Report 73-3

CHEMISTRY AND HYDROLOGY OF GROUNDWATER IN THE VICINITY OF WATERTON RESERVOIR, ALBERTA

A. Vanden Berg and K. W. Geiger

Alberta RESEARCH

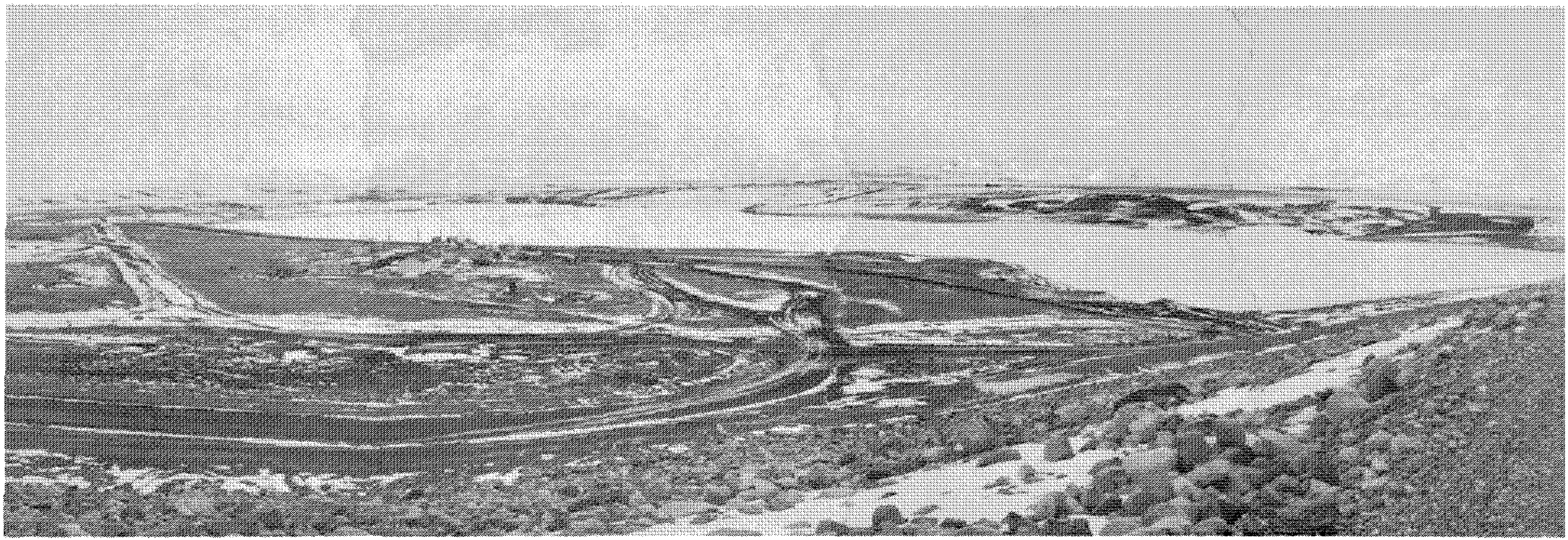

FIGURE 1. General view of Waterton Reservoir from the northeast.

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CHEMISTRY AND HYDROLOGY OF GROUNDWATER IN THE VICINITY OF WATERTON RESERVOIR, ALBERTA

Abstract

This report describes the influence of the filling of Waterton reservoir in the spring of 1965 on the chemistry and hydrology of groundwater in the vicinity of the reservoir, and attempts to estimate the amount of underground outflow from the reservoir through the Cochrane Valley aquifer – a buried bedrock valley aquifer hydraulically connected to the reservoir – into the Belly River.

Prior to the filling of the reservoir, samples of groundwater were obtained from wells and springs in an area of approximately 120 square miles and analyzed for the major constituents. Statistical analysis and trend-surface fitting of the chemical data showed that fresh groundwater predominates in the unconsolidated sediments of the topographically low areas surrounding the major water courses.

Two years after the filling of the reservoir a second set of groundwater samples was obtained from the same sources and analyzed. Although it could be shown that, on the average, a statistically significant change to a fresher groundwater had occurred, the spatial distribution of the change contradicts the assumption that it is caused by the filling of the reservoir; it is shown that the change is likely the result of an increase in precipitation in the years between the two samplings.

The hydrological investigation has been limited to the aquifer in the Cochrane buried bedrock valley; the aquifer, which reaches a thickness of over 90 feet locally, consists predominantly of clean gravel with occasional thin sand lenses and is covered by approximately 100 feet of till.

The rise of the level of the reservoir to its operating level during the spring of 1965, as well as later fluctuations, was followed almost instantaneously by a rise in the hydraulic head in the aquifer, which was monitored by a set of 5 piezometers.

From the measured average gradient of the hydraulic head and the estimated total overflow through flowing wells and springs, an average hydraulic conductivity of 2,755 igpd/ft (imperial gallons per day per foot) and an average transmissibility of 143,000 igpd/ft were calculated; these figures are close to those determined from a pump test.

In 1969 a number of large relief wells had been drilled to lessen the pressure in the aquifer and the danger of uncontrolled outflow; the combined free flow from these wells and other flowing wells in the area was 4,750 gallons per minute.

A comparison of the average hydraulic gradients in the portion of the aquifer between the reservoir and the flowing wells in 1965 and 1969 indicated that a sizeable portion of the total outflow was not derived from the reservoir, but supplied by leakage from the till or bedrock or both; it is further concluded that there is no significant underground outflow into the Belly River.

INTRODUCTION

Although plans for the construction of Waterton Dam (Fig. 2) were initiated by the Prairie Farm Rehabilitation Administration (PFRA) in the 1950's, it was not until 1962 that the Research Council of Alberta became aware of the project. Mapping of the bedrock topography in the area (Geiger, 1965) had revealed the existence of a buried bedrock channel containing permeable gravel deposits which would intersect the proposed reservoir on the northeast, almost exactly where the dam was to be built. From this point the buried valley — the Cochrane Valley — was known to trend eastward across the divide between the Waterton and Belly Rivers, joining the latter approximately 5 miles eastward. As the permeable deposits in the channel lie at a much lower elevation than the designated reservoir level, a sharp rise of the hydrostatic pressure in these deposits could be foreseen. Such an increase could be expected to cause outbreaks of groundwater to the surface through permeable zones and fractures in the overburden and through existing wells in areas where the surface elevation is less than the reservoir level; these outbreaks could furthermore occur at such high rates as to make them difficult to control.

Secondly, the possibility existed that large quantities of water from the reservoir would leak through the channel into the Belly River system to the east, with possible adverse effects on the maintenance of useful levels in the reservoir.

Officials in the PFRA were contacted and a combined study of the situation was initiated, in which the Research Council of Alberta agreed to undertake the installation of several piezometers into the buried channel, to carry out a survey of wells in the area, and to obtain samples of groundwater for chemical analysis both prior to and two years after the completion of the reservoir. The purpose of this study was to observe any changes caused by the presence of the reservoir to the amount and configuration of groundwater flow and to the composition of the groundwater, particularly in the buried channel aquifer.

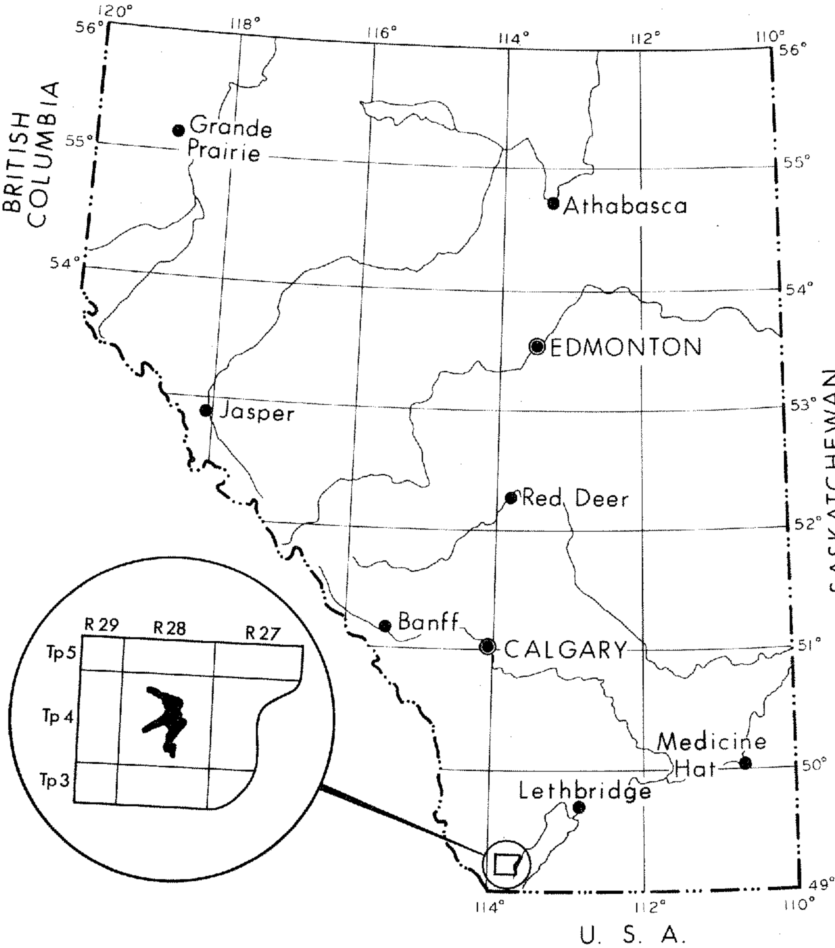


FIGURE 2. Location of Waterton Dam and study area.

Five piezometer nests were installed during 1964, each nest consisting of a shallow well to approximately 40 feet below the surface for observation of the depth of the water table, and a deep well to the top of the bedrock to measure the hydraulic pressure in the gravel aquifer. All piezometers were located in the area occupied by the buried valley and all penetrated from 30 to 90 feet of gravel and sand.

Waterton Reservoir began to fill on March 18, 1965, at a level of 3,773.3 feet above sea level, and reached its peak at 3,876.3 feet on June 18. Water levels in the deep piezometers and in those domestic wells that were completed in the Cochrane Valley aquifer began to rise almost instantaneously; a number of farm wells and deep piezometers P3, P4, and P5 started to flow

(Fig. 3a), and a spring developed in the middle of a small slough on the property of Mr. H. Wolf (NE 1/4, Sec. 32, Tp. 4, R. 27, W.4 Mer.) (Fig. 3b). The casing of one farm well was forced out of the hole by the pressure increase and the well flowed uncontrolled.

In order to prevent any additional problems that might arise from high pressures in the aquifer, PFRA undertook the construction of a number of pressure-relief wells, located in the lowest parts of the area and designed for large flow rates. The first of these wells was completed in February 1967, and at present eight wells have been successfully completed; most (Fig. 4) are located around Cochrane Lake into which they discharge. Their combined output averages 5,400 igpm (imperial gallons per minute). The resulting cone of depression in the piezometric surface is approximately 50 feet deep and several of the flowing wells in the area have since stopped flowing.

As one of the main objectives of the study was to determine the amount of water flowing out of the reservoir through the channel, the fieldwork on the project was concluded in August 1969 with a pump test to determine the aquifer constants. Test wells W1 and W2 (Fig. 4) were constructed for this purpose. The transmissibility of the aquifer was determined as 140,000 igpd/ft and the permeability as approximately 2,700 igpd/ft^2 (imperial gallons per day per square foot).

Location and Extent of the Area

The Waterton Reservoir is located at the confluence of Waterton River and Drywood Creek at latitude 49° 18′ north and longitude 113° 41′ west (Figs. 2, 5). On the basis of the Dominion Land System of survey it lies in township 4, range 28, west of the fourth meridian. The area over which groundwater samples and well data were collected is bounded on the east by the Belly River and meridian of longitude 113° 32′, on the west by meridian of longitude 113° 50′, and by parallels of latitude 49° 13′ and 49° 23′ north.

Topography and Drainage

The area is part of the Foothills belt of Alberta and lies approximately 20 miles east of the eastern limits of the Canadian Rockies. The surface slopes generally eastward (Fig. 5); the highest point is on the northern flank of Pine Ridge in the extreme southwest corner of the area, where the surface rises to 4,650 feet above mean sea level; the lowest point is in the Belly River valley in the northeast corner at an altitude of 3,600 feet.

The area is almost bisected by the Waterton River, flowing in a northerly direction. To the west of the Waterton River the terrain is strongly undulating, affected profoundly by the consequent preglacial bedrock valley and intervening ridges of the Foothills belt; the most prominent of these features are Pine Ridge,

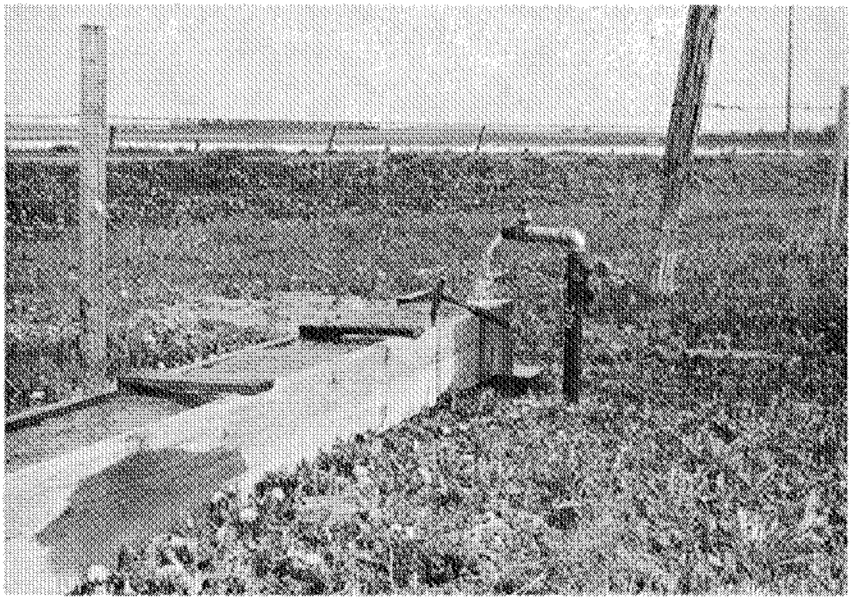


FIGURE 3. (a) Flowing piezometer P3 used by farmer to water stock.

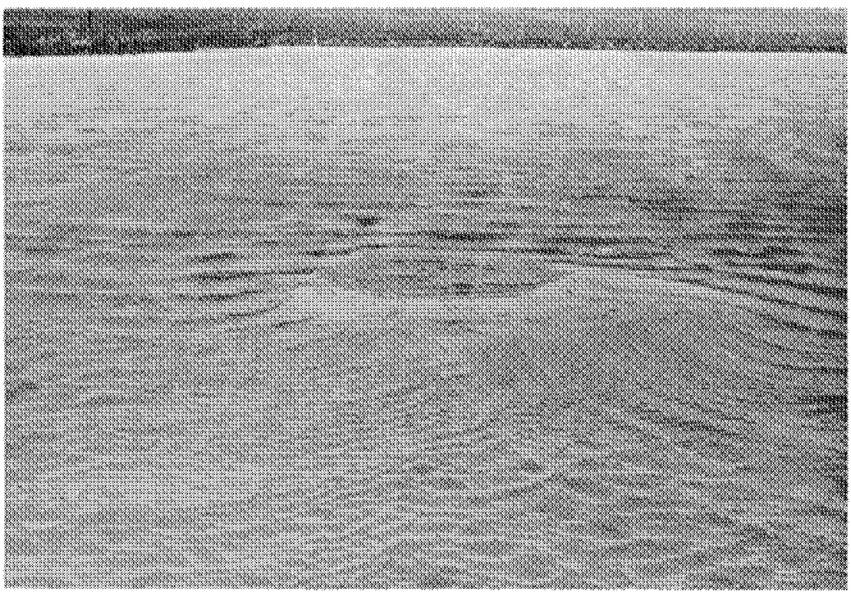
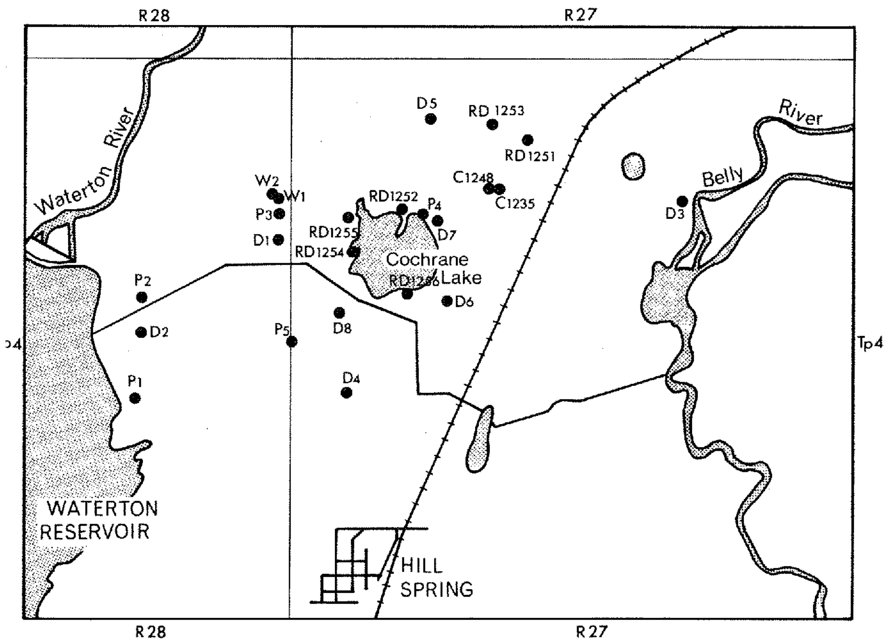


FIGURE 3. (b) Mud boil in slough caused by spring discharge.



LEGEND

C PFRA cable-tool hole

RD PFRA rotary hole

D Domestic well

P RCA (1965) piezometer

W RCA (1969) test well

FIGURE 4. Location of wells and testholes in Cochrane Valley aquifer.

Hatfield Hill, and the deeply incised valley of the Drywood Creek. To the east of the Waterton River the country is flat to very gently rolling, a reflection of the change from Foothills to Plains topography. It has an average eastward slope of 50 feet per mile; the only prominent topographic feature, west of the village of Hill Spring, is a hill which rises 250 feet above the plain to an altitude of 4,150 feet above mean sea level.

The western part of the area is drained by Waterton River and Drywood Creek; the region between the Waterton River and the Belly River drains mostly internally into a large number of small lakes and sloughs, of which the most prominent are Cochrane Lake, Dipping Vat Lake, and Strawberry Lake. This part of the area is one of the oldest irrigation districts in Alberta, and numerous canals bring in and distribute water from the St. Mary Reservoir.

Acknowledgments

The authors wish to express their appreciation to Mr. J. McKenzie, who measured the water levels and recorded the gauge and flow readings on which a large part of this report is based; moreover, it was always a pleasure to find necessary data in his extremely well-organized filing system. Appreciation is also expressed to Mr. R. B. Wells, Project Engineer, PFRA, for his assistance and interest; to Mr. A. Keibel, who very ably carried out the drilling and completion of the wells for the pump test; and to Mr. R. Brown and Mr. D. Withers, who carried out the well survey and water sampling program.

GEOLOGY

The Upper Cretaceous bedrock strata subcropping in the area are disturbed by a number of faults and thrust faults striking approximately 315° azimuth and dipping westerly (Douglas, 1951). The more important of these are the Cochrane Lake Thrust, the Hillspring Fault and the Okey Ridge Fault (Fig. 6a).

