other hydrologic quantities that have been mentioned, only transmissibility presents some additional problem when conversion is desired. In this case, however, the additional problem is simply one of converting one length unit to another and the fact that the systems of units involved may have different standard temperatures causes no difficulty. The permeability factor in the coefficient of transmissibility is, of course, adjusted by using table 14.

The other hydrologic quantities, such as storage coefficient and porosity are dimensionless and, therefore, remain the same regardless of the system of units employed.

APPENDIX B. MATHEMATICAL MODELS

by
W. A. Meneley

Introduction

A number of mathematical models have been developed that can be used to determine the transmissibilities and storage coefficients of artesian aquifers. Some of these may be applied with discretion to the estimation of the aquifer coefficients of unconfined aquifers. All have appeared in textbooks on hydrology, or in technical publications. For this reason only those working equations that are common to more than one of the reports in this bulletin are given in this appendix. Others are developed or quoted in the reports in which they are used.

All of the mathematical models described here are based on the following assumptions:

(1) the aquifer is homogeneous, isotropic, and of infinite areal extent,

(2) the well penetrates and receives water from the entire thickness of the aquifer,

(3) the pumping well has an infinitesimal diameter,

(4) water is discharged instantaneously with a decline in head, and

(5) the transmissibility is constant at all times, and places.

While these assumptions are never completely satisfied in real aquifers, the models commonly provide reasonable approximations to the aquifer coefficients, and are widely applied to the solution of groundwater problems.

Nonequilibrium Equation

Theis (1935) developed an equation to express the nonsteady state flow of groundwater in the vicinity of a discharging or recharging well. This equation was developed from the analogy between heat flow and groundwater flow. Jacob (1940) derived the same expression from hydraulic concepts.

The Theis equation may be written

$$s = \frac{Q}{4\pi T} W(u) \qquad (14)$$

where

$$u = \frac{r^2 S}{4Tt} \qquad (15)$$

$$W(u) = \int_{u}^{\infty} \frac{e^{-x}}{x}\, dx$$

$$= -\ .577216 - \ln u + u - \frac{u^2}{2.2\,!} + \frac{u^3}{3.3\,!} - \frac{u^4}{4.4\,!} + \cdots\cdots \qquad (16)$$

s = drawdown or buildup at the point of observation

Q = rate of discharge or recharge at the discharging or recharging well

T = coefficient of transmissibility

S = storage coefficient

t = time since start of discharge or recharge

r = distance from discharging or recharging well to point of observation

ln u = logarithm of u to the base e

W(u) is usually called the exponential integral by mathematicians, but in hydrologic usage it has become known as the well function of u. A useful table giving values of W(u) for values of u ranging from 10^{-15} to 9.9 was prepared by Wenzel (1942) and has been reproduced in a number of other publications (Bruin and Hudson, 1955; Todd, 1959; Wisler and Brater, 1959).

Equation (14) is an expression of the dependence of the variation of the drawdown, with time and distance, on the aquifer constants T and S. Because of the way in which T and S are contained in equation (14), however, no direct analytical solution for them is possible. It is necessary, instead, to use a method of graphical superposition developed by Theis (1935) and subsequently described by Wenzel (1942), Wisler and Brater (1959), and Todd (1959).

Modified Nonequilibrium Equation

Cooper and Jacob (1946) recognized that, as time increases or distance decreases, the numerical value of u decreases and that, when u is equal to or less than 0.02, the terms beyond ln u in the series for W(u) (Eq. 16) are no longer significant. Equation (14) may then be simplified to:

$$s = \frac{2.30\ Q}{4\pi T} \log \frac{2.25\ Tt}{r^2 S} \qquad (17)$$

where the notation "log" denotes the logarithm to the base ten.