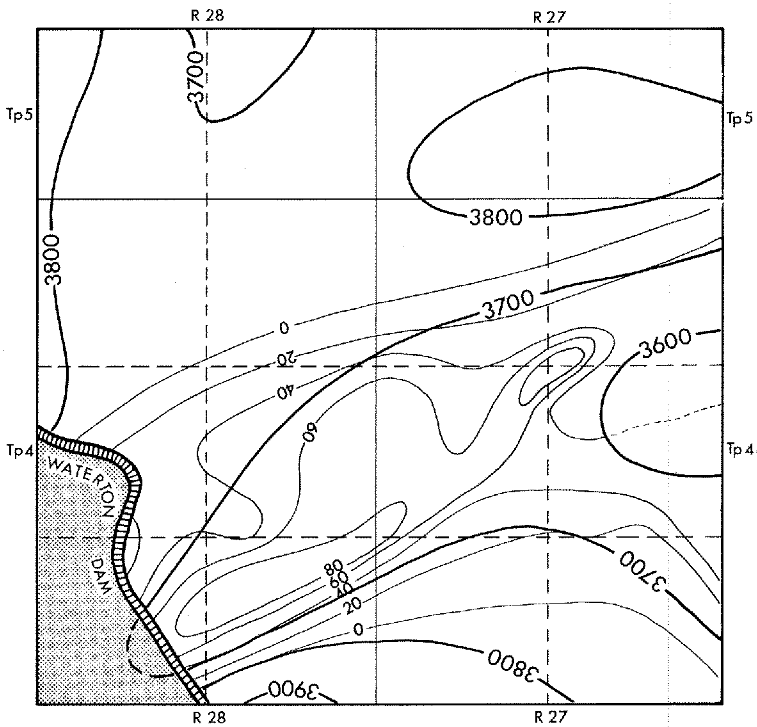
The bedrock is overlain by glacial deposits, mainly till, although a major preglacial drainage channel, the Cochrane Valley (Geiger, 1965), is incised in the bedrock surface, and gravel and sand have been deposited in it to a maximum proven thickness of 95 feet. The upper part of the sand and gravel deposit is commonly silty and clayey; the middle section consists of clean gravel with occasional thin sand lenses; and the lower 5 to 10 feet are of a coarser, locally cemented gravel. These coarse deposits have been covered by till, and from the present land surface the presence of the buried channel can hardly be suspected.

The sand and gravel deposits form what in this report is called the Cochrane Valley aquifer, which is the major topic of the study. The course of the Cochrane Valley is outlined on figure 5, which has been compiled from seismic and other borehole data. The isopach map of the Cochrane Valley aquifer (Fig. 7) and the cross sections (Fig. 8) show that the greatest thickness is reached in a narrow band less than 1,000 feet wide, where a thickness of over 90 feet has been recorded; over most of the remainder of the valley the thickness of the aquifer ranges between 35 and 60 feet.

CHEMISTRY OF GROUNDWATER

Introduction and General Remarks

In 1965, shortly before the filling of the reservoir took place, approximately 90 samples of groundwater were obtained from wells and springs in the area (Fig. 9). At 50 of the observation points samples were again collected in the summer of 1967, so that changes in composition resulting from the change in the hydrologic regime could be observed.



LEGEND

3800 — Bedrock elevation (100 ft contours)

80 — Isopach of sand and gravel (20 ft contours)

FIGURE 7. Isopachs of sand and gravel deposits in Cochrane Valley.

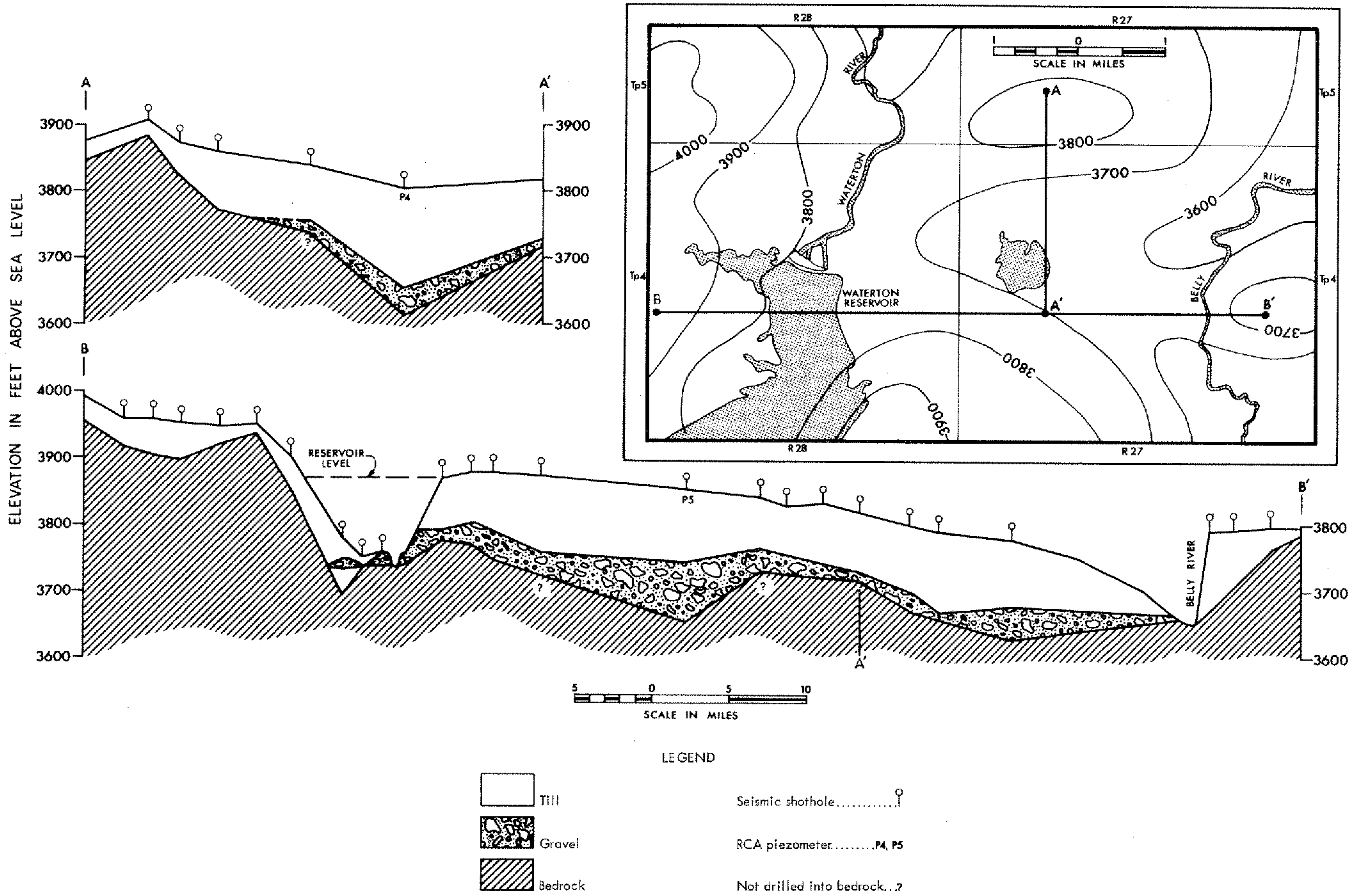


FIGURE 8. Sections through Cochrane Valley.

All samples were analyzed by the Provincial Analyst for total solids, ignition loss, calcium, magnesium, chloride, total alkalinity (or carbonate plus bicarbonate), nitrate, and iron. Sulfate and sodium, the only undetermined major components, were estimated from the ionic balance equation:

Sum of all anions = sum of all cations . . . . (1)

and the mass balance equation:

Total solids – ignition loss = sum of known and unknown components (2)

where anions and cations are expressed in epm (equivalents per million), and total solids, ignition loss, and the sum of all components are in ppm (parts per million). The ignition loss in (2) is the ignition loss as reported minus the weight of the carbon dioxide given off by the calcium and magnesium carbonates when these are converted into the oxides on ignition. The carbon dioxide loss is estimated as 0.44 times the lesser of alkalinity and hardness, but not exceeding the reported ignition loss.

By inserting the conversion constants to obtain epm values from ppm values in (1) and (2) and by solving for sulfate (in epm), the following equation is obtained:

Sulfate = 0.0140 (total solids + carbon dioxide loss – ignition loss) + 0.0021 calcium + 0.0124 magnesium – 0.0231 chloride – 0.0149 alkalinity – 0.0854 nitrate . . . . (3)

where total solids, ignition loss, and carbon dioxide loss are in ppm; calcium, magnesium, and chloride are in ppm calcium, ppm magnesium and ppm chloride, respectively; alkalinity is in ppm calcium carbonate; and nitrate is in ppm nitrogen. These constituents are customarily reported in this form by the Provincial Analyst. After sulfate has been calculated, sodium is found from the ionic balance equation.

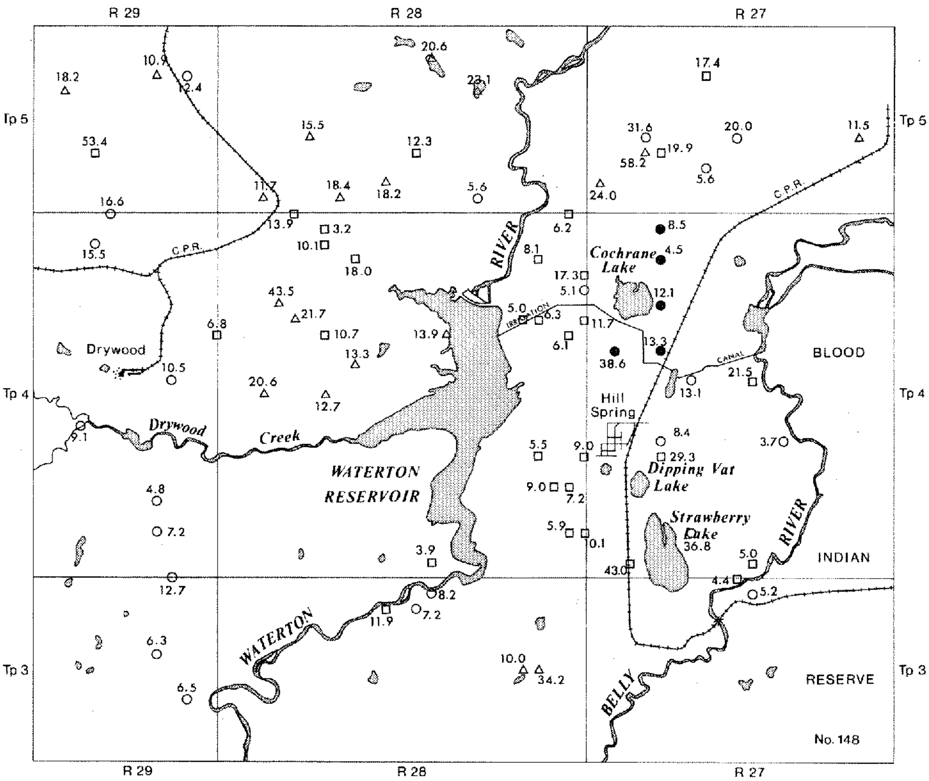
Regional Variation in Composition

Maps of total anions, per cent sodium, per cent bicarbonate plus carbonate, and per cent chloride are presented in figures 9 to 12, respectively. Although in general there seems to be little correlation between the values, the following rules seem to emerge:

1) the waters in the Cochrane Valley aquifer do not appear as a clearly distinctive group;

2) the areas surrounding Belly River, Waterton River, and Drywood Creek are relatively low in total anions and high in bicarbonate plus carbonate;

3) the percentage sodium is higher in the northern part of the area.



LEGEND

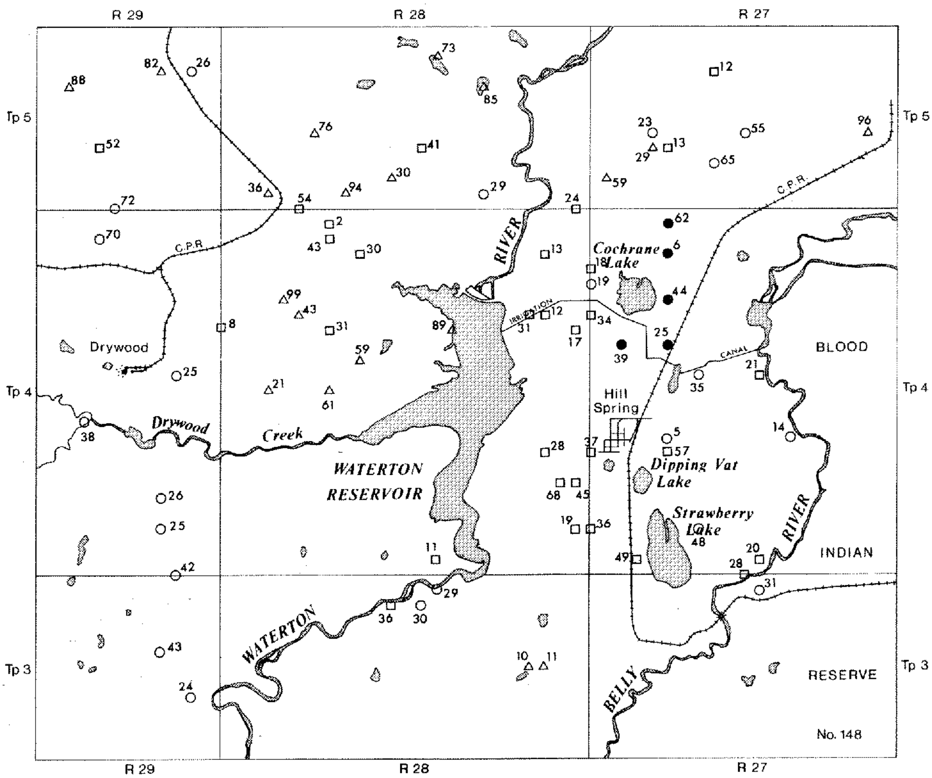
○ Sample from surficial deposits

● Sample from Cochrane Valley aquifer

△ Sample from bedrock aquifer

□ Sample from unknown depth and formation

FIGURE 9. Distribution of total anions.



LEGEND

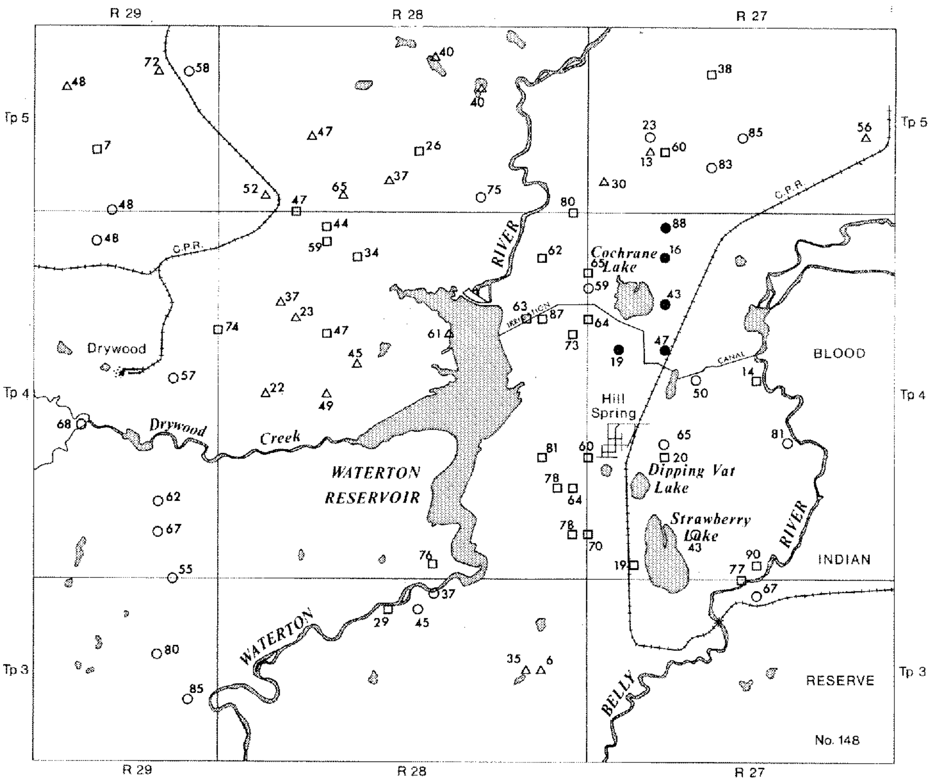
○ Sample from surficial deposits

● Sample from Cochrane Valley aquifer

△ Sample from bedrock aquifer

□ Sample from unknown depth and formation

FIGURE 10. Distribution of per cent sodium.



LEGEND

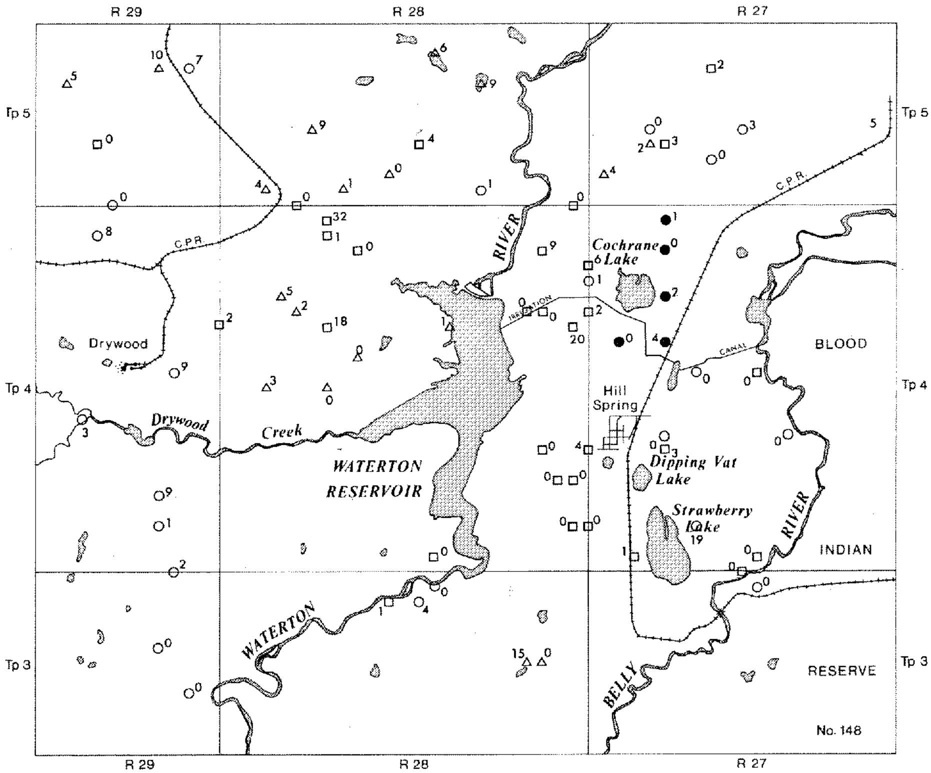
○ Sample from surficial deposits

● Sample from Cochrane Valley aquifer

△ Sample from bedrock aquifer

□ Sample from unknown depth and formation

FIGURE 11. Distribution of per cent bicarbonate plus carbonate.



LEGEND

○ Sample from surficial deposits

● Sample from Cochrane Valley aquifer

△ Sample from bedrock aquifer

□ Sample from unknown depth and formation

FIGURE 12. Distribution of per cent chloride.

In order to eliminate the subjectivity which is always inherent in the interpretation and hand contouring of poorly distributed data with a high random variation, the method of fitting a polynomial trend surface by the least-square criterion was employed to obtain contour maps on the chemical data. This method consists of approximating the function $z(x,y)$, which describes the value of the chemical parameter to be mapped – the dependent variable z – in terms of the map coordinates – the independent variables x and y – by the polynomial

$$z'(x,y) = \sum_{j=o}^{n} \sum_{i=o}^{n-j} a_{ij} x^i y^j \quad \ldots\ldots\ldots\ldots \quad (4)$$

where z' is the approximate value of z at the point (x,y).

The coefficients a_{ij} are such that the sum of squares of $z-z'$, i.e:

$$\sum_{k=l}^{m} (z_k - z_k')^2,$$

where m is the number of data points, is a minimum. They can be uniquely determined, provided that the number of data points is greater than or equal to $(n+1)(n+2)/2$ and that they do not all lie on a curve of order n; the determination of the coefficients requires the inversion of a symmetric matrix of order $(n+1)(n+2)/2$.

Surfaces up to order 7 were fitted to the data represented in figures 9 to 12. As could be expected, the correlation coefficients were quite low; for the maps of total anions, for example, they ranged from 0.248 (order 3) to 0.7ll (order 7). For orders 6 and higher, although double precision arithmetic was used, a loss of significance occurred in the last steps of the matrix inversion; this indicates that the coefficients obtained do not necessarily represent the best fitting polynomial surface for that order. For areas on the map where data are sparse, particularly along the boundaries, the behaviour of the polynomial is unpredictable, leading to extremely high or low values of the dependent variable. To remedy such shortcomings of this method several alternatives have been proposed. For example, extra data may be "eyeballed in" in the "open" areas and outside the boundaries of the area to improve the behaviour of the polynomial there, or a preliminary smoothing of the data may be performed (McIntyre, Pollard and Smith, 1968). A simple method in the latter category was devised during the present study, which worked well on the data on hand, and which, to the authors' knowledge, has not been described in the literature. This method consists in dividing the area by a regular grid into a number of rectangular subareas such that most subareas contain three or more data points. In each subarea a regular grid of auxillary control points is then created, at which values for the dependent variable are assigned as follows: if no data points are located in a subarea, no values are assigned in that subarea; if one data point is found, that value is assigned to all grid points; if two data points are found, the value assigned to each grid is the average of the values at the two data points; if from three to five data points are found, a best fitting plane surface is fitted to these points and grid points are calculated accordingly; if six or more data points are found, a best fitting 2nd order surface is constructed and values at the grid points are assigned accordingly. The thus-calculated grid values are then used as input to the routine which determines the regional surface.

Figure 13 shows maps of the sum of anions for orders 3 to 7 derived by both methods. The correlation coefficient R is consistently higher for the method with preliminary smoothing; the 7th order surface of the first method, which might be termed the actual trend, is already vaguely outlined on the 3rd order surface of the second method and is quite well defined there from the 5th order onward. On the whole, all five maps produced by the second method show a striking similarity, a feature that is entirely lacking in the set produced by the first method.

The matrix inversion routine also performed better in the second method; loss of significance did not occur until the 7th order instead of the 6th.

Figure 14 shows the 9th order surface representing the sum of anions, with the low trends marked by a dashed line. Of particular interest is the low trend through Waterton Reservoir; its northeastern continuation, and the branch to the south-southeast, closely follows the low trend of the bedrock surface which is the Cochrane Valley. The map corroborates the statement that total anions are low in the topographic lows around the rivers, but shows that the Cochrane Valley aquifer tends to be low in total anions as well. That the latter feature is not noticeable on figure 8 is possibly due to the fact that a large proportion of the sample points in this area, although listed as being of unknown origin, are actually located in the Cochrane Valley aquifer, and have, therefore, a low total solids content. The few sample points that are known to be in the aquifer naturally do not stand out against these.

The 8th order surface of per cent sodium is shown in figure 15, where the dashed lines again indicate the low trends. In contrast to the map of the sum of anions, there is no indication of a correlation between topography and chemistry. Instead, the picture is approximately one of high values in the northern quarter of the area, where values are generally over 40 per cent, and the low values in the remainder of the area – substantially the conclusion drawn from the raw data distribution map.

The contour map of the 8th order surface of per cent bicarbonate is shown in figure 16, where the high trends are indicated by the dashed lines. Although to a much lesser degree than on the map of total anions, the dashed line also follows the course of the Waterton River and the Cochrane Valley.

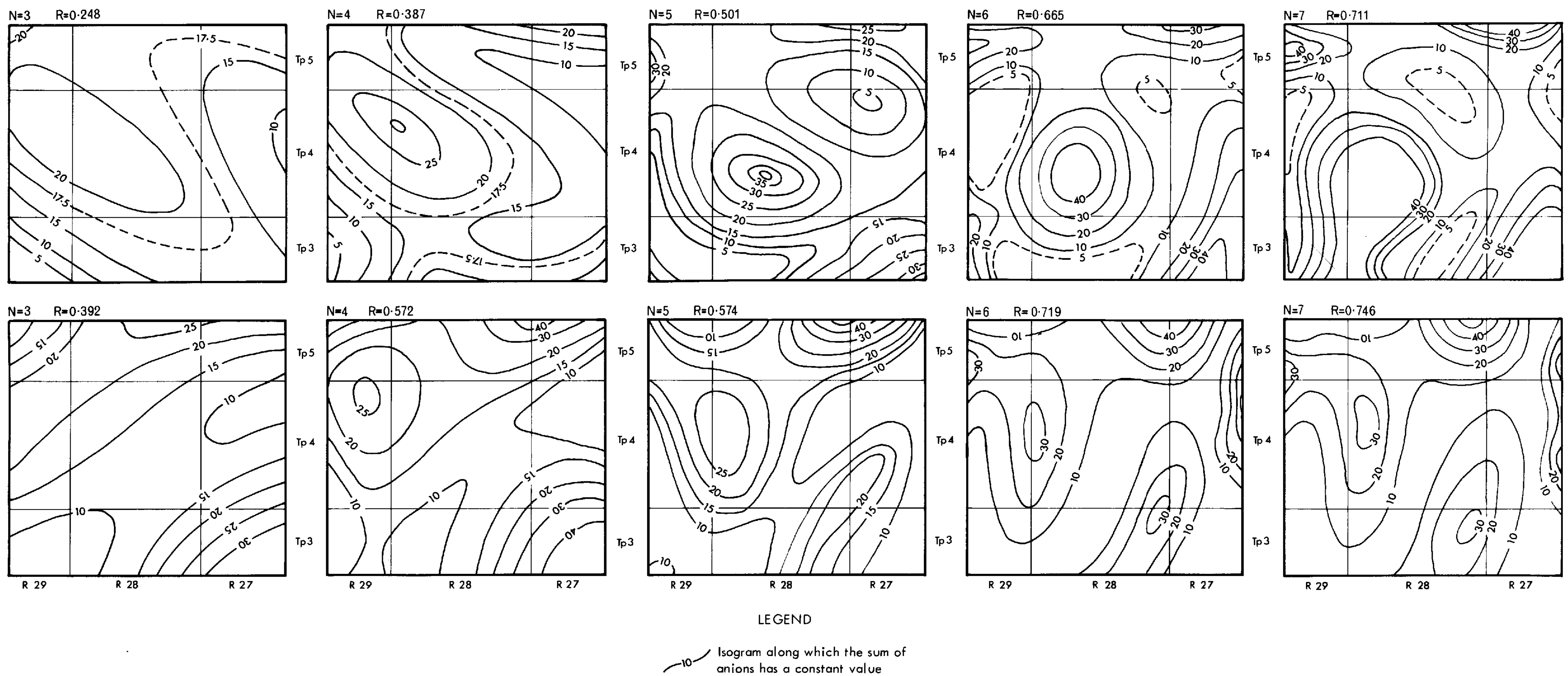


FIGURE 13. Polynomial surfaces of the sum of anions: a comparison of two methods.

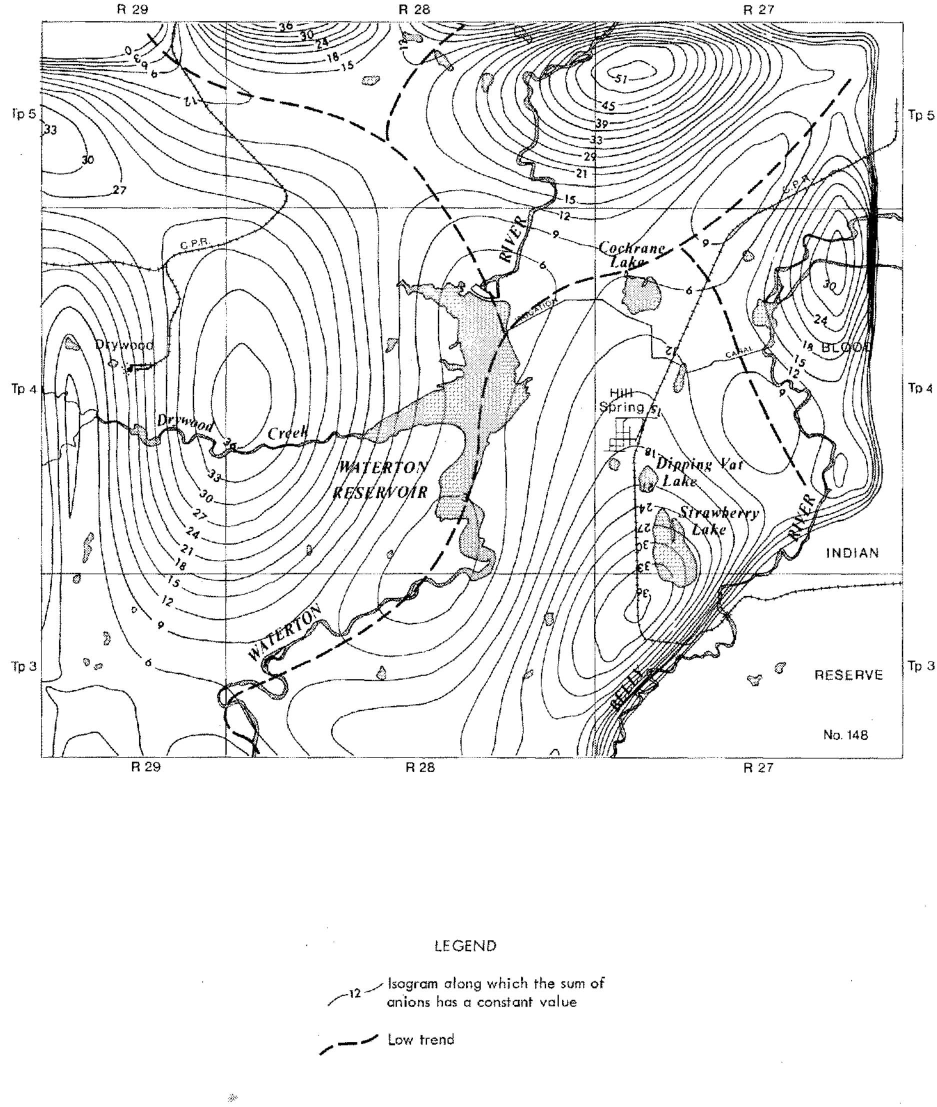
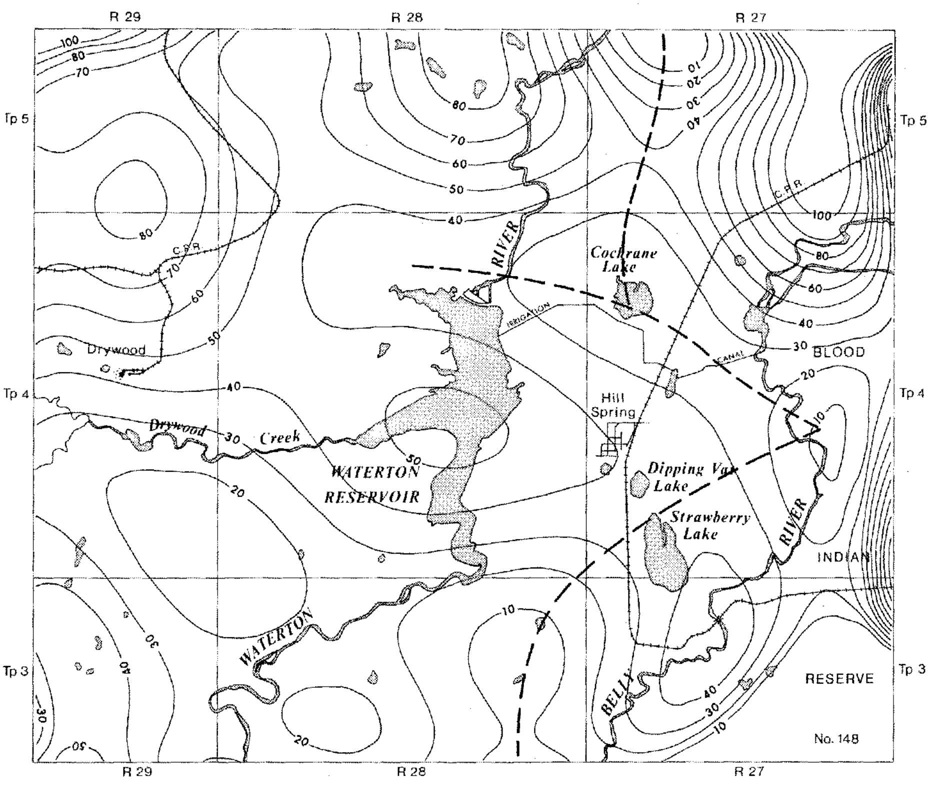


FIGURE 14. Ninth order surface of sum of anions.

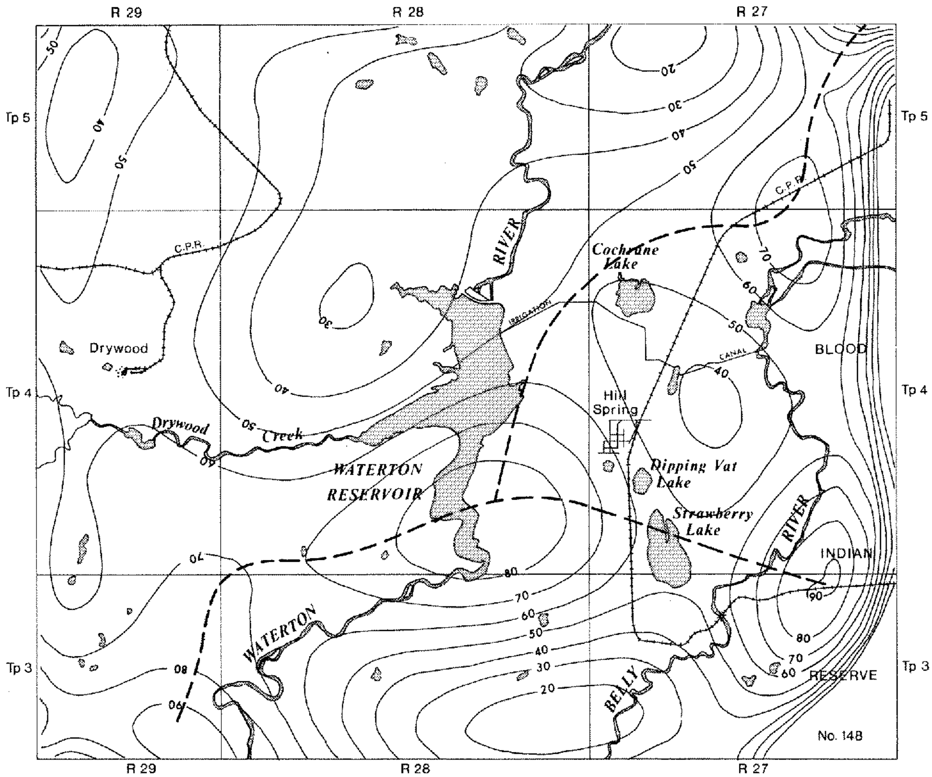


LEGEND

Isogram along which per cent sodium has a constant value

Low trend

FIGURE 15. Eighth order surface of per cent sodium.

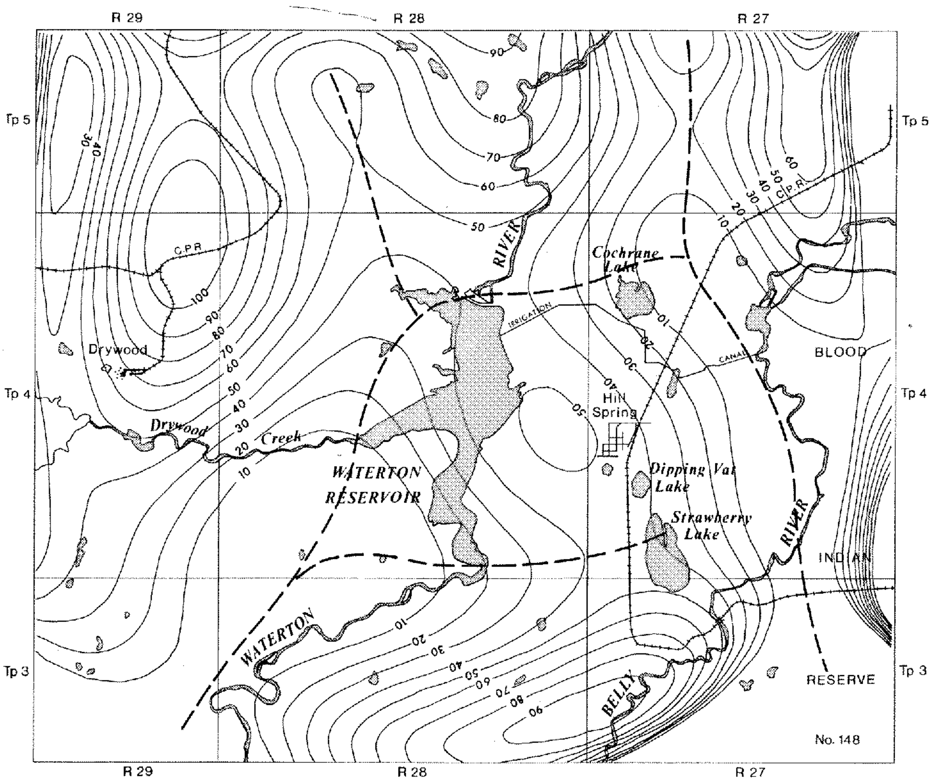


LEGEND

20 Isogram along which per cent carbonate plus bicarbonate has a constant value

Low trend

FIGURE 16. Eighth order surface of per cent carbonate plus bicarbonate.



LEGEND

10 Isogram along which per cent chloride has a constant value

Low trend

FIGURE 17. Eighth order surface of per cent chloride.

On figure 17, representing the 8th order surface of per cent chloride, the low trends, marked by the dashed lines, are very weakly outlined, but similar elements to those in the previous maps seem to be present: a low at the upstream section of Waterton River and a weak low along the Cochrane Valley.

The total anions map, the per cent bicarbonate map, and to a lesser extent the per cent chloride map, all indicate that relatively fresh groundwater is present in the low areas and in the Cochrane Valley; the groundwater there is low in total solids, high in bicarbonate, and low in chloride. The map of per cent sodium, on the contrary, may be a reflection of the bedrock geology of the area. The part of the area north of the Cochrane Lake Thrust, consisting of predominantly nonmarine sandstone and shale of the St. Mary River Formation, gives rise to waters high in sodium, while the area south of the same thrust consists of marine shales of the Bearpaw Formation and sandstones and shales of the Belly River Formation.

However, care should be taken not to interpret the maps on a regional basis only, for there is ample indication that the water in the bedrock is chemically quite different from the water in the overburden. This is especially true for per cent sodium, where the high values clearly coincide with those areas where most of the wells are drilled into the bedrock. It is well known that bedrock water is generally softer than water from the overburden, due to the base-exchange capacity of the clay minerals in the bedrock (Schoeller, 1962; Le Breton and Jones, 1962). The observed pattern of the per cent sodium map may, therefore, be solely a reflection of the relative abundance of bedrock wells in the northern part of the area.