The drawdowns s_1 and s_2 in a given observation well, at times t_1 and t_2 respectively, satisfy the relation

$$s_2 - s_1 = \frac{2.30\ Q}{4\pi T} \log \frac{t_2}{t_1} \tag{18}$$

If t_1 and t_2 are selected one log cycle apart or, in other words, so that t_2 is ten times t_1

$$s_2 - s_1 = \frac{2.30\ Q}{4\pi T} \tag{19}$$

Writing $\Delta s = s_2 - s_1$, and solving for T, the modified nonequilibrium equation is obtained

$$T = \frac{2.30\ Q}{4\pi \Delta s} \tag{20}$$

The storage coefficient may be determined from equation (17) in the form it assumes at time t_0, that is, at the time for which drawdown is zero:

$$S = \frac{2.25\ Tt_0}{r^2} \tag{21}$$

Theis Recovery Equation

The aquifer coefficients may also be determined from observations of the rate of recovery of the water level in a well after pumping ceases. The equation governing recovery is derived by considering that the discharging well continues to function, but that a recharge well of equal magnitude begins to operate at the same location at the time that pumping stops. The residual drawdown s' is given by:

$$s' = \frac{Q}{4\pi T} [W(u) - W(u')] \tag{22}$$

where

$$u' = \frac{r^2 S}{4Tt'}$$

t' = the time since pumping stopped

and the other symbols are as previously defined.

When sufficient time has elapsed so that both u and u' are less than, or equal to 0.02

$$s' = \frac{2.30\ Q}{4\pi T} \log \frac{t}{t'} \tag{23}$$

If two time-ratios are selected one log cycle apart

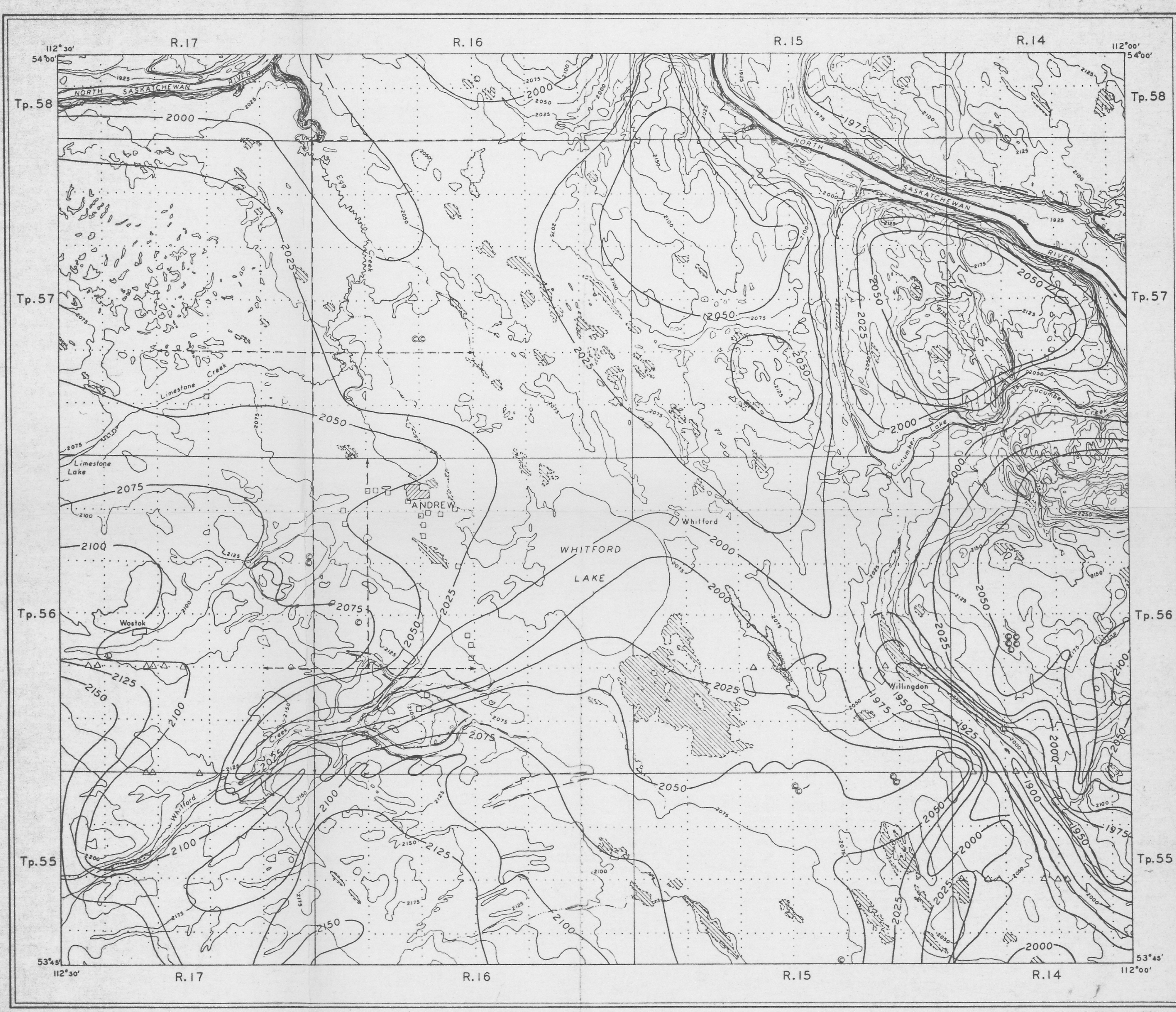
$$T = \frac{2.30\ Q}{4\pi \Delta s'} \tag{24}$$