In order to investigate the dependence of chemical composition on the type of source rock statistically, the available water analyses were divided into three categories: bedrock source, overburden source, and unknown source, i.e. those for which no data on well depth or source material was available. In the case of wells for which the depth of the hole, but no driller's log, was available, the analysis was placed in either the first or the second category on the basis of the topographic map and the constructed bedrock topography map; waters from natural sources (springs, seepages, etc.) were placed in the third category.

Table 1 gives the number of samples and the means and standard deviations for three chemical parameters: sum of anions, per cent sodium, and per cent bicarbonate plus carbonate for each category. It shows a significant difference between the bedrock waters on the one hand and the overburden waters and the waters from unknown source on the other. Z-scores and confidence levels for the statistical comparison between bedrock and overburden, between bedrock and unknown, and between overburden and unknown are given in table 2 for the three chemical parameters. The most significant difference is in per cent sodium, for which the hypothesis that bedrock and overburden samples come from populations with identical mean percentage sodium can be rejected at the 99.9 per cent confidence level.

Table 1. Means and Standard Deviations of Chemical Parameters for Groundwater from Bedrock, Overburden, and Unknown Sources

	Number of samples	Sum of anions		% Na		% (HCO_3 + CO_3)	
		mean	std. dev.	mean	std. dev.	mean	std. dev.
Bedrock	21	20	12	64	30	45	19
Overburden	32	13	9	40	18	59	20
Unknown	34	13	11	37	19	61	22
All sources	87	14.7	10.6	44.6	21.9	56.4	20.6

Highly significant differences are also shown for the per cent bicarbonate plus carbonate, while the differences in the means of the sum of anions between bedrock and overburden, and between bedrock and unknown sources may be termed probably significant. The difference between the waters from the overburden and from unknown sources is very small, and it can safely be assumed that in the latter category the majority of the samples are from the overburden.

The general picture indicates that relatively fresh groundwater, low in total solids, low in sodium, and high in bicarbonate plus carbonate, is present in the overburden material of the lower parts of the area surrounding the major rivers, while the highlands between the rivers and to the north and west contain older groundwater in the bedrock. Apparently, the overburden material, unconsolidated and therefore probably more permeable, contains water whose origin cannot solely be the bedrock high areas immediately to the west and northwest of the area or the

Table 2. Z-scores and Per Cent Confidence Levels for the Difference in the Mean of Chemical Parameters for Waters from Different Sources

	Sum of anions		% Na		% (HCO_3 + CO_3)	
	z	% conf.	z	% conf.	z	% conf.
Bedrock-overburden	2.28	97.7	3.30	99.9	2.57	99.0
Bedrock-unknown	2.17	97.0	3.69	99.98	2.85	99.6
Overburden-unknown	0.0	-	0.66	49.1	0.53	40.4

$$z = \frac{\bar{x}_1 - \bar{x}_2}{\sqrt{s_1^2/N_1 + s_2^2/N_2}}$$

$$\left.\begin{array}{l} \bar{x}_{1,2} = \text{mean} \\ s_{1,2} = \text{standard deviation} \\ N_{1,2} = \text{number of samples} \end{array}\right\} \text{for each group}$$

local bedrock high between the Belly and Waterton Rivers. It seems probable that within these materials there is a strong component of flow in a downstream direction in the valleys of Waterton River, Drywood Creek, and Belly River, the aquifers being recharged mainly from the rivers upstream at higher elevations. Similar flow regimes have been postulated by other investigators (Meyboom, Van Everdingen and Freeze, 1966; Vanden Berg and Lennox, 1969). Groundwater in the bedrock, however, is probably mainly directed transversely to the valley axes, and a flow regime there can be most adequately described following Tóth (1962). An application of Tóth's method to the Waterton Dam area is contained in the section on hydrology.

Groundwater Composition before and after Construction of Waterton Reservoir

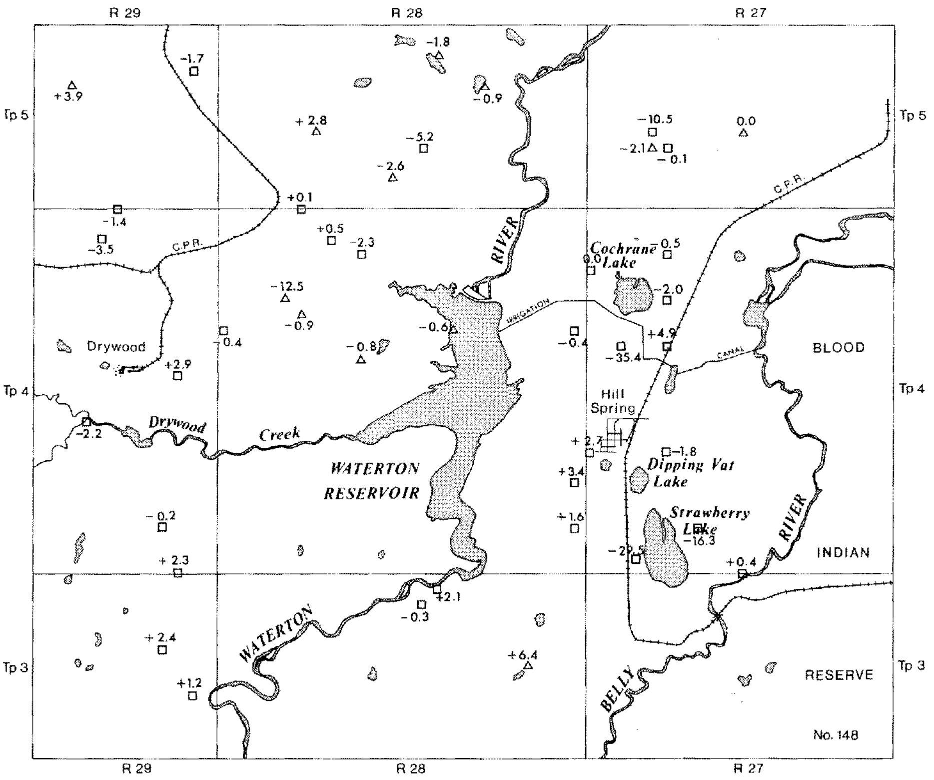
In 1967, two years after the reservoir filled, 50 of the sources sampled in 1965 were resampled and analyzed. Figure 18 shows the increase or decrease in total anions content of the water at these 50 sample points: at 44 of the sample points the total anions content had changed by 5 epm or less, and 23 of these showed a decrease in total anions. All six sample points at which the total anions content had changed by more than 5 epm showed a decrease, of which the two largest were – 35.4 and – 29.5 epm.

The regional distribution of the change in total anion content does not seem to be related to the topography of the area or the presence of the Waterton Reservoir.

Influence of Geologic Age of the Source Material

The influence of the geologic age of the source material on the change in the groundwater composition between 1965 and 1967 is again best borne out by a statistical analysis of the data. The 50 samples were divided into the three groups: bedrock waters, overburden waters, and waters from unknown source. The means and standard deviations for the sum of anions, the per cent sodium, and the per cent bicarbonate are listed in table 3 for each year, while the average change in composition for each group is shown in table 4 as the percentage of the 1965 average.

Although the average changes observed are small, they consistently show a change to fresher or younger water, that is, a water with less total anions, lower per cent sodium, and higher per cent bicarbonate plus carbonate. This freshening is most noticeable in the overburden waters and least noticeable for the bedrock waters, and is more clearly shown in the table of z-scores and per cent confidence levels (Table 5). The figure entered here under per cent confidence indicates the confidence with which the assumption can be made that the 1967 samples represent a fresher water than the 1965 analyses; the confidence levels are,



LEGEND

△ Well in bedrock

□ Well in till or unknown formation

FIGURE 18. Change in total anions content between 1965 and 1967.

Table 3. Means and Standard Deviations of Chemical Parameters for Samples Collected in 1965 and 1967

			Sum of anions	% Na	% (HCO_3 + CO_3)
Bedrock n = 11	1965	mean	23.6	64	40
		std. dev.	14.0	28	12
	1967	mean	22.6	62	41
		std. dev.	11.8	27	17
Overburden n = 23	1965	mean	14.9	43	57
		std. dev.	9.1	19	19
	1967	mean	12.1	34	62
		std. dev.	5.2	24	19
Unknown origin n = 16	1965	mean	13.2	36	60
		std. dev.	10.2	14	20
	1967	mean	12.1	32	64
		std. dev.	5.7	18	17

Table 4. Per Cent Change in Average Composition Between 1965 and 1967

	Sum of anions	% Na	% (HCO_3 + CO_3)
Bedrock	- 4.2	- 3.1	+2.5
Overburden	-18.8	-20.9	+8.7
Unknown origin	- 8.3	-11.1	+6.7

Table 5. Z-scores and Per Cent Confidence Levels for Significance of the Difference Between Averages of Composition in 1965 and 1967

	Sum of anions		% Na		% (HCO_3 + CO_3)	
	z	% conf.	z	% conf.	z	% conf.
Bedrock	0.18	57.1	0.17	56.8	0.16	56.4
Overburden	1.28	90.0	1.41	92.1	0.89	81.3
Unknown origin	0.38	64.8	0.70	75.8	0.61	72.9

therefore, based on a one-tailed t-test. Table 5 shows that this assumption is warranted for the surficial waters, but cannot be accepted with any confidence for the other two categories.

Influence of the Elevation of the Sample Point

In order to investigate the dependence of the observed changes in composition on the elevation of the sample points relative to the elevation of the water in the reservoir, those samples from wells of known depth were separated into three groups: samples from points at elevations below the pre-1965 water level of the Waterton River at the dam site, samples from points at elevations between the level of Waterton River prior to 1965 and the highest level of Waterton Reservoir after 1965, and samples from points at elevations above the highest level of Waterton Reservoir. It was naturally anticipated that freshening of the groundwater would be most pronounced in the second group and to a lesser extent in the first group, while no freshening at all was expected in the third group. As the first and second groups contain only 7 and 5 samples, respectively, a one-tailed t-test was employed to determine the levels of significance. Tables 6, 7, and 8 summarize the statistics for the three groups. Contrary to expectations, the highest

Table 6. Comparison of Groundwater Composition in 1965 and 1967 for Sample Points Below 3,750 Feet Above Mean Sea Level

Sample point	Elevation (feet)	Sum of anions		% Na		% (HCO_3 + CO_3)	
		1965	1967	1965	1967	1965	1967
7[1]	3730	38.86	3.47	39	17	19	88
9	3750	13.35	18.20	25	20	47	47
12[1]	3650	12.16	10.16	44	51	43	39
13[1]	3720	6.67	6.14	37	37	74	67
14[1]	3710	8.55	11.70	62	53	88	39
19	3740	20.07	20.04	55	38	85	66
76	3640	23.20	22.28	85	91	40	45
Mean		17.55	13.14	50	44	57	56
Standard deviation		10.27	6.66	18	23	24	17
t		0.8825		0.5032		-0.0833	
% Confidence		80-85		65-70		45-50	

[1] Sample point in the Cochrane Valley aquifer.

confidence that the 1967 waters are fresher than those of 1965 is obtained in the third group, while the first and second groups show little difference. It must be concluded, therefore, that the observed change to a fresher groundwater is to a large part, if not completely, caused by influences other than the filling of Waterton Reservoir.

Of the other possible causes, fluctuation in the yearly precipitation is one of the most likely as it affects the high recharge areas most directly, where infiltration is rapid. Continuous records of precipitation over a sufficient number of years are available for Pincher Creek, 16 miles northwest of Waterton Dam, and for Caldwell, 8 miles south of Waterton Dam, and bar graphs of the total annual precipitation over the years 1941 to 1969 for these localities are presented in figure 19. The Caldwell record shows a period of below-average precipitation from 1955 to 1963, followed by above-average precipitation to the present time; the Pincher Creek record has a similar period of below-average precipitation between 1955 and 1964, followed by close to average precipitation to the present time. On both records, the year 1965, in which the first set of samples was taken, has above-average precipitation, most of which fell in the first six months of that year (see the bar graph of total monthly precipitation for Caldwell, 1965, figure 19). The samples were obtained during the latter half of the year, but it seems reasonable to assume that in the short period between the high precipitation and the sampling the effect

Table 7. Comparison of Groundwater Composition in 1965 and 1967 for Sample Points at Elevations Between 3,750 and 3,850 Feet Above Mean Sea Level

Sample point	Elevation (feet)	Sum of anions		% Na		% (HCO_3 + CO_3)	
		1965	1967	1965	1967	1965	1967
48	3785	13.95	13.33	95	93	65	69
28	3835	17.37	17.26	21	21	65	77
17	3835	31.71	21.21	24	18	23	27
18	3835	13.41	11.59	27	9	61	60
77	3780	20.71	18.93	73	79	40	34
Mean		19.43	16.46	48	44	51	53
Standard deviation		6.68	3.54	30	35	17	20
t		0.7857		0.1735		0.1524	
% Confidence		75-80		55-60		55-60	

Table 8. Comparison of Groundwater Composition in 1965 and 1967 for Sample Points Above 3,850 Feet Above Mean Sea Level

Sample point	Elevation (feet)	Sum of anions		% Na		% (HCO_3 + CO_3)	
		1965	1967	1965	1967	1965	1967
30	3895	10.11	16.53	11	26	35	60
32	3935	7.27	6.96	32	8	46	97
33	3889	8.27	10.32	26	26	38	61
80	4278	12.76	15.04	45	30	58	62
2	3943	36.97	20.64	49	45	43	46
46	4000	19.28	18.40	94	93	59	60
47	3940	13.34	12.55	60	56	45	40
58	4030	43.84	31.39	100	98	38	49
81	4220	7.25	7.04	26	29	69	34
83	4140	8.15	5.97	39	23	69	80
84	4180	10.64	13.54	27	8	60	73
85	4240	6.84	6.40	11	7	83	82
86	4110	15.54	12.00	72	76	50	44
88	4180	16.60	15.19	75	89	50	70
70	3930	12.01	6.80	33	26	35	46
71	4070	18.30	15.73	31	22	37	31
75	3900	15.58	18.38	77	45	47	55
91	4040	18.22	22.14	88	80	48	31
92	3990	12.74	10.89	57	82	43	72
Mean		15.46	14.00	50	46	50	58
Standard deviation		9.45	6.34	27	30	13	18
t		0.5443		0.4204		1.5286	
% Confidence		70-75		65-70		90-95	

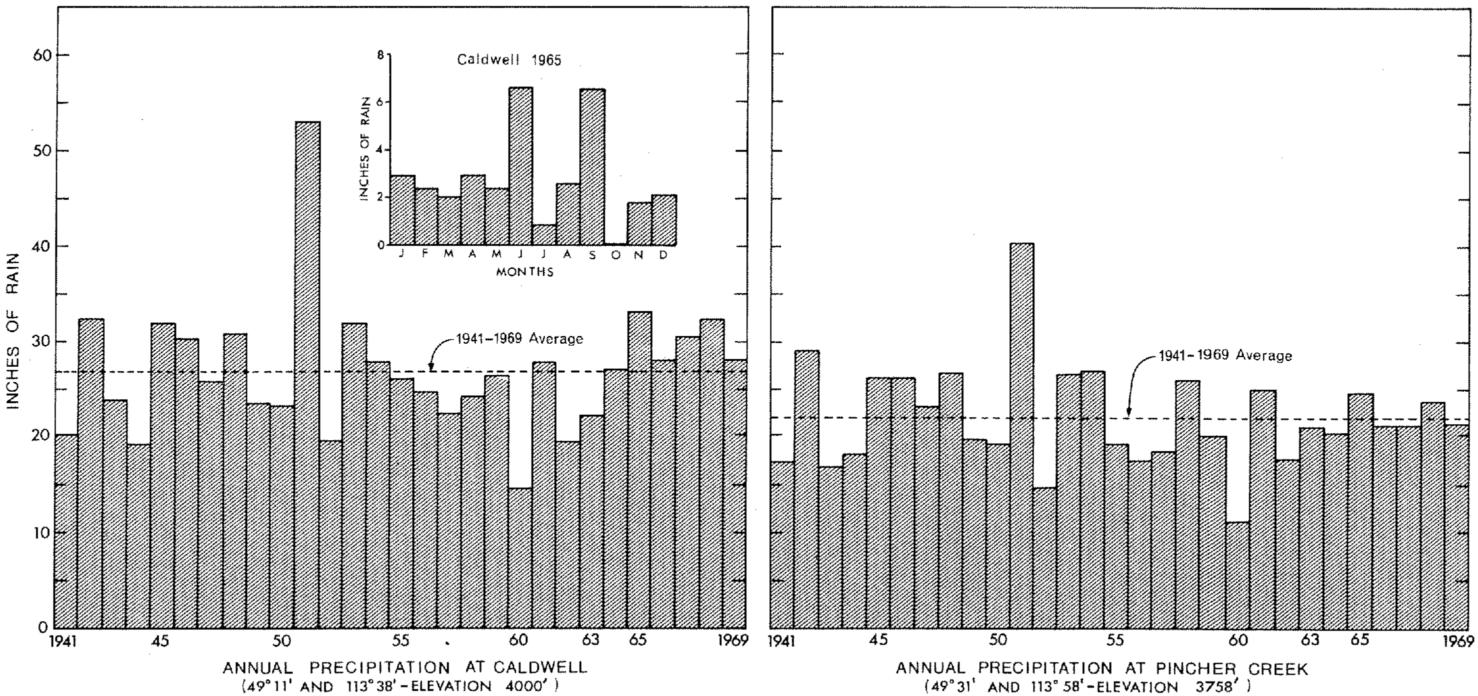


FIGURE 19. ***Annual precipitation at Caldwell and Pincher Creek.***

of the precipitation could not yet have influenced the composition of the groundwater in the majority of the wells. There is, therefore, sufficient ground to postulate that the difference in chemical composition between the 1965 and 1967 samples is largely a consequence of the increasing precipitation during these years.

Four of the samples listed in table 6 represent points in the Cochrane Valley aquifer; of these, only sample 7 indicates replacement of water from the regional flow system by fresh water from the reservoir. The well from which the sample was taken is located near the southern edge of the aquifer, and although the aquifer may be assumed to be very thin at this point (Fig. 7), the well responds rapidly to fluctuations of the reservoir level (Fig. 20, well of D. Gibb).

Sample points 12 and 13 indicate only a slight decrease in total anions, but it must be remembered that fresh water from Waterton River was already recharging the central, deeper part of the aquifer prior to the construction of the reservoir, and little change in composition could be expected.

Sample point 14 is anomalous, showing an increase in total anions and a sharp drop in per cent bicarbonate plus carbonate. This point was resampled in 1969 with the following results: sum of anions – 8.49 epm, per cent sodium – 48 epm, per cent bicarbonate plus carbonate – 58 epm. The 1967 analysis, therefore, may have been in error.

Chemical Composition of the Surface Water

In the spring of 1969 samples of water were obtained from Waterton River where it flows into Waterton Reservoir, from Waterton Reservoir at the outlet through Waterton Dam, and from Drywood Creek, approximately 4 miles upstream from the point where it flows into Waterton Reservoir.

Although the range in total solids content between the three waters is certainly significant (Table 9), the percentage values for the constituents are noticeably similar, except for chloride; Drywood Creek is richer in several major constituents (calcium, magnesium, bicarbonate, and sulfate) and the mixed water at the dam is, with regard to these constituents, intermediate between Drywood Creek and Waterton River. If, however, the ratio is calculated for each constituent, of the requisite volumes of water discharged from both streams into the reservoir to give the observed composition of the mixed water, a number of greatly different answers are obtained (Table 10). This indicates that either a substantial amount of water of entirely different composition is also admixed – possibly from groundwater discharge into the reservoir – or that chemical reactions occur between the standing water of the reservoir and its surroundings.

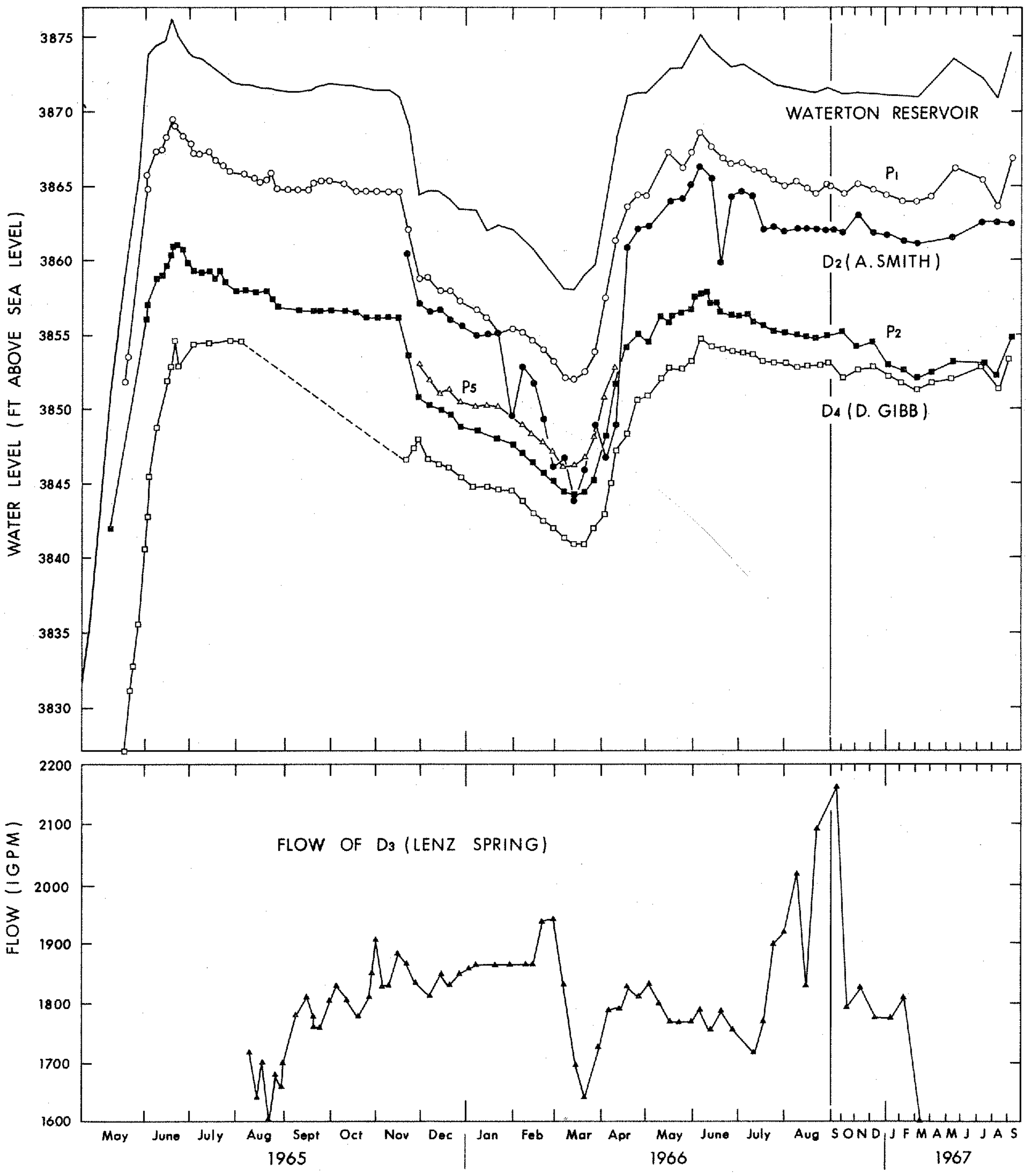


FIGURE 20. Water levels of wells in Cochrane Valley aquifer.

Table 9. Chemical Analyses of Surface Water[1]

	Drywood Creek Lsd. 5, Sec. 14, Tp. 4, R. 29, W.4th Mer.			Waterton River above Reservoir Lsd. 15, Sec. 34, Tp. 3, R. 28, W.4th Mer.			Waterton Reservoir at Outlet Lsd. 10, Sec. 26, Tp. 3, R. 28, W.4th Mer.		
Total solids (ppm)	186			140			166		
Total hardness (ppm $CaCO_3$)	170			116			133		
Total alkalinity (ppm $CaCO_3$)	152			110			126		
Residue on evaporation (ppm)	112			74			92		
pH	8.20			8.05			7.80		
	ppm	epm	% of total	ppm	epm	% of anions	ppm	epm	% of anions
Calcium (Ca^{++})	42.0	2.10	58.5	28.0	1.40	56.6	32.5	1.62	55.5
Magnesium (Mg^{++})	15.8	1.30	36.3	11.2	0.92	37.3	12.5	1.03	35.2
Sodium (Na^{+})	03.7	0.16	04.5	02.7	0.12	04.7	05.4	0.23	08.0
Potassium (K^{+})	01.0	0.03	00.7	01.4	0.04	01.4	01.4	0.04	01.2
Carbonate (CO_3^{--})	00.0	0.00	00.0	00.0	0.00	00.0	00.0	0.00	00.0
Bicarbonate (HCO_3^{-})	152.0	3.04	85.4	110.0	2.20	89.9	126.0	2.52	84.0
Sulfate (SO_4^{--})	21.0	0.44	12.3	11.0	0.23	09.3	19.0	0.40	13.3
Chloride (Cl^{-})	02.0	0.06	01.6	00.0	0.00	00.0	02.0	0.06	01.9
Nitrate (NO_3^{-})	01.6	0.03	00.7	01.2	0.02	00.8	01.4	0.02	00.8
Ca/Mg	01.61			01.57			01.52		
Cl/SO_4	00.136			00.150			00.00		

[1] Analysed in chemistry laboratory, Research Council of Alberta.

Table 10. The Ratio: Discharge of Drywood Creek/Discharge of Waterton River, Calculated from the Relative Abundance of Major Components of the Water in the Mixture

Component	Ratio = $\frac{p_1 - p_2}{p_2 - p_3}$
Total solids	0.77
Total hardness	2.18
Calcium	2.11
Magnesium	2.54
Total alkalinity	1.62
Sulfate	0.25
Nitrate	1.00

p_1 = concentration of constituent in analysis of Drywood Creek water
p_2 = concentration of constituent in analysis of mixed water
p_3 = concentration of constituent in analysis of Waterton River water above dam

Conclusions and Remarks

Analysis of the geochemical maps shows that the groundwaters in the lower parts of the area and in the Cochrane Valley aquifer generally exhibit the characteristics of younger, fresher water than is found in the surrounding high areas. This indicates that in the unconsolidated valley fill a strong component of groundwater flow exists parallel to the valley axis, presumably originating through river recharge at points upstream. This system is supposedly superimposed on the topographically controlled regional flow system, in which the flow is largely directed transversely to the valley bottom.

Groundwater samples obtained after the filling of Waterton Reservoir show a noticeable decrease in total anion content and in per cent sodium, and an increase in per cent bicarbonate plus carbonate; all three parameters indicate a freshening of the groundwater. The freshening is most pronounced in the overburden material, and considerably less for the bedrock waters.

Contrary to expectation, the freshening of the water is also more pronounced in those sample points which lie above the level of the water in the reservoir than for points below it. Precipitation records for the area indicate that the observed changes in the chemistry of the groundwater in the Waterton Dam area between the years 1965 and 1967 are caused by fluctuations in the precipitation rather than by the construction of the reservoir.

HYDROLOGY

Response of the Cochrane Valley Aquifer to Changes in the Level of Waterton Reservoir

At the time the Waterton Dam was completed and the reservoir was ready to be filled, the five piezometer nests constructed by the Research Council of Alberta were available for measurement of the hydrostatic head in the aquifer and the level of the water table in the overlying till. The water table wells, equipped with automatic recorders, showed only small fluctuations of the water table which bear no relation to the level of the Waterton Reservoir, but are the normal fluctuations associated with the seasonal variations in precipitation, evaporation, and transpiration.

The water level in the Cochrane Valley aquifer, prior to the filling of the reservoir, was at an elevation of 3,768 feet above mean sea level, or about 9 feet above the level of the Waterton River at the dam. It was a free water table, and a large proportion of the total thickness of the aquifer was unsaturated.

The reservoir began to fill in March 1965 and had attained a level of 3,784 feet above mean sea level by the end of the month; the hydrostatic head in the aquifer began to rise around the middle of the month when the reservoir level rose above the original water level in the aquifer. The reservoir level and the hydrostatic level in the aquifer continued to rise through April and May 1965 (Fig. 20). Piezometers P3, P4, and P5 started to flow on May 21, May 10, and May 28, respectively; domestic wells D1 (D. Dickie), D8 (W. C. Smith), D5 (R. Wynder), D6 (S. Wynder), and D7 (B. Wynder) were all flowing by May 24. Although regular measurements of the flow of these wells have been made, no useful correlation with reservoir levels could be established, and the measurements have not been included in this report. At the end of the month the total measured outflow was 143 igpm, the bulk of which was produced by D5, flowing at a rate of 99 igpm. The total outflow reached a peak on June 18, 1965 when a total outflow of 230 igpm was measured.

Beginning on August 10, 1965, flow measurements were also taken on D3, a large spring on the property of Mr. J. Lenz, in the Belly River valley (NE l/4, Sec. 28, Tp. 4, R. 27, W.4th Mer.). This spring was, at the time, the largest producer in the area, flowing at rates which lie between 1,640 and 2,170 igpm during the period from August 1965 to September 1967. Although no records of the flow of this spring were kept prior to and during the filling of the reservoir, according to local residents the flow of the spring before the construction of the reservoir was substantially less than it is at present. Furthermore, the spring is located near the axis of the bedrock low (Fig. 6) designated as the Cochrane Valley, and it is reasonable to assume that the water is flowing from the eastward continuation of the Cochrane Valley aquifer. In 1969 a chemical analysis was obtained on a sample from this spring, with the following result: sum of anions – 27 epm, per cent sodium — 56 epm, and per cent bicarbonate plus carbonate — 34 epm. Evidently the four years since the increased flow in the aquifer began have not been sufficient to replace the original groundwater by reservoir water. No test drilling has been carried out in this area; the thickness of the aquifer and the material of which it consists are unknown. From the rate at which the well is flowing it may be assumed, however, that the transmissibility is of the same order of magnitude as in the better known parts of the aquifer.

Fluctuations in the rate of flow of D3 (Fig. 20) do not seem to correspond to the fluctuations in the reservoir level, unless it is assumed that there is a relation between the low in the reservoir level between November 1955 and April 1966, and the low flow period of D3 between March and July, 1966. This assumption would involve a time lag of approximately four months, which is unlikely in view of the rapid response of water levels observed elsewhere in the aquifer. Moreover, the spring seems to have responded almost immediately to the completion of the first pressure relief well in February 1967, with a reduction of the flow of D3 from 1,800 to 1,600 igpm.

Water levels in the wells which do not flow follow the reservoir level closely and instantaneously (Fig. 20). During the period of low level in the reservoir (November 1965 to April 1966), water levels in the wells dropped approximately the same amount as the reservoir; piezometer P5 stopped flowing so that measurement of water levels could be resumed. The well started flowing again in April 1966, when the reservoir level had risen again to approximately 3,870 feet above mean sea level.

The rapid response of the whole aquifer to water level changes of Waterton Reservoir shows that there is an efficient hydraulic connection between the reservoir and the aquifer, and that the transmissibility of the aquifer must be high and its storage coefficient low.

Piezometric Surface and Hydraulic Gradient in the Cochrane Valley Aquifer

In order to eliminate the fluctuations of the reservoir level from the water levels measured in the aquifer, water levels of P1, P2, P5, D2, and D4 were replotted (Fig. 21) in terms of feet below the level of Waterton Reservoir. For all wells except P1, which is located very close to the reservoir, the relative level shows large variations with a range of approximately 8 feet. As a rule, the lower the reservoir level, the less is the difference between the reservoir level and the piezometric surface; this is to be expected since the lowering of the reservoir levels causes a smaller pressure differential in the flowing wells and a reduced flow, resulting in a reduced drawdown in the aquifer.

The average relative levels for the period from August 1966 to February 1967, when the reservoir was reasonably constant, are mapped in figure 22. The

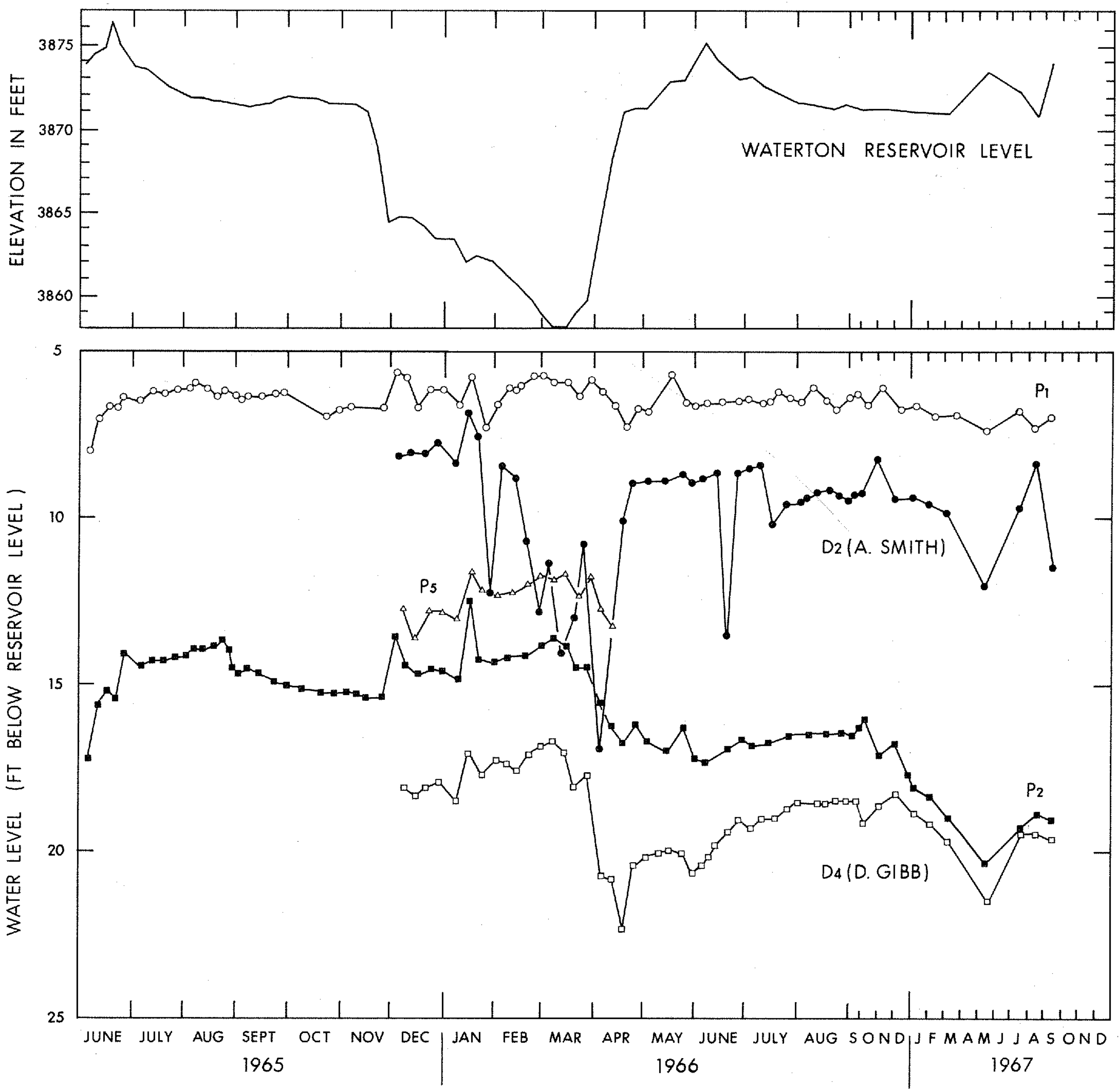


FIGURE 21. Water levels in feet below reservoir level.

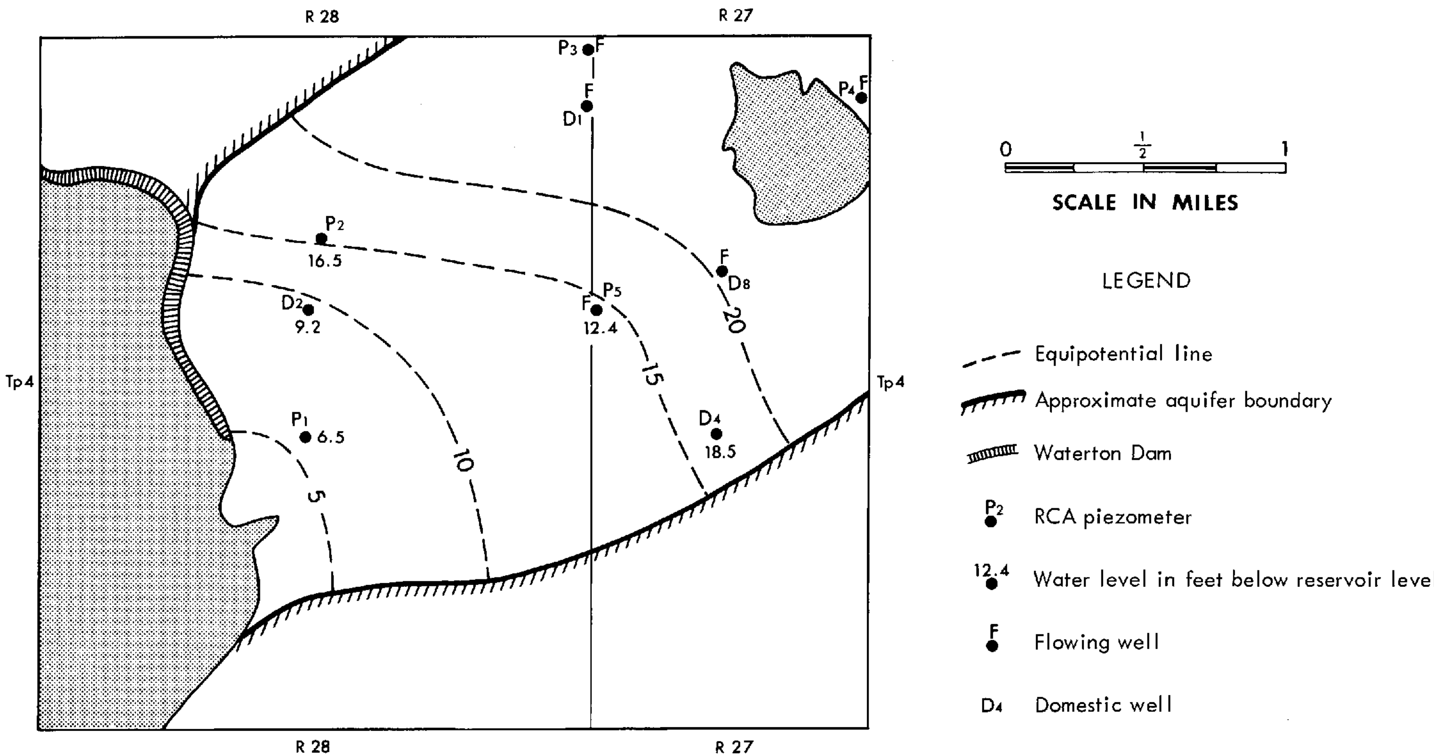


FIGURE 22. Water levels in Cochrane Valley aquifer for August 1966.

average piezometric level of P5 is also shown on this map, but it must be kept in mind that this value may be several feet low, as the observations were made during the period of low reservoir levels when the piezometer was not flowing. Although the observation points are too few to permit the construction of the complete piezometric surface, a few observations and conclusions can nevertheless be made from the available data.