where

$\Delta s'$ = the change in residual drawdown during one log-cycle change in time-ratio.

Other Mathematical Models

Besides the equations described above, dealing with drawdown and recovery in an infinite nonleaky artesian aquifer, others are available which treat drawdown in a water table aquifer (Boulton, 1954a, 1954b) or in a leaky artesian aquifer (Hantush and Jacob, 1955; Walton, 1960). The latter is an aquifer that is bounded, either above or below, by a semipermeable bed. The solutions for water table and leaky artesian conditions also assume the aquifers to be infinite in areal extent. Aquifers with boundaries may be analyzed by replacing the boundaries by a system of image wells in an infinite aquifer that will give the same effect at the boundary locations as the boundaries do in the real aquifer.



To accompany Research Council of Alberta Bulletin 12

Published 1963

FIGURE 3

BEDROCK TOPOGRAPHY
ANDREW AREA, ALBERTA

WEST OF FOURTH MERIDIAN

SCALE IN MILES

1 0 1 2 3

CONTOUR INTERVAL 25 FEET

Elevations in Feet above Mean Sea Level

LEGEND

- Water well, lithologic log available
- Water well, chemical analysis of water available
- Seismic shot hole, drilled into sand and gravel
- Seismic line
- Resistivity station
- Contour, bedrock surface 2125
- Stream trench

LEGEND

- Lake, intermittent, indefinite
- Stream, intermittent
- Contour, surface elevation 2125
- Contour, depression

Base map compiled from Willingdon sheet No 83 H/I East and West, Department of Mines and Technical Surveys 1958