1) On the basis of the three best defined points, P1, P2, and D4 (D. Gibbs), the average piezometric gradient in the aquifer is calculated as 15.8 ft/mile (0.003 ft/ft) and its direction as 210°.

2) The large difference between the water levels at P2 and D2 (A. Smyth) of 21.2 ft/mile (0.004 ft/ft) suggests that the flow is directed approximately northward in this part of the aquifer, indicating that the contact between the reservoir and the aquifer west of these locations is relatively impermeable, and that water is flowing into the aquifer in the neighbourhood of P1, and from there spreading out into the whole width of the aquifer. The lowered permeability of the aquifer material in the neighbourhood of the points P2 and D2 could possibly have been caused during construction of the dam, through sealing and compaction of the aquifer.

3) Based on the estimate of the average gradient, the cross-sectional area of the aquifer as derived from figure 7, and the measured outflow from wells, piezometers and springs, the hydraulic conductivity of the aquifer can be estimated from Darcy's equation:

$$Q = KA(dh/ds)\cos(\beta) \qquad (5)$$

where Q = measured outflow in gallons per day,

K = hydraulic conductivity in gallons per day per foot per foot,

A = total cross-sectional area in square feet,

$(dh/ds)\cos(\beta)$ = component of the gradient in the direction perpendicular to the cross section in foot per foot,

β = angle between the gradient and the direction perpendicular to the cross section.

By adding the average outflow from spring D3 of 1,900 igpm to the outflow from all the other flowing wells (230 igpm), $Q = 2{,}130 \times 1{,}440 = 3.067 \times 10^6$ igpd. From figure 7 the cross-sectional area of the aquifer along A–A′ is estimated at 4.657×10^5 ft^2; $\beta = 35°$, and the component of the gradient is $0.8 \times 0.00295 = 0.00239$, from which it follows that $K = 2{,}755$ igpd/ft^2. The average thickness of the aquifer is 52 feet which gives an average transmissibility of 143,000 igpd/ft.

In August 1969, after the big relief wells had been drilled, the total outflow from all known sources in the area was 6,060 igpm; the Lenz spring, D3, was still the largest single contributor, although its flow was then reduced to 1,310 igpm. The pressures measured at the flowing relief wells, after a 10-minute shut-in period, are shown on figure 23, expressed in feet below reservoir level; also shown are the water levels measured in piezometers P3 and P5, and wells W1 and W2. P3, which had been flowing at a rate of 8 igpm, had been shut in for a period of 12 hours prior to the measurement, and the water level was measured in a 15-foot long plastic hose, mounted vertically on top of the casing. P5, W1, and W2 did not flow, and water levels, therefore, could be measured inside the casing. For all four wells water levels could be measured to the nearest 0.01 foot. The magnitude and direction of the gradient at the sites of W1, W2, and P3 could therefore be measured accurately and were found to be 0.0103 ft/ft and 201°, respectively.

Water levels at P1, P3, and P5 were used to determine the regional average gradient; it was assumed that the hydraulic head at the site of P1, which at that time had been abandoned, could not have changed appreciably from what it was in 1967 because of its closeness to the constant head boundary. The regional average gradient was thus found to be 0.00485 ft/ft in a direction 253°.

Rough estimates of the average gradient can also be obtained from the piezometric head measured in the relief wells; these estimates, however, are on the low side, as each well has to be shut off in turn to obtain the measurement; values obtained for the gradient in this way are given in table 11. Based on the established gradient, the estimated cross-sectional area of the aquifer, the known total outflow, and assuming an average porosity of 20 per cent for the aquifer material, the average velocity of the groundwater was calculated to be in the order of 5 to 10 feet per day. In other words, after the aquifer was saturated, it would take from 1 1/2 to 3 years for the fresh water to penetrate a mile into the aquifer. Although the velocity must certainly be above average along direct flow paths between the intake area and the discharge points, this explains the lack of noticeable freshening of the groundwater observed in 1967, and in the case of the spring D3 as late as 1969.

Transmissibility and Permeability of the Cochrane Valley Aquifer from Pump Test Data

To determine the transmissibility, permeability, storage coefficient, and leakage factor of the aquifer, a 3-day constant-rate pump test was carried out in August 1969. Water levels were measured in the pumping well W1, and in three observation wells; the already existing piezometers P3 and P5, and W2 which was constructed specifically for the purpose of the pump test. The pumping well was located 186 feet south of the northeast corner of section 25, township 4, range 28, west of the fourth meridian; the locations of the observation wells with respect to the pumping well were: P3, 500 feet south; P5, 5,150 feet south; W2, 150 feet west and 15 feet north.

Table 11. Estimates of the Average Hydraulic Gradient Based on the Shut-in Pressures of Relief Wells

Well No.	Distance from reservoir (feet)	Shut-in pressure (feet of water below reservoir level)	Component of the gradient (foot/foot)
C1248	16,600	65	0.00391
RD1252	13,200	52	0.00394
RD1253	17,900	70	0.00391
RD1254	10,600	46	0.00434
RD1255	11,300	47	0.00416
RD1256	12,300	46	0.00374
		Average	0.00400

The well was pumped for 70 hours at a constant rate of 200 igpm; recovery was measured in the pumped well for 5 hours after the end of the drawdown test.

Unfortunately, two days before the start of the test, the reservoir level, which during the drilling and completion of wells W1 and W2 had been steady at 3,886 feet above mean sea level, began to drop at an average rate of 0.8 ft/day, because water was being diverted to the Medicine Hat area for irrigation. In the two days before the test an attempt was made to establish a usable relationship between the nonpumping water levels in the wells and the reservoir level, and as a result a correction factor of 0.32 times the drop in the reservoir level was applied to all water level measurements obtained during the test. In figure 24 the drawdown of the pumped well has been plotted against the logarithm of time; the residual drawdown, which is the difference between the corrected nonpumping water level and the actual water level after the well has been allowed to recover for a period of time t′, has been plotted against the logarithm of t/t′, where t is the time since pumping started.

Both the drawdown and the residual drawdown should plot as a straight line if the aquifer were homogeneous and of infinite areal extent; the transmissibility is then determined from the slope of the straight line by means of the modified Theis nonequilibrium formula (Cooper and Jacob, 1946):

$$T = 264\, Q/\Delta s \qquad (6)$$

where Q = the pumping rate in igpm, Δs is the drawdown per log cycle, and T is the transmissibility in imperial gallons per day per foot.

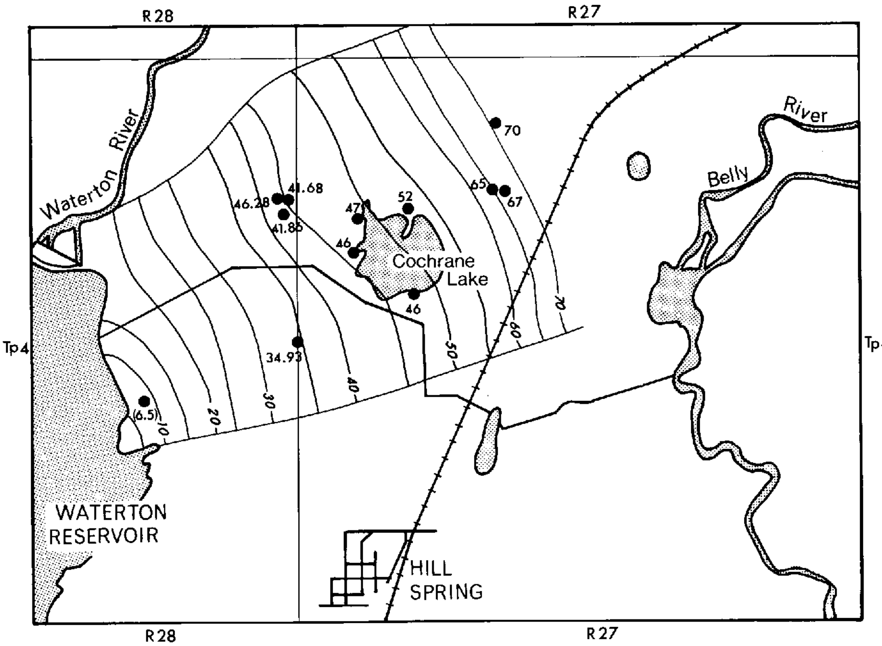


FIGURE 23. Piezometric head in Cochrane Valley aquifer, 1969, in feet below reservoir level.

The transmissibility as derived from the drawdown curve and the residual drawdown curve are 7,040 igpd/ft and 110,000 igpd/ft, respectively. The extreme discrepancy between these two results is noteworthy; as the measurements in the observation wells confirm the estimate of the transmissibility derived from the residual drawdown measurements, the estimate from the drawdown measurements is apparently too low. The unusually rapid drawdown in the pumped well during the pumping period may well be caused by a gradual increase in the well loss; that is, the sum of all the head losses associated with the rapid movement of the water through the aquifer in the immediate vicinity of the well, through the screen openings and in the well bore. In a properly developed and stabilized well, the well loss is a constant factor at a given pumping rate and therefore does not affect the slope of the time-drawdown curve and the determination of the transmissibility; if

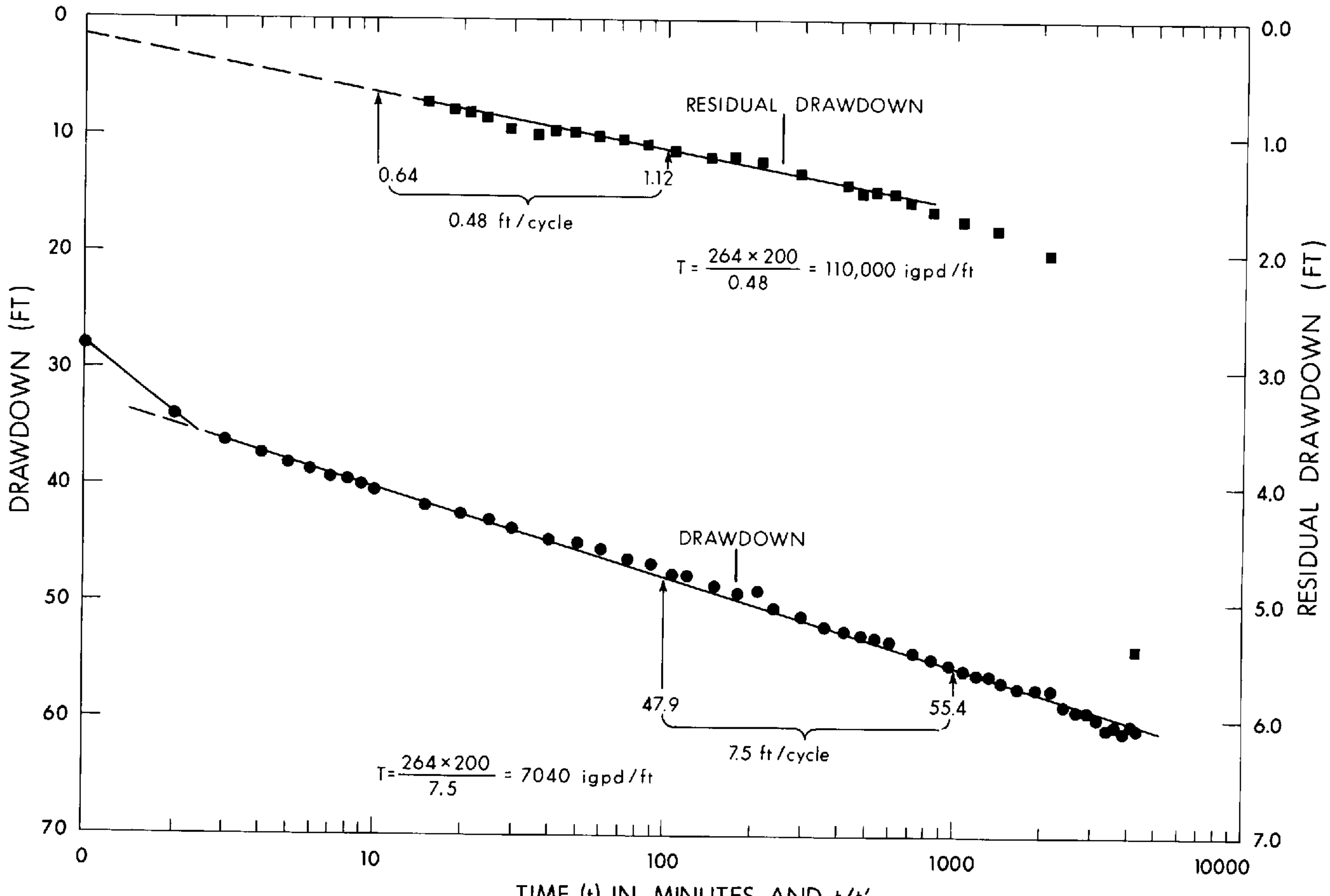


FIGURE 24. Drawdown and recovery curves of pumped well.

the well is incompletely developed or if development took place at pumping rates substantially lower than the rate of subsequent production runs, movement of the aquifer material may be induced during pumping of the well, which can cause either additional development or deterioration by plugging of the well screen or the part of the aquifer immediately adjacent to the screen.

Well W1 had been developed by alternately jet pumping and backwashing the well over a period of two days. During this process large quantities of very fine sand were removed from the formation; development was discontinued only when no more sand appeared and the formation seemed stable. The pumping rate during jetting was less than 100 igpm; therefore, it is likely that, at the pumping rate of the pump test, renewed movement of sand took place in the formation resulting in a gradually less effective well.

For the three observation wells the logarithm of the drawdown was plotted against the logarithm of time (Fig. 25); the transmissibility and storage coefficient of the aquifer were then determined using the nonequilibrium equation of Theis (1935) by means of the graphical solution given by Walton (1962):

$$T = 114.6\ QW(u)/s \qquad (7)$$

where

$$u = 2{,}244\ r^2S/Tt \qquad (8)$$

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

s = drawdown in feet

r = distance to the pumping well in feet

S = storage coefficient

t = time since pumping started in minutes.

The solid curves on figure 25 represent the fitted type curve of W(u) versus u. They represent the theoretical drawdowns that would occur if the aquifer were homogeneous, of infinite areal extent, and if the aquifer constants T and S had values as shown opposite each curve; the aquifer constants are determined by first obtaining the position of the type curve of W(u) that best coincides with the drawdown curve and then applying (7) to matching values of s, t, W(u), and u.

The solution of T and S from the drawdown curve of W2 seems to be straightforward, in the sense that the type curve can be matched very closely to the drawdown data over the whole length of the test; T and S obtained from this curve are 99,700 igpd/ft and 1.3×10^{-5}, respectively. The drawdown curve for P3 deviates very severely from the type curve at the end of the test and somewhat less between t = 25 and t = 400 minutes; the deviation towards the latter part of the test is most likely caused by insufficient correction of the data for a drop in the nonpumping level caused by the lowering of the reservoir level. As the latter is almost linear with time, its effect is felt more severely towards the end of the test, where the time increments between successive observations are in the order of several hours. T and S determined from the drawdown curve of P3 are 131,000 igpd/ft and 1.3×10^{-4}, respectively.

The dashed lines shown alongside the drawdown curves for P3 and P5 are the theoretical drawdown curves that would occur in a model rectangular aquifer of the same width, with homogeneous T and S and with an impermeable boundary at 24,000 feet (top curve) and 36,000 feet (bottom curve) from the constant head boundary at Waterton Reservoir. These curves were obtained by the ADI (Alternating Direction Implicit) finite difference method which is briefly discussed in the section on hydraulic potential distribution and groundwater flow in the bedrock on page 51.

For the transmissibility of the model aquifer the value of 143,000 igpd/ft was used, as derived in the discussion on the average gradient of the piezometric surface in the aquifer; the storage coefficient used was 1.3×10^{-4}.

Except for the first 70 minutes the fit between the actual drawdown and the drawdown predicted by either model is poor; the good fit for the early portion of the curve for P3 indicates that the storage coefficient and transmissibility are correct; however, the less than predicted drawdowns actually observed during the later part of the test suggest that the actual constant head boundary is closer to P3 than was assumed in the model.

The drawdown curve for P5 shows deviations from the matched curve in the interval between 25 and 100 minutes and again between 1,600 minutes and the end of the test at 4,200 minutes; the latter is probably caused by insufficient correction for the falling nonpumping level. The early portion of the curve is badly scattered, probably because the drawdown values are very small and hard to measure accurately, and, therefore, they have been given less weight in the fitting of the type curve. T and S determined from the shown type curve fit are 104,000 igpd/ft and 8.6×10^{-5}, respectively.

The drawdown predicted from the model with an impermeable boundary at 36,000 feet coincides closely with the actual drawdown data; in fact the type curve solution is misleading in that it substitutes a lower apparent transmissibility for the impermeable boundary immediately south of P5. If early drawdown data had been of a better quality, the boundary would probably have been recognized as a definite break in the drawdown curve, and the transmissibility could have been calculated from the early part of the curve where the boundary had not yet a measurable effect. This danger of underestimating the permeability is clearly present in the interpretation of all drawdown data from wells located at a large distance from the pumped well, if boundaries are present at a short distance from the observation well, and if their presence is not known from other sources.

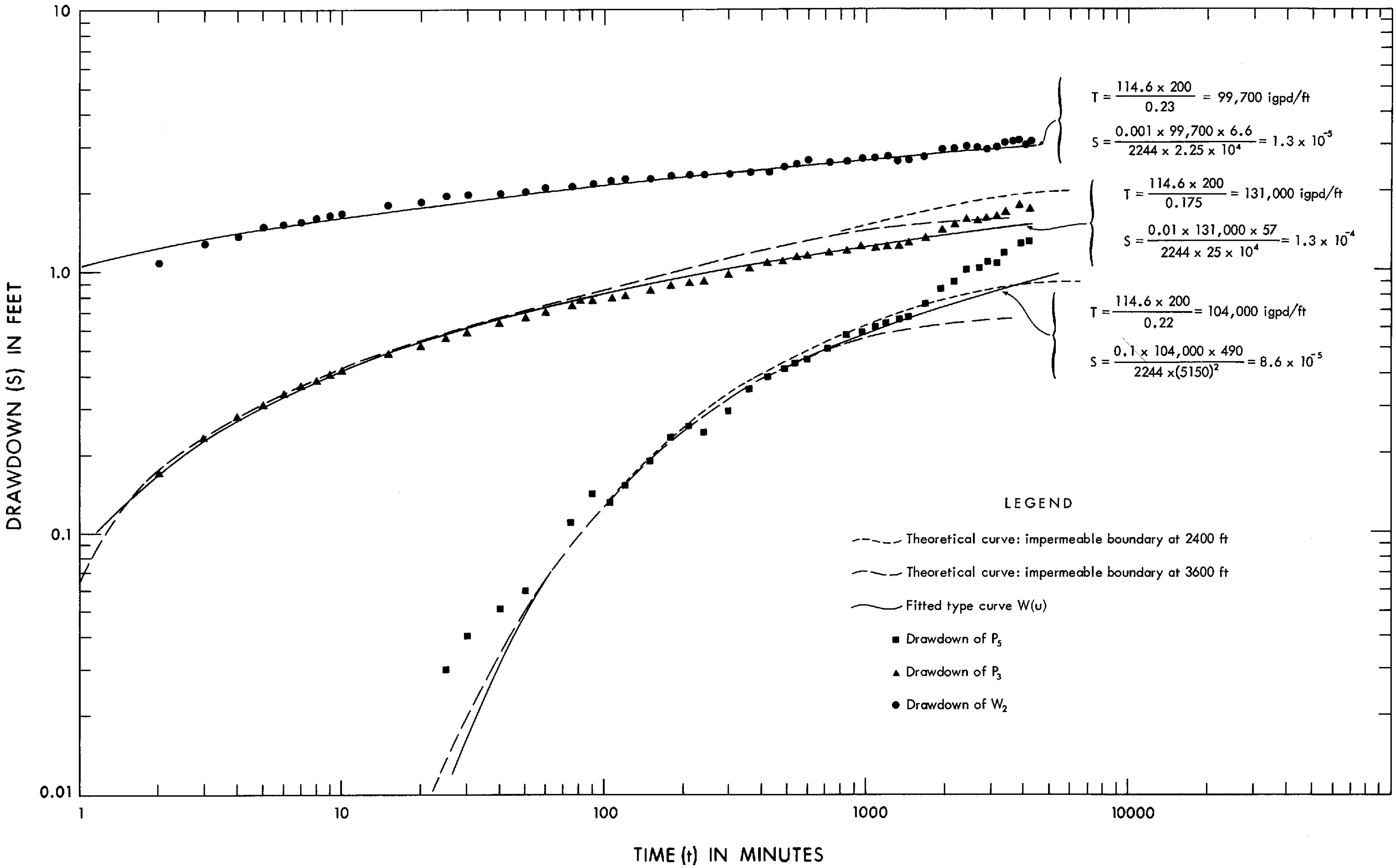


FIGURE 25. Time-drawdown curves of observation wells.

Whenever drawdown data from observation wells are analyzed with the Theis equation for infinite aquifers, and a boundary can be inferred from geologic data, it is advisable to make a quick check on the order of magnitude of the ratio:

$$\frac{\text{drawdown in the assumed infinite aquifer}}{\text{drawdown in the real bounded aquifer}}$$

for at least one point of the time-drawdown graph. When the boundary can be approximated by a straight line, boundary analysis can make use of the theory of images (Ferris *et al.*, 1962; Walton, 1962), which shows that the drawdown at any point in a semi-infinite aquifer, bounded by a straight impermeable boundary in one direction, is the same as the drawdown that would occur if the aquifer were infinite, and if each real pumping well were supplemented by an imaginary pumping well withdrawing water at the same rate as the real well and located at the image point of the real well with respect to the boundary.

Therefore, when the location of the boundary is known, the location of the image well is known, and the drawdown s_i caused in the observation well by the image well can be calculated for any time t, the ratio s_i/s_p, where s_p is the drawdown caused by the real pumping well only, is given by:

$$s_i/s_p = W(u_i)/W(u_p) \quad \ldots\ldots \quad (9)$$

where $u_i = 2{,}244\, r_i^2 S/Tt$, $u_p = 2{,}244\, r_p^2 S/Tt$, and r_i and r_p are the distance between the observation well and the image well, and between the observation well and the real well, respectively.

In the simple case where the real well, the image well, and the observation well lie on a straight line, a simple relationship exists between r_i and r_p (Fig. 26): let r_b be the distance from the boundary to the observation well or to the pumping well, whichever is the farther, and let the distance to the other well be ar_b, where $0 < a < 1$; then the distance between the observation well and the real well $= (1 - a)r_b$, and the distance between the observation well and the image well $= (1 + a)r_b$.

Substituting $u_b = 2{,}244\, r_b^2 S/Tt$ in (9), gives

$$s_i/s_p = \frac{W[u_b(1 + a)^2]}{W[u_b(1 - a)^2]} \quad \ldots\ldots \quad (10)$$

For values of $u < 0.01$ the function W(u) may be approximated by $W(u) = -0.5772 - \log_e u$; therefore, for values of $u_b(1 + a)^2 < 0.01$, or $u_b < 0.0025$ one may write

$$s_i/s_p = \frac{-05772 - \log_e u_b - 2\log_e(1 + a)}{-05772 - \log_e u_b - 2\log_e(1 - a)}$$

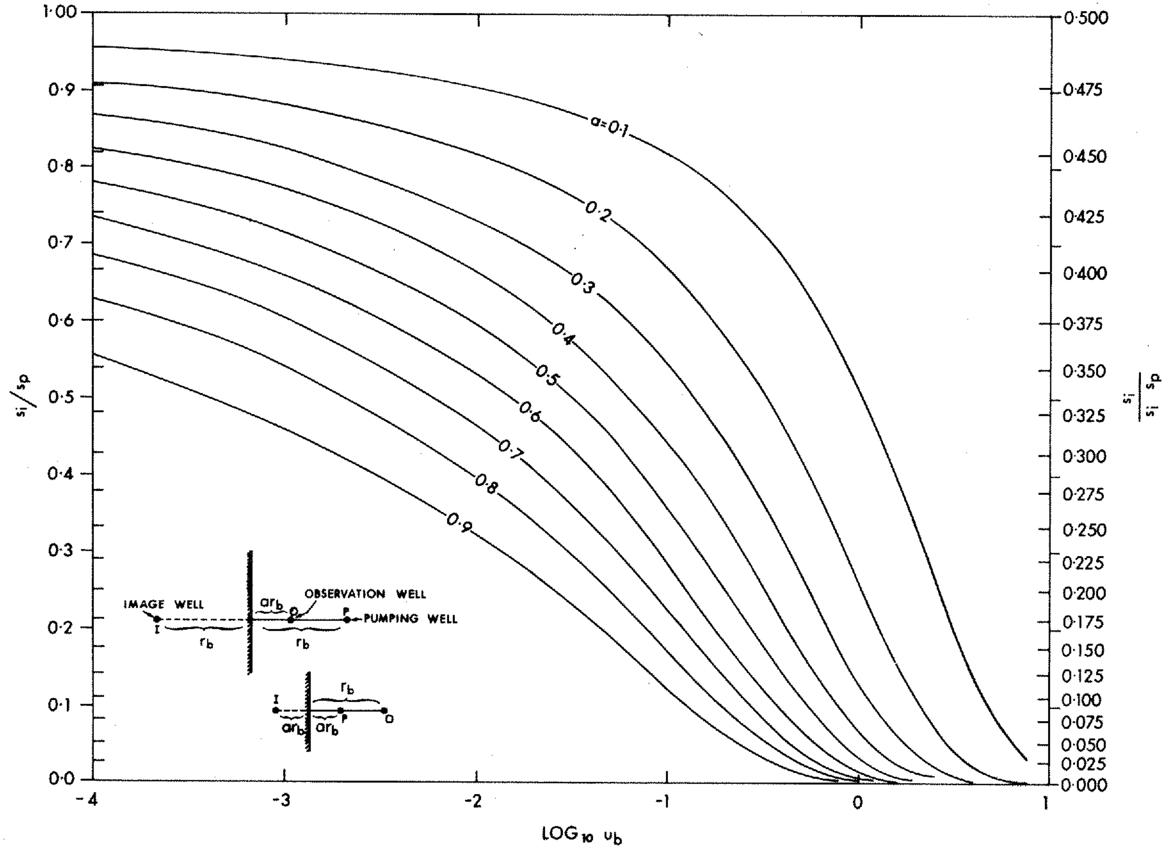


FIGURE 26. Values of s_i/s_p versus log u for different values of the ratio a.

or replacing the natural logarithm by Briggs' logarithms,

$$s_i/s_p = \frac{-0.2504 - \log_{10}u_b - 2\log_{10}(1+a)}{-0.2504 - \log_{10}u_b - 2\log_{10}(1-a)} \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots(11)$$

Equations (10) and (11) were used in preparing figure 26, which gives the ratio s_i/s_p against the logarithm of u_b, for various values of a.

The use of figure 26 in estimating the value of the component of drawdown due to the boundary still requires the calculation of u_b; a more rapid method is possible by calculating the time t_i at which the drawdown due to the image well equals the drawdown due to the real well, s_p, at time t_p. The ratio t_i/t_p is independent of time, S, T, and r_b, and therefore independent of u_b, for it follows from (7) and (8) that, if $s_p = s_i$, $t_i/t_p = (1+a)^2/(1-a)^2$. Figure 27 is a plot of t_i/t_p against a. For example, for observation well P5, a = 0.4 for each of the boundaries of the Cochrane Valley aquifer, and therefore $t_i/t_p = 5.4$. From figure 24, at t = 100 minutes, the drawdown s, which may be taken as a rough approximation to s_p, is 0.13 feet; therefore, at t_i = 540 minutes, s_i = 0.13 feet, and s_p is approximately 0.39 feet; s_i/s_p = 0.33, indicating that the boundary effect accounts for a sizeable portion of the total drawdown, and that the estimate of T will be too low.

Using T = 1.4 x 10^5 igpd/ft, S = 1.3 x 10^{-4}, t = 540 minutes, r_b = 10,000 feet, u_b is found to be 3.86 x 10^{-1} and $\log_{10}u_b$ = −041. From figure 26, s_i/s_p = 0.25, also indicating the necessity of considering the presence of the boundaries in the calculation of T and S.

For P3, a is approximately 0.95 and the ratio s_i/s_p is therefore very small, unless u becomes very small or t becomes very large.

An estimate of leakage from overlying or underlying confining beds can be made from pump tests provided the duration of the test is long enough, as the effect of leakage on the drawdown is insignificant during the early stages of a test. Hantush and Jacob (1955) describe the modifications to the theory of aquifer tests and the basic assumptions that are made to take leakage into account. They show that the drawdown s, is given by

$$s = (114.6\ Q/T)\ W(u,r/B) \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots (12)$$

where $B = Tb'/K'$, the leakage factor .... (13)

$W(u,r/B)$ = the well function for leaky aquifers,

K' = the vertical permeability of the leaking bed,

b' = the thickness of the leaking bed.

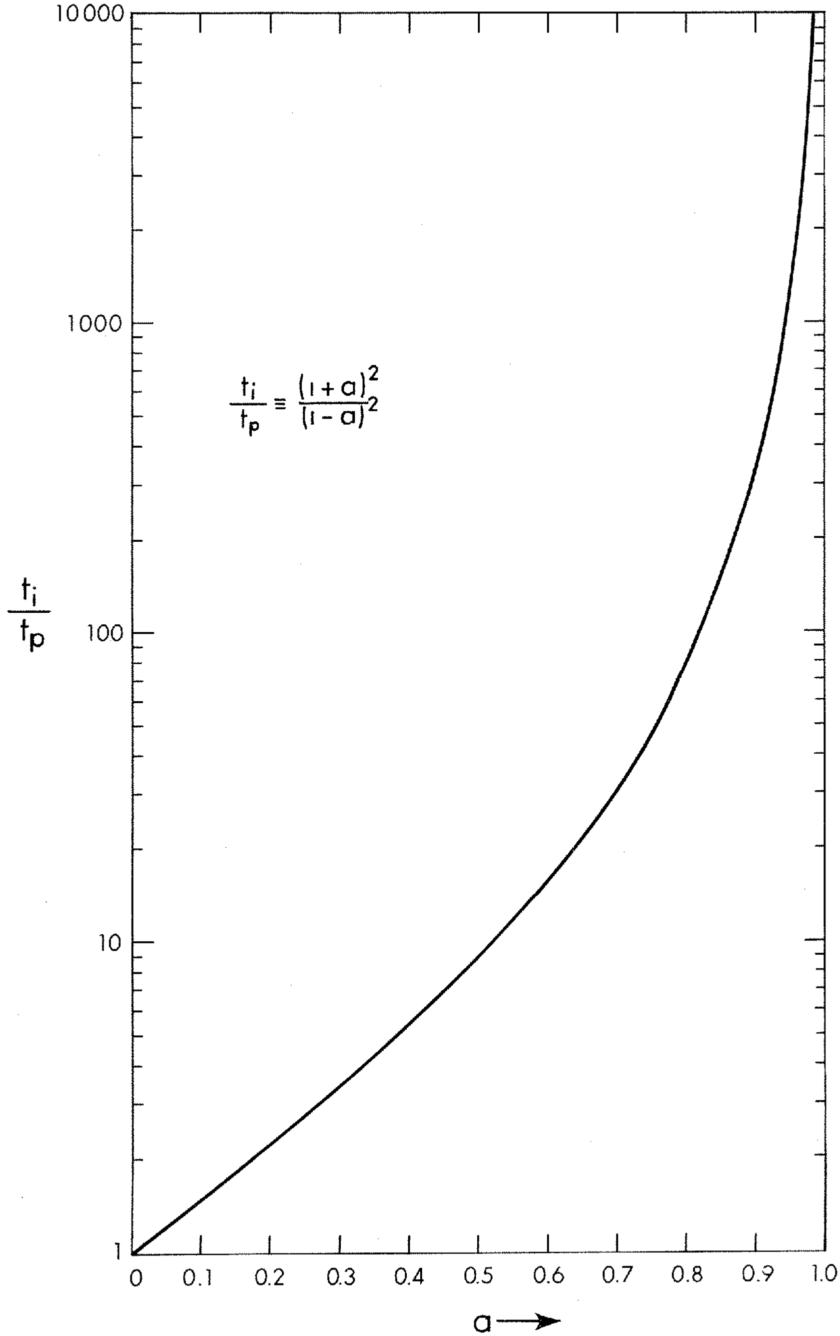


FIGURE 27. Values of t_i/t_p versus a.

Table 12. Limiting Values for r/B, B, and K

Observation Well No.	r (feet)	Maximum r/B (feet)	Minimum B	Maximum $K' = b'\ T/B^2$ (igpd/ft^2)
W2	150	0.005	3×10^4	0.013
P3	500	0.025	2×10^4	0.03
P5	5000	0.3	1.7×10^4	0.04

For large B, leakage is small, and W(u,r/B) tends to W(u). The family of curves W(u,r/B) is given by Walton (1962).

Although none of the three drawdown curves of figure 25 show the deviation from the type curve of W(u) which is characteristic of leakage, the method may still be used to estimate the maximum value of r/B or the minimum value of B, on the assumption that the deviation would have become noticeable at the time of the end of the test. Knowing the transmissibility of the aquifer, which is approximately 120,000 igpd/ft, and the thickness of the overlying leaking bed of approximately 100 feet, equation (13) can be used to find an upper limit for the vertical permeability K′ of the leaking bed (Table 12).

The distance drawdown curves for t = 60,600, and 1,680 minutes are shown in figure 28, on which the logarithm of drawdown is plotted against the logarithm of $1/r^2$. As in the case of the time drawdown curve, the three points for identical t should fall along the type curve of W(u), provided the aquifer is homogeneous and of infinite areal extent; from matching position the transmissibility and storage coefficient can then be calculated from equations (7) and (8). Because of the existing boundaries and inhomogeneities of the Cochrane Valley aquifer, the three points do not fall along the type curve. By matching pairs of points with the type curve in all possible combinations, 9 pairs of values for T and S were calculated (Table 13), and 3 pairs of values obtained from a "best" fit through all three points.

The fit through the three points gives values in reasonable agreement with those obtained from the time drawdown curves; while the pair W2, P5 yields values for T which are slightly lower, the pair W2, P3 yields values for T which are substantially lower and values for S which are higher; and for the pair P3, P5 the value of T is high.

Although there is a certain amount of variability in the values for the storage coefficient and the transmissibility of the aquifer, this should not be regarded as too surprising considering the known variation in the thickness of the aquifer and the curvature of the boundaries, as well as the fact that the boundary locations are

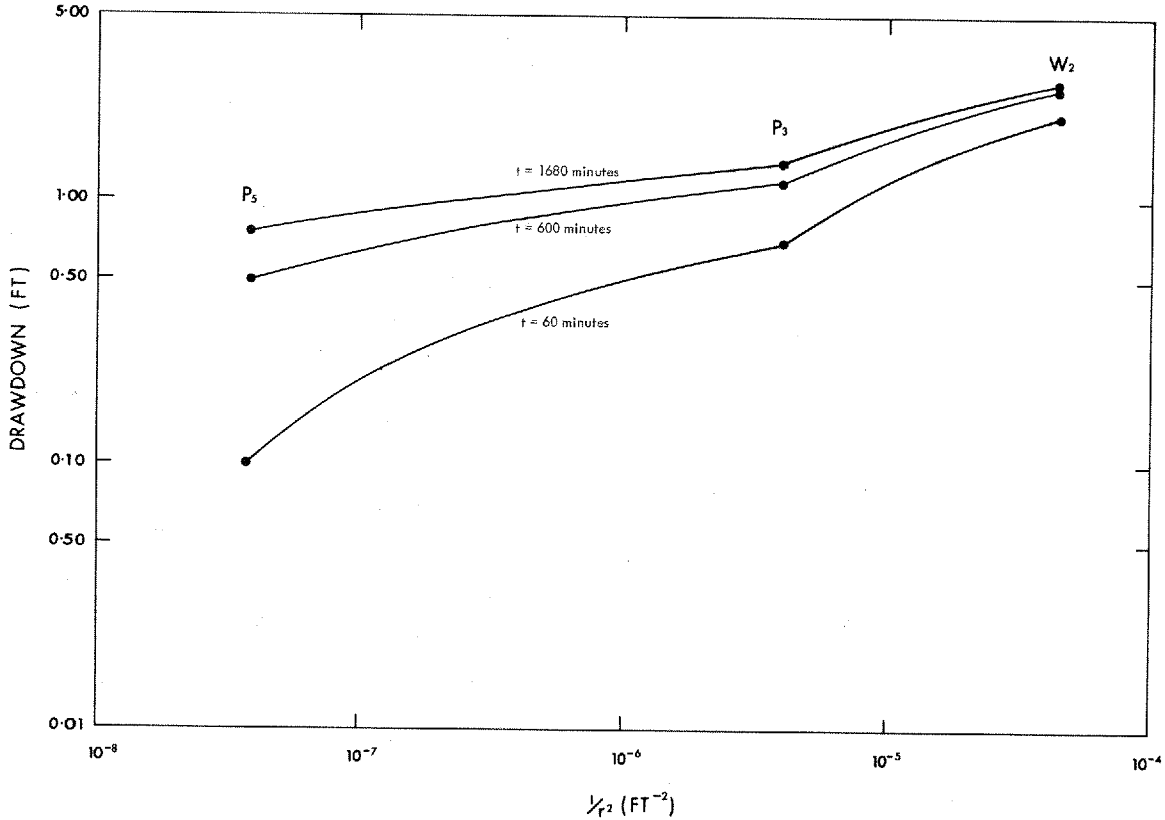


FIGURE 28. Distance-drawdown curves of observation wells.

Table 13. Values of Transmissibility and Storage Coefficient Derived from Distance-Drawdown Curves

t (min) \ Well pair used in match	W2, P3		P3, P5		W2, P5		3-point fit	
	T (igpd/ft)	$S \times 10^4$	T (igpd/ft)	$S \times 10^4$	T (igpd/ft)	$S \times 10^4$	T (igpd/ft)	$S \times 10^4$
60	37,600	9.2	176,000	0.64	76,400	0.59	115,000	0.65
600	38,200	37.0	158,000	3.2	76,400	1.0	94,000	1.3
1680	41,000	65.0	171,000	1.3	81,800	0.98	135,000	0.12

not known with sufficient precision. Moreover, hydrologic boundaries as determined from pump test data seldom coincide with actual geologic boundaries, because such boundaries are never absolutely impermeable as is assumed in the mathematical model on which the theoretical drawdown curves are based. In view of these complicating factors, the results obtained can be considered valid, and may be summarized as follows:

$$100{,}000 < \text{Transmissibility (igpd/ft)} < 143{,}000$$

$$1.3 \times 10^{-5} < \text{Storage coefficient} < 1.3 \times 10^{-4}$$

$$2{,}000 < \text{Permeability (igpd/ft}^2\text{)} < 2{,}750$$

According to De Wiest (1965), the permeability of gravels ranges from 10^4 to 10^6 igpd/ft^2 and the permeability of clean sands (good aquifers) from 10 to 110^4 igpd/ft^2; these figures match the permeability of the Cochrane Valley aquifer reasonably well, as the latter can be classified as a poorly sorted, sandy gravel with characteristics between those of the two classes defined by De Wiest.