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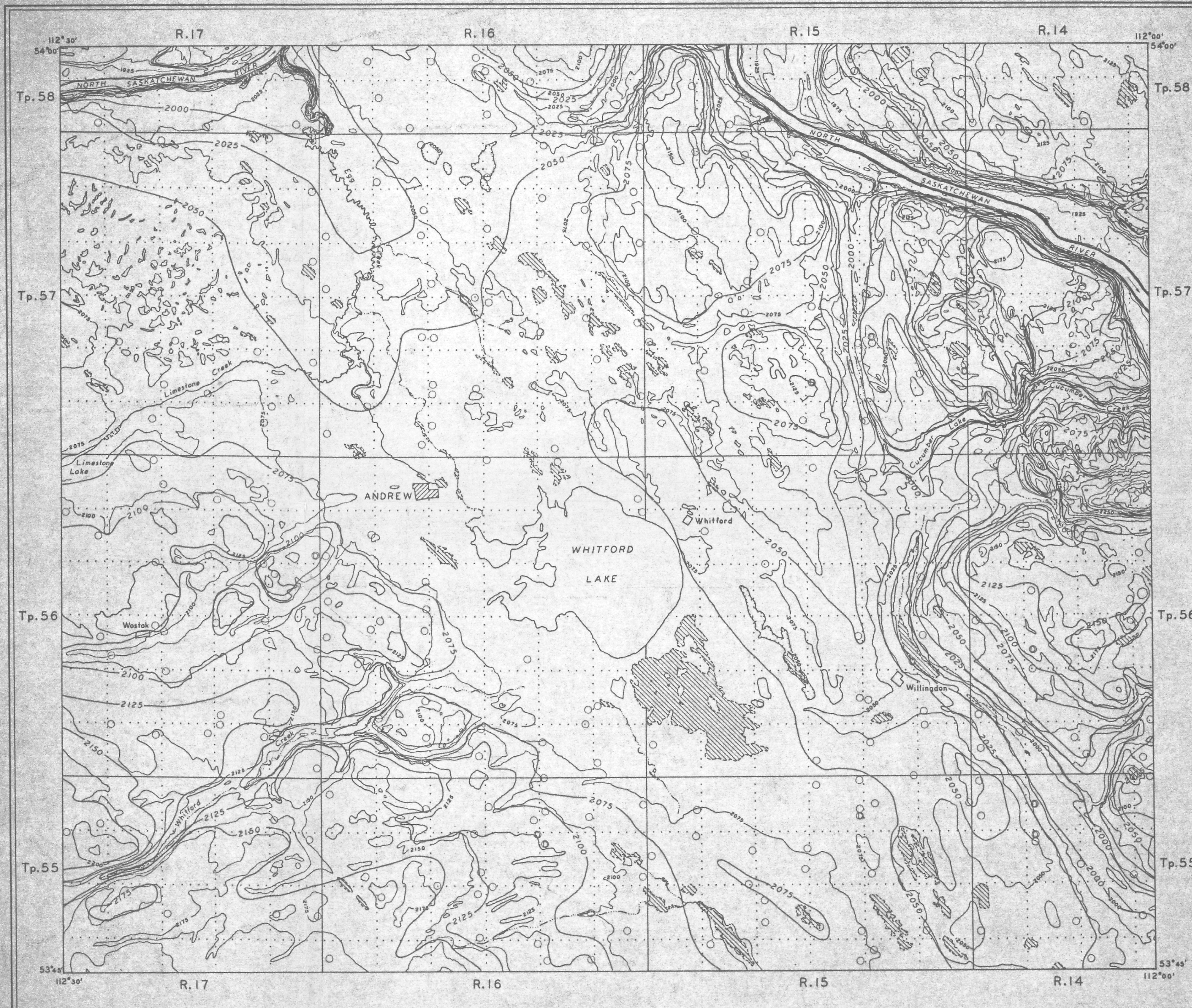