Leakage of Groundwater from Adjacent Beds and Outflow into Belly River

When the average gradients and the total measured outflows from the aquifer in 1966 and in 1969, that is, before and after the completion of the relief wells, are compared, it appears that although the average gradient in the aquifer has increased only by a factor of approximately 1.6, the total known outflow increased from 2,130 to 6,060 igpm, almost a threefold increase. However, before the completion of the relief wells, spring D3 virtually controlled the gradient in the whole aquifer, because then it was by far the largest outlet; but after the completion of the relief wells the flow from spring D3 hardly influences the gradient in the part of the aquifer between the reservoir and the relief wells, as it is masked by the relief wells. This, however, is the area where the measurements of the gradient were obtained. In comparing the gradients and flows, therefore, the flow from D3 should not be included in the total flow after the completion of the relief wells. Even so, the increase in flow was from 2,130 to 4,750 igpm, or more than a twofold increase. This seems to indicate that not all of the outflow from the wells and springs is supplied by inflow from the reservoir but that a substantial part is supplied from other sources, be it leakage from the overlying till or the underlying bedrock or both, or from changes in the storage within the aquifer itself.

From Darcy's law it follows that the inflow from the reservoir is proportional to the gradient, and because all other factors remain constant between 1966 and 1969, the total inflow from the lake can be estimated at $1.6 \times 2{,}130 = 3{,}400$ igpm, which means that approximately 1,350 igpm is derived from other sources.

The order of magnitude of changes in storage in the aquifer can be assessed as follows: from experience during the pump test it is known that the rate of change of the piezometric head in the aquifer in response to the normally occurring changes in the level of the reservoir is of the order of one foot per day; as the storage coefficient of the aquifer, which is approximately 10^{-4}, is the amount of water (in cubic feet) released per square foot of the aquifer surface when the piezometric head drops one foot, the rate of groundwater release from storage is $SA(dh/dt) = (10^{-4} \times 8 \times 5{,}300^2 \times 6.23 \times 1)/1{,}440 = 97$ igpm, where 6.23 is the factor for converting cubic feet to imperial gallons, and 1,440 the factor for converting days to minutes. Evidently changes in the storage in the aquifer account for only a small fraction of the rate sought.

If the difference between the rate of flow into and out of the aquifer is to be accounted for by the leakage from the overlying till and the bedrock, the order of magnitude of the vertical permeability, K′, of these beds can be found from Darcy's law (5) in which now:

Q = 4,750 – 3,400 = 1,350 igpm = 1,350 x 1,440 igpd;

A = twice the surface area of the aquifer, estimated at 8 square miles = $2 \times 8 \times 5{,}300^2$ square feet;

dh/ds = the vertical hydraulic gradient; if it is assumed that the piezometric head at the top of the overlying till bed, which is approximately 100 feet thick, is not affected by the cone of depression created by the pressure relief wells, and that the average drawdown in the aquifer is 20 feet, then the hydraulic gradient is 0.2 ft/ft.

Thus from Darcy's law (5),

$$K' = 1{,}350 \times 1{,}440/2 \times 8 \times 5{,}300^2 \times 0.2 = 0.02 \text{ igpd/ft}^2.$$

This compares favorably with the limiting value of K′ obtained from the pump test. De Wiest (1965) gives 0.047 igpd/ft^2 as the representative value for silty sandstone, and a range of 10^{-2} to 10 igpd/ft^2 for clayey sands; therefore, the assumption that the 1,350 igpd is derived from leakage from and through the adjacent beds seems justified.

From the foregoing discussion, it appears most unlikely that a significant amount of water other than the flow from spring D3 finds its way into the Belly River at present; the Cochrane Valley aquifer either wedges out to the east before it crosses the Belly River, or is effectively sealed off everywhere except at the location of D3.

Hydraulic Potential Distribution and Groundwater Flow in the Bedrock

In order to describe the groundwater flow and the hydraulic potential distribution such as in the subsurface of the Waterton Dam area, which has a complicated geologic structure and topography, a number of simplifying assumptions have to be made to arrive at a workable model. Tóth (1962, 1963) based all of his models on the following assumptions:

1) the area of interest is underlain by a horizontal, impermeable stratum;

2) the surface topography is sufficiently close to being linear that a two-dimensional model suffices;

3) the surface topography follows an infinitely repeating pattern of symmetrical elements, ensuring that no flow will occur across a vertical through any of the symmetry points;

4) the medium is homogeneous and isotropic with respect to its hydraulic properties;

5) the magnitude of the undulations of the surface is minor relative to the dimensions of the elements in which the verticals through the symmetry points divide the model, and the elements of the model can be considered rectangular;

6) the potential along the water table, which is practically identical with the surface, can be described as a simple mathematical function.

Under these assumptions a subsurface element between two successive symmetry points becomes a closed flow system, in which the hydraulic potential is subject to the Laplace equation $\delta h/\delta x^2 + \delta h/\delta z^2 = 0$, and the boundary conditions $\delta h/\delta z = 0$ along the base of the element, $\delta h/\delta x = 0$ along the vertical sides, and $h = h_s(x)$ along the top of the element. In these equations h is the hydraulic potential, x and z are the horizontal and vertical coordinates respectively, and $h_s(x)$ is the known function representing the height of the water table.

Tóth (1962) solved the boundary value problem for the uniformly sloping water table, and later (Tóth, 1963) for a uniformly sloping water table on which sinusoidal oscillations are superimposed.

Freeze (1966) and Pinder and Bredehoeft (1968) describe how the Laplace equation can be solved by means of numerical methods. Compared to Tóth's analytical solution they have the disadvantage that they do not give a general solution in the form of an equation; instead, each particular case has to be solved independently. On the other hand, some of the restrictions of Tóth's model – homogeneity and isotropy, replacement of the surface by a horizontal plane, simple functional representation of the value of the piezometric head along the surface – are removed, and the method can be expanded to three-dimensional models.

The numerical method that has been employed most successfully is the Alternating Direction Implicit (ADI) method (Pinder and Bredehoeft, 1968). It results in a relatively fast scheme for the digital computer to calculate the piezometric head at a number of equally spaced nodal points in the model, and can be used equally well for the solution of Laplace's equation and of Poisson's equation for time-dependent problems such as are encountered in well and aquifer evaluation.

The ADI method was employed to find the potential distribution in a simplified model of the Waterton Dam area along an east-west cross section (profile A–A′ in figure 5). The model corresponds to the Tóth model in the following points:

1) linear surface topography, justifying a two-dimensional model;

2) an infinitely repeating symmetrical pattern of the topography, of which A–A′ is an element;

3) the undulations of the surface are minor.

Further assumptions are as follows:

1) the Wapiabi Formation is assumed impermeable, and everywhere to form an impermeable base of the closed flow system underlying profile A–A′;

2) the relative permeabilities of the formations overlying the Wapiabi are assigned as follows:

St. Mary River Formation	–	100
Bearpaw Formation	–	0.1
Belly River Formation	–	20

Figure 6(b) shows the potential distribution and some flow lines. In the west half of the model the groundwater has a noticeable tendency to move back to the surface, well before the midline as defined by Tóth (1962), due to the rise of the impermeable lower boundary; the strong upward movement under Waterton River is basically in agreement with Tóth's models, but is modified by the presence of the low-permeability section of Bearpaw Formation near the surface, causing the groundwater to flow predominantly through the adjacent, more permeable Belly River strata. The large segment of St. Mary River beds at the east end of the profile acts as a smaller, practically closed system within the larger system, as it is almost surrounded by the low-permeability sediments of the Bearpaw Formation.

SUMMARY

This report is an attempt to interpret data on water levels and chemistry of groundwater which were collected prior to, during, and after the sudden rise in the level of surface water behind Waterton Dam during the spring of 1965. Basically, the answers were sought to two questions:

1) Is the amount of water that finds its way through the Cochrane Valley aquifer from Waterton Reservoir into the Belly River significant in comparison with the total capacity of the reservoir?

2) Is there a noticeable influence due to the creation of the freshwater reservoir on the chemistry of the groundwater in the area?

In answer to the first question, the study seems to indicate that, apart from the spring on the Lenz property in the Belly River valley, no significant natural discharge takes place either through underground or surface outflow; this result follows from the measurement of the average hydraulic gradient in the aquifer and the permeability determined during the pump test in 1969. It seems to follow, furthermore, that the 6,100 igpm total measurable outflow from the aquifer in 1969, after the big relief wells were installed, is not derived from inflow from the reservoir alone, but that 45 per cent of this amount is derived from leakage through the adjacent bedrock and till. Provided this assumption is correct, the permeability of the adjacent beds is in the order of 0.01 igpd/ft^2.

On a first analysis of the data, the answer to the second question seemed to be positive, as the comparison between 1965 and 1967 analyses indicated that the average groundwater of 1967 was fresher than in 1965, containing less total solids, less per cent sodium, and more per cent carbonate plus bicarbonate. On further analysis, however, it appeared that the most significant change to a fresher water had occurred in those samples obtained from sources above the highest level of Waterton Reservoir; from this the conclusion has to be drawn that the observed change to a fresher groundwater is primarily caused not by the reservoir but by fluctuations in annual precipitation, which conclusion is substantiated by precipitation records from two nearby stations. Secondly, calculations based on the observed gradient of the piezometric surface and the measured throughflow indicate that even in the highly permeable deposits of the Cochrane Valley aquifer the fresh water could have penetrated only approximately one mile between 1965 and 1967.

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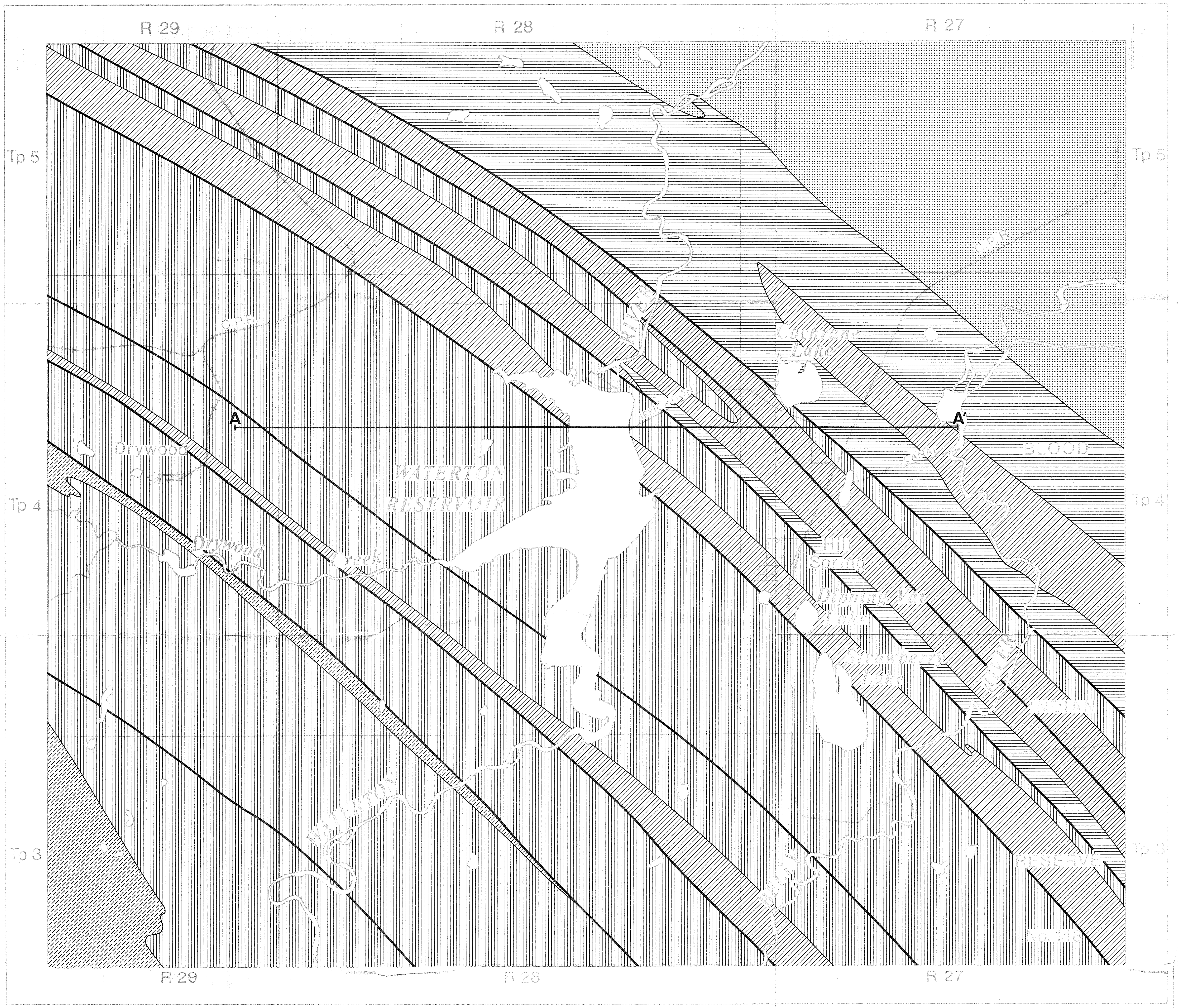

LEGEND

Willow Creek Formation

St. Mary River Formation

Bearpaw Formation

Belly River Formation

Wapiabi Formation

Fault

Formation boundary

A A' Line of cross section

Flow lines

Equipotential lines

SCALE IN MILES

1 0 1 2 3

FIGURE 6(a). BEDROCK GEOLOGY

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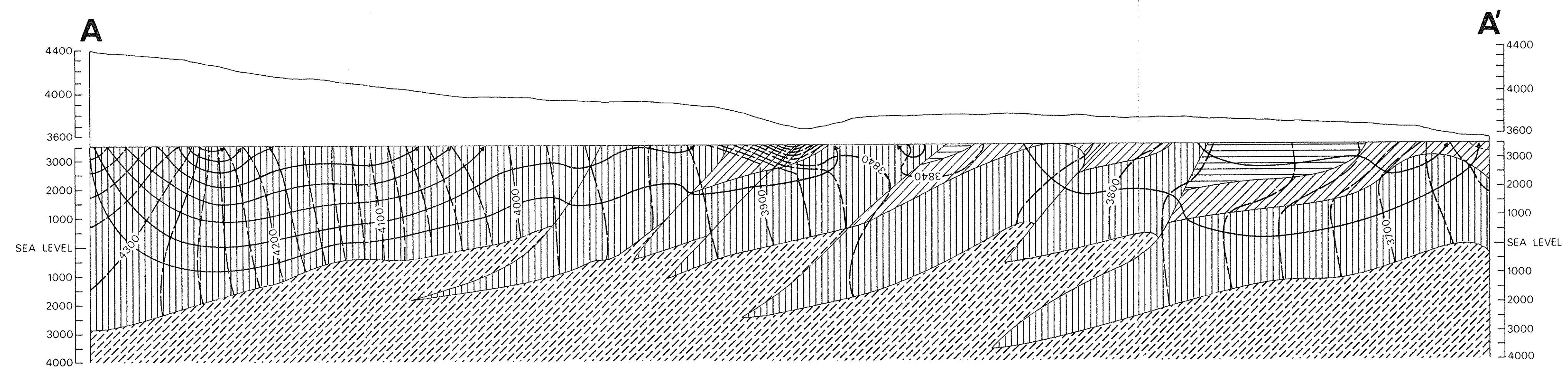

FIGURE 6(b). CROSS SECTION AND FLOW DISTRIBUTION

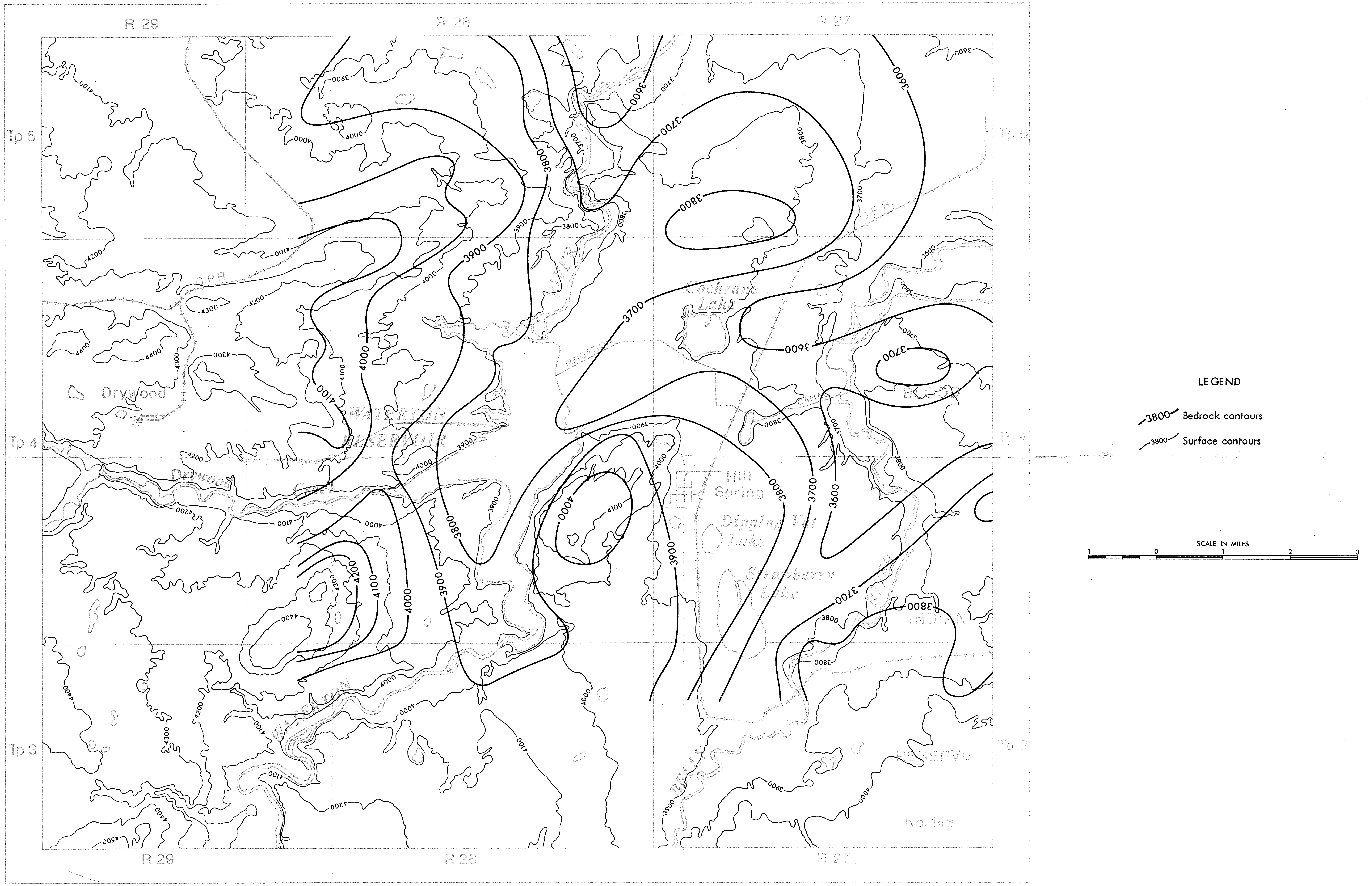

FIGURE 5. RELIEF, DRAINAGE AND BEDROCK TOPOGRAPHY

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