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FIGURE 4

PIEZOMETRIC SURFACE

ANDREW AREA, ALBERTA

WEST OF FOURTH MERIDIAN

SCALE IN MILES

Water well, dug or bored

Water well, drilled

Contour, piezometric surface 2075

LEGEND

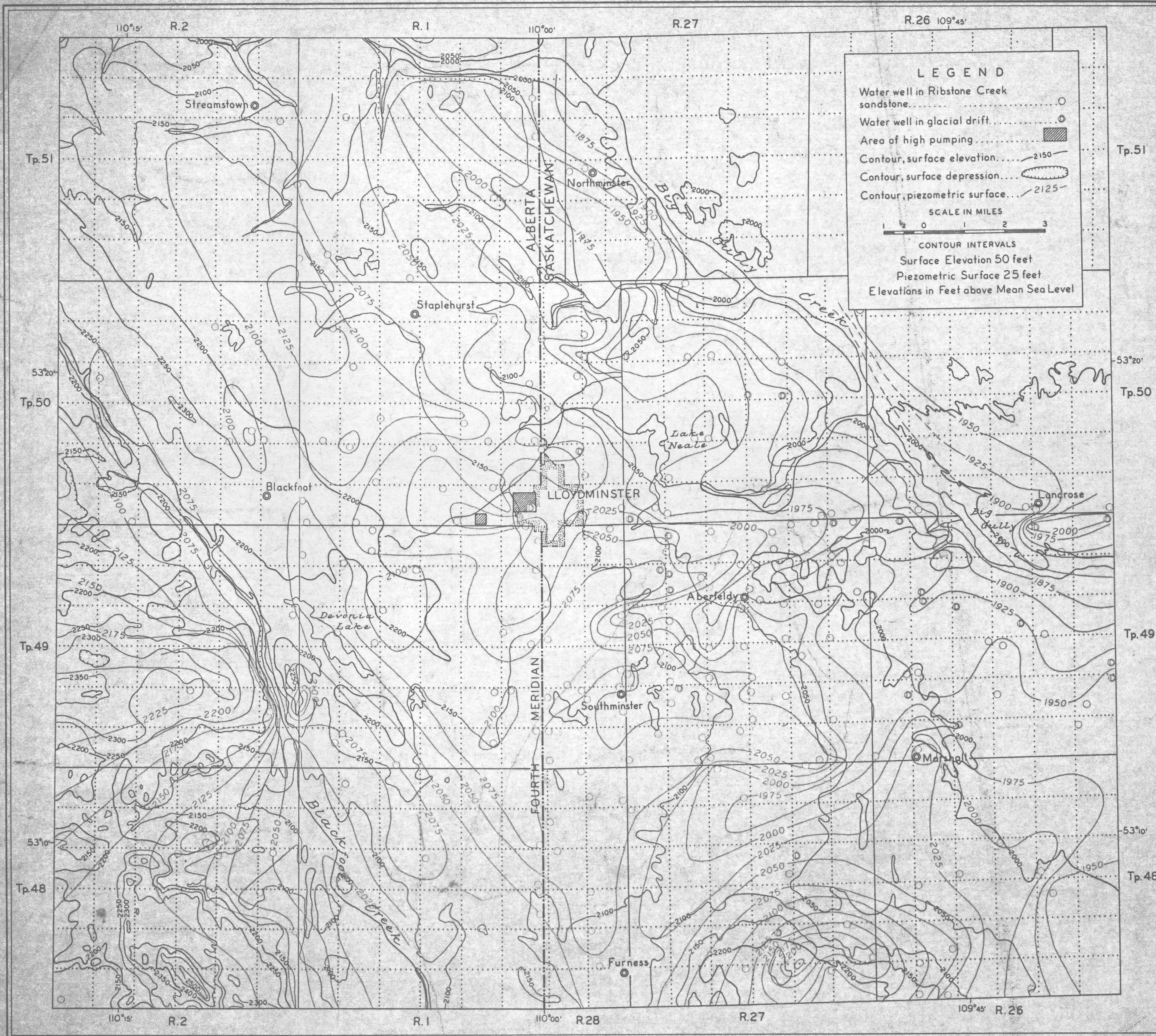
Lake, intermittent, indefinite

Stream, intermittent

Contour, surface elevation 2125

Contour, depression

Base map compiled from Willingdon sheet Nº83 H East and



LEGEND
Water well in Ribstone Creek sandstone
Water well in glacial drift
Area of high pumping
Contour, surface elevation 2150
Contour, surface depression
Contour, piezometric surface 2125
SCALE IN MILES
1 ½ 0 1 2 3
CONTOUR INTERVALS
Surface Elevation 50 feet
Piezometric Surface 25 feet
Elevations in Feet above Mean Sea Level
110°15'
R.2
R.1
110°00'
R.27
R.26
109°45'
R.28
Tp.51
Tp.50
Tp.49
Tp.48
53°20'
53°10'
Streamstown
Northminster
Staplehurst
Blackfoot
LLOYDMINSTER
Aberfeldy
Southminster
Marshall
Landrose
Furness
Lake Neale
Devonia Lake
Big Gully Creek
Blackfoot Creek
ALBERTA
SASKATCHEWAN
FOURTH MERIDIAN

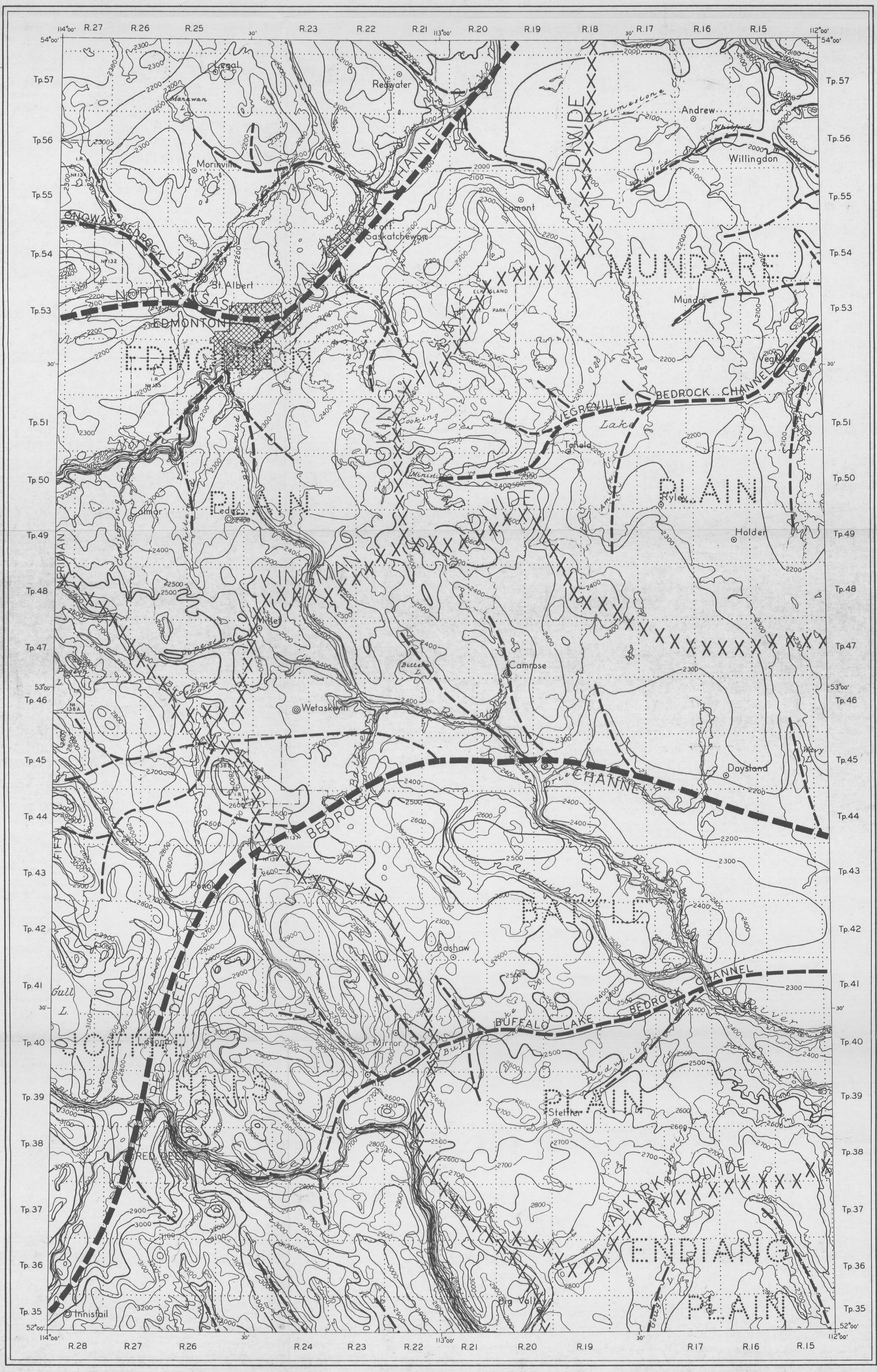


FIGURE 14. BEDROCK TOPOGRAPHY, EDMONTON-RED DEER MAP AREA, ALBERTA

EXPLANATION

Bedrock channel, center line

Bedrock physiographic boundary XXX

Contour, bedrock surface 3000

SCALE IN MILES

6 4 2 0 6 12

CONTOUR INTERVALS 100 FEET

Elevations in Feet above Mean Sea Level

REFERENCE

City; Town; Village

Stream, intermittent

Township boundary

Contour, surface elevation 2500

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